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Thermal and hydrologic attributes of rock glaciers and periglacial talus landforms: Sierra Nevada, California, USA

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ABSTRACT

To explore thermal regimes and hydrologic capacity of rock glaciers and related periglacial talus landforms, mini-thermochrons were deployed in and around potentially ice-embedded features of the Sierra Nevada. Results from pilot studies at 13 rock glaciers and 7 taluses indicate that outlet springs from these landforms generally do not desiccate but persist year-round as ice (frozen) in winter and flowing water in the warm season. Temperatures of water (liquid and ice) in rock-glacier outlet springs had an annual mean of −0.2 °C and mean of 0.6 °C during the warm season with very low diurnal fluctuation. These and other attributes suggest the existence of internal ice and/or permafrost supplying the springs. Air temperatures of rock-glacier matrices (1 m below the surface) versus surface air corroborate the periglacial nature of internal environments: annual air temperatures of matrices were below freezing (mean, −0.8 °C). Compared to surface air, especially during the warm season, matrix air temperatures were significantly colder and fluctuated less. Talus landforms followed a similar pattern, although water- and matrix air temperatures were warmer, and contrasts with surface air were not as strong as for rock glaciers. For rock glaciers and talus slopes, matrix air temperatures showed resistance (buffering) to changes in external air temperatures. Unique geomorphic conditions of rock glaciers and periglacial taluses in the Sierra Nevada appear to maintain cool-buffered thermal regimes at least partly decoupled from external air. Springs support persistent wetlands and lakes at their snouts, retaining water in otherwise semi-arid high cirques, and contribute as hydrologic reserves and critical habitat for alpine biota. Daily and seasonal lags and buffering effects suggest that ice within these landforms might resist surface warming on the longer term, which could make these landforms increasingly important as regional climates change.

1. Introduction

Thermal regimes within and near the surface of the alpine regolith influence important mountain processes, including permafrost and glacial development, snowpack retention, groundwater accumulation and discharge, wetland habitat functions, and alpine biotic interactions. Increasingly, climates at local scales in mountain environments are recognized as being at least partly decoupled from free-atmosphere meteorology (Clements et al., 2003; Pepin et al., 2011), making extrapolation from distant weather stations or synoptic patterns less reliable (Scherrer and Koerner, 2009).

Lesser studied and often overlooked relative to typical ice glaciers, rock glaciers and related periglacial talus landforms are coming into attention of climatic research. A continuum of alpine features, ranging from cirque-origin, debris-covered rock glaciers to hillslope block-fields and talus slopes, is common in mountain regions. This group of coarse rocky landforms is distinguished from rockslides and eroding alpine slopes by the role of embedded ice (persistent, transient, or relict), by distinguishing geomorphology that results from freeze-thaw and other cryogenic processes (e.g., clast sorting, over-steepened fronts, discrete sectors and boundaries; Whalley and Martin, 1992), and by distinctive thermal processes that ensue within the features (Harris and Pedersen, 1998; Juliussen and Humlum, 2008). Ice contained in these features can range from massive sedimentary blocks (glacial) to ice lenses and/or interstitial/congelation-ice in a matrix of rock and soil (permafrost or secondary refreezing; Clark et al., 1996; Haeberli, 2005; Williams et al., 2006; Leopold et al., 2011). The thick debris mantle characterizing these landforms creates challenges to evaluating the origin, presence, and extent of ice within these features; new geoelectrical methods are beginning to distinguish internal structure and dynamics (Hauck and Kneisel, 2008).

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These studies point to the contributions of rock glaciers and periglacial taluses to mountain hydrology from stored ice and permafrost. Unlike typical ice glaciers and annual snowpacks, ice and groundwater in rock glaciers and taluses are protected from direct effects of solar radiation by debris blankets, which insulate underlying ice and retard ablation relative to ambient air temperatures (Clark et al., 1994). Amplifying the insulation effect of rock mantling are internal thermal regimes created by air circulation within the rock-matrix (Juliussen and Humlum, 2008; Leopold et al., 2011), which can create conditions cold and persistent enough to depress local permafrost elevation at temperate latitudes by 1000 m (Delaloye and Lambiel, 2005). As a result, thaw of ice in rock glaciers lags behind typical ice glaciers, and these landforms appear to be in disequilibrium with climate, especially when climates are changing rapidly, such as at present (Brenning, 2005). For this reason, rock glaciers are likely to become increasingly critical alpine water reservoirs under global-warming conditions, especially as other sources of water decline, and drought severity increases (Schrott, 1996).

In the Sierra Nevada of California, hydrologic output during the critical dry season (late spring through autumn) derives primarily from annual snowpack, although the scattered and small ice glaciers also contribute to discharge, and are important for stabilizing late season flow (Raub et al., 2006). These primary sources of water storage and delivery are in severe decline in many parts of the range, and projections suggest declines to continue through the century (Maurer, 2007; Basagic, 2008). Rock glaciers and related landforms are even more abundant throughout cirques and valleys of the central and southern parts of the range (Millar and Westfall, 2008, 2010), but have been overlooked in analyses of hydrology for the region. As yet, no studies have investigated the nature or extent of ice within these features in the Sierra Nevada, the contribution of ice or groundwater to local or regional hydrology, or thermal processes related to these features. This paper summarizes results from pilot studies toward those goals. Mini-thermochrons were used to monitor air and water temperature from rock glaciers and periglacial talus fields in the central Sierra Nevada to infer the presence of ice or groundwater, assess seasonality of thermal regimes, investigate the relationship of microclimates with surface and regional climates, and explore relationships to wetland habitat.

2. Study area and landforms

The study concentrated on canyons of the alpine and upper subalpine zones of the central Sierra Nevada, including sites on the east and west sides of the hydrologic crest in a region extending from Sawtooth Ridge near Bridgeport to Mammoth Crest, south of Mammoth Lks, California (Fig. 1). The landforms occur most often on slopes and cirques that were glaciated during the Pleistocene, and thus typically are inset close to the range crest or under high peaks on escarpment ridges. The focus was placed on features that appear to be modern (*sensu* Millar and Westfall, 2008), that is, those where persistent or transient permafrost or embedded ice is likely to be present. Modern features were inferred from indirect evidence. For rock glaciers, this included presence of oversteepened fronts with reverse sorting; minimal lichen and plant growth on rock-glacier surfaces; U-shaped furrows (indicating movement) on the upper surface; running water heard below the surface; outlet streams, springs or terminal lake (if the latter, presence of glacial flour), and/or forefield wetland or terminal lake present (Fig. 2A). Periglacial taluses were differentiated from rockfall or rockslides by their lack (or minor amount) of fine sediments mixed with the blocky debris; regular clast sizes; rounded edges of clasts rather than fractured; discrete boundaries of landform with deep organic soil at the snout; flat or oversteepened front; running water heard below the rocky surface; and presence of springs and/or wetlands at the snout (Fig. 2B). Unlike rock glaciers, many talus fields in the Sierra Nevada seem unlikely to support massive, persistent ice, although seasonal/discontinuous ice and/or permafrost may play an important role (Millar and Westfall, 2008). As elsewhere, however, groundwater discharge from these features can be significant, implying important hydrologic reservoirs (Clow et al., 2003; Roy and Hayashi, 2009).

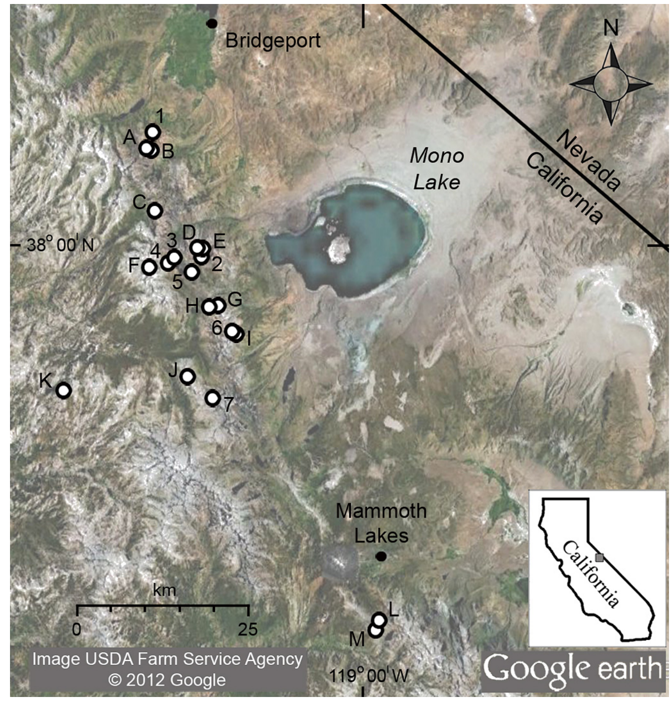

Fig. 1. Map of the study area, Sierra Nevada, eastern California, USA, showing location of rock glaciers and boulder-stream taluses where temperature measurements were taken. Location codes as in Table 1. Inset shows regional location.

