JOURNAL OF QUATERNARY SCIENCE (2009)

Published online in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/jqs.1252

# Inactive and relict rock glaciers of the Deboullie Lakes Ecological Reserve, northern Maine, USA

**AARON E. PUTNAM[1]* and DAVID E. PUTNAM[2]**
[1] Department of Earth Sciences and Climate Change Institute, University of Maine, Orono, Maine, USA
[2] School of Science and Mathematics, University of Maine at Presque Isle, Presque Isle, Maine, USA

Putnam, A. E. and Putnam, D. E. Inactive and relict rock glaciers of the Deboullie Lakes Ecological Reserve, northern Maine, USA. *J. Quaternary Sci.*, (2009). ISSN 0267-8179.

Received 28 May 2008; Revised 20 November 2008; Accepted 23 November 2008

ABSTRACT: Lobate talus slopes in the Deboullie Lakes Ecological Reserve (DLER) of northern Maine exhibit parabolic profiles characteristic of inactive and relict talus-derived rock glaciers. Vegetated rock glacier surfaces suggest that the landforms are no longer active, and lobes comprising two DLER rock glaciers document periods of past growth. Observations of perennial subsurface ice are supported by datalogger temperature measurements, indicating that sporadic permafrost exists throughout the DLER. We compare the DLER rock glaciers, along with similar features elsewhere in New England and adjacent Québec, to the modern alpine permafrost distribution. Results indicate that a mean annual temperature cooling of $\sim 6^{\circ}C$ is required to promote active rock glacier growth. Ages of plant remains recovered from the basal sediments of a local pond constrain deglaciation to before 11 320 $^{14}C$ a BP, and core stratigraphy and organic content reveal that a periglacial environment persisted during the early postglacial era. Thus, we hypothesise that the DLER rock glaciers were active during Lateglacial time despite the lack of glacier activity in the region. We take this to suggest that north-eastern US rock glaciers formed in response to mean annual temperatures skewed towards the frigid winters of the Younger Dryas chronozone. Copyright © 2009 John Wiley & Sons, Ltd.



KEYWORDS: rock glacier; talus; permafrost; Lateglacial; New England.

## Introduction

Relict permafrost-related landforms, particularly permafrost-derived rock glaciers, have become important tools for deciphering the climatic manifestation of the abrupt climate changes of the last deglaciation (Kerschner, 1978; Sailer and Kerschner, 1999; Paasche *et al.*, 2007). At present, active rock glaciers are typically associated with rugged glacial and periglacial landscapes where the climate maintains air temperatures sufficiently low to sustain continuous permafrost (Wahrhaftig and Cox, 1959; Barsch, 1996; Humlum, 1998). Most active rock glaciers occur at sites where mean annual air temperature (MAAT) is less than $-4^{\circ}C$ (Humlum, 1998; Paasche *et al.*, 2007). Therefore, inactive and relict (or 'fossil') rock glaciers can be used to reconstruct previous episodes when the MAAT was favourable for active rock glacier growth. Rock-glacier-inferred MAAT can also complement summertime temperature proxy records, such as those based on glacier behaviour and palaeo-snowlines, to gauge the seasonality associated with past climate changes (Kerschner, 1978; Denton *et al.*, 2005).

The north-eastern USA and adjacent Canada form the western boundary of the North Atlantic, and is an important region for tracking the spatial influence of Lateglacial abrupt climate events that are thought to have originated from rapid changes in the North Atlantic thermohaline circulation (e.g. Broecker, 1998; Alley, 2000). Although much research has focused on the Lateglacial history of relict rock glaciers in Europe, there is a paucity of recent information from the north-eastern USA. In this study, we document previously unrecognised relict and inactive permafrost-derived rock glacier landforms in the remote Deboullie Lakes Ecological Reserve (DLER) of northern Maine, north-eastern USA (Fig. 1). Several talus slopes located in the DLER have the characteristic lobate topography of landforms that are variably defined as 'protalus lobes' (White, 1976; Washburn, 1980; Harrison *et al.*, 2008), 'lobate rock glaciers,' (Wahrhaftig and Cox, 1959), 'talus glaciers' (Smith, 1973), 'talus-derived rock glaciers' (e.g. Kirkbride and Brazier, 1995; Brazier *et al.*, 1998; Ikeda and Matsuoka, 2002) and 'valley-wall rock glaciers' (e.g. Millar and Westfall, 2008). Over the course of four field expeditions conducted between 2004 and 2007, we carried out topographic surveys, temperature datalogger measurements and geomorphic mapping to describe these landforms and determine possible linkages to present and past climate. In addition, a new radiocarbon-dated lacustrine record of sedimentation and organic content affords insight into the timing of deglaciation, the age of rock glacier initiation, and the potential environmental conditions that may have accompanied rock glacier formation. The rock glaciers of the north-eastern USA and adjacent Canada offer important regional palaeoclimatic

* Correspondence to: A. E. Putnam, Department of Earth Sciences and Climate Change Institute, University of Maine, 228 Bryand Global Sciences Center, Orono, ME 04468, USA.
E-mail: aaron.putnam@umit.maine.edu

information, and together with European studies may afford a pan-Atlantic perspective on the seasonality of the abrupt climate changes that punctuated the last deglaciation.

## Previous work

Nearly 50 years of discussions regarding rock glacier classification based on origin and genesis, distribution and climatic sensitivity have refined the general understanding of the geomorphic and climatic significance of rock glaciers (e.g. Wahrhaftig and Cox, 1959; Potter, 1972; Johnson, 1983; Hamilton and Whalley, 1995; Kirkbride and Brazier, 1995; Haeberli and Vonder Mühll, 1996; Brazier *et al.*, 1998; Clark *et al.*, 1998; Humlum, 1998; Steig *et al.*, 1998; Konrad *et al.*, 1999; Mitchell and Taylor, 2001; Janke and Frauenfelder, 2008; Millar and Westfall, 2008). As knowledge of periglacial landform–climate relationships progresses, rock glaciers are gaining increasing attention for their utility as recorders of past environmental changes (e.g. Kerschner, 1978; Konrad *et al.*, 1999; Humlum, 1998; Aoyama, 2005; Paasche *et al.*, 2007; Millar and Westfall, 2008). Some studies have used rock glaciers as a tool for reconstructing regional climate histories. For example, Birkeland (1982), Kirkbride and Brazier (1995), Ackert (1998), Aoyama (2005), and Paasche (2007), among others, showed that rock glaciers are (1) sensitive to climate changes and (2) useful for deriving palaeoclimatic information.

Rock glacier descriptions in New England are few, and limited to relatively high-elevation regions. Goldthwait (1940, 1970) mapped a prominent 'block glacier' occupying the valley floor of King Ravine in the White Mountains of New Hampshire. Also in the White Mountains, Thompson (1999) suggested a rock glacier origin for a tongue-like talus slope occurring along the north wall of Tuckerman's Ravine. A prominent tongue-shaped rock glacier occupies the floor of North Basin, Mt Katahdin, Maine (Thompson, 1961; Davis, 1989), but this feature has not yet been the subject of close investigation.

In addition, alpine permafrost and active permafrost-related landforms occur on the summits of Mt Washington in the White Mountains of New Hampshire, Mt Katahdin in Maine, and possibly other high alpine peaks in New England (Goldthwait, 1940; Thompson, 1962; Péwé, 1983; Walegur and Nelson, 2003; Nelson *et al.*, 2007). A number of investigators also reported rock glaciers and permafrost throughout the relatively low-elevation Chic Choc Mountains and coastal areas of the Gaspé Peninsula, Québec (Gray and Brown, 1979; Richard *et al.*, 1997; Hétu and Gray, 2000a,b; Hétu *et al.*, 2003). For example, Hétu and Gray (2000b) identified at least eight relict talus-derived rock glaciers occurring in valleys along the north coast of the Gaspé Peninsula.

## Study area geology and ecology

The DLER is located in northern Maine between the Allagash and Fish River drainages (46° 57′ N, 68° 53′ W). Small, rugged mountains composed of quartzite and hornfels encircling the Deboullie syenite and granodiorite attain a maximum elevation of 599 m a.s.l. (Boone, 1958). Incipient cirques, paternoster lakes, U-shaped valleys with precipitous cliffs and numerous large talus slopes give the area its name in the local Acadian French patois: *d'eboulis* – to tumble down.

The DLER supports a rare ecological assemblage that is typically associated with cold environments. The persistence of an isolated population of landlocked arctic char (*Salvelinus alpinus oquassa*), locally known as blueback trout, is attributed to the abnormally cold lake water, where summer deep-water temperatures range between 6 and 8°C (Bernatchez *et al.*, 2002; Wilkerson, 2007). Other rare boreal animal species found in the DLER include the northern bog lemming (*Synaptomys borealis*) and a freshwater amphipod (*Gammarus lacustris*). Several rare species of boreal plants occur in the DLER as well. For instance, the DLER is home to the only known Maine population of *Minuartia rubella* (Arctic sandwort; Wilkerson, 2007).

At least five talus slopes in this region exhibit lobate rock glacier characteristics as described by Wahrhaftig and Cox (1959), White (1976) and Gordon and Ballantyne (2006), among others. We refer to these landforms as: 'Gardner Pond Rock Glacier' (GPRG), 'Crater Pond Rock Glacier' (CPRG), 'Deboullie Pond Rock Glacier' (DPRG), 'Galilee Pond Rock Glacier' (GalRG) and 'The Pushineer Pond Rock Glacier (PPRG). Figure 1 gives landform locations. This study focuses primarily on the three most prominent rock glaciers: GPRG, CPRG and DPRG.

