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An ice shelf mechanism for Heinrich layer production

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Abstract. The effect of an ice shelf in the Labrador Sea on ice-rafted sediment delivery to the glacial North Atlantic is investigated using a finite element numerical model of ice shelf flow. Discharge into the shelf from Hudson Strait creates a thick central core, extending downstream into the shelf, that is flanked by relatively thin ice. Melting at the base of the deep keel would produce cool, fresh water which would rise to refreeze along the keel's flanks. Debris deposited by melting deep central ice could create the Heinrich layers observed in the Labrador Sea while the debris-rich ice protected by basal freezing on the flanks of the thick ice plume could transport sediment over large distances. Heinrich layers would be produced solely by external climate forcing, without changes in ice sheet flow. The mechanism is plausible if the lifetime of the ice shelf is about 1000 years.

1. Introduction

The sedimentary record of the glacial North Atlantic (GNA) is punctuated by sudden, brief increases in ice-rafted debris (IRD) deposition [*Bond et al.*, 1992; *Bond and Lotti*, 1995]. The most distinctive IRD layers are the terrigenous carbonate-rich Heinrich layers, whose provenance is the lower Paleozoic carbonate rocks of Hudson Bay and Hudson Strait in eastern Canada [*Bond et al.*, 1992; *Gwiazda et al.*, 1996]. Heinrich layers are found throughout the North Atlantic as far afield as the continental slope west of Portugal [*Robinson et al.*, 1995]. Heinrich layers are also made conspicuous by changes in foraminifera assemblage that indicate a shift to cooler sea surface conditions. Estimates of the duration of each depositional event range from about 100 to 2000 years [e.g., *Dowdeswell et al.*, 1995; *MacAyeal*, 1993], although the apparent rapidity of sedimentation [*Higgins et al.*, 1995] makes shorter durations more likely. Heinrich layer deposition has been correlated with culminations of Daansgaard-Oeschger (D-O) cycles (progressive cooling cycles followed by abrupt warming, recorded as $\delta\,^{18}O$ variations in Greenland ice cores) [*Bond et al.*, 1993]. Not all D-O cold culminations coincide with Heinrich layers but when they do, the cold-warm oscillation is especially large. The sedimentary and chronologic evidence suggests that a combination of changes in the mode of ice delivery from Hudson Strait to the North Atlantic and in the local climate led to the production of Heinrich layers.

Two contrasting hypotheses have been proposed to account for the dramatic increases in the volume of debris ice rafted across the GNA. One hypothesis suggests that reduced sea surface temperature would increase the lifetimes of ocean-going icebergs sufficiently to create Heinrich layers without variation in ice sheet or ice stream flow [e.g., *Dowdeswell et al.*, 1995]. That scenario, in which an unchanged supply of carbonate debris is distributed over a larger seafloor area during creation of a Heinrich layer, implies reduced deposition near the debris source (the mouth of Hudson Strait) and concurrent increases in IRD from other (noncarbonate) sources. However, *Andrews et al.* [1994] find large increases in carbonate sedimentation in the Hudson Strait outlet that correspond with Heinrich layers 1 and 2 (about 14.5 and 20.5 ka, respectively), and the Heinrich layers contain almost exclusively sediment derived from Hudson Bay and Strait. An alternative is the ice stream surge model in which a growing ice dome over Hudson Bay reaches a critical thickness at which the base of the ice is warmed to the melt temperature and rapid ice discharge is triggered [e.g., *Alley and MacAyeal*, 1994; *MacAyeal*, 1993]. Periodic surges of an ice stream draining the Laurentide ice sheet through Hudson Strait would increase the volume of carbonate IRD delivered to the GNA, although that model offers no explanation for the coincidence of Heinrich layers and cold culminations of D-O cycles. A third scenario that considers the survival of debris within ocean-going icebergs is proposed here.

The survivability of IRD depends on the distribution of debris within an iceberg. Most of the sediment entrained in an ice sheet or ice stream is contained within the basal ice [*Hambrey*, 1994], so it forms a veneer on the bottom surface of icebergs calved into the ocean at the ice edge. IRD thus exposed at the surface of an iceberg is immediately vulnerable to melting out of the ice and is deposited close to the iceberg source. Even if the iceberg capsizes, thereby moving debris-rich ice above the sea surface, the debris-rich ice is exposed to subaerial melting. Englacial debris-rich layers are protected from immediate melting but such layers are typically thin and contain small amounts of sediment [*Hambrey*, 1994].

