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Authors: Basagic, H. J., and Fountain, A. G.

Source: Arctic, Antarctic, and Alpine Research, 43(3) : 317-330

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1938-4246-43.3.317

Arctic, Antarctic, and Alpine Research, Vol. 43, No. 3, 2011, pp. 317–330

Quantifying 20th Century Glacier Change in the Sierra Nevada, California

H. J. Basagic† and*
*A. G. Fountain**

*Department of Geology, Portland State University, P.O. Box 751, Portland, Oregon 97207-0751, U.S.A.
†Corresponding author: basagic@gmail.com

Abstract

Numerous small alpine glaciers occupy the high elevation regions of the central and southern Sierra Nevada, California. An inventory based on 1:24,000 topographic maps revealed 1719 glaciers and perennial snowfields for a total area of 39.15 ± 0.13 km^2. The number of 'true' glaciers, versus non-moving ice, is estimated to be 122 covering 14.89 ± 0.08 km^2 or 38% of the ice-covered area. Historic photographs, geologic evidence, and field mapping were used to determine the magnitude of area change over the past century at 14 glaciers. The area change between 1903 and 2004 ranged from −31% to −78%, averaging −55%. Based on these values rough estimates of volume change suggest an ice volume loss from 1903 (1.09 km^3) to 2004 (0.43 km^3) of 0.66 km^3 (0.59 km^3 water). Rapid retreat occurred over the first half of the 20th century beginning in the 1920s and continued through the 1960s after which recession ceased by the early 1980s and some glaciers advanced. Since the late 1980s glaciers resumed retreat with a rapid acceleration starting in the early 2000s. The relatively uniform timing of area changes in the study glaciers is a response to regional climate whereas the magnitude of change is influenced by local topographic effects. Area changes correlate significantly with changes in summer and winter air temperatures. Warmer winter temperatures warm the snowpack lengthening the summer melt season. Spring air temperatures and precipitation may also play an important role. The occurrence of spring snowfall can delay the onset of melt due to the increased surface albedo. Examining the influence of topographic variables we only found headwall height at the top of the glacier to show an influence on glacier change. Higher headwalls shadow the glacier from solar radiation reducing melt and enhancing snow accumulation via avalanching. If the glaciers continue to shrink at current (1972–2004) rates, most will disappear in 50–250 years.

DOI: 10.1657/1938-4246-43.3.317

Introduction

Alpine glaciers respond to changes in climate and are important indicators of climate change (e.g. Meier, 1965; Porter, 1981). As such they play an important role in reconstructing paleoclimate (Porter and Denton, 1967; Oerlemans, 2005), and in understanding variations in alpine climate. Few historic climate records exist for alpine regions because they are remote, difficult to access, and the weather typically extreme, making it difficult to monitor these environments. Therefore knowledge of changes in glacier area, preserved in landscape modifications, photographs, and maps are proxy records of past climates. Over the 20th century, with few exceptions, alpine glaciers have been receding globally in response to a warming climate (Kaser, 1999; Dyurgerov and Meier, 2000; Oerlemans, 2005). The mass wasting of these glaciers is important to alpine hydrology (Clow et al., 1996; Brittain and Milner, 2001; Dougall, 2007) and to global sea level rise (e.g. Meier, 1984; Meier et al., 2007). Glaciers delay peak runoff from spring to summer, when less water is available and demand is high, and also reduce variations in seasonal runoff (Fountain and Tangborn, 1985; Moore et al., 2009). Glacier shrinkage reduces their buffering capacity resulting in earlier spring runoff and drier summer conditions. Glacier melt provides water to streams long after the snowpack has melted, a particularly important process in low snow years.

In this report, we examine the glaciers of the Sierra Nevada, California, compile an inventory, quantify the magnitude and rate of change in glacier extent, and examine the climate drivers. While knowledge of glacier shrinkage in the Sierra Nevada is commonly known, little quantitative information exists on the magnitude or rate of glacier shrinkage. Such information is important for assessing current changes in alpine hydrology and to complement other data sets monitoring the effects of climate change in this region.

Study Area

The Sierra Nevada is situated along the eastern edge of the state of California. The range stretches about 640 km, varies in width from 65 to 130 km with peak elevations >4000 m. The range rises gradually eastward from the Central Valley to the Sierra crest where elevations drop quickly into the high desert valleys of California and Nevada. The high alpine climate along the Sierra crest is characterized by cold (−6 °C), wet (750 mm weq) winters, and warm (9 °C), dry summers (25 mm weq) (PRISM, 2006; CDEC, 2009). Precipitation is controlled by the migration of the Pacific High pressure system, such that in summer, the Pacific High moves north, steering westward-moving storms northward (NOAA, 1985; Major, 1990). Summer precipitation is from convective storms with a moisture source in the Gulf of Mexico

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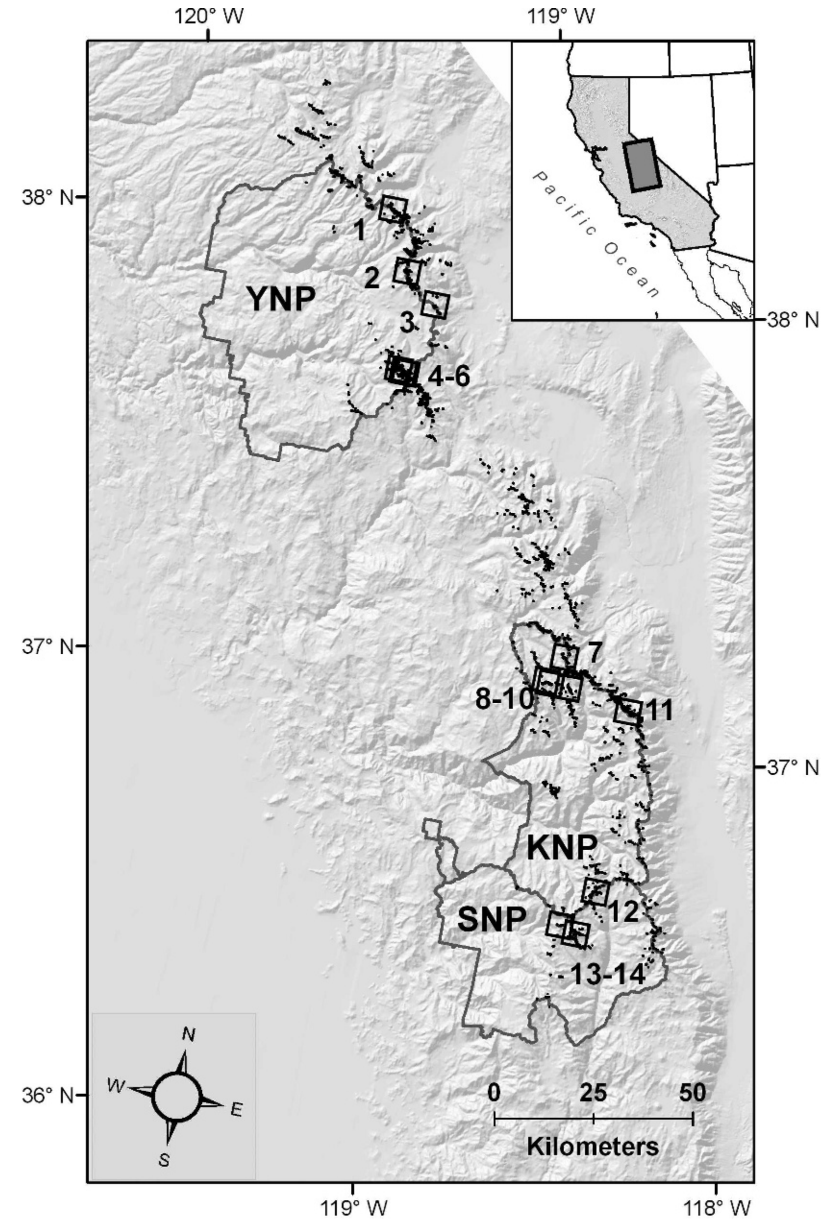

FIGURE 1. Glacier inventory results with glacier population shown in dark gray. Gray lines indicate park boundaries: YNP = Yosemite National Park, KNP = Kings Canyon National Park, SNP = Sequoia National Park. The numbered boxes indicate the location of 14 glaciers listed in Table 2 for assessing long-term climate change.

and Gulf of California. In winter, the Pacific High moves south, enhancing westerly flow and moisture into the Sierra Nevada, which present an effective barrier to westerly flow resulting in heavy precipitation on the western slope and creates a rain shadow to the east.

Numerous small alpine glaciers (<1 km^2) occupy mountain cirques in the central and southern Sierra Nevada, between 36.38° and 38.86°N latitude (Fig. 1) at high elevations (mean: 3575 m; min: 2763 m, max: 4267 m) (Raub et al., 2006). Typically, these glaciers occur on north- and northeast-facing slopes because average equilibrium line altitudes exceed peak topographic elevations, ~4500 m (Flint, 1957), suggesting the importance of local topography modifying atmospheric conditions to favor glacial environments.

