VULNERABILITY OF GLACIERS AND PERMAFROST

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Introduction

The cryosphere responds to and amplifies climate variability through positive feedbacks such as the ice–albedo feedback and greenhouse gas release from melting permafrost. At high northern latitudes the effects of recent warming are manifested by rapid wastage of glaciers and ice caps, and a startling decrease in the thickness and extent of Arctic sea ice. Accelerated rates of permafrost degradation are producing increasing thaw depths and northward migration of the southern boundary of discontinuous permafrost. Cryospheric processes dominate the fresh water budget of the Arctic Ocean and through this can affect processes of North Atlantic deep water formation and hence the global thermohaline circulation. Changes to the cryosphere affect the climate system at local-to-global scales.

Vulnerability of Glaciers

The worldwide retreat of mountain glaciers is one of the most conspicuous manifestations of recent climate change. In western Canada, glaciers represent a substantial source of renewable energy, contribute to the sustainability of ecosystems, and bolster the tourism economy. The impacts of ongoing and projected changes in glacier volume are likely to be substantial.

Properties, processes and feedbacks

To appreciate the vulnerability of the cryosphere to accelerated warming it is useful to consider the temperature range of Earth's surface and the unique, physical properties of water. Earth's average surface temperature ranges from around –55°C in central Antarctica to around 25°C near the equator. Among substances that are abundant at Earth's surface, water is exceptional in having a solid-to-liquid phase change within this environmental range. The large latent heat (333.5 kJ/kg) associated with this phase change introduces a switch in Earth's response to climate change: A large amount of energy is required to traverse the threshold from a frozen to a liquid state. This implies that a relatively small amount of thermal energy is required to warm Earth's surface from –1.5 to –0.5°C or from +0.5 to +1.5°C but a great deal of energy to cross the threshold from –0.5 to +0.5°C.

A second important property of ice is that, relative to water, it has high *albedo* (a measure of surface reflectivity) so that it has a strong tendency to return incoming solar radiation to space rather than absorb it and raise its temperature. This property underlies the *ice–albedo feedback*, a positive feedback that has the effect of keeping ice-covered regions cold and ice-denuded regions warm. At several times in geologic history, ice–albedo feedback ruled the day, leading to a planet that was almost totally ice-covered, a "snowball Earth". A final essential property of ice is that it is less dense than water. This density difference influences Earth's winter-season albedo: the surfaces of frozen lakes and oceans are bright and reflective rather than dark and absorptive. These three physical switches (involving thermal energy, albedo and density) are all activated at the ~0°C melting temperature of ice and account for the unique role of the cryosphere in the global climate system.

Glacier response to climate change

Glaciers are nourished by snowfall and depleted by melting. For glaciers in the Canadian Cordillera, snowfall occurs primarily in winter and melting in summer so they have separate sensitivities to changes in winter and summer climate. Glaciers accumulate ice mass at higher elevation, transport it downslope under the influence of gravity and ablate it at lower elevation. The boundary between the zones of net accumulation and net ablation is the Equilibrium Line Altitude (ELA). In a stable climate, the processes of accumulation, ablation and flow will tend to produce a glacier geometry where summer melting is balanced by winter snowfall. In a warming climate the ELA rises and, by doing so, reduces the catchment area of the accumulation zone and increases the ablation zone area. In response to the resulting imbalance between accumulation and ablation, the ablation zone shrinks through retreat of the terminus. If the long-term ELA rises above the regional land surface, glaciers cannot persist.

Earth's atmosphere is thermally stratified with a general tendency to cool with increasing elevation. This temperature gradient, termed the *environmental lapse rate*, is of the order of 6ºC per kilometre of elevation. Thus, for example, if the sea-level temperature (SLT) in some region is +12ºC the temperate at an elevation of 1 km should be around 6ºC and that at 2 km around 0ºC. The elevation of the 0ºC isotherm defines the *freezing level* and marks the boundary between regions where precipitation falls as either rain or snow. In mountainous regions, such as the Canadian Cordillera, the freezing level lies at a lower elevation than the summits of many mountains and allows snow to persist through the summer. Over time this snow accumulates, forms ice at depth and flows under the influence of gravity. The effect of a warming climate is to increase the SLT and, by doing so, to increase the elevation of the annually-averaged freezing level thereby reducing the "habitat" area where glaciers can exist.

