Glacial geology and chronology of Bishop Creek and vicinity, eastern Sierra Nevada, California

Fred M. Phillips[1,†], Marek Zreda[2], Mitchell A. Plummer[1,§], David Elmore[3], and Douglas H. Clark[4]
[1]*Earth & Environmental Science Department, New Mexico Tech, Socorro, New Mexico 87801, USA*
[2]*Department of Hydrology & Water Resources, University of Arizona, Tucson, Arizona 85712, USA*
[3]*Purdue Rare Isotope Measurement Laboratory, Physics Department, Purdue University, West Lafayette, Indiana 47907-1396, USA*
[4]*Department of Geology, Western Washington University, Bellingham, Washington 98225-9080, USA*

ABSTRACT

The valley of Bishop Creek, which drains part of the eastern flank of the Sierra Nevada, California, contains an unusually well-preserved set of middle to late Quaternary moraines. These deposits have been mapped by previous investigators, but they have not been quantitatively dated. We used the accumulation of cosmogenic ^{36}Cl to assign a chronology to the maximal glacial positions mapped in the valley. Our results indicate that the terminal moraines mapped by previous investigators as Tahoe were all deposited between ca. 165 and ca. 135 ka, during marine isotope stage (MIS) 6. Moraines mapped as Tioga were deposited between 28 and 14 ka, during MIS 2. These can be subdivided into Tioga 1 (28–24 ka), Tioga 3 (18.5–17.0 ka), and Tioga 4 (16.0–14.5 ka) advances (no moraines dated to Tioga 2 [21–19 ka] were found, presumably because the Tioga 3 advance either overrode or fluvially eroded them). At 15.0–14.5 ka, the Tioga 4 glacier retreated abruptly to the crest of the range. This was followed by the brief and fairly minor Recess Peak advance at ca. 13.4 ka. No Holocene advances extended beyond the very restricted limits of ice during the Matthes (Little Ice Age) advance. All preserved terminal moraines at lower elevations were deposited during either the Tahoe or Tioga stades. The Tahoe terminal moraines are extensive and voluminous, whereas the Tioga moraines are relatively narrow and have small volumes. However, this notable difference may be more a result of idiosyncrasies in the local glacial history than the result of differences in the length or intensity of glaciation between the two glacial episodes. The history of glacial advances at Bishop Creek exhibits a strong correspondence to global climate cycles, and to paleoclimate events in the North Atlantic in particular.

INTRODUCTION

Along the eastern escarpment of the central Sierra Nevada, the mouths of virtually all large canyons are distinguished by impressive sets of moraines. These deposits constitute a record of major climatic events in the region and have been the subject of numerous geological investigations over the past 100 yr. Multiple glaciations of the eastern Sierra Nevada were first recognized by Russell (1889). Knopf (1918) subsequently mapped glacial deposits of two ages in the Owens Valley region. The standard classification of glacial deposits in the region was laid out by Blackwelder (1931). In conformity with the conceptual framework for Pleistocene glaciations of his time, he proposed four glacial stages (from oldest to youngest): McGee, Sherwin, Tahoe, and Tioga. The McGee glaciation is probably Pliocene or early Pleistocene in age (Huber, 1981) and will not be discussed in this study. The type Sherwin deposits date to ca. 800 ka, based on their stratigraphic relationship (Sharp, 1968) with the well-dated Bishop Tuff (Izett and Obradovich, 1991; Sarna-Wojcicki et al., 2000). Some glacial deposits in our study area may correlate with the McGee glaciation (Bateman, 1965), but our methods cannot confirm this correlation. This study will therefore focus on evaluating the significance of Blackwelder's Tahoe and Tioga designations in this study area.

Blackwelder (1931) did not propose a quantitative chronology for his glacial sequence because numerical dating of geological materials was in its infancy when he published his study. However, he did tentatively correlate the Tioga glaciation with the midcontinental Wisconsin glaciation, the Tahoe with the Iowan (now early Wisconsin), and the Sherwin with the now-abandoned Kansan. Subsequently, Sharp and Birman (1963) proposed two additions: the Tenaya (between the Tioga and Tahoe) and the Mono Basin (between the Tahoe and Sherwin). Burke and Birkeland (1979), however, argued that these new subdivisions were not actually distinguishable from the original classification of Blackwelder (1931), based on the semiquantitative relative-weathering parameters that were the principal criteria at that time. Birman (1964) further proposed three additional Holocene advances: the Hilgard (early Holocene or early neoglacial), the Recess Peak (late Holocene), and the Matthes (Little Ice Age). Clark (1976) and Clark and Gillespie (1997), however, demonstrated that the type Hilgard moraines are actually Tioga recessional deposits, and Clark and Gillespie (1997) showed that the Recess Peak advance occurred in the late Pleistocene shortly after retreat of the Tioga glaciers. The glacial stratigraphy of the eastern Sierra Nevada has been reviewed and critically evaluated by Warhaftig and Birman (1965), Porter et al. (1983), Fullerton (1986), Gillespie et al. (1999), Osborn and Bevis (2001), Clark et al. (2003), and Kaufman et al. (2004).

In 1955, the study of Quaternary glacial history was revolutionized by the work of Emiliani (1955) on oxygen isotopes in marine sediments. In contrast to the conceptual framework at the time of Blackwelder (1931), it is now generally accepted that there have been at least seven major global glacial episodes since the eruption of the Bishop Tuff at 759 ka (Sarna-Wojcicki et al., 2000), and that each major glaciation probably contained multiple advances and retreats (Imbrie et al., 1984; Shackleton, 2000). In the Sierra Nevada, as elsewhere, this new chronology is incompatible with the classical sequence of two to four glacial advances since the beginning of

†E-mail: phillips@nmt.edu.
§Current address: Idaho National Laboratory, P.O. Box 1625, Idaho Falls, Idaho 83415, USA.

GSA Bulletin; July/August 2009; v. 121; no. 7/8; p. 1013–1033; doi: 10.1130/B26271.1; 6 figures; 2 tables; Data Repository item 2009022.

the Brunhes chron. Gibbons et al. (1984) proposed that "undercounting" of mountain glacial advances can be attributed to "obliterative overlap": the tendency of younger glaciations that are occasionally more extensive to override and obliterate the evidence of older ones that are less extensive. Modern glacial chronology studies must address the question of the extent to which the classical relative classifications are compatible with contemporary numerical chronology.

In the 25 yr since the publication of the study by Burke and Birkeland (1979), there have been numerous advances in data and in methodology. Among the most notable of these are high-resolution lacustrine records recovered from Owens Lake (the present terminus of the Owens River, to which Bishop Creek is tributary) and the advent of quantitative surface exposure dating employing cosmogenic nuclides (e.g., Cerling and Craig, 1994; Gosse and Phillips, 2001). The paleoenvironmental reconstructions from Owens Lake sediments (Benson et al., 1996; Bischoff et al., 1997; Menking et al., 1997; Smith and Bischoff, 1997; Bischoff and Cummins, 2001; Phillips, 2008) indicate that there were indeed repeated significant glacial advances during each major glacial cycle. The cosmogenic surface exposure dating studies (Phillips et al., 1990, 1996a; Poreda and Cerling, 1995; James et al., 2002) have also supported numerous advances, even during a single glacial-interglacial cycle. Based in part on cosmogenic nuclide chronologies, Gillespie and Molnar (1995) proposed that mountain-glacier advances do not necessarily correspond, in magnitude at least, to the global glacial stages recorded in marine sediments.

In this study, we attempted to resolve some of these questions, specifically, how the classical relative-age classification relates to the quantitative chronology of the glacial deposits. Our main tool for addressing these questions is surface exposure dating using cosmogenic ^{36}Cl (Phillips et al., 1986). In addition, we attempted to address a number of outstanding questions that have been posed over the years regarding Sierra Nevada glacial history:

Relative Size of Tahoe and Tioga Moraines

Blackwelder (1931, p. 884) noted that the "glaciers of the Tioga epoch were smaller than their predecessors," estimating, for example, that the Tahoe moraines in the Bridgeport Valley had 50 times the volume of the Tioga moraines there. However, Gillespie et al. (1999) argued that that this apparent discrepancy in size was not as common as Blackwelder had indicated and that Blackwelder may have lumped Tioga deposits into his Tahoe classification, thus artificially diminishing the inferred volume of Tioga deposits and enhancing that of Tahoe deposits. Mapping by Bateman (1965) of the terminal moraines at Bishop Creek would appear to support Blackwelder's position. Following Gillespie et al. (1999), we offer three hypotheses to explain this disparity: (1) Deposits mapped as Tahoe moraines at Bishop Creek (and elsewhere) are not actually the deposits of a single glaciation, but are rather superposed moraines of several glaciations that cannot be distinguished by relative dating (i.e., partial obliterative overlap). (2) The apparent relative volume of the two deposits is real and can be explained by the Tioga glaciation being of significantly shorter duration than the Tahoe. (3) There may have been a substantially longer interval between the Tahoe and the preceding glaciation than between the Tioga and Tahoe glaciations, allowing for the production of a greater amount of weathered debris for transport to the terminal moraines. We tested these hypotheses by intensively sampling the mapped Tahoe subunits in order to determine whether they resulted from separate episodes of deposition.

Rate of Retreat of the Tioga Glaciers

Clark (1976), based on geomorphic evidence, hypothesized that the final retreat of the Tioga glaciers had been very rapid. We tested this hypothesis by detailed sampling of erratics and bedrock exposed during deglaciation in order to determine a chronology of retreat.

Age of the Recess Peak and "Hilgard" Glaciations

Birman (1964) mapped in detail the glacial deposits across the Sierra crest ~30 km north of our study area. Based on his observations, he proposed three late Holocene advances: Hilgard, Recess Peak, and Matthes (Little Ice age), in decreasing order of age. Birman's age assignments, and even the existence, of these advances have remained controversial. Burbank (1991), based on an analysis of the equilibrium line altitudes (ELA) of various moraines mapped as Recess Peak, surmised that they included moraines deposited during separate glacial advances of differing magnitude. Gillespie (1991) and Clark and Gillespie (1997), however, based on additional ELA analysis and radiocarbon dates on sediments in lakes behind Recess Peak moraines, concluded that the Recess Peak advance can be attributed to the late Pleistocene, dating to ca. 13 ka. With reference to Clark's (1976) work, they inferred the "Hilgard" moraines to be Tioga recessionals. We tested these findings by dating Recess Peak moraines and a continuous sequence of glacial features between the Recess Peak and Tioga moraines in order to isolate any possible intermediate-age advances.

STUDY AREA

Bishop Creek has the largest drainage basin on the eastern slope of the southern Sierra Nevada. Most of the streams drain directly eastward into the Owens River, but Bishop Creek has been forced into a northeasterly course by the Coyote Plateau to the east (Plate 1 in GSA Data Repository[1]; Figs 1 and 2) and hence collects runoff from a 35 km interval of the Sierra Nevada crest. As a result of this large collection area, the Pleistocene glaciers at Bishop Creek descended to an elevation of 1500 m, among the lowest in the eastern Sierra Nevada. At this elevation, an arid Great Basin climate prevails (150 mm annual precipitation at the nearby Bishop airport, compared to 440 mm at South Lake, 2780 m and about half the distance to the range crest). The Tahoe terminal moraine complex was isolated from glacial advances during the Last Glacial Maximum (Tioga glaciation) by an avulsion of the glacier course through the Tahoe right-lateral moraine. The combination of dry climate and diversion of Tioga ice has resulted in an unusually complete and uneroded sequence of Tahoe glacial deposits. The comparatively great length of the Bishop Creek glacier has also provided a large area from which to select samples to document the chronology of the maximum extent and retreat of the Tioga glaciation. This combination of unusually abundant and well-preserved glacial landforms motivated us to focus this cosmogenic surface exposure dating study on the Bishop Creek drainage.

Previous Investigations

The moraines at Bishop Creek were first described by Knopf (1918), who distinguished two glacial advances. He mapped old moraines on the northwest side of Birch Creek, and he placed particular emphasis on the downcutting of the creek subsequent to the deposition of the moraines (Plate 1 [see footnote 1]) as evidence of the antiquity of the older glaciation. He placed the boundary between the older and younger moraines at the locations of samples

[1]GSA Data Repository item 2009022, Table 1S, complete listing of data pertaining to ^{36}Cl samples, and Plate 1, and map of glacial and neotectonic geology of the Bishop Creek area, Inyo and Fresno Counties, eastern California, 1:37,000, is available at http://www.geosociety.org/pubs/ft2008.htm or by request to editing@geosociety.org.

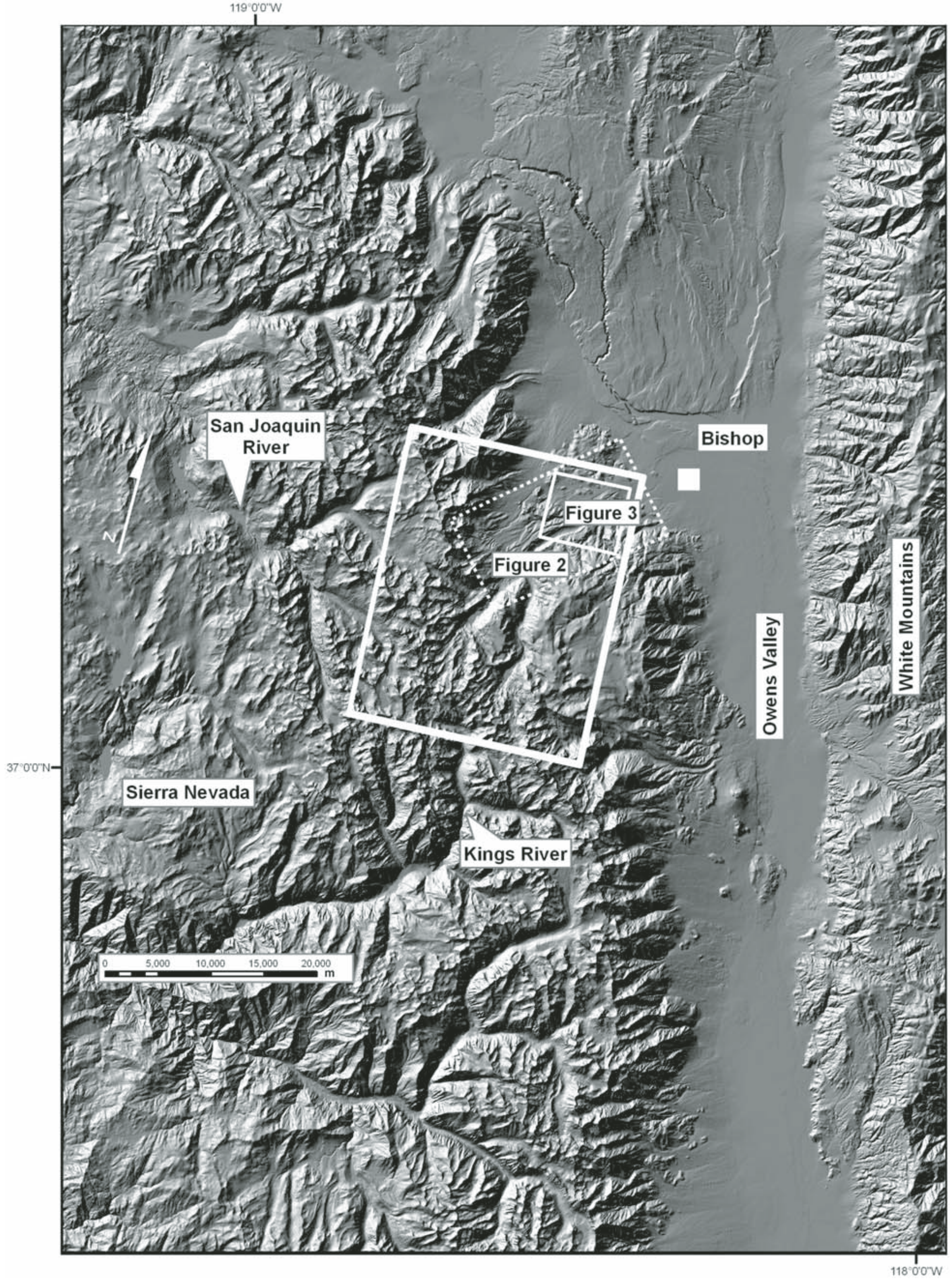

Figure 1. Location of the study area. The areas covered by other figures in this paper are indicated (Digital Elevation Model North American Datum 1983 ALBERS).

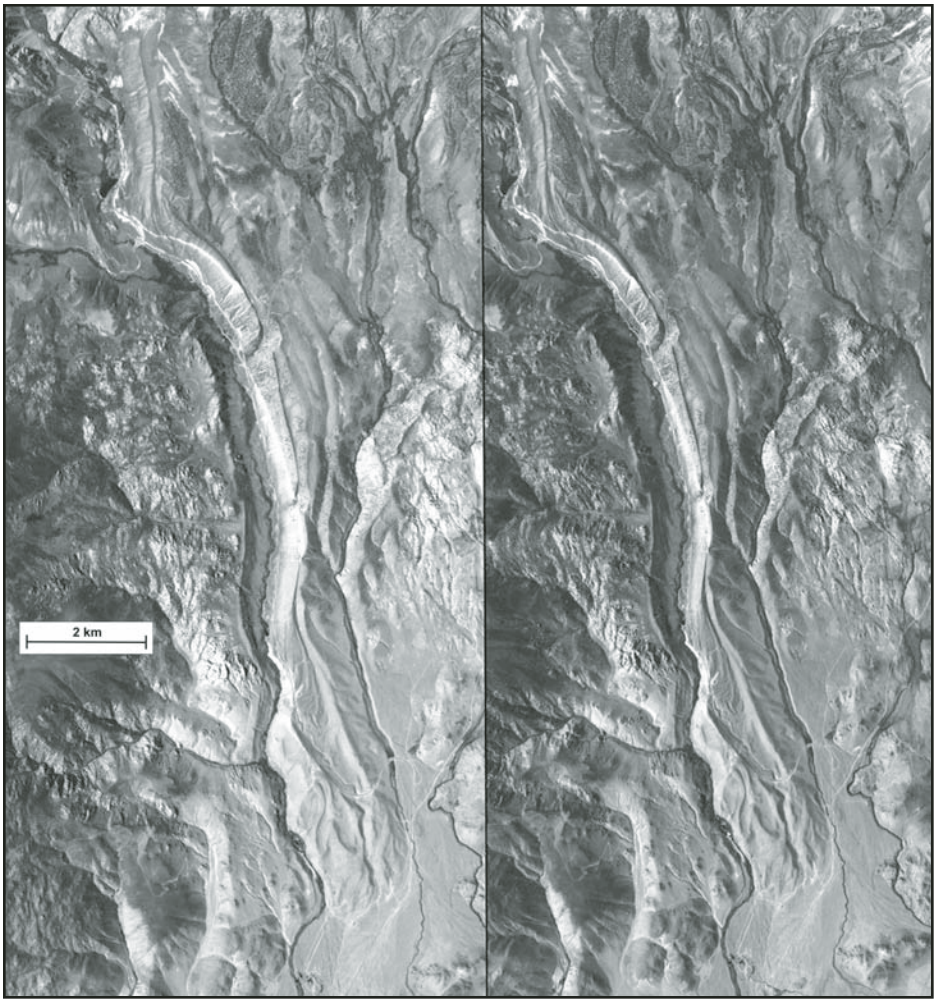

Figure 2. Vertical high-altitude stereopair photographs of the Bishop Creek terminal complex and valley up to the confluence of the North/Middle and South Forks. Photographs are oriented with west at the top. Photographs USAF 374V-174 and 374V-175 are from U.S. Geological Survey, Menlo Park, California, courtesy of M.M. Clark. The upper left corner of the left photo is at 37.225° N, 118.600° W and the lower right corner of the right photo is at 37.363° N, 118.493° W. The location of the stereo pair is shown on Figure 1.