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Large outlet glaciers often terminate as floating ice tongues (for example, Jacobsavns Isbræ in Greenland and Mackay Glacier in Antarctica). The Hudson Strait outlet glacier may also have emerged as an ice tongue. Basal ice conditions have been observed alongside 1.8 km of grounded ice and beneath floating regions of the Mackay Glacier Tongue [*Powell et al.*, 1996], a fast flowing glacier that drains the East Antarctic Ice Sheet through the Transantarctic Mountains and terminates as an ice tongue floating in -1.8° C seawater. The debris-rich basal layer thins due to melting, from 20 m to 10 m, as grounded ice flows toward the grounding line. Small icebergs calved near the grounding line contain some debris-rich basal ice, but larger icebergs calved from the end of the tongue are observed to be debris-free (R. Powell, personal communication, 1997). The majority of icebergs produced in such a setting carry a small volume of debris, which because of its location within the basal ice, is poorly suited to travel far from its source.

Preservation of debris-rich basal ice may improve if a glacier flows into a confined ice shelf instead of terminating as an ice tongue. Ice shelves form where grounded ice goes afloat on the sea surface. Ice flows from an inland source, through the ice shelf, and calves into the ocean, forming icebergs at the ice shelf front. Large slopes in the basal topography of an ice shelf may encourage freezing of water onto the base of the shelf (i.e., underplating) as meltwater plumes rise up-gradient and become supercooled [e.g., *Bombosch and Jenkins,* 1995; *Robin,* 1979]. Such processes appear to be at work today beneath the Filchner-Ronne Ice Shelf (FRIS), where meltwater plumes deposit thick ice layers in shallow regions between ice stream outlets [*Bombosch and Jenkins,* 1995]. As new ice is accreted, the original basal ice moves upward into the ice shelf, away from the ice-ocean interface. Any debris within the basal ice will thus be protected and may survive for long periods of time before being melted out of ocean-going icebergs. Sediment inclusions have been observed in thin layers near the boundary between glacier and accreted ice in two cores retrieved from the Filchner-Ronne and Amery Ice Shelves [*Eiken et al.,* 1994]. The total sediment content at those locations is small, but it could be larger at other locations depending on the interval over which the ice experiences basal melting before basal freezing begins.

Cyclic growth and decay of an ice shelf in the Labrador Sea could explain the cyclic increases in IRD volume and distribution implied by Heinrich layers. In this scenario, ice shelf basal processes provide the mechanism for increased IRD abundance, and distribution and atmospheric temperature regulation of ice shelf growth and decay (cold and warm phases of the D-O cycles) provide the required periodicity. A finite element numerical model is used to test the hypothesis.

2. Ice Shelf Model

Ice shelf thickness and velocity are computed using stress-balance and mass-balance equations. The equations are represented by piecewise linear functions on a finite element mesh that represents the Labrador Sea and Baffin Bay. Kinematic and dynamic boundary conditions represent the influence of ice flow from North America and Greenland and the influence of the ocean at the calving front. Beginning with an arbitrary initial ice thickness, the model steps through time until a steady state solution is found.

Ice shelves flow by gravity-driven horizontal spreading. This is described in the model by a set of stress-balance equations, simplified by assuming that horizontal flow is depth-independent, as observed in modern ice shelves. The equations are derived by *MacAyeal and Thomas* [1982]. The constitutive relation for ice (Glen's flow law) is embodied by an effective viscosity, which depends on a temperature-dependent rate constant and a flow law exponent (the usual value of 3 is used). The stress-balance portion of this model has successfully reproduced the flow of the Ross Ice Shelf as part of a European Ice Sheet Modelling Initiative intercomparison experiment [*MacAyeal et al.*, 1996].

Ice thickness is computed using a mass balance equation in which thickness change depends on the rate of new ice accumulation and the rate of ice thinning or thickening due to ice flow. The volume and velocity of ice flowing into the ice shelf at grounding lines is specified, and ice arriving at the shelf front is allowed to flow through it. That is, a fixed shelf-front position is specified and calving processes are not described.