Thirteen rock glaciers and seven boulder-stream talus fields were selected for study (Table 1). These were relatively small features, with the longest extending 1223 m from headwall to snout and the largest covering 40.3 ha. Aspects were northerly for the rock glaciers, and mixed north- and southerly for the taluses; substrates were diverse granitics and metamorphics.

3. Methods

Water and air temperatures were measured with Maxim iButton mini-thermochrons (www.maxim-ic.com/products/ibutton/), using DS1921 and 1922 instruments. Thermochrons were programmed to record temperature at 4-hr intervals, starting at noon local time on the day they were deployed; most ran 11 months and were switched with new instruments as near as possible to terminal dates. Temperature accuracy was set to 0.5 °C; this was checked for a period of 2 weeks by comparing iButton records to a high-resolution thermometer and placing the instrument in diverse thermal air and water conditions. From experience and that of other users, when iButtons fail in wildland settings, they stop recording rather than record errant temperatures.

For air-temperature measurements, the instrument was wrapped in wire mesh, inserted in white PVC T-fittings, which had three ~2 cm diameter openings. The PVC fittings served as sun shelters, and allowed airflow to the instrument. The PVC-instrument packages were tethered with wire to rocks and sheltered from direct sunlight in all but one situation, as described below. For both rock

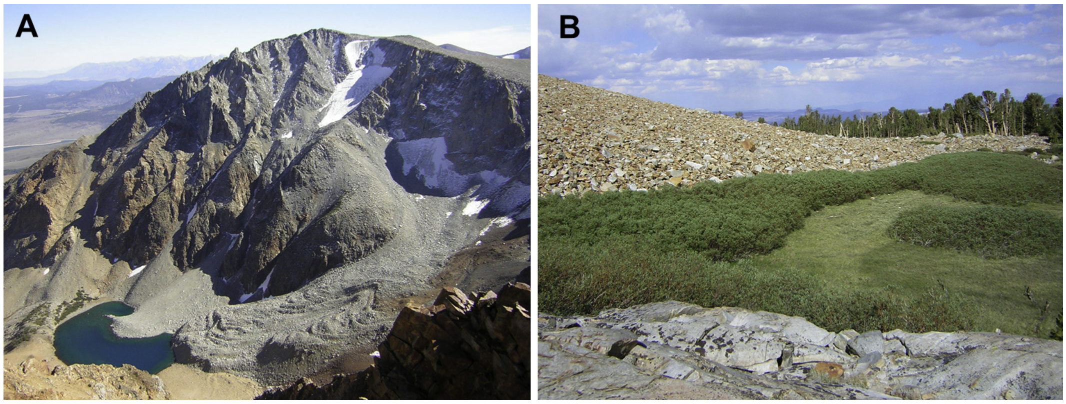

Fig. 2. Representative rock glaciers and boulder-stream taluses. A. Mt Gibbs rock glacier showing oversteepened fronts, furrowed surface, minimal plant growth, icefield on cirque wall above the landform, and persistent lake at the snout. B. Gibbs Canyon boulder-stream talus, showing forefield wetland.

glaciers and taluses, instruments measuring air temperature were placed in rock matrices ~1 m below the surface. In later years, to compare matrix temperatures to adjacent surface air temperatures, instruments were added at these positions at some locations. In those cases, the thermochrons were located near the rock surface (sheltered from direct sunlight) above the matrix locations. At one location (Barney Lk rock glacier) air temperature was also measured at 2 m above the landform surface, with the iButton-package tied to the tops of PVC poles, which served as masts to measure temperature above snow. In this case, no further protection from sunlight was available beyond the PVC fitting.

iButtons were also deployed in rock glacier and talus outlet springs. The goals in this case were to measure water temperature directly, and to infer from the temperatures and the daily fluctuations whether the spring was flowing (liquid), frozen (ice), or desiccated. For rock glaciers and two talus features, pairs of iButtons were deployed at the outlet springs, so that water temperature profiles could be compared with adjacent air temperature. These air iButtons were deployed within crowns of riparian vegetation 1 m above the ground and directly adjacent to the springs (≤1.5 m from the spring iButton); in two cases there was no riparian context, and the air instruments were positioned among rocks directly adjacent to the springs. For water iButtons, instead of wire mesh, the instruments were wrapped in impervious plastic, the package was inserted into PVC fittings and weighted with rocks in the deepest pool at the outlet spring. During the first year of study (winter 2005–2006), levelloggers (Solinst, www.solinst.com/) were used to record water temperature, deployed directly into spring waters. Not all features were measured every year and dates varied by year and location. Gaps in data also resulted from instrument failures and/or inability to access locations due to lingering snowpack. Different times of deployment and retrieval for the various locations mean that not all data can be reported in parallel format.

Temperature data were downloaded from iButtons using 1-Wire Viewer (version 3.15.50, Maxim). Descriptive statistics and significant differences in means among years, all features in total, and among landform types were computed by analysis of variance (ANOVA) and by comparison of confidence intervals. To test consistent differences between surface and matrix and between air and water temperatures, a paired test was used. However, because the data were temporally autocorrelated, the correlation was

Table 1
Environmental context of landforms studied, Sierra Nevada, CA. A. Rock glaciers and B. Boulder-stream taluses.

Type	Name	Map code	River basin	Aspect	Elevation (m)[a]	Substrate	Size (ha)[b]
Rock glaciers	Tamarack Cyn East	A	East Walker	N	3157	Metamorphic	2.8
	Tamarack Cyn West	B	East Walker	NE	3125	Metamorphic	21.4
	Mt Excelsior	C	Walker	N	3410	Metamorphic	8.7
	Oneida Lk	D	Mono	NW	3018	Metamorphic	5.5
	Crystal Lk	E	Mono	NW	3200	Metamorphic	3.7
	Mt Conness	F	Mono	NE	3378	Granitic	13.0
	Karolyn Cyn	G	Mono	N	3115	Granitic	36.4
	Ellery Lk	H	Mono	N	3198	Granitic	8.7
	Mt Gibbs	I	Mono	NE	3129	Metamorphic	1.7
	Kuna Pk	J	Tuolomne	NE	3410	Granitic	4.1
	Mt Tenaya	K	Merced	NW	2973	Granitic	6.1
	Barney Lk	L	Owens	NE	3328	Metamorphic	14.6
	Deer Lk	M	San Joaquin	N	3413	Granitic	3.2
Means					*3220*		*10.0*
Boulder-stream taluses	Tamarack Lk	1	East Walker	NW	3015	Metamorphic	0.8
	Warren Fork East	2	Mono	SW	3335	Granitic	4.8
	Saddlebag Lk	3	Mono	S	3184	Granitic	17.2
	Greenstone Lk	4	Mono	N	3165	Granitic	2.1
	Warren Fork West	5	Mono	NE	3084	Metamorphic	16.2
	Gibbs Cyn	6	Mono	SE	3173	Metamorphic	2.2
	Helen Lk	7	Tuolumne	N	3544	Metamorphic	40.3
Means					*3214*		*11.9*

[a] At mid-point of landform.
[b] Includes only the rock-mantle portion of the feature, excluding ice headwall, if present.

removed prior to the test. Paired analyses, including the adjustment for autocorrelation, were done in SAS PROC MIXED (SAS, 2011). For air–water paired comparisons among sites, analyses were done on time period means by ANOVA.

In that little is known mechanistically about heat transfer in rocky blockfields and that the processes appear to differ from those in alpine soils, the strategic approach recommended by Box et al. (1978) was followed. First, the time–series relationships of changes in air temperatures within rock matrices relative to those at the surfaces were explored by using several regression models of surface temperature against matrix to find best statistical fits. In most situations, third-order polynomials fit the time-series (seasonal) data best. When third-order models were significant, the data were fitted by non-linear modeling with a four-parameter sigmoid function to better describe the heat-transfer processes. Linear and non-linear modeling was done in JMP (SAS, 2011). In that the temperature time series are autocorrelated, statistical tests on models are not valid and such models are over-fitted. Using SAS PROC ARIMA, we removed the autocorrelation from the time-series data using autoregressive integrated moving average (ARIMA) models, and then regressed a third-order model of daily mean surface air temperatures that also included 1-2 day lags to address heat-transfer from the surface to the matrix.

