MASSIVE ICE FORMATION IN THE EUREKA SOUND LOWLANDS: A LANDSCAPE MODEL

Wayne H. Pollard1, Trevor Bell2

1. Department of Geography and Centre for Climate and Global Change Research, McGill University, 805 Sherbrooke St. W., Montreal, Qc. Canada, H3A 2K6
e-mail: pollard@felix.geog.mcgill.ca

2. Department of Geography, Memorial University of Newfoundland, St. John’s, Newfoundland, Canada, A1B 3X9
e-mail: tbell@morgan.ucs.mun.ca

Abstract

Massive ice is a common constituent in fine-grained marine sediments situated below the Holocene marine limit in the Eureka Sound area of Ellesmere and Axel Heiberg Islands. Its stratigraphic setting and character suggest it is intrasedimental in nature and formed as permafrost aggraded following Holocene emergence. This paper describes the relationship between ground ice occurrence and landscape evolution on Fosheim Peninsula, with particular emphasis on the duration and extent of glacier ice cover, sea level history, and the style and pattern of glaciomarine sedimentation in formerly submerged basins. A two-phase model is proposed to explain the nature of ground ice distribution. In this model permafrost conditions associated with late Quaternary glacial, periglacial and emergent environments are considered. The evolution of permafrost conditions is divided into two phases corresponding with: (a) prior to ~8-9 ka BP, and (b) post-glacial emergence after ~8 ka BP.

Introduction

Ground ice forms an important component of surficial deposits below the Holocene marine limit on Fosheim Peninsula, west-central Ellesmere Island (Figure 1). It is abundant in the Lowlands surrounding Eureka Sound and Slidre Fiord and frequently occurs as massive ice bodies. Field investigations identified seven sites characterized by numerous exposures of massive ice which were the focus of stratigraphic and topographic analysis. It is argued in this paper that the massive ground ice observed on the peninsula is intrasedimental in origin, as opposed to buried surface ice, and that its characteristics relate to local environmental factors, such as ground temperature, water source and surficial geology during ice segregation. The distribution of ground ice on the peninsula is therefore primarily a function of permafrost aggradation, which is closely linked to late Quaternary glacial and sea level events. The main objective of this paper is to describe the relationship between ground ice occurrence and landscape evolution on Fosheim Peninsula, with particular emphasis on the duration and extent of glacier ice cover, sea level history, and the style and pattern of glaciomarine sedimentation in formerly submerged basins.

Figure 1. Map of the Fosheim Peninsula showing the distribution of thick marine sediments, Holocene marine limit, present ice cover extent, ground ice sites (numbered 1-7) and locations of massive ice exposures.
Background

Fosheim Peninsula is characterized by a cold polar desert climate. It has a mean annual air temperature of -19.7°C and a mean annual temperature range of 43°C. The area receives an average precipitation of 64 mm/yr (60% as snow) making it one of the driest places in Canada. An oil well drilled in the middle of the peninsula revealed a permafrost thickness of over 500 m and a geothermal gradient of roughly 0.038°C/m (Taylor, 1991).

Fosheim Peninsula lies on the axis of the hypothesized Inuillian Ice Sheet of the last glaciation (Blake, 1970). Although the debate on the existence of this ice sheet is ongoing, Bell (1992, 1996) argued that extensive regional ice advances across the peninsula likely predated the last glaciation (Late Wisconsinan). Surficial mapping and radiocarbon dates reveal that at 9 ka BP currently ice-free uplands supported small ice caps, local cirque glaciers were more extensive and the sea occupied broad coastal and interior lowlands. Deglaciation of the peninsula was underway by 9.5 ka BP and sea level was at, or close to, marine limit (142-149 m ASL) from at least 10.6 to 8.7 ka BP (Bell, 1996). Submerged Lowlands acted as shallow marine basins, collecting an extensive record of sedimentation related to local glacial and sea level history. From oldest to youngest, these sediments have been interpreted to represent: (1) a deepening of the submerged basins as sea level transgressed to marine limit; (2) marked increases in sedimentation rate in response to increased ablation on local ice caps (>9.5 ka BP); and (3) marine regression and shallowing of the marine basins (Bell 1996). Initial rapid emergence caused relative sea level to fall to ~80 m ASL by 7 ka BP, but this was followed by a steadily declining rate of emergence in the middle to late Holocene (Figure 2).

