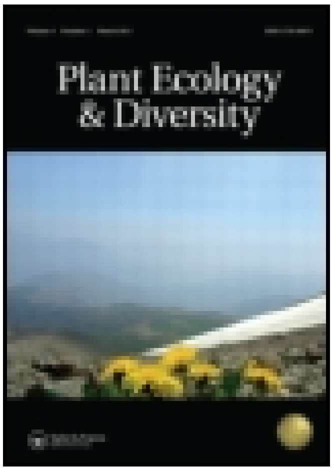

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Cryosphere: ice on Niwot Ridge and in the Green Lakes Valley, Colorado Front Range

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Cryosphere: ice on Niwot Ridge and in the Green Lakes Valley, Colorado Front Range

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Background: Subsurface ice preserved as ice lenses and within rock glaciers as well as glacial and lake ice provides sensitive indicators of climate change and serve as a late-season source of meltwater.
Aims: We synthesise the results of geomorphological, geophysical and geochemical studies during the period of 1995–2014, building on a long history of earlier work focused on ice and permafrost studies on Niwot Ridge and the adjacent Green Lakes Valley (GLV), which is part of the Niwot Ridge Long-term Ecological Research Site.
Methods: These studies are discussed in the context of how bodies of ice and rock glaciers reflect changing local climate. We review recent results from geophysical investigations (resistivity, seismic refraction and ground-penetrating radar) of the shallow subsurface, ongoing monitoring of the Arikaree Glacier, three rock glaciers and lake ice in the GLV, and interpretations of how subsurface ice melt regulates the flow and chemistry of alpine surface water after seasonal snowfields melt.
Results and conclusions: Permafrost conditions reported from Niwot Ridge in the 1970s are generally absent today, but ice lenses form and melt seasonally. Ice is present permanently within the Green Lakes 5 rock glacier and at nearby favourable sites. The Arikaree Glacier has shown a marked decline in cumulative mass balance during the past 12 years after a 30-year period when net mass balance was ca. 0. Duration of seasonal lake ice increases with elevation in GLV, but duration has decreased at all seven lakes that have been monitored during the last three decades. This decrease has been most marked at the lowest elevation where it amounted to a reduction of about 1 d year^{-1} and least at Green Lake 5 where the loss has been at a rate of 0.5 d year^{-1}. Surface temperature measurements from rock glaciers have not shown strong trends during the past 15 years. It has been suggested that almost all of the 2.5-mm year^{-1} increase in stream discharge from the upper GLV in September and October has been derived from melting of subsurface ice.

Keywords: Arikaree Glacier; climate change; Green Lakes Valley; ice; Niwot Ridge; permafrost; rock glacier

Introduction

The cryosphere – the portion of the Earth's surface where water is in solid form for at least 1 month of the year – has been shrinking in response to climate warming. Most alpine glaciers are retreating, permafrost has been melting, the area and thickness of Arctic sea ice has been shrinking, the depth and duration of winter snowpacks are diminishing, and the seasonal ice cover on lakes and rivers has been appearing later in the year and melting out earlier (Fountain et al. 2012).

Alpine glaciers

Alpine glaciers form where snow accumulation during the winter exceeds ablation in the following summer. If net accumulation lasts for several years, the snow compresses and recrystallises into dense ice that flows slowly to lower elevation, sculpting characteristic glacial landscapes. Alpine glaciers in temperate climates are sensitive to short- and long-term climate variability. In addition to changes of their terminal location along a valley, glacier mass balance is a significant measure of glacier 'health' and release of meltwater. The database of the World Glacier Monitoring Service shows that, on a global scale, alpine glaciers have had an average cumulative mass loss of 13.6-m ice over the past 30 years (Pelto 2012), with a consistent trend. Most alpine glaciers have not yet reached equilibrium, and additional mass losses are predicted for the near future. New monitoring techniques using remote sensing, such as the Gravity Recovery and Climate Experiment (GRACE) satellite (Rodell et al. 2011), have confirmed terrestrial measurements and estimated a total contribution of ice melt to sea level rise of 0.41 ± 0.08 mm, excluding the ice shields of Antarctica and Greenland (Jacob et al. 2012).

Lake ice

Long-term records of ice phenology, the timing of freeze and break-up, ice cover thickness and duration and other seasonal measures of lake ice potentially are useful indicators of climate change, particularly that related to air temperature. Ice formation and break-up dates also depend on cloud cover, snow cover, wind and other local variables. Long records of lake ice in northern Europe and North America generally show trends of later freezing and

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earlier break-up since about 1900 (Magnuson et al. 2000); with rates of change of ca. 1 day per decade (Benson et al. 2012). Many lakes with long ice records are in Europe or North America, but modelling and sparse records from central and eastern Asia suggest that lake ice trends may be broadly comparable throughout the northern hemisphere (Walsh et al. 1998). Weyhenmeyer et al. (2011) demonstrated that latitudinal trends of lake ice phenology are complex and that decadal rates of changes decrease with increasing latitude. Records of alpine lake ice are necessarily shorter, but no less important since the duration of ice cover directly affects lake ecology and is likely to affect geochemical and biological characteristics of the outflows further downstream.

Permafrost and patterned ground

Permafrost occurs when ground temperatures remain <0°C for two consecutive years (e.g., Williams and Smith 1989). Permafrost can be categorised as continuous, discontinuous or sporadic; Smith and Riseborough (2002) provided an overview of its worldwide distribution. Mountain permafrost composes only a small fraction of the worldwide permafrost distribution (Haeberli et al. 1993), but it is known from high altitude regions throughout the world, for example the Himalaya in Asia (Jin et al. 2000), the Alps in Europe (Harris et al. 2003) and the Rocky Mountains in North America (Janke 2005a). During the summer months, the upper part of the permafrost thaws, producing the active layer. Thickness and temperature of the active layer and underlying permafrost provide excellent indicators of temperature variations. Christiansen et al. (2012) highlighted the continuing warming of the permafrost, especially in the Northern Hemisphere over the past few decades. As a consequence, active layer thicknesses have increased and large areas of sporadic and discontinuous permafrost have disappeared. Similar trends have been observed in the European and Asian zone of alpine permafrost (Haeberli et al. 2010; Zhao et al. 2010; Haeberli 2013).

The definition of permafrost is independent of ice but where sufficient water is present, ice lenses grow in the subsurface. Their seasonal expansion and contraction and saturation of the near-surface produce a wide array of periglacial features, including patterned ground and gelifluction lobes (Benedict 1970). In cold climates where ground temperatures >0°C occur during the summer, the growth and decay of seasonal ground ice may also produce a variety of periglacial features.

Rock glaciers and late season hydrology and geochemistry

Rock glaciers are lobate, steep-fronted landforms of periglacial landscapes (Barsch 1996). They develop in steep terrain, usually adjacent to active sources of talus. Rock glaciers resemble small glaciers and their slow downslope motion reflects deformation of a buried ice mass or interstitial ice. Compared with glaciers, rock glaciers are less well studied, in part because they are covered with rocks and appear superficially similar to moraines, rockfall and talus slopes. Their hydrologic significance in high-elevation catchments may be substantial (Millar and Westfall 2008) because ice stored in rock glaciers and as ice lenses in nearby areas may provide notable amounts of water storage and run-off during the summer (Clow et al. 2003) and influence the geochemistry of streamflow. Their ecological significance also is high. Recently, Azócar and Brenning (2010) and Brenning and Azócar (2010) described the hydrological role of rock glaciers in the dry Andes. Their hydrologic significance may become increasingly important for alpine ecology in the future, as seasonal snow and ice melt earlier under rising temperatures (Harris et al. 2003; Millar and Westfall 2008) and there is increased competition by the flora and fauna for scarce water resources.

