Canadian Journal of Earth Sciences
Revue canadienne des sciences de la Terre

**Late-Glacial and Holocene evolution as a driver of diversity
and complexity of the northeastern North American alpine
landscapes: A synthesis**

Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2016-0004.R1
Manuscript Type:	Article
Date Submitted by the Author:	31-Mar-2016
Complete List of Authors:	Martin, Jean-Philippe; Universite du Quebec a Montreal, Institut des Sciences de l'environnement GERMAIN, Daniel; Universite du Quebec a Montreal, Département de Géographie; Universite du Quebec a Montreal, Institut des Sciences de l'environnement
Keyword:	Holocene, Alpine environment, Paraglacial, Resilience, Synergy

SCHOLARONE™
Manuscripts

**Late-Glacial and Holocene evolution as a driver of diversity and complexity of the
northeastern North American alpine landscapes: A synthesis**

Jean-Philippe Martin¹ and Daniel Germain^{1,2}

¹Institute of Environmental sciences, Université du Québec à Montréal, Canada.

martin.jean-philippe.3@courrier.uqam.ca

²Geography Department, Université du Québec à Montréal, Canada

germain.daniel@uqam.ca

Corresponding author

Jean-Philippe Martin, Institute of Environmental sciences, Université du Québec à Montréal,
P.O. Box 8888, Station Centre-Ville, Montreal, Quebec, Canada, H3C 3P8.

Phone: 1 514 987-3000 #6455

Email: martin.jean-philippe.3@courrier.uqam.ca

Abstract

Mid-altitude, mid-latitude mountains are complex environments due to their Pleistocene glacial heritage, the importance of geomorphic processes on the steep slopes, and the climatic conditions which are often close to periglacial. These factors, along with the fragmentation of the alpine habitats, enhance the topographic and floristic diversity of these environments in northeastern North America. Through case studies, this synthesis underlines the interactions between the geosphere (glacial, paraglacial, and periglacial processes), the atmosphere (climatic fluctuations) and the biosphere (vegetation establishment and evolution to the present-day) that explain the low elevation of the northeastern North American alpine environment and that testify to its complexity. Vegetation established earlier in the southern ranges, following the same general trend as the Laurentian Ice Sheet recession. However, local factors such as ice retreat, response to global scale climate changes and paraglacial processes acted in synergy to increase the resilience and to influence the occurrence of alpine landscapes. The establishment of the latter environment can therefore be considered to be azonal. Finally, our findings highlight the lack of a conceptual framework, systemic studies and multi-proxy reconstructions of alpine environments located at the limit of bioclimatic zones controlled by the equilibrium between biostatic and rhexistatic regimes.

Keywords

Holocene, alpine environment, paraglacial, resilience, synergy

1. Introduction

The diverse and combined influences of many orogenies – some in the recent past on a geological timescale – with the many subsequent erosion agents, which depend on climatic variations at local and regional scales, increase the complexity of the mountain environments and their spatiotemporal evolution. If alpine glaciers and permafrost recession are at the heart of recent scientific interest because of ongoing climatic change (e.g., Harris et al. 2009; Jomelli et al. 2011), the ecological evolution of alpine and subalpine environments at lower altitudes is equally concerning (Grabherr et al., 2010; Engler et al., 2011; Dullinger et al., 2012). Indeed, mid-altitude mountains are also foreseen to undergo profound modifications in the next decades and century, hence the interest to better understand the past and ongoing evolution of these unique environments which support a rich biodiversity, generate many natural georisks, and sustain numerous anthropogenic activities and needs such as agriculture, water resources, tourism and outdoor activities (Messerli and Ives 1997; Sagoff, 2002; Price et al. 2004).

While mid-altitude mountains often lack distinct glacial and periglacial levels, the presence of a topographic gradient influences the development of a variety of environments with distinct biophysical characteristics. In light of these observations, mid-altitude mountains should be thought of as the outcome of a long and complex evolutionary process, the latter being driven by non-linear feedbacks in space and time between many components of the geosphere (tectonics, pre-Quaternary and glacial heritages, geological structure, sediment transfer processes), the atmosphere (climatic variations at different spatial and temporal scales) and the biosphere (vegetation cover, disturbance regime, ecosystem resilience). While recent recognition of potential global-scale interactions between tectonics, climate, surface processes and vegetation

dynamics is well documented (e.g., Winkworth et al. 2005; Bishop 2007; Whipple 2009), very few examples are available in the scientific literature concerning mid-altitude mountains. Moreover, the general lack of fine-scale temporal and spatial resolution in existing studies impedes the provision of detailed information regarding the patterns and processes of these systems' interactions.

In northeastern North America (i.e., latitude 43 to 50° N), where the elevation of the treeline is well below the worldwide pattern established by Körner (1998) (Figure 1), the low level of human disturbance of the alpine and subalpine belts provides a unique opportunity to study and document the synergistic influence of causal factors responsible for the current mountain landscapes. At a regional scale, Cogbill and White (1991) reported the mean July temperature of 13 °C isotherm as a climatic control of the treeline in these mountain ranges. However, fine-scale observations show significant discrepancies related to, among other factors, the proximity of the ocean and climate variability, the toxicity of the serpentine bedrock, the importance of geomorphic processes on steep slopes, and the ecological disturbance regime. It is the aim of this paper to use case studies to illustrate the diversity of northeastern North American subalpine-alpine environments as a complex evolution based on the timing of deglaciation, the pattern of vegetation establishment, the frequency-magnitude of geomorphic processes and the ecological disturbance regime, and their responses to climate variability throughout the Holocene epoch. Future challenges and opportunities in the context of the ongoing global change will also be discussed.

2. Study area

2.1 Geographic setting

Mountain ranges with considerable alpine environments are spread throughout the northeastern United States and eastern Canada (Fig. 2). The ranges included in this synthesis are the Catskills (New York), the Adirondacks (New York), the Green Mountains (Vermont), the White Mountains (New Hampshire) and the Mount Katahdin range (Maine) from New England (USA). The ranges from the province of Quebec in Canada are the Charlevoix Highlands, as well as the Chic-Chocs of the Gaspé Peninsula, the latter also including the McGerrigle Mountains if not specified otherwise. The ranges from the Atlantic provinces of eastern Canada include the Cape Breton Highlands (Nova Scotia) and the Long Range Mountains (western Newfoundland).

2.2 Physical characteristics

Most of northeastern North America’s current alpine environments were uplifted during the Taconic or the Alleghenian phases of the Appalachian orogeny – 450 M yrs ago and 325 to 260 M yrs ago respectively (Titus 2004). The difference in lithology between certain Appalachian ranges has a profound ecological impact. For example, the Chic-Chocs Mountains are composed of Ordovician limestone, shale, argillite, greywacke, as well as siltstone with occasional layers of sandstone (Enos 1969). The Mont Albert plateau, one of the highest peaks of the range, can be considered a distinct unit because of its high serpentine content which supports a unique flora within the studied region (Sirois 1984). Intruded Devonian magmatic material forms the highest summits of the range adjacent to the Chic-Chocs, the McGerrigle Mountains, which are

composed mainly of granite and granitoid material (Gray and Brown 1979). The proximity of these three distinct units (i.e., limestone, serpentine and granite) in an otherwise similar environment in terms of topography and climate led early scientists to conclude that the geology controlled the alpine floral diversity (Fernald 1907). The High Peaks of the Adirondack Mountains and the Charlevoix Highlands are the only ranges from the Grenville Province, composed mainly of Precambrian gneiss, granite and anorthosite (Hargraves and Roy 1974). Table 1 presents the maximal elevation of the aforementioned ranges. The highest summit of the study region is Mount Washington (1 971 m) in the White Mountains.

2.3 Alpine environment

The high mountains of northeastern North America follow a classical altitudinal zonation. Montane mixed-coniferous forest covers the lower elevations, followed by a subalpine spruce-fir or boreal forest, which spreads upward to the treeline. The alpine environment is defined as the treeless, cold and windy high-altitude area above treeline on mountains (Löve 1970). In northeastern North America, the alpine environment presents a mosaic of patches on the highest summits, and not a continuous territory (Table 1). At the regional scale, there are gaps of many hundreds of kilometers between the alpine ranges. At the range scale, alpine environments are also distributed in patches, separated by forested valleys. The elevation of the treeline follows the same decreasing trend along a latitudinal gradient that is observed worldwide (Figure 1). The Chic-Chocs and the Long Range have the largest areas of land above the treeline. This is consistent with the flat summit topography of these ranges, which are post-orogenic planation surfaces (Jutras and Schroeder 1999), along with their higher latitude.

3. Methods

The last decade has seen an increasing scientific interest in these altitude environments for various reasons such as the increase of extreme hydrometeorological events (Germain and Hétu, *in press*), the decrease of the woodland caribou populations in Quebec (Johnson et al., 2015) or the response of the arctic-alpine flora to the ongoing climate changes (Bailey et al., 2015). The onset of different initiatives demonstrate this renewed interest: a yearly occurrence of the workshop entitled Northeast Alpine Stewardship Gathering, the publication of the Eastern Alpine Guide (Jones and Willey, 2012) — a book about wildness and conservation with the contribution of twenty authors, most of them scientists — and a scientific paper highlighting research priorities and the need for a network of long-term alpine monitoring sites where a standardized protocol would be used to collect data on biotic and abiotic parameters (Capers et al., 2013). In that regards, two sites located in the Chic-Chocs Mountains (Quebec) and the White Mountains (New Hampshire) were implemented following the international protocol of GLORIA (GLobal Observation Research Initiative in Alpine environments) in the last few years. However, as recently reported by Capers et al. (2013, p.559), *research in alpine areas of northeastern North America has been poorly coordinated, with minimal communication among researchers, and it has rarely been multidisciplinary.*

As a first attempt to address the aforementioned gap, the following results lean on a meta-analysis of the existing literature about the glacial, palaeoecological and geomorphological evolution of the mountain ranges of Northeastern North America. Indeed, there is a wide corpus of papers in regional and international scientific journals that provide chronologies with regard to the glacial retreat, the vegetation settlement, the evolution of geomorphic processes and systems.