The glacier record for the late Pleistocene is well known in the Sierra Nevada (summarized in Warhaftig and Birman, 1965; Clark et al., 2003). Multiple glacier advances occurred throughout the late Pleistocene with the Last Glacial Maximum dated at 22 ± 1 ka B.P. (Phillips et al., 1996), and in the northwestern Sierra at 18.6 ± 1.2 ka B.P (James et al., 2002). Glaciers of the late Pleistocene Recess Peak advance (Birman, 1964), 14.2 and 13.1 ka B.P. (Clark and Gillespie, 1997), were less extensive than earlier Pleistocene advances, with glaciers reaching lengths of about 2 km. The Sierra Nevada may have been free of glaciers between 10.5–5.4 ka B.P. and 4.8–3.2 ka B.P. (Konrad and Clark, 1998; Bowerman, 2006). Following the onset of Neoglaciation at 3.2 ka B.P. there were several glacial maxima at ~2200, ~1600, ~700, and ~250–170 yr B.P. (Bowerman, 2006). The "Little Ice Age" (LIA) (Matthes, 1939) generally refers to the cool period 700 to 100 yr B.P. when glaciers advanced throughout the northern hemisphere (Grove, 1988). The most recent advance, sometimes referred to as the "Matthes Glaciation" (Birman, 1964), reached a glacial maximum between 100 and 200 yrs B.P. and is preserved by the "Matthes moraines" (Matthes, 1940; Birman, 1964; Clark and Gillespie, 1997).

Previous Work

Glaciers were first identified in the Sierra Nevada in 1871 (Muir, 1873) and were confirmed the following summer (LeConte, 1873). In the following decades more observations were recorded and the first maps that depicted glacier extents were drawn (Russell, 1885). The first glacier change study (1883–1903) noted little change in position of Lyell Glacier over the 20-year period (Gilbert, 1904). However, Curry (1969), using historic photographs, snow records, and written accounts of the late 1800s, showed that several glaciers, including Lyell Glacier, retreated from their Matthes moraines after Russell's visit in 1883 and rapidly readvanced by the time Gilbert arrived in 1903. Glacier retreat was first noted in 1919 at Lyell Glacier (Farquhar, 1920) and continued to retreat through the 1930s (Matthes, 1940). By 1931 the National Park Service initiated annual glacier observations at several glaciers in Yosemite National Park (YNP), which included photographs, surface profiles, and terminus position relative to a fixed marker, usually a boulder (Harwell, 1931). The observations continued until 1960 after which they were intermittent to 1975 (White, 1976). Although several YNP reports indicated thickening in the upper portions of glaciers in the 1940s, and confirmed by repeat photographs of Lyell Glacier (Harrison, 1956), glacier retreat continued throughout the 1950s and early 1960s (Hubbard, 1954; YNP, 1960). During the early 1970s glacier area decreased slightly compared to the 1960s (White, 1976). The only published glacier mass balance data were collected at Maclure Glacier for six years beginning in 1967 (Dean, 1974; Tangborn et al., 1977). Following the mid-1970s, the observational record becomes scarce. A repeat photograph survey conducted in 1987 of Lyell, Maclure, Dana, and Conness glaciers found the glacier extent to be similar to that in 1975 (Hardy, unpublished).

A number of partial glacier surveys have been completed in the Sierra Nevada. Kehrlein (1948) identified 49 glaciers and 12 'dying remnants' and Meier (1961) reported 70 glaciers, 69 of which had areas <0.05 km^2. The first comprehensive inventory of Sierra Nevada glaciers was completed by Raub et al. (2006). They used 1972 aerial photography and, relying on their glaciological expertise, identified 497 "glaciers" with surface areas ranging between 0.01 and 1.58 km^2. They defined a glacier as having a minimum area of 0.01 km^2 and "*any perennial ice exhibiting one or more of the following: (1) snow or ice accumulated over several years, (2) a crevasse, (3) heavily debris-covered ice which exhibits evidence of flow, and (4) moraines or trim lines.*" An additional 788 'ice patches' were identified; features >0.005 km^2 which did not fit the glacier definition and which include snowdrifts, snowfields, and ice and snow patches (Raub et al., 2006). Our inventory differs from previous analog studies in that we created a digital database at a finer scale as compared to paper maps, which allowed for precise mapping, and topographic analysis.

Methods

We conducted a glacier inventory of the Sierra Nevada as part of a comprehensive effort to document glacier cover in the Western U.S.A., exclusive of Alaska (Fountain et al., 2007). The inventory is based on U.S. Geological Survey (USGS) 1:24,000-scale topographic quadrangle maps produced from aerial photographs. For the Sierra Nevada, the photographs were acquired between 1975 and 1984. Glaciers and perennial snowfields are depicted on the maps as white polygons with blue perimeters and contour lines, and in California, do not distinguish between glaciers and perennial snowfields. Digital hydrographic line data (e.g. rivers, lakes, glaciers) were acquired from the U.S. Forest Service. The line data were converted from their native format to shapefile polygons and imported into a geographic information system (GIS) database. Glacier polygons were extracted using an assigned numerical code within the source database. The glacier polygons were overlaid onto the digital image files of the maps to correct errors of shape, omission, and misidentification. Once corrected, a number of values were calculated for each polygon, including location (latitude and longitude of centroid) and area. The polygons were overlaid on a 30 m digital elevation model, or DEM (USGS, 1998), and the topographic characteristics of elevation, aspect, and slope were calculated. Detailed summary of the methods can be found in Fountain et al. (2007).

We calculated the uncertainty of the polygon area based on the horizontal accuracy for 1:24,000 maps (12.2 m) (USGS, 1999). A GIS buffer has been commonly used to quantify the positional uncertainty of the polygon area (e.g. Nylen, 2004; Granshaw and Fountain, 2006; Jackson and Fountain, 2007). However this method assumes complete correlation of the errors among all vertices of a polygon, resulting in unnecessarily large estimate of the uncertainty, especially for very small glaciers where the buffer area can be equivalent to the glacier area. Instead, we applied a simplified coordinate method for calculating the glacier polygon area uncertainty as used by Hoffman et al. (2007), based on Ghilani (2000). This method calculates the polygon uncertainty in an area equivalent square. While this method ignores the polygon's shape and the covariance of the error between vertices, the method generates results comparable to more rigorous techniques and is independent of the number of polygon vertices (Ghilani, 2000).

To estimate the population of 'true' glaciers, perennial snow and ice that moves, we estimate whether the feature may move. The threshold for movement is defined as whether the shear stress at the base of the feature exceeds the critical shear stress for ice, 10^5 Pa (Cuffey and Paterson, 2010). The basal shear stress is given by

$$\tau_b = \rho_i g h \sin\gamma \quad (1)$$

where ρ_i is ice density (900 kg m^{-3}), g is gravitational acceleration (9.81 m s^{-2}), h is ice thickness (m), and, γ is the ice surface slope. The surface slope was one of the topographic characteristics derived by overlaying the polygons on the DEM. We used the maximum slope of all grid cells within the feature boundary as a conservative estimate. The thickness of the feature at any one location is unknown; however, we can estimate the average thickness of the feature, $<h>$, from the known feature area and its estimated volume. A scaling relation between feature area and volume is used to estimate volume (Chen and Ohmura, 1990; Bahr et al., 1997).

$$V = \alpha A^{\beta} \quad (2)$$

where, V is glacier volume (km^3), α and β are constants, and A is the glacier area (km^2). Dividing (2) by A yields

$$\langle h \rangle = \alpha A^{\beta - 1} \quad (3)$$

The parameters of α and β may be derived from empirical data or theoretical considerations. Chen and Ohmura (1990) empirically determined the parameters based on 63 alpine glaciers from various regions in the northern hemisphere. Their study included only 11 glaciers <1 km^2, with the smallest 0.08 km^2. Several small cirque glaciers were included from the Western U.S.A. that included Maclure Glacier in the Sierra Nevada. Alternatively, Bahr et al. (1997) determined the values of the parameters from theoretical considerations. These methods result in a large uncertainty in glacier volume, as much as 50% (Granshaw and Fountain, 2006), and we apply different parameters to roughly estimate the uncertainty. Equation (2) can also be used to total estimated ice volume for the population of glaciers and perennial snowfields. In this case compensating errors reduce the uncertainty. We do not include an uncertainty with volume estimates due to the deviation from the regression of individual glaciers, uncertainty in the method (small sample size) for small glaciers, and compensating errors when applying the method to a large population of glaciers and perennial snowfields. Instead, we qualify our volume estimates as 'rough' estimates. We do not report estimated volumes for individual glaciers given the large uncertainty described by Granshaw and Fountain (2006).

To examine glacier change ideally, one would draw on a record of glacier mass balance over time, which is the key process between climate forcing and glacier change (Meier, 1965; Cuffey and Paterson, 2010). Unfortunately, such data are nonexistent in the Sierra Nevada except for a six-year record in the late 1960s to early 1970s (Dean, 1974; Tangborn, et al., 1977). Instead, we assess changes in glacier extent based on a photographic record starting in the late 1800s and use area changes as a proxy for mass changes. We focus on a subset of glaciers that have abundant historical photographs and minimal rock debris covering the ice surface. Rock debris obscures the glacier margins and the insulating properties of debris alters the glacial response to climate change (Konrad and Humphrey, 2000).