The model domain comprises the Labrador Sea and Baffin Bay between eastern Canada and the west coast of Greenland (Figure 1). The grounding line is specified at modern sea level. Isostatic adjustment and sea level fluctuations are ignored for simplicity, so the model does not include sea floor elevation, as is appropriate for those simplifications. The ice shelf front is arbitrarily selected between Labrador and the southern tip of Greenland. The domain is represented by 2018 triangular elements which range in area from 101 to 5800 km^2.

Ice input is specified for last glacial maximum (LGM) conditions under the assumption that the regions of the Laurentide and Greenland Ice Sheets draining into the ice shelf are at equilibrium. That input corresponds to steady state, not surging, ice stream flow through Hudson Strait. Topographic divides constructed for an 18 ka Laurentide Ice Sheet [*Dyke and Prest*, 1987] and an average ice accumulation rate of 0.2 ma^{-1} [e.g., *Fisher et al.*, 1985] are used to estimate the volume of ice flowing from drainage basins on the western side of the model domain (Table 1). An ice flux one third of the observed modern flux [*Reeh*, 1989] is distributed evenly along the Greenland coast. A uniform ice accumulation rate of

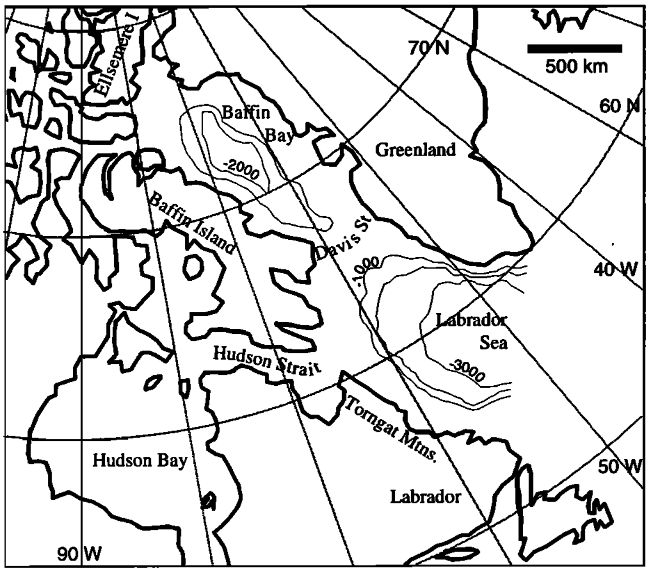

Figure 1. Labrador Sea and Baffin Bay region (Lambert equal-area projection). Coastlines (heavy lines) are at modern sea level. Isobaths beneath the model domain (light lines) are given in meters below sea level. The Hudson Bay region is the source of the terrigenous carbonate sediment that is characteristic of Heinrich layers.

0.2 ma^{-1} is specified at the ice shelf surface. The accumulation rate is zero (no model-specified melting or freezing) at the bottom surface of the shelf.

Steady state is achieved when ice shelf volume remains constant through successive time steps, that is, when the flux of ice through the grounding line is equal to the flux of ice through the shelf front. The modeled shelf is composed of only meteoric ice, and its thickness and velocity are determined by internal dynamics alone.

3. Labrador Sea Ice Shelf

An ice shelf in the Labrador Sea and Baffin Bay may begin as perennial sea ice which grows late in the cooling phase of a D-O cycle. That situation is approximated in the model by allowing the ice shelf to grow from an initial uniform 5 m ice thickness. Possible effects of wind on the ice pack are not addressed. Figure 2 chronicles the first 400 years of ice shelf growth. The thick tongue of ice emerging from Hudson Strait spreads quickly to develop large basal topographic gradients. The ice shelf could grow and decay on millennial timescales, comparable to climate-change timescales inferred from Greenland ice cores. Ice cover develops much more slowly in Baffin Bay because the ice influx is much smaller than into the Labrador Sea.

Flow of the Labrador Sea ice shelf is similar to that observed in other large ice shelves. Ice speed (Figure 3) increases toward the shelf front, and spatial variations in velocity are due mainly to spatial variations in discharge through the grounding line. The majority of ice flows from Labrador Sea grounding lines to the shelf front in 250 - 600 years. Short residence times mean that the ice shelf can adjust rapidly to climate signals, such as changes in air temperature or precipitation rate.