According to IPCC projections, the likely range of increase in global mean temperature for the period 1990 to 2100 is +1.4 to +5.8ºC. Reckoned as a change in freezing level elevation, this translates to a rise in the ELA between 230 and 970 m. Thus, for example, if the 1990 ELA was 2000 m and by 2100 it rose to 2970 m, glaciers could only exist on mountains that exceeded 2970 m in elevation. Furthermore, because of the topographic hypsometry (area vs. elevation relationship) the area and therefore the volume of glacier ice would decrease substantially.

The foregoing, ostensibly quantitative, account is an oversimplification. Warm air can hold more moisture than cold air so that, in general, the rate of precipitation should tend to increase with increasing warmth. If this precipitation falls as snow then there is a negative feedback that tends to favour glacier growth. It is this feedback that underlies the differing response of the Greenland and Antarctic ice sheets to present warming. Ice volume in Greenland is decreasing (because of increasing temperature) while ice volume in Antarctic is roughly constant or increasing slightly (because of increasing precipitation). For most regions of the world, however, the net effect of contemporary warming is to decrease the volume of mountain glaciers.

Consequences

Significant consequences of glacier recession include rising sea level, decreased ocean salinity, an altered hydrologic cycle, and a potential increase in glacier-associated natural hazards. These changes will affect Earth's regions differently. For example, melting of the Greenland Ice Sheet would decrease the sea-surface salinity of the North Atlantic (with possible effects on the Gulf Stream and convection in the Labrador Sea) and would raise sea level as much as 7 m. In contrast, complete melting of the glaciers in the Canadian Cordillera would raise sea level only slightly, but changes in the water cycle might have major impacts on hydroelectric power production, silviculture, agriculture and fisheries.

Sea level rise. The volume of all Earth's glaciers and ice sheets, expressed in terms of the equivalent sea level rise, is ~70 m of which only 0.5 m can be attributed to mountain glaciers and ice caps. However, relative to the Greenland and Antarctic ice sheets, these small ice masses are more responsive to climate warming and are projected to make the dominant contribution to sea level rise over the next 100 years. Excluding thermal expansion of the oceans under a warmer climate, the second largest contribution to sea-level rise will originate from downwasting of the Greenland Ice Sheet (7.2 m sea-level rise equivalent). However, the importance of the Greenland Ice Sheet contribution to sea-level rise over the next 100 years are currently debated in light of accelerated downwasting that is occurring today and the potential slow response-time bias in current ice dynamic models. In addition, most current models of continental ice dynamics completely ignore or parameterize hydrologically-activated subglacial processes that are expected to be important during deglacial episodes.

Changes in the water cycle. Freshwater is one of Canada's most important natural resources and glaciers, which function as frozen freshwater reservoirs, have a special role in maintaining the quality of that resource. Glaciers have special significance for western Canada because they are a substantial source of renewable energy, contribute to the sustainability of ecosystems, and provide financial returns to the regional economy through tourism.

Unlike many countries, Canada has an adequate supply of freshwater. However, recent warming of Earth's surface and the lower troposphere is altering the timing and magnitude of surface runoff. In western Canada, recent trends in snowmelt-dominated catchments indicates that nival runoff is occurring earlier in the year, and this early-season nival melt can extend the duration of glacier runoff. In British Columbia alone, glaciers cover 10% of the landmass (100,000 km^2) and they supplement streamflow in snowmelt-dominated catchments during times when demands for surface water is greatest. Western Canadian glaciers have retreated dramatically in the 20th century due to unusual warmth and many of these glaciers remain far out of balance with contemporary climates. It is clear that glaciers are important to Canada, but their health under future climates remains uncertain. Glaciers worldwide receded markedly in the twentieth century and those in western Canada have retreated at a rate that is unprecedented in the last 8,000 years. Warming over the last 150 years has been more severe in western Canada than anywhere else on the globe outside very high latitudes. Under the 2×CO_2 scenarios, western Canada will likely experience large temperature increases in this century, with drastic consequences for glaciers, streams and ecosystems. Future warming will only quicken the pace at which these glaciers retreat and disappear in many mountain catchments of the region.

In western Canada, glacier runoff maintains discharge and regulates stream temperatures during periods of low flow when aquatic ecosystems are most vulnerable. Because the magnitude of glacier-derived runoff is proportional to the fraction of ice cover in the drainage basin, glacier contraction lowers the buffering capacity afforded by glacier runoff. The importance of this buffering capacity is highlighted in the Columbia River Basin where 10–20% of annual flows and 50% of summer flows are glacier fed.