## Modern climate

Regional climate data recorded from the nearby Clayton Lake and Allagash weather stations between 1971 and 2000 give MAAT of 2.3°C and 2.2°C, respectively; mean summer temperatures of 15.8°C and 15.9°C, respectively; and mean winter temperatures of −13.2°C and −12.5°C, respectively (National Climatic Data Center, 2002). Mean annual precipitation (MAP) recorded between 1971 and 2000 is 903 mm $a^{-1}$ water equivalent (w.e.) and 912 mm $a^{-1}$ w.e. at the Allagash and Clayton Lake weather stations, respectively (National Climatic Data Center, 2002). The Allagash and Clayton Lake climate normals provide a close estimate for the climate of the Deboullie Lakes, where no permanent weather stations exist. These data reveal that central northern Maine experiences some of the coldest winter conditions in New England, and is affected by seasonality characteristic of a continental setting.

## Methods

### Rock glacier classification

We used a field-based geomorphology approach to identify and describe several talus slopes with parabolic profiles in northern Maine, which we classify here as rock glaciers. The nomenclature of Wahrhaftig and Cox (1959), as refined by more recent studies (e.g. Millar and Westfall, 2008), is used to classify rock glacier activity and categorise these talus-derived rock glaciers (hereafter referred to as 'rock glaciers') as 'active' 'inactive' or 'relict' (see Table 1 for definitions). This classification is based on topographic data, observed distributions of persistent subsurface ice, as well as vegetation and lichen cover (after Wahrhaftig and Cox, 1959; Ikeda and Matsuoka, 2002; Cannone and Gerdol, 2003).

### Long profile survey

We used a traditional optical transit and metric stadia rod to measure longitudinal profiles of the DPRG, CPRG and GPRG, and to classify rock glacier activity. Transects were from the top of each slope to the distal lobe where the rock glaciers intersect



DOI: 10.1002/jqs

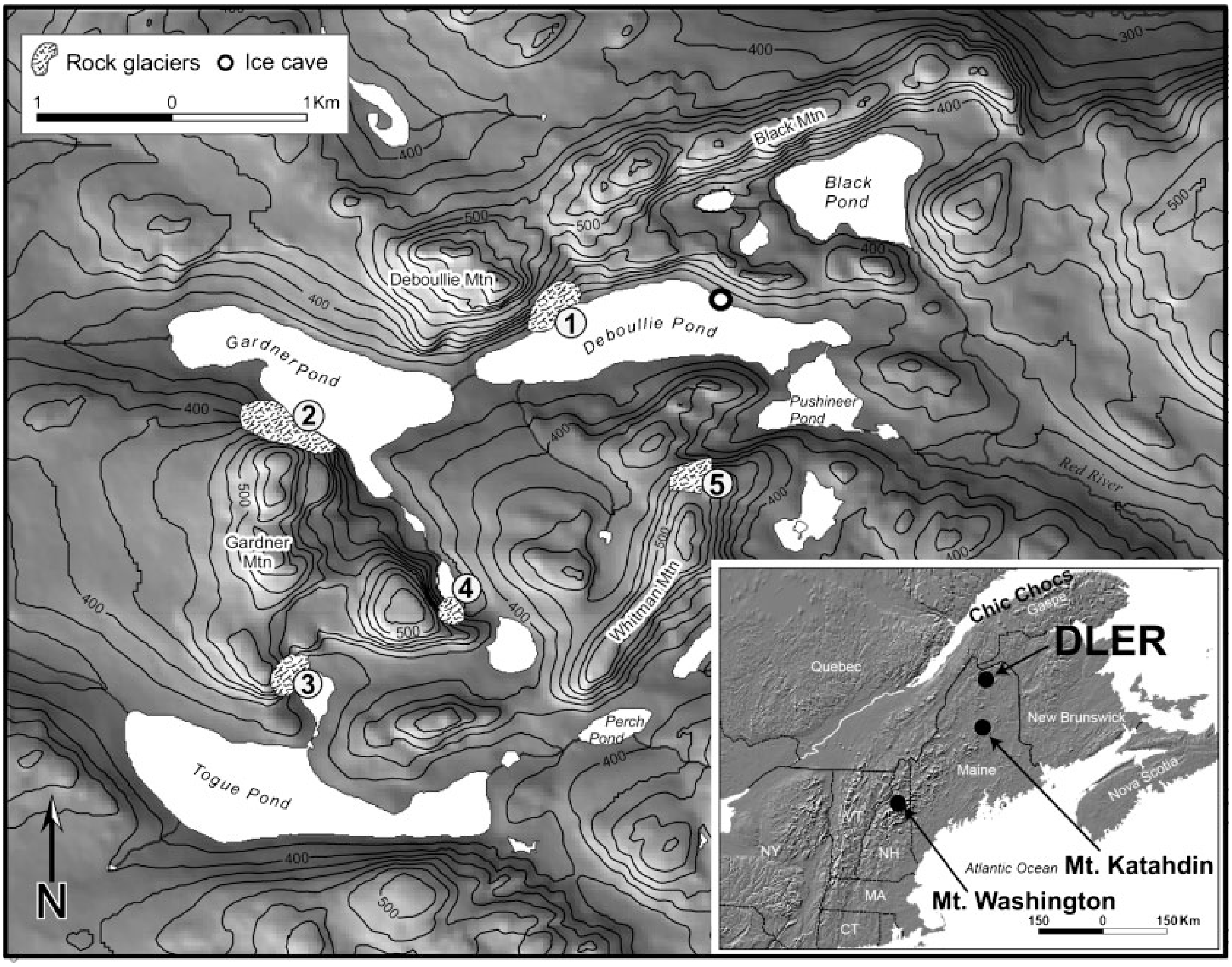


**Figure 1** Map of the DLER study area. Stippled pattern indicates rock glaciers. Open dot shows ice cave. Circled numbers give the following rock glacier locations: (1) DPRG; (2) GPRG; (3) CPRG; (4) GalRG; (5) PPRG. Inset shows regional map with locations mentioned in the text

a lake shoreline. Subsurface temperature was measured at 1 m intervals along each profile. During the 2005 summer field season we developed a high-resolution contour map of the GPRG surface using a total-station EDM to aid in geomorphic interpretation.

## Temperature dataloggers

To collect continuous temperature records, we placed three HOBO™ temperature dataloggers at different elevations along the longitudinal profile on the GPRG that recorded temperatures 1 m above and 1 m below the rock glacier surface. The subaerial probes were fastened to the top of wooden tripods, and the lower probes were suspended into the interstices between the large boulders of the GPRG surface. We left the subsurface temperature sensors uncovered by soil or rock so that we could gain an accurate assessment of the depth-to-surface air temperature gradient. We programmed the instruments to collect data at 0.5 h intervals for the summer/fall of 2004, and reduced the resolution to record once every hour over the fall/winter/spring of 2005/2006.

**Table 1** Rock glacier activity classification (adapted from Wahrhaftig and Cox, 1959; Cannone and Gerdol, 2003; Ikeda and Matsuoka, 2002; Millar and Westfall, 2008)

| Rock glacier activity | Description |
|---|---|
| Active | Embedded ice, active landform movement, oversteepened frontal slope, active clast sorting, no vegetation or lichens. Often associated with glaciers and persistent snowfields |
| Inactive | Embedded ice, steep frontal slope, no landform movement, vegetation and lichen cover |
| Relict (fossil) | No embedded ice, no landform movement, 'deflated' appearance, vegetation cover |

## Lake coring

Galilee Pond is the second highest pond in a paternoster chain located in the central DLER, at an elevation of 348 m a.s.l. (Fig. 1). Galilee Pond has an area of 0.03 km$^2$, a maximum depth of about 9 m and occurs adjacent to a small rock glacier located beneath the steep, east-facing valley wall at its upper, southern end (GalRG; see Fig. 1). We used a modified Livingstone piston corer to extract a 7.92 m core from the sediments 8.4 m beneath the surface of Galilee Pond, for the purpose of determining a minimum age for deglaciation and to provide a general constraint on the age of rock glacier formation. At the University of Maine palaeoecology laboratory, core-bottom sediments were sieved for organic macrofossils for radiocarbon dating, and loss-on-ignition (LOI) analyses were performed using the standard methods of Bengtsson and Enell



DOI: 10.1002/jqs

(1986) and Heiri *et al.* (2001). We collected 2 $cm^3$ samples at 4 cm intervals in the lower 432 cm and at 2 cm intervals in the upper portion of the core. Samples were dried overnight by oven, and combusted at 550°C. Percentage LOI was determined from the ratio of the mass lost during combustion to the dry mass of the sample.

## Rock glacier morphology: results and interpretations

### Gardner Pond Rock Glacier

The GPRG is located beneath the north-east-facing cliff of Gardner Mountain, and has a parabolic slope that is comprised of large, tabular boulders between 0.5 and 2.0 m in length. At least three longitudinal furrows and three transverse lobes (see Figs. 2 and 3) characterise the slope surface. The largest lobe terminates at a steep slope ~11 m in height, is soil covered, and supports a thick 'spruce talus woodland', including *Picea mariana* (black spruce) and *Betula papyrifera* (paper birch) trees (Wilkerson, 2007). Pockets of ice were observed in boulder crevices during the late summer of each field season, and the July subsurface temperature profile, shown in Fig. 2, reflects these observations. Mature vegetation indicates that the lobe is not actively advancing (Cannone and Gerdol, 2003), but the steep topography suggests that the rock glacier front is not significantly deflated. Post-depositional slumping has removed parts of the steep frontal slope. The resultant sections reveal that large, stacked angular boulders comprise the exposed portions of the rock glacier interior. Given the thick forest cover, parabolic topography, steep frontal slope and persistent subsurface ice, we classify this lobe as 'inactive'.

