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ARTICLE

Genesis of dispersal plumes in till[1]

Roger LeB. Hooke, Don I. Cummings, Jerome-Etienne Lesemann, and David R. Sharpe

Abstract: In regions formerly covered by continental ice, till sheets may contain distinctive clastic particles derived from local bedrock sources such as ore bodies. Such particles, especially in thicker tills, may be distributed in three-dimensional dispersal trains or plumes. Developments in our understanding of glacial erosion, entrainment, and deposition over the past two or three decades help clarify formation of these plumes. Much of the debris in the basal ice of ice sheets is likely incorporated at places where water is freezing onto the base of the ice sheet. This water, largely a product of melting of basal ice further upglacier, has migrated downglacier under the influence of a gradient in the hydraulic potential controlled primarily by the ice surface slope and secondarily by basal topography and thermal regime. Refreezing occurs over a substantial distance along flow, so as the ice moves away from an ore body, material eroded from the body is later elevated above the bed by refreezing of more meltwater, incorporating additional material derived from the country rock. Broadening and mixing of the plume, both vertically and horizontally, can occur by shear in the ice, by shifting ice flow directions, by collisions between particles, and by folding of basal ice layers. Further downflow, where net basal melting resumes, the debris is deposited by meltout or lodgement. Many idiosyncrasies of dispersal plumes are likely caused by non-steady-state changes in basal thermal regime.

Résumé : Dans des régions anciennement recouvertes de glace continentale, les couches de till peuvent contenir des particules clastiques distinctives provenant du socle rocheux local, p. ex des amas minéralisés. De telles particules, surtout dans les tills à grande épaisseur, peuvent être distribuées dans des trains ou des panaches de dispersion à trois dimensions. Au cours des deux à trois dernières décennies, une meilleure compréhension de l'érosion glaciaire, du transport et de la déposition a aidé à expliquer la formation de ces panaches. Une grande partie des débris dans la couche de glace basale est sans doute incorporée à des endroits où l'eau gèle à la base de la nappe glaciaire. Cette eau, provenant en grande partie de la fusion de la couche de glace basale plus en amont du glacier, a migré vers l'aval sous l'influence d'un gradient du potentiel hydraulique, lequel est surtout contrôlé par la pente de la surface glaciaire et puis par la topographie à la base du glacier et le régime thermique. Le regel a lieu sur une grande distance le long de l'écoulement; donc, à mesure que la glace s'éloigne d'un amas minéralisé, le matériel érodé de l'amas est par la suite élevé au-dessus du lit par le regel d'autre eau de fonte, incorporant du matériel additionnel provenant de la roche encaissante. L'élargissement du panache et le mélange à l'intérieur de celui-ci, horizontalement et verticalement, peuvent se produire par le cisaillement de la glace, le changement de direction d'écoulement de la glace, des collisions entre les particules et le plissement de couches basales. Plus encore vers l'aval, lorsque recommence la fonte basale nette, les débris sont déposés dans un till de fusion ou de fond. Plusieurs particularités des panaches de dispersion proviennent probablement de changements d'états non stationnaires au régime thermique de base. [Traduit par la Rédaction]

Introduction

For over two centuries, it has been recognized that ore bodies could be located by tracing clasts in glacial till back to their origin in the bedrock. The process by which the clasts had been moved, however, was initially unknown (see Grip 1953 for historical details). The "clasts" to which we refer herein vary in size from microscopic grains, identifiable only by chemical analysis, to cobbles and boulders with visible mineralogy. If these can be associated with a known source, all of them, from chemical signatures to macroscopic particles, are called *indicators* (Flint 1971, p. 175; Shilts 1993; Neuendorf et al. 2005). Drake (1983) believed that the distribution of indicators in till must inherently be three-dimensional (3-D) (Fig. 1), and thus that a 3-D term like "plume" should be used. Plume is normally used to refer to 3-D buoyancy-driven flows; Drake, however, used the term, and we use it herein, only to emphasize the 3-D nature of the dispersion (not the driving force).

The 3-D character can be recognized only if the till sheet sampled is thick, and if samples are taken at different depths. Most tills are sampled using shallow pits dug at the surface (e.g., Larson and Mooers 2005; McClenaghan et al. 2002), which precludes identifying the plume geometry. In addition, many till sheets are relatively thin; plumes in these are also likely 3-D, but this would be difficult to demonstrate.

That clasts higher in the till came from further up-ice was apparently first recognized by Sauramo (1924, cited by Puranen 1988). The full 3-D nature of plumes, however, was likely first demonstrated by geologists tracing a train of boulders containing rich silver ore near Harmsarvet, Sweden, in 1935–1937 (Grip 1953). There, a boulder train was followed northward, first on the surface and then in subsurface by trenching, until the mother lode was found.

Drake (1983) cites several examples which he reinterprets to be evidence for a 3-D geometry, and noted some difficulties in explaining this geometry. He was particularly concerned by the fact

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R.LeB. Hooke. School of Earth and Climate Sciences and Climate Change Institute, University of Maine, Orono, ME 04469, USA.
D.I. Cummings. DC Geosciences, 12 Decarie Street, Aylmer, QC J9H 2M3, Canada.
J.-E. Lesemann. Department of Earth Sciences, University of Ottawa, Ottawa, ON K1N 6N5, Canada.
D.R. Sharpe. Geological Survey of Canada, 601 Booth St. Ottawa, ON K1A 0E8, Canada.
Corresponding author: Roger LeBaron Hooke (e-mail: rogerhooke@gmail.com).

