Catastrophic ice shelf breakup as the source of Heinrich event icebergs

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[1] Heinrich layers of the glacial North Atlantic record abrupt widespread iceberg rafting of detrital carbonate and other lithic material at the extreme-cold culminations of Bond climate cycles. Both internal (glaciologic) and external (climate) forcings have been proposed. Here we suggest an explanation for the iceberg release that encompasses external climate forcing on the basis of a new glaciological process recently witnessed along the Antarctic Peninsula: rapid disintegrations of fringing ice shelves induced by climate-controlled meltwater infilling of surface crevasses. We postulate that peripheral ice shelves, formed along the eastern Canadian seaboard during extreme cold conditions, would be vulnerable to sudden climate-driven disintegration during any climate amelioration. Ice shelf disintegration then would be the source of Heinrich event icebergs. *INDEX TERMS:* 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 1827 Hydrology: Glaciology (1863); 4267 Oceanography: General: Paleoceanography; *KEYWORDS:* Heinrich events, ice shelf disintegration

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1. Introduction

1.1. Signature of Heinrich Events

[2] Six major pulses of ice-rafted detritus (IRD) mark North Atlantic sediment cores between 14,000 and 70,000 years ago in the last glaciation [*Heinrich*, 1988; *Broecker et al.*, 1992; *Bond et al.*, 1992, 1997; *Grousset et al.*, 1993]. This detritus consists of limestone, dolomite and other lithic fragments of Canadian origin [*Bond et al.*, 1997; *Hemming et al.*, 1998, 2003] and is commonly spread across the North Atlantic ice-rafting belt to just off the coast of Portugal, a remarkably long distance of more than 3000 km. The carbonate layers have sharp basal contacts [*Bond et al.*, 1992, 1997]. Together, these characteristics suggest short-lived, catastrophic discharges of icebergs from the Laurentide Ice Sheet (LIS) that tracked across the North Atlantic to the European seaboard. These outbursts are referred to as Heinrich events, following terminology established by *Bond et al.* [1999, p. 48], where Heinrich layers in marine sediment cores were defined as "percentage increases in detrital carbonate above ambient values."

1.2. Previously Proposed Heinrich Event Mechanisms

[3] The origin of the Heinrich events remains a major unsolved problem in Pleistocene climatology. At issue is whether dynamics within the LIS itself [*MacAyeal*, 1993] or external climatic forcing on the ice sheet or its environment caused these dramatic events [*Bond and Lotti*, 1995; *Bond et al.*, 1999; *Dowdeswell et al.*, 1995]. One possibility is that switches in basal thermal conditions of the LIS led to periodic purges of interior ice through Hudson Strait into the North Atlantic Ocean [e.g., *MacAyeal*, 1993; *Payne*, 1995; *Greve and MacAyeal*, 1996; *Marshall and Clarke*, 1997]. The purge mechanism is consistent with the sharp onset of detrital-carbonate deposition and with the copious flush of icebergs necessary to depress North Atlantic deepwater formation implied by proxy records from both the Greenland ice core and North Atlantic sediment cores. However, because the timing of such purges would be tied solely to a Laurentide ice-dynamics clock, this mechanism does not readily explain why sediment from entirely separate glacial regimes in Europe and Iceland occurs in the same ice-rafted debris (IRD) pulses as the material of Canadian origin. In fact, the arrival of Icelandic and European detritus precedes the deposition of carbonate-rich debris in cores SU90-09, VM23-81, and DSDP 609 from the central and eastern portions of the North Atlantic ice-rafting belt [*Grousset et al.*, 2000, 2001; *Bond et al.*, 1999; *Snoeckx et al.*, 1999]. It is unlikely that the Icelandic sources of icebergs could have anticipated the basal-melting trigger of LIS purges, thus the presence of Icelandic detritus immediately prior to the deposition of Canadian carbonate material suggests an external climate forcing for both the Laurentide and Icelandic sources.

[4] A marine ice sheet instability mechanism [*Thomas*, 1977] operating in the inland-deepening Hudson Strait

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could also generate iceberg outbreaks. This mechanism is similar to the tidewater-glacier retreat mechanism of Alaskan glaciers, and could be activated at cold-climate extremes if a grounding line extending seaward of Hudson Strait pushed a protecting moraine shoal over the edge of the continental shelf, as implied by *Alley* [1991]. However, a retreating ice stream grounding line alone would be unlikely to produce a sufficient population of icebergs to distribute the large sediment load implied by each Heinrich layer. An alternative is simply that falling atmospheric temperatures drove the LIS to advance onto the continental shelf, producing calving margins and thus a relatively large fluxgate for icebergs [e.g., *Hindmarsh and Jenkins*, 2001]. However, this accounts for neither the sudden onset of Canadian carbonate deposition nor its great volume. None of these previously considered mechanisms easily account for both the sediment distribution and the timing of Heinrich events.

1.3. Revision of an Ice Shelf Source for Heinrich Event Icebergs

[5] Here we revisit the Labrador Sea ice shelf hypothesis proposed by *Hulbe* [1997], in light of a disintegration mechanism recently recognized along the Antarctic Peninsula. In the original hypothesis, an ice shelf, proposed to grow during a cold extreme of climate, provided a means to sequester terrigenous debris within icebergs (refreezing meltwater beneath the shelf) in a manner that would favor its subsequent survival against immediate melt-out at the start of transatlantic iceberg drift trajectories. As originally proposed, the IRD supply was relatively steady over at least a 1000-year interval. The hypothesis lacked an abrupt onset for carbonate deposition.

[6] In the revision proposed here, the sudden breakup of an ice shelf would deliver Canadian-source icebergs and their stored sediment load rapidly across the Atlantic Ocean. Ice shelf expansion occurs during the cold-culmination of a series of increasingly high-amplitude Dansgard-Oeschger (D-O) oscillations and disintegration is the result of summer-season warming. That warming could signify the termination of the cold-extreme, or it could simply reflect intrastadial variability. The sudden breakup mechanism also facilitates debris protection and dispersal over long distances. Both the modern analogue and our application of the disintegration process to Heinrich events are discussed in detail in section 2, below. We make note of the fact that the original ice shelf mechanism called for a large, Ross Ice Shelf style embayed ice shelf that would have covered the entire Labrador Sea. This was necessary to stabilize the ice-thickness patterns within the ice shelf so as to promote sufficient marine-ice underplating by oceanographic circulations particular to embayed ice shelves [e.g., *MacAyeal*, 1985; *Bombosh and Jenkins*, 1995]. In the revised mechanism, an embayed ice shelf is no longer necessary. Rather, we posit a series of fringing ice shelves along the Canadian coast, able to expand only when climate cools substantially and vulnerable to rapid disintegration due to climate amelioration.

[7] We shall argue that, with the addition of the climate-triggered (meltwater-induced) breakup process, the ice shelf mechanism meets all the criteria for Heinrich-layer deposition. In our new hypothesis, Heinrich events are understood to be a response to climate forcing, with sedimentation that is amplified by glaciological processes. The hypothesis relies on a known mechanism that is extreme in its suddenness and amplitude, but which is regulated by an external climate forcing that need not be extreme, e.g., a mild increase in the length or intensity of the surface meltwater producing summer season. The ultimate cause of the Heinrich events, that is, the cause of the underlying climate changes that yield ice shelf buildup followed by sudden disintegration, remains unknown.

2. Breakup of Larsen A and B Ice Shelves, Antarctica

2.1. Fragmentation Process

[8] Our revival of the ice shelf hypothesis for Heinrich events is motivated by the series of extreme ice shelf disintegrations witnessed along the Antarctic Peninsula since 1995 [*Rott et al.*, 1996, 1998; *Vaughan and Doake*, 1996; *Vaughan et al.*, 2001; *Scambos et al.*, 2000, 2004]. Of these disintegrations, those of the Larsen A and B ice shelves, occurring at the end of the summer melt seasons in 1995 and 2002, respectively, are the best studied. A sequence of satellite images of the Larsen B ice shelf acquired in early 2002 by the Moderate Resolution Imaging Spectrometer (MODIS) flying aboard the U.S. Terra satellite capture that year's large collapse event [*Scambos et al.*, 2004]. Three of the images are reproduced here (Figure 1). A 3198 km^2 region of the ice shelf (all areas reported here are computed from the MODIS imagery by O. Sergienko, personal communication, 2002), nearly its entire area, broke catastrophically into thousands of icebergs over the course of two weeks beginning on about 23 February 2002. Most of the breakup activity spanned the three-day period from 3 to 6 March.