The cryosphere at the Niwot Ridge LTER programme

The cryosphere is an important component of the Niwot Ridge (NWT) LTER programme. For the past 2.5 M years, glaciation has been a major force in shaping the topography and landscape types characteristic of the NWT site, the Colorado Front Range and the southern Rocky Mountains. Today, the cryosphere helps to support ecosystem processes in the alpine landscape (Seastedt et al. 2004). Loss of seasonal and long-term ice mass in alpine areas has both important ecological implications and is of significant concern for regional water managers. Research on cryospheric topics dates back many decades at the NWT LTER and in nearby areas. In 1965, Outcalt and MacPhail (1965) studied the Arikaree Glacier and produced a detailed map of its extent. Evidence of frozen ground and associated characteristic geomorphic features such as solifluction lobes and terraces, stone stripes or patterned ground were recorded in the 1970s at several locations along NWT and nearby areas (Benedict 1970; Caine 1986). Rock glacier investigations in Colorado began in the 1940s, when Ronald Ives described several rock glaciers and other periglacial landforms in the Front Range (Ives 1940).

Over that past 15 years, research on NWT and in the Green Lakes Valley (GLV) has led to a greater understanding of the extent and changes in the cryosphere. Here we review this research, placing it in the broader context of the alpine landscape and relative to other alpine and non-alpine systems. We focus on (1) the present mass and mass balance of Arikaree Glacier; (2) ice thickness and duration on lakes in the GLV system; (3) evaluation of permafrost distribution and periglacial features, such as solifluction lobes; (4) rock glacier morphology; and (5) importance of the cryosphere for late-season hydrology. We take advantage of many different field techniques that have been employed over the last 15 years to improve the understanding of the cryosphere, including mapping, excavation, thermal measurements and modelling, and various geophysical methods (Janke 2005a, 2005b). In particular, we highlight the use of non-invasive geophysical techniques, such as shallow seismic refraction, ground penetrating radar (GPR) and electronic resistivity tomography (ERT), which have become increasingly popular methods for detailed observation of the near-subsurface (Leopold et al. 2008; Leopold, Dethier, Völkel, Raab, Rikert, et al. 2008; Leopold, Voelkel, et al. 2013; Leopold, Völkel, et al. 2013).

Site description

The NWT area comprises a series of high mountain surfaces and slopes and the adjoining formerly glaciated GLV to the south. Bedrock exposed in the area consists of fractured Precambrian biotite gneisses and granitic rock and Paleogene quartz monzonite, locally covered by Pleistocene glacial and periglacial deposits and Holocene talus. During the latest Pleistocene, ice filled two-thirds of the GLV and adjacent valleys, but NWT and adjacent slopes did not support glacial ice. Deglaciation of GLV was completed by ca. 14 ka and only the Arikaree Glacier (9 ha) persists in the valley (Dühnforth and Anderson 2011). Areas of NWT do not support a snow cover during most of the winter due to strong winds; snowfall accumulates mainly by wind drifting in topographic hollows (Erickson et al. 2005).

As a result of the glacial history, most of the surficial sediment on the floor of the GLV formed during ice retreat or is of Holocene age, deposited by ongoing slope and periglacial activity (e.g. Caine 2001; Williams et al. 2006; Leopold et al. 2011). Talus creep, rock fall and local fluvial activity also modify the surface. Three small rock glaciers (Figure 1) extend from steep, north-facing slopes to the floor of GLV. In contrast, most of NWT and its adjacent slopes were above the glacial limit and layered slope deposits as thick as 15 m, and an early Pleistocene diamict cover the bedrock in most areas (Gable and Madole 1976; Madole 1982). Study sites at NWT are located in the alpine vegetation zone, where mean annual air temperature (MAAT) at 3740 m is −3.8°C and mean annual precipitation (MAAP) is 993 mm (Barry 1973; Greenland 1989; Williams et al. 1996; Greenland and Losleben 2001) at site D1 (Figure 1). MAAT at the Saddle site, at an elevation of 3528 m, is −2.1°C, based on a temperature record from 1982 to 2012, and is warmer than at D1 since start of measurements there. On a linear regression model, positive degree-day values (+DD) at D1 rose from 1953 to 2012. In contrast, negative degree-day values (−DD) showed some variation during 4 years in the early 1980s, but the linear trend does not indicate warming (Figure 2). To generalise, summers appear to have become warmer, whereas winters have shown rather stable temperatures over the past 60 years.

Over the past 10s of 1000s of years, processes related to alpine climate led to distinct surface morphology characterised by terraces, lobes, taluses and areas of patterned ground (Figure 3).

Glacier mass and mass balance

Horizontal changes in the boundaries of the Arikaree Glacier since the study by Outcalt and MacPhail (1965) have been small. However, the change in thickness has been massive, as discussed below, using mass balance data. Measurements using GPR showed recent depth and subsurface boundaries of the glacier. Several lines were

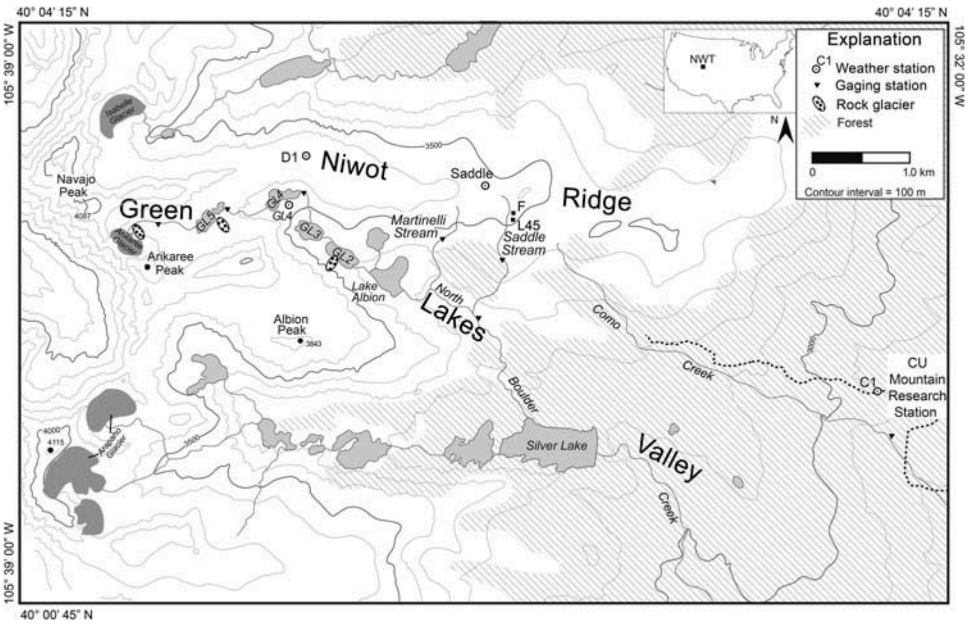

Figure 1. Map of Niwot Ridge, the Green Lakes Valley and surrounding area. Site include Silver Lake, Lake Albion, Green Lake 1 (Gl1), Green Lake 2 (Gl2), Green Lake 3 (Gl3), Green Lake 4 (Gl4), and Green Lake 5 (Gl5), Fahey site (F), Lobe 45 site (L45), and the three rock glaciers, including RG5 and RG2, adjacent to Gl5 and Gl2, respectively (modified from Miller and McKnight, D (this volume)).