A smaller concave lobe occurs above the larger, lower lobe (Fig. 3). The upper lobe terminates at a near-vertical slope ~3 m high. The slope is comprised of large boulders, and is covered by a 'Labrador tea talus dwarf shrubland' community consisting of a ~30 cm thick 'cryophilic' floral mat containing numerous species of lichen (Wilkerson, 2007). Small, stunted *Picea mariana* (black spruce) is sparsely distributed on the upper lobe. We also observed persistent summer ice in the upper-lobe subsurface. For the same reasons given above, we consider this lobe inactive.

Finally, a forested terrace of talus boulders projects from the base of the GPRG lower lobe and terminates at the Gardner Pond shore where, underwater, the bouldery slope dips steeply toward the lake bottom. At its northern end, the GPRG overlies a diamicton interpreted to be a lodgement till that occurs as a thin drift throughout the DLER. The till is composed of rounded

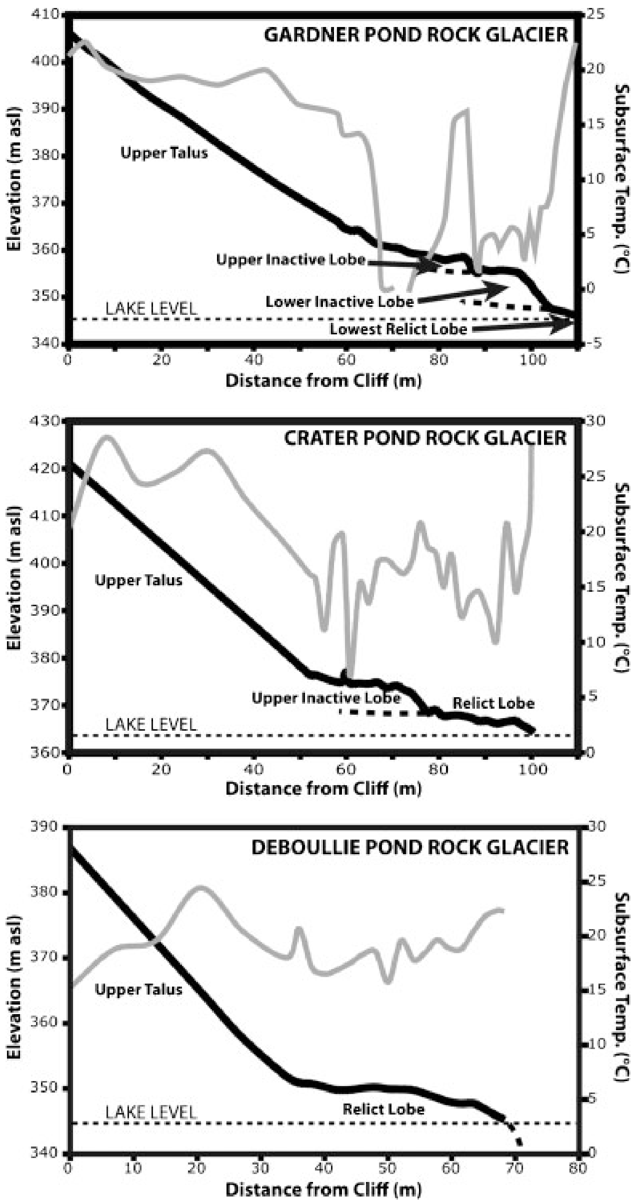


Figure 2 Longitudinal slope profiles of the GPRG (top panel), CPRG (middle panel) and DPRG (bottom panel). Black solid line represents slope surface, grey line gives July temperature profile along slope. Thick dotted black lines show inferred contacts between lobes, and thin, horizontal dotted lines indicate modern lake level

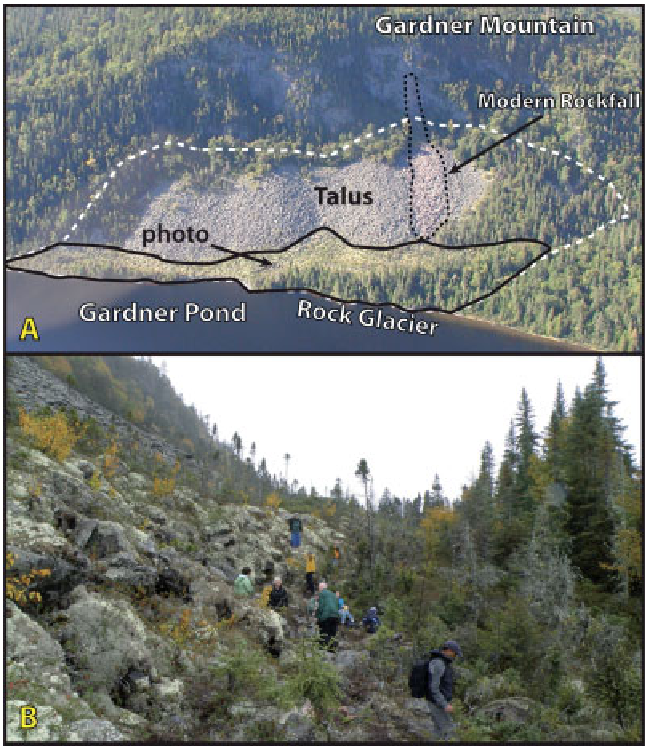


Figure 3 (A) Oblique air photograph taken south towards the GPRG. Dark line indicates rock glacier portion of slope; white dashed line outlines talus slope; black dashed line shows modern rockfall scar (cliff) and deposit (talus). Arrow indicates direction of photograph in panel B. (B) Photograph taken looking west along the frontal slope of the GPRG upper lobe. Note the steep slope angle and the thick tundra-like vegetation cover. Members of the UMaine Climate Change Institute for scale. This figure is available in colour online at www.interscience.wiley.com/journal/jqs

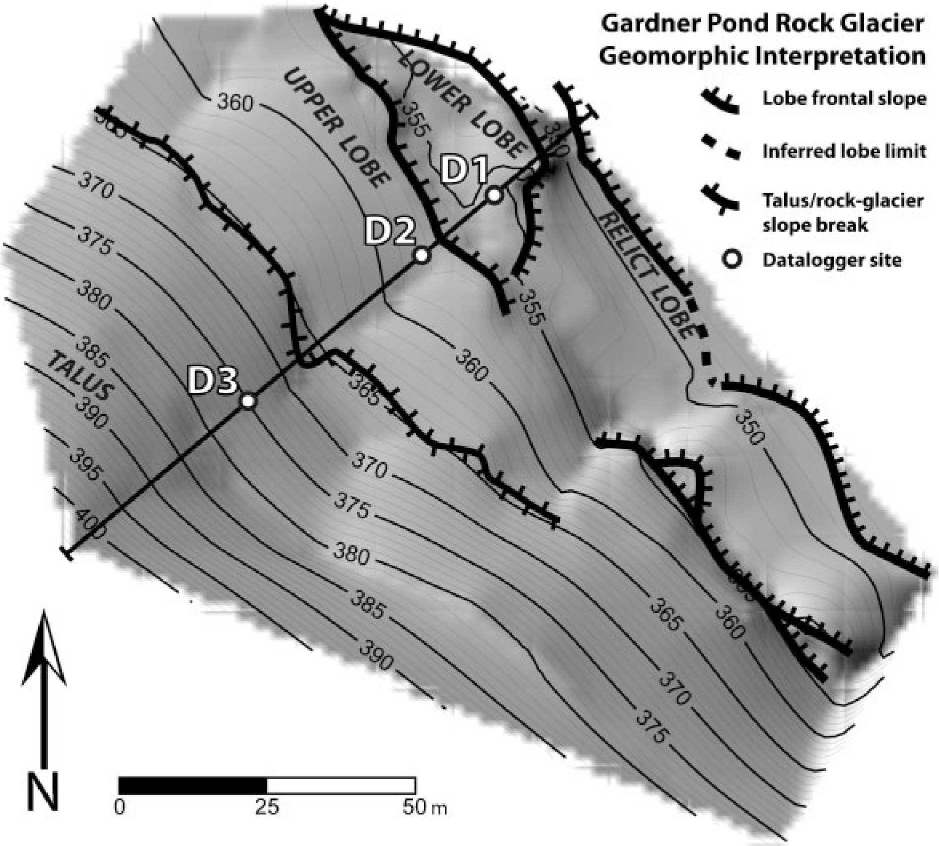


**Figure 4** Geomorphic map of GPRG. Light contours are spaced at 1 m intervals, dark contours at 5 m intervals. Inset shows key to map symbols

to subrounded cobbles and boulders of allocthonous lithologies. Similar outcrops of small, rounded, unsorted clasts outcrop at one location near the modern shoreline, indicating that the terminus of the rock glacier is roughly associated with the lakeshore. We interpret the lower rock glacier lobe as the oldest relict feature of the GPRG (Fig. 4).

