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# Centurial-millenial ice-rafted debris pulses from ablating marine ice sheets

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**Abstract.** We use an ice-sheet model to show that (i) margins of marine ice-sheets can be expected to be frozen to the bed, except where ice-streams discharge; (ii) 20-50km retreats induced by ablation rates of 2 m/yr provide sufficient debris flux through the grounding line to produce large sedimentation events. Such ablation would reduce ice-shelf extent markedly, permitting debris to reach the calving front and be transported by icebergs leading to ice-rafted debris (IRD) events. Ice shelf break-up takes around a century (start of IRD pulse), while the creation of warm-based conditions (end of IRD pulse) due to upwards motion of warm ice takes a few more centuries. Such IRD pulses are unlikely to explain Heinrich events, which are associated with relatively cold periods within glaciations. Surges are not necessary conditions for the production of large IRD events.

## Introduction

We present a non-surge mechanism for producing centurial-millenial ice-rafted-debris (IRD) pulses from marine ice sheets. It is a climatically-driven model, but the required forcing (most likely a warming) is not consistent with evidence for the climatic associations of Heinrich or Bond Events [*Heinrich*, 1988; *Bond and others*, 1993a, b]. We do however show that the occurrence of large ice-rafted debris events is not a sufficient condition for surges and that climatic forcing may be the explanation for other IRD events, for example those emanating from Antarctica [*Kanfoush & others*, 2000]. Our model also avoids some problems of surge models of ice-rafted debris events [*MacAyeal*, 1993; *Alley & MacAyeal*, 1994], which have yet to explain how a surging ice-stream with a melting base can transport frozen sediment into the ice-shelf. Modern fast-flowing ice-streams appear for the most part to to have clean basal ice [*Kamb*, 2001].

We suppose that the ice is initially frozen to the base, and that retreat of the grounding line incorporates frozen sediment into the floating ice-shelf. Frozen sediment, perhaps weakened by the presence of brine pockets, can be lifted off by superjacent ice or incorporated into the ice by fracturing and folding as happens at the terminus of the Mackay Glacier [*Powell and others*, 1996] and other cold-based terrestrial glaciers [*Robinson*, 1984]. Frozen marine margins are found in Antarctica, where relatively broad stretches of slow cold-based margins are separated by narrower fast-flowing warm-based stream mouths. Net positive annual accumulation occurs at the grounding line and on the adjacent ice shelf.



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We investigate conditions under which a frozen bed occurs using a physically-based ice sheet model. We then show that when the marine margin experiences ablation it retreats and warms at the base. Since we require a frozen bed to incorporate debris into the shelf, we suppose that once the glacier base is warmed to melting point, sediment supply ceases. This usually happens after a few centuries. We also find that the ice shelf is no longer viable, and that this happens before the base of the grounded ice reaches melting point. Frozen-in debris crossing the grounding line may therefore reach the calving front and be released in icebergs rather than being melted off underneath the shelf. These two processes, shelf destruction and the attainment of basal melting under the grounded ice mean that sediment supply begins around a century after the start of the climate warming, and stops because of internal ice dynamics, unrelated to external forcing.

The model requires a band of erosion of sediment along a broad zone of the continental shelf. Our proposed mechanism can produce large IRD events. Using the Heinrich Events as an example of a large IRD event, each of Events 1 and 2 requires a sediment influx of $3.7 \times 10^{11}$ m$^3$ [*Dowdeswell and others*, 1995]. If the source were the continental shelf off Eastern Canada and glaciated New England (2000 km $\times$ 100km), it would require an average erosion of nearly 2 m, assuming that the porosity of distal IRD sediments and shelf sediments was roughly the same. If erosion were concentrated over the Labrador coast (1000 km $\times$ 50km), then the erosion would need to be 7.5 m. This is compatible with observations of glacially-moved erratic blocks of rock and of sediment [*Sugden and John*, 1976]. The required amounts are upper limits, as a proportion of the sediment was deposited by turbidite processes [*Andrews and Tedesco*, 1993], presumably activated by IRD deposition onto the continental margin. Freezing of sediment to a depth of a few metres over a thousand years or more is readily conceivable if we can demonstrate that cold-based conditions existed under marine ice-sheet margins. Erosion of sediment is not likely to be the rate limiting process.