The spatial pattern in ice thickness is the most tantalizing feature of the ice shelf. Most (about 90%) of the ice floats below the sea surface, so the ice thickness map in Figure 4 can be viewed as an inverted topographic map of the ice shelf base. A deep central core emerging from Hudson Strait is flanked by relatively shallow regions. The core is a result of the large discharge from the Hudson Bay drainage basin through Hudson Strait. Thin ice originates at the corners of the Hudson Strait outlet, where large horizontal strain rates between fast and slow flowing ice cause vertical thinning. Rifting and thinning is a common process at fast-slow boundaries in Antarctic ice shelves [*MacAyeal et al.,* 1986]. Large thickness gradients may encourage melting and freezing processes at the shelf base [e.g., *Bombosch and Jenkins,* 1995; *Robin,* 1979].

4. Debris Packaging

Melting and freezing at the base of an ice shelf depend on the temperature and salinity of water at the ice-ocean interface and on basal slope. Where water at the interface is warmer than the local (pressure- and salinity-dependent) freezing temperature, ice melts. Freezing occurs when water is supercooled with respect to ambient conditions. Cool water may originate as meltwater produced beneath the ice shelf. The salinity of meltwater is low compared to other sub-ice-shelf watermasses, so it is positively buoyant and rises. As a meltwater plume flows up the ice shelf basal gradient, the water moves to regions of lower pressure and therefore higher local freezing temperature. Where the ambient freezing temperature is above the plume water temperature, the plume water must freeze. Once initiated, meltwater plumes exert a local control on basal melting and freezing that supersedes regional oceanography.

Ice-ocean interaction beneath the Labrador Sea ice shelf may be inferred by comparing its basal topographic relief with that of the FRIS. This simple approach is preferable to developing a meltwater plume flowline model for the Labrador Sea because such a model would require poorly constrained assumptions

Table 1. Steady State Ice Flux Into Model Domain

Drainage Basin	Area km^2	Ice Flux m^3 a^{-1}
Torngat Mountains	0.47 x 10^6	94 x 10^9
Hudson Bay/Hudson Strait	2.30 x 10^6	460 x 10^9
Cumberland Sound	0.02 x 10^6	4.2 x 10^6
Baffin Island	0.22 x 10^6	44 x 10^9
Ellsemere Island	0.12 x 10^6	24 x 10^9
Greenland	n.a.	33 x 10^9