## Crater Pond Rock Glacier

The east-facing CPRG (Fig. 2) has a parabolic form similar to that of GPRG. The most extensive lobe is a low-profile concave forested apron of talus-derived boulders lacking a steep frontal slope. Vegetation is classified as 'spruce talus woodland' (Wilkerson, 2007). As with the GPRG, this lower lobe overlies a diamicton comprised of allocthonous cobbles. No ice was observed during the 2004–2006 field seasons in this portion of CPRG. Therefore, we suggest that this deflated landform is a relict rock glacier lobe.

The upper CPRG lobe occurs upslope of the lower lobe, and terminates in a steep front 5 m in height. We observed persistent summer ice in the subsurface of the upper CPRG lobe, which is complemented by the July 2004 subsurface temperature profile (Fig. 2). A recent slump on the frontal slope has exposed the internal structure of a small portion of the rock glacier. As with the GPRG, in this section the CPRG appears to be entirely clast supported, comprised almost exclusively of large, angular boulders. The upper lobe is in part forested, particularly on the steep frontal slope as well as on the upper talus. 'Labrador tea talus shrubland' covers the broad top surface of the lobe (Wilkerson, 2007). We classify the upper CPRG lobe as inactive on the basis of steep, lobate topography, the persistence of subsurface ice and vegetated surfaces.

## Deboullie Pond Rock Glacier

The DPRG (Fig. 2) is comprised of large, unsorted tabular clasts about 1–2 m in diameter. The slope faces due south, and consists of one parabolic lobe that contains a more subdued frontal bulge with no obvious transverse ridges or furrows. The DPRG surface is mostly bare, with vegetation occurring only in a few small depressions. No visible ice could be found within the boulder interstices, and the July subsurface temperature profile does not show abnormally cool conditions (Fig. 2). We interpret the DPRG to be a 'relict' rock glacier landform.

A man-made hole, known historically as the 'ice cave', occurs in till at the foot of a forested lobate talus slope east of the DPRG. The ground subsurface at this site generally remains frozen throughout the summer, indicating that despite the relict form of the DPRG perennial ice does occur sporadically in the south-facing slopes of the region.

## Interpretations of relative age and genesis

The three described landforms exhibit distinct parabolic profiles that are classically associated with 'talus-derived' or 'valley-wall' rock glaciers, and these features consist of both relict and inactive lobes. We observed no distinct transverse ridges with well-defined hillside troughs, or clast sorting to indicate that these features formed as protalus (or pronival) ramparts (Gordon and Ballantyne, 2006). The inactive transverse lobes retain steep fronts indicating that ice may still exist beneath the surface. However, established vegetation communities suggest no recent rock glacier activity (Cannone and Gerdol, 2003). In contrast, relict features are deflated, do



DOI: 10.1002/jqs

not contain observable ice and are forested on the distal surfaces of CPRG and GPRG. The DPRG is a conspicuous feature of the north shore of Deboullie Pond because it faces due south, is not vegetated and consists of one deflated, yet extensive relict lobe.

The deflated form of the DPRG may be due to its southerly aspect, which exposes the landform to more direct solar radiation than GPRG or the CPRG, the latter two being situated beneath steep, north-east- and east-facing cliffs, respectively. The lack of vegetation may indicate relatively recent settling compared to the relict lobes of CPRG and GPRG. It is worth noting, however, that ice exists among the forested talus slopes that occur adjacent to the DPRG, and therefore may point to another reason for the deflated form of the rock glacier. Middle to late Holocene lake-level rise of northern Maine ponds has been by documented by Dieffenbacher-Krall and Nurse (2006). In addition, young littoral features around Deboullie Pond indicate elevated lake level associated with a historical dam at the Pushineer Pond outlet (Wilkerson, 2007). Lake level changes, either climate-driven or anthropogenic, may thus explain the relict form of the DPRG, and perhaps the lower relict lobes of GPRG and CRPRG. With the exception of the DPRG, the other rock glaciers exhibit evidence for multiple advances based on the presence of multiple overlapping lobes.

Rock glacier advances occur when climate conditions are favourable for permafrost growth and creep (Haeberli and Vonder Mühll, 1996), as well as talus production resulting from freeze–thaw cycles in the overhead cliffs. Cooler climate and stronger seasonality enhance both of these processes. Vegetated cliff faces of the DLER indicate relative stability, with only one apparently recent rock avalanche scar and deposit on the north end of the upper GPRG talus (Fig. 3). The fresh appearance of the recent rockfall deposit is a marked contrast to the nearby weathered lichen-covered talus. Distinct characteristics of ancient talus, such as weathering and lichen cover, indicate a low frequency of recent talus accumulation. Infrequent rockfall, taken together with the vegetated inactive and relict lobes of the various DLER rock glaciers, implies that the lobes represent earlier periods of climate conditions more favourable to rock glacier development, including both temperature and debris supply. Due to the high shear stress required to deform coarse rock–ice mixtures, thick ice lenses are necessary to facilitate downslope creep of talus-derived rock glaciers (Whalley and Azizi, 2003; Harrison *et al.*, 2008). Present stability of the talus slopes implies that ice lenses are not sufficiently thick or actively accreting in the upper talus to facilitate downslope deformation. Therefore, frontal bulges of the inactive lobes of GPRG and CPRG indicate previous periods of downslope extension. The deflated, concave profile of the DPRG exhibits the result of the loss of internal ice. Overall, the DLER rock glaciers appear to be inactive or relict landforms formed during multiple episodes of prior growth.

## Temperature dataloggers: results and interpretations

The temperature datalogger results for summer 2004 and winter 2005–06 are shown in Figure 5. Field observations suggest that as summer progresses the visible ice, frozen among the surface interstices, gradually recedes into the depths of the rock glacier. Given that the subsurface temperature signatures for D1 and D2 show little resemblance to their respective above-surface counterparts (as well as the D3 signal), and display relatively cool or near-freezing temperatures throughout the summer, we assume that the subsurface temperature reflects the relative proximity of the sensor to the surface of the rock glacier ice.

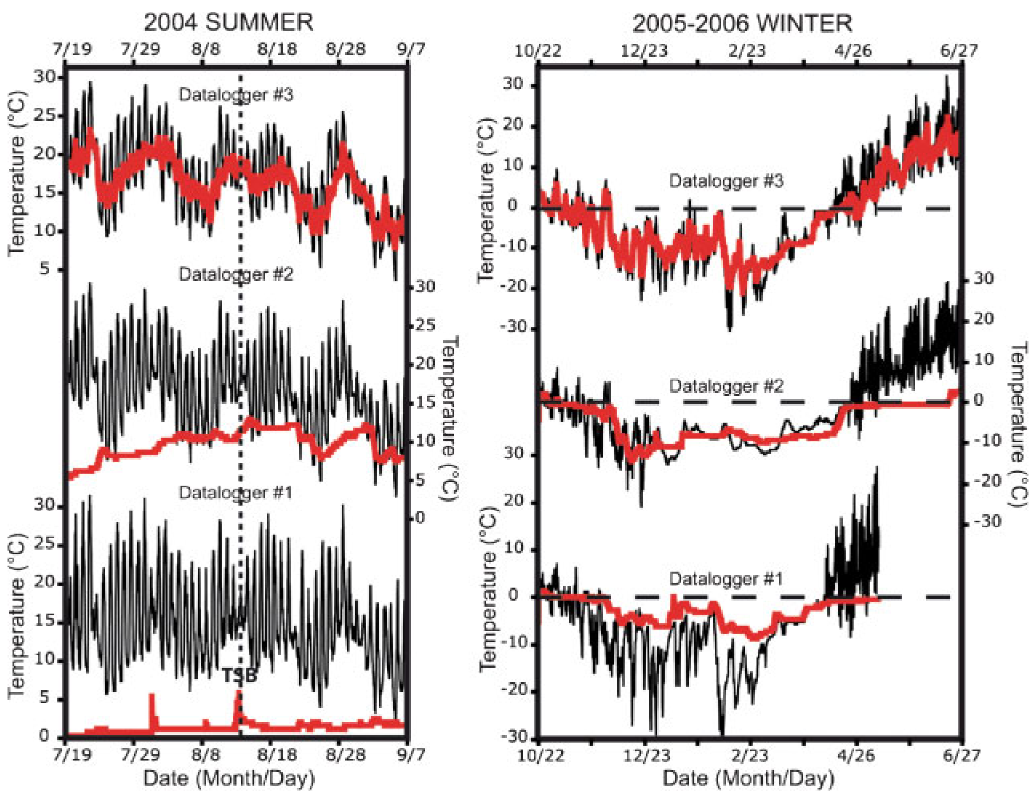


**Figure 5** GPRG datalogger results for summer 2004 (left panel) and winter 2005–2006 (right panel). Thin, solid black lines show GPRG above-surface temperature; solid, lighter lines indicate subsurface temperatures. Vertical dashed line marks the date when the remnants of tropical storm Bonnie (TSB) reached northern Maine. This figure is available in colour online at www.interscience.wiley.com/journal/jqs



DOI: 10.1002/jqs

Temperatures registered in the subsurface at sites D1 and D2 display a gradual rise throughout the summer of 2004, as well as in the late spring of 2006. In contrast, subsurface temperatures measured at D3 co-vary with above-surface temperatures throughout the year. We note that no ice has been observed in the interstices at the location of D3 in the upper talus. These results show that in portions of the slope that contained ice subsurface temperatures rose slowly throughout the duration of the summer season, despite considerable variability in air temperature 1 m above the surface. During the late summer of 2004, the rising trend in rock glacier subsurface temperature actually opposes the declining trend of above-surface temperature. From this we infer that warmth registered 1 m above the rock glacier is not apparent immediately below the surface. The extreme summer temperature–depth gradient is perhaps a result of the inefficient downward conduction of surface heat through the high-albedo granite boulders forming the surface of the rock glacier, and through the spongy vegetation mat insulating the surface. This may also indicate the presence of deeper ice that may facilitate the persistence of ice so close to the rock glacier surface during the summer. Thus, we interpret temperature data recorded by D1 and D2 to indicate the presence of permafrost in the GPRG.