## Marine ice-sheet model

We investigate the physical plausibility of our glaciological hypotheses using a vertical plane-flow numerical model of a marine ice-sheet [*Hindmarsh*, 1996, 1999] coupled to an ice-shelf model [*Oerlemans and Van der Veen*, 1984]. The model is actually intended to represent a dome such as the Siple Dome which is flanked by ice streams. We compute the flow along the centre line (divide) of the dome, and include a scale correction term to account for lateral divergence. The model tracks the grounding line position explicitly, permitting accurate computation of margin motion. An equivalent

**Table 1.** Summary of results of model simulations. Scenarios refer to *initial* conditions A to E; some are used more than once with different ablation rates forcing the transient simulations. Symbols $a, b, \theta^s, L$ represent the accumulation rate in m/yr, depth of the continental shelf below sea-level in km, the approximate upper surface temperature of the ice-sheet at the grounding line in °C and inter-stream spacing in km. $\Delta S$ refers to change in span in km in 1000 yr, $a_b$ is ablation at grounding line in m.yr$^{-1}$, $T_f$ is time to shelf break up in years, $T_m$ is time to first basal melting in years. Time to first basal melting refers to anywhere under the ice sheet and the grounding line could still be cold based.

| Scenario | $a, b, L, \theta^s$ | $\Delta S$ | $a_b$ | $T_f$ | $T_m$ |
|---|---|---|---|---|---|
| A | (0.05,-0.25,$\infty$,-30) | 25 | 2.3 | 50 | 100 |
| B | (0.05,-0.25,250,-20) | 41 | 2.3 | 100 | > 1000 |
| C | (0.1,-0.25,250,-20) | 18 | 2.3 | 100 | > 1000 |
| D | (0.05,-0.5,250,-20) | 122 | 2.3 | 100 | 800 |
| E | (0.1,-0.5,250,-20) | 74 | 2.3 | 100 | 400 |
| D | (0.05,-0.5,250,-20) | 24 | 0.5 | 700 | > 1000 |
| E | (0.1,-0.5,250,-20) | 10 | 0.5 | > 1000 | > 1000 |
| D | (0.05,-0.5,250,-20) | 200 | 4.7 | 30 | 200 |

map-plane model does not exist. We carry out two numerical studies, considering (i) whether an ice-sheet with appropriate dimensions and accumulation distribution can exist with a frozen base and (ii) examining the fate of the ice-sheet margin, the basal temperature and the ice-shelf when ablation starts.

The divergence is incorporated into the continuity equation, which relates the $x$-direction gradient of the flux $q$ and the accumulation rate $a$ by

$$\partial_t H + \partial_x q = a + D, \tag{1}$$

where

$$D = -2\bar{A}(\rho_i g)^n \frac{H^{n+2} s^n}{(n+2)\, L^{n+1}}, \tag{2}$$

represents a scale estimate of the flux divergence into streams with spacing $2L$ [*Hindmarsh*, 1990]. Here $x$ is the distance along the dome, $t$ is the time, $H$ is the thickness of ice, $s$ the height above sea-level, and $A, n$ are the rate factor and index in the Glen relationship

$$e_{ij} = A\tau_{ij}^n, \tag{3}$$

where $e_{ij}$ is the strain rate and $\tau_{ij}$ is the deviator stress tensor. The quantity $\bar{A}$ represents the vertical average of $A$ [*Hindmarsh*, 1996], while $\rho_i$ is the density of ice and $g$ is the acceleration due to gravity. The flux is given by the usual relationship

$$q = -2(\rho_i g)^n \bar{A} \frac{H^{n+2} |\partial_x s|^{n-1} \partial_x s}{2(n+2)}, \tag{4}$$

and in the shelf the longitudinal stress $\tau_{xx}$ is computed by using the relationship

$$\tau_{xx} = \left(1 - \frac{\rho_i}{\rho_w}\right) \rho_i g H, \tag{5}$$

where $\rho_w$ is the desnity of water. Using this relationship in the consititutive relationship for ice gives

$$\partial_x u = \tilde{A}\tau_{xx}^n, \tag{6}$$

to solve for the velocity. Here $\tilde{A}$ is the appropriate average for the shelf [*Hindmarsh*, 1993]. The temperature equation is

$$\partial_t \theta + u\partial_x \theta = \kappa \partial_z^2 \theta + \partial_z u \tau_{xz}/\rho_i c, \tag{7}$$