A bibliographic database containing 109 references was created (Supplementary material 1), but to increase the readability, only the most relevant are integrated to this paper. We are aware that the present analysis does not include every aspects of the alpine environment that is covered by the scientific literature, such as the insects, the lichens, the mammals (especially the woodland caribou), local climatology or the geology (lithology and past orogenies). Rather than an exhaustive catalogue of every previous study done in the alpine environment of northeastern North America, this work represents the first step of a transdisciplinary, large-scale analysis of alpine environments since the Late-Glacial. It is the hope of the authors that this synthesis serves as a starting point for more detailed analysis and scientific work about the long-term biophysical conditions that characterise these mid-altitude mountains.

Tables 2 and 3 detail the sites and types of records that were used in the following glacial and paleoecological reconstructions. Most of the chronological data are therefore derived from ^{14}C dates of pollen assemblages and macrofossils. Unless specified otherwise, all dates are given in calendar years. CALIB 6.0 software (Reimer et al. 2009) was used to calibrate radiocarbon dates with a standard deviation of 100 years when calendar dates were unavailable. Calibrated dates were rounded to the nearest 100 years and are labeled Before Present (BP).

4. Results and discussion

The meta-analysis of the literature allowed the reconstruction of the chronology of the general deglaciation and vegetation establishment throughout the study region. These general patterns provide an overlook of the inception of alpine environments on the higher summits of Northeastern North America. Different case studies are then presented in order to propose

concepts that encompass the complex relationships explaining the local discrepancies in the altitudinal and latitudinal localisation of the treeline. First, from a geomorphological standpoint, the elasticity of the paraglacial period, explicated by local topographical, geographical and geological factors, will influence nowadays equilibrium between rheostatic and biostatic processes on the steep slopes. Next, the late-Holocene fluctuations in climate acted in synergy with punctual ecological disturbances to create localized alpine environments on previously forested summits. Finally, the inception and preservation of permafrost in the study regions, along with the fact that the alpine treeline is controlled by disturbances rather than temperature, explain the delicate balance between the sensitivity and the resilience of the alpine landscape.

4.1 Deglaciation pattern, vegetation establishment and regional climatic changes

4.1.1 Deglaciation

Although the patterns of deglaciation of the Appalachian Highlands and the supposed presence of nunataks have been debated for nearly a century across the study area (Fernald 1925; Bradley 1981; Gosse et al. 2006; Olejczyk and Gray 2007), it is now well-established that the Laurentian Ice Sheet (LIS) flowed across the mountain ranges and highlands of New England and Quebec, and connected with regional ice caps in the Atlantic Provinces and in the Chic-Chocs Mountains during the Last Glacial Maximum (LGM). The recession of the LIS in New England followed a general South to North trend. The onset of the glacial recession is poorly dated; the best estimate of its beginning is the onset of varved sedimentation at the south end of Glacial Lake Hitchcock in southern Connecticut, at 15,600 ¹⁴C yr BP (Dyke et al., 2002). The southwestern ice masses,

such as the White Mountains ice cap, remained connected with the LIS, while the Gaspé Peninsula and Newfoundland ice caps were isolated.

The approximate date of ice recession in the different alpine ranges covered in this paper is presented in Figure 3. These data show the general trends, but any comparison between them remains uncertain, since the recessional dates were determined by various proxies (Table 2) that represent a range of spatial scale dynamics from regional (e.g. pollen) to local (e.g. gyttja). Moreover, there is much variation in the elevation at which the data were derived. At the global scale, New England was deglaciated before the Younger Dryas (12,900-11,400 BP). There is then a long time gap before deglaciation begins in the northernmost ranges, which suggests that the Younger Dryas delayed the melting of alpine ice caps located further north. In most ranges, the timing of ice recession is consistent with the deglaciation isochrones map of Southern Quebec and adjacent territories (Richard, 2009), with the exception of the White Mountains, for which the local data are older. This implies that the regional scale pattern might not reflect the local specificities of ice recession in the alpine ranges.

In the Gaspé Peninsula, variations in basal dates from different records are important to consider. Some are much younger than others, which led Richard et al. (1997) to propose the presence of late small glaciers in some of the cirques of the Chic-Chocs. While some hypotheses have been proposed to explain the discrepancy (e.g., presence of an alpine desert or late ice), the absence of direct evidence dictates that caution must be taken when interpreting regional trends from a single or few basal dates.

4.1.2 Vegetation establishment

The postglacial vegetation establishment in northeastern North America followed three stages: tundra, afforestation and closed-forest stages. The general chronology for the vegetation dynamics of different mountain ranges is presented in Figure 3, which must, however, be treated with caution for several reasons:

- 1) *Contamination from organic carbon and radiocarbon accuracy.* The dates from the Chocoma Highlands are from Richard et al. (1997), who demonstrated that the basal date of a core sampled in 1979 was too old, due to the contamination of the dissolved CO₂ by inactive carbon from the bedrock. Based on this evidence, other dates are potentially too old at other sites as well;
- 2) *Altitude.* The altitude at which the studies were conducted varies greatly, from the mixed-coniferous montane forest to the upper limit of the spruce-fir subalpine belt;
- 3) *Availability of accelerator mass spectrometry (AMS) dating.* As for old carbon contamination, AMS dating of macrofossils instead of total organic matter has allowed more accurate chronologies. However, northeastern North American alpine paleoecology was mainly published before 1995, when these dating techniques still were not widespread. However, above the treeline in the alpine environment, the pollen influx is low. The pollen from distant sites is over-represented and therefore allows only the reconstruction of the vegetation at the regional level and makes the interpretation of treeline evolution particularly difficult; hence the need for macrofossil analysis in treeless environments (Birks and Birks 2000).

At the regional scale, the vegetation establishment pattern in alpine environments followed the ice recession during the Late-Wisconsinan (Figure 3), the southernmost ranges being colonized

first. At Mount Katahdin, the tundra stage was synchronous with the Adirondacks and the White Mountains, but the afforestation stage was longer and potentially delayed by the onset of the Younger Dryas. Moreover, there is a wide temporal gap in the vegetation establishment between the alpine ranges of New England and of Quebec (Figure 3). The proximity of the Quebec ranges with the Gulf of St. Lawrence and the Atlantic Ocean would have increased the cooling effect of the Younger Dryas compared with the southern, more continental ranges (Cwynar and Spear 2001), therefore making it impossible for vegetation establishment to begin during this climatic period. Afterward, the recurrence of cold episodes following the Younger Dryas (Atlantic Preboreal Oscillation and the 8,200 BP cooling period), enhanced by the proximity with the LIS and a high volume of freshwater runoff in the St. Lawrence River, would have delayed it further (Anderson et al. 2007).

There are however some discrepancies between the deglacial and paleoecological records. For example, in the Charlevoix region, deglaciation is thought to have occurred around 12,500 BP at low altitudes from the presence of the St. Narcisse moraine. However, the oldest record from organic material in the Highlands dates from 10,500 BP. Two hypotheses could explain the gap of 2,200 years between ice recession and vegetation establishment; i) like in the Gaspé Peninsula, the Younger Dryas period might have been favourable to the build-up of alpine glaciers on the highest summits of Charlevoix; ii) it is likely that an alpine desert occurred between the ice recession and the tundra stage, such as reported for the Chic-Chocs Mountains.

This regional picture allows depicting the general trend of ice retreat and alpine environment establishment along a SW – NE gradient. At the local scale, the following case studies illustrate that complex interactions between geomorphological, ecological and climatic systems create a set

of conditions for the establishment, the diversity and the preservation of the alpine environments throughout the Holocene.

4.2 Complexity as the basis of the inception and conservation of alpine landscapes

4.2.1 Paraglacial time magnitude and elasticity impact on the alpine landscape

The alpine landscapes of northeastern North America have been geomorphologically active since the Late-Glacial period, but the frequency, magnitude and efficiency of the processes involved have varied following deglaciation and the Holocene climatic fluctuations. The recession of the LIS or the Appalachian ice caps exposed the landscape to paraglacial processes corresponding to isostatic instability and high sediment availability, providing a context of high geomorphic activity on the steep slopes. However, the scientific literature on Holocene hillslope geomorphology in the alpine ranges of northeastern North America comes almost exclusively from the Gaspé Peninsula (regional synthesis are provided by Hétu and Gray 2000; Germain et al. 2010), with few studies focusing on the present-day processes in the Adirondack Mountains (e.g. Bogucki, 1977), the White Mountains (e.g. Kull and Magilligan, 1994), the Cape Breton Highlands (e.g. Wahl et al. 2007) and the Newfoundland Long Range (e.g. Hiscott and James, 1985).

The paraglacial processes were central in shaping the alpine landscape in this region through their dynamic interactions with ecological processes. Hillslope processes in the White Mountains during the Early-Holocene rendered the vegetation establishment more complex. The beginning of pollen accumulation was approximately synchronous at low (200 and 650 m) and high (1,325

and 1,542 m) altitudes (Spear et al. 1994). Before 13,500 BP, all these sites recorded tundra vegetation. During this period, there is an absence of vegetation in the pollen records at mid-elevation sites (1,140 and 1,275 m), the earliest ones corresponding to 11,500 BP. Spear et al. (1994) suggested that the mid-altitude active steep slopes prohibited vegetation establishment until their stabilization, which points to the role of paraglacial geomorphic agents in the ecological evolution of the White Mountains.