Paleoclimatic records for the Canadian High Arctic reveal three broad trends: a continuously warm period (with temperatures similar to the present) that prevailed from roughly 10 ka to 5 ka BP; followed by a cooling trend for 4-5 ka; and a marked warming since the early 1900s, involving a temperature rise of 3-4°C (Bradley, 1990; Koerner and Fisher, 1990; Taylor, 1991).

Massive ice origin

All of the massive ice exposures observed on Fosheim Peninsula lie below the Holocene marine limit (Figures 1 and 2) and nearly all (94%) display an upper contact with fine-grained marine sediments (Figure 3). Typically, these sediments consist of massive to weakly laminated mud with occasional thin sand layers. They are interpreted to represent suspension settling from turbid plumes, generated in marine basins by inflowing meltwater streams (Bell, 1996; Aitken and Bell, in press). Marine sediments reach a maximum thickness of ~36 m in lower Slidre River valley and have a minimum age range of 5.9 to 9.6 ka BP.

The contact between marine sediments and underlying massive ice is conformable and gradational over a few 10s of centimeters. Commonly, it grades from laminated silt through icy silt and silty ice to pure ice, or from massive clay through clay with reticulate ice veins to white ice containing clay pods or to pure ice.
Sediment and gas inclusion layers within the ice parallel the laminated structure of the overlying marine deposits and multiple bands of clear, sediment-rich and bubble-rich ice give massive ice bodies their foliated appearance (Pollard, 1991). Sediment inclusions in the ice have similar mineralogical and textural characteristics to the overlying marine units (Figures 4 and 5). Lower contacts are rarely visible, but in the limited number of cases where they are visible the massive ice unit conformably overlies weathered Tertiary bedrock. The bedrock unit contains ice veins and lenses. Figure 5 presents a stratigraphic column for a massive ice body exposed near Slidre River. In this section two marine facies (units A and B) conformably overlie massive ice (unit C) which in turn overlies weathered Tertiary sandstone (unit D). Backwasting of the massive ice cause thawed debris to collect at the base of the section, hence in this case the lower contact is exposed near the base of an adjacent section. This stratigraphy and these contact relationships suggest that the massive ice grew by segregation at the base of the marine sediments.

Chemical continuity between ice veins in the lowermost marine sediments and the uppermost massive ice is demonstrated by similarities in conductivity measurements and δ18O values across unit boundaries (Figure 5). Electrical conductivity measurements of thawed ice provide a proxy for total dissolved solids concentrations within a sample. Conductivity profiles and measurement of anions and cations were used to compare chemical signatures of ice from different stratigraphic units. Conductivities of ground ice on the Fosheim range between 500 and 2300 µS/cm, higher than those reported in the literature for other massive ice bodies (cf. Mackay and Dallimore, 1992). In addition, values for massive ice are comparable with those from ice contained within confining sediments (Figure 5), suggesting a common water source. This relationship was also reflected in the concentrations of major ions.

Isotopic analysis from the site shown in Figure 5 also reveal a similar pattern. δ18O values range between -28.89‰ and -34.84‰ for reticulate ice, -33.01‰ and -36.84‰ for massive ice, and -36.01‰ for an ice vein in a nearby exposure of Tertiary sandstone. The δ18O
values of the enclosing sediments are similar to those for the massive ice; the more positive value higher in the section is thought to reflect a surface water input. The continuity of these samples also suggests a common water source and isotopic signatures indicate it was precipitated under very cold conditions.

In general, massive ground ice is either (i) intrasedimental ice, formed by the combined processes of ice segregation and ground water intrusion, or (ii) buried ice, formed by the burial and preservation of a surface ice body (e.g., glacier ice). Massive ice on Fosheim Peninsula is interpreted as intrasedimental ice for the following reasons:

1. The massive ice is conformably overlain by marine sediments and contains internal structures which parallel the upper ice contact. The ice does not show evidence for primary thaw or erosional contacts which would indicate that the ice predates burial (cf. Mackay, 1989; Mackay and Dallimore, 1992).