[9] After the ice shelf breakup (7 March image; see panel 3 of Figure 1), a dense mass of icebergs covered a sea-surface area approximately twice the original area of the shelf, or 6742 km^2. This mass appears in the image as a mixture of two size classes of icebergs. The first class, comprising a total area of 1629 km^2, is larger than the 250-m resolution of the MODIS imager, and thus appears as bright, white tabular bergs with distinct outlines in the imagery. The second class is smaller than the 250-m resolution, and appears in the original color imagery as a blue-colored material filling the space between the bright tabular icebergs. The area of this blue-colored material is 5113 km^2 and it contains the fragments created from 1569 km^2, or about half, of the original prebreakup ice shelf area. (Original imagery is available at the National Snow and Ice Data Center (NSIDC) of the U.S. Web site, http://nsidc.org/iceshelves.) The blue color (represented by darker gray in panel 3 of Figure 1) is interpreted to indicate previously englacial ice exposed at the surface as a result of iceberg capsize. Also seen within the dense mass of small icebergs are trails of darker pixels. These colors are associated with englacial debris, now exposed on the sides of the capsized fragments. The debris must have originated

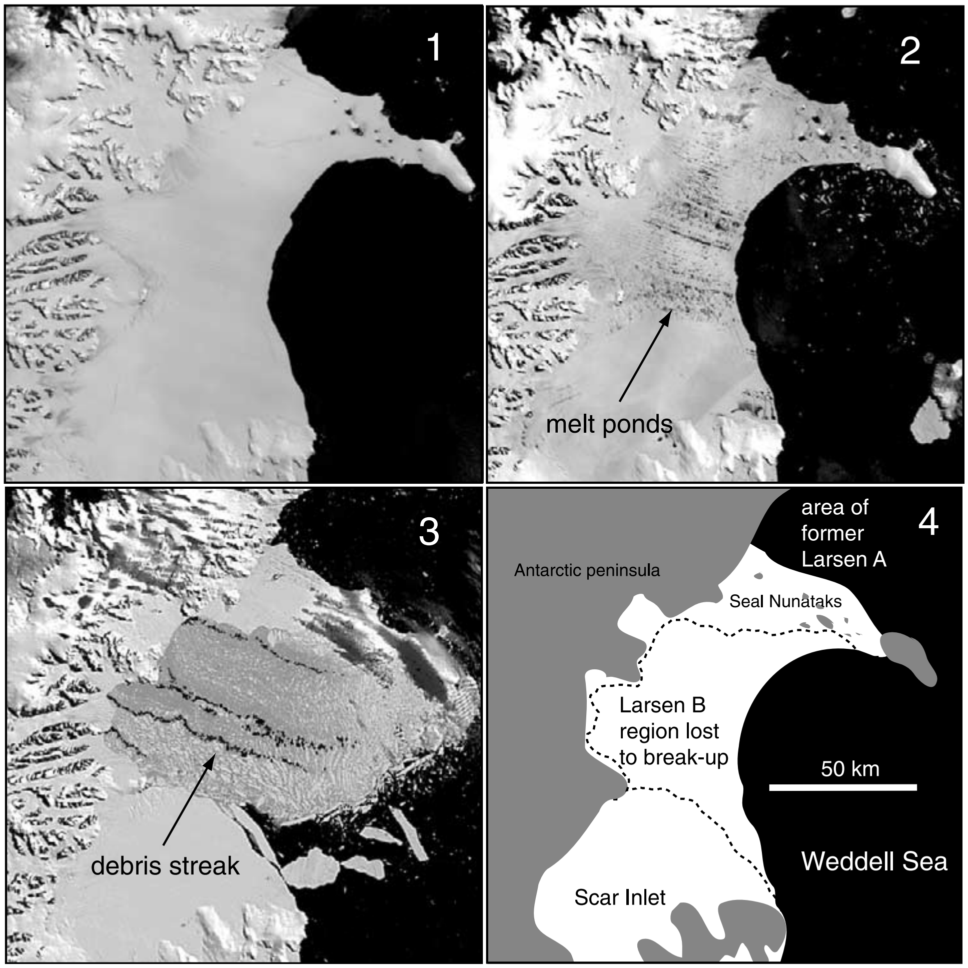

Figure 1. Satellite images (MODIS) depicting the 2002 disintegration of Larsen B ice shelf, Antarctica, on 22 November 2001 (panel 1), 31 January 2002 (panel 2), and 7 March (panel 3). Melt ponds and meltwater-filled surface crevasses are visible as dark patches on the white ice surface (panel 2). Large, uncapsized icebergs appear as distinct bright pieces and capsized fragments (or ice mélange) appear as darker gray matrix (panel 3). Bands of englacial debris exposed by capsize, more readily visible in the true color imagery, are colored black for clarity (panel 3). Location map (panel 4).

upstream as lateral and medial moraines in the mountain glaciers that feed the ice shelf. Pixels in which debris is exposed at the surface are colored black in panel 3 of Figure 1.

[10] The catastrophic breakup of the Larsen B Ice Shelf is remarkable because it reveals an iceberg production mechanism far different from those previously thought to determine the extent of Antarctic ice shelves. Typically, calving is considered to be a process limited to the infrequent shedding of large, tabular icebergs at the seaward ice front and more frequent, smaller events. What distinguishes the surface-meltwater-induced iceberg production mechanism is the fact that innumerable icebergs are created simultaneously through the entire breadth of the ice shelf, i.e., not at a restricted seaward ice-front boundary. Smaller-scale, but similar, disintegrations have also removed large sections of ice shelves on the western side of the peninsula in recent years [*Scambos et al.*, 2000]. This recently recognized process allows iceberg production to be sudden and high-volume, and to be initiated at a distinct atmospheric climate threshold. In addition, the calving process produces icebergs of a wide variety of sizes, including those that readily capsize.

2.2. Climate Connection

[11] The notion that the nearly instantaneous breakup of ice shelves can be a response to a modest change in summer atmospheric temperature is motivated by observations of annual average surface temperature warming throughout the Antarctic Peninsula and by the fact that this warming is responsible for the accumulation of meltwater on fringing ice shelves over the last two decades [*Vaughan and Doake*, 1996; *Rott et al.*, 1998; *Skvarca et al.*, 1998; *Doake et al.*, 1998; *Scambos et al.*, 2000]. Statistical analysis of temperature data from recording stations located along the Antarctic Peninsula yields an annual average surface warming of about 2.5°C over the last half-century [*Vaughan et al.*, 2001]. Coincident with this warming, in which summertime temperatures rise to near the melt temperature at some

recording locations, is an increase in the duration (number of days) of the melt season. On Larsen A and B, the melt season lengthened from about 40 to about 80 days over the last two decades [*Scambos et al.*, 2000]. Following in step with those changes, meltwater ponds often appeared on the surface of the shelves during the height (January–February) of the melt season.

[12] Severe surface melting preceded the disintegration of Larsen B. The image of 23 November 2001 (panel 1 of Figure 1) displays the ice shelf immediately prior to the onset of summer melting. As with virtually all ice shelves in Antarctica, the surface appears homogeneous and white, a result of the accumulation of snow through the previous winter. By 23 February 2002, (panel 2 of Figure 1) surface meltwater ponds are widespread and many crevasses are filled to the brim with meltwater. These surface melt accumulation features are commonly aligned in streets that follow tracks of crevasses extending from the grounding line to the calving front. This observational record, along with similar surface conditions observed in prior summer seasons, and preceding prior breakup events, is compelling evidence that the breakup was ultimately caused by climate warming and the meltwater it produced.

[13] The impetus for ice shelf disintegration is the mechanical effect of meltwater filling surface crevasses (panel 1 of Figure 2). As described by *Weertman* [1973], *Hughes* [1983] and *van der Veen* [1998] and applied to the circumstances of Larsen A and B ice shelves by *Scambos et al.* [2000], surface crevasses that are filled with liquid water can propagate farther downward toward the ice shelf bottom than can air-filled counterparts. This is possible because the density of water is greater than the density of ice so the water can act as a mechanical wedge that pushes apart the vertical walls of the crevasse. The ice overburden pressure that would normally limit the depth of crack-tip propagation is overcome when the water-filling depth is about 90% of the initial crack depth (the percentage derives from the ratio of ice-column density to water density). Sufficiently deep initial crevasses with an adequate water supply can crack through the full ice shelf thickness [*Scambos et al.*, 2000].

[14] Ice shelves on which observations of surface meltwater are rare do not experience iceberg production events