Historic ground-based and aerial photographs including digital orthophotograph quadrangles (DOQ) were collected from a number of different sources with much of the material from U.S. government agencies (Appendix). The DOQs were particularly important because they include encoded positional data that were used to rectify aerial photographs from other years. The seasonal timing of the photographs was important because residual seasonal snow hides glacier boundaries, making delineation difficult, if not impossible. Therefore, we only used photographs taken in late summer when seasonal snow cover was minimal. Contrast, color, and texture were helpful to differentiate between snow, ice, and rock. The earliest glacier extents were defined by the location of the Matthes moraines. As mentioned previously, the early photographs showed the glaciers in contact with the moraines in 1903 (Gilbert, 1904). We assumed that all glaciers in the Sierra Nevada were in contact with the moraines at this time and we used the crest of the moraine to infer its position. The moraine crest was digitized from the earliest aerial photographs to reduce uncertainty caused by erosion.

Glacier extents were derived from vertical aerial and ground-based photographs, and measured with global positioning system (GPS). The aerial photographs were georeferenced to the DOQ imagery using ArcGIS (ESRI, Inc) software. Seven to ten control points, typically boulders and bedrock features, were used to georeference the photographs. Control points were chosen close to each glacier, especially near the glacier terminus where change is typically the greatest. Once a photograph was georeferenced, the glacier perimeter was digitized to create a polygon for that year. Ground-based photographs provided some temporal detail but were problematic because of scale, perspective, and inherently large uncertainties from such an oblique angle. Therefore we only used them to confirm early glacier contact with the moraines and in situations where the glacier extent could be reasonably constrained within 10 m by other temporal glacier boundaries established from aerial photographs.

The most recent glacier perimeters were measured in the field using GPS. A Garmin Legend GPS unit was used in 2003 and a more accurate Trimble Geo3 GPS in 2004; with accuracies of ± 5 m and ± 2.5 m, respectively. In those cases where the glacier boundary was impossible to define because of rock debris, we defined the glacier edge where ice could no longer be observed. During this field work we photographed the glaciers from the same locations as Gilbert in 1903 (Gilbert, 1904), in 1908 (Gilbert, 1908), and Russell in 1883 (Russell, 1885) to provide a qualitative record of glacier change following methods outlined by Harrison (1960) and Klett et al. (1984). The GPS coordinates of each photographic location were recorded for future reference.

The planimetric area was calculated for each glacier extent (polygon) within a GIS. The area uncertainty was estimated for each glacier extent using the simplified coordinate method used by Hoffman et al (2007), described previously. Calculation of the positional accuracy was dependent on the data source. For aerial photographs, the uncertainty included the positional uncertainty of the original DOQ imagery and the georeferencing error. For terrestrial ground-based photographs, we estimated the uncertainty based on the distance between glacier depictions, typically 10 m, and combined this error with the positional error of the DOQ imagery. For the GPS data, we used the positional accuracy of the GPS unit to estimate the area uncertainty.

Results

GLACIER INVENTORY

Comparison of the digital polygon data and the map images revealed significant errors in the digital data files. These errors included 53 glaciers missing, over 200 hydrographic features (e.g., lakes) were mistakenly coded as glaciers, and 9 polygons showed uniform glacier cover but the source maps showed bedrock 'islands' within the glacier boundary. These errors, with examples, are summarized in Fountain et al. (2007) and were corrected before completing the inventory.

For the 9-year period from 1975 to 1984, 1719 glaciers and perennial snowfields were identified with 13 named glaciers (Fig. 1). Areas ranged from 0.0007 ± 0.0005 to 0.8352 ± 0.0192 km^2, a mean of 0.02 km^2, and total ice-covered area of 39.15 ± 0.13 km^2. The largest glacier is Palisade Glacier (0.8352 ± 0.0192 km^2), but the population is dominated by small features, 90% of which are <0.05 km^2, but account for 19.54 km^2 (50%) of the total ice-covered area (Fig. 2). The mean elevations of the glaciers and perennial snowfields ranged from 2739 m to 4263 m, with a population mean of 3419 m. Mean elevation increases with decreasing latitude, about 480 m over 238 km, a rate of 2.02 m km^{-1} (Fig. 2d). Several anomalously low elevation features at $<$2800 m were examined in the central Sierra and determined to be perennial snow in small protected topographic niches. Most features (64%) face north to northeast, less than 4% face south, and mean slope was 28°.

The number of glaciers and perennial snowfields identified is much larger than the previous inventory by Raub et al. (2006), who identified 1285. However, if we match criteria between inventories by removing features <0.005 km^2, that Raub et al. (2006) did not count, our estimate decreases to 1313 and only 28 (2%) more than Raub et al. (2006). The total area of these features in our inventory is 37.9 ± 6.9 km^2, or 2.7 km^2 (8%) greater than Raub et al. (2006) and well within our uncertainty. The number of 'true' glaciers were estimated from the shear stress criterion (Equations 1–3) and Chen and Ohmura (1990) values for α and β of 0.028524, 1.357 for their entire data set. Results yielded a total of 38 features, 2% of the population (Table 1). Other Chen and Ohmura (1990) values for different alpine regions yielded similar results, except those from "Cascades, small glaciers," a region north of the Sierra Nevada (α and β of 0.021346, 1.145, based on 15 glaciers), which yielded 122 'true' glaciers. We consider this estimate to be more accurate because visual inspection of the 122

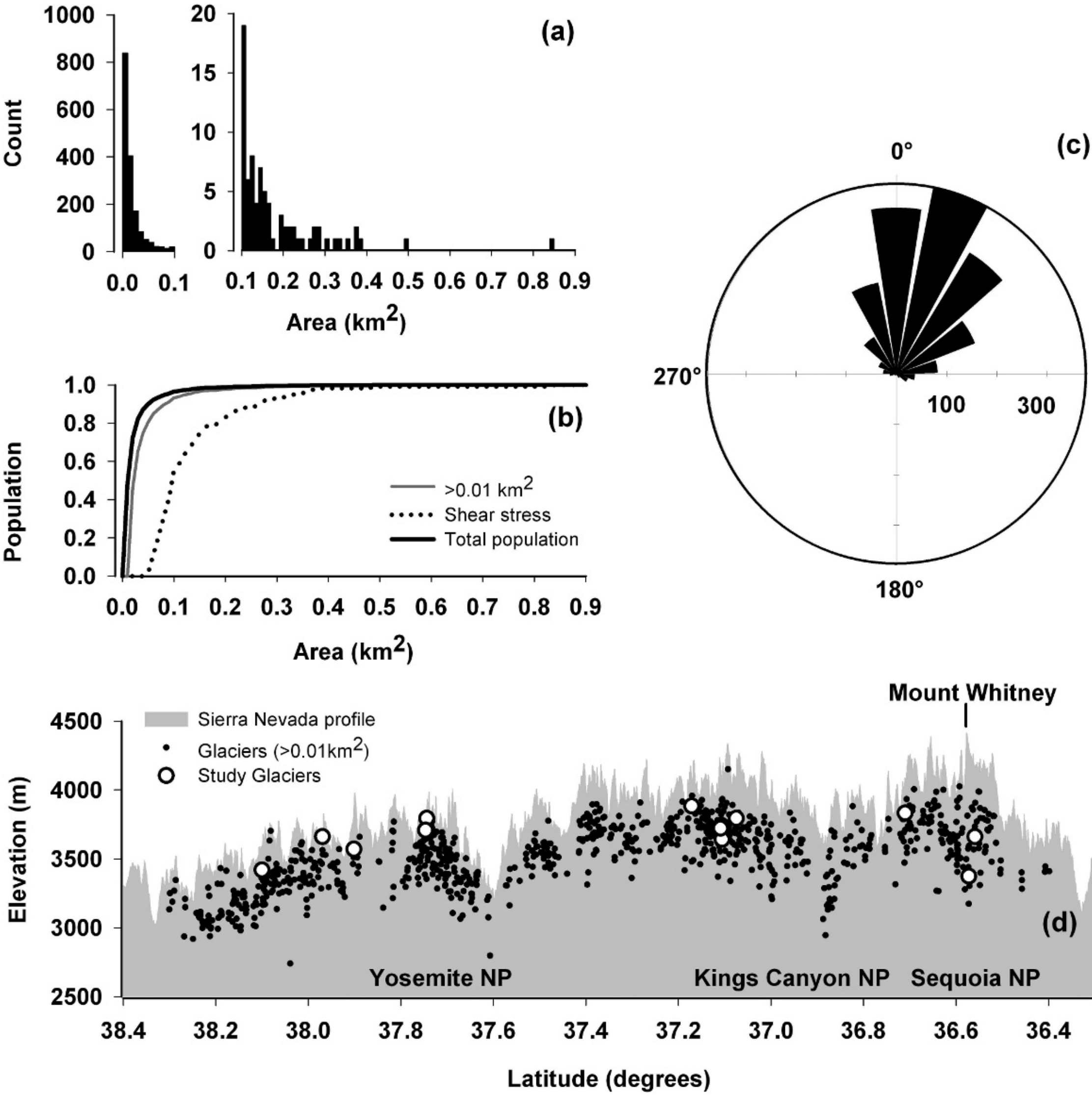

FIGURE 2. Characteristics of the population of glaciers and perennial snowfields in the Sierra Nevada, California: (a) population as a function of area (km^2), note change in *y*-axis for glaciers >0.01 km^2; (b) cumulative sum with area for the total population, for glaciers >0.01 km^2, and for the population of features with an estimated basal shear stress >10^5 Pa; (c) total population as a function of aspect, using 20° bins; and (d) the north-to-south distribution of the glacier population >0.01 km^2 (black dots) and study glaciers (white circles) overlaid on elevation of the Sierra Nevada (gray).