The duration of the paraglacial period is directly related to the glacially conditioned sediment availability and varies from 10 to > 10,000 years. It is assumed that for an isolated landsystem, the paraglacial period ends at the exhaustion of this sediment stock, the latter being characterized by an exponential decline. However, secondary processes such as debris flows occurring in primary paraglacial stock in scree slopes, can rework glacial sediments stored in primary paraglacial stocks, further extending the duration of this period (Ballantyne 2002). Hétu and Gray (2000) define the end of the paraglacial period in the northern Gaspé Peninsula as the complete burial of the rock walls by rockfalls, and later reworking of the debris by dry grain flows, frost-coated clasts flow, debris flows, snow avalanches and niveo-aeolian transport. At present, the coastal valleys of the northern Gaspé Peninsula have a diachronous pattern of paraglacial influences. For example, in the Mont Saint-Pierre valley, forest vegetation covers the east-facing slopes where rockwalls have been almost completely eliminated, while on the west-facing slopes, where summit rockwall segments are still active, the upper part of the scree slope is still devoid of vegetation. As many studies highlight (*cf.* Hétu and Gray 2000), the upper limit of the forest on these slopes has been changing constantly throughout the Holocene, in response to hydroclimatic conditions. A warmer and drier climate generally tends to favour an upward movement of the treeline via a decrease in the frequency and magnitude of mass wasting

processes; this tendency reverses in colder and wetter periods. However, the treeline on the scree slopes of Mont Saint-Pierre has been extending upward from the end of the Little Ice Age until 1950. Since then, geomorphic processes have increasingly fragmented the treeline despite a positive trend in the temperature (Lafortune et al., 1997). Therefore, vegetation dynamics and paraglacial processes reach a dynamic equilibrium conditioned by complex geomorphic responses to the changes in hydroclimatic conditions.

The differences in the duration of the paraglacial period are even more striking on a larger scale. The Long Range and the Chic-Chocs contain similar drift-mantled landsystem based on similar processes and with similar lithologies, and have been deglaciated for approximately the same length of time. However, there seems to be an important difference in the efficiency and the magnitude of the present-day geomorphic processes. For example, on the Tablelands in Newfoundland, debris fans of a few hundred meters wide are common and still active, since they are not colonized by vegetation at an altitude close to sea level. On Mont Albert in the Chic-Chocs, debris fans of this magnitude are relict nowadays, but are superimposed by active debris cones with a surface and debris size smaller by close to an order of magnitude (Germain 2005). This hydrosedimentological transition appears related to hydroclimatic conditions, which were favourable to the definitive closure of the regional forest cover as established by pollen analysis (Hétu and Bail, 1996). This example underlines the importance of local ecological and hydroclimatic conditions in controlling the duration of exhaustion of glacially conditioned sedimentary stocks and consequently the delay of paraglacial responses of geomorphologic landsystems. Moreover, the overlapping of landforms, such as the debris cones of different scales on Mont Albert, is evidence of a more mature, postglacial landscape, with similar processes driving the sediment cascade, but to a different magnitude.

This section highlights the importance of the paraglacial concept to understanding the formation of the northeastern North American montane landscape, as well as the ecological evolution of the steep slopes by the rheostatic forcing that influenced the vegetation establishment and vice versa. Considering that the essence of the paraglacial concept is to define recently deglaciated areas that are in an unstable geomorphological state, vulnerable to erosion processes, some slopes of the mountain regions of northeastern North America are still influenced by paraglacial processes more than 10,000 years after the onset of the Holocene period. This dichotomy between active and relict paraglacial periods is visible at the local and regional scales. Therefore, not only is there an overlapping of temporal scales amongst paraglacial periods from different landsystems (Mercier 2008), but also an overlapping of spatial scales controlled by external variables such as vegetation and hydroclimatic conditions.

4.2.2 Synergistic effects of climate change and ecological disturbances

The data in Figure 1 shows the expected negative relationship between the altitude of the treeline and the latitude of the mountain range for northeastern North America. However, the altitude of the treeline in some ranges, such as the White Mountains and the Charlevoix Highlands, is lower than would be expected from this figure, as well as from a comparison with other mountain ranges at similar latitudes around the world. While regional climatic latitudinal gradient as well as deglaciation and vegetation establishment processes explain the general trend, local dynamics are responsible for such discrepancies. As the following case studies demonstrate, it is not one system that is directly responsible for the lower elevation of the treeline in certain ranges but rather the synergistic effects of different systems.

384
385 The subalpine spruce-fir forest composition of the Appalachians differs from that of the boreal
386 forest in disturbances controlling their dynamics, among other factors (Cogbill and White 1991).
387 The subalpine spruce-fir forest is controlled by wind and individual tree mortality (Sprugel
388 1976), whereas the boreal forest dynamics are mainly controlled by ecological disturbances such
389 as fire and insect outbreaks (Cogbill 1985).

390
391 The Charlevoix Highlands are located within the boreal forest belt. Charcoal found on a summit
392 was dated to 3,150 BP, which shows that it was previously forested and then decimated by fire
393 (Bussires et al., 1996). Today, open bogs occupy depressions in the upper slopes of these
394 summits. The low regeneration rate following these fires is probably due to the deterioration of
395 climatic conditions, since evidences of synchronous climatic cooling responsible for the increase
396 in fire activity have been recorded in Charlevoix and in the Gasp Peninsula. The paludification
397 of the alpine zone is also indicative of a generally positive moisture surface balance during the
398 Neoglacial. This case highlights the azonal nature of alpine and subalpine altitudinal belts of the
399 Charlevoix Highlands, since they do not result from a climatic gradient associated with
400 topography, but rather from a modification in the ecological dynamics caused by the large-scale
401 Late-Holocene climatic deterioration.

402
403 In the White Mountains, the treeline of the subalpine spruce-fir forest moved upslope during the
404 Younger Dryas (between 12,900 and 11,500 BP), reaching altitudes higher than in the present-
405 day. This caused scientists to state that the Younger Dryas had not influenced the high-elevation
406 vegetation of the White Mountains (Spear et al. 1994). However, later studies based on loss-on-
407 ignition and chironomids suggest that even if there was an attenuation of the effects of this

cooling episode westward from the Maritime provinces, the temperature of mid-altitude lakes in the White Mountains dropped by approximately 5 °C during the Younger Dryas (Cwynar and Spear 2001). Therefore, the treeline was highest in the White Mountains during one of the coldest periods since ice recession. The beginning of Younger Dryas saw the occurrence of a shift from northwesterly winds during the Bolling-Allerod period to northeasterly katabatic prevailing winds in New England (Thorson and Schile 1995) and Quebec (Filion 1987). Present-day winds at the top of Mount Washington are extremely strong (Gordon 1989), which is the result, among other causes, of the orientation of the mountain range perpendicular to the prevailing winds. The upward progression of the treeline at Mount Washington, controlled primarily by these extreme winds and rime episodes, could be caused by this shift in prevailing winds to an axis more parallel to the range, and thus decreasing their velocity. This anomaly in treeline position was preserved during the warmer, Mid-Holocene climatic optimum to finally regress to its present elevation during the Neoglacial, showing that the lower altitudinal limit of the alpine environment in this area does not respond to cold episodes or strong winds individually, but rather to a synergistic relationship between these two variables.

4.2.3 Sensitivity and Resilience of the Alpine Landscape

Permafrost. Permafrost patches are still present in the Charlevoix Highlands, Mont Jacques-Cartier in the McGerrigles, Mount Washington in the White Mountains and in western Newfoundland (Figure 4; Clark and Schmidlin, 1992). Except for the latter, these ranges are well beyond the latitudinal limit of discontinuous permafrost distribution in Canada. Based on temperature extrapolation, Walegur and Nelson (2003) estimated that the presence of permafrost would be possible on the highest summits of the High Peaks of the Adirondacks (Mount Marcy

and Whiteface) and of northern Maine (Sugarloaf Mountain, Snow Mountain, Crocker Mountain and Mount Katahdin).

Permafrost is considered to be a system that is sensitive to ongoing and anticipated climate change (Camill and Clark 2000). In the Chic-Chocs, however, 30 years of monitoring a permafrost patch led to the conclusion that the latter is stable (Gray et al., 2009). The permafrost of Mont du Lac des Cygnes in the Charlevoix Highlands is well below the elevation trend of the other summits where permafrost bodies were observed. The approximate mean annual temperature of around 0°C for adjacent summits would not be sufficiently cold for the inception of permafrost. This suggests that it was formed during a colder period. In Charlevoix, the climatic conditions during the Neoglacial were not cold enough to trigger the formation of perennially frozen ground (Zimmerman and Lavoie 2001). Therefore, the permafrost islands in Charlevoix are estimated to have formed during the Little Ice Age, in the last 500 years, in response to a unique set of climatic and ecological conditions. Fires were very frequent in the Charlevoix Highlands in the last 3,000 years (Bussires et al. 1996). The opening of the vegetation, combined with the strong winds characteristics of this region and the cooling of the climate, led to little snow accumulation on the summits and to deeper ground penetration of the frost wave during the winter. The cool climate reduced evaporation and was favourable to the accumulation of moss, which isolated the ground ice. These cases show that permafrost patches in this study area are not climate driven, but their inception rather occurred in a precise eco-climatic context. Since present-day ecological conditions are responsible for their resilience, they evolved into an ecosystem-protected permafrost, which means that they are unlikely to recover after a significant disturbance with the ongoing change in climate (Shur and Jorgenson, 2007).