4. Results

4.1. Outlet springs

4.1.1. Rock glaciers

Water in rock-glacier outlet springs was on average colder than adjacent air throughout the year and especially in the warm season (Figs. 3A and 4A, Table 2A). All but one comparison of differences between means within and across years and locations was significant. Mean annual temperature of water (liquid and ice) was

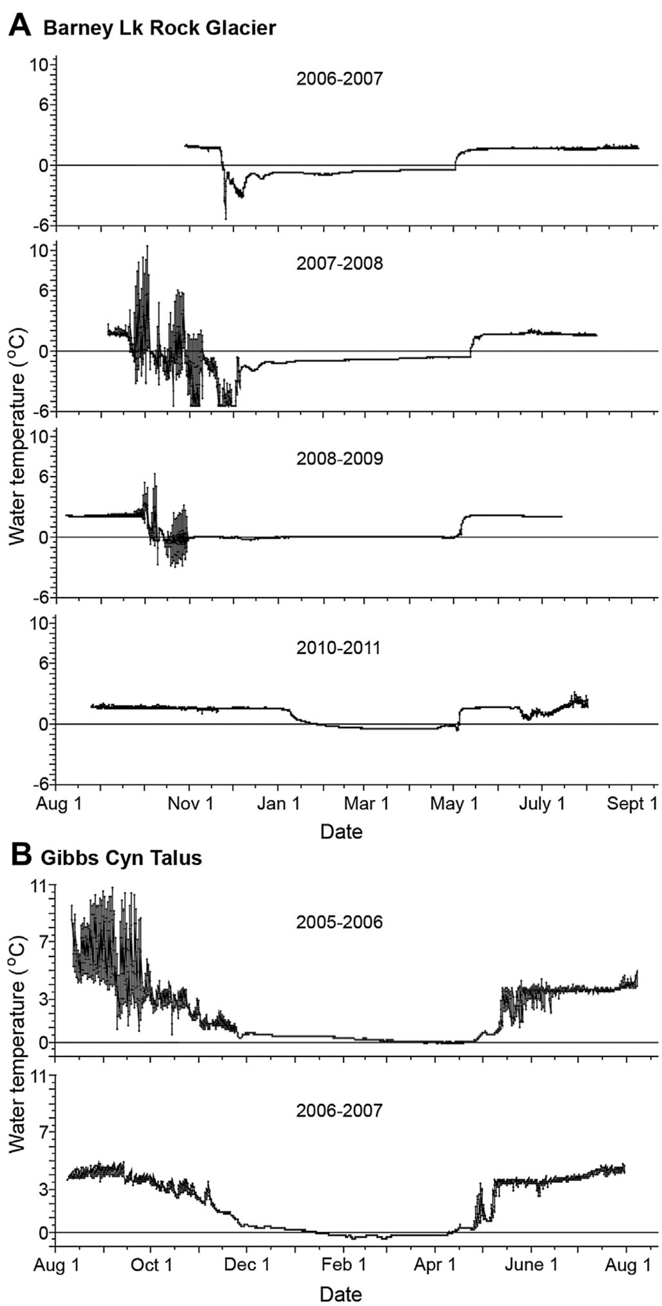

Fig. 3. Temperature recorded by thermochrons placed in representative outlet springs below snouts ("water temperature" can imply liquid water or ice). A. Barney Lk rock glacier, 2006–2011 (excluding 2009–2010). B. Gibbs Canyon boulder-stream talus, 2005–2007.

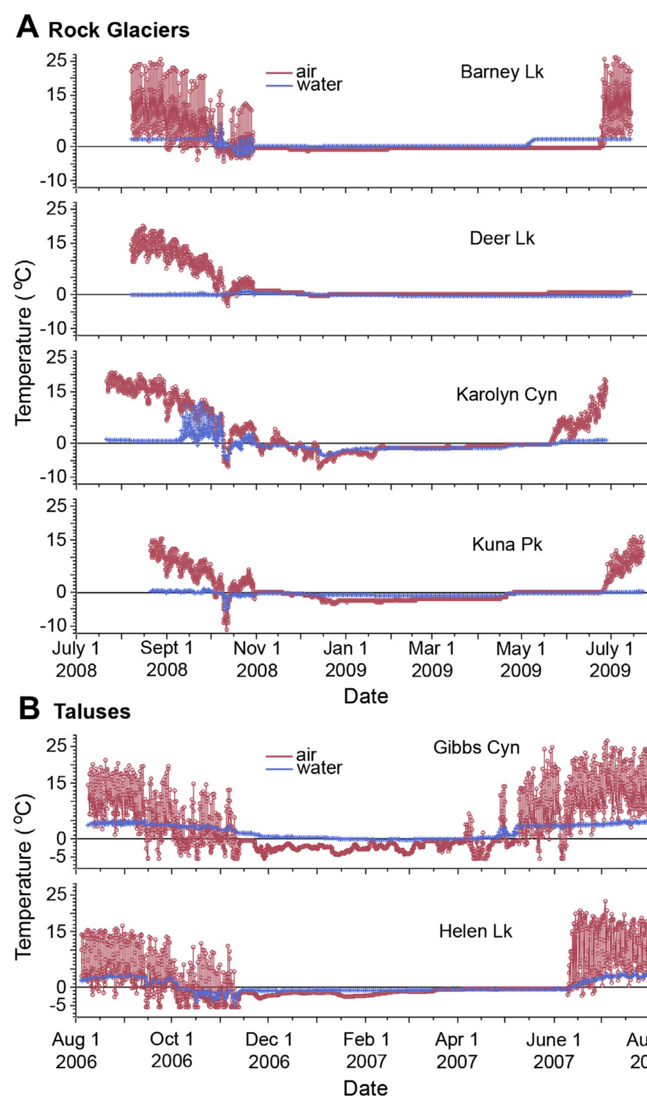

Fig. 4. Outlet springs paired water- and adjacent air temperatures for representative rock glaciers and taluses ("water temperature" as in Fig. 3). A. Four rock glaciers, 2008–2009. B. Two boulder-stream taluses, 2006–2007.

Table 2
Water temperatures of outlet streams, measured at springheads, compared to adjacent air temperature (°C). A. Rock glaciers. B. Boulder-stream taluses. Only features with complete records for the respective time periods are included. All differences between mean water and air temperatures within year were significant ($p < 0.05$) except one comparison, noted as *ns*.