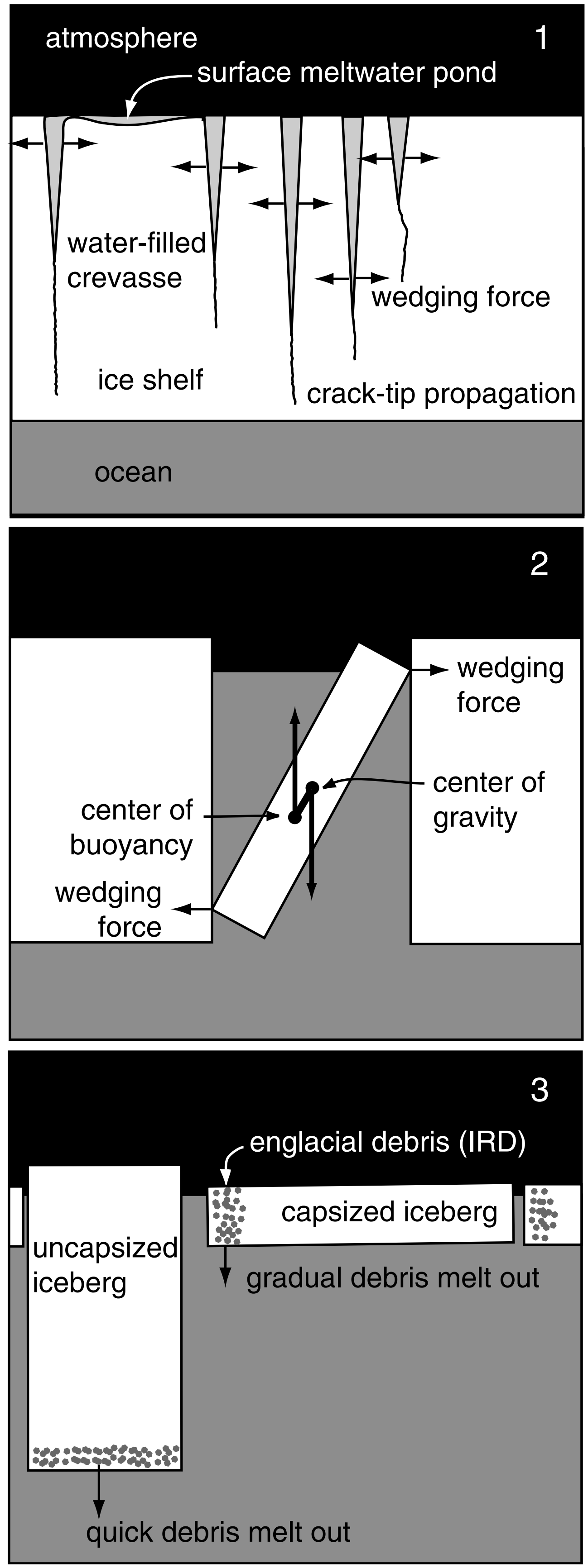

Figure 2. Ice shelf cross sections depicting processes associated with sudden ice shelf disintegration. Climate warming initiates the process via surface meltwater (panel 1). Water-filled crevasses propagate downward to the ice shelf bottom, eventually allowing discrete, independent ice shelf fragments to develop. Some fragments will begin to capsize spontaneously, converting buoyancy-driven capsize torque into forces that push apart the surrounding ice shelf apart (surrounding ice depicted for simplicity as unfractured, panel 2). Approximately 50% of the original, prebreakup area of Larsen B Ice Shelf, for example, broke into icebergs that subsequently capsized by rolling 90° onto their sides (panel 3). Gravitational potential energy released by capsize drives the final stages of the ice shelf's explosive disintegration, and yields icebergs with debris bands (as shown in panel 3) that are favorably oriented for long-term survival in transatlantic drift.

like those observed on Larsen B. For example, the Ross and Ronne-Filchner ice shelves release gigantic tabular icebergs on an episodic basis that appears to have little to do with environmental conditions [*Lazzara et al.*, 1999]. Instead, these icebergs result from the interaction of slowly growing ice shelf rifts with the geometry of coasts that embay the ice shelves [*Rignot and MacAyeal*, 1998]. These gigantic tabular icebergs calve on multidecadal timescales and maintain the location of a mechanically favorable ice shelf front.

[15] Larsen-style ice shelf disintegration occurs within a particular climate context. Mean annual temperatures are well below freezing, about −9°C, while in summer the air temperature warns to near the freezing point, about −1°C [*Morris and Vaughan*, 2004; *Skvarca and De Angelis*, 2004], Farther south, the Antarctic continent is experiencing a slight cooling [*Vaughan et al.*, 2001]. Thus we can view ice shelf breakup as a threshold event, requiring a few years to decades of seasonal warming within a regionally cold regime.

2.3. Iceberg Configuration

2.3.1. Iceberg Size and Dispersal

[16] Taking as a constraint imposed by the sedimentary record that in each Heinrich event icebergs must be delivered rapidly over large distances, we assert that ice shelf fragments created by Larsen-style disintegrations are well-suited to the task. Simply put, tabular icebergs calved from ice shelves in the normal, nonexplosive manner are too massive to drift efficiently downwind [*Lichey and Helmer*, 2001; D. R. MacAyeal et al., unpublished manuscript, 2003]. This has been demonstrated repeatedly by giant Antarctic icebergs, such as B9, B15, and A22, calved from various Antarctic ice shelves over the last decade) [*Keys et al.*, 1990; *Long et al.*, 2001]. (Scatterometer data animations of giant iceberg drift are available at http://www.scp.byu.edu/data/iceberg/database1.html.) These large icebergs lingered for decades to years near their sources [e.g., *Nøst and Østerhus*, 1998]; and once beyond the coastal ocean, exhibited curious immobility despite the strong winds of the Southern Ocean. Sea ice is observed to pass around the large tabular bergs, indicating how the iceberg motion diverges from the direction and speed with which smaller, wind-driven elements of the ocean surface move. Another example, Iceberg B15A, has remained steadfast against Ross Island, Antarctica, for two years despite a prevailing southerly wind (D. R. MacAyeal et al., unpublished manuscript, 2003; see also http://amrc.ssec.wisc.edu/amrc/iceberg.html). In contrast, the relatively small icebergs produced by the breakup of Larsen A in 1995 exited the Weddell Sea and dispersed across the South Atlantic in a matter of two years. These small icebergs respond immediately and directly to wind stress, as does sea ice [*Smith and Banke*, 1983; *Matsumoto*, 1999]. At the time of this writing, the icebergs from Larsen B have not yet had sufficient time to drift, so their long-term trajectories are not discussed here.

[17] The merit of small icebergs as agents of IRD dispersal is a simple matter of iceberg-drift dynamics [e.g., *Gladstone et al.*, 2001; *Lichey and Hellmer*, 2001; *Matsumoto*, 1999; *Bigg et al.*, 1997]. Three forces are involved: wind stress generated by the iceberg freeboard and subaerial surface area, F_w, ocean friction generated by both viscous forces and wave drag induced by iceberg drift, F_o, and the Coriolis force, F_c, which in the Northern Hemisphere pushes icebergs to the right of the direction of drift. Although all three forces affect iceberg drift, the balance between F_w and F_o is most important for small icebergs, and the balance between F_w and F_c is most important for large icebergs. To demonstrate this contrast, we compare the scales of ocean friction and Coriolis force, with the aim of determining which is largest for icebergs of horizontal-scale L and thickness scale H. We assume that ocean resistance by the combination of skin friction and wave-drag, i.e., "hull drag," is given by

$$F_o = L\,H\,k\,u, \tag{1}$$

where u represents the iceberg drift velocity and k represents a coefficient of friction. Both L and H appear in the expression for F_o because motion through the ocean is resisted by both skin friction and wave drag. The Coriolis force on an iceberg is given by

$$F_c = L^2 H\,\rho f\,u, \tag{2}$$

where ρ is ice density and f is the Coriolis parameter [e.g., *Pedlosky*, 1979]. The ratio of F_o to F_c is proportional to $1/L$. This suggests that as the length scale L gets large, hull drag becomes less important in balancing the wind stress. An iceberg that responds to wind-forcing by generating a balancing Coriolis force drifts at right angles to the wind direction. For the prevailing westerly winds on the North Atlantic Ocean, large tabular icebergs must move south, not east. Small icebergs, particularly those which are irregularly shaped, are instead likely to balance wind stress with friction induced by downwind drift; much like the force balance involved in a ship or sailing yacht. Thus icebergs produced by Larsen-style disintegration are better suited to produce the observed Heinrich layer distribution than are icebergs produced by calving from an ice shelf in steady mass balance.

2.3.2. Iceberg Capsize and Debris Transport

[18] Larsen-style ice shelf disintegration upends the ice stratigraphy. Many, possibly a majority, of the icebergs created during the catastrophic disintegration of Larsen B capsized, rolling 90° onto their sides (panels 2 and 3 of Figure 2). The gravitational potential energy transmitted between adjacent toppling blocks may in fact be an important dynamic feedback in a surface-meltwater-induced ice shelf breakup [*MacAyeal et al.*, 2003]. Here we are interested in the effect of capsize on debris distribution within the drifting icebergs.

[19] Capsize may aid the survival of debris within icebergs over the long drift trajectories across the North Atlantic. When debris-laden ice is at or near the iceberg base, as is the case for icebergs that fail to capsize, it is vulnerable to basal melting, which can be as large as 10 m/a or more [e.g., *Rignot and Jacobs*, 2002]. When icebergs capsize (panel 3 of Figure 2), debris is moved away from the fast-melting, and thus fast-deposition, interface. The Larsen B capsize appears to be primarily by 90° but rotation