In contrast to the summer months, the colder above-surface temperature was matched by general cooling in the rock glacier subsurface at all datalogger locations during the winter. Thus, it appears that the GPRG subsurface is directly sensitive to cooling episodes, but seems to respond more slowly to warmth. Wintertime subsurface air sensitivity may be associated with the 'Balch' or 'chimney' effect, where winter convection in the upper talus draws cold air deep into the talus interstices, and supercools the base of the deposit (Delaloye and Lambiel, 2005). Thompson (1962) suggested that Balch ventilation may play an important role in the growth of subsurface ice in the King Ravine rock glacier of the White Mountains, and this may also be true for the DLER rock glaciers. Delaloye and Lambiel (2005) suggested that Balch ventilation might preserve relict permafrost bodies in the lower portions of relict and inactive rock glacier landforms at or below the altitudinal limit of the discontinuous permafrost zone. We suspect this may be the mechanism allowing permafrost to persist in the CPRG and CPRG.

Finally, precipitation patterns appear to play a secondary role in elevating summer subsurface temperatures. The subsurface sensor of D1 registers a series of conspicuous peaks during the summer of 2004 that result in slightly elevated subsurface temperatures. These peaks correlate with thunderstorm events that occurred in northern Maine, as well as precipitation from the remnants of tropical storm Bonnie. The August 2004 'unusual weather phenomena storm data' (World Meterological Society, 2004) reported heavy rain, flooding and mudslides as tropical storm Bonnie swept over northern Maine between 13 and 14 August. This event was registered as an abrupt warming in the subsurface temperature of D1, perhaps related to rain percolating into the talus and rock glacier interstices, and over the temperature sensor. However, D1 subsurface temperature returned to near-zero values following each event. Though rainfall must play a role in the ablation of subsurface ice at D1, it is less noticeable against the overall warming trend registered in D2.

## Palaeotemperature reconstruction

To estimate the MAAT favourable to initiate rock glacier expansion in the DLER, we provide a simple conceptual model that uses the regional modern permafrost elevation to infer the climatic cooling necessary to initiate active rock glacier growth (after Walegur and Nelson, 2003; Nelson *et al.*, 2007). The modern mountain permafrost–elevation gradient is constructed based on the known permafrost localities of Mt Washington (e.g. Thompson, 1962; Péwé, 1983; Spear, 1989; Clark and Schmidlin, 1992) and the Gaspé Peninsula (Hétu and Gray, 2000b), as well as the geological and ecological indicators of active permafrost presence on the summit of Mt Katahdin (Péwé, 1983; Walegur and Nelson, 2003; Fig. 6). The theoretical elevation of continuous permafrost at the latitude of the DLER was interpolated from the modern permafrost–elevation gradient, and from this we determined an elevation difference of ~900 m between the modern permafrost elevation and the toe of the GPRG. Assuming an adiabatic lapse rate of 6.5°C km$^{-1}$, we estimate that the DLER rock glaciers were

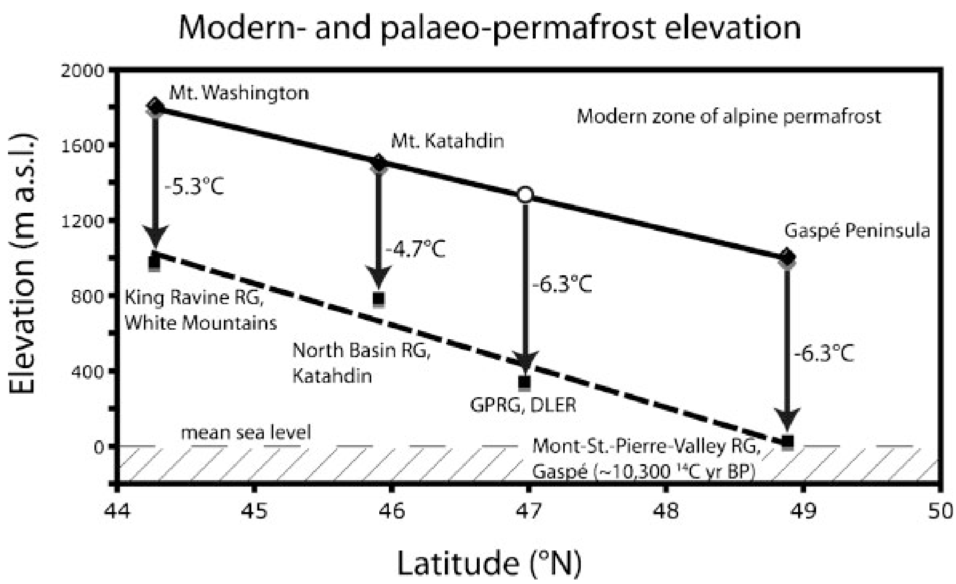


**Figure 6** Diagram shows modern permafrost elevation trend with latitude as solid line, and reconstructed palaeo-permafrost gradient as dark dashed line. Filled diamonds represent regions of modern alpine permafrost occurrence, filled squares indicate rock glacier occurrence, open circle represents the point interpolated from the modern permafrost trendline above DLER and arrows show calculated MAAT changes inferred from modern vs. palaeo-permafrost elevation difference



DOI: 10.1002/jqs

formed at a time when MAAT was ~6.3°C cooler than present, yielding a palaeo-MAAT of about −4°C. Following the same procedure, the palaeo-MAAT difference was calculated using the King Ravine rock glacier in the White Mountains, New Hampshire, and the North Basin rock glacier of Mt Katahdin, Maine. The difference between the modern permafrost elevation and the toe of the King Ravine rock glacier is ~800 m, and indicates a MAAT temperature difference of ~5.3°C. The toe of the North Basin rock glacier occurs ~720 m below the modern permafrost elevation, indicating a ~4.7°C difference in MAAT. These temperature estimates range with respect to the adiabatic lapse rate used. For instance: 6.0°C $km^{-1}$ (moist) ~ 8% warmer, 7.0°C = ~8% cooler, and 8.0°C $km^{-1}$ (dry) ~ 23% colder.

Our palaeotemperature estimates from the DLER and elsewhere in New England are very similar to values obtained by Hétu and Gray (2000b), who attributed relict rock glaciers occurring in the lowlands of the Gaspé Peninsula to a ~6°C lowering of MAAT. The observations of Hétu and Gray (2000b), together with the data from this study, were used to calculate a palaeo-permafrost–elevation gradient, shown in Fig. 6. Since rock glacier length can be limited by debris input and moisture availability for ice lens growth, this trend probably reflects a minimum estimate for the lower elevation of continuous permafrost at the time the rock glaciers formed. This does not preclude the possibility that permafrost occurred at lower elevations; however, the close agreement among the results from these four localities indicates that the rock glaciers of New England and maritime Canada may represent a close estimate for the former lower limit of continuous permafrost in the region.

This simple regional model should be considered a preliminary framework that requires future testing and refinement. Uncertainties involve topoclimatic effects that may complicate the relationship of permafrost distribution to regional climatic trends, as well as lapse rate choice (Walegur and Nelson, 2003). Also, the interpreted palaeotemperature lowering for the New England rock glaciers slightly underestimates the palaeo-MAAT value calculated using the classification of Humlum (1998), who suggested that active rock glaciers are most common in environments with MAAT less than −6°C. Despite these caveats, our results reflect simple and consistent regional trends, and thus we hypothesise that the rock glaciers of New Hampshire, Maine, and the Gaspé Peninsula reached their maximum postglacial extent at roughly the same time, during a period when MAAT was at least ~6°C cooler than today in New England and maritime Québec.

## Galilee Pond sediment core: results and implications

The Galilee Pond core consists of three sedimentary units: 0.1 m of unconsolidated sand, gravel and pebbles at the very bottom, 1.4 m of blue-grey sandy clay and silt at the base and 6.42 m of gyttja at the top (see Fig. 7). The bottom 10 cm contains striated, rounded pebbles of the Seboomook formation and angular granules from the local granodiorite bedrock. We infer the basal unit to be till, probably continuous with the glacial drift that surrounds the pond. Two terrestrial plant macrofossil samples sieved from the basal sediments 76 cm and 49 cm above the base of the core returned basal accelerator mass spectrometry (AMS) dates of 10 930 ± 40 and 11 320 ± 40 $^{14}C$ a BP, respectively (Table 2). Every 4 cm sample from the lower 1.5 m of the core contained some plant material, although in most cases the total dry weight of the material was insufficient for AMS analysis. *Dryas* leaves and twigs occurred in all samples. *Betula* flowers appear 50 cm above the base of the core and occur in each sample upwards into the gyttja.