Name	Measurement years	Annual temp, mean & sd, water	Annual temp, mean and sd, air	Warm-season temp, mean & sd, water	Warm-season temp, mean & sd, air	Cold-season temp, mean & sd, water	Cold-season temp, mean & sd, air
A. Rock glaciers							
Barney Lk Upper	2006–2007	−0.74 (0.89)		−0.28 (0.09)		−0.99 (1.03)	
	2007–2008	−0.08 (1.06)		0.49 (2.13)		−0.24 (1.13)	
Barney Lk Lower	2006–2007	1.16 (3.74)	3.58 (7.16)	1.62 (0.07)	9.07 (7.59)	0.90 (4.61)	0.67 (4.84) *ns*
	2007–2008	−0.27 (1.88)	1.49 (5.58)	1.58 (0.88)	6.05 (7.35)	−1.17 (1.55)	−0.71 (2.29)
	2008–2009	0.74 (1.11)	1.70 (5.40)	2.08 (0.13)	5.83 (7.38)	0.07 (0.69)	−0.41 (1.73)
	2010–2011	0.90 (0.90)	1.58 (4.27)	1.52 (0.38)	3.93 (6.29)	0.58 (0.92)	0.30 (1.39)
Mt Conness	2005–2006	0.49 (1.27)		0.88 (1.32)		0.26 (1.18)	
Deer Lk	2006–2007	−0.76 (0.42)	1.11 (6.64)	−0.32 (0.19)	6.96 (7.38)	−1.02 (0.27)	−2.29 (1.17)
	2007–2008	−0.61 (0.60)	−0.85 (3.34)	−0.31 (0.32)	2.39 (3.83)	−0.74 (0.65)	−2.42 (1.37)
	2008–2009	−0.33 (0.30)	2.28 (4.60)	−0.40 (0.14)	5.93 (6.30)	−0.30 (0.35)	0.42 (1.23)
Karolyn Cyn	2005–2006	−0.16 (1.15)		0.90 (0.33)		−0.70 (1.04)	
	2006–2007		4.35 (8.40)		13.30 (6.47)		−1.22 (2.73)
	2008–2009	−0.55 (1.86)	2.98 (6.77)	0.96 (1.79)	10.76 (5.66)	−1.31 (1.36)	−0.99 (2.52)
	2009–2010	−1.18 (2.13)	1.68 (5.98)	0.78 (1.63)	7.93 (6.29)	−2.18 (1.60)	−1.51 (1.93)
	2010–2011	0.72 (1.49)	4.37 (6.11)	1.39 (1.47)	9.26 (5.89)	0.16 (1.26)	0.34 (1.90)
Kuna Pk	2005–2006	0.24 (0.57)		0.80 (0.27)		−0.10 (0.40	
	2006–2007		2.21 (7.84)		9.47 (6.36)		−2.75 (3.90)
	2007–2008	−1.19 (1.19)		−0.03 (0.12)		−1.58 (1.13)	
	2008–2009	−0.58 (0.54)	0.84 (4.34)	−0.11 (0.23)	5.13 (4.82)	−0.81 (0.50)	−1.25 (1.83)
	2009–2010	−1.11 (0.59)	−0.50 (3.94)	−0.48 (0.26)	3.32 (4.54)	−1.43 (0.43)	−2.44 (1.32)
	2010–2011	−0.41 (0.19)	0.10 (3.38)	−0.23 (0.20)	2.72 (4.59)	−0.49 (0.11)	−1.23 (1.12)
Means		*−0.20 (1.15)*	*1.79 (5.58)*	*0.57 (0.63)*	*6.68 (6.05)*	*−0.58 (1.10)*	*−1.03 (2.08)*
B. Boulder-stream taluses							
Helen Lk	2006–2007	0.03 (1.59)	1.43 (5.52)	1.65 (1.39)	6.05 (6.41)	−0.93 (0.65)	−1.33 (1.96)
Gibbs Cyn	2005–2006	2.04 (2.19)		4.26 (1.83)		0.71 (0.96)	
	2006–2007	1.86 (1.82)	3.27 (7.56)	3.85 (0.38)	10.97 (6.61)	0.72 (1.25)	−1.16 (3.32)
Greenstone Lk	2005–2006	1.41 (1.45)		2.28 (1.86)		0.97 (0.93)	
Tamarack Lk	2006–2007	1.76 (3.13)		5.13 (2.12)		−0.26 (1.46)	
Warren Fork West	2005–2006	2.82 (0.28)		2.74 (0.34)		2.86 (0.23)	
Means		*1.65 (1.74)*	*2.35 (6.54)*	*3.32 (1.32)*	*8.51 (6.51)*	*0.68 (0.91)*	*−1.24 (2.64)*

−0.2 °C and fluctuated little (SD, 1.2 °C) compared to adjacent air temperatures, which had a mean of 1.8 °C and fluctuated more diurnally and seasonally (SD, 5.6 °C). Warm-season (May 15 to Sept 30) water/ice temperatures remained between −1.5 °C and 2 °C, with a mean of 0.6 °C, and varied little (SD, 0.6 °C). This contrasted with the warm-season air temperature mean of 6.7 °C, which also had higher variation (SD, 6.0 °C), and ranged from −15 °C to >25 °C. Cold-season (Oct 1 to May 14) mean water/ice temperature (−0.6 °C, SD 1.1 °C) of rock-glacier springs was only slightly lower than those in warm seasons. A common pattern in late autumn and early winter was for the water-iButton temperature to fall below 0 °C, attaining the coldest minimum temperatures of the year during this time, recording a slight annual cold-spike (Figs. 3A and 4A). Temperatures in the cold season thereafter remained between −1 °C and −3 °C with minor fluctuations, interpreted to be ice, having verified this with field observations. This was further assumed to be the snow-covered period. Temperatures continued to increase very slightly until about April, when many temperature profiles jumped to 0 °C and remained at this temperature until thaw. Air temperatures in the cold-season remained near freezing (mean −1.0 °C; SD, 2.1), in some situations hovering at 0 °C and in others dipping to −12 °C during late autumn, documenting a more abrupt but coincident annual cold spike than the instruments in the springs. Thereafter the temperatures remained a few degrees below 0 °C for the rest of the season. The abrupt rise in air temperature and onset of diurnal fluctuation of the air iButtons is interpreted to mark the end of snow cover.

In a few situations water temperatures in early autumn (October to mid-November) showed fluctuations above 2 °C or below −2 °C (Fig. 4A) coinciding with air temperature patterns (Fig. 5A), suggesting that the springs had desiccated. Although in one situation this seems most likely (Barney Lk rock glacier 2007–2008, Fig. 3A), close inspection of the relationship between air and water in Fig. 4A indicates that the water temperatures were generally far lower, and fluctuations far less, than air temperatures even during these times. This suggests either transient drying (e.g., Karolyn Cyn) or that spring pools become shallow where the instrument was located, and the water warmed with the surface air.

4.1.2. Boulder-stream taluses

Water in boulder-stream talus springs was on average cooler year-round with mean annual temperature 1.7 °C and lower daily fluctuations (SD, 1.7 °C) than adjacent air, which had mean temperature of 2.4 °C and greater fluctuations (SD, 6.5 °C; Table 2B, Figs. 3B and 4B). All relevant comparisons of mean differences were significant. A common annual pattern in water temperatures was a stable warm-season temperature ranging from 2 °C to 4 °C and cold-season temperatures hovering near freezing, 0 °C to −1 °C (frozen). During the warm-season, mean air temperatures were significantly higher than water temperatures, with only occasional night temperatures colder than water temperatures. During the beginning and ends of the cold-season, air temperatures fluctuated above and below water temperatures, with minima to −7 °C. During winters, mean air temperatures were similar to water temperatures, although in some situations air was significantly colder. In a few locations and years, water temperatures rose above 4 °C during daytimes in late autumn and summer (e.g., Gibbs, 2005–2006, Fig. 3B), suggesting that the springs desiccated. This situation was observed mostly during the summer of 2007, which was an exceptionally dry and warm year.

Considered by type of landform, mean annual- and warm-season water temperatures of boulder-stream taluses were significantly

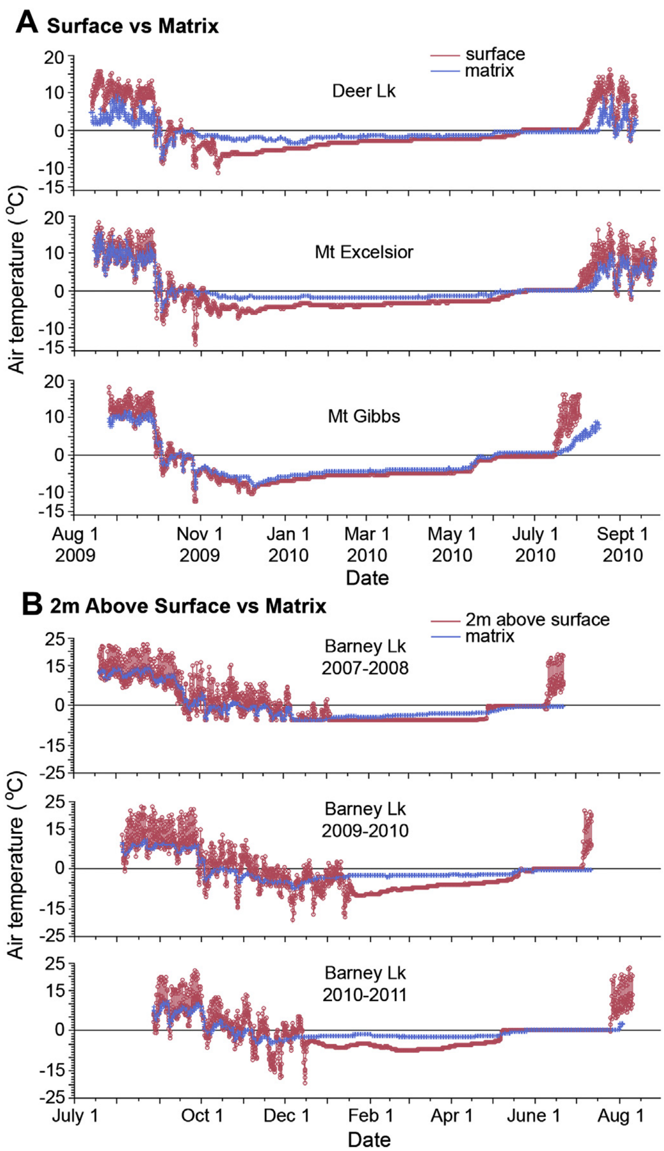

Fig. 5. Air temperatures above rock-glacier surfaces and in rock matrices, 1 m below surface, three locations, 2009–2010. A. Surface-air- and rock-matrix comparisons. B. Air at 2 m height above rock-glacier surface and rock-matrix comparisons at Barney Lk rock glacier over three years.

warmer ($p < 0.05$) than those of rock glaciers taluses and had significantly higher variance (Table 2).