where $\theta$ is the temperature, $u$ is the velocity, $\kappa$ is the diffusivity of ice, and under the shallow ice approximation we have [*Hutter*, 1983]

$$\tau_{xz} = -\rho_i g(s-z)\partial_x s, \tag{8}$$

$$\partial_z u = 2A\tau_{xz}^n, \tag{9}$$

The temperature equation is solved on a moving grid with first-order upwinding for horizontal advection and pseudo-spectral methods for vertical heat transport and a flux-gradient based computation of the vertical velocity [*Hindmarsh & Hutter*, 1988; *Hindmarsh*, 1999]. Using 41 grid points means that the error of the numerical model is 3%. We use $A = 1 \times 10^{-16}$ Pa $^{-3}$ yr $^{-1}$, $(\rho_i, \rho_w) = (917, 1030)$ kg m $^{-3}$, $g = 9.81$ ms$^{-2}$, $c = 2000$J kg$^{-1}$ and $\kappa = 34.4$ m$^2$yr$^{-1}$.

## Modelling

The Siple Dome shows diversion of ice into streams [*Scambos and others*, 1998]. The ice is roughly 500m thick at the grounding line, where we infer the base to be frozen from the large break in surface slope at the grounding line, which indicates a large change in basal traction. Without lateral diversion, our model predicts that the Siple Dome would be warm-based at the grounding line, but with lateral diversion, we obtain a frozen base. Basal freezing at the very margin is favoured by low accumulation rates and sea-level temperatures, and by thin ice at the grounding line. The diversion of ice flow into streams, which reduces the flux and heating, thus has a significant effect.

Considering more general situations, we find that without lateral divergence we need quite low sea-level temperatures (-30°C) , thin ice (250m) and low accumulation rates (0.05 m/yr) to obtain cold-based grounding lines for an ice-sheet with divide-margin distance of 400km. (This is the distance from the Labrador Coast to the divide running south from Ungava Bay, and is comparable with distance (500km) from the Siple Coast to the inland divide). With lateral divergence included, many more scenarios with cold-based margins are possible. Figure 1 shows computed basal temperature against distance along the centre-line for various scenarios described in Table 1. Figure 2a shows the initial geometry and internal temperature field for Scenario D. The ice sheet is very flat at high elevations because most ice is being diverted laterally rather than flowing towards the grounding line. The Siple Dome is also very flat in this

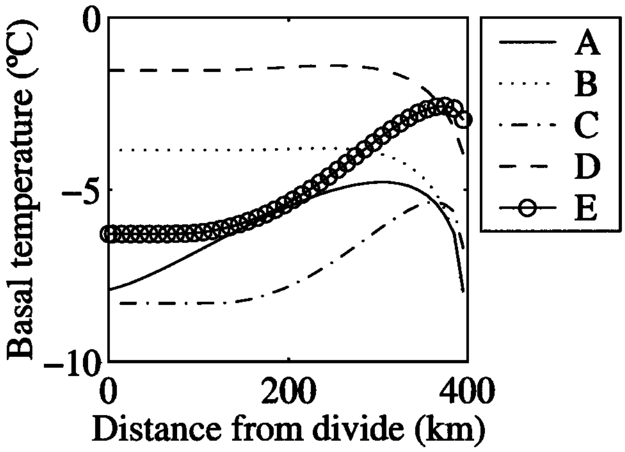


**Figure 1.** Computed steady basal temperature profiles for the indicated scenarios. See Table 1 for scenario parameters.

longitudinal sense. The temperature field is essentially determined by conduction for the low accumulation rate prescribed in this experiment.