2. Massive ice exposures have a gradational contact between the ice and the overlying marine sediments which in this case can only develop where the ice is younger (i.e., epigenetic) than the marine sediments.

3. The petrographic and gas inclusion patterns are typical of segregated ice (Mackay, 1989).

4. Sediment inclusions in the massive ice are similar to the overlying marine sediments, indicating segregated ice (cf., Mackay, 1989).

5. There is chemical continuity between ice from marine sediments, massive ice and adjacent Tertiary sandstone suggesting a common cold water source. Since in most cases buried ice is derived from a source other than local groundwater (or at least precipitated under different temperature conditions), there is usually a physical and chemical unconformity between the massive ice and ice in the enclosing sediments (Mackay and Dallimore, 1992). However segregated ice formed from glacier meltwater will have a chemical signature similar to glacier ice.

A buried glacier origin is rejected for the following reasons: (i) since all of the massive ice lies below the Holocene marine limit, it is unlikely that tabular bodies of glacier ice would have survived prolonged marine submergence (see below), (ii) there are no glacial landforms or sediments associated with these massive ice bodies, and (iii) preservation of submerged and buried glacier ice during high sea level stands would have produced significant thaw degradation structures which are not observed in exposures.

Landscape model

An intrasedimental origin for massive ground ice on Fosheim Peninsula implies that ice segregation was primarily a function of permafrost aggradation. The distribution of ground ice and the timing of its formation is therefore best explained within the context of late Quaternary paleoenvironmental change and landscape evolution. Two distinct phases of permafrost history are proposed: Phase I existed prior to ~8-9 ka BP when local environmental parameters were not conducive to large-scale ice segregation, while Phase II postdates initial glacio-isostatic emergence, which created suitable landscape (e.g., host sediments) and environmental (e.g., ground temperature, water supply) conditions for ground ice formation.

Phase I

During Phase I, three main landscape systems existed on Fosheim Peninsula: (a) ice-covered uplands and cirque valleys, (b) ice-free lowlands, and (c) recently submerged lowlands. The geomorphic record of upland ice caps (lateral and proglacial meltwater channels) strongly suggests that they were entirely cold-based for most of their history (cf. Dyke, 1993). Although basal ice temperatures are not known, it is likely that subglacial permafrost far exceeded central ice-cap thicknesses but may have been thinner than beneath adjacent ice-free areas, due to strain heating in the ice which would cause some permafrost degradation (Dyke, 1993).

Ignoring local variations in heat flow and geology, the deepest and oldest permafrost would be expected in areas above Holocene marine limit that remained ice-free during the last glaciation. Assuming a geothermal gradient similar to present (0.038°C/m, Taylor, 1991) and dry periglacial conditions 3-5°C colder than present then equilibrium permafrost depths for ice-free Lowlands were probably 80-130 m deeper than present (500+ m, Taylor, 1991). Although permafrost was well developed, there is no evidence to suggest that ground ice was ever an important component of surfaces located above Holocene marine limit. The extreme aridity of the region combined with the extensive exposure of weathered Tertiary bedrock likely limited the potential for ground ice development.

Radiocarbon dates on shells from raised marine deposits in Slidre River valley suggest that Holocene marine limit was likely established by 10.6 ka BP (GSC-4784), with sea level remaining high (within 10 m of marine limit) until 8.7 ka BP or later (Figure 2, Bell, 1996). The history of marine transgression prior to 10.6 ka BP is poorly known. England (1992) proposed two hypothetical glacio-isostatic loading curves for adjacent Greely Fiord: hypothesis A produced a slow rise in relative sea level from 60 m ASL to marine limit in 8.9 ka,
whereas hypothesis B had relative sea level rising from -80 m to marine limit in 4-5 ka. Bell (1996) considered that hypothesis A was more consistent with the sea level chronology from Fosheim Peninsula, with the implication being that areas lying between 60 and ~150 m were possibly submerged for 3 to 10 ka, depending on their proximity to marine limit, whereas lower-lying areas were likely submerged for much longer. Consequently, Pleistocene permafrost that existed beneath lowlands prior to marine transgression likely experienced degradation, even near marine limit, due to the abrupt surface temperature shift (-19°C) and prolonged submergence (cf. Taylor, 1991). Whether the period of marine transgression was long enough to completely degrade permafrost is not important; the critical point is that ice in superficial deposits would not have survived the last marine transgression.