features using aerial photography included 40 glaciers that exhibited crevasses (a sign of motion), none of which were included in the original scaling values. The total area of the 122 'true' glaciers is 14.89 ± 0.08 km^2 representing 38% of the total ice-covered area. Individual glacier areas ranged from 0.02 to 0.84 km^2 with a mean of 0.12 km^2. Using an area threshold of >0.01 km^2, a common definition used to define a glacier (Raub et al., 2006; Paul et al., 2009), 881, or 51% of the total population, would be considered glaciers, with a total area of 34.76 ± 0.12 km^2 and a mean glacier area of 0.04 km^2. The estimated volume of ice during this 9-year period is 0.55 km^3, based on the parameters for "Cascades, small glaciers." Using the parameters for the entire

TABLE 1

Topographic characteristics of the glaciers and perennial snowfields of the Sierra Nevada, based on aerial imagery acquired during the period 1975–1984. 'All' refers to the entire population of features in the inventory; 'Area' refers to those features larger than 0.01 km^2; and 'True Glaciers' are those features considered to be a glacier because the shear stress is estimated to equal or exceed the critical threshold of 10^5 Pa. Two sets of parameters are shown for the area-volume scaling relations. Values for elevation, slope, and aspect are arithmetic means. The last set of 'True' glaciers is considered more accurate.

Criteria	Count	Elevation (m)	Slope (deg)	Aspect (deg)	Mean area (km^2)	Total area (km^2)
All	1719	3419	28	19	0.0228	39.15 ± 0.13
Area >0.01 km^2	881	3524	28	18	0.0395	34.76 ± 0.12
'True' Glaciers (α,β of 0.028524, 1.357)	38	3672	28	21	0.2379	9.04 ± 0.06
'True' Glaciers (α,β of 0.021346, 1.145)	122	3638	30	24	0.1221	14.89 ± 0.08

TABLE 2

Spatial and topographic variables and area change for 14 glaciers. The 7 glaciers in bold are a subset used for detailed temporal changes. Names in italics are informal glacier names. Glacier topographic values are based on 2004 glacier extents, except Maclure and Dana (2003), Middle Palisade and Brewer (2005), and Dragtooth (2006). Elev = elevation, Asp = aspect, Hw = headwall height (meters).

				Topographic				Area			Percent
	Glacier	Latitude	Longitude	Elev (m)	Slope (deg)	Asp (deg)	Hw (m)	1903 (km^2)	2004 (km^2)	change (km^2)	Change
1	*Dragtooth*	38.1009	−119.3838	3419	38	7	162	0.2094 ± 0.0042	0.0548 ± 0.0021	−0.1546 ± 0.0055	−74
2	**Conness**	37.9695	−119.3192	3660	26	16	171	0.3339 ± 0.0046	0.1600 ± 0.0002	−0.1739 ± 0.0046	−52
3	*Dana*	37.9013	−119.2173	3569	30	30	190	0.3193 ± 0.0045	0.1153 ± 0.0004	−0.2040 ± 0.0063	−64
4	Maclure	37.7465	−119.2806	3707	27	352	165	0.2930 ± 0.0060	0.1553 ± 0.0004	−0.1377 ± 0.0061	−47
5	***West Lyell***	37.7439	−119.2640	3794	22	357	85	0.4643 ± 0.0082	0.2795 ± 0.0003	−0.1849 ± 0.0076	−40
6	***East Lyell***	37.7430	−119.2722	3707	30	3	44	0.6334 ± 0.0096	0.1402 ± 0.0002	−0.4932 ± 0.0089	−78
7	**Darwin**	37.1712	−118.6768	3882	33	359	238	0.2536 ± 0.0042	0.1172 ± 0.0002	−0.1365 ± 0.0056	−54
8	*Middle Goddard*	37.1091	−118.7129	3721	35	5	155	0.1106 ± 0.0057	0.0347 ± 0.0001	−0.0759 ± 0.0037	−69
9	***Goddard***	37.1075	−118.7030	3742	31	358	153	0.3787 ± 0.0023	0.1734 ± 0.0002	−0.2053 ± 0.0069	−54
10	*Black Giant*	37.1056	−118.6470	3640	37	48	305	0.1120 ± 0.0031	0.0688 ± 0.0003	−0.0432 ± 0.0037	−39
11	Middle Palisade	37.0751	−118.4679	3792	31	38	261	0.3071 ± 0.0051	0.1734 ± 0.0038	−0.1336 ± 0.0073	−44
12	*Brewer*	36.7100	−118.4820	3832	35	16	123	0.1456 ± 0.0035	0.0418 ± 0.0019	−0.1037 ± 0.0046	−71
13	**Lilliput**	36.5723	−118.5532	3373	34	336	261	0.0699 ± 0.0021	0.0485 ± 0.0001	−0.0214 ± 0.0029	−31
14	***Picket***	36.5593	−118.5082	3660	26	16	113	0.1555 ± 0.0030	0.0798 ± 0.0002	−0.0757 ± 0.0044	−49
	14 glacier mean			3678	31	10	173	0.2705	0.1173		
	7 glacier mean			3688	29	1	152	0.3270	0.1427		
	14 glacier total							3.7863 ± 0.0212	1.6427 ± 0.0048	−2.1436 ± 0.0218	−55
	7 glacier total							2.2893 ± 0.0163	0.9986 ± 0.0006	−1.2909 ± 0.0163	−56

data set, the volume is smaller, 0.42 km^3. If we assume that only features >0.01 km^2 are to be included, the volume based on "Cascades, small glaciers" parameters yields 0.51 km^3.

GLACIER CHANGE

We identified 14 glaciers that had sufficient data to define the glacier extent for two time periods at least 100 years apart (Table 2, Fig. 1). The relatively large and well known Palisade Glacier was not included because the terminus is extensively covered with rock debris. Lyell Glacier forms two separate lobes, each with distinct area-elevation distributions (not shown) that do not interact dynamically. We treat each lobe, East Lyell and West Lyell, separately. Details of the data sources are summarized in the Appendix. Mapping the 1903 extents using the ridgeline of the Matthes moraines was difficult due to the deflation of the moraines, a probable result of their ice cores melting (Beatty, 1939). We too observed ice present in the Matthes moraines of Conness and Goddard glaciers in 2004. The ridgeline was estimated from a combination of the earliest aerial photographs available and field mapping. Rock debris on the glacier surface made field mapping of the boundary on Conness, West Lyell, and Goddard glaciers difficult. We also had problems with obtaining suitable aerial photographs with low snow cover. This was a significant problem in the 1980s and 1990s due to a combination of above average winter snow accumulation and early season aerial photography. For the aerial photographs found suitable, relative ground control was abundant limiting the georectification uncertainty (root mean squared error) to 4.9–12.1 m.

The selected glaciers have a larger mean area and higher elevation (+259 m) compared to the glacier population. The topographic characteristics of the study glaciers are more similar to those of the 'true' glaciers than to the total population. All 14 glaciers decreased in area over the past century with a total loss of 2.14 ± 0.02 km^2 (55%) since 1903 (Table 2). The magnitude of glacier change can be qualitatively appreciated using ground-based photographic comparisons (Fig. 3). Individual losses ranged from 0.0214 ± 0.0029 km^2 (31%) at Lilliput Glacier to 0.4932 ± 0.0089 km^2 (78%) at the East Lyell Glacier, with an average of 0.15 km^2. Four of the 14 glaciers were located on the east side of the Sierra crest. No significant difference in the magnitude of area change was found between the east and west sides of the crest, although the east side had a slightly higher mean area loss of 59% versus 53%.

Within the set of 14 glaciers a subset of 7 had at least three estimates of area over the 100-year period and we used this subset to examine glacier variation over time. The 7 glaciers showed that no change in glacier area appeared prior to 1916 at West Lyell, 1914 at East Lyell, or 1908 at Goddard or Darwin (Fig. 4), suggesting that the glaciers began receding from their Matthes moraines about 1920. Glacier retreat was rapid through the 1930s and 1940s; the highest loss rate was at East Lyell Glacier between 1931 and 1944, −8360 m^2 yr^{-1}, totaling 0.11 km^2 (17% of the initial area). West Lyell and Conness glaciers lost about 25% of their area over the first half of the century and East Lyell and Darwin glaciers lost about 50%. Overall, glaciers continued to retreat between the 1950s and early 1970s, but at a slower rate. Three glaciers, East Lyell, Darwin, and Lilliput, enlarged (by 1%, 5%, and 5%, respectively) between the mid-1970s and 1990s, but the limited data preclude a more precise timing. The glaciers continued their retreat in the early 2000s. While the rate of change was variable between glaciers there is broad agreement in the temporal pattern of change among the glaciers.