90 through 94 in Figure 3. He also noted that the Tioga moraines from the Middle Fork of Bishop Creek cut across the mouth of the South Fork at the confluence of the two forks, ascribing this to the greater "alimentation" of the Middle/North Forks (i.e., greater accumulation area).

Blackwelder (1931) mentioned Bishop Creek as containing both Tioga and Tahoe moraines, but he did not publish any maps or figures delineating these deposits. Since he cites Knopf (1918) and does not raise any differences in interpretation, he presumably accepted Knopf's "older" and "younger" map units as Tahoe and Tioga moraines. Later, Rahm (1964) subdivided both the Tahoe and Tioga moraines into older and younger units, but did not publish a map showing these distinctions. He also named a "Wonder Lakes" advance that was probably equivalent to the Recess Peak, for which he estimated a Holocene age.

The first detailed glacial geology map of the area since Knopf (1918) was published by Bateman (1965) in the course of a comprehensive geological investigation of the Bishop area. Bateman distinguished moraines he termed Sherwin, Older Tahoe, Younger Tahoe, Older Tioga, and Younger Tioga. The "Sherwin" moraines consist of remnants of till lodged along the northwestern edge of the Birch Creek canyon, and correspond to the "Pliocene-Pleistocene till" map unit in Plate 1 (see footnote 1). Bateman's "Older Tahoe" and "Younger Tahoe" deposits generally, although not completely, correspond to our "older pre-Tioga" and "younger pre-Tioga" units in Plate 1. Bateman differed from Knopf (1918) in mapping Tioga deposits in the terminal moraine area as being more extensive. He placed the Tahoe-Tioga boundary about halfway down Sand Canyon and also down Bishop Creek past the confluence with Coyote Creek. Bateman's mapping of the "Older Tioga–Younger Tahoe" boundary corresponds closely to the Tioga 1–Tahoe boundary in Plate 1 (see footnote 1). Bateman's "Older Tioga" corresponds to our Tioga 1 and 3 combined, and his "Younger Tioga" corresponds to our Tioga 4.

An observation on basalt clasts in the moraines that is useful for identifying the source and age of till was made by Bateman (1965) (p. 150): "Patches of basalt exposed along the North Fork of Bishop Creek… probably are remnants of once extensive flows; older glacial till on the north side of Bishop Creek contains abundant boulders of basalt that must have come from this area… Inasmuch as adjacent moraines assigned to the Tahoe glacial stage contain no basalt boulders, the dissection of the basalt at North Lake must have been completed at an early time." We, however, have observed basalt clasts in both Tahoe and Tioga moraines, although the abundance decreases markedly with moraine age.

Work in the Bishop Creek drainage subsequent to Bateman (1965) has been limited and specialized. Sheridan (1971) published short descriptions of moraines in the terminal complex and some limited weathering data. Berry (1990, 1994, 1997) described soil development on moraines of various ages in the area north and west of the confluence of the Middle and South Forks of Bishop Creek. She was able to distinguish three relative age classifications corresponding to pre-Tahoe, Tahoe, and Tioga. Bach (1995) examined eolian modifications to boulder surfaces on the Tahoe terminal complex and related areas of ventifaction and dust deposition to past positions of the glacial terminus. Clark et al. (1994) and Clark and Gillespie (1997) mapped Recess Peak and Matthes moraines in the headwaters of Bishop Creek. Phillips et al. (1996a) briefly surveyed ^{36}Cl ages for Tioga moraines in the Bishop Creek area and compared them with other regional glacial chronologies and the sedimentary record from Owens Lake (Benson et al., 1996). They concluded that there appeared to be a correlation between Heinrich Events (Bond et al., 1992) and Sierra Nevada glacial advances. Phillips et al. (1996a) used values for ^{36}Cl production parameters that have now been superseded, so ages reported in this paper may differ from theirs. Plummer and Phillips (2003) developed and employed a numerical model of glacier energy–mass balance to help

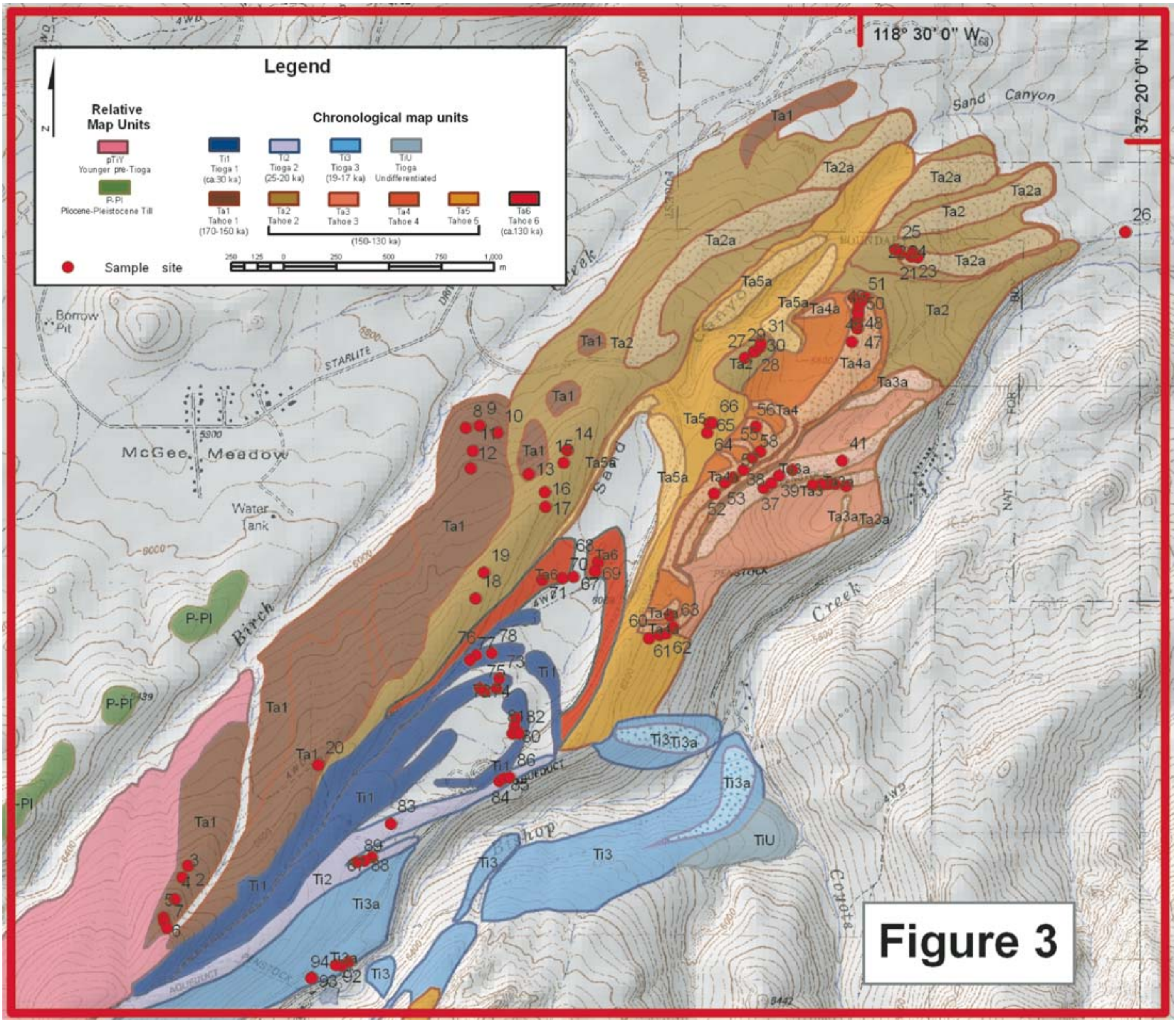

Figure 3. Chlorine-36 sample locations in the Bishop Creek terminal moraine area.

understand climate conditions in the Bishop Creek basin during glacial episodes.

METHODS

Moraine Mapping and Sampling

Prior to sampling for surface exposure dating, the glacial geology of the drainage basin was mapped. A simplified version of the mapping is shown in Plate 1 (see footnote 1). Mapping was based on moraine morphology observed in the field and on aerial photographs. Features on the map in Plate 1 can be visualized by use of the stereopair high-altitude photographs in Figure 2. The mapping initially distinguished all separable morphological features (Zreda, 1994). This initial mapping was used to guide the sampling program so as to address the hypotheses described in the introduction to this paper. Based on the results of the dating, the morphological units were then lumped into chronostratigraphic units.

For the purposes of this study, map units for which we had numerical age control were distinguished from those for which we had only relative age assignments. Those that could be directly dated, or reasonably securely correlated with dated moraines, were given names based on chronological assignment, as follows: 190–130 ka (i.e., MIS 6)—Tahoe; 75–60 ka (i.e., MIS 4)—Basin Mountain; ca. 30 ka—Tioga 1; 25–20 ka—Tioga 2; 19–17 ka—Tioga 3; 16–15 ka—Tioga 4; ca. 13 ka—Recess Peak; late Holocene—Matthes. (No glacial units that dated to MIS 4 were found in the Bishop Creek drainage, and they will not be discussed in this paper.) Pre-Tioga moraines or tills for which ages could be estimated based only on weathering/erosion characteristics and position relative to dated moraines were assigned names in a relative dating scheme referenced to the oldest dated moraine against which they are juxtaposed (i.e., moraines older than dated Tioga deposits are referred to as "pre-Tioga," and those older than dated Tahoe are referred to as "pre-Tahoe"). In the Bishop Creek area, most moraines mapped as "pre-Tioga younger" are probably correlative to the Tahoe (MIS 6). Those mapped as "pre-Tioga older" are likely younger than mid-Pleistocene but older than MIS 6. Finally, the patches of strongly eroded bouldery diamicton that were noted by previous investigators, as described above, have been mapped as "Pliocene-Pleistocene till." Based on comparison with old glacial deposits elsewhere in the Sierra Nevada, and local factors discussed later, they could range in age from late Pliocene to mid-Pleistocene.

After mapping, sampling sites were selected on the crest of each mapped moraine unit. Sampling sites were chosen on the basis of availability of suitable boulders and apparent stability of the moraine surface. Boulders within the site area were selected for sampling based on three criteria: their height and size, surface texture, and degree of weathering. Taller boulders are preferable because they are more likely to have projected above the original moraine surface, prior to erosional lowering. Random sampling at Bishop Creek has shown a strong correlation between boulder height and apparent cosmogenic exposure age (Zreda et al., 1994). Glacially smoothed or ventifacted surfaces were preferred for sampling, while spalled surfaces or those undergoing granular disintegration were avoided, unless no other boulders were available. Relatively fresh boulders were preferred over weathered ones. Virtually all boulders sampled were crystalline Sierra Nevada batholith rock, and a large majority was Lamark granodiorite. A much smaller number was made up of diorite or gabbro, and one sample was from a basaltic boulder. Samples of 300–500 g were chiseled from the top 2–5 cm of each boulder.

Chlorine-36 Processing and Analysis

Surface exposure dating using cosmogenic nuclides relies on the accumulation of rare radioactive or stable nuclides produced by reactions of cosmic-ray neutrons and muons with the nuclei of atoms in rocks at the surface of Earth (Gosse and Phillips, 2001). Glacial erosion brings rocks to the surface that have previously been shielded by many meters of rock or ice and exposes them to cosmic radiation on the tops of moraines, at which point the cosmogenic "clock is set," and the boulders begin to accumulate cosmogenic ^{36}Cl and other nuclides. The rock samples are ground and dissolved in a mixture of hydrofluoric and nitric acid, and the chloride that is liberated is precipitated as AgCl by the addition of $AgNO_3$. The ratio of ^{36}Cl to total stable Cl of the AgCl is measured by accelerator mass spectrometry (AMS) (Elmore et al., 1979; Finkel and Suter, 1993). Based on the ^{36}Cl/Cl ratio and the measured Cl concentration, the concentration of ^{36}Cl can be determined, and from this, the exposure age can be calculated (Zreda and Phillips, 1994).

This paper reports ^{36}Cl exposure ages calculated from data obtained over the span of a decade. During this period, we implemented significant advances in analytical methodology. The most important of these is with regard to the determination of the Cl concentration of the rock samples. Until 1994, the Cl concentration was measured on a separate aliquot of rock powder by ion-specific electrode (ISE) in a Teflon diffusion cell apparatus, according to the procedure of Aruscavage and Campbell (1983) and Elsheimer (1987). After 1994, we employed isotope-dilution mass spectrometry (ID-MS). After the rock is dissolved in acid, a weighed spike of 99% ^{35}Cl is added to the solution. During the AMS analysis, both the $^{36}Cl/^{35}Cl$ and the $^{35}Cl/^{37}Cl$ ratios are measured. Based on the mass of spike added, the known spike and natural $^{35}Cl/^{37}Cl$ ratios, and the measured sample ratios, the ^{36}Cl/Cl and Cl concentrations can be calculated. ID-MS has many advantages over the ISE method, particularly for samples with low Cl concentration.

Reproducibility of the ISE method is mediocre (~20%) for samples with low Cl concentration (<15 ppm), unless very large numbers of replicate analyses are performed, and this propagates directly into the ^{36}Cl calculation. Furthermore, by comparing analyses on comparable samples using both methods, we have observed that there sometimes appears to be a bias toward erroneously low calculated ^{36}Cl concentrations, and thus young ages, for low-Cl samples measured by ISE. Given the improvement in analytical methodology, more confidence should be placed in the post-1994 results, especially for low-Cl samples.

Calculation of the cosmogenic ^{36}Cl production rate requires a fairly complete chemical analysis of the samples. Major elements were analyzed by X-ray fluorescence (XRF). The concentrations of Gd and B (which are significant competitors with Cl for absorption of low-energy neutrons) were measured by prompt-gamma emission spectrometry and U and Th (which produce neutrons that result in a small background concentration of ^{36}Cl) by XRF. Prior to 1995, Gd and B were measured on only ~20% of the samples, and U and Th were not measured. After 1995, these elements were measured on all samples. Values of these elements were estimated for pre-1995 samples in which they were not measured based on lithology-dependent averages of the post-1995 data. Complete chemical data can be accessed in the GSA Data Repository (see footnote 1)

Ages were calculated using the program CHLOE (chlorine-36 exposure age) (Phillips and Plummer, 1996). This version of CHLOE employed the thermal and epithermal neutron distribution equations of Phillips et al. (2001) and production of ^{36}Cl by muons according to Stone et al. (1998). The ^{36}Cl production parameters of Phillips et al. (1996b) were used, as corrected by Phillips et al. (2001) for improved neutron distribution equations and the incorporation of production from muons. The values of the three critical production parameters were 66.8 atoms ^{36}Cl (g Ca)$^{-1}$ yr^{-1}, 154 atoms ^{36}Cl (g K)$^{-1}$ yr^{-1}, and 626 epithermal neutrons (g air)$^{-1}$ yr^{-1}. Chlorine-36 production rates were scaled for elevation and latitude according to Lal (1991). Ages were corrected for snow shielding, shielding by surrounding topography, and effects of nonhorizontal surfaces. Snow-shielding calculations were based on snow-survey data from the California Department of Water Resources Snow Course Data Archive (CDWR, 2006). These shielding corrections are tabulated in the GSA Data Repository (see footnote 1). All analytical uncertainties are reported as plus-or-minus one standard deviation and incorporate only the reported analytical uncertainty in the ^{36}Cl measurement. Consideration of all sources of uncertainty would probably result in 10% to 15% standard deviations (Phillips et al., 1996b), but the magnitude of systematic uncertainties has not been quantified sufficiently to specify them for individual samples. A more complete description of the dating methodology and shielding calculations can be found in Gosse and Phillips (2001).

Parameterization of the production reactions for ^{36}Cl has proved to be more difficult than most other terrestrial cosmogenic nuclides due to the multiple production reactions (Gosse and Phillips, 2001). The most widely used production constants are those published by Phillips et al. (2001) and the set from Stone et al. (1996a, 1996b). Swanson and Caffee (2001) have published a third alternative parameter set. The production constants estimated by Phillips et al. (2001) were calibrated in the same region as this study, and many of the calibration samples were in the same general age range. This reduces the likelihood that the production rates contain biases due to inadequate spatial and temporal scaling corrections. In order to evaluate the magnitude and direction of the systematic uncertainties indicated by these alternative parameterizations, we recalculated our ^{36}Cl ages using the values given in these two alternative production-rate sets, and we compared the results of all the production parameterizations with independent chronological constraints.

RESULTS

Sample locations and ^{36}Cl ages are given in Table 1. (Complete data on sample chemistry, $^{36}Cl/Cl$ ratios, and calculated ages are given in Table 1S in the GSA Data Repository [see footnote 1].) Samples are grouped according to unit or geomorphic setting. Sample numbers are keyed to Plate 1 (see footnote 1). For purposes of visual clarity, the sample locations in the Bishop Creek terminal moraine area are shown in larger scale in Figure 3. The age distributions are graphically presented in Figures 4 and 5. The inventory of any in situ cosmogenic nuclide in a sample is a function of the time of exposure at the surface of Earth, the rate of erosion (or, in cases where eolian or other deposition dominates, accretion) of the surface, and the half-life of the nuclide. We do not independently know the erosion rate of each rock surface, and we have therefore calculated apparent ages based on a range of assumed erosion rates: no erosion, 1.1 mm/k.y., and 3.3 mm/k.y. We believe that the zero erosion and 3.3 mm/k.y. rates bound the actual erosion values in nearly all cases, and that most boulders sampled are probably in the range 0 to 1.1 mm/k.y. Several lines of evidence support this range of erosion rates. One is that in areas where bedrock polished by Tioga glaciers is now exposed, patches of polished granite are frequently observed, surrounded by rough areas of flaked-off surface. The height differential between the preserved polish (virtually zero erosion) and the weathered surfaces is generally less than 10 mm. Given 15 k.y. since the retreat of the Tioga glaciers, this indicates rock-surface erosion rates of ~0.7 mm/k.y. or less. Surfaces of granite inselbergs close to the floor of the Owens Valley (which are weathered and thus eroding faster than fresh glacial boulders) yielded erosion rates based on ^{10}Be accumulation of 5.4 ± 2.7 mm/k.y. (Nichols et al., 2006), while similarly located boulders gave an erosion rate of <0.4 mm/k.y. based on ^{21}Ne accumulation (Bierman and Gillespie, 1991a). Studies of boulder weathering over both a short time scale in an environment in the Rocky Mountains (Benedict, 1993) similar to Sierra Nevada moraines and over the 10^4 yr time scale in Lapland (André, 1996) estimated boulder-surface erosion rates of <1.5 mm/k.y. Using the combined $^{36}Cl/^{10}Be$ method on glacial boulders from the Wind River Basin, Phillips et al. (1997) obtained erosion rates less than 0.2 mm/k.y. for most samples, including those on Illinoian-age moraines. Thus, we believe that for most samples, the best estimate of the exposure age lies between 0 and 1.1 mm/k.y., and 3.3 mm/k.y. provides a probable upper limit. Unless otherwise stated, ^{36}Cl ages in this paper are calculated for an assumed 1.1 mm/k.y. erosion rate. We note that these surface lowering rates are intended to be applied to large granitic boulders and are not appropriate for soils, weathered bedrock, or for bedrock under soil. Furthermore, although we observed that mass loss from the surface of Tioga-age boulders was almost entirely by granular disintegration, many Tahoe-age boulders showed evidence of loss by spalling of thin surficial layers (probably most commonly attributable to fire spalling; Bierman and Gillespie, 1991b), which probably results in a higher loss rate.