**Phase II**

Phase II (after ~8-9 ka BP) coincides with a period of dramatic warming and increased runoff from glaciers in the Canadian High Arctic (Bradley, 1990; Koerner and Fisher, 1990; Taylor, 1991). Upland ice caps on Fosheim Peninsula were rapidly melting as indicated by the relatively high rates of marine rhythmite sedimentation in the formerly submerged Slidre River basin (6 mm/a between 9.1 and 7.9 ka BP; Bell, 1996). Perhaps these sedimentation rates also indicate the partial development of warm-based thermal conditions in local glaciers and upland ice caps (cf. Stewart, 1991). Eventually, permafrost conditions beneath deglaciated areas would have gradually changed to correspond with the surrounding ice-free areas which, due to the higher temperatures of the early Holocene, may have experienced some permafrost warming and degradation.

Initial rapid emergence between 8.5 ka and 7 ka BP caused relative sea level to fall by ~70 m, exposing fine-grained marine sediments (Figure 2). This marked the beginning of permafrost formation and ground ice aggradation which continued more or less simultaneously with shoreline regression. A steadily declining rate of emergence in the middle to late Holocene exposed the deepest parts of the former marine basins, where some of the most extensive ground ice is now located.

The most extensive and laterally continuous marine deposits are contained in the lower Slidre River basin (~250 km²; Figure 1), where they form discrete wedges in tributary valleys, reaching maximum thickness of 36 m in the main valley (Bell, 1996). These deposits are commonly associated with deep, tabular ground ice bodies (Figure 3). Interfluves and sites more distal to meltwater outlets consist of weathered bedrock, either marine washed or draped by a thin veneer (<0.5 m thick) of marine sediment. The resulting ground ice bodies are typically shallow, irregularly shaped and of uneven thickness (e.g., south Slidre Fiord, Hot Weather Creek, Blue Man Cape). This form of ground ice plays an important role in thermokarst and active layer detachment activity.

In both cases, as permafrost reached the base of the marine unit, a thick uniform layer of ground ice formed from the ground water contained in the underlying weathered bedrock. There are two possible sources of water: meltwater from glaciers and snow, or sea water. Electrical conductivity, major ions and oxygen isotope ratios tend to suggest a groundwater system supported by meltwater mixed with saline porewater and precipitates.

Of the three main landscape types - ice covered, ice free and emergent - the latter is the most likely to contain significant ground ice because of the availability of groundwater and the frost susceptible nature of the fine-grained marine sediments. For the most part, upland surfaces which were either ice-free or ice-covered are composed of coarse-grained Tertiary bedrock which is unsuitable for massive ground ice formation.

**Conclusions**

This study offers three main conclusions regarding ground ice origin and landscape evolution. (1) Massive ice occurs widely below the Holocene marine limit and has an intrasedimental origin. (2) Permafrost is closely related to early Holocene glacial and sea level histories and can be divided into two phases; Phase I occurred while ice caps were fully developed, and Phase II corresponds with ice cap retreat and glacio-isostatic emergence. Only the last phase is associated with the occurrence of massive ground ice. And (3) massive ice deposits are time transgressive and formed as permafrost aggraded into emerging marine sediments during the Holocene.

**Acknowledgments**

This work was supported by the Natural Science and Engineering Research Council of Canada, the Department of Energy Mines and Resources (EMR Research Agreement Program) and the Atmospheric Environment Service. The authors wish to express sincere appreciation to the Polar Continental Shelf Project (PCSP/EPCP manuscript 02197) for its generous field support and Sylvia Edlund and Doug Hodgson for their helpful discussions. The field assistance of Craig Forcese, Peter Barry and Ray Shannon is also acknowledged. Logistical support for student assistants was provided by the Northern Scientific Training Program (DIAND). Helpful comments were provided by Val Sloan and an anonymous reviewer.
References