The resilience of mountain top permafrost in marginal environmental conditions at latitudes lower than the regional pattern of permafrost distribution is surprising and must be caused by the interaction of the different systems. While the mean annual air temperature following the Little Ice Age got warmer, the absence of trees on the summit allowed the strong winds to cause enough snow drifting to clear the snow or compact it sufficiently to increase its thermal conductivity. Both cases would result in a negative mean annual ground temperature.

Rock glaciers. Rock glaciers are widespread in the study region. Many were identified at low (20 m) and high (>600 m) elevations in the Chic-Chocs and almost all of these rock glaciers were located on a scree slope at the base of a rock wall. These objects form by the combination of paraglacial processes and periglacial conditions (Humlum, 1998) that influenced the landscape during the Late-Holocene. The distribution of the altitudinal limit of rock glaciers followed the same latitudinal trend as present-day summit permafrost (Figure 4). Nowadays, most of these features are relict, but there are some exceptions at surprisingly low altitudes in the Gaspé Peninsula or in the Laurentian Highlands. Putnam and Putnam (2009) identified two inactive rock glaciers with an ice core in Maine, at an altitude of less than 500 m. However, these authors only recorded the underground temperature for one year; therefore one cannot conclude that the ice core lasted for a sufficient duration to be considered to be permafrost.

While there is no evidence of the sustainability of these ice cores beyond one year under present-day conditions, their resilience to summer melting is still surprising at such low altitudes. Cold talus scree slopes exhibiting a cold ventilation mechanism have been found at elevations close to or even lower than the Early-Holocene permafrost altitudinal limits (Figure 4). Therefore, even if it is impossible to conclude the interannual sustainability of these ice cores, the existence of

sufficient winter frost ground penetration for the onset of a ventilation mechanism in some rock glaciers and scree slopes at low elevations can be concluded, contributing to the thermal inertia of the cores and delaying the melting, at least, to the end of the thawing season. Because of the sporadic distribution of these ventilation systems in northeastern North America, we can conclude that they are located at the boundary of the climatic conditions necessary for their development. Hence, their onset must be controlled by the local topographic, geomorphologic, and climatic conditions.

Given the latitude of the study region, it is unlikely that discontinuous permafrost is frequent at such moderate elevations. However, these cold scree slopes create favourable conditions for the annual conservation of ice, as reported elsewhere in mountainous environments under the discontinuous permafrost limit (e.g., Morard et al. 2010; Gądek 2012). This dynamic testifies to the increased complexity associated with local processes such as the ground thermal regime among the paraglacial coarse-grained sediment stocks. More long-term data could permit insight into whether there is an actual (discontinuous) permafrost regime at mid- to low altitudes in the mountains of northeastern North America.

Treeline position. Altitudinal and latitudinal treelines and ecotones are often considered to be sensitive to temperature, and hence to ongoing climate warming (Harsch et al. 2009). However, the increase in the elevation of the treeline in response to climate change is heterogeneous at the global scale. Harsch et al. (2009) reported an increase in elevation for 52% of the treelines at the global scale and failed to associate the probability of treeline advance with either mean annual or summer temperature increase, the latter being considered to be the main control over treeline position (e.g., Cogbill and White 1991). At the Holocene temporal scale, since the establishment

of alpine tundra at high elevations (ca. 9,000 BP), climatic fluctuations had significant impacts on lower elevation vegetation assemblages, but they had little influence on alpine plant communities (Spear 1989). At a smaller temporal scale, diachronic comparisons of treeline position all yield the same results: the treeline is generally stable but the shrub species cover in the lower alpine tundra increases (Robinson et al. 2010; Capers and Stone 2011; Dumais et al. 2014). However, the few studies undertaken on the climate of mountain ranges in our study region did not find significant trend in temperature or precipitation for high elevation sites (Seidel et al., 2009; Fortin and Hétu 2014). At the regional scale, over the last four decades there has been a notable warming trend in the mean annual temperature in Gaspé (0.11 °C per decade; Yagouti et al. 2006) and in the mean winter temperature in the northeastern United States (0.1 °C per decade; Burakowski et al. 2008). However, one has to exercise extreme caution in the extrapolation of general temperature trends to local alpine environments, especially with the quasi-absence of high altitude weather stations with reliable long-term data in the current study region.

Other environmental conditions must control the treeline elevation resilience to climate warming. Carlson et al. (2011) report that the alpine environment of the Adirondacks exhibits a spatial pattern that corresponds with the prevailing wind direction, providing further evidence that the alpine treeline of northeastern North America is primarily controlled by winds and winter icing events. The treeline of our study region does exhibit the presence of a wind dominated disturbance regime, since almost every tree community located in the subalpine-alpine ecotone grows in krummholz. At the global scale it can be observed that, contrary to the diffuse treelines, the krummholz treelines fail to exhibit a significant relationship between their upward progression (or lack thereof) and temperature warming (Harsch et al. 2009). The krummholz growth form of trees at the treeline forbidding the production of viable seeds, the lack of suitable

germination sites on the mountaintop bedrock and felsenmeers, or the absence of a soil layer have also been mentioned as factors restricting the upward progression of the treeline (Dumais et al. 2014). The presence of permafrost bodies on the concerned summits may also prevent the progression of the treeline to higher elevations. The permafrost system and the treeline system therefore mutually increase the resilience of one another. The permafrost limits the upward progression of the treeline. The absence of trees above the treeline allows for higher wind speeds and greater snow drifting from the summit, which, as was stated earlier, is a key factor in allowing sufficient frost ground penetration for the conservation of the permafrost. The permafrost lying on the summit of these mid-altitude mountains is thus particularly vulnerable and depends on complex interactions between snow cover, vegetation and local climate (Johansson et al. 2013).

Hence, as long as there are rime or ice storms episodes during the winter, the wood abrasion by wind transported ice crystals and the mechanical breakage will prevent the upward progression of the treeline (Rocheport et al. 1994). However, the Northern Hardwood Boreal Ecotone is controlled by temperature, as reported by Beckage et al. (2008) who estimate an upslope shift of 2.1 to 2.8 m yr⁻¹ in the Green Mountains. It is therefore the subalpine spruce-fir belt that is the most vulnerable. The ongoing warming is expected to contribute to the narrowing of the altitude range of the subalpine spruce-fir belt with an upward displacement of the underlying ecotone (Carlson et al. 2011).

*4.3 Holocene climate reconstruction in northeastern North American alpine landscapes –
discrepancies and the need for a multi-proxy approach*

Figure 5 compares different indicators of Late-Glacial and Holocene climatic changes in the alpine ranges of northeastern North America. From this synthesis, several statements can be made:

- 1) Among different ranges across the studied region, there seem to be important differences in the response to global scale Holocene climate fluctuations. There seem to be regional differences between some of the climatic periods across the studied regions. For example, from the vegetation records, there is a lag of approximately 2,000 years for the occurrence of a Mid-Holocene climatic warming between New England and the Gaspé Peninsula.;
- 2) At local scale, there has been no clear evidence of the Younger Dryas episode from the pollen records of the White Mountains (Spear et al. 1994). Indeed, these records suggest an opposite trend, with the upward progress of the treeline due to shifting prevailing winds mentioned above. Other indicators (i.e., % organic matter, chironomids) showed clearer evidence of the Younger Dryas in the White Mountains synchronously with the Atlantic Provinces, but to a lesser extent (Cwynar and Spear 2001). The steep topography in the mountain ranges of northeastern North America could reinforce these climatic variations, supported by the orographic effect. There is, however, a lack of data to distinguish how the impacts of the different climatic fluctuations are related to zonal (e.g., latitude) and azonal (e.g., distance to the ocean, topography) variables;
- 3) On a regional scale, there is also an important difference in the quantity of data available, with some ranges having been studied more thoroughly than others. Alpine environments of

northeastern North America are characterized by non-linear responses due to interactions and feedbacks between geological, geomorphological, ecological and climatic factors. This complexity results in singular environments exhibiting a resilient permafrost state despite being at boundary conditions, a unique arctic-alpine flora on the summits, original geomorphic processes such as the frost-coated clast flows (Hétu et al. 1994), climatic particularities like the extreme winds on Mount Washington and puzzling ecological dynamics like the lowering of the treeline in the Gaspé Peninsula despite the warming trend of the climate. Therefore, a global and thorough understanding of the alpine environment of northeastern North America remains a scientific challenge. In the context of the ongoing climate change, which could result in profound alterations of these systems, it is the opinion of the authors that an integrated and multi-proxy approach (Lotter, 2003) would be necessary to get a valuable prospective view of the evolution of these altitudinal bio-geo-ecosystems.

5. Conclusions

Through different case studies, this synthesis underlined the imbricated interactions of Pre-Holocene history with the Late-Glacial and Holocene evolution of climate, biosystems and geosystems leading to the modern alpine landscapes. Regional ice recession and postglacial vegetation establishment followed a south to north temporal trend, with each mountain range displaying a distinct evolution. However, there is a need for further studies on mountain environments to better understand, among others, the origin and evolution of the felsenmeers on the summits of many ranges discussed in this paper, the impacts and duration of cold-based vs warm-based glaciers on the shaping of the alpine landscapes, and the complexity of this environment at the regional scale. Further studies using innovative datation techniques in this

study region, such as cosmogenic nuclide dating (e.g. Gosse et al. 2006), Schmidt hammer (e.g. Matthews and Wilson, 2015) or optically stimulated luminescence (e.g. Rémillard et al. 2015) could refine the resolution and enhance our understanding of the evolution of the alpine environment of the Northeast. The case studies presented above show how the complex interactions between the different systems underlie the presence of alpine environments at such low elevations, their different responses in terms of landscape evolution and of their resilience to climate change.