Table 1 also shows results from the time-dependent response of ice-sheets to ablation occurring over the shelf and at the grounding line. Ablation at the grounding line was set, declining with elevation, so that a substantial accumulation area remained. When grounding line ablation rates were set at 2.3 m/yr, the ice-shelf melted away in between 50 and 100 years. This is likely to be an overestimate of the time taken to destroy an ice-shelf, as ablation promotes ice-shelf break-up [*Doake & Vaughan*, 1991]. The removal of the ice-shelf eliminates its role as a location for melting-off of frozen basal debris [*Jenkins & Doake*, 1991]. (There have been suggestions [*Hulbe*, 1997] that special conditions could protect sediment already frozen in). Simultaneously, the margin retreats, and the base at the grounding line begins to warm up because the dynamics of the glacier change such that warm ice is forced upwards, reducing vertical conduction of heat. There is a period of at least a few hundred years prior to melting of basal ice, which permits the outflow of frozen debris across the grounding line. The occurrence of basal melting is a glacial (i.e. non-climatic) mechanism for the termination of IRD events and implies that cessation of the IRD flux can predate the cessation of the increased iceberg flux. For lower ablation rates (0.5 m/yr) the retreats were less (less debris released), warming was slower (more time for the debris to be released across the grounding line) but the ice-shelf existed for a much longer period (debris melted off at the base of the shelf). Opposing effects occurred for high ablation rates. These effects of retreat, warming and shelf disappearance are illustrated in Figures 2b and 2c. Figure 2b shows upwarping of temperature fields near the margin arising from upwards advection of warm ice.

If the ablation were due to sublimation and thus associated with *cold* conditions, the amount required to produce large IRD events is somewhat in excess of present observations (e.g. the Taylor Glacier in East Antarctica experiences up to 0.5m/yr of ablation [*Robinson*, 1984].)

There is a separate question of whether a warming would allow the long distance movement of icebergs in the presumably warmer oceans. We suggest that the influx of large numbers of icebergs could have locally cooled and freshened the ocean. Even if warm conditions were already prevalent over the adjacent ocean during the ice-rafting event, colder oceanic conditions would have been maintained along the iceberg drift tracks.

## Concluding discussion

We have shown that a mechanism which includes an external forcing (a climate warming) and internal glacial response can create pulses of ice rafted debris which occur sometime after the climatic warming started, because of the time required to break up the ice shelf, and can cease before the warming ceases, owing to the melting of basal ice. In consequence, a fresh water flux can occur without there being a debris flux. We require ice at the base of the glacier to be frozen in order that ice rafted debris is created.

The climatic associations of this mechanism do not permit it to be considered as an explanation for Heinrich or Bond events with our present understanding of the glacial climate. However, there is no requirement that all IRD

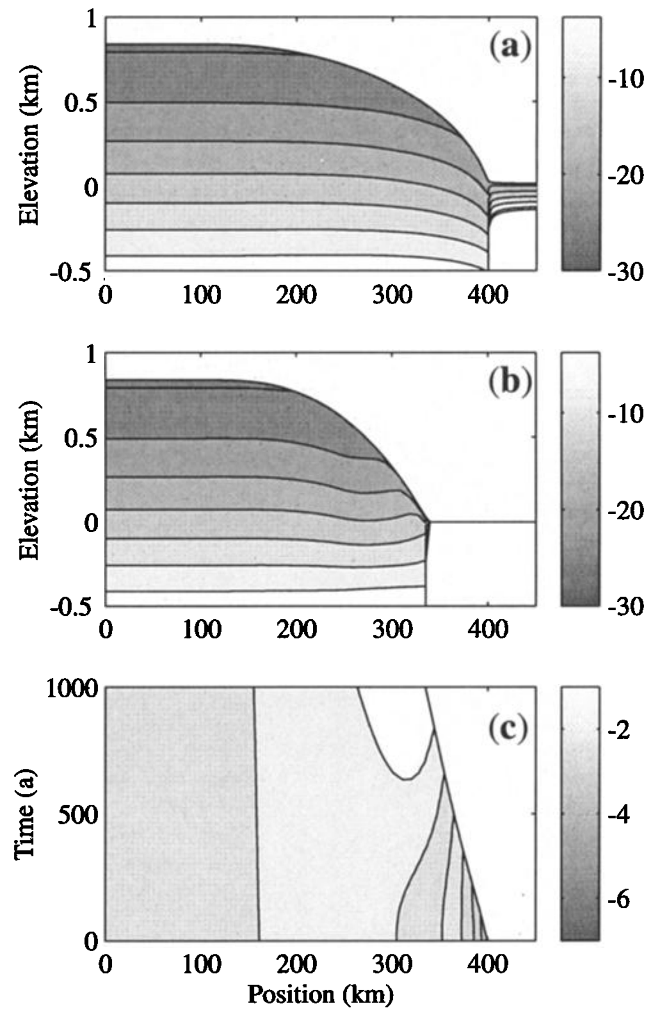


**Figure 2.** Ice sheet geometry, internal temperature and evolving basal temperatures for Scenario D. Panel (a) shows initial ice sheet geometry and temperature field with contour intervals of 3.75°C; (b) shows geometry and temperature field after one thousand years with same contouring, while (c) shows the evolving basal temperature with contour intervals of 0.5°C.

events are caused by the mechanisms which create Heinrich and Bond events, and we provide a mechanism which can produce similar pulsations but is climatically forced.