4.2. Air- and rock-matrix temperatures

4.2.1. Rock glaciers

For rock-glacier locations and years with data for pairs of instruments measuring temperature of the matrix (1 m below surface) and adjacent rock surfaces, annual means of matrices were significantly colder than surface air and had lower daily variance (Table 3A, Fig. 5A). Considering data from all locations and years, the mean annual matrix temperature was below freezing (−0.8 °C). During the warm season, surface-air temperatures varied daily between 6°C and 18 °C, with large diurnal fluctuations; matrix temperatures were significantly lower ($p < 0.05$), at or below the range of the surface-air for the respective location, and had lower variance. Transitions from warm- to cold-season in autumn were abrupt and characterized by rapid drop of daily temperatures below freezing with lower diurnal fluctuation (Fig. 5A). For short periods in early autumn, surface-air- and matrix temperatures were often similar in value and diurnal variance. By mid-late autumn, relative order changed in nearly all cases, such that matrix temperatures were warmer than surface-air temperatures.

During the cold season, many pairs of mean temperatures between surface and matrix were significantly different, but orders changed by location, suggesting that effects were determined more by location than by position (Table 3A) Diurnal variance for the cold season was low at both surface and matrix positions. Commonly mid-winter temperatures were below freezing, with surface-air temperatures dipping to −12 °C and matrix temperatures to −5 °C (Fig. 5A). By the end of the cold season a common pattern was for temperature at surface- and matrix positions to remain unvarying at 0 °C. This freezing flat-line condition extended from two weeks to (more commonly) two-three months. This situation terminated abruptly at the beginning of the warm season, varying in date by year and location, when daytime surface air temperatures abruptly rose above freezing, with high annual variance, interpreted as the end of snow cover at these sites. This was followed within one to two weeks by similar transitions, although attenuated, in matrix temperatures.

Although the goal for placing iButtons at 2 m height on poles was to record air temperature above the snow, this appears not to have been achieved. The 2 m temperature trace fell to and remained at 0 °C to −5 °C from winter through early summer, with very little daily fluctuations, indicating snow cover. Nonetheless, matrix temperatures were compared to those 2 m above surface temperatures, understanding that uncovered air temperatures would be much different. When matrix temperatures were compared with temperatures on poles at 2 m above the rock-glacier surface, annual mean temperatures were significantly different for most comparisons; relative order varied, however, with half the cases having overall mean matrix temperature colder than external air (Fig. 5B, Table 3B). The overall mean annual matrix air temperature was below freezing (−0.6 °C). The 2 m above surface temperature fluctuations were greater than when measured at the rock-glacier surface, with extremes ranging between 22 °C and −20 °C. A similar pattern of transitions between warm- and cold-seasons was recorded, however, as for the surface-air versus matrix comparisons, with an abrupt cooling spike in autumn, and freezing flat-line temperatures in late cold season for both 2 m above the surface and matrix temperatures that persisted up to three months. Warm- and cold-season trends also were similar. For the warm season, matrix air temperatures were significantly colder ($p < 0.05$) than external air temperatures. For the cold season, many individual comparisons were significant, but order varied, and overall means were not significantly different.

Explorations using statistical modeling of the relationship between surface and matrix temperatures over time yielded non-linear third-order polynomial functions to have the best fit for many rock-glaciers; talus features showed second-order best fits. An example of the behavior of matrix versus surface temperature is shown for two years of warm-season temperatures at the Barney Lk rock glacier, using the sigmoid fit of the data (Fig. 6). Both show asymptotes in matrix temperatures at high and low surface temperatures. These suggest resistance (hysteresis) in matrix temperatures to warming and cooling of surface temperatures for specific ranges of external thermal conditions. Although there are differences in asymptotes between years, the shape parameters of the models are similar.

The attempt to improve the process base of the model by removing time-series autocorrelations showed similar patterns over the season. The best fit in the autoregressive model used was ARIMA (2,1,2); i.e., 2 days lag in autocorrelation and moving

Table 3
Air temperatures of rock glaciers (°C). A. Surface-air- and rock-matrix (1 m below surface) temperatures. B. Air at 2 m above surface and matrix temperatures. Differences between mean surface- and matrix air temperatures marked with bold font are significant ($p < 0.05$).

Name	Measurement years	Annual temperature				Warm-season				Cold-season			
		Surface		Matrix		Surface		Matrix		Surface		Matrix	
		Mean	Std	Mean	Std	Mean	Std	Mean	Std	Mean	Std	Mean	Std
A. Surface-air and matrix													
Deer Lk snout	2006–2007	1.11	6.38			6.96	7.38			−2.29	1.17		
	2007–2008	−0.85	3.34			2.39	3.83			−2.24	1.37		
	2008–2009	2.28	4.60			5.93	6.30			0.42	1.23		
	2009–2010	−0.48	5.43	−0.66	2.24	**3.97**	**5.40**	**1.04**	**2.35**	**−3.80**	**1.98**	**−1.93**	**0.95**
Deer Lk left lateral	2007–2008			−0.47	2.50			1.45	3.50			−1.39	0.85
	2008–2009			−0.39	2.39			1.57	3.19			−1.29	1.05
	2009–2010	**0.87**	**3.61**	**−1.04**	**4.10**	**3.52**	**4.13**	**2.23**	**4.29**	**−1.11**	**0.91**	**−3.48**	**1.31**
Deer Lk right lateral	2007–2008			0.27	3.45			2.26	4.56			−1.50	1.71
	2008–2009			−0.05	2.23			1.46	3.31			−0.74	0.82
	2009–2010	**0.83**	**3.55**	**0.52**	**3.50**	**3.68**	**4.97**	**3.10**	**4.61**	**−0.59**	**0.62**	**−0.99**	**0.97**
Mt Excelsior snout	2007–2008			−2.39	3.19			0.68	3.65			−3.94	1.22
	2008–2009			−0.07	4.19			3.99	5.35			−2.14	1.48
	2009–2010	**−0.17**	**5.52**	**0.69**	**3.90**	**4.21**	**5.49**	**3.51**	**4.38**	**−3.66**	**1.79**	**−1.55**	**0.81**
Mt Excelsior left lateral	2007–2008			−0.39	6.12			6.43	6.14			−3.80	1.59
	2008–2009			1.19	5.17			6.25	5.99			−1.39	1.56
	2009–2010	**0.06**	**5.33**	**1.88**	**5.95**	**5.34**	**6.08**	**6.87**	**5.69**	**−5.83**	**1.79**	**−4.44**	**1.36**
Mt Excelsior right lateral	2007–2008			−1.74	5.00			4.04	4.60			−4.65	1.31
	2008–2009			−0.96	4.93			4.70	4.30			−3.84	1.65
	2009–2010	**−2.30**	**6.32**	**−1.63**	**5.13**	**4.63**	**6.29**	**3.88**	**5.35**	**−2.63**	**1.61**	**−2.10**	**1.43**
Mt Gibbs upper	2007–2008			0.42	6.81			5.67	6.35			−3.92	3.31
	2008–2009			−2.91	3.97			0.97	3.85			−4.87	2.88
	2009–2010			−3.07	3.99			0.59	3.85			−4.85	2.60
Mt Gibbs lower	2007–2008			−0.56	5.55			5.16	5.87			−3.44	2.07
	2008–2009			−1.77	3.59			2.14	3.53			−3.26	2.23
	2009–2010	**−2.00**	**6.29**	**−1.53**	**4.83**	**4.57**	**6.45**	**3.46**	**4.31**	**−5.35**	**2.32**	**−4.38**	**1.90**
	2010–2011	−0.21	5.18			4.48	6.24			−2.61	1.93		
Karolyn Cyn	2006–2007	4.35	8.40			13.30	6.47			−1.22	2.74		
	2008–2009	2.98	6.77			10.76	5.65			−0.99	2.52		
	2009–2010	1.68	5.98			7.93	6.29			−1.51	1.93		
	2010–2011	4.37	6.11			9.26	5.89			0.34	1.90		
Kuna Pk snout	2006–2007	2.21	7.83			9.65	6.36			−2.75	3.90		
	2008–2009	0.84	4.34			5.13	4.82			−1.25	1.83		
	2009–2010	−0.50	3.94			3.32	4.54			−2.44	1.32		
	2010–2011	0.10	3.38			2.72	4.59			−1.23	1.12		
Kuna Pk left lateral	2007–2008			−3.95	2.05			−1.06	1.03			−4.93	1.22
	2008–2009			−2.89	4.75			2.34	3.52			−5.43	2.80
	2009–2010			−1.93	3.99			2.46	3.74			−4.16	1.46
	2008–2009			0.60	2.70			2.66	3.71			0.42	0.97
Mt Conness snout	2008–2009			−1.37	4.34			2.85	4.44			−3.46	2.28
Mt Conness medial	2008–2009			0.60	2.70			2.66	3.71			−0.42	0.97
	2009–2010			−1.66	4.28			2.44	4.75			−3.76	1.77
Oneida Lk left lateral	2006–2007	3.08	6.91			10.19	6.28			−2.93	2.77		
Oneida Lk snout	2006–2007			2.59	7.99			11.74	4.73			−2.93	2.77
Oneida Lk left lateral	2006–2007	2.59	6.91			11.74	4.73			−1.27	2.05		
Crystal Lk	2006–2007	2.04	5.72			8.83	6.35			−0.68	1.97		
Mt Tenaya	2007–2008	2.81	5.04			6.41	6.14			0.72	2.52		
Means		*1.12*	*5.52*	*−0.76*	*4.18*	*6.47*	*5.68*	*3.25*	*4.29*	*−1.95*	*1.88*	*−2.95*	*1.64*