To provide a generalized and more detailed time series of glacier change we assumed that the timing of advance and retreat of all seven glaciers was identical and we could interpolate a synthetic glacial time series. For example, if all glaciers had data for 1920 and 1940 showing retreat, and one glacier also had a datum for 1930, showing that the glacier area did not change between 1920 and 1930, we interpolated a constant glacier area until 1930. For dates where data were available for all or most of the glaciers, the fractional area change was simply an area-weighted average of all the values. We contend the result represents the general pattern of glacier change in the Sierra Nevada with four main phases (Fig. 5): (I) 1900–1920, the glaciers are at the Matthes moraines; (II) rapid retreat 1921–1960; (III)

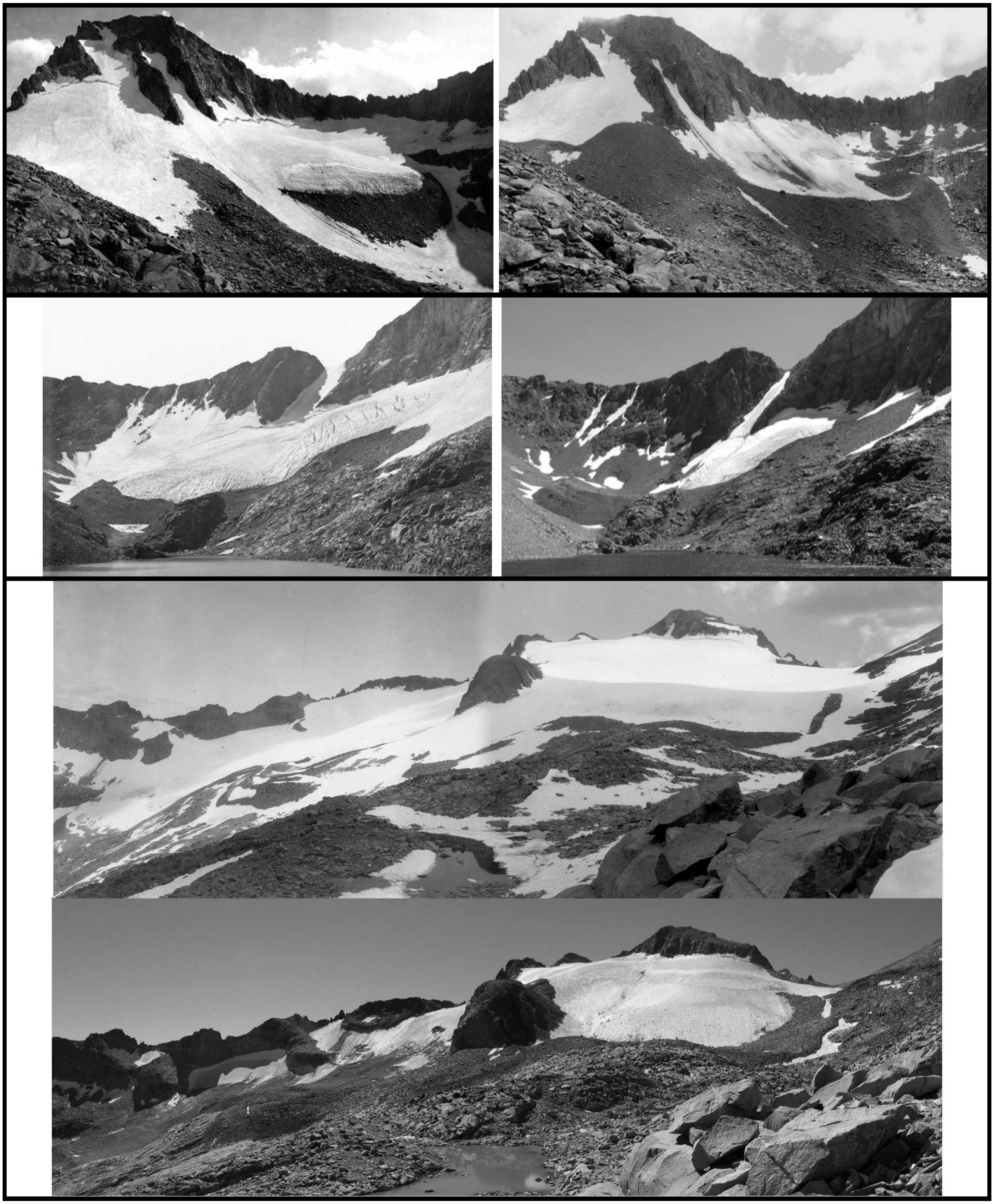

FIGURE 3. Repeat photographs of Darwin Glacier (top set); left photograph taken on 14 August 1908 by G. K. Gilbert (USGS), right photograph on 14 August 2004 by H. Basagic. Dana Glacier (middle set); left photo taken in 1883 by I. C. Russell, right photo taken on 9 September 2004 by H. Basagic. The East and West Lyell Glacier (bottom set); top photo taken on 7 August 1903 by G. K. Gilbert (USGS), bottom photo taken on 5 September 2004 by H. Basagic.

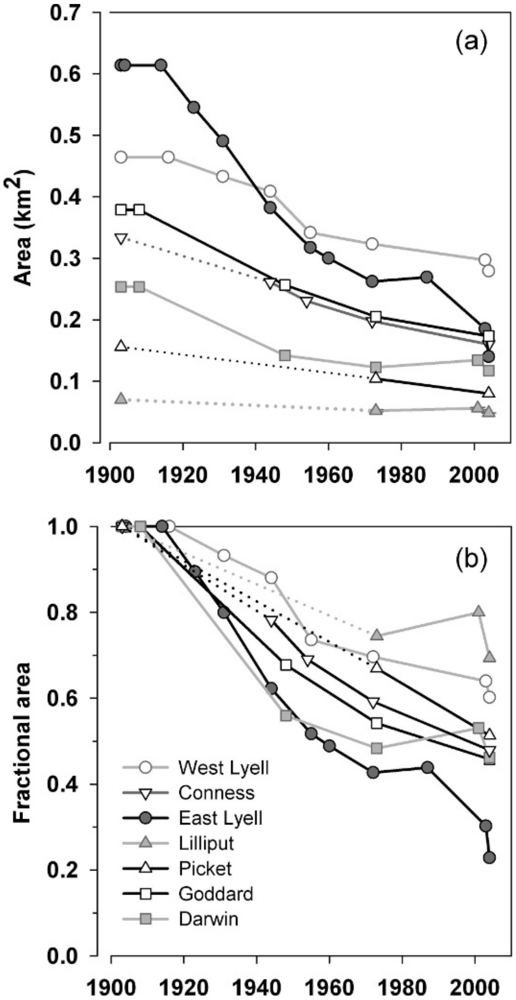

FIGURE 4. (a) Surface area and (b) fractional area changes at the seven glaciers in the Sierra Nevada over the past century (1903–2004). Dotted lines represent assumed change for glaciers lacking early 1900s photographic evidence. Error bars are omitted as error is smaller than symbol.

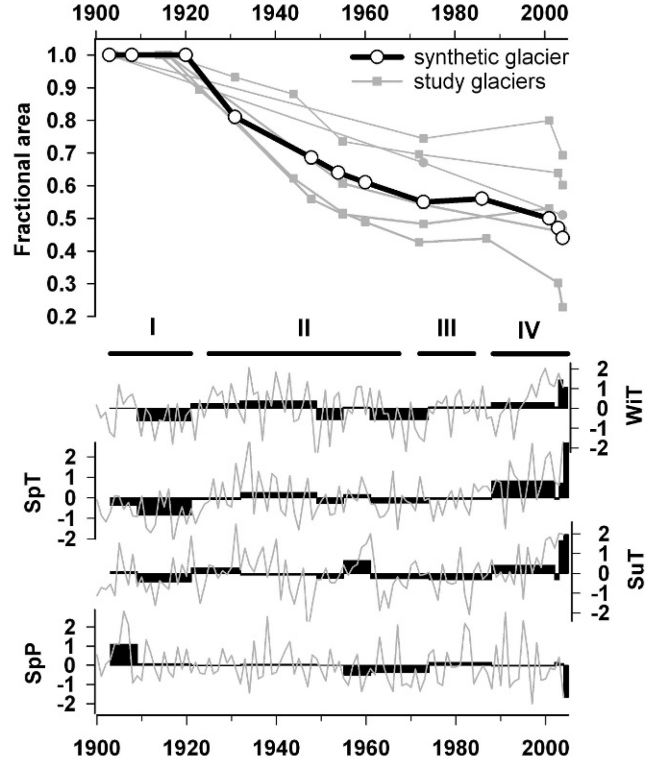

FIGURE 5. A comparison between (top) fractional area glacier change and (bottom) seasonal climate variables over the past century. Glaciers include the area weighted synthetic glacier constructed from the seven study glaciers. Horizontal lines (I–IV) define the four periods of glacier activity. Climate variables are normalized (light gray lines) and include winter (WiT), spring (SpT), and summer (SuT) temperatures, and spring precipitation (SpP). Black climate bars represent average conditions for time periods based on the synthetic glacier.

retreat slowed and slight advance in the late 1960s–late 1980s; (IV) late 1980s–present, retreat resumed. A possible fifth phase may have started since 2000 with accelerated retreat rate.