The synergistic interactions between the different systems demonstrate the need for an interdisciplinary approach in order to better understand how this unique environment will evolve in the coming decades. Capers et al. (2013) recognized this need by doing a survey of scientists from diverse disciplinary fields to establish alpine research priorities in the Northeast, but the bias towards ecology is still apparent in the proposed research projects. The paraglacial concept seems to be a promising framework within which to achieve this interdisciplinary need, since it considers the landscape as a complex system that integrates many feedbacks at different temporal scales (Mercier 2008). There is a need to broaden its definition further by considering ecological dynamics not only as variables influencing the sedimentary budget, but as an inherent subsystem of the paraglacial environment, and by recognizing the overlapping spatial scales with regards to the temporal duration of the paraglacial period and to its dominant processes. Finally, the synthesis of the climate proxies presented in Figure 5 shows that alpine ranges responded differently to major climatic changes and highlights the relevance of a multi-proxy approach in reconstructing past alpine environments. Some hypotheses were proposed to explain the discrepancies between the different records. The fact that the studies were only conducted locally and many decades apart might be the primary cause of these inconsistencies. The complex

systems of alpine environment of the Northeast are characterized by their nonlinear responses. As an example, in some fine-grained scree slopes of the Gaspé Peninsula, due to an increased sediment flux related to frost-coated clast flows over the last decades, the altitude of the treeline is decreasing since the middle of the twentieth century even considering the ongoing climate warming favourable to the growth of the tree cover. In this regard, there is a need for broader-scale multi-proxy research since the uniqueness of the alpine landscapes of northeastern North America holds keys to better understanding past and upcoming environmental changes (Germain and Hétu, *in press*).

Acknowledgements

This research was financially supported by the Fonds québécois de Recherche sur la Nature et les Technologies through a scholarship granted to J.P. Martin. The authors would like to thank Gabriel Magnan, Pierre Richard, Patrick Lajeunesse and especially Michelle Garneau for their thoughtful insights that contributed to this review, as well as André Parent who assisted in the design of the figures.

References

Anderson, R.S. and Davis, R.B. 1986. History of late- and post-glacial vegetation and disturbance around Upper South Branch Pond, northern Maine. *Canadian Journal of Botany*, **64**: 1977-1986.
doi: 10.1139/b86-262

- Anderson, T.W., Levac, E. and Lewis, M.C.F. 2007. Cooling in the Gulf of St. Lawrence and estuary region at 9.7 to 7.2 14C ka (11.2–8.0 cal ka): Palynological response to the PBO and 8.2 cal ka cold events, Laurentide Ice Sheet air-mass circulation and enhanced freshwater runoff. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **246**: 75-100. doi: 10.1016/j.palaeo.2006.10.028
- Asnong, H. and Richard, P.J.H. 2003. La végétation et le climat postglaciaires du centre et de l'est de la Gaspésie, au Québec. *Géographie physique et Quaternaire*, **57**(1): 37-63. doi: 10.7202/010330ar
- Bailey, S.W., Hoy, J. and Cogbill, C.V. 2015. Vascular flora and geocology of Mont de la Table, Gaspésie, Québec. *Rhodora*, **117**(969): 1-40. doi: 10.3119/14-07
- Ballantyne, C.K. 2002. Paraglacial geomorphology. *Quaternary Science Reviews*, **21**: 1935–2017. doi: 10.1016/S0277-3791(02)00005-7
- Beckage, B., Osborne, B., Gavin, D.G., Pucko, C., Siccama, T. and Perkins, T. 2008 A rapid upward shift of a forest ecotone during 40 years of warming in the Green Mountains of Vermont. *Proceedings of the National Academy of Science*, **105**(11): 4197-4202. doi: 10.1073/pnas.0708921105
- Birks, H.H. and Birks, H.J.B. 2000. Future uses of pollen analysis must include plant macrofossils. *Journal of Biogeography*, **27**(1): 31-35. doi: 10.1046/j.1365-2699.2000.00375.x

- 670 Bishop, P. 2007. Long-term landscape evolution: linking tectonics and surface processes. *Earth*
671 *Surface Processes and Landforms*, **32**; 329-365 doi: 10.1002/esp.1493
672
- 673 Bogucki, D.J. 1977. Debris slide hazard in the Adirondack Province of New York State.
674 *Environmental Geology*, **1**: 317-328. doi: 10.1007/BF02380500
675
- 676 Bradley, D.C. 1981. Late Wisconsinan Mountain Glaciation in the Northern Presidential Range,
677 New Hampshire. *Arctic and Alpine Research*, **13**(3): 319-327. doi: 10.2307/1551038
678
- 679 Burakowski, E.A., Wake, C.P., Braswell, B., Brown, D.P. 2008. Trends in wintertime climate in
680 the northeastern United States: 1965-2005. *Journal of Geophysical Research*, **113**(D20). doi:
681 10.1029/2008JD009870
682
- 683 Bussieres, B., Payette, S. and Filion, L. 1996. Deboisement et entourage des hauts sommets
684 de Charlevoix l'Holocene suprieur: origine des tages alpin et subalpin. *Gographie physique*
685 *et Quaternaire*, **50**(3): 257-269. doi: 10.7202/033099ar
686
- 687 Camill, P. and Clark, J.S. 2000. Long-Term Perspectives on Lagged Ecosystem Responses to
688 Climate Change : Permafrost in Boreal Peatlands and the Grassland/Woodland Boundary.
689 *Ecosystems*, **3**(6): 534-544. doi: 10.1007/s100210000047
690
- 691 Capers, R.S., Kimball, K.D., McFarland, K.P., Jones, M.T., Lloyd, A.H., Munroe, J.S., Fortin,
692 G., Mattrick, C., Goren, J., Sperduto, D.D. and Paradis, R. 2013. Establishing alpine research

- 693 priorities in Northeastern North America. *Northeastern Naturalist*, **20**(4): 559-577. doi:
694 10.1656/045.020.0406
695
- 696 Capers, R.S. and Stone, A.D. 2011 After 33 years, trees more frequent and shrubs more abundant
697 in Northeast U.S. alpine community. *Arctic, Antarctic and Alpine Research*, **43**(4): 495-502. doi:
698 10.1657/1938-4246-43.4.495
699
- 700 Carlson, B.Z., Munroe, J.S. and Hegman, B. 2011 Distribution of Alpine Tundra in the
701 Adirondack Mountains of New York, U.S.A. *Arctic, Antarctic and Alpine Research*, **43**(3): 331-
702 342. doi: 10.1657/1938-4246-43.3.331
703
- 704 Clark, G.M. and Schmidlin, T.W. 1992. Alpine periglacial landforms of eastern North America:
705 A review. *Permafrost and Periglacial Processes*, **3**(3): 225-230. doi: 10.1002/ppp.3430030309
706
- 707 Cogbill, C.V. 1985. Dynamics of boreal forest of the Laurentian Highlands, Canada. *Canadian*
708 *Journal of Forest Research*, **15**: 252-261. doi: 10.1139/x85-043
709
- 710 Cogbill, C.V. and White, P.S. 1991. The Latitude-Elevation Relationship for Spruce-Fir Forest
711 and Treeline along the Appalachian Mountain Chain. *Vegetatio*, **94**(2): 153-175. doi:
712 10.1007/BF00032629
713
- 714 Cwynar, L.C. and Spear, R.W. 2001. Lateglacial climate change in the White Mountains of New
715 Hampshire. *Quaternary Science Reviews*, **20**: 1265-1274. doi: 10.1016/S0277-3791(00)00151-7
716

- 717 Davis, P.T. and Davis, R.B. 1980. Interpretation of minimum-limiting radiocarbon dates for
718 deglaciation of Mount Katahdin area, Maine. *Geology*, **8**(4): 396-400. doi: 10.1130/0091-7613
719
- 720 Dullinger, S., Gatttringer, A., Thuiller, W., Moser, D., Zimmermann, N.E., Guisan, A., Willner,
721 W., Plutzer, C., Leitner, M., Mang, T., Caccianiga, M., Dirnböck, T., Ertl, S., Fischer, A., Lenoir,
722 J., Svenning, J.C., Psomas, A., Schmartz, D.R., Silc, U., Vittoz, P. and Hübner, K. 2012.
723 Extinction debt of high-mountain plants under twenty-first century climate change. *Nature*
724 *Climate Change*, **2**: 619-622. doi: 10.1038/nclimate1514
725
- 726 Dumais, C., Ropars, P., Denis, M.P., Dufour-Tremblay, G. and Boudreau, S. 2014. Are low
727 altitude alpine tundra ecosystems under threat ? A case study from the Parc National de la
728 Gaspésie, Québec. *Environmental Research Letters*, **9**: 1-10. doi: 10.1088/1748-9326/9/9/094001
729
- 730 Dyke, A.S., Andrews, J.T., Clark, P.U., England, J.H., Miller, G.H., Shaw, J. and Veillette, J.J.
731 2002. The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary*
732 *Science Reviews*, **21**: 9-31. doi: 10.1016/S0277-3791(01)00095-6
733
- 734 Dyke, A.S. 2004. An outline of North American Deglaciation with an emphasis on central and
735 northern Canada. *In* *Quaternary Glaciations – Extent and Chronology, Part II: North America*.
736 *Edited by J. Ehlers and P.L. Gibbard. Developments in Quaternary Science*, **2b**. Elsevier,
737 Amsterdam, Netherlands. pp. 373-424.
738
- 739 Engler, R., Randin, C.F., Thuiller, W., Dullinger, S., Zimmermann, N., Araujo, M.B., Pearman,
740 P.B., Le Lay, G., Piedallu, C., Albert, C.H., Choler, P., Coldea, G., De Lamo, X., Dirnböck, T.,