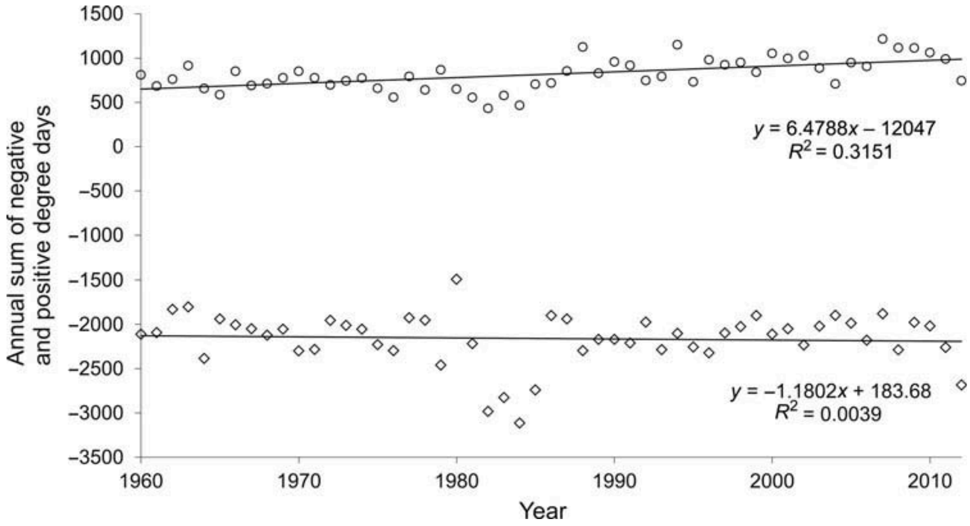

Figure 2. Annual sum of negative (diamonds) and positive-degree days (circles) based on mean daily air temperature at D-1 (3740 m elevation), Niwot Ridge, 1960–2012 (modified from Leopold, Völkel, et al. 2013 with some updated values). *Reproduction licence obtained 14-Jan-2015, www.schweizerbart.de*

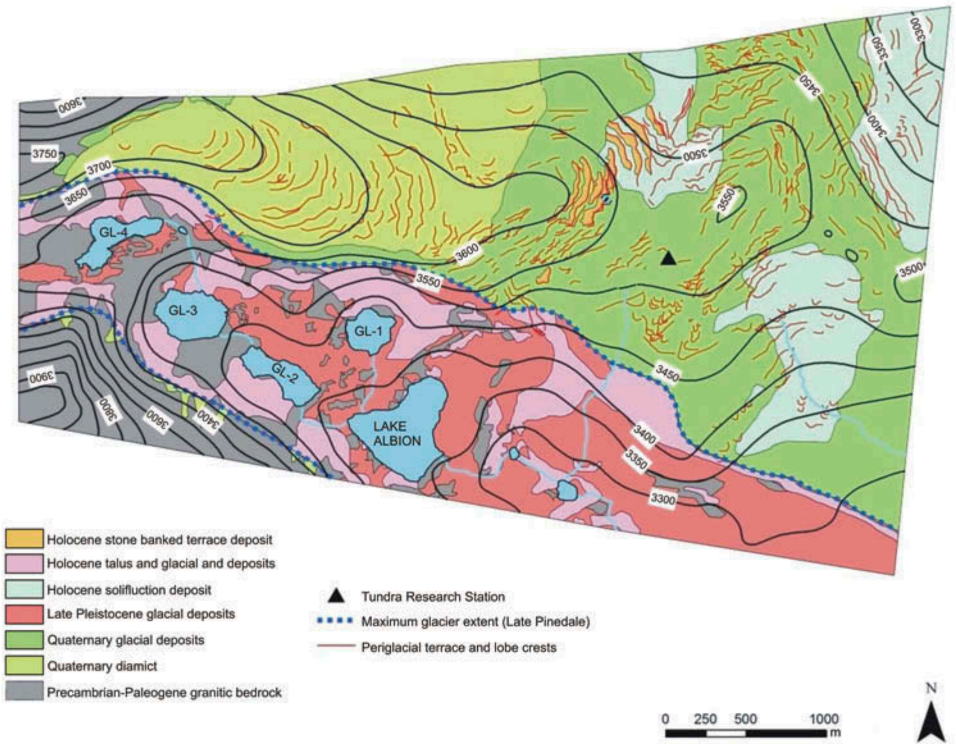

Figure 3. Map of surficial deposits and periglacial landforms, Niwot Ridge and Green Lakes Valley. Periglacial features interpreted from 1-m LiDAR base (Anderson et al. 2012) and geology modified from Benedict (1970) and from Gable and Madole (1976). Note that geologic mapping did not extend west into the upper Green Lakes Valley.

collected during the summer 2009 and the position of one is displayed in Figure 4. Glacier thickness was estimated by applying different velocities, which were calculated from common midpoint analysis to the time-dependent electromagnetic GPR signal (Figure 5). A maximum depth of 20–25 m was calculated with more shallow parts at the top and the toe of the glacier. The velocity of the electromagnetic signal of 0.16 m ns^{-1} indicates pure ice as shown in other studies (Davis and Annan 1989; Eder et al. 2008).

Mass balance of the glacier is based on field measurements of accumulation at the end of winter and ablation measurements taken approximately weekly in summer (Figures 4, 6). Years with missing observations have been estimated