Power generation in British Columbia depends heavily on glacier and snowmelt runoff. Electricity produced from surface runoff accounts for approximately 90% of BC's and 17% of Alberta's current power. The domestic electricity demand in BC is increasing about 1.8% annually and system peak load requirements are close to the capacity of the system. In dry years, glacier runoff constitutes most of late summer inflows to several large reservoirs. The sustainability of these flows in the face of future climate change remains uncertain.

The sustainability of freshwater ecosystems may be compromised by changing freshwater resource availability and associated changes in stream temperature. Maximum stream temperature typically occurs in July or August during extended periods of dry, warm weather. Reductions in glacial meltwater will result in increases in summer water temperatures, with potential negative impacts on cold-water species such as salmonids.

Hazards. Relative to most parts of the world, the population density of northwest North America is low in glaciated terrain so that the hazard posed by glaciers (for example, glacier outburst floods) is also low. Thus impacts that are deemed to be significant in the Himalaya and the European Alps are less so in the Canadian Cordillera.

Vulnerability of Permafrost

Permafrost is defined as ground that remains below 0ºC for two or more years. Where it occurs, it usually remains frozen for centuries or millennia. At the surface of permafrost terrain there is a seasonally-thawed active layer. The base of the active layer is at the permafrost table, where the maximum temperature during the year is 0°C. In unconsolidated sediments, including peat and occasionally in bedrock, the uppermost portion of permafrost (immediately beneath the active layer) is ice-rich. This ice-rich ground is primarily responsible for the sensitivity of permafrost terrain to disturbance. Ice-rich ground has a water content when thawed that is greater than saturation; the volume above saturation is the excess ice content. Ground with excess ice is sensitive to thawing, but if the ice content is less than or at saturation, the ground is thaw stable.

More than half of Canada and Russia, most of Alaska, a quarter of China and considerable parts of cold mountain ranges worldwide are underlain by permafrost. The terrain is especially sensitive to disturbance where permafrost is ice-rich, usually in fine-grained sediments, particularly glaciolacustrine and loessal deposits. Coastal and river erosion accelerate in ice-rich ground, while thaw settlement occurs in association with surface disturbance. Ice-rich soil and deeper materials on hill slopes exhibit long-term, down-slope movement, while deep-seated deformation of ice-supersaturated debris on mountain slopes leads to the development of rock glaciers**.** The construction of settlements, airfields, pipelines, roads, mines and power dams in such areas requires specialized design techniques. Permafrost in forested areas is usually discontinuous, underlying less than 80% of the ground. Over large areas of such discontinuous permafrost, the perennially-frozen ground is warmer than –5°C and in many cases warmer than –2°C. Such ground may be particularly sensitive to climate warming which threatens to raise its temperature above 0°C.

Properties, processes and feedbacks

Active layer. The thickness of the active layer is a function of the ground surface temperature, the thermal properties of soil materials and the temperature in permafrost. In general, the active layer is deepest in warm, dry soils or in bedrock, and where the mean annual temperature in permafrost is close to 0°C, while thin active layers are measured in wet peat above cooler ground. The active layer generally thickens from north to south through the permafrost regions of the Northern Hemisphere. Beneath the boreal forest of North America, the active layer varies in thickness from about 50 cm to 150 cm; north of tree line the thickness may decline, but it is rarely less than 30 cm.

Changes in climate that influence active-layer development concern variations in air temperature, snow cover, and precipitation. In northern regions, variations in air temperature and precipitation are not independent, most importantly in winter when snowfall is associated with relatively warm conditions. In central Yukon Territory, warm snowy winters during the 1980s led to permafrost warming by 1° to 1.5°C, but subsequently the ground has cooled beneath thin snow covers. A similar trend has been observed in eastern Alaska. On the North Slope of Alaska, active-layer depth increases inland, partly in response to greater trapping of snow by taller vegetation. Indeed, a substantial thinning of the active layer occurs across tree line associated with the reduction in thickness and increase in density of snow cover at tundra sites, so that cooling of the ground is greater in winter and mean annual ground temperatures are lower.

Ice-rich zone at the base of the active layer. The ice-rich zone commonly extends more than a metre below the base of the active layer. Individual ice lenses in this zone are up to several cm thick and ice masses decimetres thick can occur but are not common. Ice wedges, V-shaped bodies of massive ice, also extend downwards from the base of the active layer. In continuous permafrost, ice wedges are ubiquitous, forming polygons 10–15 m in diameter in lowlands and about 50 m wide on slopes in uplands, but in discontinuous permafrost, which underlies most of the boreal forest, they are less evident.