- 741 Gégout, J.C., Gomez-Garcia, D., Grytnes, J.A., Heegard, E., Høistad, F., Nogués-Bravo, D.,
742 Normand, S., Puşcaş, M., Sebastià, M.T., Stanisci, A., Theurillat, J.P., Trivedi, M.R., Vittoz, P.
743 and Guisan, A. 2011. 21st century climate change threatens mountain flora unequally across
744 Europe. *Global Change Biology*, **17**(7): 2330-2341. doi: 10.1111/j.1365-2486.2010.02393.x
745
- 746 Enos, P.C. 1969. Cloridorme Formation, middle Ordovician Flysch, Northern Gaspé Peninsula,
747 Québec. *Geological Society of America Special Paper*, **117**. doi: 10.1130/SPE117-p1
748
- 749 Fernald, M.L. 1907. The soil preferences of certain alpine and subalpine plants. *Rhodora Journal*
750 *of the New England Botanical Club*, **9**(105): 149-193.
751
- 752 Fernald, M.L. 1925. Persistence of plants in unglaciated areas of Boreal America. *Memoirs of the*
753 *American Academy of Arts and Sciences*, **15**: 558-572. doi: 10.2307/25058128
754
- 755 Filion, L. 1987. Holocene Development of Parabolic Dunes in the Central St. Lawrence Lowland,
756 Québec. *Quaternary Research*, **28**: 196-209. doi: 10.1016/0033-5894(87)90059-7
757
- 758 Fortin, G. and Hétu, B. 2014. Estimating winter trends in climatic variables in the Chic-Chocs
759 Mountains, Canada (1970-2009). *International Journal of Climatology*, **34**: 3078-3088. doi:
760 10.1002/joc.3895
761
- 762 Gądek, B. 2012. Debris slopes ventilation in the periglacial zone of the Tatra Mountains (Poland
763 and Slovakia) : The indicators. *Cold Regions Science and Technology*, **74-75**: 1-10. doi:
764 10.1016/j.coldregions.2012.01.007

- 765
- 766 Germain, D. 2005. Dynamique des avalanches de neige en Gaspésie, Québec, Canada. Ph.D.
- 767 thesis, Department of Geography, Laval University, Québec, Canada.
- 768
- 769 Germain, D. and Héту, B. (*in press*) Hillslope processes and related sediment fluxes on a fine-
- 770 grained scree slope of Eastern Canada. *In* Source-to-sink fluxes in Undisturbed Cold
- 771 Environments. *Edited by* A.A. Beylich, J.C. Dixon and Z. Zwolinski. Cambridge University
- 772 Press, Cambridge, UK.
- 773
- 774 Germain, D., Héту, B. and Filion, L. 2010. Tree-ring based reconstruction of past snow avalanche
- 775 events and risk assessment in Northern Gaspé Peninsula (Québec, Canada). *In* Tree rings and
- 776 natural hazards. *Edited by* M. Stoffel, M. Bollschweiler, D.R. Butler and B.H. Luckman.
- 777 Springer, Dordrecht, Netherlands. pp. 51-73.
- 778
- 779 Gordon, G. 1989. The Worst Weather in the World. *Windswept*, **30**: 66-69.
- 780
- 781 Gosse, J.C., Bell, T., Gray, J.T., Klein, J., Yang, G. and Finkel, R. 2006. Using cosmogenic
- 782 isotopes to interpret the landscape record of glaciation : Nunataks on Newfoundland ? *In* Glacier
- 783 science and environmental change. *Edited by* P.J. Knight. Blackwell, Oxford, UK. pp. 442-446.
- 784
- 785 Grabherr, G., Gottfried, M. and Pauli, H. 2010. Climate change impacts in alpine environments.
- 786 *Geography Compass*, **4**(8): 1133-1153. doi: 10.1111/j.1749-8198.2010.00356.x
- 787

- 788 Gray, J.T. and Brown, R.J.E. 1979. Permafrost presence and distribution in the Chic-Chocs
789 Mountains, Gaspésie, Québec. *Géographie physique et Quaternaire*, **33**(3-4): 299-316. doi:
790 10.7202/1000366ar
791
- 792 Gray, J.T., Godin, É., Masse, J. and Fortier, D. 2009. Trois décennies d'observation des
793 fluctuations du régime thermique du pergélisol dans le parc national de la Gaspésie. *Le*
794 *naturaliste canadien*, **133**(3): 69-77.
795
- 796 Hargraves, R.B., Roy, D.W. 1974. Paleomagnetism of Anorthosite in and around the Charlevoix
797 Cryptoexplosion Structure, Quebec. *Canadian Journal of Earth Sciences*, **11**: 854-859. doi:
798 10.1139/e74-085
799
- 800 Harris, C., Arenson, L.U., Christiansen, H.H., Etzelmüller, B., Frauenfelder, R., Gruber, S.,
801 Haeberli, W., Hauck, C., Hözle, M., Humlum, O., Isaksen, K., Kääb, A., Kern-Lütschg, M.A.,
802 Lehning, M., Matsuoka, N., Murton, J.B., Nötzli, J., Phillips, M., Ross, N., Seppälä, M.,
803 Springman, S.M. and Mühll, D.V. 2009. Permafrost and climate in Europe: Monitoring and
804 modelling thermal, geomorphological and geotechnical responses. *Earth-Science Reviews*, **92**;
805 117-171. doi: 10.1016/j.earscirev.2008.12.002
806
- 807 Harsch, M.A., Hulme, P.E., McGlone, M.S., Duncan, R.P. 2009. Are treelines advancing ? A
808 global meta-analysis of treeline response to climate warming. *Ecology Letters*, **12**: 1040-1049.
809 doi: 10.1111/j.1461-0248.2009.01355.x
810

- 811 Hétu, B. and Bail, P. 1996. Évolution postglaciaire du regime hydrosédimentaire et vitesse de
812 l'ablation dans un petit bassin-versant des Appalaches près de Rimouski (Bas-Saint-Laurent,
813 Québec). *Géographie physique et Quaternaire*, **50**(3): 351-363.
- 814
- 815 Hétu, B. and Gray, J.T. 2000. Effects of environmental change on scree slope development
816 throughout the postglacial period in the Chic-Choc Mountains in the northern Gaspé Peninsula,
817 Quebec. *Geomorphology*, **32**: 335-355. doi: 10.1016/S0169-555X(99)00103-8
- 818
- 819 Hétu, B., Van Steijn, H. and Vandelac, P. 1994. Les coulées de pierres glacées: un nouveau type
820 de coulées de pierraille sur les talus d'éboulis. *Géographie physique et Quaternaire*, **48**:3-22. doi:
821 10.7202/032969ar
- 822
- 823 Hiscott, R.N. and James, N.P. 1985. Carbonate debris flows, Cow Head Groupe, western
824 Newfoundland. *Journal of Sedimentary Research*, **55**(5): 735-745.
- 825
- 826 Humlum, O. 1998. The climatic significance of rock glaciers. *Permafrost and Periglacial*
827 *Processes*, **9**(4): 375-395. doi: 10.1002/(SICI)1099-1530
- 828
- 829 Ibe, R.A. 1985. Postglacial Montane Vegetational History Around Balsam Lake, Catskill
830 Mountains, New York. *Bulletin of the Torrey Botanical Club*, **112**(2): 176-186. doi:
831 10.2307/2996414
- 832
- 833 Jackson, S.T. and Whitehead, D.R. 1991. Holocene vegetation patterns in the Adirondack
834 Mountains. *Ecology*, **72**(2): 641-653. doi: 10.2307/2937204

- 835
- 836 Johansson, M., Callaghan, T.V., Rosiö, J. and Akerman, H.J. 2013. Rapid responses of
837 permafrost and vegetation to experimentally increased snow cover in sub-arctic Sweden.
838 Environmental Research Letter, **8**(3): 035025. doi: 10.1088/1748-9326/8/3/035025
- 839
- 840 Johnson, C.J., Ehlers, L.P.W. and Seip, D.R. 2015. Witnessing extinction – Cumulative impacts
841 across landscapes and the future loss of an evolutionarily significant unit of woodland caribou in
842 Canada. Biological Conservation, **186**: 176-186. doi: 10.1016/j.biocon.2015.03.012
- 843
- 844 Jomelli, V., Khodri, M., Favier, V., Brunstein, D., Ledru, M.P., Wagnon, P., Blard, P.H., Sicart,
845 J.E., Braucher, R., Grancher, D., Bourlès, D.L., Braconnot, P. and Vuille, M. 2011. Irregular
846 tropical glacier retreat over the Holocene epoch driven by progressive warming. Nature, **474**:
847 196-199. doi: 10.1038/nature10150
- 848
- 849 Jones, M.T. and Willey, L.L. 2012. The Eastern Alpine Guide. Beyond Ktaadn and Boghaunter
850 Books, New Salem, Massachusetts.
- 851
- 852 Jutras, P. and Schroeder, J. 1999. Geomorphology of an exhumed Carboniferous paleosurface in
853 the southern Gaspé Peninsula, Québec; paleoenvironmental and tectonic implications.
854 Géographie physique et Quaternaire, **53**(2): 249-263. doi: 10.7202/005690ar
- 855
- 856 Körner, C. 1998 A re-assessment of high elevation treeline positions and their explanation.
857 Oecologia, **115**: 445-459. doi: 10.1007/s004420050540
- 858