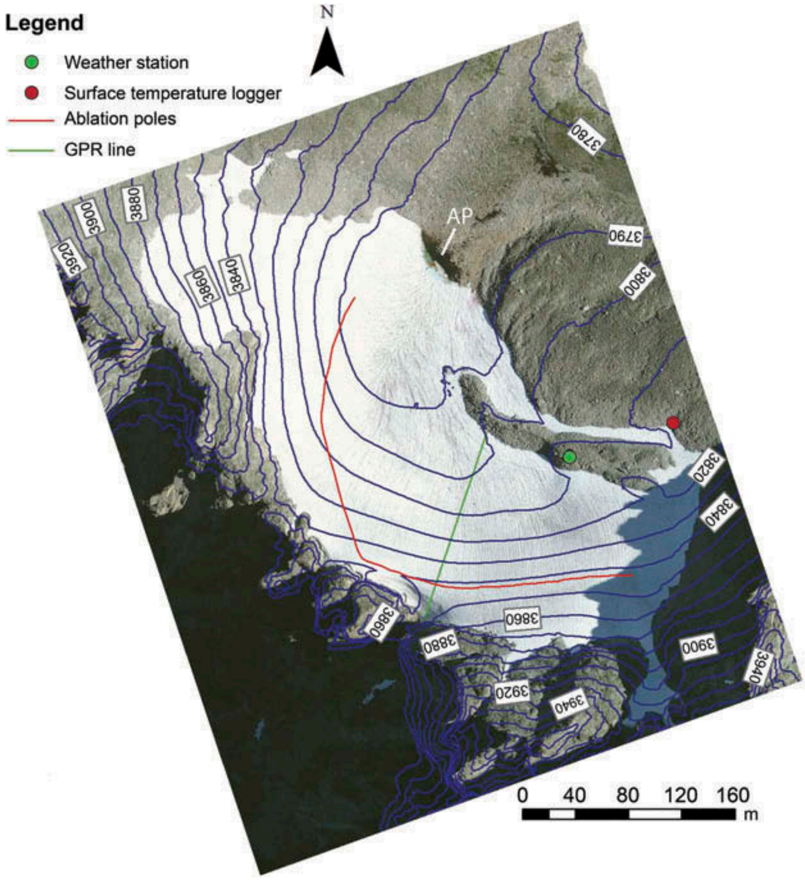

Figure 4. Aerial photograph view of Arikaree Glacier (http://www.bing.com) in 2011, showing the location of ablation measurements and GPR profile. AP, Arikaree Pond; Navajo Pond is ca. 250 m downstream near the upper right hand side of the image.

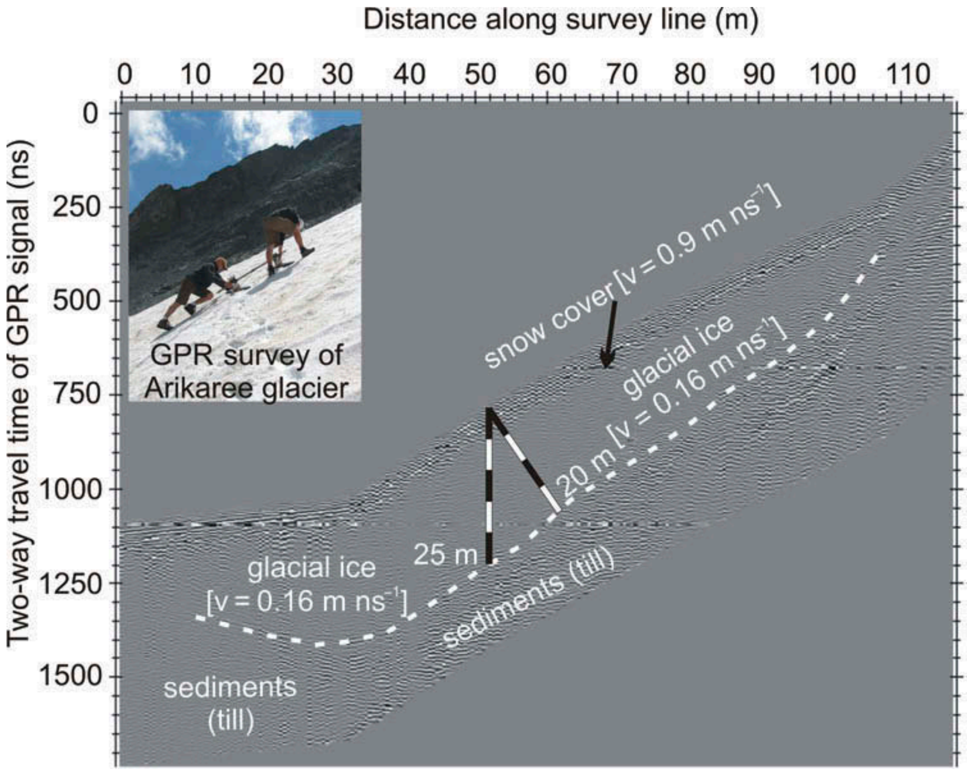

Figure 5. Bottom: Radargram along the indicated line in Figure 4. Glacial ice is represented by a low reflection zone, above the strong reflections from till and probably bedrock. On top the recent snow cover, partly compressed, is indicated by strong reflections parallel to the slope. Glacial ice thickness reaches 25 m as a maximum. The inset image shows GPR data collection during the summer of 2009.

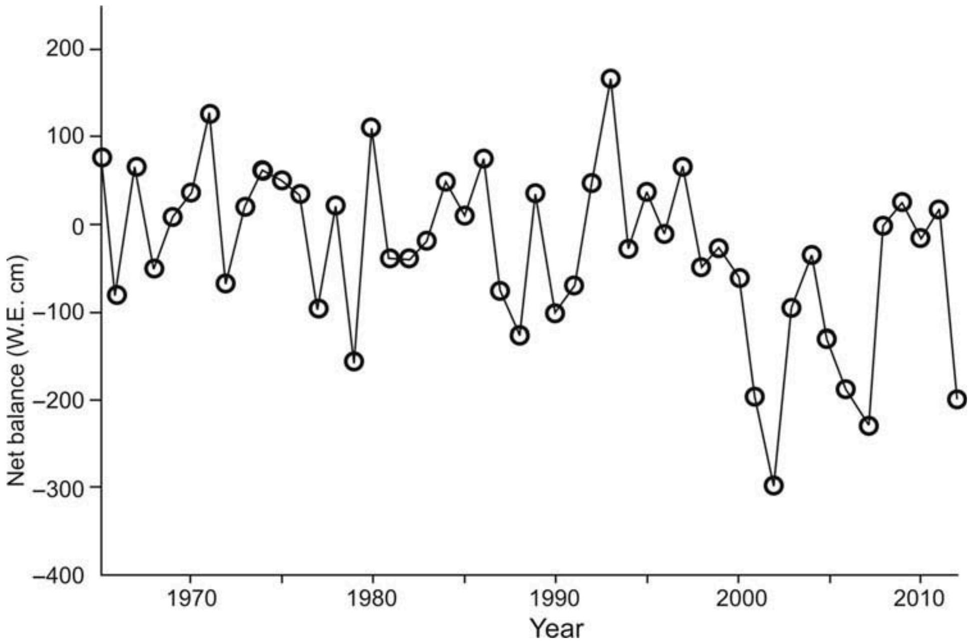

Figure 6. Annual mass balance of Arikaree Glacier, 1965–2012 (from Hoffman et al. 2007, updated to 2012).

indirectly from seasonal temperature (for ablation) and precipitation (for accumulation) records at D-1. The record shows a marked decline in the cumulative mass balance in the last 12 years. The cumulative change in annual balance from 1965 to 1997 was about zero (mean = 3.7 cm water equivalent (W.E.) with S.E. = 13.1 cm; with 19 of those years showing a positive balance and 14 a negative balance). In the following 16 years, 1998–2012, the mean balance has been clearly negative (mean = −100.8 cm W.E. with S.E. 26.3 cm) with only two positive years and 14 negative years, including 2002 when the balance was ca. −300 cm W.E. If the average of the last 16 years was maintained in the future, the Arikaree Glacier could melt completely in about 25 years, based on the 25-m measured ice thickness.