Effects of Moraine Erosion

The boulders on some portions of the Tahoe moraines show spreads in exposure age approaching 100 k.y. This is much greater than the variation that would be expected due solely to variable rock-surface erosion rates of between 0 and >3 mm/k.y., which are typical for surfaces with ages on the order of 15 ka. The boulder age distribution exhibits a strong tailing toward young ages. It is well established that this variation largely arises from progressive exposure of boulders during erosion of the unconsolidated moraine (Hallet and Putkonen, 1994; Zreda et al., 1994; Putkonen and Swanson, 2003). It is reasonable to expect that the erosion rate of the loose sand and gravel comprising the till matrix will be one or more orders of magnitude greater than the erosion rate of the surfaces of granite boulders (Birkeland and Burke, 1988). Such rates are high enough that even large boulders now exposed on the moraine crest may have originally been buried within the till and gradually exposed by erosion. The dispersion of the ages is approximately proportional to the soil-erosion depth (Zreda et al., 1994).

In order to assess soil-erosion rates in the study area, we collected three soil samples from a moraine crest in the Tahoe right-lateral complex (samples BCS92–2, 3, and 4; Fig. 3). (These samples were previously discussed by Zreda et al. [1994], but the use of now-superseded production parameters gave somewhat different results.) Using standard cosmogenic nuclide age and erosion equations (Gosse and Phillips, 2001), the surface erosion rate can be calculated if the age of the feature is known. Assuming an actual moraine age of 140 ka (discussed later) and bulk density of 1.9 g cm^{-3}, these three samples yielded soil erosion rates ranging from 25 to 37 mm k.y.$^{-1}$, which would have resulted in stripping of 3.5–5.2 m of soil from the crests over this period. The typical heights of boulders sampled were 1.0–1.5 m, so clearly a large proportion of these boulders could have been buried during the initial portions of the moraine history. Modeling of boulder exposure during progressive erosion of the moraine crest by Zreda et al. (1994) showed a good match with actual boulder data from the moraine, and when updated estimates of ^{36}Cl production parameters (Phillips et al., 2001) and moraine age (140 ka) are used, the match remains good. These results support the hypothesis that differences in the degree of dispersion of ages on different landforms largely reflect the depth of erosional stripping of the landforms. These results indicate that for the Tahoe moraines, the typical height of the boulders sampled in this study was too low, which emphasizes the importance of selecting moraines that possess large (>2 m) boulders on their crests. This finding does not apply to the much younger Tioga moraines, for which boulders in the 1.0–1.5 m height range generally gave quite consistent results. This consistency is not in conflict with the erosion scenario; over ~20 k.y., less than 1 m of soil would be stripped from the moraine crests at these erosion rates.

GLACIAL DEPOSITS AND CHRONOLOGY

Pre-Tahoe Moraines

We have mapped the glacial sediments to the northwest of Bishop Creek as Pliocene-Pleistocene till. These closely follow the "Sherwin" and "Qtu" (till of unknown age) units of Bateman (1965), except that we have excluded a large area of Bateman's Qtu between Horse and McGee Creeks that we interpret as glacial outwash. These pre-Tahoe units consist of heavily eroded bouldery diamictons. Due to the high degree of erosion, they are unlikely to yield useful cosmogenic exposure ages, and hence they were not sampled. They are distinguished from younger mapped units largely on the basis of minimal morainal topography. As noted by Bateman, and many other authors describing analogous units in the eastern Sierra Nevada (e.g., Clark et al., 2003), without the aid of constructional topography, it is very difficult to conclusively differentiate glacial till from glacial outwash or debris-flow deposits. However, the geographical distribution of these deposits, in a band paralleling the outer margin of the younger left-lateral moraines of Bishop Creek, strongly suggests that they are the eroded remnants of much older lateral moraines. The strongly eroded morphology of the old moraines and their much stronger soil development than the Tahoe moraines (Berry, 1994) support a long time interval since their deposition. These inferences suggest that the position of the Bishop Creek glacier has migrated to the southeast (toward the base of the Coyote Plateau) over the middle to late Quaternary.

TABLE 1. IDENTIFICATION NUMBERS, ^{36}Cl SURFACE EXPOSURE AGES, SAMPLE TYPE, AND MAP UNITS FOR ROCKS SAMPLED IN THIS STUDY

Map number	Sample number	Erosion age 0 mm/k.y. (ka)	Erosion age 1 mm/k.y. (ka)	Erosion age 3 mm/k.y. (ka)	Material sampled	Map unit
	BCS92-2CR	41.3 ± 3.9	40.7 ± 3.8	41.1 ± 4.0	Soil	Ta4
	BCS92-3CR	37.7 ± 2.4	36.3 ± 2.3	35.0 ± 2.2	Soil	Ta4
	BCS92-4CR	35.6 ± 1.6	34.3 ± 1.5	33.1 ± 1.5	Soil	Ta4
1	BCR89-1	224 ± 15	248 ± 21	∞	Boulder	PTiY
2	BPCR90-27	213 ± 16	226 ± 20	824 ± 176	Boulder	Ta1
3	BPCR90-28	155 ± 18	169 ± 23	267 ± 8	Boulder	Ta1
4	BPCR90-29	160 ± 11	172 ± 13	272 ± 55	Boulder	Ta1
5	BPCR90-30	151 ± 7	165 ± 9	264 ± 33	Boulder	Ta1
6	BPCR90-31	126 ± 8	130 ± 9	165 ± 18	Boulder	Ta1
7	BPCR90-32	167 ± 12	152 ± 11	187 ± 23	Boulder	Ta1
8	BPCR90-5	136 ± 9	119 ± 7	127 ± 11	Boulder	Ta1
9	BPCR90-6	90.2 ± 4.7	95.6 ± 5.4	114 ± 8	Boulder	Ta1
10	BPCR90-7	105 ± 5	101 ± 5	109 ± 7	Boulder	Ta1
11	BPCR90-8	68.0 ± 5.2	69.2 ± 5.4	74.7 ± 6.7	Boulder	Ta1
12	BPCR90-9	143 ± 12	116 ± 9	115 ± 12	Boulder	Ta1
13	BPCR90-12	70.6 ± 3.9	72.6 ± 4.2	80.0 ± 5.3	Boulder	Ta1
14	BPCR90-1	135 ± 6	122 ± 5	133 ± 7	Boulder	Ta1
15	BPCR90-2	135 ± 8	122 ± 7	132 ± 10	Boulder	Ta2
16	BPCR90-3	105 ± 5	95.8 ± 4.4	98.1 ± 5.5	Boulder	Ta2
17	BPCR90-4	129 ± 9	133 ± 10	166 ± 19	Boulder	Ta2
18	BPCR90-10	124 ± 12	120 ± 12	146 ± 23	Boulder	Ta2
19	BPCR90-11	136 ± 23	124 ± 20	136 ± 33	Boulder	Ta2
20	BCV-1	105 ± 4	83.3 ± 2.5	75.5 ± 2.5	Boulder	Ta2
21	BPCR90-58	120 ± 7	119 ± 8	139 ± 12	Boulder	Ta2
22	BPCR90-59	140 ± 7	124 ± 6	133 ± 9	Boulder	Ta2
23	BPCR90-60	126 ± 9	123 ± 9	139 ± 14	Boulder	Ta2
24	BPCR90-61	94.4 ± 6.4	86.1 ± 5.5	86.2 ± 6.4	Boulder	Ta2
25	BPCR90-62	117 ± 5	99.4 ± 3.7	97.0 ± 4.4	Boulder	Ta2
26	BPCR91-10	145 ± 16	152 ± 18	208 ± 46	Boulder	Ta2
27	BPCR90-53	150 ± 9	115 ± 6	111 ± 7	Boulder	Ta 2
28	BPCR90-54	66.0 ± 4.4	62.1 ± 4	60.6 ± 4.2	Boulder	Ta2
29	BPCR90-55	82.5 ± 5.5	78.8 ± 5.1	79.9 ± 5.9	Boulder	Ta2
30	BPCR90-56	139 ± 5	138 ± 5	169 ± 10	Boulder	Ta2
31	BPCR90-57	140 ± 5	142 ± 5	181 ± 11	Boulder	Ta2
32	BPCR90-38	135 ± 23	126 ± 21	142 ± 36	Boulder	Ta2
33	BPCR90-39	88.0 ± 5.7	91.4 ± 6.3	105 ± 9	Boulder	Ta3
34	BPCR90-40	117 ± 5	117 ± 5.6	136 ± 9	Boulder	Ta3
35	BPCR90-41	94.5 ± 8.8	81.4 ± 6.8	76.2 ± 7.1	Boulder	Ta3
36	BPCR90-42	101 ± 9	84.5 ± 6.5	77.6 ± 6.6	Boulder	Ta3
37	BPCR90-43	129 ± 5	117 ± 4	126 ± 6	Boulder	Ta3
38	BPCR90-44	69.6 ± 6.0	64.4 ± 5.2	61.7 ± 5.3	Boulder	Ta3
39	BPCR90-45	99.6 ± 3.2	98.2 ± 3.2	107 ± 4	Boulder	Ta3
40	BPCR90-46	108 ± 4	102 ± 4	108 ± 5	Boulder	Ta3
41	BPCR90-47	57.1 ± 3.4	51.8 ± 2.9	48.1 ± 2.7	Boulder	Ta3
42	BPCR90-63	104 ± 6	89.3 ± 4.7	85.4 ± 5.2	Boulder	Ta3
43	BPCR90-64	104 ± 6	88.3 ± 4.7	83.5 ± 5.0	Boulder	Ta4
44	BPCR90-65	95.4 ± 5.4	84.1 ± 4.4	81.0 ± 4.7	Boulder	Ta4
45	BPCR90-66	83.7 ± 3.3	85.0 ± 3.4	94.1 ± 4.6	Boulder	Ta4
46	BPCR90-67	99.4 ± 3.4	86.6 ± 2.7	82.9 ± 2.9	Boulder	Ta4
47	BPCR91B-5	111 ± 10	109 ± 9	122 ± 14	Boulder	Ta4
48	BPCR91B-8	48.7 ± 4.0	448 ± 3.4	41.6 ± 3.2	Boulder	Ta4
49	BPCR91B-9	89.2 ± 7.7	78.5 ± 6.2	76.3 ± 7.0	Boulder	Ta4
50	BPCR91B-16	117 ± 10	97.3 ± 7.6	91.7 ± 8.3	Boulder	Ta4
51	BPCR91-19	57.7 ± 4.7	54.6 ± 4.3	53.0 ± 4.4	Boulder	Ta4
52	BPCR90-48	131 ± 8	119 ± 7	130 ± 11	Boulder	Ta4
53	BPCR90-49	71.9 ± 4.8	64.7 ± 4.0	61.6 ± 4.1	Boulder	Ta4
54	BPCR90-50	68.0 ± 4.6	64.9 ± 4.2	61.3 ± 4.5	Boulder	Ta4
55	BPCR90-51	131 ± 10	109 ± 8	109 ± 10	Boulder	Ta4
56	BPCR90-52	127 ± 7	127 ± 7	151 ± 12	Boulder	Ta4
57	BPCR92-1	75.1 ± 7.7	62.6 ± 5.6	55.8 ± 5.3	Boulder	Ta4
58	BPCR92-2	83 ± 14	74 ± 11	70 ± 11	Boulder	Ta4
59	BPCR90-68	80.3 ± 6.6	70.0 ± 5.2	64.8 ± 5.2	Boulder	Ta4
60	BPCR90-69	82.0 ± 5.2	70.5 ± 4.0	64.9 ± 3.4	Boulder	Ta4
61	BPCR90-70	103 ± 11	87.7 ± 8.7	83.4 ± 9.6	Boulder	Ta4
62	BPCR90-71	117 ± 8	95.9 ± 6.0	92.3 ± 6.9	Boulder	Ta4
63	BPCR90-72	94.1 ± 5.5	87.7 ± 4.9	88.4 ± 5.7	Boulder	Ta4
64	BPCR90-76	74.6 ± 4.5	67.8 ± 3.8	65.4 ± 4.0	Boulder	Ta4
65	BPCR90-78	55.5 ± 1.9	53.3 ± 1.7	52.2 ± 1.8	Boulder	Ta5
66	BPCR90-79	58.8 ± 7.0	52.8 ± 5.8	48.6 ± 5.4	Boulder	Ta5
67	BPCR90-33	134 ± 7	126 ± 7	142 ± 11	Boulder	Ta5
68	BPCR90-34	108 ± 4	102 ± 4	108 ± 5	Boulder	Ta6
69	BPCR90-35	95.6 ± 0.9	102 ± 1	124 ± 2	Boulder	Ta6
70	BPCR90-36	131 ± 10	135 ± 11	170 ± 21	Boulder	Ta6
71	BPCR90-37	129 ± 7	117 ± 6	128 ± 9	Boulder	Ta6
72	BPCR96-12	28.1 ± 1.5	25.8 ± 1.3	23.4 ± 1.1	Boulder	Ta6

(*continued*)

TABLE 1. IDENTIFICATION NUMBERS, ^{36}Cl SURFACE EXPOSURE AGES, SAMPLE TYPE, AND MAP UNITS FOR ROCKS SAMPLED IN THIS STUDY (*continued*)

Map number	Sample number	Erosion age 0 mm/k.y. (ka)	Erosion age 1 mm/k.y. (ka)	Erosion age 3 mm/k.y. (ka)	Material sampled	Map unit
73	BpCr96-13	22.2 ± 0.7	20.9 ± 0.7	19.4 ± 0.6	Boulder	Ti1
74	BpCr96-14	22.9 ± 0.9	22.4 ± 0.9	21.8 ± 0.9	Boulder	Ti1
75	BpCr96-15	27.3 ± 0.6	27.9 ± 0.7	29.2 ± 0.7	Boulder	Ti1
76	BPCR90-19	21.0 ± 1.3	21.2 ± 1.3	21.5 ± 1.4	Boulder	Ti1
77	BPCR90-20	20.0 ± 1.3	20.2 ± 1.3	20.5 ± 1.4	Boulder	Ti1
78	BPCR90-21	17.5 ± 1.5	17.6 ± 1.6	18.0 ± 1.6	Boulder	Ti1
79	BPCR91-5	15.7 ± 2.1	15.3 ± 2.0	14.8 ± 1.9	Boulder	Ti1
80	BPCR91-6	20.7 ± 1.4	19.9 ± 1.3	18.9 ± 1.2	Boulder	Ti1
81	BPCR91-7	30.4 ± 3.2	27.9 ± 2.7	25.2 ± 2.4	Boulder	Ti1
82	BPCR91-8	24.0 ± 1.1	24.4 ± 1.1	25.3 ± 1.2	Boulder	Ti1
83	BPCR91-4	17.8 ± 1.6	17.3 ± 1.5	16.7 ± 1.4	Boulder	Ti3
84	BPCR90-73	17.2 ± 0.8	16.8 ± 0.8	16.3 ± 0.7	Boulder	Ti3
85	BPCR90-74	17.9 ± 2.1	17.3 ± 2.0	16.4 ± 1.8	Boulder	Ti3
86	BPCR90-75	19.3 ± 0.9	18.9 ± 0.9	18.3 ± 0.8	Boulder	Ti3
87	BPCR91-1	17.5 ± 1.2	17.2 ± 1.1	16.6 ± 1.1	Boulder	Ti3
88	BPCR91-2	6.7 ± 1.1	6.6 ± 1.1	6.4 ± 1.0	Boulder	Ti3
89	BPCR91-3	18.5 ± 1.1	18.1 ± 1.1	17.5 ± 1.0	Boulder	Ti3
90	BPCR90-22	18.1 ± 1.7	18.3 ± 1.7	18.7 ± 1.8	Boulder	Ti3
91	BPCR90-23	10.3 ± 0.7	10.4 ± 0.7	10.4 ± 0.7	Boulder	Ti3
92	BPCR90-24	17.5 ± 0.9	16.9 ± 0.8	16.0 ± 0.8	Boulder	Ti3
93	BPCR90-25	18.1 ± 2.2	18.2 ± 2.2	18.5 ± 2.3	Boulder	Ti3
94	BPCR90-26	18.0 ± 2.6	18.1 ± 2.6	18.5 ± 2.7	Boulder	Ti3
95	BpCr96-20	19.0 ± 0.7	19.0 ± 0.7	19.1 ± 0.8	Boulder	Ti3
96	BpCr96-18	16.4 ± 1.9	16.3 ± 1.9	16.2 ± 1.9	Boulder	Ti3
97	BpCR97-8	13.9 ± 0.8	13.5 ± 0.6	13.1 ± 0.6	Boulder	Ti4
98	BpCR97-9	16.0 ± 0.8	15.6 ± 0.7	15.1 ± 0.7	Boulder	Ti4
99	BpCR97-13	16.3 ± 0.8	16.1 ± 0.7	15.7 ± 0.7	Boulder	Ti4
100	BpCR97-14	15.8 ± 0.8	15.4 ± 0.7	14.9 ± 0.7	Boulder	Ti4
101	BpCR97-15	15.6 ± 0.8	15.3 ± 0.7	14.8 ± 0.7	Boulder	Ti4
102	BPCR91-11(1)	21.5 ± 0.9	21.2 ± 0.8	20.8 ± 0.8	Bedrock	Ti4
102	BPCR91-11(2)	22.2 ± 1.0	21.9 ± 0.9	21.5 ± 0.9	Bedrock	Ti4
103	BpCr95B-6(99)	13.9 ± 0.7	13.8 ± 0.6	13.7 ± 0.6	Bedrock	Ti4
104	BPCR96-1	14.3 ± 0.5	14.0 ± 0.5	13.5 ± 0.5	Boulder	Ti4
105	BpCr96-16	15.1 ± 1.0	15.1 ± 1.0	15.2 ± 1.0	Bedrock	Ti4
105	BpCr96-17	13.7 ± 0.5	13.5 ± 0.5	13.3 ± 0.5	Boulder	Ti4
106	BpCr95-3 (99)	14.0 ± 0.5	13.9 ± 0.5	13.6 ± 0.5	Bedrock	Ti4
107	BPCR96-10	16.0 ± 0.6	15.6 ± 0.6	15.0 ± 0.5	Bedrock	Ti4
107	BPCR96-11	13.4 ± 0.4	13.2 ± 0.4	12.9 ± 0.4	Boulder	Ti4
108	BpCr95B-1(99)	14.8 ± 0.6	14.7 ± 0.6	14.6 ± 0.6	Boulder	Ti4
109	BpCR97-3	16.3 ± 0.7	16.3 ± 0.7	16.5 ± 0.7	Boulder	Ti4
109	BpCR97-4	16.6 ± 0.6	16.3 ± 0.6	15.8 ± 0.6	Bedrock	Ti4
110	BpCR97-2	11.2 ± 0.6	11.2 ± 0.6	11.2 ± 0.6	Bedrock	Ti4
110	BpCR97-1a	15.2 ± 0.7	15.2 ± 0.7	15.3 ± 0.7	Boulder	Ti4
110	BpCR97-1b	15.2 ± 0.6	15.3 ± 0.6	15.4 ± 0.7	Boulder	Ti4
111	BpCr96-21	13.9 ± 0.6	13.9 ± 0.6	14.1 ± 0.6	Bedrock	Ti4
111	BpCr96-22	14.5 ± 1.4	14.6 ± 1.4	14.9 ± 1.5	Boulder	Ti4
112	BpCr95-2 (99)	14.4 ± 0.5	14.4 ± 0.5	14.5 ± 1.5	Bedrock	Ti4
113	HB97-1	14.6 ± 0.7	14.7 ± 0.7	14.9 ± 0.7	Boulder	Ti4
114	HB97-2	16.1 ± 0.6	16.3 ± 0.7	16.7 ± 0.7	Boulder	Ti4
115	HB97-3	14.8 ± 0.4	14.8 ± 0.4	15.0 ± 0.5	Boulder	Ti4
116	HB97-4	15.0 ± 0.5	14.9 ± 0.5	14.6 ± 0.5	Boulder	Ti4
117	HB97-5	15.4 ± 0.6	15.4 ± 0.6	15.4 ± 0.6	Boulder	Ti4
118	BpCR97-10	11.9 ± 0.6	11.5 ± 0.5	10.9 ± 0.5	Boulder	Rp
118	BpCR97-11	13.2 ± 0.7	13.0 ± 0.7	12.7 ± 0.6	Bedrock	Rp
119	BpCR97-12	13.2 ± 0.6	13.3 ± 0.6	13.4 ± 0.6	Boulder	Rp
120	BPCR96-2	10.5 ± 0.4	10.4 ± 0.4	10.3 ± 0.4	Boulder	Rp
120	BPCR96-3	9.8 ± 0.4	9.8 ± 0.4	9.6 ± 0.4	Boulder	Rp
120	BPCR96-4	11.5 ± 0.3	11.2 ± 0.3	10.7 ± 0.3	Boulder	Rp
120	BPCR96-5	19.9 ± 0.4	19.2 ± 0.4	18.4 ± 0.3	Boulder	Rp
121	BPCR96-9	12.4 ± 0.5	12.2 ± 0.5	11.9 ± 0.5	Bedrock	Rp
122	BPCR96-6	12.4 ± 0.4	12.2 ± 0.3	11.8 ± 0.3	Bedrock	Rp
122	BPCR96-7	12.6 ± 0.3	12.5 ± 0.3	12.5 ± 0.3	Boulder	Rp
123	BPCR96-8	3.8 ± 0.2	3.8 ± 0.15	3.8 ± 0.15	Boulder	Rp
124	BigP97-1a	14.5 ± 0.7	14.6 ± 0.7	14.8 ± 0.7	Boulder	Rp
124	BigP97-2	14.6 ± 0.9	14.0 ± 0.8	13.3 ± 0.7	Boulder	Rp
124	BigP97-3	13.2 ± 0.7	12.9 ± 0.6	12.3 ± 0.6	Boulder	Rp