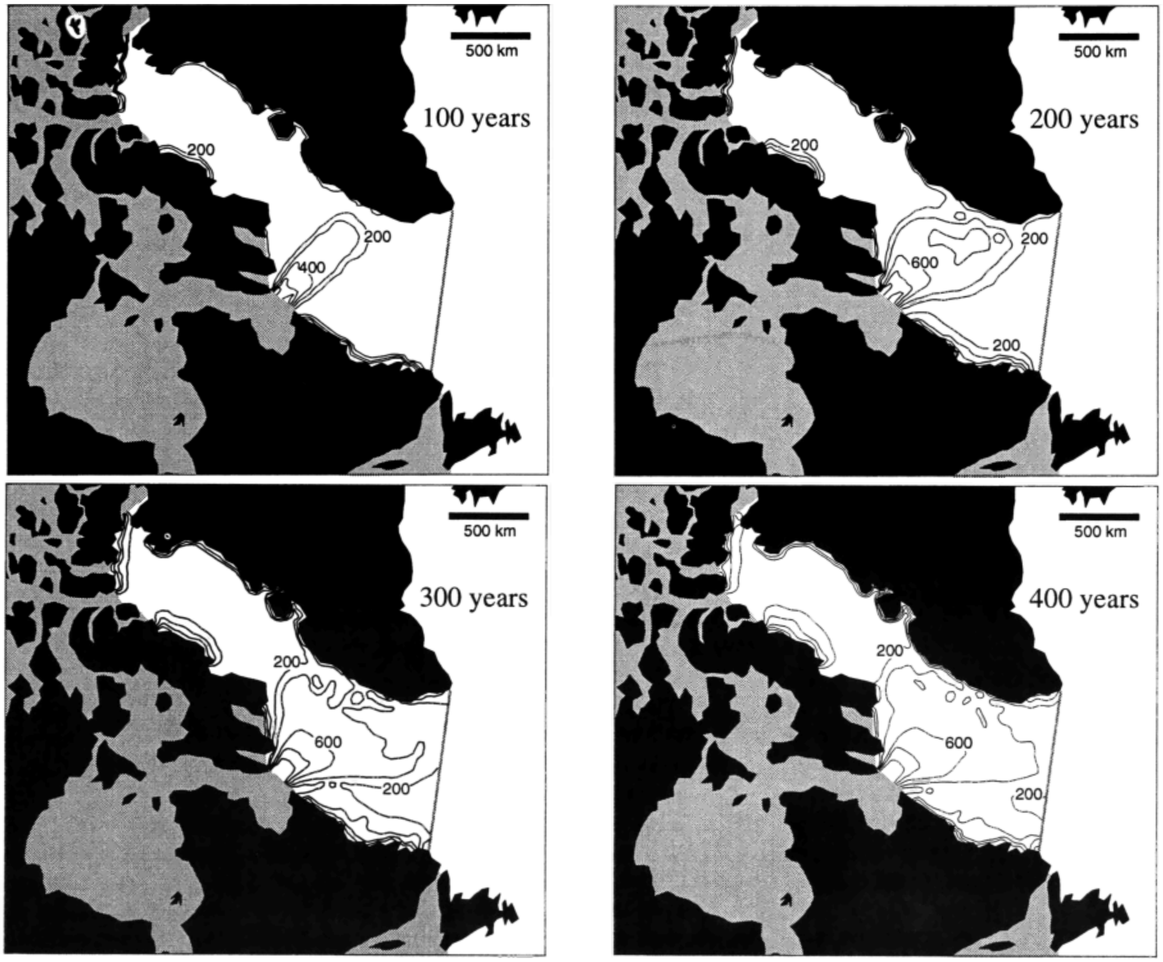

Figure 2. Ice shelf growth from an initial uniform 5-m thickness at 100 a intervals. Large thickness gradients develop rapidly in the Labrador Sea portion of the shelf, allowing basal melting/freezing processes to begin soon after ice shelf inception. The timescale of ice shelf growth fits within the millennial-scale climate oscillations of Daansgard-Oeschger cycles. The Baffin Bay portion of the shelf grows more slowly, requiring thousands of years to reach steady state. Shaded areas are outside the model domain. The heavy line between Labrador and the southern tip of Greenland represents the model's seaward boundary. The isopach interval is 200 m.

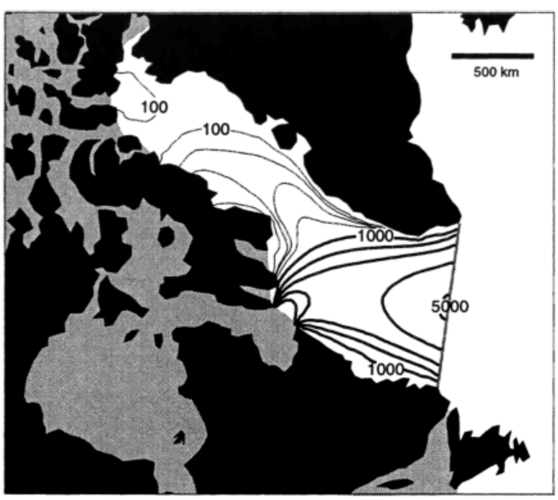

Figure 3. Steady state ice speed. Ice flows through the Labrador Sea portion of the shelf rapidly in part because of large discharge from Hudson Strait. Smaller influx from grounded ice and constriction at Davis Strait result in a uniformly sluggish ice shelf in Baffin Bay. The isotach interval is 1000 $\mathrm{m\,a^{-1}}$ in Labrador Sea (heavy lines) and 100 $\mathrm{m\,a^{-1}}$ (light lines) in Baffin Bay.

Figure 4. Steady state ice shelf thickness. Because most of the ice floats below sea level, this map can be viewed as an inverted map of basal topography. Large basal gradients in the Labrador Sea ice shelf are the basis for the Heinrich event mechanism proposed in the text. Basal melting is expected along the deep central keel of the shelf, while basal freezing is expected upslope along its flanks. The isopach interval is 200 m.