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## References

Alley, R.B. and D.R. MacAyeal, Ice-rafted debris associated with binge purge oscillations of the Laurentide ice-sheet , Paleoceanography, 9(4), 503-511, (1994)

Andrews. J.T., and Tedesco, K., Detrital carbonate rich sediments, northwestern Labrador Sea; implications for ice-sheet dynamics and iceberg rafting events in the North-Atlantic Geology, 20(12), p.1087-1090, (1993);

Bond, G. and six others (1993), Evidence for massive ice-berg discharges into the glacial Nothern Atlantic, Nature, 360(6401), p.245-249.

Bond, G. and thirteen others, Correlations between climate records from North-Atlantic sediments and Greenland ice, Nature, 365(6442), p.143-147, (1993).

Doake, C.S.M. and D.G. Vaughan, Rapid disintegration of the Wordie Ice Shelf in response to atmospheric warming, Nature, 350(6316), 328-30., (1991)

Dowdeswell, J.A., M.A. Maslin, J.T. Andrews and I.N. McCave, Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North-Atlantic sediments, Geology, 23(4), p.301-4, 1995.

Heinrich, H. , Origin and consequences of cyclic ice rafting in the Northeast Atlantic-Ocean during the past 130,000 years , Quaternary Research 29, p.143-152 (1988);

Hindmarsh, R.C.A. Timescales and degrees of freedom in the operation of continental ice-sheets. Trans. Roy. Soc. Ed. Earth Sci, 81, p.371-84, (1990);

Hindmarsh, R.C.A. Modelling the dynamics of ice-sheets, Prog. Phys. Geog., 17(4), p. 391-412, (1993).

Hindmarsh, R.C.A Stability of ice-rises and uncoupled marine ice-sheets Ann. Glaciol., 23, p.105-115 (1996),

Hindmarsh, R.C.A., On the numerical computation of temperature in an ice-sheet, J. Glaciol., 45(151), p.568-574, (1999).

Hindmarsh, R.C.A. and K. Hutter, Numerical fixed domain mapping solution of free surface flows coupled with an evolving interior field, Int. J. Numer. Anal. Meth. Geomech., 12, p.437-59.

Hulbe, C.L., An ice shelf mechanism for Heinrich layer production Paleoceanography, 12(5), 711-7, (1997).

Hutter, K., *Theoretical Glaciology*, , Reidel.

Jenkins, A and C.S.M. Doake, Ice-ocean interaction on Ronne Ice Shelf, Antarctica J. Geophys. Res, 96(C1), 791-813, (1991).

Kamb, W.B., The Lubricating Basal Zone of the West Antarctic Ice Streams, in. R.B. Alley and R.A. Bindschadler (eds), The West Antarctic Ice Sheet: Behavior and Environment, Antarctic Research Series 77 157-99, (2001)

Kanfoush, S and five others, Millennial-Scale Instability of the Antarctic Ice Sheet During the Last Glaciation, Science 288 1815-18, (2000)

MacAyeal,D.R. Binge/purge oscillations of the Laurentide ice-sheet as a cause of the North-Atlantic's Heinrich events, Paleoceanography, 8(6), p.775-784, (1993);

Oerlemans, H. and Van der Veen, C.J., Ice Sheets and Climate, Reidel, (1984)

Powell, R.D., M. Dawber, J.N. McInnes and A.R. Pyne, Observations of the grounding line area at a floating glacier terminus, Ann. Glaciology (22), p. 217-223, (1996)

Robinson, P., Ice dynamics and thermal regime of Taylor Glacier, South Victoria Land, Antarctica. J. Glaciol., 30(105) 153-160 (1984) (1982).

Scambos, T.A., Nereson N.A. and Fahnestock M.A., Detailed topography of Roosevelt Island and Siple Dome, West Antarctica Ann. Glaciol. 27, p.61-67.

Sugden, D.E and B. John (1976), Glaciers and Landscape, Arnold.

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