Most basal ages of sediment cores recovered from northern Maine lakes have been obtained from plant macrofossils or bulk organics from gyttja overlying basal minerogenic sediments (e.g. Borns *et al.*, 2004; Dieffenbacher-Krall and Nurse, 2006). In this case, we collected plant macrofossils from within the mostly inorganic basal sediments, providing a closer minimum age for when the pond became free of overlying glacial ice. Radiocarbon dates indicate that the pond was deglaciated during the late Alleröd interstade, and plant macrofossils in the basal unit imply that a shrub–tundra environment persisted for much of the Lateglacial period in the region. We infer from these results that the northern Maine ice cap identified and discussed by Lowell (1985), Kite and Stuckenrath (1986), Lowell and Kite (1986), Lowell *et al.* (1990), Newman *et al.* (1985), Borns *et al.* (2004), Pelletier and Robinson (2005) and Nurse *et al.* (2006) must have withdrawn from the DLER before the late Allerød, with no evidence for Younger Dryas reoccupation (YD; 11 000–10 000 $^{14}C$ a BP; Mangerud *et al.*, 1974). This is consistent with the deglaciation isochrones of Borns *et al.* (2004) from central northern Maine, and with findings from the White Mountains and Mt Katahdin indicating pre-YD deglaciation (Davis, 1989, 1999).

The elevation of Galilee Pond, slow sedimentation rate (~0.07 cm $a^{-1}$) and the continuous occurrence of plant macrofossils throughout the basal silty clay indicate a terrigenous, non-glacial sediment source. This conclusion contrasts with previous assumptions that a similar basal clay unit found throughout northern Maine is glaciolacustrine in origin (i.e., Dieffenbacher-Krall and Nurse, 2006). Instead, we suspect that surface runoff may have gradually winnowed the relatively fresh till around the pond, transporting glacial silts and clays and contributing to the glaciogenic appearance of the blue-grey clay unit. Cryoturbation and aeolian processes also may have augmented sediment supplied to Galilee Pond during the Lateglacial period (Mott and Stea, 1993). Meltwater from the small GalRG could have contributed to the minerogenic basal sediments in the pond; however, this explanation cannot account for the widespread occurrence of similar basal sediments in lakes throughout northern Maine (i.e., Borns *et al.*, 2004; Dieffenbacher-Krall and Nurse, 2006).

The stratigraphy and macrofossils of the relatively inorganic basal sediments indicate that a cold, possibly periglacial environment existed during Lateglacial time, yielding climate conditions favourable for rock glacier growth. Core stratigraphy and LOI data show no indications that the climate reverted to similarly cold conditions at any time prior to deposition of the basal sediments, enhancing the probability that the DLER rock glaciers formed soon after the region was uncovered by ice at the end of the Pleistocene.

## Palaeoclimate implications

For the reasons stated above, the Lateglacial was probably the most favourable period of any time throughout the postglacial era for the growth of deep permafrost lenses required for rock glacier development from the coarse talus slopes of New England and Atlantic Canada. Regional palaeoecological interpretations (e.g. Levesque *et al.*, 1993, 1997; Stea and Mott, 1998; Borns *et al.*, 2004; Dieffenbacher-Krall and Nurse, 2006), in addition to more distant ice core analyses from



DOI: 10.1002/jqs

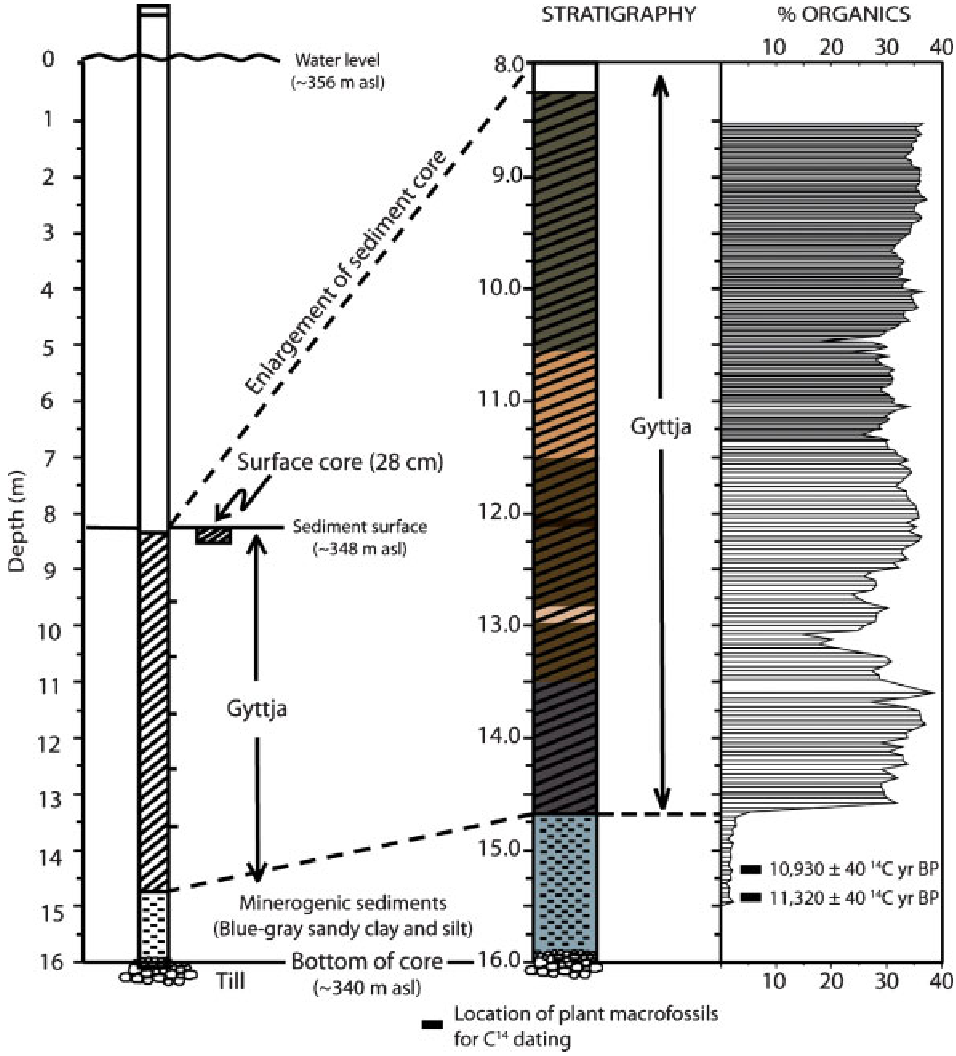


**Figure 7** Galilee Pond core diagram. Left side of figure shows water column and sediment core, and the enlarged sediment stratigraphy is displayed in the middle. Different shades of grey reflect the relative colour differences of the gyttja with depth. Diagonal lines indicate gyttja, and dot pattern is blue-grey sandy silt and clay. Percentage of organics inferred from loss-on-ignition analyses shown on right. Black squares signify depth of radiocarbon ages. This figure is available in colour online at www.interscience.wiley.com/journal/jqs

Greenland (e.g. Alley, 2000), show that the Lateglacial period was punctuated by the most recent severe cold events to occur in the North Atlantic region: the Killarney Oscillation (KO; 11 200–10 900 $^{14}$C a BP; Levesque *et al.*, 1993) and the YD cold period (Levesque *et al.*, 1993, 1997; Borns *et al.*, 2004). In New England, proxy records of postglacial climate show that the KO/YD oscillations bear the most outstanding signature of Lateglacial cooling (Borns *et al.*, 2004), which is in accord with our stratigraphic and LOI record from Galilee Pond.

Our analysis indicates that northern Maine was affected by a decrease in MAAT that led to the development of rock glaciers, but did not cause glaciers to readvance into the DLER. Additional supporting evidence for regional Lateglacial rock glacier growth comes from the Mont-St-Pierre Valley, Gaspé. In this region, a wave-cut marine terrace that incised a now-relict rock glacier has been dated to 10 300 $^{14}$C a BP (Hétu and Gray, 2000b), indicating that rock glacier expansion beyond glacier limits occurred during the YD on the Gaspé Peninsula. Kerschner (1978) and Frauenfelder *et al.* (2007) presented similar cases, in which they showed that YD-age rock glaciers in the Austrian and Swiss Alps, respectively, expanded to lower elevations than local glacier equilibrium lines. In addition, lateglacial rock-glacier expansion occurred in the British Isles well beyond the Loch Lomond moraine limits (Harrison *et al.*, 2008).

This situation is consistent with the findings of Denton *et al.* (2005), who proposed that MAATs during the YD, inferred from the isotopic composition of trapped gas in Greenland ice (e.g.

**Table 2** Radiocarbon dates from the Galilee Pond basal sediments. 'NR' indicates that there was insufficient mass for a $^{13}$C/$^{12}$C ratio measurement, and thus the AMS laboratory reported none. A conventional radiocarbon age was derived from a $^{13}$C/$^{12}$C ratio of a small aliquot of graphite that included laboratory and detector-induced fractionation, and therefore was not reported by Beta Analytic. This ratio corrects to the appropriate conventional radiocarbon age for sample GALM1516-12, however (D. Hood, personal communication, 2006)

| Sample ID | Depth (cm) | Age ($^{14}$C a BP) | $\delta^{13}$C (‰) | Sample description | Lab ID |
|---|---|---|---|---|---|
| GALM1516-12 | 676 | 10 930 ± 40 | −28.3 | Herbaceous twigs, *Betula* flowers and *Dryas* leaves | Beta-222924 |
| GALM1543-40 | 703 | 11 320 ± 40 | NR | One small herbaceous twig and several *Dryas* leaves | Beta-222923 |

Severinghaus and Brook, 1999), were most influenced by intensely cold winters that overwhelmed the summer temperature signal. They argued that intensely cold winters were a consequence of YD weakening of the thermohaline circulation, which allowed winter sea ice to spread over the North Atlantic. During that time the extreme seasonality initiated periglacial processes across northern Europe, as well as on the eastern seaboard of North America, far from active glacial ice (e.g. Borns, 1965; Stea and Mott, 1998; Hétu and Gray, 2000b). Therefore, given the generally accepted link between rock glacier occurrence and MAAT, as well as our preliminary inferences about the dampened sensitivity of the DLER rock glaciers to summer temperature, we suggest that the expansion of the New England and maritime Canada rock glaciers occurred in response to the severe winter climate of the YD period, despite comparatively mild summers.