Note: Ages are reported for three assumed rock surface erosion rates: 0 mm/k.y., 1.1 mm/k.y., and 3.3 mm/k.y. Uncertainties are one standard deviation, propagating only the analytical uncertainty of the ^{36}Cl analysis. Complete sample information is available in the GSA Data Repository (Table S1; see text footnote 1).

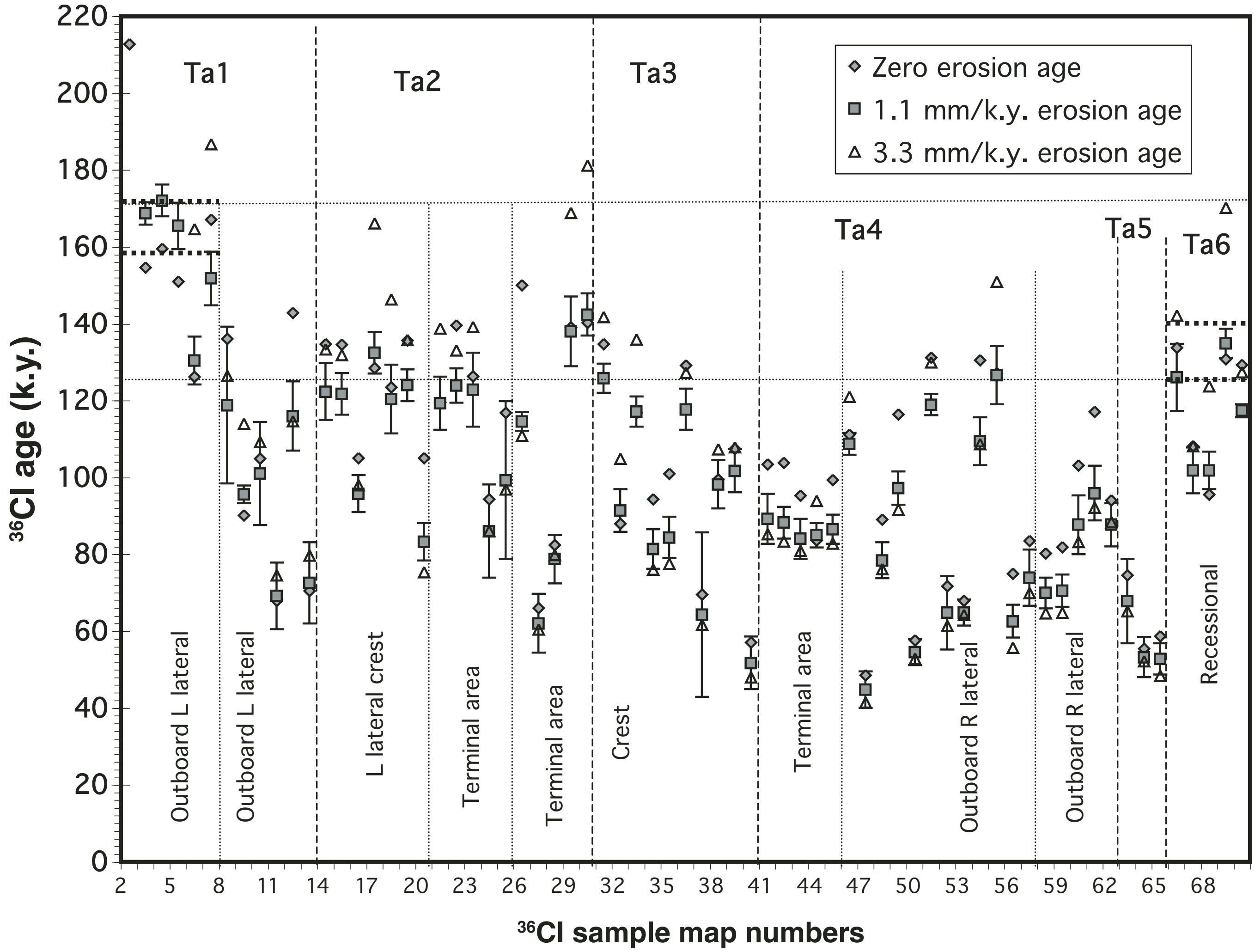

Figure 4. Chlorine-36 ages for samples from the Tahoe terminal complex at Bishop Creek. Samples are shown in order of approximate relative age of the units, from oldest on the left to youngest on the right. Ages have been calculated for three assumed erosion rates. For visual clarity, one-standard-deviation uncertainty bars are shown only on the 1.1 mm/k.y. erosion-rate symbols. The heavy horizontal lines indicate our best estimate of the maximum and minimum ages of the oldest and youngest moraines in the Tahoe complex. The upper and lower light horizontal lines indicate our best estimate of the beginning and end, respectively, of the Tahoe glaciation. As described in the text, boulders show a strong dispersion toward young ages due to generally small boulder size and high soil-erosion rates.

The mechanism for such a shift is obscure. Bateman (1965) has mapped old tills ("Sherwin") on the margin of the Coyote Plateau, well above more recent glacial limits on the South and Main Forks of Bishop Creek, that he interpreted as remnants of old right-lateral moraines. This implies that incision of Bishop Creek has been mainly vertical. Were the course of Bishop Creek to have migrated ~2 km to the southeast, any right-lateral moraines associated with the earlier position would have been destroyed.

The area immediately north of North Lake provides evidence to support an alternative hypothesis. Bateman (1965) mapped the eroded throat of a small basaltic eruption center (presumably the one responsible for the basalt clasts found in the tills downstream) on the crest of the ridge overlooking North Lake, at an elevation of 3225 m (10,600 ft) (see Plate 1 [see footnote 1]). This basalt has yielded an ^{40}Ar/^{39}Ar age of 3.46 ± 0.02 Ma (2σ) (W.C. McIntosh and F.M. Phillips, 2008, personal commun.). Field examination showed that the surface of the basalt is smoothed, apparently by glacial action, and has granitic erratics resting on it. The northeast margin of the Tioga glacier was 225 m below this point, demonstrated by a small patch of lateral moraine preserved almost directly below the basalt. Glacial overriding of this basalt outcrop thus was probably quite ancient and suggests that at that time, the current ridge crest formed part of the bottom of the valley of the North Fork. Ice flow over the (current) ridge top would have carried basalt clasts toward the north and northeast, where they comprise much of the till below Grouse Mountain.

The distribution of basalt clasts in the tills to the north and northeast supports an origin from the North Fork of Bishop Creek. There are no sources of basalt in the Birch, McGee, or Horton Creek drainages, and thus the presence or absence of basalt should be diagnostic of till coming from the North Fork of Bishop Creek. The deposit that is richest in basalt is the large patch of "Pliocene-Pleistocene till" directly south of Grouse Mountain. More than 75% of the cobble-sized clasts in this till are basalt, although virtually all the boulders >1 m diameter

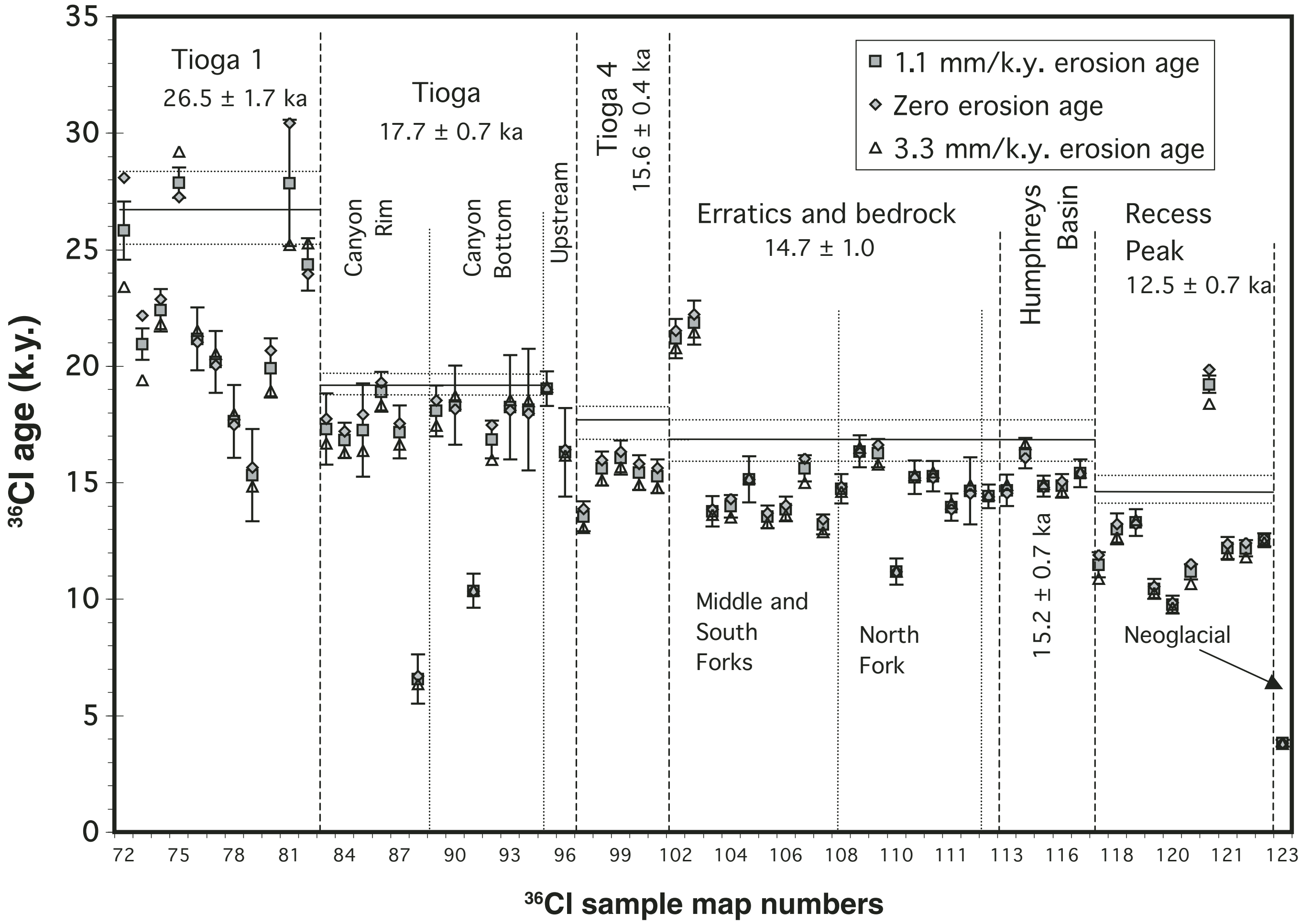

Figure 5. Chlorine-36 ages for samples from the Tioga, Recess Peak, and Holocene units at Bishop Creek. Samples are shown in order of relative age of the units, from oldest on the left to youngest on the right. Ages have been calculated for three assumed erosion rates. For visual clarity, one-standard-deviation uncertainty bars are shown only on the 1.1 mm/k.y. erosion-rate symbols. Means and standard deviations calculated from the 1.1 mm/k.y. ages for each unit are given at the top. Samples that were not used in calculating the means and standard deviations are indicated by open symbols (boxes) for the 1.1 mm/k.y. erosion rate.

are granodiorite. The deposits of Pliocene-Pleistocene till down the course of Birch Creek also contain abundant basalt clasts. The pre-Tioga older moraines to the north and west of Bishop Creek contain moderate to minor amounts of basalt. Basalt is sparse in the pre-Tioga younger moraines (the "basalt boundary" line in this area in Plate 1 separates moraines containing abundant basalt clasts to the south of the line from those with very sparse clasts to the north of the line [see footnote 1]). This distribution of basalt clasts indicates that at one time, ice flowed almost due north from the North Fork toward Grouse Mountain and then turned northeast and flowed down the present course of Birch Creek. Given the depth of incision of Birch Creek since deposition of the till (Knopf, 1918) and the 3.46 Ma age of the basalt, a late Pliocene age for this glaciation is not improbable.

The possibility of a previous glacial course in this direction is supported by the canyon-wall topography. The ridge containing the basalt outcrop (directly north of North Lake) ends in a prominent triangular facet. The facet appears to have been glacially carved, but it is not parallel with the current orientation of Tioga and Tahoe moraines downstream of it, which is 50°E of N. The facet is oriented 25°E of N and is directly aligned with the pre–Tioga 1 and Pliocene-Pleistocene till deposits. The precipitous eastern wall of the Middle Fork canyon (south of Jawbone Canyon) is oriented parallel to the triangular facet. We speculate that these canyon walls were cut at a time when only the South Fork followed the current path of Bishop Creek and the combined North and Middle Forks followed a separate course to the north, toward Grouse Mountain. Under this scenario, the degraded tills on the northwest side of Birch Creek, south of Grouse Mountain and McGee Meadow, could be the remains of left-lateral moraines from the North and Middle Forks. It is possible that some of the till on the southeast side of Birch Creek could be remnants of correlative right-lateral moraines.

Tectonic factors could have played a role in evolution of the postulated drainage pattern into the modern one. We and Bateman (1965) have mapped a series of northeast-trending faults through this area. Southwest of the current confluence of the Middle and South Forks, these faults are valley-down; northeast of the confluence, they are valley-up (antithetic). Downdrop of the central block would have tended to channel flow toward the current confluence. Some support for a relatively recent establishment of the current confluence of the Middle/North and

South Forks can be found in the stream profiles of the forks. The combined Middle/North Fork drainages supported more extensive glacial ice than the South Fork (Plummer, 2002), yet the Middle/North Forks exhibit a fairly steep step above the confluence, while South Fork has a relatively regular and gentle profile upstream of the confluence. This could be explained if the South Fork captured the Middle/North Forks in the relatively recent geological past.

Tahoe and "Older Pre-Tioga" Moraines

Tahoe moraines were mapped in detail in the terminal moraine complex. At that time (1990), we considered it likely that these might represent tills deposited over several glacial cycles. However, the ^{36}Cl dating subsequently indicated that they were all deposited during MIS 6 (see following). Nevertheless, it seems likely that, just as several distinct advances can be distinguished within the Tioga (MIS 2) moraines, multiple advances over a period of 20–40 k.y. may have produced the observed complex sequence of Tahoe terminal moraines, although quantitatively distinguishing these is at the current limit of resolution of cosmogenic surface exposure dating. The detailed mapping may help to define such events. With very limited exceptions, the Tahoe moraines were sampled for ^{36}Cl dating only in the terminal complex. Due both to erosion and to extensive overlap of the Tioga left-lateral on the Tahoe laterals, it is not possible to reliably trace the Tahoe crests up-valley from the terminal complex, and therefore the lateral moraines were simply mapped as "pre-Tioga older" (pTio) and "pre-Tioga younger" (pTiy). This mapping of the lateral moraines generally follows the subdivision of Bateman (1965).

As described already, the quite extensive set of Tahoe laterals may record a shift in the course of the Middle/North Forks from down the current course of Birch Creek to the present configuration, perhaps in several steps. Based on comparison of the surficial characteristics of the lateral crests with those of the dated crests in the terminal complex, most of the crests appear to substantially predate MIS 6. One boulder (BPCR89–1) was sampled on what appeared to be a subcrest beyond the younger pre-Tioga lateral moraine crest, and it gave an age of 233 ± 19 ka, which could be correlative with MIS 8. This limited evidence tends to support deposition of the pre-Tioga lateral moraine complex over a long period of time, ranging from possibly the mid-Pleistocene to the end of MIS 6.

A map of the terminal complex is given in Plate 1 (see footnote 1) and Figure 3, and a stereophotograph pair is shown in Figure 2. Morphological features are designated in order of apparent relative age, based on superposition and crosscutting relationships. Although these morphological units record glacial events, they do not necessarily correspond to distinct glacial advances. The most prominent and stable feature of the complex is the left-lateral crest (Ta2). This crest partially buries several subdued ridges that project northward from beneath its outer flank (Ta1). This configuration is consistent with a general tendency of the complex to grow from northwest toward the southeast. The Ta1 remnant crests are apparently left from the early stages of deposition of the complex. The main Ta2 crest was then deposited on top of, and slightly southeast of, these foundational deposits. During the latter part of this stage, the glacier appears to have constructed a relatively high and continuous terminal loop just northeast of where State Highway 168 curves across the bottom of Sand Canyon. This loop apparently presented such a barrier to subsequent ice advance toward the northeast that the ice tongue instead repeatedly overtopped the right-lateral moraine close to the main terminal loop. These events formed small tongue-like loop moraines draped over the southeast flank of the complex (Ta3 and Ta4). The right-lateral moraine seems to have been overtopped repeatedly, and the left-lateral moraine never overtopped because the left one is consistently higher than the right one, by ~30 m. Late in the depositional sequence, and perhaps during the early stages of deglaciation, meltwater cut through the main terminal loop and incised the present course of Sand Canyon northeast of Highway 168. Ice appears to have briefly advanced down this cut, leaving two small terminal moraines before it retreated (Ta5). Finally, the last Tahoe ice in Sand Canyon deposited the low Ta6 loop across the bottom of the canyon. The Ta6 loop is probably recessional, but it could possibly have resulted from a brief readvance.