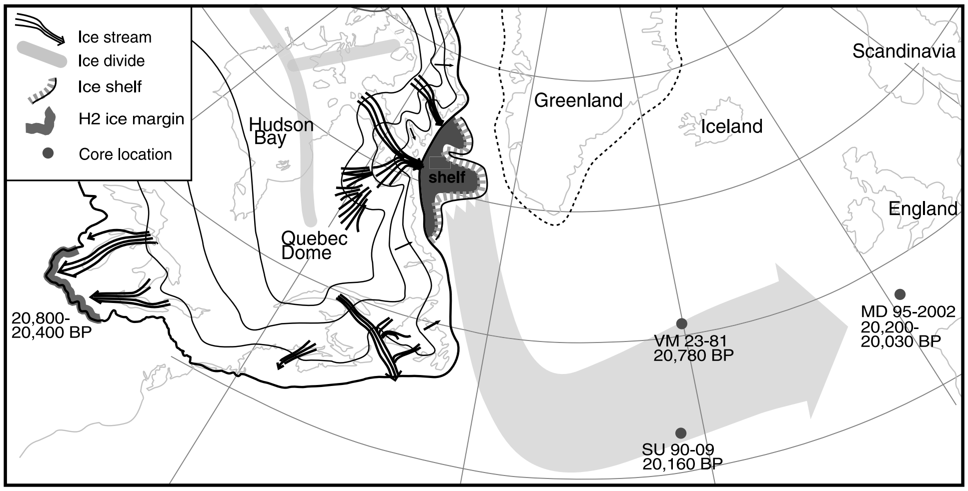

Figure 3. Sketch map of LIS at the time of H2 showing a fringing ice shelf along the eastern seaboard of Canada that is reinforced by an ice tongue extending from the outlet of Hudson Strait. Dates for the H2 advance of the ice margin in the Great Lakes region and for H2 IRD layers in deep-sea cores are given in ^{14}C years BP. Broad gray swath in the center of the North Atlantic indicates the approximate area of the ice-rafting belt. Typical radiocarbon dates (in ^{14}C years BP) used to estimate the timing of LGM Laurentide advance in the Great Lakes region are: Center School House, 20,500 ± 130 ISGS-89; 20,030 ± 150 ISGS-2921; 20,090 ± 150 ISGS-2922; 20,440 ± 160 ISGS-2923; 21,950 ± 180 ISGS-2924; 19,480 ± 100 AA-8458; 20,360 ± 110 AA-8459; 20,360 ± 110 A-8459; Charleston, 19,340 ± 180 ISGS-2918; 19,500 ± 200 ISGS-27; 19,980 ± 150 ISGS-2842; 20,050 ± 170 ISGS-2593; 20,666 ± 170 ISGS-2919; 21,300 ± 200 ISGS-28; Oxford Cuts, 20,030 ± 140 PITT-0625; 20,620 ± 180 PITT-0624; 20,820 ± 210 ISGS-2757; 20,800 ± 250 ISGS-2758; 20,800 ± 210 ISGS-2760; 20,800 ± 200 ISGS-2761; 20,770 ± 210 ISGS-2763; 20,850 ± 200 ISGS-2762; 21,240 ± 150 PITT-0765; 21,390 ± 200 PITT-0764; Cincinnati, 19,200 ± 140 PITT-0508; 19,310 ± 170 PITT-0506; 19,690 ± 150 PITT-0509; 19,960 ± 170 PITT-0227; 20,200 ± 140 PITT-0507; Sidney Cut, 22,480 ± 800 W-356; 23,000 ± 800 W-188; 22,700 ± 80 Beta-144872; 22,650 ± 80 Beta-144874; Chillicothe, 17,590 ± 210 UGa-6709; 18,240 ± 180 UGa-6711; 18,490 ± 280 UGa-6710; 18,520 ± 280 UGa-6708; 18,750 ± 260 UGa-6706; 18,800 ± 290 UGa-6707; Garfield Heights, 22,210 ± 120 DIC-32; 23,313 ± 391K361-3; 23,430 ± 410 DIC-63; 23,560 ± 610 DIC-35; 24,520 ± 695 DIC-38; 24,600 ± 800 W-71; 28,195 ± 535 K361-4. Sources of information and data reduction techniques can be found in the work of *Lowell et al.* [1999].

up to 180°, that is, rolling bottom-side up, is also possible. In the 180° case, relatively dark debris, distributed through the basal ice, would absorb solar energy and slowly melt down into the iceberg (i.e., cryoconite holes; *Gribbon* [1979]). Debris moving downward into an iceberg would be protected and carried far a field. When icebergs capsize by 90°, debris-laden ice is distributed vertically through the ice column. The debris would then be melted out and deposited gradually, over the lifetime of the iceberg. Capsize, which is common for icebergs created by the Larsen-style disintegration, promotes the widespread transport of debris far from the immediate source of the icebergs, as is required by the geologic record.

3. Source of Heinrich Layer Sediments

3.1. Northeast LIS Flow at the LGM

[20] At the last glacial maximum (LGM), discharge from the northeastern sector of the LIS was primarily through five fast-flowing ice streams (Figure 3). Guided by bedrock topography, the streams were separated by slowly flowing interstream areas. The radially arranged streams were fed by five corresponding ice sheet catchments: Boothia, Cumberland, Hudson Strait, Ungava Bay and St. Lawrence (Figure 4). It has been suggested that massive surge events originating within the entire catchment area of the Hudson Strait outlet of the LIS could have been the source of Heinrich icebergs [*MacAyeal*, 1993]. This suggestion has been the subject of much debate, at issue are both the observational and dynamical plausibility of a massive Hudson Strait surge being the mechanism behind the Heinrich event.

[21] Indeed, recent evidence from the present day, multi-stream drainage system of the West Antarctic Ice Sheet (WAIS) suggests that the thermodynamics of the system may mitigate rapid drawdown of ice sheet-interiors. The Ross Sea sector of the WAIS is channeled into broad, fast-flowing ice streams with modest topographic control. The WAIS ice streams are fed by a network of smaller tributaries which follow well-defined bedrock valleys [*Joughin et al.*, 1999]. Streaklines observed in the surface of the Ross Ice Shelf document repeated cycles of outflow redirection over the last millennium, which have been interpreted to indicate that drawdown rates at individual ice stream outlets are regulated by internal, thermal processes (C. L. Hulbe and

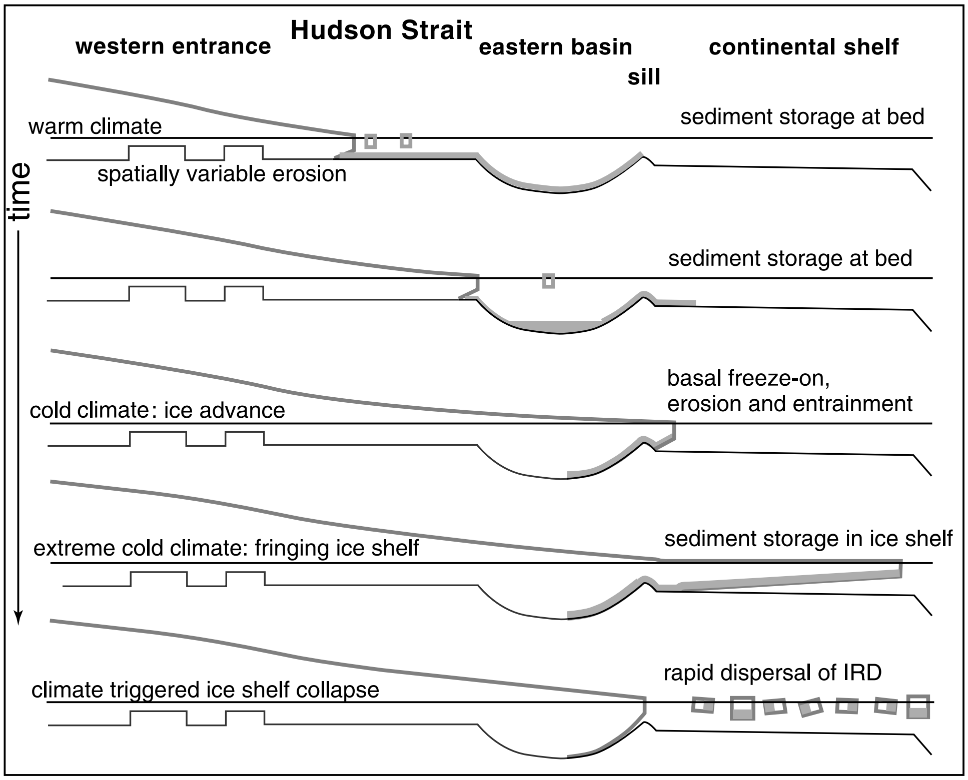

Figure 4. At the LGM, ice streams with the potential to feed ice shelves in the Labrador Sea and Baffin Bay existed in Gulf of Boothia-Lancaster Sound, Cumberland Sound and Hudson Strait. Our interpretation of ice stream trajectories are denoted with heavy black lines. Sediment-source rock types are mapped and their relative contributions to the ice shelf (shelves) are predicted in the colored bars at the downstream end of each ice stream.

M. A. Fahnestock, West Antarctic ice stream discharge variability: Mechanism, controls, and pattern of grounding line retreat, submitted to *Journal of Glaciology,* 2003, hereinafter referred to as Hulbe and Fahnestock, submitted manuscript, 2003). In brief, large downstream extension, and thus thinning, causes the basal temperature gradient to increase faster than can be accommodated by diffusion, so that the fastest-flowing, and thus fastest-thinning, ice refreezes the basal water that facilitates its fast flow, effectively limiting the drawdown rate. The underlying physics is the same as in the binge-purge paradigm [*MacAyeal*, 1993] but the relevant timescale is much shorter and a the ice stream has an internal regulation against runaway purges.

[22] With the geologic evidence (discussed in detail below) and the modern WAIS analogue in mind, we understand the northeast sector of the LIS to be relatively stable at the LGM, the time of the Heinrich 2 event (H2). Geomorphologic evidence for mosaics of frozen and thawed bed beneath northern ice sheets is well documented from Scandinavia [*Kleman et al.*, 1999; *Fabel et al.*, 2002] and similar evidence exists for the northeastern LIS [*Kleman and Borgström*, 1996]. Because the major LIS ice streams were clearly topographically guided, we regard it as likely that they were persistent features in the ice sheet, their discharge varying in concert with climate and regional mass balance and not only as a result of internally driven discharge fluctuation.