Finally, we consider the possibility that the DLER rock glaciers may have been completely free of ice, and hence relict, during the middle Holocene when summers in New England were warmer (i.e. Sanger *et al.*, 2007), and that the modern ground ice observed in the DLER region represents the regrowth of permafrost during late Holocene cooling. This hypothetical situation seems reasonable, though we have no data to confirm or deny the possibility. However, we offer the following speculation. Since rock glacier activity is apparently most sensitive to MAAT, it seems possible that the strong seasonality imposed by Earth's precessional cycle during the early and middle Holocene could have resulted in zero net effect on rock glacier activity. In this scenario, the impact of warmer summers on ground temperature would have been cancelled by equally cold winters. Therefore, insolation alone may not sufficiently account for the Holocene regrowth of rock glacier ice, and thus any Holocene climate cooling responsible for permafrost growth must be of mean-annual character. Developing a landform chronology using surface exposure dating methods, more comprehensive lacustrine records, as well as chemical analyses of deep ice, could help to inform our understanding of these issues, and develop reasonable hypotheses.

## Conclusions

A series of talus landforms located in the Deboullie Lakes Ecological Reserve of northern Maine are classified as inactive and relict talus-derived rock glaciers on the basis of parabolic topographic profiles and subsurface thermal characteristics. Inactive and relict lobes indicate past rock glacier activity. Observations and temperature datalogger results of subsurface and above-surface conditions of the GPRG show that subsurface ice persists throughout the year in the lower portion of the slope. Subsurface temperature is largely decoupled from the magnitude of summer warmth in the lower, lobate portion of the landform, whereas during the winter subsurface temperatures appear to respond more immediately to above-surface temperatures. This is particularly the case during periods of internal cooling, which may be related to Balch ventilation, and may be important for the preservation of subsurface ice in lower portions of the rock glaciers.

Using the published archive of modern alpine permafrost occurrence, we estimate that a MAAT depression of at least $\sim$6°C is required to initiate active growth of the DLER rock glaciers. This estimate is consistent with the palaeotemperature depression derived from rock glaciers in Gaspésie, Québec (Hétu and Gray, 2000b), the King Ravine rock glacier in New Hampshire and the North Basin rock glacier of Mount Katahdin. Finally, the radiocarbon-dated Galilee Pond sediments and organic content indicate that the DLER rock glaciers most likely formed during the Lateglacial period. Similar to the case in Europe, rock glacier activity in the absence of significant glacier readvance in the DLER implies that a severe wintertime climate characterised the Lateglacial period in northern Maine.

*Acknowledgements* We thank the Maine Department of Conservation, Bureau of Parks and Lands, for access, logistical support and aerial photography. John Helstrom's expertise in monitoring and maintaining dataloggers was invaluable. We thank Bert Pelletier for his EDM surveying expertise; Andrea Nurse for providing coring equipment and field expertise; and Chunzeng Wang for his help drafting the figures. Karen Marysdaughter helped with LOI analyses. We are grateful for Mike Brophy's hospitality at Red River Camps, and for the use of his canoes. The wisdom, encouragement and persistent support of Gary Boone, the late Dee Caldwell, Marvin Caldwell and Bill Forbes precipitated many of the ideas presented in this study. Terry Hughes, Gordon Hamilton, Jorge Strelin, Roger Hooke, George Denton, Tom Weddle, Brenda Hall and Tom Lowell contributed helpful insights. We thank Michael Kaplan for reviewing early drafts of this paper, and Jeffrey Munroe and Woody Thompson for providing constructive reviews that greatly improved the manuscript. Finally, we thank all of the hard-working students who assisted with fieldwork during each expedition.

## References

Ackert RP Jr. 1998. A rock glacier/debris-covered glacier at Galena Creek, Absaroka Mountains, Wyoming. *Geografiska Annaler* **80A**: 267–276.

Alley RB. 2000. The Younger Dryas cold interval as viewed from central Greenland. *Quaternary Science Reviews* **19**: 213–226.

Aoyama M. 2005. Rock glaciers in the northern Japanese Alps: palaeoenvironmental implications since the Late Glacial. *Journal of Quaternary Science* **20**: 471–484.

Barsch D. 1996. *Rock Glaciers: Indicators for the Present and Former Geoecology in High Mountain Environments*. Springer: Berlin.

Bengtsson L, Enell M. 1986. Chemical analysis. In *Handbook of Holocene Palaeoecology and Palaeohydrology*, Berglund BE (ed.) Wiley: Chichester; 423–451.

Bernatchez L, Rhydderch JG, Kircheis FW. 2002. Microsatellite gene diversity analysis in landlocked arctic char from Maine. *Transactions of the American Fisheries Society* **131**: 1106–1118.

Birkeland PW. 1982. Subdivision of Holocene glacial deposits, Ben Ohau Range, New Zealand, using relative-dating methods. *Geological Society of America Bulletin* **93**: 433–449.

Boone GE. 1958. *The geology of the Fish River Lake–Deboullie area, northern Maine*. PhD dissertation, Yale University, New Haven, CT.

Borns HW Jr. 1965. Late glacial ice-wedge casts in northern Nova Scotia, Canada. *Science* **148**: 1223–1226.

Borns HW Jr, Doner LA, Dorion CC, Jacobson GL Jr, Kaplan MR, Kreutz KJ, Lowell TV, Thompson WB, Weddle TK. 2004. The deglaciation of Maine. In *Extent and Chronology of Glaciations. Vol. 2: North America, Maine*, Ehlers J, Gibbard P (eds). INQUA Commission on Glaciation. Elsevier: Amsterdam; 89–109.

Brazier V, Kirkbride MP, Owens IF. 1998. The relationship between climate and rock glacier distribution in the Ben Ohau Range, New Zealand. *Geografiska Annaler* **80A**: 193–207.

Broecker WS. 1998. Paleocean circulation during the last deglaciation: a bipolar seesaw? *Paleoceanography* **13**: 119–121.

Cannone N, Gerdol R. 2003. Vegetation as an ecological indicator of surface instability in rock glaciers. *Arctic, Antarctic, and Alpine Research* **35**: 384–390.

Clark DH, Steig EJ, Potter N Jr, Gillespie AR. 1998. Genetic variability of rock glaciers. *Geografiska Annaler* **80A**: 175–182.

Clark GM, Schmidlin TW. 1992. Alpine periglacial landforms of eastern North America: a review. *Permafrost and Periglacial Processes* **3**: 225–230.

Davis PT. 1989. Late Quaternary glacial history of Mt Katahdin and the nunatak hypothesis. *Maine Geological Survey: Studies in Maine Geology* **6**: 119–134.

Davis PT. 1999. Cirques of the Presidential Range, New Hampshire, and surrounding alpine areas in the northeastern United States. *Géographie physique et Quaternaire* **53**: 25–45.

Delaloye R, Lambiel C. 2005. Evidence of winter ascending air circulation throughout talus slopes and rock glaciers situated in the lower belt of alpine discontinuous permafrost (Swiss Alps). *Norsk Geografisk Tidsskrift* **59**: 194–203.

Denton GH, Alley RB, Comer GC, Broecker WS. 2005. The role of seasonality in abrupt climate change. *Quaternary Science Reviews* **24**: 1159–1182.

Dieffenbacher-Krall AC, Nurse AM. 2006. Late-glacial and Holocene record of lake levels of Matthews Pond and Whitehead Lake, Northern Maine, USA. *Journal of Paleolimnology* **34**: 283–309.

Frauenfelder R, Haeberli W, Hoelzle M, Maisch M. 2007. Using relict rockglaciers in GIS-based modeling to reconstruct Younger Dryas permafrost distribution patterns in the Err-Julier area, Swiss Alps. *Norsk Geografisk Tidsskrift* **55**: 195–202.

Gray JT, Brown RJE. 1979. Permafrost presence and distribution in the Chic-Choc Mountains, Gaspésie, Québec. *Géographie Physique et Quaternaire* **33**: 299–316.

Goldthwait RP. 1940. *Geology of the Presidential Range* Bulletin No. 1. New Hampshire Academy of Science: Hanover, NH; 1-43.

Goldthwait RP. 1970. Mountain glaciers of the Presidential Range in New Hampshire. *Arctic and Alpine Research* **2**: 85–102.

Gordon LS, Ballantyne CK. 2006. 'Protalus ramparts' on Navajo Mountain, Utah: reinterpretation as blockslope-sourced rock glaciers. *Permafrost and Periglacial Processes* **17**: 179–187.

Haeberli W, Vonder Mühll D. 1996. On the characteristics and possible origin of ice in rock glacier permafrost. *Zeitschrift für Geomorphologie* **104**: 43–57.

Hamilton SJ, Whalley WB. 1995. Rock glacier nomenclature: a reassessment. *Geomorphology* **14**: 73–80.

Harrison S, Whalley B, Anderson E. 2008. Relict rock glaciers and protalus lobes in the British Isles: implications for Late Pleistocene mountain geomorphology and palaeoclimate. *Journal of Quaternary Science* **23**: 287–304.