The distribution of ^{36}Cl ages (Fig. 4) indicates that this entire sequence of events took place during MIS 6. The oldest ages are, as the relative age sequence would indicate, from samples collected on Ta1. The oldest Ta1 ages cluster in the range 170–140 ka. The oldest ages on the Ta2 crest and terminal loop overlying Ta1 are in the range 150–120 ka. The right-lateral moraine and overlying overflow loops (Ta3 and Ta4) scatter widely between 130 and 80 ka. This age distribution might be used to infer that some of these features were younger than MIS 5e, except that the ages from the Ta6 recessional loop, which must provide a limiting minimum age for all of the other features, cluster close to 130 ka. As described above, we attribute the wide scatter of ages for the Ta3 and Ta4 features to extensive erosion. We believe that erosion was deeper on the southeast flank partly because of later extensive undercutting and incision by Bishop Creek during the Tioga glaciation, and partly because the till deposited there appears to be more sandy and less bouldery than the Ta2 moraines. Large numbers of boulders have been demonstrated to armor slopes and greatly slow soil erosion rates (Granger et al., 2001).

We collected a sample from the flat top of a very large granite boulder (10 × 5 × 5 m high; BPCR91–10) that was embedded in outwash ~500 m east of the northeastern tip of the Tahoe terminal complex. Part of the reason for sampling it was to test the hypothesis that the boulder was not transported by a debris-flow event, but was rather directly deposited by a glacier on the crest of a pre-Tahoe moraine that was subsequently buried in Tahoe outwash. However, the boulder produced a ^{36}Cl age of 146 ± 17 ka, which is completely consistent with the Tahoe moraine ages. We surmise that the boulder was transported across the outwash fan by a catastrophic proglacial debris-flow event such as those described in this region by Blair (2001).

Tioga Glacial Features

Mapping of the Tioga moraines is illustrated in Plate 1 (see footnote 1), and detail of the moraines in the terminal complex area is shown in Figure 3. Our mapping follows closely that of Bateman (1965), with additional subdivisions based mainly on ^{36}Cl chronology. Phillips et al. (1996a) combined ^{36}Cl chronologies from several localities in the eastern Sierra region to delineate the following age groupings: Tioga 1 at ca. 30 ka, Tioga 2 at 25–20 ka, Tioga 3 at 19–17 ka, and Tioga 4 at 17–15 ka. Of these, we have measured ^{36}Cl exposure ages that correspond to Tioga 1, 3, and 4 at Bishop Creek. (We note that current ages for the same samples may vary slightly from those of Phillips et al. [1996a] due to revision of the ^{36}Cl production formulation and parameters by Phillips et al. [2001]). Tioga 1 and 3 correspond to Bateman's "Tioga" (Qti), and Tioga 4 corresponds to his "younger advance of the Tioga" (Qtiy).

Tioga 1 Deposits

Tioga 1 was identified only at the upper (southwest) end of Sand Canyon. These moraines apparently resulted from an initial advance of ice close to the beginning of the Tioga glaciation into the terminal complex vacated by the Tahoe glacier. The ^{36}Cl ages from these moraines showed an unusually wide scatter toward young ages. The cause may possibly be unusually high erosion in an exposed position close to the active glacier during the remainder of the Tioga glaciation (~15 k.y.). Inasmuch as very few outliers with anomalously old ages due to inheritance were observed in the remainder of the data set (Fig. 4), the average of the four oldest ^{36}Cl ages, 26.5 ± 1.7 ka, was accepted as the best age for the moraines. The succession of terminal loops on the floor of Sand Canyon records an episodic withdrawal of Tioga ice from the valley floor. No subsequent Tioga glaciers occupied Sand Canyon; instead they flowed down the current course of Bishop Creek to the southeast. The Tioga (and modern) valley of Bishop Creek at the confluence with Coyote Creek is ~120 m lower in elevation than the floor of Sand Canyon at a corresponding distance from the point where the Tahoe and Tioga glacial courses diverged. The Tioga 3 glacier flowed through a gap excavated in the right-lateral Tahoe moraine. The most probable explanation for these observations is that during the initial advance of the Tioga glacier (Tioga 1), the ice surface overtopped the right-lateral Tahoe moraine, and the marginal meltwater stream was diverted across the crest, cutting through it, and thus initiating an avulsion of the course of the glacier. The Tioga 1 morainal loops on the floor of Sand Canyon probably record successive positions of the glacier snout as the gap was enlarged and ice flow was increasingly diverted into the new course. The pronounced break in slope of the southeast flank of the Tahoe terminal complex as Bishop Creek is approached indicates that the area between the Tahoe complex and the Coyote Plateau was incised after the glacier changed course. The rapid withdrawal of Tioga 1 ice from Sand Canyon, indicated by the small volume and good preservation of the Tioga 1 terminal loops, indicates that this incision was accomplished penecontemporaneously with the avulsion through the right lateral.

Mechanism for Avulsion. The fundamental explanation for this avulsion appears fairly straightforward: sedimentation by the Tahoe glacier had remained constrained within the limits fixed by its lateral and terminal moraines until the terminal complex aggraded well above the surrounding topography, and when a lateral moraine was breached, the glacier was permanently diverted into the steeper path that the breach provided. This explanation is completely analogous to that for classical river avulsion. However, the mechanism by which the glacial avulsion was accomplished is more difficult to reconstruct, largely because the evidence for events leading up to the avulsion was mostly destroyed by the erosion that accompanied the diversion.

Avulsion is accomplished when a lateral moraine is breached and the glacier moves through the breach rather than down its previous course. Given the relatively low erosivity of glacial ice, particularly toward the glacier terminus, it is likely that the actual erosion of the breach is mostly by meltwater. However, this requires a significant proportion of the melt from the glacier to flow over a single point on the lateral moraine, and it is the circumstances that produce such a diversion that explain the occurrence of avulsions. Prior hypotheses regarding mechanisms for glacial avulsion in the area have tended to focus on interglacial fluvial erosion of lateral moraines (Blackwelder, 1929) or on the deposition of large terminal loops that diverted glacial meltwater over the crests of lateral moraines (Russell, 1889; Kesseli, 1941), allowing subsequent advances to proceed through the gap created by the overflow. However, no remains of any such massive loop are evident at Bishop Creek, nor do other valleys in the area contain massive early Tioga terminal loops. Rather, we suggest that the following factors may have contributed to the avulsion.

(1) Reduced topographic gradient in the terminal area. The average topographic gradient down the axis of Bishop Creek from South Lake to Dutch Johns Meadow is 0.07. The reconstructed gradient from the lower end of Dutch Johns Meadow to Sand Canyon is 0.03. This difference is almost entirely due to aggradation in the terminal area; the gradient of the topographic surface of McGee Meadow, on which the terminal complex is constructed, is also 0.07. As the gradient driving flow decreases, the glacier thickness must increase to compensate (Paterson, 1995). As described previously, the right-lateral moraine in the terminal area was incapable of fully constraining the ice at the end of the Tahoe glaciation and was overridden by numerous small overflows. This incapacity was probably in large part a result of the reduced gradient. The ability to constrain the advancing Tioga glacier would have been even less, due to deposition of till and recessional moraines on the valley floor by the retreating Tahoe glacier. Ice overflow on the lateral moraine would have diverted marginal meltwater that could initiate incision.

(2) Steep profile of the advancing Tioga glacier. Advancing glaciers normally exhibit much steeper longitudinal profiles at the snout than do retreating ones (Jóhannesson et al., 1989; Paterson, 1995). The advancing Tioga glacier would thus have tended to overtop the lateral moraine more readily than the retreating Tahoe glacier.

(3) Upstream fluvial incision of the right-lateral moraine. The Tioga glacier avulsed through the right-lateral moraine at the point where it diverged from the bedrock slope of the Coyote Plateau. Outboard of the lateral crest, this area would have formed a small, blind, tributary valley of Coyote Creek prior to the avulsion. Although the drainage area is too small to have produced effective incision of the lateral moraine, runoff from Coyote Plateau above, or headward incision from Coyote Creek, may have produced at least minor undercutting of the base of the right-lateral moraine.

(4) Increased pore pressure at the bedrock–lateral moraine junction. As the advancing Tioga glacier began to overtop the right-lateral moraine, meltwater from the surface and flanks of the glacier would tend to saturate the moraine. Upstream of the point of the avulsion, the lateral moraine is plastered against the relatively impermeable granite bedrock of the Coyote Plateau. This would have tended to channel meltwater downstream to the point where the lateral moraine diverged from the bedrock, creating a high degree of saturation and large water flux toward the outer base of the moraine. This could have led to failure of the lateral moraine as the load of the advancing ice was imposed. This failure could have taken the form of a sudden collapse (i.e., a landslide) or that of rapidly self-propagating sapping. Either type of failure could have been promoted by fluvial undercutting, both due to incision of the lateral moraine by diverted marginal meltwater and due to prior fluvial erosion, as hypothesized in point 3. In summary, we suggest that some, or all, of the factors described here, probably acting in concert, resulted in the creation of a gap in the right-lateral moraine through which ice was able to advance early in the Tioga glaciation and ultimately divert the course of the glacier.

Tioga 3 Deposits

The Tioga 1 loop moraines at the upper end of Sand Canyon are crosscut along the edge of Bishop Creek canyon by younger lateral moraines (Plate 1 [see footnote 1]; Fig. 3). The crests of these lateral moraines are slightly higher (~10 m) than the Tioga 1 crests they bury. The ^{36}Cl data from these moraines clearly demonstrate that the lateral moraines were not formed during the Tioga 1 readjustment event, or during the following Tioga 2 phase (25–20 ka), but rather during the Tioga 3 phase (19–17 ka). Two lateral moraines were sampled: the main lateral moraine on the canyon rim, and an apparent earliest recessional-phase lateral moraine that inclines slightly down the canyon wall. One age (BPCR91–2) was clearly anomalously young, but the other six ages were in good agreement, yielding a mean of 17.6 ± 0.8 ka. These lateral moraines can be traced into subtle terminal loops that trend steeply down the canyon walls just above the confluence of Coyote Creek with Bishop Creek. Because of the poor preservation and potential for postdepositional erosion of these terminal features, we did not sample them for surface exposure dating. We instead focused on more stable recessional loops that surround Power Plant 3 on Bishop Creek. The ^{36}Cl ages from these loops also contained one young outlier (BPCR90–23), but the remaining four samples gave a combined age of 17.9 ± 0.7 ka, indistinguishable from the lateral moraine above. The combined age of the lateral and loop samples is 17.7 ± 0.7 ka. Two large boulders on small recessional moraines between Power Plant 3 and the confluence of the Middle and South Forks were also sampled. These were ~1.7 and 5.2 km upstream of Power Plant 3, and they yielded ages of 19.0 ± 0.7 and 16.3 ± 1.9 ka. Although these ages do not closely constrain retreat of the Tioga 3 glacier, they are consistent with relatively rapid withdrawal of ice following the maximum advance at ca. 17.7 ka. This inference is also consistent with the regular spacing and relatively small volume of the Tioga 3 recessional loops on the canyon floor between Power Plant 3 and the confluence of the Middle and South Forks.

Given that the maximum dated advance of the Tioga glacier was during Tioga 3, we also mapped the upstream lateral moraines of clearly Tioga age as Tioga 3. However, it is also possible that they could have been deposited during the Tioga 2 period. We mapped an interesting sequence of Tioga moraines ~1 km north of Intake No. 2 (Plate 1 [see footnote 1]). In this area, the Tioga glacier flowed over a preexisting Tioga left-lateral moraine and down a gap between the Tioga and pre-Tioga (probably Tahoe) lateral moraines. An intricate sequence of recessional moraines was deposited as the overflow stabilized and retreated. The gap between the probable Tahoe and the Tioga lateral moraines is apparently the latest manifestation of the southeastward migration of the lateral moraines in the area between Aspendell and the confluence of the North/Middle and South Forks, which we hypothesize was due to relatively recent capture of the North/Middle Forks by the South Fork.

Tioga 4 Deposits and Tioga Retreat

The well-defined lateral and terminal moraines resulting from the latest Tioga readvance out of the Middle Fork and projecting across the canyon of the South Fork (Plate 1 [see footnote 1]) are among the most striking glacial geologic features of Bishop Creek and were discussed at some length by the earliest workers (Knopf, 1918; Bateman, 1965). Bateman emphasized the steepness of the descent of the lateral moraines toward the terminal area and the strong evidence that the moraines were the result of a readvance rather than a recessional pause, and both authors commented on the apparent lack of any correlative moraines in the South Fork valley. However, our mapping did identify low mounds of till in the vicinity of Aspen Campground on the South Fork, and other more subtle patches of till and groupings of erratic boulders for some distance downstream of the campground, which we interpret as probably correlative with the distinct moraines at the confluence of the Middle and South Forks (hereafter referred to as "the confluence"). Upstream of the Aspen Campground till on the South Fork, and upstream of the area of Aspendell on the Middle Fork, there is virtually no till, only scattered erratic boulders on bedrock, until the Recess Peak moraines are encountered.

For this sequence, the objective of our sampling was to establish the age of the confluence moraine and to sample features spread over a large area above the confluence moraine in order to determine the chronology and rate of deglaciation. We sampled both glacially polished and striated bedrock and erratic boulders on the bedrock. If suitable sites were available, we collected paired erratic/bedrock samples from the same locality. In addition to samples from the Bishop Creek basin, we also crossed Piute Pass to Humphreys Basin and sampled boulders on low, broad moraines we had mapped in that basin. Part of the motivation for this was because moraines at similar high elevation, but outside the Recess Peak moraine limits, are not found in the main drainage of Bishop Creek, and we wished to investigate whether there was an additional advance we had not identified and sampled in Bishop Creek. Another motivation was that, as previously noted by Matthes (1960, 1965), striations and chattermarks on bedrock exposures at Piute Pass indicated that ice had flowed over the divide from west to east, and we wished to determine the additional accumulation area provided by this flow and whether it had influenced the chronology of deglaciation.

All five boulders sampled on the confluence moraine gave consistent ages, averaging 15.2 ± 1.0 ka. This age is within the grouping of the Tioga 4 advance (Phillips et al., 1996a). The ~2 k.y. difference in age from the Tioga 3 moraines tends to confirm Bateman's (1965) conclusion that the confluence moraine resulted from a readvance rather than a stillstand during Tioga 3 recession.

The samples upstream of the confluence moraine also gave consistent results. Only one sample, BPCR91–11, gave an anomalous (old) age of 21.2 ± 0.8 ka. This sample was from the summit of a large roche moutonnée in the bottom of the Middle Fork valley 1.2 km above the confluence. We replicated this analysis on sample material left over from the previous sample preparation, yielding an age of 21.9 ± 0.9 ka. The first analysis employed the ion-selective electrode–diffusion-cell method for Cl analysis, and the second employed the ID-MS method. This result indicates that the anomalously old age was due to inadequate removal of rock from the surface of the roche moutonnée to completely reset the cosmogenic clock from the time of previous exposure. Roche moutonnées are created because the rock of which they are composed is less erodible than the surrounding rock, and very low abrasion rates on their tops have been observed elsewhere using cosmogenic nuclides (Briner and Swanson, 1998).

The remaining 15 samples gave a mean age of 14.7 ± 1.0 ka. Paired erratic/bedrock samples (see Table 1S [see footnote 1]) are in good agreement and yielded no evidence that bedrock samples (with the exception of the one large roche moutonnée) are anomalously old due to inheritance. There is no clear spatial distribution to the ages (e.g., increasing in age with increasing elevation), probably because the deglaciation time was too short to be clearly resolved by cosmogenic dating at its current status. Given the relatively large numbers of samples analyzed on both the confluence moraine and the upstream area, the 500 yr difference in the means of the two groups may be a reasonable estimate of the deglaciation interval, and it probably was not longer than 1 k.y. This rapid deglaciation is in agreement with the geomorphic evidence for rapid deglaciation in the range detailed by Clark (1976) and Clark and Clark (1995).

Two samples (BPCR96–1 and BPCR95–3) from the South Fork, at or below South Lake, yielded ^{36}Cl ages between 14 and 13 ka, which are in good agreement with the rest of the basin above the confluence moraine (Fig. 4), and they are not close to 18 ka as would be expected if the Tioga 4 readvance had failed to move down the South Fork valley. The lack of a prominent moraine is thus presumably because the snout of the glacier in the South Fork failed to stabilize long enough in any one place to deposit a recognizable moraine. This difference from the Middle/North Fork valley may have been caused by the glacial dynamics of tributaries such as the Green Lakes and Tyee Lakes, which have distinctly different hypsometries, and thus glacial budgets, than the main basin above South Lake. The glacier modeling of Plummer and Phillips (2003) has confirmed that the equilibrium position of ice in the South Fork is short of the confluence, even when North/Middle Fork ice has moved past the confluence. In any case, the virtual absence of recognizable moraines in the South Fork, even though the moraines in the Middle/North Fork demonstrate was clearly a major readvance, is a salutary reminder of the inherent variability of the glacial record and the necessity of examining a number of localities before drawing conclusions regarding the glacial history of a region.

In Humphreys Basin, on the west side of the Sierra Crest, three samples were collected from a low, broad body of till that was probably a ground moraine resulting from rapid melt-back of the Humphreys Basin ice field. This moraine must have been deposited late in the final Tioga deglaciation. Additionally, two boulders were sampled on the crests of bedrock ridges ~1.5 km west of Mt. Humphreys. These boulders must also have been deposited very close to the end of the deglaciation. All five samples gave ages in good agreement, with a mean of 15.2 ± 0.6 ka. This mean age is virtually identical to that from the Tioga 4 maximum moraine at the confluence and to the up-valley samples from Bishop Creek, and it strongly supports a very rapid deglaciation that was synchronous on both sides of the crest.

In order to assess the extent of glacial overflow from Humphreys Basin to the North Fork of Bishop Creek, glacial flow indicators (striations and chattermarks) were mapped (Plate 1 [see footnote 1]). These indicators clearly demonstrate that the crest of the ice divide in Humphreys Basin lay 0.5–1.0 km to the west of the topographic crest of the range. This offset of the glacial divide can be attributed to doming of ice to the west of the pass, produced by the relatively flat topography of Humphreys Basin and higher snowfall rates on the west side of the crest, combined with the low barrier offered by Piute Pass and the steep gradient for ice flow down the North Fork of Bishop Creek. This cross-divide flow added 3.7 km^2 to the accumulation area of the Bishop Creek glacier, which was ~75 km^2. Flow across Piute Pass thus added only ~5% to the total Bishop Creek glacier accumulation area, but it constituted a 15%–20% addition to the accumulation area of the North Fork glacier. Over longer time scales, progressive addition of this ice flux as Piute Pass was gradually worn down would have substantially increased the size of the North Fork glacier and may have thus contributed to the hypothesized pre-Tahoe integration of the North and Middle Forks with the South Fork.