3.2. The Ice Streams

[23] At the LGM, Laurentide ice streams with the potential to feed ice shelves in Baffin Bay and the Labrador Sea existed in Gulf of Boothia-Lancaster Sound, Cumberland Sound and Hudson Strait (Figure 4). By our hypothesis, ice derived from these streams would eventually deposit the H2 layer. The evidence for these streams is mainly in the form of striae on the bedrock surface and is of variable quality.

[24] The prime evidence for the Gulf of Boothia ice stream is the strong convergence of flow traces on the landmasses adjacent to southern and central Gulf of Boothia (J. Kleman, unpublished map, 2002). The traces define the upstream convergence of a northeast flowing ice stream [*Stokes and Clark*, 1999]. This outlet would have received a minor contribution from a smaller ice stream in Admiralty Inlet, which is defined by a scoured zone with lineation convergence at the head of the inlet. At the eastern, downstream, end of Lancaster Sound, the Gulf of Boothia Ice Stream would have been free to expand laterally into a Baffin Bay ice shelf. A zone of strong scouring and east trending glacial lineations in the lowland between Cumberland and Hall Peninsulas is direct evidence for a Cumberland Sound ice stream. The floor of Cumberland Sound is

heavily eroded, with water depths up to 1300 m [*Jennings*, 1993].

[25] Direct geological evidence for a Hudson Strait ice stream is sparse, consisting of relatively few observations of glacial striae on well-separated islands along the margins of the Strait. We present our preferred interpretation of those features but recognize that other interpretations are possible [e.g., *Veilette et al.*, 1999; *Clark et al.*, 2000, *Jansson et al.*, 2002]. East trending striae are documented from Nottingham and Salisbury islands at the western end of the Strait [*Laymon*, 1992], from Charles and Big Islands in the central section [*Gray*, 2001], and from Resolution and Button Islands at the mouth of the Strait. Glacial geological mapping of southern Baffin Island [*Manley*, 1996; *Kleman et al.*, 2001] gives no evidence for east trending striae or till lineations. However, *Gray* [2001] suggests that striae at some mid-Strait locations indicate that southward flow off Baffin Island and northward flow off Ungava Peninsula was deflected eastward into a Hudson Strait ice stream.

[26] Striae cannot be assigned absolute dates, so it is not possible to state definitively that an ice stream extended along the entire length of Hudson Strait at any given time. Again, we may learn from a modern analogue for a multi-stream ice sheet, the Ross Sea sector of the WIAS. Several lines of evidence, including the Byrd surface camp ice core and ice sheet internal layers, indicate that the present ice stream drainage configuration, or one much like it, has existed since at least the LGM [*Nereson and Raymond*, 2001; *Steig et al.*, 2001]. The thermodynamic outlet regulation noted above would promote a steady, nonpurging discharge system. Given the topographical context, we favor the existence of a persistent ice stream in the Eastern Basin of the Strait, occasionally extending headward to Nottingham and Salisbury Islands. Upstream, the western entrance to Hudson Strait is topographically complex. Several islands and shoals created by block-faulting have relative relief up to 600 m. A tributary system similar to the present west Antarctic system may have flowed through this relatively rugged terrain and further headward propagation of the ice stream into the LIS may not have been possible.

[27] Other, smaller outflows also drained the northeast sector of the ice sheet at the LGM. The direct evidence for a Cabot Strait ice stream is in the seafloor morphology, primarily a very wide and streamlined trough leading to the shelf edge. Onshore evidence is yet to be found, and possibly has been eroded or was never created. Small ice streams in marine inlets or embayments, in Admiralty Inlet, Eclipse Sound, Frobisher Bay and Notre Dame Bay, are defined on the basis of head convergence zones and topographical context. Only two minor ice streams appear to have traversed high ground. One on the Labrador coast appears to have existed inland of Hopedale. Large till lineations and conformable striae, overprinted obliquely by deglacial flow traces, can be traced to south of Smallwood Reservoir. A single minor ice stream traversed central Baffin Island, northwest of Cumberland Peninsula. It exploited a low point in the topographic backbone of Baffin Island, with the prime evidence for its existence being plumes of carbonate-bearing till derived from the western coast of Baffin Island or Foxe Basin [*Tippett*, 1985].

[28] In contrast to the large, apparently fast-flowing ice streams found elsewhere, flow across the Baffin Island Coast between Bylong Island and Cumberland Peninsula and the Labrador Coast was accomplished by numerous small streams in fjords and inlets. At the scale of the present analysis, ice motion in these regions can be thought of as diffuse ice sheet flow.

3.3. Interstream Areas and Frozen-Bed Zones

[29] The relatively high ground between ice stream corridors corresponds to areas of frozen basal conditions where we expect a high ice surface, convex shape, and slow, divergent flow. Such conditions are depicted on Borden and Brodeur peninsulas, along most of the Baffin Island backbone, on Cumberland, Hall and Meta Incognita Peninsulas, and in the Torngat Mountains. Central Newfoundland and the Gaspé Peninsula experienced similar conditions. Relict landscapes, indicative of little or no glacial erosion, are found in all these areas, with the exception of central Newfoundland [*Sugden and Watts*, 1977; *Grant*, 1989]. In many cases, delicate nonglacial landforms such as fesenmeer and tors survived at least the last glaciation without any sign of glacial modification, a clear indication of frozen-bed conditions [*Kleman and Borgström*, 1996].

3.4. Sediment Transport to Ice Shelves in the Labrador Sea and Baffin Bay

[30] In order to produce a Heinrich layer upon disintegration, our hypothetical ice shelves must be charged with lithic detritus of suitable origin. We make a qualitative prediction the provenance and relative proportion of basal debris in the major northeast LIS using a map of bedrock provinces (compiled from *Gwiazda et al.* [1996], *Grant* [1989], and *MacLean* [2001]) and three simple assumptions. First and foremost, we assume that the mapped bedrock distribution is a proxy for the lithological composition of the material to be entrained in basal ice. Because ice sheet drainage pathways were guided by bedrock topography, flow direction along each ice stream path can be expected to have been stable and the problem of multicycle transport and remobilization of old till is modest. Old tills available for recycling would reflect primarily the same source areas as freshly eroded material. Second, we recognize that the effective catchment size for basal debris is smaller than the corresponding ice sheet catchment size, because ice near catchment boundaries is relatively inactive. Third, although modest erosion and entrainment does occur under frozen-bed conditions [*Cuffey et al.*, 2000], the bed beneath the main channel of a fast-flowing stream would yield more sediment per unit area than would zones peripheral to it.

[31] The conclusion of our provenance analysis is simple: the debris stored by Labrador Sea ice shelves would have been dominated by Churchill province rocks with isotope ages in the range 1.7–1.9 Ga, and Paleozoic carbonate rocks (see insets of Figure 3 depicting sediment source compositions). Only in Central Quebec and on the Ungava Peninsula did the effective catchment basins reach into the Superior province. The Boothia and Cumberland Sound Ice streams were likely to have carried exclusively Churchill

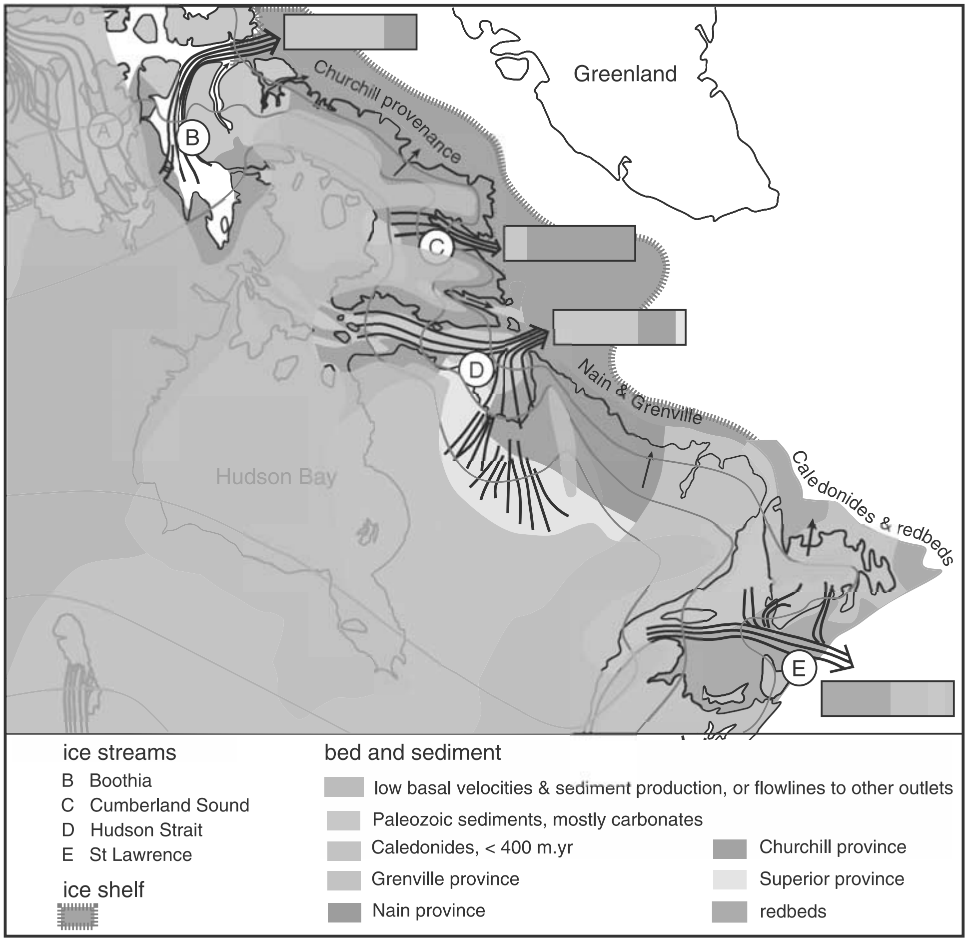

Figure 5. Ice shelf advance and collapse scenario, in cross-section view through Hudson Strait. During warm climates (upper panel) the Hudson Strait ice stream accumulates basal sediment that is subsequently dropped proximally, in a fore-deepened basin near the mouth of the strait. During cold and extreme cold climate (lower panel), ice advances across the area of accumulated sediments, picking up basal debris, and forms an ice shelf in the Labrador Sea. However, the shelf is vulnerable to summer-season surface melting. When climate ameliorates enough to allow repeated seasons of pervasive surface melting, the ice shelf disintegrates, sending an armada of debris bearing icebergs into the North Atlantic (bottom panel). See color version of this figure at back of this issue.