Ice thickness and duration on lakes in the Green Lakes Valley (GLV) system

Caine (2002) reported a steady decline at about 2.0 cm year^{-1} in the maximum ice thickness on Green Lake 4 during 1982–2000, associated with an increase in flows through the lake in September and October. This trend in ice thickness has not continued since 2000 and may even have reversed (Figure 7); however, the trend of

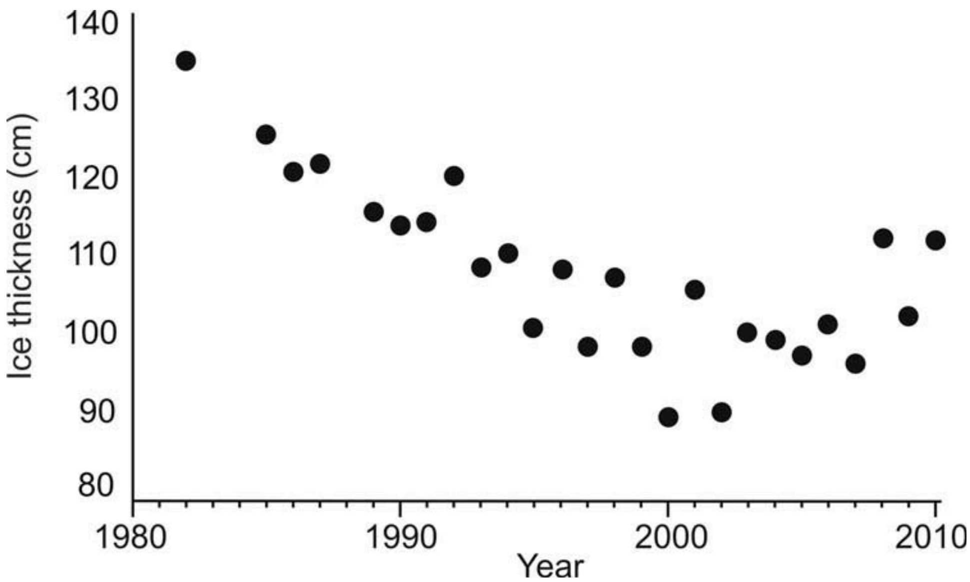

Figure 7. Ice thickness on Green Lake 4 in late March (from Caine 2002, with updated record to 2010). *Reproduction licence received 10-Feb-2015, reprinted from the Annals of Glaciology with the permission of the International Glaciological Society.*

Table 1. Changes in ice cover on the lakes of Green Lakes Valley. All trends are in days year^{-1}.

Site	Elevation (m)	Melt* trend	Formation trend	Duration trend
		(days year^{-1})		
Silver L	3128	−0.52	0.54	−1.06
L. Albion	3346	−0.11	0.26	−0.37
Green Lake 1	3426	−0.36	0.37	−0.73
Green Lake 2	3402	−0.18	0.48	−0.66
Green Lake 3	3454	−0.29	0.34	−0.63
Green Lake 4	3553	−0.28	0.38	−0.66
Green Lake 5	3605	−0.20	0.34	−0.54
Arikaree pond	3800	**	0.23	**

Notes: *Melt trends are consistently earlier and formation trends are consistently later, while a negative duration trend suggests shorter ice cover duration. Trends based on 1982–2013.
**Since 1990, the pond at Arikaree Glacier remained frozen throughout the year except in 2002 and 2012.

increased late-season flows has continued. A discontinuous record of the dates of complete new ice cover and of ice clearance on the eight lakes in the Green Lakes system shows that the duration of ice cover increases with elevation (Table 1). It also suggests that the duration of ice cover on all lakes has declined at rates of up to 1.0 days year^{-1}. With the exception of Lake Albion and Green Lake 2, which are manipulated for water supply during winter and spring, the declining trend in lake ice duration decreases with increased elevation in the valley: from 0.9 days year^{-1} at Green Lake 1 to 0.5 days year^{-1} at the head of the valley (Table 1). In general, the reduced duration of ice cover is due more to a delay in ice formation than to earlier ice melt (Table 1).

Permafrost, subsurface ice and periglacial landforms

Permafrost was thought to exist throughout regions of Colorado and New Mexico at elevations above ca. 3400 m (Marker 1990). Ives (1974) placed the lower level of discontinuous permafrost between 3500 and 3750 m, which corresponds with the −1°C mean annual temperature (MAT) isotherm on NWT. Greenstein (1983) concluded that (1) continuous permafrost extended as low as 3550 m on north-facing slopes and down to 3600 m on south-facing slopes; (2) discontinuous permafrost extended as low as 3200 m on north-facing slopes but only to 3300 m on south-facing slopes; and (3) lower elevations (3470–2930 m) did not contain permafrost. Recent research (Leopold, Völkel, et al. 2013) reported only minor permafrost on NWT. Studies between 3700–3500 m along Trail Ridge Road in Rocky Mountain National Park compared three different permafrost distribution models, but the resulting spatial distributions showed only an 8% overlap (Janke et al. 2012). Temperature loggers at depths down to 7 m did not provide any indication for freezing conditions throughout the year, indicating that these permafrost models might be incorrect (Janke et al. 2012). Permafrost extent and thickness on NWT and in Rocky Mountain National Park may have decreased due to rising temperatures over the last few decades.

Gelifluction lobes, terraces and patterned ground on NWT provide textbook examples of periglacial features (compare Figure 3), and their early mapping and measurement by the late James Benedict and Ives (Benedict 1966, 1970; Ives and Fahey 1971) were significant contributions to characterising North American alpine environments. Benedict's work demonstrated that most of the lobes were inactive at about 1970, but were moving at rates of 1.9 cm year^{-1} and 3.4 cm year^{-1} about 2000 years ago. Even though freeze–thaw processes were active on NWT in the 1970s and1980s, few locations supported active layers capable of maintaining movement of large features, such as transverse lobes and large sorted polygons (Marker 1990). Sorted polygons in ephemeral ponds on the valley floor are frozen seasonally and remain active above annual ice lenses and recent re-measurement has shown that Lobe no. 45 is still moving ca. 1.1 cm year^{-1}, similar to rates measured 40 years ago (Leopold, Völkel, et al. 2013).

Subsurface temperatures on Niwot Ridge

At the NWT sites where data have been collected recently, neither the temperature record nor the geophysical measurements show evidence for permafrost or permanent ice lenses. In the 1970s, subsurface temperature data were collected at several sites on NWT by Benedict (1970), Ives and Fahey (1971) and Ives (1973, 1974) using thermistor strings that were read manually two or three times each month. Data provided by Ives (1974) depicted a permafrost table at 3.5 m depth on a north-facing slope at 3750 m (Figure 8).

In contrast to records collected in the early 1970s, recent subsurface temperature data, including a nearly complete temperature record from the Saddle area in 2008, show that the ground thaws during the summer (Figure 8). Subsurface temperatures have been logged recently at two boreholes on and beside the gelifluction lobe studied by Ives and Fahey (1971) at depths ranging from 0.05 to 2.6 m at the East logger and of 0.05 to 7.0 m at the West, both located near the Saddle site (Figure 1). Thermistors were calibrated using a two-point calibration at 0°C and −22°C. Temperatures were collected every 10 minutes and averaged to hourly and daily values between 2007 and 2009 (compare Leopold, Völkel, et al. 2013).

Records and field notes from the monitoring wells at 3500 m show that ice forms in the late autumn and that ice persists at around 2 m depth (written communication from Hills Buchanan, NWT LTER programme, 2013) in several wells before completely melting in late July or early August each year. Thus, subsurface temperatures do not suggest permafrost conditions at 3500 m at the Saddle location.