- 859 Kull, C.A. and Magilligan, F.J. 1994. Controls over landslide distribution in the White
860 Mountains, New Hampshire. *Physical Geography*, **15**(4): 325-341. doi:
861 10.1080/02723646.1994.10642520
862
- 863 Labelle, C. and Richard, P.J.H. 1984. Histoire Postglaciaire de la Végétation dans la Région de
864 Mont-Saint-Pierre, Gaspésie, Québec. *Géographie physique et Quaternaire*, **38**(3): 257-274. doi:
865 10.7202/032567ar
866
- 867 Lafortune, M., Fillion, L. and Hétu, B. 1997. Dynamique d'un front forestier sur un talus
868 d'éboulis actif en climat tempéré froid (Gaspésie, Québec). *Géographie physique et Quaternaire*,
869 **51**(1): 1-15. doi: 10.7202/004840ar
870
- 871 Lavoie, M. and Richard, P.J.H. 2000. Paléoécologie de la tourbière du Lac Malbaie, dans le
872 Massif des Laurentides (Québec): Évaluation du rôle du climat sur l'accumulation de la tourbe.
873 *Géographie physique et Quaternaire*, **54**(2): 169-185. doi: 10.7202/004843ar
874
- 875 Lebus, J. and David, P.P. 1977. La stratigraphie et les événements du Quaternaire de la partie
876 occidentale de la Gaspésie, Québec. *Géographie physique et Quaternaire*, **31**(3-4): 275-296. doi:
877 10.7202/1000278ar
878
- 879 Lotter, A.F. 2003. Multi-proxy climatic reconstructions. *In* *Global change in the Holocene*.
880 *Edited by* A. Mackay, R. Battarbee, J. Birks and F. Oldfield. Arnold, London, UK. pp. 373-383.
881

- 882 Löve, D. 1970. Subarctic and Subalpine: Where and What? *Arctic and Alpine Research*, **2**(1): 63-
883 73. doi: 10.2307/1550141
884
- 885 Matthews, J.A. and Wilson, P. 2015. Improved Schmidt-hammer exposure ages for active and
886 relict pronival ramparts in southern Norway, and their palaeoenvironmental implications.
887 *Geomorphology*, **246**: 7-21. doi: 10.1016/j.geomorph.2015.06.002
888
- 889 Mercier, D. 2008. Paraglacial and paraperiglacial landsystems : concepts, temporal scales and
890 spatial distribution. *Géomorphologie : relief, processus, environnement*, **4**: 223-234. doi:
891 10.4000/geomorphologie.7396
892
- 893 Messerli, B. and Ives, J.D. 1997. *Mountains of the world: A global priority*. Parthenon Publishing
894 Group, UK.
895
- 896 Morard, S., Delaloye, R. and Lambiel, C. 2010. Pluriannual thermal behaviour of low elevation
897 cold talus slopes (western Switzerland). *Geographica Helvetica*, **65**(2): 124-134. doi: 10.5194/gh-
898 65-124-2010
899
- 900 Olejczyk, P. and Gray, J.T. 2007. The relative influence of Laurentide and local ice sheets during
901 the last glacial maximum in the eastern Chic-Chocs Range, northern Gaspé Peninsula, Quebec.
902 *Canadian Journal of Earth Sciences*, **44**: 1603-1625. doi: 10.1139/e07-039
903
- 904 Price, M., Jansky, L. and Iatsenia, A.A. 2004. *Key issues for Mountain Areas*. United Nations
905 University press, Tokyo.

- 906
- 907 Putnam, A.E. and Putnam, D.E. 2009. Inactive and relict rock glaciers of Deboullie Lakes
908 Ecological Reserve, northern Maine, USA. *Journal of Quaternary Science*, **24**(7): 773-784. doi:
909 10.1002/jqs.1252
- 910
- 911 Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey,
912 C., Buck, C.E., Burr, G.S., Edwards, R.L., Friedrich, M., Grootes, P.M., Guilderson, T., Hajdas,
913 I., Heaton, T., Hogg, A., Hughen, K., Kaiser, K., Kromer, B., McCormac, F., Manning, S.,
914 Reimer, R., Richards, D., Southon, J., Talamo, S., Turney, C., van der Plicht, J. and
915 Weyhenmeyer, C. 2009. INTCAL09 and MARINE09 Radiocarbon age calibration curves, 0-
916 50,000 years cal BP. *Radiocarbon*, **51**: 1111-1150.
- 917
- 918 Rémillard, A.M., Hétu, B., Bernatchez, P., Buylaert, J.P., Murray, A.S., St-Onge, G., Geach, M.
919 2015. Chronology and palaeoenvironmental implications of the ice-wedge pseudomorphs and
920 composite-wedge casts on the Magdalen Islands. *Boreas*, **44**(4): 658-675. doi: 10.1111/bor.12125
- 921
- 922 Richard, P.J.H. 2009. Chronologie de la déglaciation: de l'importance des années étalonnées
923 (calibrées). Association Québécoise pour l'étude du Quaternaire. Available from:
924 <http://www.er.uqam.ca/nobel/aquual/articles/ChronoDeglaciation.pdf>
- 925
- 926 Richard, P.J.H., Veillette, J.J., Larouche, A.C., Hétu, B., Gray, J.T. and Gangloff, P. 1997.
927 Chronologie de la déglaciation en Gaspésie: nouvelles données et implications. *Géographie*
928 *physique et Quaternaire*, **51**(2): 163-184. doi: 10.7202/033116ar
- 929

- 930 Robinson, S.C., Ketchledge, E.H., Fitzgerald, B.T., Raynald, D.J. and Kimmerer, R.W. 2010. A
931 23-year assessment of vegetation composition and change in the Adirondack alpine zone, New
932 York State. *Rhodora*, **112**: 355-377. doi: 10.3119/09-03.1
933
- 934 Rochefort, R.M., Little, R.L., Woodward, A. and Peterson, D.L. 1994. Changes in sub-alpine tree
935 distribution in western North America: a review of climatic and other causal factors. *The*
936 *Holocene* **4**(1): 89-100. doi: 10.1177/095968369400400112
937
- 938 Sagoff, M. 2002. On the Value of natural Ecosystems: The Catskills Parable. *Politics and the Life*
939 *Sciences*, **21**(1): 19-25.
940
- 941 Seidel, T.M., Weihrauch, D., Kimball, K.D., Pszenny, A., Soboleskil, R., Crete, E. and Murray,
942 G. 2009. Evidence of climate change declines with elevation based on temperature and snow
943 records from 1930s to 2006 on Mount Washington, New Hampshire, USA. *Arctic, Antarctic and*
944 *Alpine Research*, **41**: 362-372. doi: 10.1657/1938-4246-41.3.362
945
- 946 Shur, Y.L. and Jorgenson, M.T. 2007. Patterns of permafrost formation and degradation in
947 relation to climate and ecosystems. *Permafrost and Periglacial Processes*, **17**: 7-19. doi:
948 10.1002/ppp.582
949
- 950 Sirois, L. 1984. Le plateau du mont Albert: étude phyto-écologique. Master thesis, Biology
951 Department, Laval University, Québec, Canada.
952

- 953 Spear, R.W. 1989. Late-Quaternary History of High-Elevation Vegetation in the White
954 Mountains of New Hampshire. *Ecological Monographs*, **59**(2): 125-151. doi: 10.2307/2937283
955
- 956 Spear, R.W., Davis, M.B. and Shane, L.C.K. 1994. Late Quaternary History of Low- and Mid-
957 Elevation Vegetation in the White Mountains of New Hampshire. *Ecological Monographs*, **64**(1):
958 85-109. doi: 10.2307/2937056
959
- 960 Sprugel, D.G. 1976. Dynamic structure of wave-regenerated *Abies balsamea* forests in the
961 northeastern United States. *Journal of Ecology*, **64**: 889-905. doi: 10.2307/2258815
962
- 963 Thorson, R.M. and Schile, C.A. 1995. Deglacial eolian regimes in New England. *Geological*
964 *Society of America Bulletin*, **107**(7): 751-761. doi: 10.1130/0016-7606(1995)
965
- 966 Titus, R. 2004. *The Catskills: A Geological Guide*, 3rd Edition. Purple Mountain Press,
967 Fleischmann, New York.
968
- 969 Wahl, K., Spooner, I. and Colville, D. 2007. Thin-Skinned Debris Flows in Cape Breton
970 Highlands National Park, Nova Scotia, Canada. *Atlantic Geology*, **43**: 45-56. doi:
971 10.4138/4213
972
- 973 Walegur, M.T. and Nelson, F.E. 2003. Permafrost distribution in the Appalachian Highlands,
974 northeastern USA. *In Permafrost. Edited by Philips, Springman and Anderson*. Liss, Swets and
975 Zeitlinger, pp. 1201-1206.
976

Whipple, K.X. 2009. The influence of climate on the tectonic evolution of mountain belts. *Nature geoscience*, **2**: 97-104. doi: 10.1038/ngeo413

Winkworth, R.C., Wagstaff, S.J., Glenney, D. and Lockhart, P.J. 2005. Evolution of the New Zealand mountain flora: Origins, diversification and dispersal. *Organisms, Diversity & Evolution*, **5**: 237-247. doi: 10.1016/j.ode.2004.12.001

Yagouti, A., Boulet, G. and Vescovi, L. 2006. Projet-MNEV-7 : Homogénéisation des séries de températures et analyse de la variabilité spatio-temporelle de ces séries au Québec méridional. Rapport #4 – Homogénéisation des séries de températures du Québec méridional et analyse de l'évolution du climat à l'aide d'indicateurs. Consortium Ouranos, Québec, Canada.