Name	Measurement years	Annual temperature				Warm-season				Cold-season			
		2 m Air		Matrix		2 m Air		Matrix		2 m Air		Matrix	
		Mean	Std	Mean	Std	Mean	Std	Mean	Std	Mean	Std	Mean	Std
B. Air at 2 m above surface and matrix													
Barney Lk snout	2007–2008	**0.64**	**7.33**	**1.86**	**7.02**	**8.16**	**7.08**	**5.91**	**5.86**	−3.05	3.78	−3.03	1.77
	2008–2009			−0.07	4.84			4.91	4.92			−2.61	2.01
	2009–2010	**−1.08**	**7.59**	**−0.64**	**4.16**	**6.14**	**7.18**	**3.59**	**4.41**	**−4.77**	**4.53**	**−2.81**	**1.52**
	2010–2011	**−0.96**	**6.46**	**−0.51**	**3.09**	**4.33**	**6.15**	**2.11**	**3.47**	**−3.83**	**4.54**	**−1.85**	**1.72**
Barney Lk upper	2007–2008	**−0.09**	**5.57**	**−1.38**	**4.58**	**8.87**	**6.30**	**3.81**	**4.26**	**−1.57**	**4.27**	**−3.92**	**1.66**
	2008–2009	**1.71**	**8.27**	**−1.84**	**4.46**	**9.69**	**5.45**	**2.71**	**4.06**	**−2.35**	**6.25**	**−4.16**	**2.39**
	2009–2010	**2.06**	**8.58**	**−1.42**	**3.67**	**9.89**	**6.10**	**2.69**	**2.98**	−3.51	5.01	−3.53	1.63
	2010–2011	0.34	7.92			7.90	6.37			−3.81	5.12		
Means		*0.37*	*7.39*	*−0.57*	*4.55*	***7.85***	***6.38***	***3.68***	***4.28***	*−3.27*	*4.79*	*−3.13*	*1.81*

average; data were adjusted for the decreasing trend in temperature over season. After removing the autocorrelations, the data were fit with third-order models of air temperatures (2 m above surface) that included two-day lags in temperatures for those locations. Model fits for both years were very high ($R^2 > 0.98$). The form of the fit of the data in the time-series analysis (Supplemental Fig. 1), although more complicated, was similar to the sigmoidal forms shown in Fig. 6A,B. During the warm and cool periods for temperatures 2 m above the rock-glacier surface, matrix temperatures lagged by 1–3 days. Air temperatures 2 m above the surface

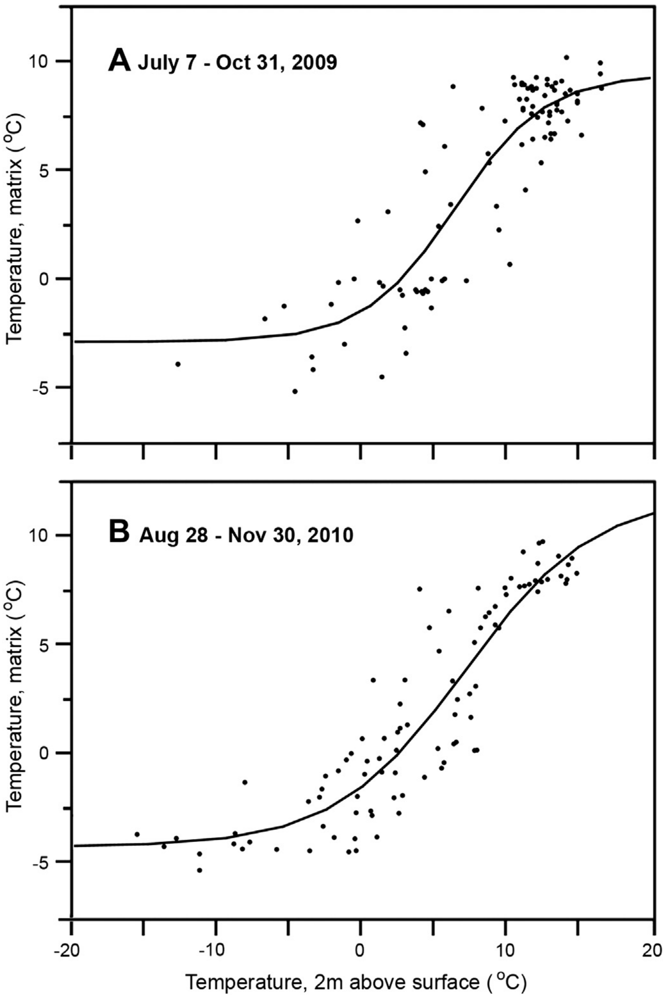

Fig. 6. Regressions of warm-season external air temperature against rock-matrix temperature for two representative years, with best-fit polynomial curves, Barney Lk rock glacier. A. 2009. B. 2010.

were in an approximate weekly cycle of cooling then re-warming (Figs. 5 and 7). Matrix temperatures would remain unchanged for 2–3 days, then respond to the surface temperatures, resulting in cycling behavior in the matrix temperatures (Supplemental Fig. 1). Occasionally during these cycles, 2 m above the surface air temperatures cooled as much as 8–10 °C, the matrix temperatures would cool as much as ∼2–3 °C and then cool no more (Fig. 5B). Subsequently, as the 2m air temperatures switched to a warming pattern, the matrix temperatures would follow, lagged by about two days, but rise only to the mean for the past several weeks, whereas 2 m air temperatures continued to rise above these values. These patterns from the time-series analysis and from raw temperature profiles suggest that rock-glacier matrices resist (or are buffered against) changes in surface temperatures, at least during autumn and for limited time periods, as suggested below.