province and Paleozoic sedimentary rocks. The Hudson Strait ice stream would have carried a minor component of Superior province material due to flow from the Ungava Bay drainage. The proportion of Superior sediments would have increased at times of headward expansion of Ungava Bay ice streams. The effective catchment for the Cabot Ice Stream comprises comparably sized areas of the Greenville province, Caledonides, Paleozoic sediments, and red beds, yielding a provenance signature that is quite distinct from the drainages to the north.

4. Revised Ice Shelf Mechanism for Heinrich Events

[32] The Heinrich event scenario proposed here derives from the four-step process originally described by *Hulbe* [1997]. We are able to improve upon that hypothesis thanks to the ice shelf disintegration events recently observed in Antarctica and to glacial geologic evidence for the state of the cryosphere during the last glaciation. A schematic diagram of events leading to a Heinrich event, as seen near the Hudson Strait outlet, is provided in Figure 5.

[33] We posit that the extremely cold episodes during the last glaciation would have favored enhanced seaward flow of Laurentide ice across the grounding line. Similarly, low sea-surface temperatures would have aided the expansion of floating ice beyond the grounding line. As a consequence, ice tongues, fringing ice shelves, and perhaps an embayed ice shelf, would have grown in the Labrador Sea and Davis Strait. The coastal sedimentary record supports the existence of coastal ice shelves at Heinrich event times [*Rashid et al.*, 2003]. The relatively slow rate of influx into Baffin Bay makes it unlikely that more than coastal fringing ice shelves grew there. It is important to note that given appropriate sea-surface conditions, ice shelves could have been sustained by steady drainage from the LIS, without

appeal to ice stream surges. Model simulations indicate that the timescale for ice shelf expansion in the due to steady state discharge from the LIS would have been several hundred years [*Hulbe*, 1997, Figure 2]. While not required by the present scenario, it is glaciologically possible for an embayed shelf to grow within a century to millennial timescale that could be accommodated by the foraminiferal record [e.g., *Aksu and Mudie*, 1985].

[34] Once delivered, ice remains within the shelf for centuries before it arrives at the seaward front to be recycled by iceberg calving. The floating ice would have received and stored englacial (mainly basal) sediment from its source catchment basins on land, as discussed in section 3. Debris-rich basal ice within a shelf would have been either lost to melting or protected by basal freeze-on. Underplating by marine ice may occur by two mechanisms. First, where meltwater rises along the underside of the shelf, it may become supercooled, and refreeze [*Bombosh and Jenkins*, 1995]. Second, rapid thinning of grounded ice in a fast-flowing outlet glacier steepens the temperature gradient deep in the ice, thereby favoring basal freezing once the ice is afloat (following ideas presented by Hulbe and Fahnestock (submitted manuscript, 2003)). During the extreme cold stadial conditions under which the ice shelf advanced, large tabular icebergs, calved episodically from a stable ice shelf front, would have drifted slowly southward, dropping debris along the region proximal to the North American glacial system.

[35] Ice shelves along the eastern Canadian seaboard would likely have been susceptible to the same disintegration mechanism now active along the Antarctic Peninsula. Catastrophic disintegration event(s) would have been initiated whenever summer warming was sufficient to produce surface melting. Over the course of a few years to decades, surface meltwater would flood the snow and firn, fill crevasses, and eventually cause the crevasses to penetrate to the bottom of the floating ice. By this mechanism, fringing ice shelves in the Labrador Sea could disintegrate rapidly, soon after climate warming increased the length and intensity of summer melt season. The disintegration may have been en masse or could have advanced stepwise along prevailing climate gradients (e.g., from south to north) over a few decades, as is the case today, for the Larsen ice shelves [*Scambos et al.*, 2000].

[36] Ice shelf disintegration events would have generated a large quantity of small, capsized icebergs, nearly instantaneously. So produced, the icebergs would have been of a size and shape to drift readily with the prevailing winds; eastward to the limits of the North Atlantic gyre. Capsize would have redistributed englacial debris such that it would survive long transport distances within continually melting icebergs. Laurentide detrital carbonate sediment, accumulated slowly in the ice shelf, would thus have been dispersed abruptly across the North Atlantic Ocean in an armada of innumerable icebergs, delivering a widespread deposit to the seafloor.

5. Discussion

[37] Does the proposed mechanism of sudden ice shelf disintegration described in the preceding sections fit what is now known about Heinrich events? There are two particularly important aspects to this question. One involves the physical characteristics of Heinrich layers. The other is the relation of Heinrich events to the climate oscillations of the last glaciation.

5.1. Sedimentary Record

[38] The distribution of dolomite and limestone bedrock outcrops shows that the ultimate source of Heinrich sediments lies beneath the former LIS [*Andrews and Tedesco*, 1992; *Bond et al.*, 1993; *Gwiazda et al.*, 1996; *Hemming et al.*, 1998], particularly in the region drained through Hudson Strait. Ice flowing from that ice sheet outlet would have charged the ice shelf with basal sediment derived from regions of Churchill rocks, as well as detrital dolomite and limestone carbonate. Sediments melted off the base of the ice shelf close to the grounding line would have cascaded down the outer slope of the continental shelf. The resulting turbidites and nepheloid flows explain the exceptional thickness and sedimentary characteristics of Heinrich layers on the floor of the Labrador Sea [*Hasse and Khodabakhsh*, 1998]. The debris-laden icebergs produced from disintegration of the ice shelf would have also contributed to deposition in the Labrador Sea. Most important, icebergs produced from Hudson Strait outflow into the fringing ice shelf would have be the sole source for Heinrich carbonate layers in the open North Atlantic Ocean [*Hasse and Khodabakhsh*, 1998]. Thus our ice shelf hypothesis fits well with the provenance and source-proximal deposition of the Heinrich detrital carbonate layers.

[39] The formation of a Heinrich carbonate layer requires melting across the North Atlantic of thousands of icebergs produced by a massive, short-lived glaciological discharge operating in eastern Canada and the Labrador Sea [*Bond et al.*, 1992]. The explosive formation of thousands of debris-laden bergs by disintegration of a Labrador Sea ice shelf, and their rapid wind-driven transit across the North Atlantic Ocean, would produce the sharp basal contacts of the carbonate layers. Further, the capsize of the icebergs, a characteristic of the newly recognized, Larsen-style ice shelf breakup, would rotate englacial debris toward a vertical distribution, so that the bergs would have to melt entirely before all the debris loads were discharged onto the ocean floor. The improved survivability of the entrained IRD, together with the rapid transatlantic passage of the bergs, explains the extraordinary length of the ice-rafted carbonate tongues.

[40] Our hypothesized ice shelf must also be capable of delivering the volume of IRD now resting on the north Atlantic seafloor. *Alley and MacAyeal* [1994] calculated that, in a typical Heinrich event, about 100 km^3 of terrigenous material was transported by floating ice. Typical debris content in basal glacier ice ranges from 5 to 35%, although concentrations of up to 60% have been measured where freeze-on rates are large [*Lawson et al.*, 1998]. Using the more conservative range, we estimate that 285 to 2000 km^2 of dirty ice is needed but note that the smaller volume may be more likely, as freeze-on would have been vigorous along the adverse slope beneath ice flowing seaward through Hudson Strait. Meltwater moving rapidly

upslope may become supercooled and refreeze, entraining sediment in the process [*Alley et al.*, 1998]. Rapid downstream thinning and ensuing steepening of the basal temperature gradient may also drive basal freeze-on, as has been speculated about West Antarctic ice streams (Hulbe and Fahnestock, submitted manuscript, 2003). Accreted ice observed at the bases of those fast-flowing ice streams reaches 2% of the total ice thickness (H. Engelhardt, personal communication, 2001). Recognizing that both freezing and melting occurred at the base of the stream and shelf into which it flowed, we estimate that accreted dirty ice made up 1% of the total ice thickness. The minimum required ice shelf volume is then in the range of 28,000 to 200,000 km^3. For comparison, the volume of ice converted to icebergs in the 2002 Larsen B breakup was about 748 km^3 and the volume of ice contained in the very large Ross Ice Shelf is about 500,000 km^3. *Hulbe* [1997] estimated that a steady state ice shelf covering the Labrador Sea, fed by steady LGM discharge of the LIS, would have had a volume of 750,000 km^3. This accounting implies that ice shelves fringing the eastern Canadian seaboard would be a sufficient source for the armada of icebergs necessary to produce a Heinrich event.