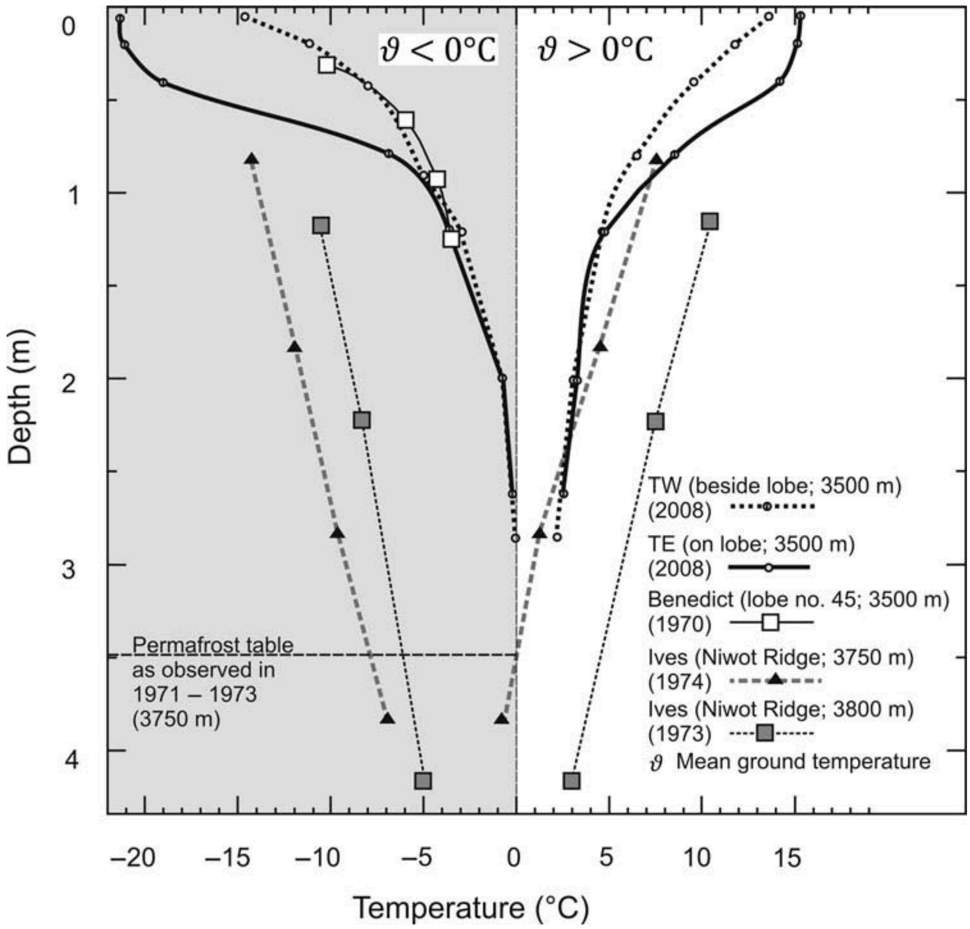

Figure 8. Annual variations of the minimum and maximum subsurface temperatures at various depths from different years and sites at Niwot Ridge. Shaded side symbolises temperatures below 0°C. The curve given by Benedict (1970) and the data measured in 2008 are similar (modified from Leopold, Völkel, et al. 2013). *Reproduction licence obtained 14-Jan-2015, www.schweizerbart.de*

Geophysical measurements using various techniques at different locations on NWT show that the shallow subsurface consists of layered, mainly gravel-rich deposits with composite thickness of 1 to >10 m above bedrock. Measurements using ERT portray the seasonal development and disappearance of an ice-rich layer (Figure 9). With the exception of a small area on a north-facing slope at 3750 m, none of the subsurface imaging suggests that permanent ice lenses persist in the present climate on NWT below this elevation. However, the growth and decay of seasonal ice produces sorted polygons and contributes to the continued downslope movement of gelifluction lobes in shallow ponds and areas that remain wet in autumn above 3500 m.

Leopold et al. (2010, 2011), and Leopold, Völkel, et al. (2013) reported that geophysical measurements below NWT showed that the south-facing slopes of the valley above and below the glacial limit (Figure 3) are underlain by 5–7-m thick, coarse alpine blockfields containing finer sediments near the surface and no ice lenses. The Green Lake 5 rock glacier (see below) is underlain by ice-rich sediments and flanked by a local area of wet permafrost with a ca. 2-m-thick active layer. Other geophysical measurements suggest that permafrost in GLV below 3700 m is only present locally at north- and east-facing positions (Leopold, Voelkel, et al. 2013).

Rock glaciers

Rock glacier investigations in Colorado began in the 1940s, when Ronald Ives described several rock glaciers and other periglacial landforms in the Front Range (Ives 1940). Three kilometres south of the GLV, White (1971) studied the Arapaho rock glacier, which is separated from the Arapaho Glacier by an ice-cored terminal moraine and several lateral troughs. Outcalt and Benedict (1965) had reported a core of glacier ice within the rock glacier beneath the surficial debris cover. White (1971) measured the motion of the rock glacier and its discharge of material from the cirque (ca. 215 m^3 $year^{-1}$), and also calculated its shear stress and viscosity.

Year-round ground surface temperatures over the last 12–15 years near the heads of the three largest rock glaciers on the north-facing wall of GLV are summarised in Table 2. Ground surface temperatures have been monitored by a variety of loggers since September 1996 at