Post-Tioga Glacial Features

Recess Peak Deposits at Bishop Creek

Evidence showing that the Recess Peak glaciation was terminal Pleistocene, rather than late Holocene as thought by Birman (1964), who originally defined the advance, was obtained from the Baboon Lakes in the Middle Fork drainage by Clark and Gillespie (1997). We sampled four boulders on the Recess Peak terminal moraine at Baboon Lake (BPCR96–2 through BPCR96–5), glacially polished bedrock from a low roche moutonnée ~0.5 km upstream of the Baboon Lake terminal (BPCR96–9), and a boulder/bedrock pair (BPCR96–6 and BPCR96–7) from near the outlet of Sunset Lake, ~1.7 km upstream of the terminal. We also sampled other Recess Peak features within the Bishop Creek basin: a boulder/bedrock pair (BPCR97–10 and BPCR97–11) at Topsy Turvy Lake (Middle Fork) and a very large boulder (BPCR97–12) on the Recess Peak terminal moraine at Treasure Lakes (South Fork).

The four boulders from the terminal moraine at Baboon Lake yielded a mean age of 12.6 ± 4.4 ka. The cause of the large boulder-to-boulder variability of the ^{36}Cl ages from this site is unknown. Fortunately, the samples from the other sites in the basin yielded more consistent ^{36}Cl ages. The mean of the remaining six samples was 12.5 ± 0.7 ka. The oldest ages from the terminal moraine areas (13.3 ± 0.6 ka; BPCR97–12) are ~1 k.y. older than those closest to the cirque headwall (12.4 ± 0.3 ka: BPCR96–6 and 7), but given the limited number of samples and the uncertainties in the ages, it is difficult to know whether this is a realistic estimate of the duration of deglaciation. Based on field assessment, sample BPCR97–12 (Treasure Lakes) was considered the best sample, and it also gave the oldest age: 13.3 ± 0.6 ka.

Recess Peak Deposits at Big Pine Creek

Big Pine Creek is the next major drainage south of Bishop Creek. One interesting and enigmatic feature of that basin is a large furrowed moraine (Plate 1 [see footnote 1]) that issues from the north slope of Contact Pass into the North Fork of Big Pine Creek and that extends to form the southern limit of Second and Third Lakes. We sampled three boulders from the crest of this apparent inactive debris-covered glacier (BigP97–1 through BigP97–3). These yielded a mean ^{36}Cl age of 13.9 ± 0.9 ka. This mean age is very similar to, although slightly older than, that of the most reliable ^{36}Cl ages for the Recess Peak deposits in the Bishop Creek basin. This age confirms the Big Pine Creek feature as a fossil Recess Peak debris-covered glacier. It also strongly supports the hypothesis that a very similar feature just west of North Lake is probably also a fossil Recess Peak debris-covered glacier. Although Bateman (1965, p. 170) considered the low elevation and southward aspect of the source area to be strong evidence against an origin as a Holocene rock glacier, the similar characteristics of the clearly Recess Peak–age deposit in Big Pine Creek support a glacial, rather than landslide, origin for the North Lake feature.

Holocene Glacial Deposits

Birman (1964) named the youngest Holocene glacial deposits in his study area the Matthes moraines. These are generally small moraines that are found only very close to the headwalls of well-shaded cirques. Despite their limited extents from the cirques, they are commonly voluminous relative to the size of the glacier feeding them. Birman (1964) attributed these deposits to historical advances during the Little Ice Age. Dating these deposits was not part of our study; if they were formed during the Little Ice Age, they are too young to resolve using cosmogenic nuclide accumulation. However, we did collect one sample from the snout of the large active rock glacier west of Sunset Lake (BPCR96–8) as a test of whether the dating method would yield the expected late Holocene age. This sample returned an age of 3.8 ± 0.15 ka. Although the age is much older than the Little Ice Age (<1 ka), it is consistent with lichenometric and radiocarbon evidence suggesting that neoglaciation in the Sierra Nevada was initiated ca. 3.4 ka (Konrad and Clark, 1998; Bowerman and Clark, 2005). The position of the sampled boulder near the toe of a relatively elongate rock glacier is also consistent with Konrad and Clark's conclusion that slow-developing rock glaciers in the area probably initiated more than 1000 yr ago. However, given that it is a single sample and that most surface debris on these rock glaciers is derived from rockfall, inheritance cannot be ruled out, and the age should be viewed as a limiting maximum.

ACCURACY OF THE GLACIAL CHRONOLOGY

This study has relied entirely on cosmogenic ^{36}Cl ages to establish the glacial chronology of Bishop Creek. However, at present, there exist substantial uncertainties with regard to the systematics of cosmogenic surface exposure dating, including spatial scaling of the cosmic-ray flux, temporal variability of the flux, production rates and reactions, and factors associated with surface stability, coverage, and geometry of rocks sampled (Gillespie and Bierman, 1995; Gosse and Phillips, 2001). In particular, several different production-rate parameterizations have been proposed for ^{36}Cl (Stone et al., 1996a, 1996b; Phillips et al., 2001; Swanson and Caffee, 2001). Due to these uncertainties, we conducted a careful comparison between our results and ages provided by independent constraints, and between our ^{36}Cl ages and those calculated using alternative production-parameter sets, with the goal of assessing the systematic uncertainties that might affect our results.

Independent glacial chronology results from Bishop Creek itself, with which the ^{36}Cl chronology presented in this paper can be compared, are scarce, being limited to the minimum ^{14}C age for the Recess Peak glaciation from a lacustrine core at Baboon Lakes (Clark and Gillespie, 1997). This date is corroborated by ^{14}C data from other lacustrine cores in the Sierra Nevada (Clark, 1997). However, there are also a number of chronological data points available from elsewhere in the mountain range. Given that climatic events within an area of this size were probably nearly synchronous, these are also useful for testing the Bishop Creek chronology. Comparison with these independent constraints will be discussed next, going from strongest to weakest constraint.

The most direct comparison is with the Baboon Lakes radiocarbon date. The mean of six Recess Peak glacial features was 12.5 ± 0.7 ka; the age of the oldest and probably best sample was 13.3 ± 0.6 ka. These ages compare favorably with the limiting minimum calibrated ^{14}C age of 13.1 ± 0.09 ka measured on a plant macrofossil found just above the Recess Peak glacial outwash in the sediment core. The length of time required to reestablish vegetation in the lake basin subsequent to glacial retreat is not known; therefore, the extent to which the ^{14}C age postdates the time of deposition of the terminal Recess Peak moraine is uncertain, but it is likely in the range of 100 or more years. Several of the ^{36}Cl samples were from positions behind the terminal moraines and therefore also postdate the Recess Peak maximum, again by uncertain amounts. Given all of these variables, and also taking into account the sample-to-sample variability in ages, it seems probable that the ^{36}Cl ages are systematically younger than the best estimate for the glacial event based on the calibrated ^{14}C age, but only by a very small amount, perhaps 2%–4%.

Clark et al. (1995) tabulated ^{14}C ages that provide minimum limits on the initiation of Tioga deglaciation. Evaluation of the three oldest calibrated ages indicates that deglaciation was initiated shortly before 18.1 ± 0.5 ka (note that this refers to the initiation of retreat from the maximal position, not the final retreat that occurred several thousand years later). Assuming that this corresponds to the retreat of the Tioga 3 glacier at Bishop Creek, the ^{36}Cl data yield a range of 18.4–17.0 ka for the deposition of the Tioga 3 terminal moraine. The overlap in these two age ranges would allow them to be completely consistent, but the comparison again suggests that the ^{36}Cl ages may be biased slightly young.

There are no firm and precise independent age constraints for the Tahoe glaciation in the Sierra Nevada, but the ages may be compared to regional and global paleoclimatic records, which are well-established through U/Th dating (Winograd et al., 1997; Gallup et al., 2002; Spötl et al., 2002). The distribution of Tahoe ^{36}Cl ages from Bishop Creek indicates that the Tahoe terminal moraines there were deposited between ca. 170 and 130 ka. MIS 6 spanned the interval 175 to 135–130 ka, probably terminating closer to 135 than 130 ka (Spötl et al., 2002; Siddall et al., 2006), although the exact timing and sequence of events are somewhat controversial. This comparison involves some degree of circularity, since the Tahoe glaciation is not otherwise firmly constrained to MIS 6 (Gillespie, 1984; Gillespie and Molnar, 1995), but nevertheless, identification of the Tahoe deposits at Bishop Creek within the global MIS 6 interval appears highly plausible. Comparing the age ranges, the general agreement is excellent, but there is once again a suggestion that the ^{36}Cl ages may be biased slightly young by a small amount (~3%). In summary, a variety of comparisons with independent age constraints indicate that the ^{36}Cl ages from the eastern Sierra Nevada, calculated using the production constants of Phillips et al. (2001), are probably quite close to the actual ages, but may be biased toward the young side by ~3%. This 3% value should not be interpreted as a measure of the uncertainty associated with either individual ^{36}Cl exposure ages or interpreted landform ages, but rather as an estimate of the systematic bias in calculated ages (relative to the best available independent values) that is associated with this particular set of samples and analytical methodology.

Ideally, random uncertainties associated with the processing and chemical/isotopic analysis of a single sample should be addressed by measuring individual samples repeatedly. This was not done in this study; thus, the best estimate available is from the comparison of multiple ages obtained from a single moraine. Such a comparison introduces additional sources of uncertainty (e.g., all samples my not have been exposed simultaneously, may not have had exactly the same exposure history, may have had differing amounts of inheritance, and may have differing chemical compositions), and thus such a comparison should provide an upper limit on the uncertainty due to processing and analysis. The three sample sets showing the best reproducibility (and thus the least likelihood of additional variability due to differences in exposure history) were the Tioga 3 (10 samples), the Tioga 4 ("confluence moraine"; 5 samples), and the Tioga 4 moraine in Humphreys Basin (5 samples). Two clear outliers (91–2 and 90–23) were excluded from the complete Tioga 3 data set; all samples were included in the other two data sets. The Tioga 3 data set yielded a coefficient of variation (1σ) of 3.3%, that for the Tioga 4 confluence moraine was 6.1%, and that for the Humphreys Basin moraine was 3.9%. These values enable an estimate of the random analytical uncertainty at probably less than 5%, and perhaps even less than 4%.

Comparison with ^{36}Cl Chronology Based on Alternative Production-Rate Parameterizations

As mentioned already, a unique parameterization for the production systematics of cosmogenic ^{36}Cl has proved difficult to obtain, probably due to the relatively large number of production reactions that are involved (Gosse and Phillips, 2001). Given this uncertainty, we also calculated ages for the various glacial advances at Bishop Creek using the two most commonly employed alternative sets of production parameters: those published by Stone et al. (1996a, 1996b) and Swanson and Caffee (2001). A comparison of all three sets of ages is shown in Figure 6. The best estimates for the beginning and end of each of the glacial advances, according to all three production schemes, is given in Table 2.

Ages calculated by the parameterization of Stone et al. (1996a, 1996b) are, on average, younger than those calculated according to Phillips et al. (2001) by 11.3% ± 5.5%. For the most part, this difference does not require a major change in the interpretation of the glacial chronology. However, for the three events for which comparison with independent and reliable chronology is possible (end of MIS 6, retreat of Tioga 3, and retreat of Recess Peak), the corresponding ^{36}Cl best-estimate ages are 10%–12% too young.

The ^{36}Cl production parameterization by Swanson and Caffee (2001) produces ages that are markedly younger than either of the alternative parameterizations. The average difference from the ages calculated according to Phillips et al. (2001) is 30.4% ± 5.8% younger. This difference greatly changes the interpretation and correlation of the Bishop Creek glacial ages from that based on Phillips et al. (2001). The Tahoe glaciation no longer correlates with MIS 6, but rather MIS 5. Although a significant MIS 5 advance is not out of the question (Gillespie and Molnar, 1995), it seems rather unlikely that all of the pre-Tioga glacial deposits at Bishop Creek can be attributed to MIS 5 and none to MIS 6. The Tioga 3 retreat and Recess Peak retreat age comparisons are 23% and 28% too young. The age of the Recess Peak (9.5–8.5 ka) is early Holocene rather than latest Pleistocene, as the data of Clark and Gillespie (1997) would indicate.

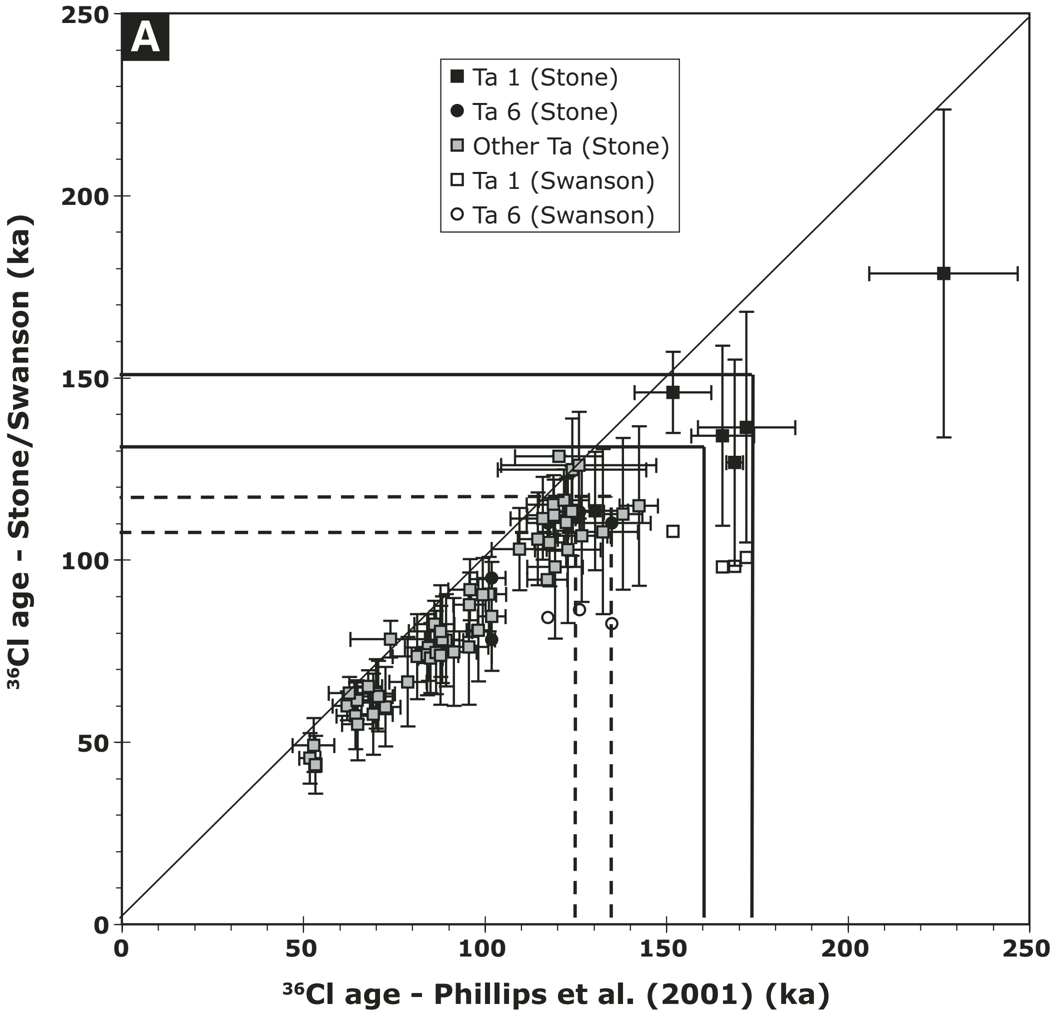

Figure 6. Comparison of ages for (A) Tahoe samples and (B) Tioga samples, calculated according to the ^{36}Cl production parameters of Phillips et al. (2001) with those of Stone et al. (1996a, 1996b) (filled symbols) and Swanson and Caffee (2001) (open symbols). The diagonal line indicates 1:1 correspondence between the ages. For visual clarity, only selected samples calculated according to Swanson and Caffee (2001) are shown, and uncertainty bars are omitted. (*Continued on following page.*)

In summary, the ^{36}Cl production parameterization published by Phillips et al. (2001) appears to give ages that agree best with independent constraints, although they may be slightly (~3%) too young. Those ages calculated according to Stone et al. (1996a, 1996b) are similar, but ~10% younger than independent constraints would indicate. Those based on Swanson and Caffee (2001) appear too young to be geologically reasonable. Although the parameterization of Phillips et al. (2001) apparently gives the best result in this study, this result should not necessarily be generalized. The principal reason for the apparent accuracy of the ages calculated using the Phillips et al. (2001) parameters is probably that many of the samples used for calibrating these parameters were from latitudes and elevations similar to that of the Bishop Creek study area. Cosmogenic nuclide production depends in a complex fashion on altitude, geographical position, and exposure age, as well as the elemental composition of the target rock. Until the effects of all of these factors are adequately constrained, the possibility remains that different parameterizations may inconsistently yield better or worse results depending on the combination of variables associated with a particular set of samples.

CONCLUSIONS

Chronological Significance of Traditional Mapping Units

The classification proposed by Blackwelder (1931) has provided a set of criteria for differentiation of glacial features in the field that has been corroborated by many investigators over the subsequent years. However, Blackwelder did not provide any definitive correlations or absolute ages for his units, and their significance within the framework of regional and global Quaternary climatic events has remained uncertain (Gillespie et al., 1999). A major objective of this study was to assess the relation between this traditional relative classification of glaciations, as applied to the deposits at Bishop Creek, and modern global climate chronology.