[41] If the ice shelf hypothesis presented here is correct, Heinrich sediments may be used to infer which boundaries of the LIS supported ice shelves and which did not. An ice shelf restricted to areas north of the Churchill-Nain contact, which is located approximately 100 km south of Hudson Strait, would not incorporate any Nain or Greenville province material. Thus we may speculate that either ice influx rates were insufficient or local climate and sea temperature were too warm to support a significant ice shelf south of the Churchill-Nain contact.

5.2. Climate Record

[42] Heinrich events are consistently locked to the cold culminations of the saw-toothed Bond climate cycles (Figure 6). This connection takes on increased significance now that these cycles are known to have a hemisphere-wide footprint. First recognized in Greenland ice core records, the cycles have been identified in many paleoclimate proxy records. These include: northern North Atlantic sea-surface temperature [*Bond et al.*, 1993], Cariaco Basin major element chemistry (which tracks the ITCZ in the tropical Atlantic [*Peterson et al.*, 2000]), lake-level fluctuations in the northern Great Basin [*Zic et al.*, 2002], and circulation changes in the Santa Barbara Basin on the California margin [*Hendy and Kennett*, 1999].

[43] Also important is the recognition that Heinrich events correlate with ice sheet and mountain-glacier maxima in North America, Europe and southern South America. For the southern margin of the Cordilleran Ice Sheet, the Coguitlam Stade, dated between 21,300 ± 250 (GSC-3305) and 18,700 ± 170 ^{14}C years BP (GSC-2344), corresponds to the time of H2, and the Vashon Stade, dated between 15,610 ± 130 (Beta-11057) and 13,500 ± 220 ^{14}C years BP (GSC-3124) [*Hicock et al.*, 1999]. Moreover, the Vashon Stade reached its maximum extent between 14,450 ± 90 (CAM-23160) and 13,700 ± 150 ^{14}C years BP (QL-4067 [*Porter and Swanson*, 1998] coeval with a pulse of ice-rafted debris in core PAR-85-01 from west of Vancouver Island which shows a marked increase sand grains after 15,280 ± 170 ^{14}C years BP (RIDDL-808) [*Blaise et al.*, 1990].

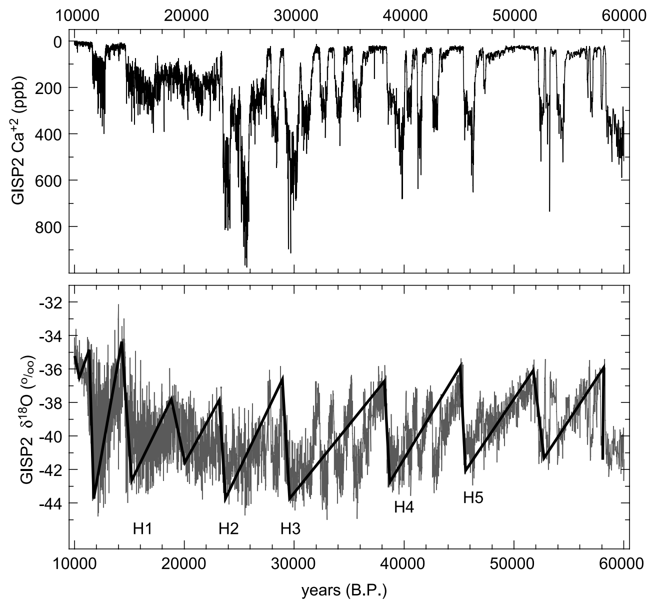

Figure 6. The Ca^{+2} and the $\delta^{18}O$ records from the GISP2 ice core (upper panel and lower panel, respectively). Data are adapted from *Mayewski et al.* [1997] and *Stuiver and Grootes* [2000] using the GISP2 timescale [*Meese et al.*, 1997]. Superimposed on the $\delta\ ^{18}O$ record (heavy black line) is the qualitative pattern of change referred to as the Bond cycle [*Bond et al.*, 1993]. The placement of the Heinrich events, labeled H1-H5, in the lower panel follows *Bond et al.* [1993]. Note that H2 and H3 are accompanied by particularly strong Ca^{+2} spikes. Cooling events between H4 and H3 and between H3 and H2, some coeval with pronounced Ca^{+2} spikes, are not accompanied by ice-rafted debris pulses from the LIS.

[44] The LGM advance of the LIS south of Lake Michigan in central Illinois was achieved at 20,251 ± 53 ^{14}C years BP, as is evident from 6 newly obtained ^{14}C dates (Figure 3). Farther east, the Laurentide Ice Sheet overran two sites in Oxford, Ohio, at 20,770 ± 58 ^{14}C years BP (average of 10 dates, *Lowell et al.*, 1999). For comparison, the H2 detrital-carbonate peak in North Atlantic core VM-23-81 dates to 20,780 ± 170 ^{14}C years BP [*Bond et al.*, 1997, 1993], and in core SU-90-09 it has an age of 20,160 ^{14}C years BP [*Grousset et al.*, 2000]. In North Atlantic core OMEX-2K near Ireland, the peak of Canadian ice-rafted debris dates to 20,360 ± 140 ^{14}C years BP (KIA-8072) [*Scourse et al.*, 2000], and in North Atlantic core MD-95-2002 it is bracketed by ages of 20,200 ± 80 and 20,030 ± 80 ^{14}C years BP [*Grousset et al.*, 2000].

[45] Evidence of Andean piedmont glacier advance into the outer Llanguihue moraine belt in the Chilean Lake District is unequivocal at the times of the H1 and H3 events in the North Atlantic Ocean, and exists for H2 in somewhat less certain form. The ages of these Andean maxima are

14,550 ± 54 to 14,882 ± 72 ^{14}C years BP (sites 23, 26, 28, 30, 47, 98 of *Denton et al.* [1999, Table 2]); 20,800 ^{14}C years BP, and 26,437 ± 84 ^{14}C years BP (site 16 of *Denton et al.* [1999, Table 2]). As a reminder, core VM-23-81 places the H1 carbonate peak at 14,100 ^{14}C years BP, the H2 peak at 20,780^{14}C years BP, and the Heinrich 3 peak at 26,270 ^{14}C years BP [*Bond et al.*, 1997, 1999].

[46] Other cold-climate indicators also keep pace with Heinrich events. Snow line depression in the Chilean Andes during the time of the North Atlantic H2 event was about 1000 m from its present elevation of 1900–2100 m [*Porter*, 1981; *Hubbard*, 1997]. Numerous pollen sites within 200 m elevation of present-day sea level along nearly 2° of latitude uniformly show an open Subantarctic Parkland environment during the time of the H2 event [*Heusser et al.*, 1999]. The paucity of trees suggests that these low-lying sites were all then near the tree line. The implication is that tree line elevation lowered about 1000 m or more from its present-day limit at close to 1250 m in the adjacent Andes. From the long and detailed pollen record at Taiquemó on Isla Grande de Chiloé, *Heusser et al.* [1999] showed that the maximum lowering of tree line elevation, and by inference the coldest mean summer temperatures, occurred close to 21,430 ± 200 ^{14}C years BP (AA-17985). Thus the maximum advances of Andean piedmont glaciers in the southern Lake District (and hence maximum snow line depression), the greatest lowering of Andean tree line elevation, and the greatest temperature depression at Taiquemó all correlate with the H2 interval in the North Atlantic Ocean.

[47] A plausible interpretation of these paleoclimate records is that Heinrich events are the final result, not the cause, of the culminating cold episode of the long declining limb of Bond climate cycles (Figure 5). This conclusion is reinforced by the massive snow line lowering in the Chilean Andes that accompanied the last three North Atlantic Heinrich events. As pointed out by *Bond et al.* [1992, 1999], a climate driver for Heinrich events is also consistent with the early ocean surface cooling and precursor ice-rafting events associated with detrital-carbonate layers in the North Atlantic Ocean. In marine sediment cores VM-23-81, DSDP-609, and MD-95-2002 near the eastern end of the North Atlantic ice-rafting belt, the detrital carbonate layers are embedded in a sediment pulse derived not only from Laurentide but also from Icelandic and/or European sources [*Bond and Lotti*, 1995; *Bond et al.*, 1999; *Grousset et al.*, 2000]. In all three cases, the deposition from non-Laurentide sources preceded that of Laurentide detrital carbonate. In addition, *Bond et al.* [1992] pointed out that cooling of the ocean surface preceded each Heinrich event.