Heiri O, Lotter AF, Lemcke G. 2001. Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Paleolimnology* **25**: 101–110.

Hétu B, Gray JT. 2000a. Les étapes de la déglaciation dans le nord de la Gaspésie (Québec): les marges glaciaires des dryas ancien et recent. *Géographie Physique et Quaternaire* **54**: 5–40.

Hétu B, Gray JT. 2000b. Effects of environmental change on scree slope development throughout the postglacial period in the Chic-Choc Mountains in the northern Gaspé Peninsula, Québec. *Geomorphology* **32**: 335–355.

Hétu B, Gray JT, Gangloff P, Archambault B. 2003. Postglacial talus-derived rock glaciers in the Gaspé Peninsula, Québec (Canada). In *Proceedings of the 8th International Conference on Permafrost* Vol. 8, Phillips SM, Springman SM, Arenson LU (eds). Swetts & Zeitlinger: Lisse; 389–394.

Humlum O. 1998. The climatic significance of rock glaciers. *Permafrost and Periglacial Processes* **9**: 375–395.

Ikeda A, Matsuoka N. 2002. Degradation of talus-derived rock glaciers. *Permafrost and Periglacial Processes* **9**: 375–395.

Janke J, Frauenfelder R. 2008. The relationship between rock glacier and contributing area parameters in the Front Range of Colorado. *Journal of Quaternary Science* **23**: 153–163.

Johnson PG. 1983. Rock glaciers: a case for a change in nomenclature. *Geografiska Annaler* **65A**: 27–34.

Kerschner H. 1978. Paleoclimatic inferences from late Würm rock glaciers, eastern central Alps, western Tyrol, Austria. *Arctic and Alpine Research* **10**: 635–644.

Kirkbride M, Brazier V. 1995. On the sensitivity of Holocene talus-derived rock glaciers to climate change in the Ben Ohau Range, New Zealand. *Journal of Quaternary Science* **10**: 352–365.

Kite JS, Stuckenrath R. 1986. Postglacial history of the upper St John drainage basin. In *Contributions to the Quaternary Geology of Northern Maine and Adjacent Canada*, Kite JS *et al.* (eds). *Maine Geological Survey Bulletin* **37**: 113–128.

Konrad SK, Humphrey NF, Steig EJ, Clark DH, Potter N Jr, Pfeffer WT. 1999. Rock glacier dynamics and paleoclimatic implications. *Geology* **27**: 1057–1184.

Levesque AJ, Mayle FE, Walker IR, Cwynar LC. 1993. A previously unrecognized late-glacial cold event in eastern North America. *Nature* **361**: 623–626.

Levesque AJ, Cwynar LC, Walker IR. 1997. Exceptionally steep north-south gradients in lake temperatures during the last deglaciation. *Nature* **385**: 423–426.

Lowell TV. 1985. Late Wisconsin ice-flow reversal and deglaciation, northwestern Maine. In *Late Pleistocene History of Northeastern New England and Adjacent Québec*, Borns HW Jr, LaSalle P, Thompson WB (eds). Special Paper 197. Geological Society of America: Boulder, CO; 59–70.

Lowell TV, Kite JS. 1986. Deglaciation of northwestern Maine. In *Contributions to the Quaternary Geology of Northern Maine and Adjacent Canada*, Kite JS *et al.* (eds). *Maine Geological Survey Bulletin* **37**: 113–128.

Lowell TV, Kite JS, Calkin PE, Halter EF. 1990. Analysis of small-scale erosional data and a sequence of late Pleistocene flow reversal, northern New England. *Geological Society of America Bulletin* **102**: 74–85.

Mangerud J, Andersen ST, Berglund BE, Donner JJ. 1974. Quaternary stratigraphy of Norden: a proposal for terminology and classification. *Boreas* **3**: 109–128.

Millar CI, Westfall RD. 2008. Rock glaciers and related periglacial landforms in the Sierra Nevada, CA, USA: inventory, distribution, and climatic relationships. *Quaternary International* **188**: 90–104.

Mitchell WA, Taylor PJ. 2001. Rock glaciers in the Northwestern Indian Himalaya. *Glacial Geology and Geomorphology* **rp01**: 1–26.

Mott RJ, Stea RR. 1993. Late-glacial (Allerød/Younger Dryas) buried organic deposits, Nova Scotia, Canada. *Quaternary Science Reviews* **12**: 645–657.

National Climatic Data Center/NESDIS/NOAA. 2002. Monthly station normals of temperature, precipitation, and heating and cooling degree days 1971–2000. *Climatography of the United States* 81.

Nelson KJP, Nelson FE, Walegur MT. 2007. Periglacial Appalachia: palaeoclimatic significance of blockfield elevation gradients, eastern USA. *Permafrost and Periglacial Processes* **18**: 61–73.

Newman WA, Genes AN, Brewer T. 1985. Pleistocene geology of north-eastern Maine. In *Late Pleistocene History of Northeastern New England and Adjacent Québec*, Borns HW Jr, LaSalle P, Thompson WB (eds). Special Paper 197. Geological Society of America: Boulder, CO; 59–70.

Nurse AM, Borns HW, Hall BL, Dieffenbacher-Krall AC, Fastook JL. 2006. Geologic and paleoecologic evidence of ice-mass advance in northern Maine during the Younger Dryas cold reversal. *Geological Society of America Abstracts with Programs* **38**: 236.

Paasche Ø, Dahl SO, Løvlie R, Nesje A. 2007. Rockglacier activity during the Last Glacial–Interglacial transition and Holocene spring snowmelting. *Quaternary Science Reviews* **26**: 793–807.

Pelletier BG, Robinson BS. 2005. Tundra, ice and a Pleistocene cape on the Gulf of Maine: a case of paleoindian transhumance. *Archaeology of Eastern North America* **33**: 163–176.

Péwé TL. 1983. Alpine permafrost in the contiguous United States: a review. *Arctic and Alpine Research* **15**: 145–156.

Potter N Jr. 1972. Ice-cored rock glacier, Galena Creek, northern Absaroka Mountains, Wyoming. *Geological Society of America Bulletin* **83**: 3025–3058.

Richard PJH, Veillette J, Larouche AC, Hétu B, Gray JT, Gangloff P. 1997. Chronologie de la déglaciation en Gaspésie: nouvelles données et implications. *Géographie Physique et Quaternaire* **51**: 163–184.

Sailer R, Kerschner H. 1999. Equilibrium line altitudes and rockglaciers in the Ferwall-Group (western Tyrol, Austria) during the Younger Dryas cooling event. *Annals of Glaciology* **31**: 275–280.

Sanger D, Almquist H, Dieffenbacher-Krall A. 2007. Mid-Holocene cultural adaptations to central Maine. In *Climate Change and Cultural Dynamics: A Global Perspective on Mid-Holocene Transitions*, Anderson DG, Maasch KA, Sandweiss DH (eds). Elsevier: Amsterdam; 435–456.

Severinghaus JP, Brook EJ. 1999. Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* **29**: 930–934.



DOI: 10.1002/jqs

Smith HTU. 1973. Photogeologic study of periglacial talus glaciers in northwestern Canada. *Geografiska Annaler* **55A**: 69–84.

Spear RW. 1989. Late-Quaternary history of high-elevation vegetation in the White Mountains of New Hampshire. *Ecological Monographs* **59**: 125–151.

Stea RR, Mott RJ. 1998. Deglaciation of Nova Scotia: stratigraphy and chronology of lake sediment cores and buried organic sections. *Géographie Physique et Quaternaire* **52**: 1–19.

Steig EJ, Fitzpatrick JJ Jr, Potter N Jr, Clark DH. 1998. The geochemical record in rock glaciers. *Geografiska Annaler* **80A**: 277–286.

Thompson DJ. 1999. Talus fabric in Tuckerman Ravine, New Hampshire: evidence for a tongue-shaped rock glacier. *Géographie physique et Quaternaire* **53**: 47–57.

Thompson WF. 1961. The shape of New England mountains. *Appalachia, Pt III* **27**: 458–478.

Thompson WF. 1962. Preliminary notes on the nature and distribution of rock glaciers relative to true glaciers and other effects of climate on the ground in North America. *International Association of Scientific Hydrology* **58**: 212–219.

Walegur MT, Nelson FE. 2003. Permafrost distribution in the Appalachian highlands, northeastern USA. In *Proceedings of the 8th International Conference on Permafrost*, Vol. 2 Phillips M, Springman SM, Arenson LU (eds). AA Balkema: Lisse; 1201–1206.

Wahrhaftig C, Cox A. 1959. Rock glaciers in the Alaska Range. *Geological Society of America Bulletin* **70**: 383–436.

Washburn AL. 1980. *Geocryology: A Survey of Periglacial Processes and Environments*. Halsted Press: New York.

Whalley BW, Azizi F. 2003. Rock glaciers and protalus landforms: analogous forms and ice sources on Earth and Mars. *Journal of Geophysical Research: Planets* **108**: 8032–8045.

White SE. 1976. Rock glaciers and blockfields: review and new data. *Quaternary Research* **6**: 77–97.

Wilkerson B. 2007. *Natural resource inventory of the northern Aroostook region: Deboullie*. Bureau of Parks and Lands, Department of Conservation, Augusta, ME.

World Meteorological Organization Storm Data and Unusual Weather Phenomena. 2004. http://findarticles.com/p/articles/mi_m0QRF/is_8_46/ai_n9481382 [August, 2004].