The answer to this line of inquiry is startlingly simple: moraines traditionally mapped as Tahoe at Bishop Creek were deposited throughout MIS 6 (the Illinoian) and those mapped as Tioga were deposited during MIS 2 (the Wisconsinan). In fact, the earliest and latest ages for each mapping category closely correspond to the generally accepted ages for the initiation and termination of the global glacial stages. In spite of intensive sampling, no hint was discovered that the mapped Tahoe moraines are a composite that includes earlier or later glacial episodes. Although this result would appear to permit a straightforward one-to-one mapping of the traditional Blackwelder field units onto the major divisions of modern global glacial chronology, it is very likely applicable only within this drain-

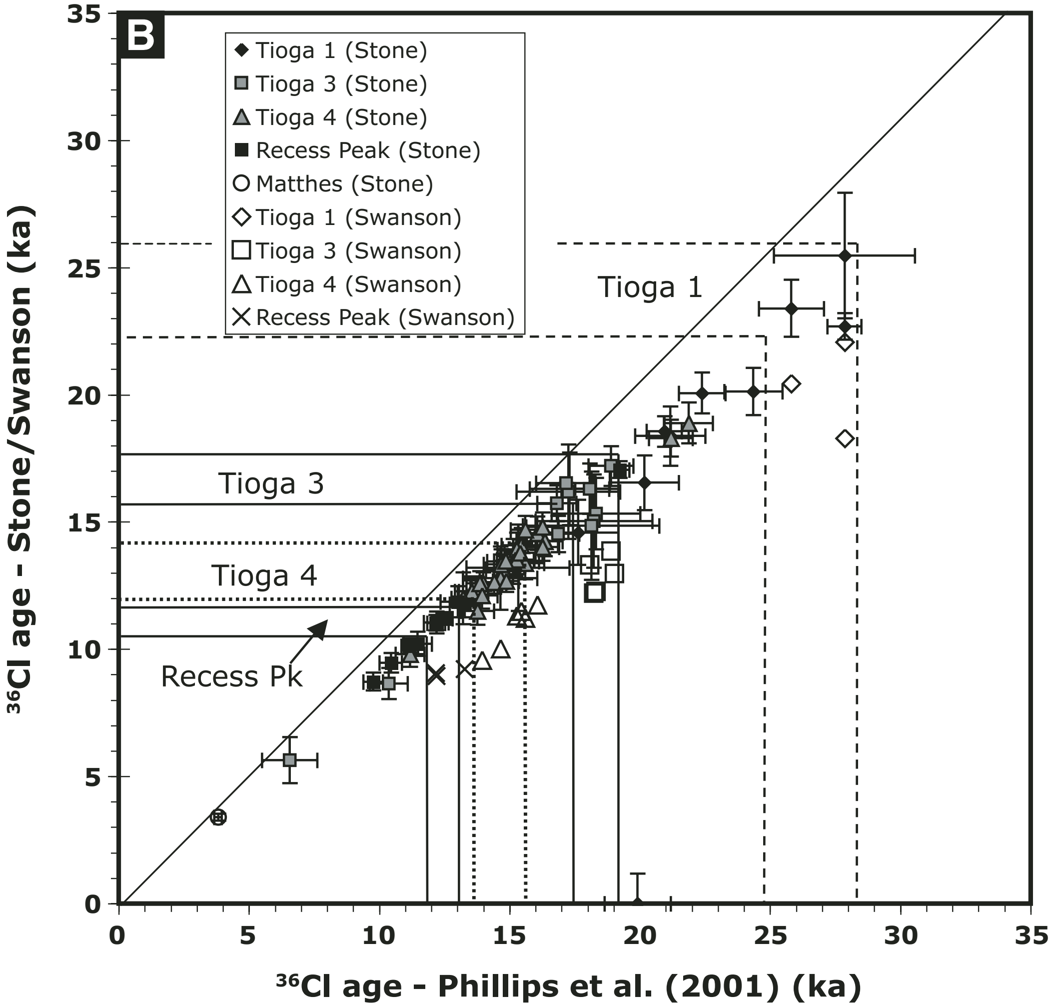

Figure 6 (*continued*).

age, because several types of evidence indicate that moraines elsewhere in the Sierra Nevada were deposited at times that do not correspond to this simplistic correlation (Gillespie et al., 1984; Phillips et al., 1990, 1996a). Obliterative overlap (Gibbons et al., 1984) may well have destroyed the evidence at Bishop Creek for glacial advances at times other than MIS 6 and MIS 2.

Evaluation of Questions Regarding Glacial History

Relative Size of Tahoe and Tioga Moraines

Blackwelder (1931) maintained that the volume of Tahoe terminal moraines was much larger (by as much as a factor of 50) than that of Tioga moraines in the same canyon. Gillespie et al. (1999) countered that Blackwelder conflated moraines of various ages and thus biased his estimates of volume. Ironically, the terminal moraines at Bishop Creek exhibit an even more extreme discrepancy in volume than Blackwelder asserted. The Tioga terminal moraines are probably less than 1/100 the volume of the Tahoe terminal moraines. Our initial hypothesis was that the Bishop Creek moraines had been incorrectly mapped and that the actual moraine volumes do not differ greatly. However, our dating has confirmed that the age assignments are largely correct and that the differential cannot be explained by misidentification. (We note that it is possible that the Tahoe moraines may be no more voluminous than the Tioga, but may be thinly spread over extensive pre-Tahoe moraines so as to completely cover them. This seems unlikely, given the small volume of the Tioga terminal moraines, but it cannot be tested using cosmogenic nuclide dating.)

Our second hypothesis was that the apparent relative volume of the two deposits is real and is explained by the Tioga glaciation being of much shorter duration than the Tahoe. This has found some support from our data. The period of deposition of the Tahoe moraines

TABLE 2. BEST-ESTIMATE UPPER AND LOWER LIMIT AGES FOR THE MORAINE UNITS SAMPLED IN THIS STUDY, ACCORDING TO THE ^{36}Cl PRODUCTION-RATE PARAMETERIZATIONS OF PHILLIPS ET AL. (2001), STONE ET AL. (1996a, 1996b), AND SWANSON AND CAFFEE (2001)

Production parameterization	Phillips et al. (2001) (ka)	Stone et al. (1996a, 1996b) (ka)	Swanson and Caffee (2001) (ka)
Tahoe	170–130	140–120	110–80
Tioga 1	28–24	26–22	22–28
Tioga 3	18.5–17.0	17.5–16.0	14.0–12.0
Tioga 4	16.0–14.5	14.0–12.0	11.5–9.5
Recess Peak	13.4–12.0	12.0–10.5	9.5–8.5

was apparently between 165 and 135 ka, an interval of 30 k.y. The earliest and latest dates for the Tioga moraines near the terminal area are 28 and 18 ka, yielding a deposition interval of only 10 k.y. Although this difference might explain a significant size discrepancy between the terminal moraines, it is not likely to account for the hundredfold difference.

The third hypothesis was that the interval between the Tioga and Tahoe glaciations was shorter than that between the Tahoe and its preceding glaciation, allowing time for greater accumulation of weathering products. We attempted to test this by sampling for pre-Tahoe moraines within the Tahoe terminal complex, thus establishing the interval between the Tahoe and the preceding glaciation. However, our results indicate that the Tahoe terminal complex was entirely deposited during MIS 6. This may be taken to indicate that any such preceding advance was either long before MIS 6, or that it was less extensive and was re-eroded and redeposited by the Tahoe glacier, but it does not provide conclusive evidence. We did obtain an age of 233 ± 19 ka for a single boulder from a lateral moraine well above the terminal area, which could be consistent with an advance during MIS 8, but such a single age provides only weak evidence. The lack of any pre-Tahoe terminal moraines provides some support for this hypothesis, but confirmation from chronological data and an estimate of the extent of previous glaciations is needed.

The difference in the depositional setting of the Tahoe and Tioga terminal moraines has been discussed in detail, and we feel that this difference may have strongly influenced the relative moraine volumes. The Tahoe terminal moraines were deposited on a broad piedmont with a low topographic gradient, which easily accommodated lateral shifts in glacial position. The Tioga glacier, however, avulsed through the Tahoe right lateral moraine and debouched into a narrow canyon where it was confined between the massive Tahoe terminal complex and the bedrock of the Coyote Plateau. The topographic gradient was significantly steepened compared to that during deposition of the Tahoe complex. These factors probably caused most of the sediment load of the melting Tioga glacier to be exported with the outwash, which was focused into a narrow channel with high sediment-transport capacity. Thus, the relative difference in moraine volumes at Bishop Creek may be largely a specific, local effect.

Rate of Retreat of the Tioga Glaciers

Clark (1976), based on geomorphic evidence, hypothesized that the final retreat of the Tioga glaciers had been very rapid. We tested this hypothesis by detailed sampling of erratics and bedrock exposed during deglaciation in order to determine a chronology of retreat. Our results strongly confirm Clark's hypothesis. The rate of retreat of the Tioga 4 glaciers was barely within the time resolution of the ^{36}Cl dating method; probably 500 yr or less, and certainly less than 1000 yr. This implies that the termination of the Tioga glaciation was caused by a warming event that was both very rapid and large in magnitude. Given that the age of this sudden retreat was between 15 and 14.5 ka, it seems reasonable that the climate transition corresponds to the North Atlantic Bølling warming event. This inference is supported by paleotemperature evidence from marine cores taken off the coast of California (Hendy et al., 2002).

Age of the Recess Peak and "Hilgard" Glaciations

Clark and Gillespie (1997) presented evidence that there has been only one significant glacial advance in the Sierra Nevada since the Tioga; the Recess Peak glaciation at ca. 13.5 ka. This interpretation differs significantly from that of previous investigators (Birman, 1964; Burbank, 1991). Our ^{36}Cl ages strongly support the position of Clark and Gillespie. Bedrock and boulder ages above the Tioga 4 terminal moraine are uniformly in the range 15.0–14.5 ka, until Recess Peak moraines are reached. These yield an average age of 12.5 ± 0.7 ka, with oldest ages close to 13.5 ka. Above the Recess Peak moraines, boulder ages are consistent with deposition by the retreating Recess Peak glacier, until clearly neoglacial (Matthes) moraines are reached. There thus can be little doubt that at Bishop Creek, the Recess Peak advance at ca. 13.5 ka was the only significant one since Tioga retreat at ca. 15 ka.

Paleoclimatic Correlations

The glacial deposits at Bishop Creek record a long history of glaciation. The position of degraded moraines indicates that the three forks of Bishop Creek may have conjoined at their current location only in the relatively recent geological past, and that prior to that, the North and Middle Forks may instead have flowed down the present path of Birch Creek. A sequence of lateral moraines, apparently younger as they approach the present course of Bishop Creek, may record a lateral translation, but unfortunately most of these moraines appear to be beyond the current age range for dating.

Chlorine-36 ages show that the classic Tahoe terminal moraine complex, as mapped by Knopf (1918) and Bateman (1965), was deposited between 165 and 130 ka, or entirely within global MIS 6. The Tioga lateral and terminal moraines, as distinguished by the same authors, were entirely deposited during MIS 2, between 28 and 14.5 ka. A detailed history of advances and retreats cannot be distinguished within the Tahoe glaciation, but is possible for at least the later part of the Tioga glaciation. The Tioga glacier initially advanced at ca. 26 ka (Tioga 1). We do not have evidence for events during the following 10 k.y. The glacier stood at its maximum extent at ca. 17.7 ka (Tioga 3) and then rapidly retreated up-valley to at least the confluence of the two forks of Bishop Creek by 17 or 16.5 ka. It then readvanced to the confluence at ca. 15.6 ka (Tioga 4). Shortly after 15.5 ka, it retreated rapidly (within 500–1000 yr) back to the highest cirques.

As previously postulated by Phillips et al. (1996a), these fluctuations appear to have been strongly related to global, and especially North Atlantic, climate events. The beginning of the Tioga 3 retreat appears to have been coincident with the warming period that initiated west European ice-sheet retreat at ca. 17.5 ka, just before Heinrich event 1 (Lagerklint and Wright, 1999; Zaragosi et al., 2001), and the Tioga 4 readvance appears to have been coincident with the peak of Heinrich 1 between 16.9 and 15.8 ka. The Tioga 3 retreat was also synchronous with the global-scale termination of the Last Glacial Maximum at 17.3 ka (Schaefer et al., 2006). The final and dramatic Tioga 4 retreat can very likely be correlated with the dramatic Bølling warming in the North Atlantic. As in northern Europe, this warming appears to have begun closer to 15.5 ka (the end of Heinrich event 1) than to the 14.7 ka date generally accepted for Greenland (Genty et al., 2006). Following this termination of the large-glacier phase in the Sierra Nevada, there was a brief and relatively small Recess Peak advance at ca. 13.3 ka. Surprisingly, this event apparently did not correlate with the Younger Dryas cooling, which is inferred to have affected the ocean waters off the California coast (Hendy et al., 2002), but rather with the Inter-Allerød Cold Period at 13.3 ka, which is apparently also recorded in the Santa Barbara Basin marine sediments. The lack of glaciation during the Younger Dryas may have been related to a decrease in precipitation during that period (Haynes, 1991). Glacier modeling by Plummer and Phillips (2003) indicates that greater-than-modern precipitation is required to simulate Recess Peak glacier limits under a 3 °C temperature reduction. Although there is no clear one-to-one correspondence between North Atlantic climate events and glacial advances in the Sierra Nevada, the associations in timing clearly indicate a strong connection through the global climate system.

ACKNOWLEDGMENTS

Deborah Elliot-Fisk, Andrew Bach, Tanzhou Liu, and Sean Campbell participated in early field work. Lois Phillips provided important field logistic support. Many samples were processed by Terry Thomas, New Mexico Bureau of Geology. Lisa Majkowski-Taylor's assistance in geographic information systems (GIS) and map preparation was invaluable. Alan Gillespie and Daryl Granger significantly improved the paper through thorough reviews. This research was funded by the U.S. National Science Foundation through grants ATM-9117566 and EAR-9417810.

REFERENCES CITED

André, M.-F., 1996, Rock weathering rates in arctic and subarctic environments (Abisko Mts., Swedish Lapland): Zeitschrift fur Geomorphologie Neue Folge, v. 40, p. 499–517.

Aruscavage, P.J., and Campbell, E.Y., 1983, An ion-selective electrode method for determination of chlorine in geological materials: Talanta (Oxford), v. 30, p. 745–749, doi: 10.1016/0039-9140(83)80171-4.

Bach, A.J., 1995, Aeolian modifications of glacial moraines at Bishop Creek, eastern California, *in* Tchakerian, V.P., ed., Desert Aeolian Processes: London, Chapman & Hall, p. 179–197.

Bateman, P.C., 1965, Geology and Tungsten Mineralization of the Bishop District, California: U.S. Geological Survey Professional Paper 470, 208 p.

Benedict, J.B., 1993, Influence of snow upon rates of granodiorite weathering, Colorado Front Range, USA: Boreas, v. 22, p. 87–92.

Benson, L.V., Burdett, J.W., Kashgarian, M., Lund, S.P., Phillips, F.M., and Rye, R.O., 1996, Climatic and hydrologic oscillations in the Owens Lake Basin and adjacent Sierra Nevada, California: Science, v. 274, p. 746–749, doi: 10.1126/science.274.5288.746.

Berry, M.E., 1990, Soil-Geomorphic Analysis of Late-Quaternary Glaciation and Faulting, Eastern Escarpment of the Central Sierra Nevada, California [Ph.D. thesis]: Boulder, University of Colorado, 365 p.

Berry, M.E., 1994, Soil-geomorphic analysis of late-Pleistocene glacial sequences in the McGee, Pine, and Bishop Creek drainages, east-central Sierra Nevada, California: Quaternary Research, v. 41, p. 160–175, doi: 10.1006/qres.1994.1018.

Berry, M.E., 1997, Geomorphic analysis of late Quaternary faulting on Hilton Creek, Round Valley and Coyote Warp faults, east-central Sierra Nevada, California, USA: Geomorphology, v. 20, p. 177–195, doi: 10.1016/S0169-555X(97)00033-0.

Bierman, P.R., and Gillespie, A.R., 1991a, Lowering rates of granitic landforms determined by measurement of in situ produced cosmogenic nuclides: Eos (Transactions, American Geophysical Union), v. 72, no. 44, p. 575.

Bierman, P.R., and Gillespie, A.R., 1991b, Range fires: A significant factor in exposure-age determination and geomorphic surface evolution: Geology, v. 19, p. 641–644, doi: 10.1130/0091-7613(1991)019<0641:RFASFI>2.3.CO;2.

Birkeland, P.W., and Burke, R.M., 1988, Soil catena chronosequences on eastern Sierra Nevada moraines, California, U.S.A.: Arctic and Alpine Research, v. 20, p. 473–484, doi: 10.2307/1551345.

Birman, J.H., 1964, Glacial Geology across the Crest of the Sierra Nevada: Geological Society of America Special Paper 75, 80 p.

Bischoff, J.L., and Cummins, K., 2001, Wisconsin glaciation of the Sierra Nevada (79,000–15,000 yr BP) as recorded by rock flour in sediments of Owens Lake, California: Quaternary Research, v. 55, p. 14–24, doi: 10.1006/qres.2000.2183.

Bischoff, J.L., Menking, K.M., Fitts, J.P., and Fitzpatrick, J.A., 1997, Climatic oscillations 10,000–155,000 yr B.P. at Owens Lake, California, reflected in glacial rock flour abundance and lake salinity in Core OL-92: Quaternary Research, v. 48, p. 313–325, doi: 10.1006/qres.1997.1933.

Blackwelder, E., 1929, Moraines of Convict Lake glacier: Geological Society of America Bulletin, v. 40, p. 171.

Blackwelder, E., 1931, Pleistocene glaciation in the Sierra Nevada and Basin Ranges: Geological Society of America Bulletin, v. 42, p. 865–922.

Blair, T.C., 2001, Outburst flood sedimentation on the proglacial Tuttle Canyon alluvial fan, Owens Valley, California, U.S.A.: Journal of Sedimentary Research, v. 71, p. 657–679, doi: 10.1306/2DC4095E-0E47-11D7-8643000102C1865D.

Bond, G., Heinrich, H., Broecker, W., Labeyrie, L., McManns, J., Andrews, J., Huon, S., Jantschik, R., Clasen, S., Simet, C., Tedesco, K., Klas, M., Bonani, G., and Ivy, S., 1992, Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial period: Nature, v. 360, p. 245–249, doi: 10.1038/360245a0.

Bowerman, N.D., and Clark, D.H., 2005, Onset and culmination of neoglaciation in the Sierra Nevada, California: Geological Society of America Abstracts with Programs, v. 37, no. 7, p. 543.

Briner, J.P., and Swanson, T.W., 1998, Using inherited cosmogenic ^{36}Cl to constrain glacial erosion rates of the Cordilleran ice sheet: Geology, v. 26, p. 3–6, doi: 10.1130/0091-7613(1998)026<0003:UICCTC>2.3.CO;2.

Burbank, D.W., 1991, Late Quaternary snowline reconstructions for the southern and central Sierra Nevada, California, and a reassessment of the "Recess Peak Glaciation": Quaternary Research, v. 36, p. 294–306, doi: 10.1016/0033-5894(91)90004-O.

Burke, R.M., and Birkeland, P.W., 1979, Re-evaluation of multiparameter relative dating techniques and their application to the glacial sequence along the eastern escarpment of the Sierra Nevada, California: Quaternary Research, v. 11, p. 21–51, doi: 10.1016/0033-5894(79)90068-1.

CDWR (California Department of Water Resources), 2006, Snow course data archive (California Data Exchange Center): http://cdec.water.ca.gov/cgi-progs/snowQuery (accessed 2002).

Cerling, T.E., and Craig, H., 1994, Geomorphology and in-situ cosmogenic isotopes: Annual Review of Earth and Planetary Sciences, v. 22, p. 273–317, doi: 10.1146/annurev.ea.22.050194.001421.

Clark, D.H., 1997, A new alpine lacustrine sedimentary record from the Sierra Nevada: Implications for late-Pleistocene paleoclimatic reconstructions and cosmogenic isotope production rates: Eos (Transactions, American Geophysical Union), v. 78, no. 46, p. F249.

Clark, D.H., and Clark, M.M., 1995, New evidence of late-Wisconsin deglaciation in the Sierra Nevada, California, refutes the Hilgard glaciation: Geological Society of America Abstracts with Programs, v. 27, no. 5, p. 10.

Clark, D.H., and Gillespie, A.R., 1997, Timing and significance of late-glacial and Holocene cirque glaciation in the Sierra Nevada, California: Quaternary International, v. 38/39, p. 21–38, doi: 10.1016/S1040-6182(96)00024-9.

Clark, D.H., Clark, M.M., and Gillespie, A.R., 1994, Debris-covered glaciers in the Sierra Nevada, California, and their implications for snowline reconstructions: Quaternary Research, v. 41, p. 139–153, doi: 10.1006/qres.1994.1016.

Clark, D.H., Bierman, P.R., and Larsen, P., 1995, Improving in situ cosmogenic chronometers: Quaternary Research, v. 44, p. 367–377, doi: 10.1006/qres.1995.1081.

Clark, D.H., Gillespie, A.R., Clark, M., and Burke, R., 2003, Mountain glaciation of the Sierra Nevada, *in* Easterbrook, D.J., ed., Quaternary Geology of the United States: International Quaternary Association 2003 Field Guide Volume: Reno, Nevada, Desert Research Institute, p. 287–312.

Clark, M.M., 1976, Evidence for rapid destruction of latest Pleistocene glaciers of the Sierra Nevada, California: Geological Society of America Abstracts with Programs, v. 8, no. 3, p. 361–362.