[48] The Heinrich carbonate spikes have a distinct placement within the IRD depositional sequence that also speaks to their origin. Carbonate IRD is conspicuously absent in the North Atlantic during the time of the cold spikes in Greenland stable isotope records prior to the H2 and H3 events. The Greenland oxygen isotopic record (Figure 6) shows that these early cold spikes are nearly as extreme as the subsequent spikes accompanied by Heinrich events. These early spikes also show up as lithic pulses in marine sediment cores from the eastern end of the North Atlantic ice-rafting belt [*Bond and Lotti*, 1995; *Bond et al.*, 1997, 1999; *Grousett et al.*, 2000; *Scourse et al.*, 2000] (Figure 3). In core VM-23-81 the early pulse prior to the H2 carbonate spike is dated to 22,400 ^{14}C years BP [*Bond et al.*, 1999]. In core OMEX-2K it is dated to 22,130 ^{14}C years BP [*Scourse et al.*, 2000] and in core MD-95-2002 to 21,850 ^{14}C years BP [*Grousset et al.*, 2000]. The earlier spike is dated to 29,000 ^{14}C years BP in core VM-23-81 [*Bond and Lotti*, 1995]. In both the Great Lakes region and the Chilean Andes the best available geochronology suggests that the most significant ice expansion toward maximum conditions occurred at 22,500 ^{14}C years BP; precisely within the temporal context of the early spike before the H2 event. The implication of these Heinrich event-free early spikes is that Bond cold episodes need not be forced by interior Laurentide ice dynamics.

[49] Although we infer that Heinrich events are paced by climate events superimposed on the eastern sector of the LIS, we are quick to note that the resulting armada of icebergs can dramatically jolt the climate system by curtailing the formation of North Atlantic Deep Water (NADW). Theories have been developed around the idea that large releases of icebergs into the North Atlantic associated with Heinrich events have had a strong influence on NADW production and associated sea-surface temperature (SST) [e.g., *Rahmstorf*, 2002; *Bard et al.*, 2000]. These theories may be equally well supported by the ice shelf breakup mechanism proposed here as the source for these icebergs as by an ice stream source such as proposed in previous hypotheses described in section 1.

5.3. Climate and Ice Shelves

[50] The ice shelf disintegration hypothesis is consistent with the known climate record summarized in section 5.2. Heinrich events correspond with the most severe cold pulses of the last glacial cycle, precisely the condition that would have allowed ice shelf growth on the eastern Canadian seaboard. The seaward flow of Laurentide ice across the grounding line would have been at a maximum at that time. Sea-surface temperatures would then be at a minimum. Eastern Canadian ice shelves of significant volume would have been possible only under conditions of extreme lowering of atmospheric and sea-surface temperatures. This is crucial to our argument. There would be no Heinrich event without the formation of an ice shelf. Heinrich layers were not deposited during precursor cold spikes of the Dansgard-Oeschger oscillations simply because climate was not extreme enough to sustain ice shelves. Once in existence, the Labrador Sea ice shelf (or shelves) would have disintegrated catastrophically whenever climatic warming produced enough surface meltwater to saturate the firn and fill crevasses. Such disintegration could await terminal warming at the end of a cold pulse, i.e., the onset of the strong interstadials that follow the culminating cold event of a Bond cycle. Or the appropriate conditions could occur during a decades-long summer warming episode within an otherwise cold interval. In fact, it is possible that fringing Canadian ice shelves could form and disintegrate several times during a Heinrich event.

[51] The correlation of LGM ice advance of the southern margin of the LIS with the H2 event favors the ice shelf

hypothesis. Both the ice lobes in the Great Lakes region and the ice stream in Hudson Strait on the eastern Canadian seaboard are fed by the Labradorean sector of the LIS [*Denton and Hughes*, 1981]. A simple interpretation of chronologic data given in section 5.2 is that the Labradorean sector was at its most robust phase at 20,400 ^{14}C years BP. In the Great Lakes region, this was reflected by ice lobes reaching a maximum extent. Near the mouth of Hudson Strait, this was reflected by the advance of the grounding line onto the continental shelf and by the formation of a fringing ice shelf (or shelves) similar to present-day ice shelves along the Antarctic Peninsula. Floating ice tongues seaward of the Laurentide ice streams could have mechanically stiffened the fringing shelves, as in the modern Brunt Ice Shelf and Stancomb-Wills Glacier system on the eastern side of the Weddell Sea [*Thomas*, 1973]. Disintegration of this shelf, either en masse or in a sequential, stepwise fashion (e.g., as in the recent disintegrations of Larsen A and B ice shelves of the Antarctic) would result in deposition of the North Atlantic H2 carbonate layer.

6. Conclusion

[52] We close with a few remarks about the general implications of the ice shelf disintegration hypothesis for Heinrich events. It is important that this hypothesis does not postulate collapse of the main body of the LIS. Surges of ice streams into the Labrador Sea are not required, nor are they proscribed. This is in contrast to the "binge/purge" scenario [*MacAyeal*, 1993], in which periods of quiescence are punctuated by extraordinary draw-down events, for which there is no obvious evidence elsewhere around the Labradorean sector of the LIS. Instead, we move the catastrophic event seaward of the grounding line and Heinrich events become the product of a robust, not collapsed, ice sheet. In such a case, sea level jumps would not accompany the Heinrich events themselves. Rather, they would occur during the abrupt warming episodes that terminate each Bond cooling cycle, when equilibrium lines would rise rapidly on the surfaces of the large ice sheets that persisted through the Heinrich events.

[53] By the proposed hypothesis, climate change drives the ice shelf component of Heinrich events. This is because the ice shelves that lie at the heart of the hypothesis are dependent on atmospheric cooling for their formation and warming accompanied by surface melting for their sudden disintegration. Wholesale ice shelf breakup would affect the climate system immediately, as the resulting armada of icebergs brought a fresh water flux to the North Atlantic that could depress deep water formation [e.g., *Rahmstorf*, 2002]. Unfortunately, the proposed hypothesis leaves unanswered the fundamental question of what could have caused the dramatic climate changes associated with Heinrich events. In this regard, it should be noted that abrupt climatic change is not a prerequisite for Heinrich events. As is the case with the Larsen A and B ice shelf explosions, slow climate warming could simply reach a threshold in which sufficient surface meltwater is produced to disintegrate a fringing ice shelf on the Canadian seaboard.

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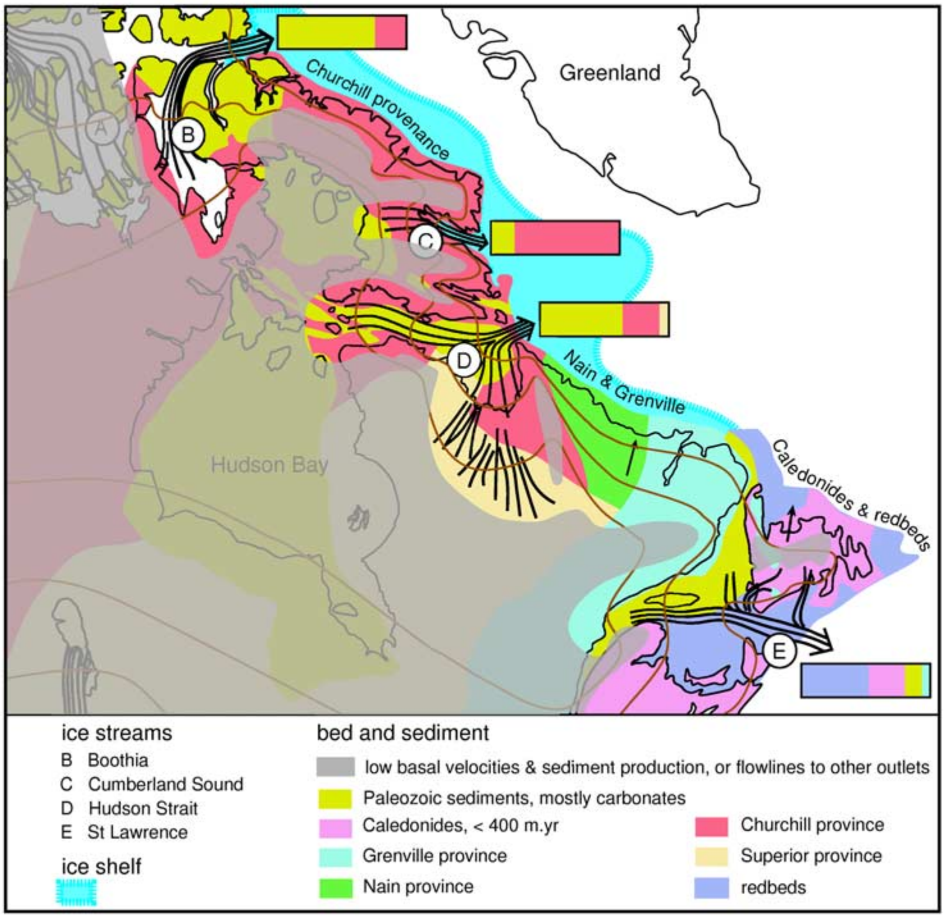

Figure 5. Ice shelf advance and collapse scenario, in cross-section view through Hudson Strait. During warm climates (upper panel) the Hudson Strait ice stream accumulates basal sediment that is subsequently dropped proximally, in a fore-deepened basin near the mouth of the strait. During cold and extreme cold climate (lower panel), ice advances across the area of accumulated sediments, picking up basal debris, and forms an ice shelf in the Labrador Sea. However, the shelf is vulnerable to summer-season surface melting. When climate ameliorates enough to allow repeated seasons of pervasive surface melting, the ice shelf disintegrates, sending an armada of debris bearing icebergs into the North Atlantic (bottom panel).