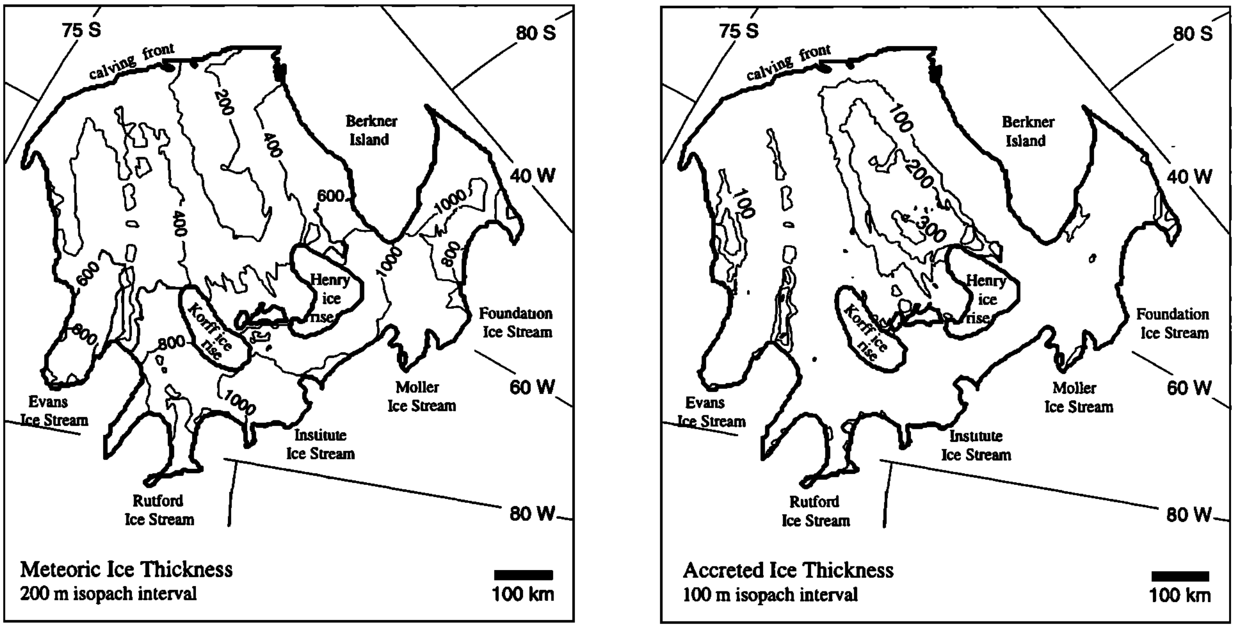

Figure 5. Meteoric (derived from ice flowing from land and top-surface snow accumulation) and accreted ice thickness of the Ronne Ice Shelf (Lambert equal-area projection). Accreted ice deposits are thickest where meteoric ice is thinnest. The maps are made with data provided by *Vaughan et al.*, [1994].

about sub-ice-shelf oceanology without affording additional insight. FRIS meteoric and accreted ice thicknesses shown in Figure 5 are mapped here using observational data from *Vaughan et al.* [1994]. The map of meteoric ice thickness in the FRIS is directly comparable with the model Labrador Sea ice thickness map in Figure 4. Both the modern FRIS and the model ice shelf are fed by large ice streams that emerge as thick, fast flowing cores flanked by thinner ice. The thickest accreted ice deposits beneath the FRIS occur where meteoric ice is thinnest, in the regions between major ice stream outlets. It appears that meltwater plumes begin at depth, where sub-ice-shelf water is warmest, and rise toward the shallowest areas possible, where the meltwater refreezes. In a modeling study of this process, *Bombosch and Jenkins* [1995] find new ice accretion along much of the plume rise: larger ice crystals are deposited where basal slope is steep, and smaller crystals are deposited where basal slope shallows. They compute accretion rates of up to 2.5 ma^{-1}. A similar situation is expected for the Labrador Sea shelf. Relatively warm North Atlantic Intermediate Water (between 2 and 3° C, according to isotopic shifts reported by *Oppo and Lehman* [1993]) circulating under the ice shelf would melt ice along the deep keel of the shelf (where the freezing point of fresh water is about -0.6° C), producing cool, fresh meltwater. The meltwater would be positively buoyant and would rise toward adjacent shallow regions and refreeze. Thus, by analogy with the FRIS it is expected that beneath the Labrador Sea ice shelf, marine ice would accrete along the shallowing flanks of the ice core discharging from Hudson Strait.