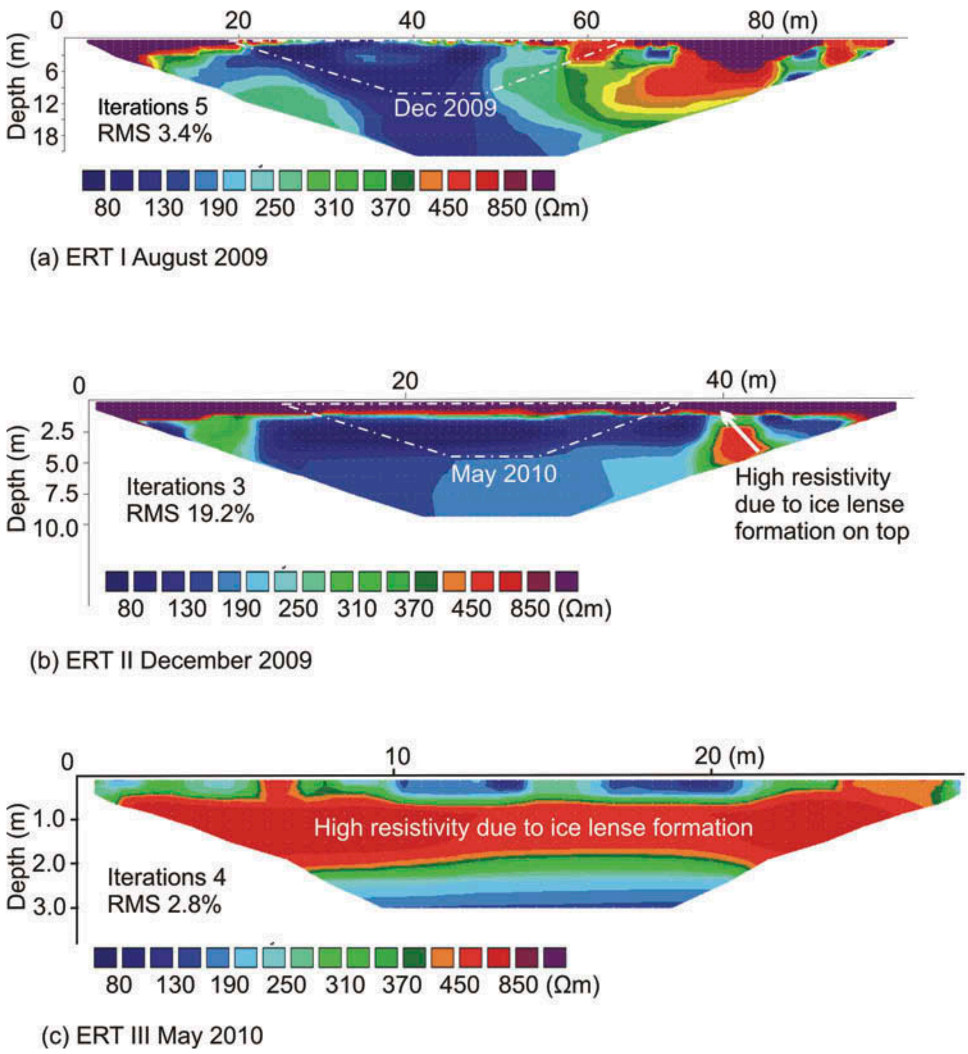

Figure 9. Electric resistivity tomograms (ERT) from the 'Fahey' site on Niwot Ridge (for location see Figure 1). ERT I-III represent the same place but with varying lengths and depths and during different times of the year at 3500 m altitude. Note the low resistivity values during summer conditions and the increased resistivity values on top during the winter freeze. In 9c, there is a layer of higher resistivity values (ice lenses) in between two unfrozen zones (low resistivity values) (modified from Leopold, Völkel, et al. 2013). *Reproduction licence obtained 14-Jan-2015, www.schweizerbart.de*

Table 2. Ground temperatures on three rock glaciers in Green Lakes Valley, 1996–2012. Frost numbers greater than 0.67 indicate continuous permafrost and one of 0.56 probable permafrost.

Site	Statistic	Annual mean temperature (°C)	Annual minimum temperature (°C)	October–May temperature (°C)	Frost number*
Arikaree					
3815 m	Average	−3.46	−9.23	−5.86	0.69
	S.D.	0.45	1.18	0.76	0.04
GL5					
3663 m	Average	−2.65	−20.01	−7.64	0.58
	S.D.	0.28	2.63	0.60	0.02
GL2					
3434 m	Average	−1.47	−7.19	−4.09	0.63
	S.D.	0.79	1.39	0.80	0.02

Note: *The frost number is estimated using the method of Nelson and Outcalt (1987).

intervals that have varied from 2 h (for the early records) to 0.5 h more recently. Most records are continuous except for gaps of 5–7 days during the summer in many years; loss of the logger on rock glacier at Green Lake 2 (RG2) means that the record from there is missing after 2010. None of these records show significant trends in ground temperatures over the period of record. However, measured ground temperatures support the interpretation of rock glaciers as indicators of probable permafrost despite the fact that two of the three examples (those at Green Lake 2 and in the Arikaree cirque) show no signs of motion at the present time and so may not have an ice core (Table 2). Part of the RG5 appears to be active and supports an unstable front slope; surface sediment overlies an internal ice mass (Leopold et al. 2011). The three temperature records vary widely, reflecting local site conditions, especially those affecting snow accumulation. In contrast to the other sites, the surface of the RG5 accumulates little winter snow cover, which controls the timing of soil warming in spring and summer, as well as winter minimum temperatures.

Williams et al. (2006) used the geochemistry of meltwater to speculate about the internal structure of this rock glacier. They hypothesised that the rock glacier had an internal ice core surrounded by interstitial ice intermixed with coarse debris. The 0°C isotherm within the rock glacier was near the surface at the initiation of snow melt. As the surface temperature increased during the summer months (Figure 2), the 0°C isotherm extended lower in the rock glacier. In September and October of 2003, the 0°C isotherm extended into the interstitial ice and some of the interstitial ice melted, contributing to base flow. Records of water temperatures of the past 6 years have shown that the outlet stream at RG5 rarely exceeded 2.5°C and reached a maximum in late August.

This model of the internal structure of the rock glacier suggested by Williams et al. (2006) was evaluated by Leopold et al. (2011). They used three different geophysical techniques – shallow seismic refraction, ground penetrating radar and electrical resistivity tomography – to develop a detailed subsurface model of RG5 (Figure 10). The model was consistent with fine sediments and soils at the surface overlying a zone of coarse debris containing large air-filled voids, which extended to about 2–3 m depth. Between 1–3 to 4–5 m depth, measurements suggested a change in the materials to finer and wetter sediments that were unfrozen in late summer. This zone corresponded to the deepest part of the active layer during the climatic conditions of recent years. Below 4–5 m, GPR signals and ERT images both suggested a fourth zone within the rock glacier body that is best