By late September to early October in each year, the 2 m above the surface temperature traces, as well as other locations described previously where surface-matrix contrasts were recorded, showed an abrupt cooling by as much as 13 °C (Figs. 4, 5 and 7). In these situations, matrix temperatures declined by about half the amount as the external air, suggesting a partial loss of matrix resistance to cooling. Prior to significant snowfall accumulation (after mid-late November, as recorded from the Mammoth Lks, CA weather station), even when surface temperatures would rise, the matrix

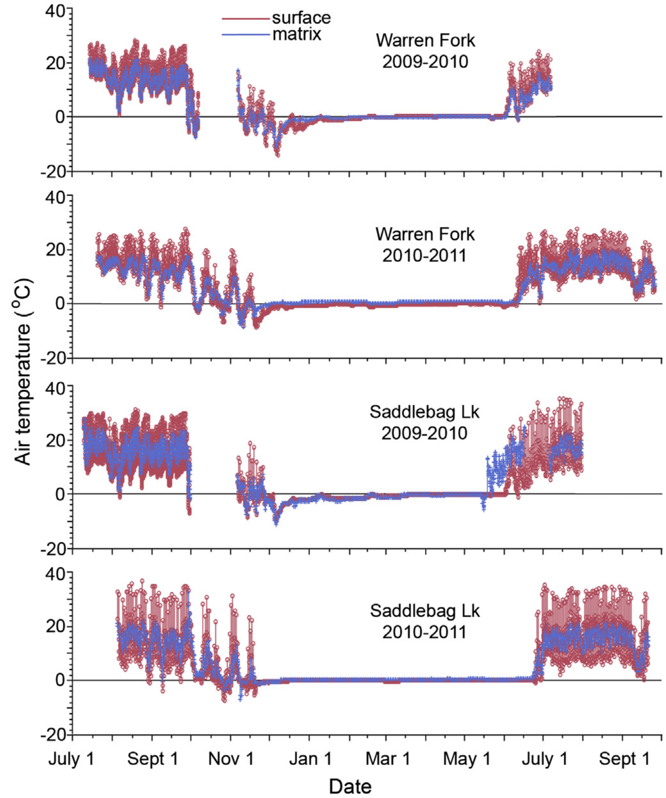

Fig. 7. Air temperatures at two boulder-stream talus surfaces and in rock matrices, 1 m below surface, 2009–2011. Interrupted lines indicate missing data.

temperatures generally remained cool. With accumulation of snow at the surface, diurnal fluctuations in the matrix temperatures declined to nil, and a winter pattern set in where little relationship was apparent between temperatures at the matrix and the surface.

4.2.2. Boulder-stream taluses

Trends and relative patterns for boulder-stream talus surface-air- and matrix-air temperatures were similar to those for rock-glaciers, although differences between air and matrix were smaller (Fig. 7). In paired comparisons, differences in mean annual temperature of matrices (5.2 °C) were not significantly lower than respective mean surface air temperature (5.5 °C) although three of five individual comparisons were significant (Table 4). Diurnal fluctuations were lower in matrices than in surface air temperatures. Similar to rock glaciers, during the warm season four of five individual comparisons showed matrix mean air temperatures significantly colder than the surface. Further, the abrupt cold spike observed in autumn for rock glaciers was reflected in matrix and surface temperatures at all talus sites as well. The late-cold-season freezing flat-line that occurred for rock glaciers also occurred in taluses, making transitions between cold- and warm seasons abrupt. Compared to rock glaciers, mean- and warm-season temperatures of boulder-stream taluses were significantly warmer. This might reflect southerly aspects of some taluses in these analyses, corroborated by the cooler temperatures at Greenstone Lk, which was the only fully northern-aspect talus available for this comparison (Tables 1 and 4) or that instruments were more distant from embedded ice.

Regressions of surface-air and rock-matrix temperature for taluses were generally second-order; none were third-order (third-order parameter estimates were near zero). In general, asymptotes were outside the range of the temperatures at the surface. Above

Table 4
Air temperatures at boulder-stream talus surfaces and in rock matrices 1 m below the surface (°C). Significant differences ($p < 0.05$) between mean surface- and matrix air temperatures are noted with bold font.

Name	Measurement years	Annual temperature				Cold-season				Warm-season			
		Surface		Matrix		Surface		Matrix		Surface		Matrix	
		Mean	Std	Mean	Std	Mean	Std	Mean	Std	Mean	Std	Mean	Std
Saddlebag Lk	2009–2010	**9.18**	**9.67**	**9.21**	**8.60**								
	2009–2010					**−1.33**	**2.32**	**−1.63**	**2.02**				
	2010									**12.23**	**9.27**	**10.64**	**5.40**
	2010–2011	4.39	8.22	4.26	6.74								
	2010–2011					0.69	3.66	0.63	2.45				
	2011									**10.19**	**9.95**	**9.56**	**7.30**
Warren Fk East	2009–2010	**7.53**	**8.65**	**6.70**	**7.50**								
	2009–2010					−0.80	3.40	−0.88	2.65				
	2010									**12.26**	**7.22**	**10.00**	**5.62**
	2010–2011	4.24	7.77	4.08	5.93								
	2010–2011					**−0.14**	**3.29**	**0.67**	**2.40**				
	2011									**10.87**	**7.54**	**9.92**	**6.08**
Greenstone Lk	2010–2011	**2.00**	**4.73**	**1.76**	**4.73**								
	2010–2011					**0.07**	**1.40**	**−0.29**	**1.52**				
	2011									4.15	5.97	4.16	5.56
Means		*5.47*	*7.81*	*5.20*	*6.70*	*−0.30*	*2.81*	*0.01*	*2.16*	*9.94*	*7.99*	*9.46*	*5.99*

~17 °C, however, rate of warming of the matrices declined with increasing external temperature.

5. Discussion

Persistent internal ice in Sierra Nevada rock glaciers is implicated by temperature of outlet-spring water remaining near 0 °C year-round, and by the persistent presence of water (or ice) at springheads. The latter is inferred from the significantly lower temperatures and attenuated daily fluctuations of water compared to adjacent air during the warm season and water temperatures remaining near 0 °C in the cold season, when air temperatures were much colder (implying under snow). Even when water temperatures showed daily fluctuations in late summer, their significantly colder mean temperatures and lower diurnal ranges than air suggest that water persisted but pools were likely shallow. Water temperatures at and near freezing throughout the warm season imply close proximity of the outlet spring to melting snow or ice, presumably within the rock-glacier matrix. During the cold-season, rock-glacier spring water appeared most often to be frozen, a situation confirmed with field observations in late autumn and spring.

In contrast to rock-glacier springs, water flowing from periglacial taluses does not appear to derive from ice in close proximity to springs, in that mean temperatures during the warm season were often above freezing (as warm as 5 °C). Ice is still possible within these taluses, but lies more distant from outlet springs than in rock glaciers, and/or is transient during the warm season. Similar to rock-glacier outlet water, most talus springs did not desiccate but remained persistent year-round (frozen in winter), as inferred by comparison to adjacent air. In a very few cases, temperatures taken in the outlet springs matched those of air above them, and had similar diurnal fluctuations, implying that the springs had dried. This occurred at a few lower-elevation taluses and during 2007, which was an unusually dry and warm year in the Sierra Nevada. During that year, many non-rock glacier and non-talus derived small springs and streams that normally flow year-round had dried by late autumn. Springs originating from rock glaciers and talus fields were among the few locations (excepting major springs) supplying output that year.

Based on interpretations of water (liquid and frozen), rock-glacier/talus matrix, and external air temperatures, snow-covered durations were estimated for a subset of landforms for the years 2006–2011 (Table 5). The annual abrupt autumn drop in temperature (cold spike) occurred in striking synchrony among years, between October 1 and October 3 in all cases. These cold spikes coincided with arrival of cold northern air masses in the region, as indicated by weather records from nearby Mammoth Lakes, CA, Coop Weather Station for these years. Snow did not necessarily coincide with the cold spikes in the years of analysis. At this time in autumn (Sept. through early November), only minor snowfall events are recorded in the Mammoth Lks, CA weather archives, and storms that deliver snow are infrequent. Relative to the cold spikes, snow, if any, came a few days later, suggesting that transfer of cool temperatures to the matrix was not dependent on snow cover. Duration of the snow cover varied by individual location (Table 5) and by year. In general, rock glaciers had longer duration of snow cover than taluses (means 227 and 175 days, respectively), reflecting their higher and more northerly situations. For the group of rock glaciers where sample sizes were greater, the winter of 2006–2007 was the year of shortest snow cover and 2008–2009 the longest.