The aspect of the modeled ice shelf most critical for this discussion is the spatial pattern of basal slope. Variations in ice shelf thickness (and thus basal slope) depend primarily on the rate of discharge from grounded ice, which in turn, depends on the assumed accumulation rate. A series of tests in which accumulation was halved and doubled for the entire domain and was halved for just the Hudson Bay/Hudson Strait outlet were performed in order to observe the sensitivity of model outcome to the rate of ice accumulation. As should be expected, increases in the accumulation rate lead to increases in net ice thickness and ice speed and vice versa. However, in all tests the deep central core emerging from Hudson Strait is preserved. It is also noted that the arbitrary placement of the calving-front boundary has little effect on model outcome because variations in ice thickness develop as a result of dynamics at the mouth of Hudson Strait (even early in the shelf's history when ice at the shelf front is very thin). The pattern of relative ice thickness, used here to propose a Heinrich layer production mechanism, is a robust feature of the model.

5. Heinrich Layer Production

A simple Heinrich "event" scenario is proposed:

1. Atmospheric and sea surface cooling near the cold culmination of a D-O cycle foster ice shelf growth in the Labrador Sea. Large ice shelf basal gradients develop

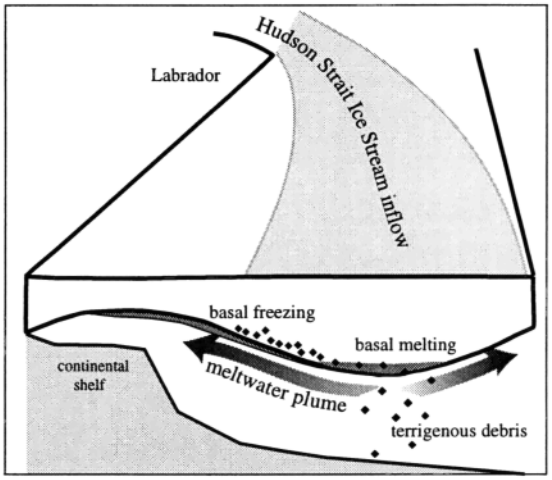

Figure 6. Cartoon of the IRD packaging mechanism. Melting along the center of the deep ice keel creates meltwater plumes, which refreeze to underplate debris-rich ice along the flanks of the keel. Melting deposits terrigenous carbonate sediments within the Labrador Sea, and underplating leads to icebergs with embedded debris-rich layers.

rapidly as a thick core of ice flows downstream from Hudson Strait.

2. Melting along the deepest part of the core ice creates buoyant meltwater plumes which rise along the basal gradient. Rising plumes become supercooled and refreeze, underplating the flanks of the thick ice core with a layer of new ice (Figure 6).

3. Ice flowing into the shelf from Hudson Bay through Hudson Strait contains carbonate sediment-rich basal ice. Melting under the deepest portions of the shelf deposits terrigenous carbonate sediment in the Labrador Sea. Underplating on the flanks of the deep ice core protects sediment-rich ice while it is in the shelf and increases its residence time in ocean-going icebergs, allowing IRD to be transported long distances in the GNA.

4. Warming at the end of a D-O cycle is transmitted quickly through the fast flowing ice shelf, so it retreats and debris-rich iceberg production ends.

The amount of IRD produced in this manner depends on the thickness of debris-rich basal ice preserved by underplating and the rate at which such ice is produced. That, in turn, depends on the proportion of carbonate debris-bearing core over which melting or freezing occurs. The rate of basal ice melting at the grounding line and along the deep keel of the shelf is difficult to estimate. The melting rate required to maintain a constant ice thickness and temperature gradient between a cold upper surface and melting at the base (following *MacAyeal* [1985]) is of the order of 0.01 ma^{-1}, an unreasonably small value. *Bombosch and Jenkins* [1995] compute melt rates of up to 2 ma^{-1} at the grounding lines of ice streams discharging into the FRIS. Carbonate sediment will be deposited downstream of the grounding line, but it is possible that some debris-rich basal ice will survive into areas along the flanks of the central thick ice keel where basal freezing can occur. The depth at which the meltwater begins to refreeze depends on the local fluid dynamics (heat and momentum transfer) of the plume, which cannot be determined without additional speculation about subshelf oceanography [cf. *Bombosch and Jenkins,*1995].