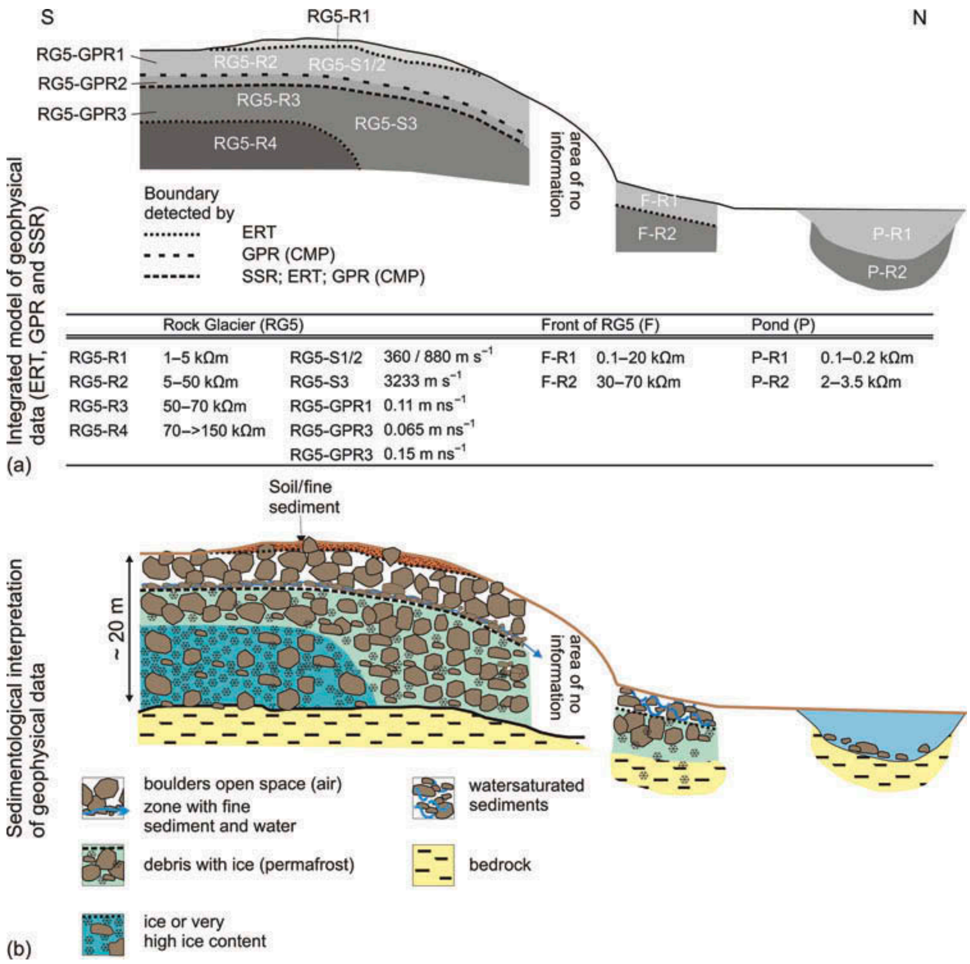

Figure 10. (a) An integrated model of the Green Lake 5 rock glacier (RG5) based on geophysical methods and (b) an interpreted stratigraphic model of the internal structure of the rock glacier. GPR, Ground-penetrating radar; ERT, electrical resistivity tomography; SSR, shallow seismic refraction; CMP, common midpoint (from Leopold et al. 2011, modified). *Reproduction under License Number 3547521009552, 14-Jan-2015, John Wiley and Sons*

interpreted as debris with very high ice content. The data provided only weak indications of a solid ice core in RG5. The authors recognised that their model had less explanatory power with increasing depth, but bedrock could be detected at a depth of about 16–18 m, consistent with its exposure on the adjacent valley floor. Much research remains to be carried out on the role of ice in rock glaciers, particularly as to whether there is a central ice core, dispersed ice throughout the rock glacier, or some combination of these scenarios (Barsch 1996).

Trends in late season hydrology and geochemistry

Late summer and early autumn water flow from the GLV has increased in the past few decades. Flow patterns (Table 3) and the geochemistry of surface waters provide evidence both for seasonal and for long-term changes.

Changes in seasonal and long-term water flow are much more evident in the high alpine sub-catchments compared to the lower elevations (compare Caine 2010). For example, the Middle Boulder Creek station at Nederland (2495 m) shows a weak negative trend (−0.2 ± 0.2 mm year^{-1}), whereas the stations at elevations between 3250 and 3730 m indicate an increase of water flow after about 1982 (Table 3). Increased melt of Arikaree Glacier is not sufficient to explain the overall increasing trend in water flow. Despite a positive trend in late season water flow from 1982 to 2012, the standard error is high and the correlation is not significant at Navajo pond, some 250 m downstream of Arikaree Glacier at 3730 m altitude. Downstream gauges between 3620 m and 3250 m elevation have shown increases in late season water flow (r > 0.61; P < 0.01). An additional water source must contribute to the increased flow rates. The high-elevation increase is best explained by an increased melt of subsurface ice (Caine 2010). Williams et al. (2006) showed that for the outflow of the rock glacier at Green Lake 5, late season water flow coincided with an increase of the hydrochemical signal (Ca^{2+} and SO_4^{2-}). They hypothesised that the increase in Ca^{2+} and SO_4^{2-} was driven by increased melt of internal ice lenses or ice bodies. Further studies will be required to confirm if the hydrochemical signature in alpine water streams can be used as a valid indicator for ongoing subsurface ice melt and as for a basis calculating the rates of ice melt.

Processes responsible for increased late season water flows are still hard to determine, as temperature is not the only factor in permafrost degradation. Despite positive trends in +DD for August as well as for the late season time span of September and October from 1982 to 2012 at the D1 station, the coefficient of determination for these trends is weak (R^2 = 0.093 for August and R^2 = 0.159 for September and October months). A thermal wave produced by the general warming trend (Figure 2) over the whole summer and autumn might be responsible for an increased melt of ice-rich permafrost in GLV. Cumulative lowering of the permafrost table over the past few decades may have increased aquifer capacity, delaying drainage response and producing slower recession of the annual hydrograph. It must be noted that such changes are within 1–2 mm year^{-1} and thus rather low. However, there have been no significant changes in the recession coefficient at GL4 since 1980. Changes in annual snow cover, another reason for changed permafrost melt, were not observed over the past decades, and there is no indication of a reduction in snowmelt peak flows at GL4.

Summary and conclusions

The cryosphere at NWT and the adjacent GLV includes a glacier and seasonal lake ice, as well as various forms of subsurface ice. Periglacial landforms, such as rock glaciers, patterned ground or gelifluction lobes are notable features in these areas, but they may be mainly a legacy of past climate. In the past 10 to 15 years, increased degradation of the cryosphere has been observed, most likely associated with a rising temperature trend during the summer and autumn season (Williams et al. 2006; Caine 2010; Leopold et al. 2010). Ice directly influenced by air temperatures, such as glaciers and lake ice, have shown clear signs of increased melt rates and shorter annual persistence. On average, mass balance of the Arikaree Glacier

Table 3. Trend of late-season flows in the Green Lakes Valley (from Caine 2010, with some updated values). Locations of gauging stations, except for Middle Boulder Creek, are noted in Figure 1.

	September–October trends in discharge				
Basin	Area (km^2)	Elevation (m)	Years	Trend# (mm yr^{-1})	Correlation
Navajo	0.42	3730	1982–2012	2.6 ± 3.0	−0.213 NS
Green L. 5	1.35	3620	1982–2009	2.1 ± 0.5	0.624 **
Green L. 4	2.21	3550	1982–2012	2.6 ± 0.7	0.634 **
Albion	7.10	3250	1981–2012	1.9 ± 0.5	0.613 **
Martinelli	0.08	3410	1982–2012	1.04 + 0.6	−0.323 NS
Middle Boulder Creek at Nederland	93.80	2495	1980–2009	−0.2 ± 0.2	−0.122 NS

Notes: #Trends estimated by least squares regression and shown with a one standard error range.
*P < 0.05; **P < 0.01.

has decreased at a rate of ca. 1 m year^{-1} over the past 15 years. Depending on elevation, the duration of ice cover of seven different lakes between 3128 and 3605 m elevation has decreased between 1.0 and 0.4 days year^{-1} on average over the past 30 years.

Subsurface ice, which is not directly in contact with the atmosphere, also seems to be affected by warmer summer and autumn temperatures. Sites that supported permafrost in the 1970s – even at the edge of permafrost definition – only support seasonal ice lenses at present. Permafrost below an elevation of 3700 m seems to exist only on north- and east-facing slopes. Increased late season water flow in high elevation GLV is indicative of an additional water source which is best explained by increased melt rates of subsurface ice.

We conclude that the cryosphere at NWT and GLV is shrinking on a broad scale, but different parts of the cryosphere shrink at different rates. If ongoing trends continue, Arikaree Glacier will have disappeared in two decades, permafrost areas will have further decreased and partly disappeared. Late season water flow will probably decrease towards its former level. However, valuable hydrological buffers during warm drought years such as glacier, rock glacier or subsurface ice will be much reduced under current trends. Flora and fauna adapted to ice-rich areas will be even more affected if the cryosphere further shrinks. Thus, the high-elevation area of NWT and the adjacent GLV fit into the global trend of decreasing cryosphere. These trends are unlikely to change under current climate conditions.

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