Elmore, D., Fulton, B.R., Clover, M.R., Marsden, J.R., Gove, H.E., Naylor, H., Purser, K.H., Kilius, L.R., Beukens, R.P., and Litherland, A.E., 1979, Analysis of ^{36}Cl in environmental water samples using an electrostatic accelerator: Nature, v. 277, p. 22–25, doi: 10.1038/277022a0.

Elsheimer, H.N., 1987, Application of an ion-selective electrode method to the determination of chlorine in 41 international geochemical reference materials: Geostandards Newsletter, v. 11, p. 115–122, doi: 10.1111/j.1751-908X.1987.tb00014.x.

Emiliani, C., 1955, Pleistocene temperatures: The Journal of Geology, v. 63, p. 538–578.

Finkel, R., and Suter, M., 1993, AMS in the earth sciences: Technique and applications: Advances in Analytical Geochemistry, v. 1, p. 1–114.

Fullerton, D.S., 1986, Chronology and correlation of glacial deposits in the Sierra Nevada, California, *in* Sibrava, V., Bowen, D.Q., and Richmond, G.M., eds., Quaternary Glaciations in the Northern Hemisphere: Report of the International Geological Correlation Programme, Project 24: Oxford, Pergamon Press, p. 161–169.

Gallup, C.D., Cheng, H., Taylor, F.W., and Edwards, R.L., 2002, Direct determination of the timing of sea level change during Termination II: Science, v. 295, p. 310–313.

Genty, D., Blamart, D., Ghaleb, B., Plagnes, V., Causse, C., Bakalowicz, M., Zouari, K., Chkir, N., Hellstrom, J., Wainer, K., and Bourges, F., 2006, Timing and dynamics of the last deglaciation from European and North African δ^{13}C stalagmite profiles—comparison with Chinese and South Hemisphere stalagmites: Quaternary Science Reviews, v. 25, no. 17–18, p. 2118–2142, doi: 10.1016/j.quascirev.2006.01.030.

Gibbons, A.B., Megeath, J.D., and Pierce, K.L., 1984, Probability of moraine survival in a succession of glacial advances: Geology, v. 12, p. 327–330, doi: 10.1130/0091-7613(1984)12<327:POMSIA>2.0.CO;2.

Gillespie, A.R., 1984, Evidence for both Wisconsin and Illinoian ages for the Tahoe glaciation, Sierra Nevada, California: Geological Society of America Abstracts with Programs, v. 16, no. 6, p. 519.

Gillespie, A.R., 1991, Quaternary subsidence of Owens Valley, California, *in* Hall, C.A., Jr., Doyle-Jones, V., and Widawski, B., eds., Natural History of Eastern California and High-Altitude Research: Proceedings of the White Mountain Research Station Symposium 3: Los Angeles, University of California, p. 356–382.

Gillespie, A.R., and Bierman, P.R., 1995, Precision of terrestrial exposure ages and erosion rates estimated from analysis of cosmogenic isotopes produced in situ: Journal of Geophysical Research, v. 100, p. 24,637–24,649, doi: 10.1029/95JB02911.

Gillespie, A.R., and Molnar, P., 1995, Asynchronous maximum advances of mountain and continental glaciers: Reviews of Geophysics, v. 33, p. 311–364, doi: 10.1029/95RG00995.

Gillespie, A.R., Huneke, J.C., and Wasserburg, G.J., 1984, Eruption age of a 100,000-year old basalt from ^{40}Ar-^{39}Ar analysis of partially degassed xenoliths: Journal of Geophysical Research, v. 89, p. 1033–1048, doi: 10.1029/JB089iB02p01033.

Gillespie, A.R., Clark, M.M., and Burke, R.M., 1999, Eliot Blackwelder and the alpine glaciations of the Sierra Nevada, *in* Moores, E.M., Sloan, D., and Stout, D.L., eds., Classic Cordilleran Concepts: A View from California: Geological Society of America Special Paper 338, p. 443–452.

Gosse, J.C., and Phillips, F.M., 2001, Terrestrial in-situ cosmogenic nuclides: Theory and application: Quaternary Science Reviews, v. 20, p. 1475–1560, doi: 10.1016/S0277-3791(00)00171-2.

Granger, D.E., Riebe, C.S., Kirchner, J.W., and Finkel, R.C., 2001, Modulation of erosion on steep granitic slopes by boulder armoring, as revealed by cosmogenic ^{26}Al and ^{10}Be: Earth and Planetary Science Letters, v. 186, p. 269–281, doi: 10.1016/S0012-821X(01)00236-9.

Hallet, B., and Putkonen, J., 1994, Surface dating of dynamic landforms: Young boulders on aging moraines: Science, v. 265, p. 937–940, doi: 10.1126/science.265.5174.937.

Haynes, C.V., Jr., 1991, Geoarcheological and paleohydrological evidence for a Clovis-age drought in North America and its bearing on extinction: Quaternary Research, v. 35, p. 438–450, doi: 10.1016/0033-5894(91)90056-B.

Hendy, I., Kennett, J.P., Roark, E.B., and Ingram, B.L., 2002, Apparent synchroneity of submillennial scale climate events between Greenland and Santa Barbara Basin, California from 30–10 ka: Quaternary Science Reviews, v. 21, p. 1167–1184, doi: 10.1016/S0277-3791(01)00138-X.

Huber, N.K., 1981, Amount and Timing of Late Cenozoic Uplift and Tilt of the Central Sierra Nevada, California—

Evidence from the Upper San Joaquin River Basin: U.S. Geological Survey Professional Paper 1197, 28 p.

Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W.L., and Shackleton, N.J., 1984, The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $\delta^{18}O$ record, *in* Berger, A., Imbrie, J., Hays, J., Kukla, G., and Saltzman, B., eds., Milankovitch and Climate: Understanding the Response to Orbital Forcing, Part 1: Norwell, Massachusetts, D. Reidel, p. 269–305.

Izett, G.A., and Obradovich, J.D., 1991, Dating of the Matuyama-Brunhes boundary based on $^{40}Ar/^{39}Ar$ ages of the Bishop Tuff and Cerro San Luis rhyolite: Geological Society of America Abstracts with Programs, v. 23, no. 5, p. A106.

James, L.A., Harbor, J., Fabel, D., Dahms, D., and Elmore, D., 2002, Late Pleistocene glaciations in the northwestern Sierra Nevada, California: Quaternary Research, v. 57, p. 409–419, doi: 10.1006/qres.2002.2335.

Jóhannesson, T., Raymond, C., and Waddington, E., 1989, Time-scale for adjustment of glaciers to changes in mass balance: Journal of Glaciology, v. 35, p. 355–369.

Kaufman, D.S., Porter, S.C., and Gillespie, A.R., 2004, Quaternary alpine glaciation in Alaska, the Pacific Northwest, Sierra Nevada, and Hawaii, *in* Gillespie, A.R., Porter, S.C., and Atwater, B.F., eds., The Quaternary Period in the United States, Developments in Quaternary Science Volume 1: Amsterdam, Elsevier Press, p. 77–103.

Kesseli, J.E., 1941, Studies in the Pleistocene glaciation of the Sierra Nevada, California: II. Changes in the courses of some Pleistocene glaciers and their related interglaciation: University of California Publications in Geography, v. 6, no. 8, p. 327–361.

Knopf, A., 1918, A Geologic Reconnaissance of the Inyo Range and the Eastern Slope of the Sierra Nevada, California, with a Section on the Stratigraphy of the Inyo Range by Edwin Kirk: U.S. Geological Survey Professional Paper 110, 130 p.

Konrad, S.K., and Clark, D.H., 1998, Evidence for an early neoglacial glacier advance from rock glaciers and lake sediments in the Sierra Nevada, California, USA: Arctic and Alpine Research, v. 30, p. 272–284, doi: 10.2307/1551975.

Lagerklint, I.M., and Wright, J.D., 1999, Late glacial warming prior to Heinrich event 1: The influence of ice rafting and large ice sheets on the timing of initial warming: Geology, v. 27, p. 1099–1102, doi: 10.1130/0091-7613(1999)027<1099:LGWPTH>2.3.CO;2.

Lal, D., 1991, Cosmic-ray labeling of erosion surfaces: In situ production rates and erosion models: Earth and Planetary Science Letters, v. 104, p. 424–439, doi: 10.1016/0012-821X(91)90220-C.

Matthes, F.E., 1960, Reconnaissance of the Geomorphology and Glacial Geology of the San Joaquin Basin, Sierra Nevada, California (prepared posthumously by F. Fryxell): U.S. Geological Survey Professional Paper 329, 62 p.

Matthes, F.E., 1965, Glacial Reconnaissance of Sequoia National Park, California: Characteristics and Distribution of the Ancient Glaciers in the Most Southerly National Park of the Sierra Nevada (prepared posthumously by F. Fryxell): U.S. Geological Survey Professional Paper 504-A, 58 p.

Menking, K.M., Bischoff, J.L., Fitzpatrick, J.A., Burdette, J.W., and Rye, R.O., 1997, Climatic/hydrologic oscillations since 155,000 yr B.P. at Owens Lake, California, reflected in abundance and stable isotope composition of sediment carbonate: Quaternary Research, v. 48, p. 58–68, doi: 10.1006/qres.1997.1898.

Nichols, K.K., Bierman, P.R., Foniri, W.R., Gillespie, A.R., Caffee, M., and Finkel, R., 2006, Dates and rates of arid region geomorphic processes: GSA Today, v. 16, no. 8, p. 4–11, doi: 10.1130/GSAT01608.1.

Osborn, G., and Bevis, K., 2001, Glaciation in the Great Basin of the Western United States: Quaternary Science Reviews, v. 20, p. 1377–1410, doi: 10.1016/S0277-3791(01)00002-6.

Paterson, W.S.B., 1995, The Physics of Glaciers (3rd edition): Oxford, Pergamon, 480 p.

Phillips, F.M., 2008, Geological and hydrological history of the paleo-Owens drainage since the late Miocene, *in* Reheis, M.C., and Hershler, R., eds., Late Cenozoic Drainage History of the Southwestern Great Basin and Lower Colorado River Region: Geologic and Biotic Perspectives: Geological Society of America Special Paper 439, p. 115–150.

Phillips, F.M., and Plummer, M.A., 1996, CHLOE: A program for interpreting in-situ cosmogenic nuclide data for surface exposure dating and erosion studies [abs.]: Radiocarbon (Abstracts of the 7th International Conference on Accelerator Mass Spectrometry), v. 38, p. 98–99.

Phillips, F.M., Leavy, B.D., Jannik, N.O., Elmore, D., and Kubik, P.W., 1986, The accumulation of cosmogenic chlorine-36 in rocks: A method for surface exposure dating: Science, v. 231, p. 41–43, doi: 10.1126/science.231.4733.41.

Phillips, F.M., Zreda, M.G., Smith, S.S., Elmore, D., Kubik, P.W., and Sharma, P., 1990, Cosmogenic chlorine-36 chronology for glacial deposits at Bloody Canyon, eastern Sierra Nevada: Science, v. 248, p. 1529–1532, doi: 10.1126/science.248.4962.1529.

Phillips, F.M., Zreda, M.G., Benson, L.V., Plummer, M.A., Elmore, D., and Sharma, P., 1996a, Chronology for fluctuations in late Pleistocene Sierra Nevada glaciers and lakes: Science, v. 274, p. 749–751, doi: 10.1126/science.274.5288.749.

Phillips, F.M., Zreda, M.G., Flinsch, M.R., Elmore, D., and Sharma, P., 1996b, A reevaluation of cosmogenic ^{36}Cl production rates in terrestrial rocks: Geophysical Research Letters, v. 23, p. 949–952, doi: 10.1029/96GL00960.

Phillips, F.M., Zreda, M.G., Gosse, J.C., Klein, J., Evenson, E.B., Hall, R.D., Chadwick, O.A., and Sharma, P., 1997, Cosmogenic ^{36}Cl and ^{10}Be ages of Quaternary glacial and fluvial deposits of the Wind River Range, Wyoming: Geological Society of America Bulletin, v. 109, p. 1453–1463, doi: 10.1130/0016-7606(1997)109<1453:CCABAO>2.3.CO;2.

Phillips, F.M., Stone, W.D., and Fabryka-Martin, J.T., 2001, An improved approach to calculating low-energy cosmic-ray neutron fluxes near the land/atmosphere interface: Chemical Geology, v. 175, p. 689–701, doi: 10.1016/S0009-2541(00)00329-6.

Plummer, M.A., 2002, Paleoclimatic Conditions during the Last Deglaciation inferred from Combined Analysis of Pluvial and Glacial Records [Ph.D. thesis]: New Mexico Institute of Mining & Technology, 308 p.

Plummer, M.A., and Phillips, F.M., 2003, A 2-D numerical model of snow/ice energy balance and ice flow for paleoclimatic interpretation of glacial geomorphic features: Quaternary Science Reviews, v. 22, p. 1389–1406, doi: 10.1016/S0277-3791(03)00081-7.

Poreda, R.J., and Cerling, T.E., 1995, Cosmogenic ^{3}He exposure ages of June Lake basalts: Applications to Sierra Nevada glacial chronology [abs.]: Eos (Transactions, American Geophysical Union), v. 76, fall supplement, p. 685.

Porter, S.C., Pierce, K.L., and Hamilton, T.D., 1983, Late Wisconsin mountain glaciation in the western United States, *in* Wright, H.E., Jr., ed., Late Quaternary Environments of the United States: Minneapolis, University of Minnesota Press, p. 71–111.

Putkonen, J., and Swanson, T., 2003, Accuracy of cosmogenic ages for moraines: Quaternary Research, v. 59, p. 255–261, doi: 10.1016/S0033-5894(03)00006-1.

Rahm, D.A., 1964, Glacial geology of the Bishop area, Sierra Nevada, California [abs.], *in* Abstracts for 1963: Geological Society of America Special Paper 76, p. 221.

Russell, I.C., 1889, Quaternary history of Mono Valley, California: Eighth Annual Report of the U.S. Geological Survey: Washington D.C., United States Government Publishing Service, p. 267–394.

Sarna-Wojcicki, A., Pringle, M.S., and Wijbrans, J., 2000, New $^{40}Ar/^{39}Ar$ age of the Bishop Tuff from multiple sites and sediment rate calibration for the Matuyama-Brunhes boundary: Journal of Geophysical Research–Solid Earth, v. 105, p. 21,431–21,443.

Schaefer, J.M., Denton, G.H., Barrell, D.J.A., Ivy-Ochs, S., Kubik, P.W., Andersen, B.G., Phillips, F.M., Lowell, T.V., and Schluchter, C., 2006, Near-synchronous interhemispheric termination of the Last Glacial Maximum in mid-latitudes: Science, v. 312, p. 1510–1513, doi: 10.1126/science.1122872.

Shackleton, N.J., 2000, The 100,000-year ice-age cycle identified and found to lag temperature, carbon dioxide, and orbital eccentricity: Science, v. 289, p. 1897–1902, doi: 10.1126/science.289.5486.1897.

Sharp, R.P., 1968, Sherwin Till–Bishop Tuff geological relationships, Sierra Nevada, California: Geological Society of America Bulletin, v. 79, p. 351–364, doi: 10.1130/0016-7606(1968)79[351:STTGRS]2.0.CO;2.

Sharp, R.P., and Birman, J.H., 1963, Additions to classical sequence of Pleistocene glaciations, Sierra Nevada, California: Geological Society of America Bulletin, v. 74, p. 1079–1086, doi: 10.1130/0016-7606(1963)74[1079:ATCSOP]2.0.CO;2.

Sheridan, M.F., 1971, Guidebook to the Quaternary Geology of the East-Central Sierra Nevada: Phoenix, Arizona, Lebean Printing, Friends of the Pleistocene Guidebook 1971, Pacific Cell, 60 p.

Siddall, M., Bard, E., Rohling, E.J., and Hemleben, C., 2006, Sea-level reversal during Termination II: Geology, v. 34, p. 817–820, doi: 10.1130/G22705.1.

Smith, G.I., and Bischoff, J.L., 1997, An 800,000-year paleoclimatic record from core OL-92, Owens Lake, southeast California: Geological Society of America Special Paper 317, 165 p.

Spötl, C., Mangini, A., Frank, N., Eichstädter, R., and Burns, S.J., 2002, Start of the last interglacial period at 135 ka: Evidence from a high Alpine speleothem: Geology, v. 30, p. 815–818, doi: 10.1130/0091-7613(2002)030<0815:SOTLIP>2.0.CO;2.

Stone, J., Evans, J., Fifield, K., Cresswell, R., and Allan, G., 1996a, Cosmogenic chlorine-36 production rates from calcium and potassium: Radiocarbon, v. 38, p. 170–171.

Stone, J.O., Allan, G.L., Fifield, L.K., and Cresswell, R.G., 1996b, Cosmogenic chlorine-36 from calcium spallation: Geochimica et Cosmochimica Acta, v. 60, p. 679–692, doi: 10.1016/0016-7037(95)00429-7.

Stone, J.O.H., Evans, J.M., Fifield, L.K., Allan, G.L., and Cresswell, R.G., 1998, Cosmogenic chlorine-36 production in calcite by muons: Geochimica et Cosmochimica Acta, v. 62, p. 433–454, doi: 10.1016/S0016-7037(97)00369-4.

Swanson, T.W., and Caffee, M.L., 2001, Determination of ^{36}Cl production rates derived from the well-dated deglaciation surfaces of Whidbey and Fidalgo Islands, Washington: Quaternary Research, v. 56, p. 366–382, doi: 10.1006/qres.2001.2278.

Warhaftig, C., and Birman, J.H., 1965, The Quaternary of the Pacific mountain system in California, *in* Wright, H.E., Jr., and Frey, D.G., eds., The Quaternary of the United States: Princeton, New Jersey, Princeton University Press, p. 299–340.

Winograd, I.J., Landwehr, J.M., Ludwig, K.R., Coplen, T.B., and Riggs, A.C., 1997, Duration and structure of the past four interglaciations: Quaternary Research, v. 48, p. 141–154, doi: 10.1006/qres.1997.1918.

Zaragosi, S., Eynaud, F., Pujol, C., Auffret, G.A., Turon, J.L., and Garlan, T., 2001, Initiation of the European deglaciation as recorded in the northwestern Bay of Biscay slope environments (Meriadzek Terrace and Trevelyan Escarpment): A multi-proxy approach: Earth and Planetary Science Letters, v. 188, no. 3–4, p. 493–507, doi: 10.1016/S0012-821X(01)00332-6.

Zreda, M.G., 1994, Development and Calibration of the Cosmogenic ^{36}Cl Surface Exposure Dating Method and its Application to the Chronology of Late Quaternary Glaciations [Ph.D. thesis]: Socorro, New Mexico Institute of Mining and Technology, 318 p.

Zreda, M.G., and Phillips, F.M., 1994, Surface exposure dating by cosmogenic chlorine-36 accumulation, *in* Beck, C., ed., Dating in Surface Contexts: Albuquerque, University of New Mexico Press, p. 161–184.

Zreda, M.G., Phillips, F.M., and Elmore, D., 1994, Cosmogenic ^{36}Cl accumulation in unstable landforms: 2. Simulations and measurements on eroding moraines: Water Resources Research, v. 30, p. 3127–3136, doi: 10.1029/94WR00760.

Manuscript Received 12 June 2007
Revised Manuscript Received 12 February 2008
Manuscript Accepted 5 September 2008

Printed in the USA