Zimmermann, C. and Lavoie, C. 2001. A paleoecological analysis of a southern permafrost peatland, Charlevoix, Quebec. *Canadian Journal of Earth Sciences*, **38**: 909-919. doi:

10.1139/e00-110

Range Name	Surface area (km ²)	Maximal elevation (m)	Area of alpine environment (km ²)	Proportion of alpine environment (%)	Minimal elevation of alpine environment (m)
Adirondacks (High Peaks)	1200	1629	0.7	0.06	1350
White Mountains	3500 (e)	1917	11.3	0.32	1330 (e)
Mount Katahdin	809	1606	7.3	0.90	1200 (e)
Charlevoix Highlands	1000 (e)	1180	2.25 (e)	0.23	700 (e)
Chic-Chocs	950 (e)	1261	45 (e)	4.97	900 (e)
Long Range	4100 (e)	812	>500 (e)	12.19	200 (e)

Table 1: Physiographic characteristics of different alpine ranges of northeastern North America across a SW (Adirondacks) to NE (Long Range) gradient. Data with an (e) represent rough estimates made by the author using satellite imagery (Google Earth).

Range	Location	Latitude (°)	Dated material	Sources
Catskills	Balsam Lake (762 m a.s.l.)	42.1	Inferred from age- depth curve	Ibe, 1985
Adirondacks	Upper Wallface Pond (948 m a.s.l.)	44.1	Pond sediment	Jackson and Whitehead, 1991
White Mountains	Deer Lake Bog (1,325 m a.s.l.)	44.3	Pond sediment	Spear, 1989
Mount Katahdin	Lower Tongue Pond (183 m a.s.l.)	45.9	Lake sediment	Davis and Davis, 1980
Charlevoix Highlands	Lac Malbaie (800 m a.s.l.)	47.8	Peat	Lavoie and Richard, 2000
Chic-Chocs high plateaux	Lac du Diable (494 m a.s.l.)	48.9	Gyttja	Richard et al, 1997
	Lac Côté (915 m a.s.l.)		Organic matter above sand and silt	Lebuis and David, 1977
	Lac Bouliane (992 m a.s.l.)		Gyttja	Richard et al., 1997
Chic-Chocs low plateaux	Lac du Triangle (465 m a.s.l.)	49.2	<i>Dryas integrifolia</i> and <i>Salix</i> <i>herbaceae</i>	Asnong and Richard, 2003
	Lac Turcotte (457 m a.s.l.)		Marl interbedded with gyttja	Labelle and Richard, 1984
Chic-Chocs coastal valleys	60 m a.s.l.	49.2	<i>Hiatella arctica</i>	Hétu and Gray, 2000
Long Range	NA	49.6	NA	Dyke, 2004

Table 2 : Sites used in the deglaciation reconstruction.

Range	Location	Altitude (m a.s.l.)	Latitude (°)	Sources
Catskills	Balsam Lake	762	42.1	Ibe, 1985
Adirondacks	Heart Lake	661	44.1	Jackson and Whitehead, 1991
	Livingstone Pond	850		
	Upper Wallface Pond	948		
	Elk Pass Bog	1,024		
	Lake Arnold	1,150		
	Lake Tear of the Clouds	1,320		
White Mountains	Mirror Lake	213	44.3	Spear et al. 1994
	Lost Pond	625		
	Little East Pond	793		
	Lonesome Lake	831		
	Carter Notch Pond	1,004		
	Kinsman Pond	1,140		
	Eagle Lake Bog	1,275		
	Deer Lake Bog	1,325		
	Lake of the Clouds	1,542		
Mount Katahdin	Upper South Branch Pond	300	45.9	Anderson et al. 1986
Charlevoix Highlands	Lac Malbaie	800	47.8	Lavoie and Richard, 2000
	Lac à l'Empêche	950		Bussi�res et al. 1996
	Lac des Cygnes Mountain	960		Zimmerman and Lavoie, 2000
Chic-Chocs high plateaux	Lac du Diable	494	48.9	Richard et al, 1997
Chic-Chocs low plateaux	Lac Turcotte	457	49.2	Labelle and Richard, 1984
Chic-Chocs coastal valleys	Lac � L�onard	17	49.2	Labelle and Richard, 1984

Table 3: Sites used in the paleoecological reconstructions.

Figure captions

Figure 1: Worldwide altitudinal treeline elevation, showing the unique character of northeastern North America. White dots present data from ranges discussed in this article. From South to North (left to right on the figure): Adirondacks (New York), Green Mountains (Vermont), White Mountains (New Hampshire), Mount Katahdin (Maine), Charlevoix Highlands (Quebec), Chic-Chocs (Quebec), Long Range (Newfoundland). Worldwide data from Körner (1998).

Figure 2: Major alpine environments of northeastern North America used in this study. 1) Catskills; 2) Adirondack Mountains; 3) Green Mountains; 4) White Mountains; 5) Mount Katahdin; 6) Cape Breton Highlands; 7) Charlevoix Highlands; 8) Chic-Chocs and McGerrigle Mountains; and 9) Long Range.

Figure 3: Time of establishment of the vegetation stages on different ranges of the northeastern North America. The white line represents the tundra stage (unavailable for the Chic-Chocs coastal plateaux), the gray line represents the afforestation stage and the black line represents the closed-forest stage (unavailable for the Catskills). The top two are from Gaspé Peninsula lowland. The gray boxes represent Late-Glacial and Early Holocene cold episodes (from left to right: Younger Dryas, Preboreal Oscillation and 8,200 BP cooling period). The black trendline represents minimum glacial recession age, with error bars related to either different data sources or from radiocarbon calibration.

Figure 4: Modern and paleo-permafrost elevation following an elevation-latitude gradient. The dashed lines represent the modern elevation of permafrost (upper dashed line) and of cold talus

1059 scree slopes (lower dashed line). The solid line represents the Early-Holocene elevation of
1060 permafrost, as inferred from rock glacier elevations. Even today, the ventilation system of the
1061 cold scree slopes provides them with sufficient thermic resilience to exist at similar altitudes as
1062 the Early-Holocene rock glaciers (Modified from Putnam and Putnam, 2009).

1063
1064 Figure 5: Climatic indicators of the different alpine regions of northeastern North America
1065 Straight arrows represent indicators of colder episodes. Waved arrows represent indicators of
1066 warmer episodes. PBO: Pre-Boreal Oscillation; MWP: Medieval Warm Period; LIA: Little Ice
1067 Age

Draft

Figure 1: Worldwide altitudinal treeline elevation, showing the unique character of northeastern North America. White dots present data from ranges discussed in this article. From South to North (left to right on the figure): Adirondacks (New York), Green Mountains (Vermont), White Mountains (New Hampshire), Mount Katahdin (Maine), Charlevoix Highlands (Quebec), Chic-Chocs (Quebec), Long Range (Newfoundland). Worldwide data from Körner (1998).

Figure 2: Major alpine environments of northeastern North America. 1) Catskills; 2) Adirondack Mountains; 3) Green Mountains; 4) White Mountains; 5) Mount Katahdin; 6) Cape Breton Highlands; 7) Charlevoix Highlands; 8) Chic-Chocs and McGerrigle Mountains; and 9) Long Range.

Figure 3: Time of establishment of the vegetation stages on different ranges of the northeastern North America. The white line represents the tundra stage (unavailable for the Chic-Chocs coastal plateaux), the gray line represents the afforestation stage and the black line represents the closed-forest stage (unavailable for the Catskills). The top two are from Gaspé Peninsula lowland. The gray boxes represent Late-Glacial and Early Holocene cold episodes (from left to right: Younger Dryas, Preboreal Oscillation and 8,200 BP cooling period). The black trendline represents minimum glacial recession age, with error bars related to either different data sources or from radiocarbon calibration.

Figure 4: Modern and paleo-permafrost elevation following an elevation-latitude gradient. The dashed lines represent the modern elevation of permafrost (upper dashed line) and of cold talus scree slopes (lower dashed line). The solid line represents the Early-Holocene elevation of permafrost, as inferred from rock glacier elevations. Even today, the ventilation system of the cold scree slopes provides them with sufficient thermic resilience to exist at similar altitudes as the Early-Holocene rock glaciers (Modified from Putnam and Putnam, 2009).

Range	Older Dryas	Younger Dryas	PBO / 8200 BP cooling episodes	Climatic Optimum	Neoglacial	MWP	LIA / Present
Chic-Chocs	14.5-13.8 → Moraines, fluvio-glacial deltas	12.9-11.4 → Moraines, fluvio-glacial deltas	11.1-8.1 → - Glacial cirques - Rock glaciers - Slushflow fans - <i>Athys crispa</i> expansion - Variation in herbaceous vegetation	7.8-5.7 ↔ - <i>Pinus strobus</i> maximum - Podzols at high elevations - ↗ solifluxion at high elevations - Stabilisation of debris fans	1.4 - 0.9 → - Snow avalanches with geomorphic impacts - Peat accumulation in avalanche runout zones		0.15 - 0 ↔ - ↗ vegetation elevation on scree slopes
Charlevoix		12.9-11.4 → St. Narcisse moraine		6.6-4.5 ↔ Peat accumulation	3.2 - 0.0 → - High fire frequency - Vegetation low regeneration rate - Peat accumulation on summits		0.5 - 0.0 → Inception of permafrost on summits
New England	14.5-13.8 → Bethleem moraine	12.9-11.4 → - ↘ % organic matter in peat - ↘ loss on ignition - ↗ treeline altitude	10.1 - 5.7 ↔ <i>Pinus strobus</i> altitudinal maximum	5.7 - 3.2 ↔ ↘ treeline elevation to present-day altitude			

Figure 5: Climatic indicators of the different alpine regions of northeastern North America. Straight arrows represent indicators of colder episodes. Waved arrows represent indicators of warmer episodes. PBO: Pre-Boreal Oscillation; MWP: Medieval Warm Period; LIA: Little Ice Age.