We extrapolated the changes of the synthesized glacier record to estimate the change for the entire glacier population in the Sierra Nevada from 1903 to 2004. Extrapolating the results incurs two assumptions. First, the synthesized glacier record is a reasonable representation of the behavior of the glacier population such that fractional area changes directly scale to the fractional area change to the entire population (Table 3). Second, the total ice-covered area from the inventory is equivalent to the total area in 1973, a data point in our synthetic glacier record. The inventory is based on aerial photographs acquired in 1975–1984, a period of little glacier change, and we assume that the inventory is representative of the area in 1973. The uncertainty for the population is known for 1973 from the GIS inventory (Table 1), and the uncertainty of the 1900 and 2004 population area is estimated from the standard deviation of the change between the seven glaciers at the intervening time periods (1900–1972, 1972–2004). The estimated area of the population for 1903, 71.18 ± 8.53 km^2, may be overestimated because smaller features in close proximity to each other in 1973 would overlap in 1903. However, this may be offset by the potential loss of smaller features from 1903 to 1973. The estimated area by 2004 is 31.32 ± 4.40 km^2, a loss of 39.86 ± 9.59 km^2 (56%) since 1903.

Volume of ice in 1903 and 2004 was estimated by simply changing the area of each glacier and perennial snowfield by the fractional area change from 1973 and recalculating volume using the "Cascades, small glaciers" parameters in Equation (2). Area changes tend to be larger for the smaller glaciers (Granshaw and Fountain, 2006; Paul et al., 2009). Using the change in 7 glaciers, which are larger than the population of glaciers in the Sierra, as an index of change for the population probably underestimates the volume shrinkage of the population. For 1903 the total ice volume is estimated to be 1.09 km^3 and by 2004 the estimated volume shrunk to about 0.43 km^3.

Analysis

GLACIER CLIMATE COMPARISON

To determine the climate variables controlling glacier change we examined seasonal air temperatures and precipitation because these are the dominant macroscale influences on glacier mass balance (Meier, 1965; Tangborn, 1980). We included temperatures for summer (June–August) for its importance to melt, and spring (March–May) for its importance to the phase of precipitation, snow or rain. Winter (December–February) air temperatures are also included as an index of snow temperature because if the winter snow becomes progressively warmer over the decades, less energy is required to heat the snow to melting in spring thus

TABLE 3

Estimated area and volume changes for the glacier population.

Year	Synthetic Fractional Area	Population area (km^2)	Area loss since 1903 (km^2)	Volume (km^3)
1903	1	71.18 ± 8.53	—	1.09
1972	0.55	39.15 ± 0.13	32.03 ± 8.53	0.55
2004	0.44	31.32 ± 4.40	39.86 ± 9.59	0.43

expanding the summer melt season. Winter precipitation is an index of snow accumulation, and spring precipitation is included because it could add mass to the glaciers and the fresh snow increases the albedo of the glacier surface reducing melt. We also included autumn (September–November) temperature and precipitation to complete the annual cycle of climate variables. Because we are comparing climate variations to geometric change, rather than mass change, the time-scale response (Johannesson et al., 1989) of these glaciers could be important. These glaciers are small and tend to be thin, averaging 22 m as estimated from the seven glaciers with good temporal data series. The annual rate of mass loss at the terminus is >2 m yr^{-1} (Tangborn et al., 1977). Correspondingly, the time-scale response of these glaciers is fast, <11 yrs, and roughly equivalent to the intervals in our glacier time series of roughly 10 yrs.

For climate data, we used monthly data from the Parameter-elevation Regressions on Independent Slopes Model (PRISM), a spatially modeled data set interpolated from surrounding lower elevation stations for the years 1900–2004 (Daly et al., 1997), because measured high elevation data are not available for the length of the glacier record. We used data from a single grid cell (4 km resolution) centered over Lyell Glacier. Although the synthetic glacier record is constructed from glaciers as much as 170 km away from Lyell, the high alpine region experiences relatively uniform temporal variations in monthly climate, though we acknowledge that deviations between the central and southern regions occur. PRISM cells in the southern Sierra contain artifacts due to the inclusion of new stations over time so we used only the cell over Lyell which appears not to include artifacts. We compared monthly mean temperature and precipitation PRISM data to the climate record at Grant Grove (2012 m elevation, 1949–2004), located 115 km to the south, with significant results for both temperature ($r^2 = 0.96$) and precipitation ($r^2 = 0.83$). Climate trends in PRISM data over the past century were statistically significant ($p < 0.05$) for winter and spring air temperatures (warming 0.07 °C and 0.19 °C $decade^{-1}$, respectively) and summer precipitation (increasing). Summer temperatures also warmed but the trend was not significant. There was no trend in winter precipitation.

TABLE 4

Pearson product-moment correlation results for the synthetic glacier area and seasonal climate variables, 1903–2004. Significant values are shown in bold ($p < 0.05$).

	All intervals ($n = 11$)	2003–2004 absent ($n = 10$)
Temperature		
Winter	**−0.66**	**−0.64**
Spring	**−0.81**	−0.61
Summer	**−0.80**	**−0.70**
Autumn	0.31	−0.16
Precipitation		
Winter	−0.35	0.25
Spring	**0.72**	0.48
Summer	−0.04	−0.11
Autumn	0.49	0.14

To compare glacier change with climate variations we used the synthetic glacier because it had a relatively detailed time series with 11 data points. To correlate the climate data which occurs at annual time intervals and the glacier data at irregular intervals, the climate data were averaged over the glacier data intervals (Fig. 5). Results show that area change correlated significantly with winter, spring, and summer air temperatures, and spring precipitation (Table 4). The correlation results were sensitive to the 2003–2004 period, a year with large area change and above-average spring temperature and below-average spring precipitation. Without the 2003 data, only winter and summer temperatures were significant. Neither autumn air temperature nor precipitation were correlated significantly with glacier area change.

Clearly warmer summer temperatures are important because they greatly influence summer melt (Meier, 1965; Cuffey and Paterson, 2010). Winter temperatures have little or no influence on ablation through melting because of the subfreezing temperatures common to the elevations glaciers are found at in the Sierra Nevada (Tangborn et al., 1977). However, a warmer snowpack at the start of summer requires less energy to warm the snowpack to melting than a cold snowpack. Although spring temperatures/precipitation were correlated significantly when the 2003–2004 glacier record was included, it may become important in the future with continued climate warming and deserves some mention. The importance of spring temperatures may be related to the length of the melt season, in addition to the phase of precipitation. The mean temperature for spring over the 20th century is –1.49 °C, (standard deviation = 0.95 °C) at Lyell Glacier, close to the melting point. The phase of spring precipitation is important as new snow, as opposed to rain, dramatically increases the albedo of the glacier (Shea et al., 2004), which reduces energy absorption and melt. The importance of spring snow has been illustrated elsewhere in the high alpine of the western states (Hoffman et al., 2007).

We qualitatively compared the four phases of our synthetic glacier record to three climate variables (Fig. 5). The climate variables are normalized differences calculated by taking the difference from the long-term average and dividing by the long-term standard deviation. In the early part of the century, period I, glacier area was constant at roughly the Matthes extent, winter/spring/summer temperatures were up to −1 °C below the long-term averages, −6.59, −1.47, and 9.46 °C, respectively, and spring precipitation was above average. Glacier recession, period II, started about 1920 as the temperatures warmed about 0.5 °C above average and spring precipitation decreased to the long-term average. Midway through period II, temperatures oscillated by 0.5 °C cooler/warmer than average and spring precipitation had decreased. In period III, a time of relative glacier stasis and slight advance, temperatures were cool and spring precipitation above normal. In the final period, IV, temperatures warmed by at least 0.5 °C, and by the late 1990s winter and spring temperatures were warmest on record. Overall, winter, spring, and summer air temperatures appear to be driving the glacier changes and spring precipitation (snowfall), while not exhibiting any long-term trend may temporarily reinforce or buffer the effects of temperature.

LOCAL TOPOGRAPHY

The overall uniform timing of glacier change across ~200 km suggests the glaciers are responding to the temporal variations in regional climate; however, the magnitude of glacier response differs between glaciers suggesting secondary factors which we infer to be a local influence and individual to each glacier. We examine the local topography as one possible local factor. Subglacial topography exerts a strong control on glacier thickness. All other factors being equal, a glacier on a steep slope will be thinner than a glacier filling a bowl-shaped depression. Glacier advance/retreat would be more responsive to climatic changes in the former condition than the latter (Johannesson et al., 1989). Subaerial topographic influences include changes in incident solar radiation caused by slope and aspect (Graf, 1976; Allen, 1998; Chueca and Julian, 2004), and enhancement or reduction of snow accumulation (Kuhn, 1995). That most glaciers are found on north- to northeast-facing slopes is a result of topographic-induced reduction of incident solar radiation (Fig. 2c, Table 2). Of course, topography can enhance or reduce snow accumulation through avalanching and wind transport (Kuhn, 1995; McClung and Schaerer, 2006). In the Sierra Nevada, mass balance measurements at Maclure Glacier in 1967 indicated that redeposition of winter snow from the surrounding area contributed 43% of total accumulation (Tangborn et al., 1977). The area-elevation distribution of the glacier is one topographic influence which can affect glacier area response to climate (Tangborn et al., 1990). The difference in area distribution is apparent at East and West Lyell glaciers; the former had a lower area-weighted mean elevation than the latter by −84 m in 1900 and −43 m in 2004. We assert that this difference explains the difference in area change between the glaciers—East Lyell has lost 78% of its original 1903 area; West Lyell has lost 40%.