Thermokarst development. The relief produced by melting of ground ice is collectively called thermokarst. Many depressions act as basins for accumulation of water and snow, and once water depth is greater than about one half of the winter ice thickness, permafrost is unsustainable beneath the pool and permafrost degradation proceeds vertically downwards and horizontally. A characteristic of the expansion of thermokarst ponds and lakes in the boreal forest is toppling of trees into the water as permafrost surrounding the lakes thaws. Development of thermokarst lakes requires flat or rolling topography to contain the water-filled depressions.

During the Holocene, periods of warm climate have been associated with increased thermokarst development throughout the permafrost regions of Russia, Alaska, and

Canada. However, individual lakes may be initiated by a site-specific disturbance, such as the melting of an ice wedge, or uprooting of a tree. Changes in surface conditions following forest fire may also lead to ground thawing over a sufficient area for several lakes to develop simultaneously. Although the association of thermokarst development with climate change over millennia is well established, there is no evidence that climatic amelioration since the "Little Ice Age" has led to such development throughout the discontinuous permafrost zones. However, at the southerly margin of permafrost in the Northern Hemisphere, there is considerable evidence that permafrost which grew during the "Little Ice Age" is currently melting, as the limits for sustaining frozen ground move northwards. Across the northern Prairie Provinces of Canada recent degradation of permafrost in peatlands is well documented. Such thawing leads to elimination of forests that developed on peat plateaus raised by ice above the surrounding wetland. While an association with recent climate warming is proposed for many areas, such changes may also be repeated in a cyclical fashion due to the recurrence of fire at 500-year intervals.

Forest fires. Forest fires cause the most widespread disturbance to surface conditions in permafrost regions. The degree of disturbance is determined by the intensity and nature of the wild fire, which are associated with antecedent meterological conditions. Projections of the nature and frequency of forest fires in a warmer world suggest the extent of damage may increase. However, some environments are less susceptible to disturbance by fire, particularly where there is sufficient moisture to reduce the intensity of burning at the surface.

The initial impact of fire is to raise ground surface temperatures in summer due to reduction in shading. Destruction of vegetation leads to reduced evapotranspiration, raising ground temperatures further as the ground surface dries out. Deepening of the active layer usually follows, so that by the end of summer the ice-rich zone has begun to degrade. On hillslopes, the thawing leads to landslides, as the ice-rich zone provides a lubricated plane with minimal shear strength. Such disturbances may also follow particularly warm summers when there is substantial rainfall in early autumn, but are almost ubiquitous in sloping ground after fire.

Active-layer deepening is accentuated in years following fire by an increased snow depth in burned areas because interception of snow by the forest canopy is eliminated or reduced. In combination, such changes at the ground surface lead to an increase in ground surface temperature of >2°C and may lead to permafrost eradication. Where permafrost is thin, the degradation may be completed within a decade, but where the thickness is >10 m centuries are required to thaw the ice-rich ground completely. Vegetation recovery is assured over such time scales, so that permafrost degradation will likely stop. While the most substantial damage occurs in ice-rich terrain, the slowest recovery is at dry sites.

Potential positive feedback. The principal ecological role of permafrost is to provide a near-surface hydrologic barrier that directly influences soil moisture and runoff within the active layer. Most soil microbial activity and plant root development occurs in the active layer. A potential positive feedback associated with future warming is to alter high-latitude ecosystems from their current status as net sinks for greenhouse gases to methane and carbon sources as the soil warms.

Permafrost response to climate change

The response of permafrost to climatic warming occurs through melting at the top of permafrost over periods of years, with or without changes in active-layer thickness. Warming of the temperature profile within permafrost occurs over decades to centuries, and upward displacement of the base of permafrost to reach a new equilibrium thickness takes centuries to millennia. In addition to the direct effects of climate, principally through changes in air temperature and snow fall, local ground temperature is also strongly influenced by factors related to climate, such as the type of vegetation and properties of the organic layer and soil. These factors can interact in a complex fashion, making it difficult to predict the overall effect of climatic changes. Modification of permafrost conditions and subsequent adjustment of related biotic and abiotic surface processes takes place over variable and extended time periods. First attempts are now being made to monitor active-layer thickness and permafrost temperatures within global climate-related observing systems. Efforts presently concentrate on monitoring the thickness of the active layer and on obtaining borehole temperatures.