The potential consequences of rock glaciers and taluses maintaining snow cover, developing internal ice lenses, and delivering stable hydrologic output are important in semi-arid and Mediterranean-climate mountain ranges such as the Sierra Nevada, and especially during warming climates as present. Meltwater from snow, internal ice (potentially refreezing as congelation ice), permafrost, and/or stable groundwater within and below these features appears to provide an important hydrologic reservoir through the long warm-dry season as other high-elevation sources in the Sierra Nevada, such as snowpack, diminish (Raub et al., 2006; Maurer, 2007). This output contributes to streamflow and downslope recharge, and provides proximal, persistent water that supports abundant wetlands in high-elevation canyons at the base of these features. In the otherwise arid alpine environment, wetlands supported by these springs are critical habitat for a host of alpine biota, some of which, such as the American pika (*Ochotona princeps*), depend on wetland habitats that are supported by adjacent talus fields and rock glaciers (Millar and Westfall, 2010). Wetlands also act as sponges to retain water in upper-elevation basins, in contrast to meltwater from annual snowpacks and ice glaciers, which more typically exit the uplands directly in incised channels.

What enables persistent ice, permafrost, and/or high groundwater reservoirs to develop in these landforms in warm, dry mountain regions such as the Sierra Nevada? A combination of

Table 5
Seasonal thermal events interpreted from temperature measurements of rock glaciers and taluses depicted in Figs. 3–6. A. Dates estimated for the onset of autumn cold spike, beginning of snow cover and date snow-free. B. Estimates of snow-covered duration.

Type	Name	2006–2007			2007–2008			2008–2009			2009–2010			2010–2011		
		Date autumn cold spike	Date snow covered	Date snow free	Date autumn cold spike	Date snow covered	Date snow free	Date autumn cold spike	Date snow covered	Date snow free	Date autumn cold spike	Date snow covered	Date snow free	Date autumn cold spike	Date snow covered	Date snow free
A. Seasonal events																
Rock glaciers	Barney Lk		12/20	5/3	9/20	12/10	6/4	10/11	11/1	6/20	10/1	1/15	7/1	10/1	12/17	7/25
	Deer Lk							10/11	11/1	7/20	10/1	11/15	8/1			
	Karolyn Cyn							10/11	11/1	5/20						
	Kuna Pk							10/11	11/1	6/28						
	Mt Excelsior										10/1	11/15	8/2			
	Mt Gibbs										10/1	11/15	7/15			
Boulder-stream taluses	Gibbs Cyn	10/3	11/15	4/22												
	Helen Lk	10/3	11/15	6/10												
	Warren Fk										10/1	12/10	5/21	10/1	11/17	6/3
	Saddlebag Lk										10/1	12/10	5/28	10/1	11/17	6/22

Type	Name	Snow-covered duration (days)					
		2006–07	2007–08	2008–09	2009–10	2010–11	Mean
B. Snow-covered duration							
Rock Glaciers	Barney Lk	134	177	232	204	220	193.4
	Deer Lk			262	263		262.5
	Karolyn Cyn			232			232.0
	Kuna Pk			271			271.0
	Mt Excelsior				259		259.0
	Mt Gibbs			242			242.0
	Annual mean	134.0	177.0	247.8	242.0	220.0	
	Overall mean						**226.9**
Boulder-stream taluses	Gibbs Cyn	157					157.0
	Helen Lk	206					206.0
	Warren Fk			162		167	164.5
	Saddlebag Lk			169		186	177.5
	Annual mean	182		166		177	
	Overall mean						**174.5**

factors seems to contribute to this. Topographic contexts of the features are such that local climates tend to be cooler than means expected for their elevations (Millar and Westfall, 2008), and occur in locations where snow develops and lingers (Haeberli, 2005). This results partly from northerly aspects and also from local conditions that favor cold micro-sites and periglacial processes, including steep rocky slopes; narrow and deep canyons; cirque head locations; slopes below cliffs, rock chutes, and avalanche tracks; and other contexts where solar insolation is reduced and cold-air drainage, cold-air pooling, and inversions are common (Lundquist et al., 2008; Millar and Westfall, 2008). Rock glaciers and periglacial taluses often develop below persistent snow- or ice fields at the base of cliffs, a location that likely contributes to initial development of the landforms and then supplies meltwater as features enlarge. These locations also are below cliffs where freeze-thaw and other erosion forces deliver abundant rock-fall and avalanching; together these favor build-up of snow-ice-rock mixtures, which would be incorporated into the landform.

Beyond topographic setting, internal thermal regimes of rock glaciers and talus blockfields have been shown to be decoupled from external micro- and meso-climates (Juliussen and Humlum, 2008), which these results corroborate. During the warm season, negative thermal anomalies develop, wherein air temperatures within rock matrices are significantly lower than air at and above the rock surface. Anomalies between rock matrix and adjacent mineral soil have been recorded as large as 4–7 °C (Harris and Pedersen, 1998) and up to 2 °C colder than adjacent morainal till and bedrock (Juliussen and Humlum, 2008). Cold air is often felt as a persistent breeze that flows out the base of the landform (Delaloye and Lambiel, 2005). Several internal circulation mechanisms involving advection and convection have been proposed to explain these intra-matrix anomalies, including Balch effect, chimney-flue ventilation, evaporation–sublimation processes, and the effect of continuous air exchange with the atmosphere (Harris and Pedersen, 1998; Juliussen and Humlum, 2008). These processes likely also contribute to the rock-glacier matrix resistance to warming relative to free air measured, where internal temperatures lagged response to changes in external air temperature, and warmed to threshold temperatures, but did not continue to warm or cool even as external air did.

The records also document regular occurrence of a late winter to spring "zero curtain" condition that is described for coarse soils (*sensu* Outcalt et al., 1990) and recorded in the data by persistent flat-line temperature traces (~0 °C) in matrix and surface air. These conditions appear to develop as a result of heat-transfer processes during freezing and thawing, involving characteristic changes in freezing-water and meltwater ion concentrations, progression of freezing downward, and consequent effect of latent heat maintaining temperatures near 0 °C (Outcalt et al., 1990). In rock glaciers and taluses, these heat-transfer processes potentially contribute to the formation of persistent internal ice, especially given the presence of deep mantles of rock debris and open air voids. Had thermochrons been deployed deeper in the rock matrix nearer to embedded ice, colder temperatures might have been recorded.

A consequence of environmental context, decoupled thermal processes, and resistance to warming is the presence of hydrologic reservoirs related to rock glaciers and periglacial talus fields. With projections in the Sierra Nevada for warming air temperatures, diminishing snowpacks, earlier peak streamflows, and longer, drier summers (Moser et al., 2009), the occurrence of these landforms throughout high elevations could serve increasingly important roles in mountain hydrology and ecosystem function.

6. Conclusions

The open-lattice matrices of rock glaciers and periglacial taluses in the Sierra Nevada maintain cool-buffered thermal regimes that appear partly decoupled from external air during some periods of the year. A combination of factors, including topographic location and intra-rock-matrix circulation, leads to annual intra-matrix temperatures of rock glaciers below freezing, conditions that point to development of permafrost and/or retention of persistent ice. The latter are corroborated by persistent flow of outlet springs at rock-glacier- and talus snouts, with cold temperatures persisting through the warm season. Spring temperatures near 0 °C in the case of rock-glacier springs suggest proximity to ice within rock matrices. Springs derived from these features support persistent wetlands and lakes at their snouts, retaining water in the otherwise semi-arid high cirques, and contribute importantly as hydrologic reserves in high Sierran cirques and to critical habitat for alpine biota. Lag and buffering effects between surface and matrix temperatures suggest that ice within these landforms resists surface warming, and thus that rock glaciers and periglacial taluses are increasingly likely to be important to mountain hydrology and alpine ecology as regional climates change.

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Appendix A

After modeling best statistical fits with regression analysis, to better explore heat exchange between rock glacier surfaces and matrices, temperature data were assessed using SAS PROC ARIMA. The autocorrelation was removed from the time-series data using autoregressive integrated moving average (ARIMA) models, and then third-order models of daily mean surface air temperatures that also included 1–2 day lags to address heat-transfer from the surface to the matrix were regressed. Lines in each graph are predicted values of matrix temperatures, given modeled pole temperatures for the time series. Dots are the actual data. In 2009, the time series begins July 7 and ends October 31, 2009; in 2010, the time series begins August 28 and ends November 30. These time series were selected as follows: The beginning of the series is the date the matrix thawed and the series was truncated at a date when variation in the matrix temperatures declined to minimal values. Numbers on the graphs indicate the sequence of dates in the time series. Transitions from summer to autumn temperature (abrupt cold spikes) occur around September 30, 2009 and October 3, 2010. Resistance in the matrix temperatures, including looping responses, is reflected in the non-linear responses in the matrix temperatures as well as lagged responses to changes in the pole temperatures.

Appendix B. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.quaint.2012.07.019.

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