Many small glaciers require local topographic conditions to enhance snow accumulation or to reduce ablation because the entire glacier exists below the theoretical snow line (Kuhn, 1995). The magnitude of the topographic influences can change through time as a glacier changes size (Graf, 1976; López-Moreno et al., 2006). A cirque glacier that retreats into its own basin is relatively more shaded from the sun and gains a greater fraction of its winter snow accumulation from avalanching compared to one that extends well outside the basin. Clark and Gillespie (1997) suggested that Sierra Nevada glaciers have become increasingly dominated by local topographic conditions as they have become smaller. To assess the degree of local topographic influence, we examined four topographic variables: elevation, slope, aspect, and headwall height. The mean elevation, slope, and aspect of the 14 glaciers (Table 2) were similar to the 'true' glaciers of the glacier population (Table 1). We derived the mean cirque headwall height above each glacier by taking the elevation difference of the peak or ridge above the glacier, determined from a 10 m DEM, and the glacier boundary from the USGS topographic maps. Headwall cliffs above the 14 glaciers examined ranged in height between 44 m at East Lyell Glacier and 305 m at Black Giant Glacier (Table 2). No significant correlations were found between the four topographic variables and glacier area change except with cirque headwall height ($r^2 = 0.32$; $p < 0.05$) (Fig. 6). The scatter in the figure may be due to differences in surrounding morphology of the

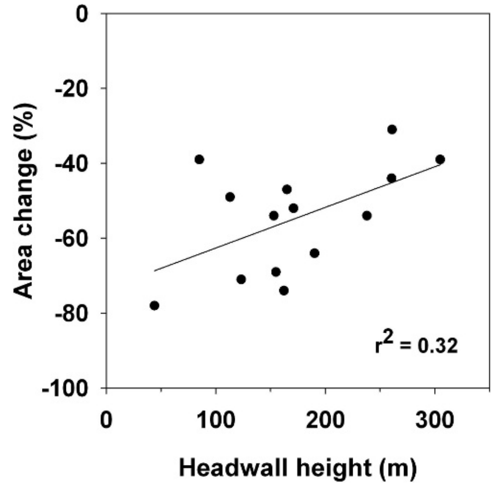

FIGURE 6. Glacier area change (1903–2004) as a function of mean headwall cliff height above glacier.

headwalls, which affects the quantity of snow available for redistribution to the glacier. A slight increase in area loss was observed with increased elevation ($r^2 = 0.03$) and increased slope ($r^2 = 0.10$). No trend was observed with aspect, likely because all 14 glaciers faced north/northeast. The local influence is likely from a combination of these variables, though a multiple linear regression did not improve the statistical significance.

We examined solar radiation to determine if glaciers have retreated into more favorable environmental conditions through time. We used the ArcView Solar Analyst extension to calculate the total annual magnitude of incoming solar radiation, or insolation (direct + diffuse radiation) for each of the 14 study glaciers. Solar Analyst calculates insolation based on an elevation model and accounts for changes in viewshed, surface orientation, elevation, and atmospheric conditions. We assumed clear skies to identify differences in potential annual insolation. Solar radiation was calculated for glacier configurations in 1903 and 2004 to determine change in annual insolation and summarized with an area weighted mean. The total mean annual insolation decreased for all 14 study glaciers over the past century by an average of 15 W m^{-2}. Dragtooth Glacier had the largest decrease in mean insolation, by 38 W m^{-2} (−28%) and West Lyell Glacier decreased the least, 5 W m^{-2} (−3%). The results show that as glaciers have retreated over the past century, they have retreated to locations that are more controlled by local topography. Greater local topographic control may lessen the glacier area response to regional climate. This has implications on the interpretation of future glacier area changes.

Discussion and Conclusions

Our digital inventory, based on the USGS 1:24,000 maps, identified 1719 glaciers and perennial snowfields with a mean area of 0.02 km^2 and total area of 39.15 ± 0.13 km^2. The inventory count and total area is likely a present-day maximum because the photography for the maps were acquired between 1975 and 1984 during a relatively cool and snowy period, and there has been significant glacial retreat since that time. However, the glacier count may be less affected than area by the splitting of glaciers as they recede, while the total area has likely decreased.

The glacier inventory results compare well with Raub et al. (2006)—who had directly examined aerial photography with glaciological expertise—once we filtered our inventory to match their size criteria for glaciers and ice patches, >0.005 km^2. Given that size threshold, our estimate of the number of features decreases to 1313, 28 (2%) more than Raub et al. (2006), and with a total area of 37.9 ± 0.1 km^2, 2.7 km^2 (8%) greater than Raub et al. (2006). We consider the correspondence between estimates to be quite good considering the differences in map scales, photography, dates, and approaches. That our inventory based on cartographic expertise compares well with an inventory based on glaciological expertise (Raub et al., 2006) underscores the accuracy of our methods and provides confidence in applying our methods to other glacier populated regions. Granshaw and Fountain (2006) found similar results between their digital inventory and a manual inventory by Post et al. (1971) for glaciers and perennial snowfields of the North Cascade Range in Washington.

A second comparison between inventories was made for 'true' glaciers. Our shear stress criteria identified 122 glaciers compared to Raub et al. 's (2006) result of 881 using an area threshold (more than twice our glacier-covered area). Matching Raub et al.'s glacier count required a reduced shear stress threshold of 0.4×10^5 Pa, but resulted in an increase in glacier-covered area of 4.5 km^2 over Raub et al.; whereas matching glacier area required a shear stress of 0.5×10^5 Pa, but increased the glacier count over Raub et al. by 187. Our shear stress criteria eliminated perennial snowfields on flat slopes and many small features ($\geq$0.05 km^2). A minimum area definition for glaciers can be useful for general inquires; however, such a definition is arbitrary and may ignore many small glaciers. We consider our estimates of glacier volumes and thickness to be rough estimates of significant differences between individual points and the regression as shown in Chen and Ohmura (1990). We attempted to account for a possible bias introduced by the data set, which is based on much larger glaciers, by using a regression that Chen and Ohmura (1990) developed for the smaller glaciers in the Cascades of the Pacific Northwest. Results showed a better correspondence with observations given the methods in our application. The differences in identifying 'true' glaciers highlight the coarse methods of approximation of glacier movement based on mapped data.

The magnitude of area change for 14 glaciers from 1903 to 2004 ranged from -31% to -78%, an average of -55%. The total ice-covered area lost over the past century for the entire Sierra Nevada is estimated at 39.86 ± 9.59 km^2. Based on these values, the ice volume loss from 1903 (1.09 km^3) to 2004 (0.43 km^3) is 0.66 km^3 (0.59 km^3 water). The fractional area loss is similar to other glacier-covered regions globally, such as -49% in the Southern Alps of New Zealand from 1850 to 2006, and -35% in the European Alps (Hoelzle et al., 2007). This change is also similar to that in the Western U.S.A. and Canada, including glaciers in the Lewis Range of Glacier National Park, Montana (-65%, 1850–1979) (Key et al., 1998; Hall and Fagre, 2003); in the Colorado Front Range (-40%, 1909–2004) (Hoffman et al., 2007); in the Northwest U.S.A. including Mount Hood (-34%, 1901–2004) (Jackson and Fountain, 2007), Mount Adams (-49%, 1904–2006) Sitt et al., 2010), and Mount Rainier (-19%, 1910–1994) (Nylen, 2004); and in Canada in the Columbia and Canadian Rocky Mountains (-5% and -15%, 1951/1952–2001, respectively) (Debeer and Sharp, 2007).

The temporal pattern of 20th century glacier retreat in the Sierra Nevada can be divided into four periods. From 1903 to about 1920 glacier area did not change and was at its fullest LIA extent. Following 1920 through the 1960s the glaciers retreated rapidly. Between the 1970s and early 1980s, the glaciers stopped retreating, and in some cases, started to advance. Finally, the glaciers began to retreat again in the late 1980s and increased their rate in the early 2000s. This pattern is similar to glacier-covered regions elsewhere in the Western United States. Most glaciers, including those in the Sierra Nevada, began retreating in the 1920s–1930s (Matthes, 1940). The retreat lasted until the late 1940s and early 1950s when many glaciers stabilized (Nylen, 2004; Hoffman et al., 2007; Jackson and Fountain, 2007). By the 1960s the glaciers in the Northwest United States began to advance (Hubley, 1956; LaChapelle, 1965). Glaciers in the European Alps showed a similar pattern with glaciers generally retreating from 1930 to 1960 (Dyurgerov and Meier, 2000; Haeberli et al., 2007). Glaciers stagnated or advanced through the mid-1970s until a return to general retreat in the mid-1980s. Most glaciers in the Northwest United States returned to retreat in the 1980s and 1990s. Overall, the glaciers in the Sierra Nevada appear consistent with global glacier change.

Loss of glacier area in the Sierra Nevada is significantly correlated with summer, and winter air temperatures. Winter temperatures do not directly influence ablation but may warm the snowpack such that less energy is required in the spring to warm the snowpack to melting temperatures. Spring temperature and precipitation may to be important to glacier ablation due to increased warming for melt and late spring snowfall which increases albedo and decreases melt. Hoffman et al. (2007) reached similar conclusions regarding glacial response in the Front Range, Colorado, to spring and summer temperatures.