Various scenarios for the extent of equilibrium permafrost following climate warming indicate that the lower limits of permafrost occurrence in mountain areas could rise by several hundred metres. However, the time for permafrost to reach new equilibrium thickness may be on the order of centuries, as the elevation of the permafrost base is accomplished by the relatively small geothermal flux. This time is extended by at least an order-of-magnitude for ice-rich permafrost, where the change in thickness may eventually be tens of meters. Convective effects from groundwater circulation may accelerate rates of permafrost degradation, but these are unpredictable at present and usually site-specific in scale.

Northward migration of forests following summer climate warming may accelerate active-layer deepening through changes to snow-pack properties. This is an example of amplification in climate/permafrost relations due to ecological effects that makes prediction of the impact of climate change on terrain conditions difficult. Field evidence of such amplification has been obtained from examination of active-layer development in western arctic Canada during the early Holocene climate optimum. About 9,000 years ago, when tree line was considerably further north, and in areas that are now tundra, the active layer was about 2.5 times its present depth. About one half of the active-layer thinning in the last 9,000 years has been attributed to the disappearance of forest from these sites and the remainder to climate cooling.

In contrast with the winter regime, warm conditions during summer are usually dry in continental permafrost regions while rain and snowfall are associated with lower temperatures. Evaporation from moist ground also consumes energy and lowers the ground heat flux, further reducing active-layer development. A result is that changes in site wetness may alter ground temperatures and active-layer thickness more than changes in air temperature *per se*.

Polar regions are expected to experience greater disturbance than the mid latitudes under enhanced greenhouse warming and changes are expected to be greater in winter than in summer. A wetter climatic regime is projected, with greater snowfall due to a more vigorous general circulation in the atmosphere. For permafrost terrain we should expect a thicker active layer due to the cumulative net heat storage caused by warmer summers and thicker, insulating snowpacks during winter.

Ground temperature records from Alaska data point to the critical role of snow cover on the ground thermal regime because, although air temperatures have not increased in the 1990s, average snow depth has and ground temperatures have warmed over large areas. Yet in eastern Alaska and in adjacent Yukon Territory, Canada, average ground temperatures cooled in the 1990s due to thinner snow cover. Modelling studies of the impact of climate warming on permafrost suggest that for the $2\times CO_2$ scenario permafrost will disappear from much of its present distribution and the continuous/discontinuous boundary will move several hundred km northward. The time required to reach equilibrium for such transformation is on the order of several millennia. The effects of climatic variability on permafrost stability have also been investigated using computational models. One such study, which compared the effects of a stable *vs*. a warming climate showed that permafrost may become established rapidly due to frost penetration in winter, but the rate of summer thawing is reduced because of insulation provided by the thawed active layer. In consequence, permafrost has a persistence that is due to the latent heat held in ground ice and the seasonally-changing thermal properties of the surficial layers.

Consequences

The most significant terrain hazards wrought by the presence of permafrost are due to melting of the ice-rich zone following deepening of the active layer. Changes in climate may tend to cause relatively gradual deepening of the active layer within the boreal forest because the active layer is relatively thick and climate changes tend to be gradual, superimposed on year-to-year variability. In contrast, surface disturbances, leading to removal of vegetation and organic soil, can cause a substantial, rapid impact on active-layer thickness. The most common and widespread agent of such change is forest fire. Anthropogenic disturbance, usually from construction or mining, may have a similar effect.

Permafrost is a persistent phenomenon and its future presence is assured within the boreal forest. The terrain hazards presented by permafrost relate to the presence of ground ice. Given that near-surface layers of permafrost terrain are characteristically ice-rich, ground subsidence and slope instability are likely the principal hazards to be expected under climatic change.

Concluding Remarks

Ironically, the challenges of climate cryospheric change have emerged in Canada following a long period of declining national commitment to climate, Northern, and cryospheric science. Only recently has this trend been reversed through a combination of influences that include the Intergovernmental Panel on Climate Change, the Canada Research Chairs Program, the NSERC/SSHRC Northern Task Force, the Canadian Foundation for Climate and Atmospheric Sciences and the International Polar Year. Research networks that have floated on this rising tide include ArcticNet and CASES, CRYSYS, the Polar Climate Stability Network and the Western Canadian Cryospheric Network.