The thickness of debris-rich ice that must survive to be underplated along the flanks of the central ice core can be estimated using the flux from Hudson Strait and the model-derived ice thickness. Taking the duration of a Heinrich event to be 1000 years and the volume of IRD in a typical Heinrich layer to be 100×10^9 m^3 [*Alley and MacAyeal,* 1994], if the shallowest 30% of the core area becomes subject to basal freezing, that ice must contain 0.72 m of sediment. If the basal debris concentration in the ice ranges from 10 to 35% by volume (*Hambrey* [1994] suggests up to 50% by weight), then between 2.1 and 7.2 m of debris-rich basal ice must be preserved to produce a Heinrich layer by the basal-freezing mechanism. If the lifetime of the shelf is shorter, say 500 years, then between 4.2 and 14.4 m of debris-rich basal ice would be required.

How deeply the debris is embedded within the ice shelf and in icebergs calved at the shelf front depends on the rate and duration of marine ice deposition. The rate of basal freezing varies along a plume flowline, because of changes in plume ascent rate and plume temperature. Unlike the FRIS situation, where plume flowlines are thought to correspond to ice flowlines, ice shelf flowlines along the flanks of the deep ice keel are perpendicular to the basal slope and therefore to the plume flow direction. Thus, along any particular ice flowline, the rate of freezing may remain fairly constant. If a typical ice accumulation rate is 1 ma^{-1} (near the smaller end of the range modelled by *Bombosch and Jenkins* [1995]), as much as 300 m of marine ice could be added to the ice shelf base. The debris would be deeply embedded within the ice column and well poised to survive traveling long distances within ocean-going icebergs.

6. Conclusion

A successful Heinrich layer deposition model must produce sufficient IRD, distribute it over a large area, and account for the apparent relationship between Heinrich layer deposition and D-O cycles. The scenario suggested by the present study may meet those requirements. It is assumed that local climate cooling prior to Heinrich layer deposition allows an ice shelf to grow in the Labrador Sea. That Heinrich layers are deposited at only the more extreme D-O cold phases may reflect some threshold in the magnitude or duration of cooling required for ice shelf growth. The ice shelf, fed by steady state flow (i.e., not surging) from surrounding ice sheet drainage basins and modified by normal basal melting and freezing processes, could produce the embedded debris-rich icebergs needed to deposit a Heinrich layer in 1000 years. It is unlikely that a shorter-lived ice shelf could accomplish the task. The scenario also becomes untenable if much less than 30% of ice emerging from Hudson Strait is underplated with new ice. However, within those constraints, an ice shelf offers a new explanation for the occurance and distribution of Heinrich layers which depends only on external (and well-documented) climate forcing without appeal to changes in continental ice sheet flow.

The model must also accommodate the pattern of sedimentation observed within the Labrador Sea. Ice melting and terrigenous carbonate sedimentation at the mouth of Hudson Strait (as observed by *Andrews and Tedesco* [1992]) would be heightened while the ice shelf exists because the ice is not free to drift away as it would be in calved icebergs. Thus the proposed mechanism can accommodate Heinrich layer deposition at locations both distal and proximal to the debris source. Basal debris in ice emerging from the eastern coast of Labrador is also of concern because such sediments are not found in Heinrich layers. It is possible that slower ice flow rates along this coast (due to small ice flux) allow sufficient time for basal melting offshore of the grounding line to remove debris before it reaches regions where basal freezing predominates. Between ice shelf growth events, IRD sedimentation in the North Atlantic should return to typical glacial-stage levels. A potentially more troubling issue is that an ice shelf should limit sea surface biologic activity. No significant foraminifera hiatuses have been reported [e.g., *Aksu and Mudie,* 1985; *Aksu et al.,* 1992], but the ephemeral nature of the ice shelf may make those changes difficult to identify, given the time resolution of the stratigraphic record from these areas (thousands of years).

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