Variations in the magnitude of individual glacier response are probably caused by differences in local topography that result in differences in mass accumulation and melting. Of the broad-scale topographic features examined we found that higher headwall cliffs above the glaciers were significantly correlated with smaller losses of glacier area. Glaciers with tall headwall cliffs presumably avalanche more winter snowfall onto the glacier surface contributing to the mass gain and shade the surface in summer, which reduces melt. Glaciers have retreated into areas with less annual insolation over the past century, which may make them less responsive in the future to climate changes.

The recent (1972–2004) mean retreat rate of the seven study glaciers is 0.0012 km^2 yr^{-1}. If the glaciers continue to shrink at this rate, most will disappear in 50 to 250 years. The loss of these frozen reservoirs means they will no longer buffer streamflow against summer droughts (Fountain and Tangborn, 1985; Moore et al., 2009) and there will be an increase in stream water temperatures. This would reduce kryal-adapted instream flora and fauna (Brittain and Milner, 2001; Dougall, 2007), but possibly increase the range of warm temperature–adapted biota.

Acknowledgments

This research was supported by funding from the U.S. National Park Service, including Sequoia and Kings Canyon National Parks, the Geologic Resource Division of the National Science Foundation (BCS-0351004), NASA (NNGO4GJ41G), and the Western Mountain Initiative, funded by the USGS. We appreciate comments received from Greg Stock and one anonymous reviewer, which contributed to improving this paper. Special thanks to Annie Esperanza, Danny Boiano, Linda Mutch, Hal Pranger, Matt Hoffman, and Rhonda Robb for their support.

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MS accepted January 2011

Appendix

Photograph sources:

Sequoia and Kings Canyon National Park
47050 Generals Highway
Three Rivers, California 93271-9700, U.S.A.

Yosemite National Park Research Archive
P.O. Box 577
Yosemite National Park, California 95389, U.S.A.

U.S. Forest Service, Bishop Field Office
351 Pacu Lane
Bishop, California 93514, U.S.A.

U.S. Geological Survey Photographic Library
URL: http://libraryphoto.cr.usgs.gov/

National Agriculture Imagery Program (NAIP), U.S. Department of Agriculture
URL: http://www.fsa.usda.gov/FSA/apfoapp?area=home&subject=prog&topic=nai

U.S. Geological Survey Earth Resources Observation Systems
URL: http://eros.usgs.gov/

TABLE A1

Compilation data of each year depicted for 14 glaciers and root mean squared errors (RMSE) for georeferenced aerial photographs (AP), terrestrial ground-based photographs (TGP), digital orthophotograph quadrangles (DOQ), and global positioning system (GPS) field mapping. Data annotated with [m] indicate a moraine was used to derive the glacier extent. For Sources, YOSE—Yosemite National Park Archive; USFS—U.S. Forest Service Bishop Office Archive; SEKI—Sequoia and Kings Canyon National Parks Archive; NAIP—National Agriculture Imagery Program; EROS—U.S. Geological Survey Earth Resources Observation Systems.

Glacier	Year	Data	Date	Source	RMSE (m)	Area (km^2)
Dragtooth	1903	AP [m]	N/A	NAIP	7	0.2094 ± 0.0042
Dragtooth	2005	DOQ	8/25/2005	NAIP	7	0.0548 ± 0.0021
Conness	1903	AP [m]	9/27/1944	YOSE	4.9	0.3339 ± 0.0046
Conness	1944	AP	9/27/1944	YOSE	4.9	0.2611 ± 0.0041
Conness	1954	AP	8/19/1954	USFS	5.9	0.2304 ± 0.0041
Conness	1972	AP	8/14/1972	USFS	9.9	0.1977 ± 0.0049
Conness	1993	DOQ	9/23/1993	EROS	—	
Conness	2004	GPS	9/23/2004	Basagic	2.5	0.1600 ± 0.0002
Dana	1903	GPS [m]	8/18/2003	Basagic	5	0.3193 ± 0.0045
Dana	2003	GPS	8/18/2003	Basagic	5	0.1153 ± 0.0004
Maclure	1903	AP [m]	N/A	geologic	10	0.2930 ± 0.0060
Maclure	2003	GPS	8/7/2003	Basagic	5	0.1553 ± 0.0004
West Lyell	1903	AP [m]	9/28/1944	YOSE	11.2	0.4643 ± 0.0082
West Lyell	1916	TGP	8/14/1916	unknown	10	0.4643 ± 0.0076
West Lyell	1931	TGP	10/1/1931	NPS, 1931	11.2	0.4329 ± 0.0079
West Lyell	1944	AP	9/28/1944	YOSE	11.2	0.4088 ± 0.0077
West Lyell	1955	AP	8/8/1955	YOSE	6.9	0.3417 ± 0.0053
West Lyell	1972	AP	8/14/1972	USFS	8	0.3233 ± 0.0056
West Lyell	1993	DOQ	9/25/1993	EROS	—	
West Lyell	2003	GPS	8/13/2003	Basagic	5	0.2970 ± 0.0006
West Lyell	2004	GPS	9/3/2004	Basagic	2.5	0.2795 ± 0.0003
East Lyell	1903	AP [m]	9/28/1944	YOSE	11.2	0.6334 ± 0.0096
East Lyell	1914	TGP	1914	Matthes	10	0.6334 ± 0.0089
East Lyell	1923	TGP	9/13/1923	YOSE	10	0.5451 ± 0.0082
East Lyell	1931	TGP	10/1/1931	NPS, 1931	11.2	0.4908 ± 0.0084
East Lyell	1944	AP	9/28/1944	YOSE	11.2	0.3821 ± 0.0074
East Lyell	1955	AP	8/8/1955	YOSE	12.1	0.3174 ± 0.0072
East Lyell	1960	TGP	9/9/1960	NPS, 1960	10	0.3001 ± 0.0061
East Lyell	1972	AP	8/14/1972	USFS	9	0.2621 ± 0.0053
East Lyell	1987	TGP	9/14/1987	D. Hardy	20	0.2691 ± 0.0099
East Lyell	1993	DOQ	9/25/1993	EROS	—	
East Lyell	2003	GPS	8/13/2003	Basagic	5	0.1855 ± 0.0005
East Lyell	2004	GPS	9/2/2004	Basagic	2.5	0.1402 ± 0.0002
Darwin	1903	AP [m]	10/7/1948	USFS	5.5	0.2536 ± 0.0042
Darwin	1908	TGP	8/14/1908	Gilbert	10	0.2536 ± 0.0056
Darwin	1948	AP	10/7/1948	USFS	5.5	0.1419 ± 0.0031
Darwin	1973	AP	9/9/1973	SEKI	5.2	0.1226 ± 0.0028
Darwin	1999	DOQ	7/31/1999	EROS	—	
Darwin	2001	AP	7/29/2001	SEKI	6	0.1345 ± 0.0031
Darwin	2004	GPS	8/12/2004	Basagic	2.5	0.1172 ± 0.0002
Goddard	1903	AP [m]	10/7/1948	USFS	7.2	0.3787 ± 0.0057
Goddard	1908	TGP	8/15/1908	Gilbert	10	0.3787 ± 0.0069
Goddard	1948	AP	10/7/1948	USFS	7.2	0.2566 ± 0.0047
Goddard	1973	AP	9/8/1973	SEKI	3.5	0.2051 ± 0.0033
Goddard	1999	DOQ	7/31/1999	EROS	—	
Goddard	2004	GPS	8/15/2004	Basagic	2.5	0.1734 ± 0.0002
M. Goddard	1903	GPS [m]	8/15/2004	Basagic	2.5	0.1106 ± 0.0023
M. Goddard	2004	GPS	8/15/2004	Basagic	2.5	0.0347 ± 0.0001
Black Giant	1903	DOQ [m]	8/27/2005	NAIP	7	0.1120 ± 0.0031
Black Giant	2004	GPS	8/17/2004	Basagic	5	0.0688 ± 0.0003
M. Palisade	1903	DOQ [m]	8/27/2005	NAIP	7	0.3071 ± 0.0051
M. Palisade	2005	DOQ	8/27/2005	NAIP	7	0.1735 ± 0.0038
Brewer	1903	DOQ [m]	8/26/2005	NAIP	7	0.1456 ± 0.0035
Brewer	2005	DOQ	8/26/2005	NAIP	7	0.0418 ± 0.0019
Lilliput	1903	AP [m]	9/9/1973	SEKI	5.2	0.0699 ± 0.0021
Lilliput	1973	AP	9/9/1973	SEKI	5.2	0.0521 ± 0.0019
Lilliput	1988	DOQ	8/7/1988	EROS	—	
Lilliput	2001	AP	7/29/2001	SEKI	3	0.0559 ± 0.0017
Lilliput	2004	GPS	7/30/2004	Basagic	2.5	0.0485 ± 0.0001
Picket	1903	AP [m]	8/29/1973	SEKI	4.1	0.1555 ± 0.0030
Picket	1973	AP	8/29/1973	SEKI	4.1	0.1041 ± 0.0025
Picket	1988	DOQ	8/8/1988	EROS	—	
Picket	2004	GPS	8/27/2004	Basagic	2.5	0.0798 ± 0.0002