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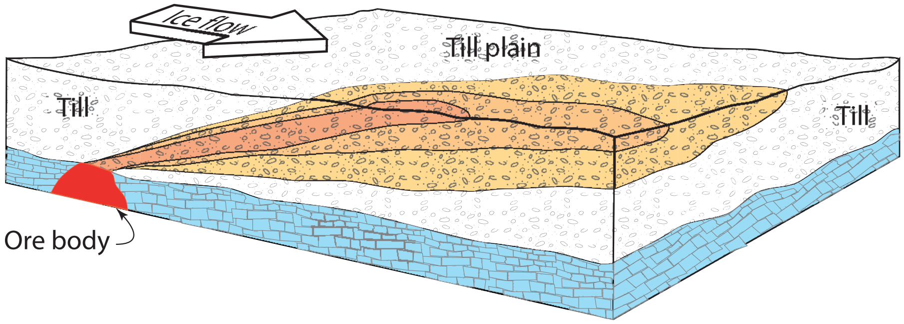
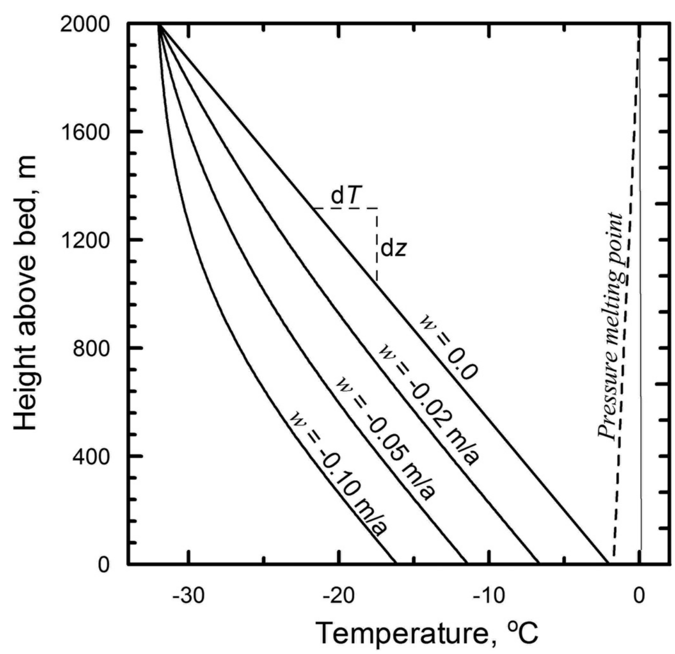
Fig. 1. Idealized model of a 3-D dispersal train in till (after Drake 1983; used with permission of the University of Chicago Press).



that once a source is covered by till, further erosion would be inhibited. However, he did not offer a solution to this problem. Miller (1984) provided data demonstrating the 3-D nature of some geochemical dispersal plumes, and attributed the rising of the plume to low-angle thrust planes. Stanley (2009, p. 42) invoked thrusting to entrain material, and he correctly recognized that entrainment would continue downflow from an ore body. However, he thought the plume was raised off the bed because shear somehow moved particles upward in the ice. He did not provide a rigorous physical explanation for this process. Puranen (1990, p. 31) went a step further, appealing to "the action of newly eroded material pushing the previously loosened grains upwards" to elevate ore particles into the ice. Both Drake and Miller focused on the observation that the maximum concentration of indicators was sometimes found hundreds of metres or kilometres downflow from the bedrock source. Drake, however, did not have subsurface data to demonstrate that the feature was 3-D, and in some cases we suspect he overinterpreted the observations.

In our effort to clarify the formation of dispersal plumes, we build on Puranen's analyses, incorporating ideas on glacial erosion that have developed over the past 30 years. These ideas, separately, are well known in the glaciological community. Some, however, do not appear to have been discussed in the exploration literature, and to the best of our knowledge they have not previously been combined into a model for formation of dispersal plumes. Our goal in this paper is to provide exploration geologists and others studying dispersal plumes with a better understanding of plume formation, and thus a firmer foundation upon which to interpret their observations.

Fig. 2. Temperature (*T*) profiles near the center of an ice sheet, 2000 m thick, with various vertical velocities (*w*). Geothermal flux: 33 mW/m² (typical of parts of the Canadian Shield; Blackwell and Richards 2004). Calculated using Robin's (1955) theory.



Temperature of an ice sheet

Glacial erosion is strongly dependent on the thermal regime at the base of an ice sheet. Where ice is frozen to the bed, erosion is negligible. Where the basal temperature is at the pressure melting point and ice is melting, erosion is possible but the thickness of the basal layer of debris-bearing ice is likely minimized by the tendency for particles to be moved toward the bed as melting occurs. Where meltwater is freezing to the base of the ice sheet, however, substantial thicknesses of debris-rich basal ice can accumulate.

Near the center of an ice sheet, at the ice surface, the mean annual accumulation rate in metres of ice equivalent is roughly balanced by the vertical velocity, *w*. This vertical flow advects cold ice downwards, thus cooling the bed (Fig. 2). In the Laurentide Ice Sheet, the ice at the divide was likely thicker than shown in Fig. 2, perhaps approaching 3 km. The accumulation rate was also probably quite low. Thus, there may well have been a small amount of basal melting beneath the divide (Fig. 3).

Outward from the divide, the horizontal velocity increases, as does the accumulation rate (and hence *w*). Thus, downward advection of cold surface ice is stronger, and the ice is also advected outward (dashed flowline in Fig. 3). Although this advection of cold ice is partially offset by higher surface temperatures at lower elevations and by strain heating near the bed, it still leads to a downglacier increase the basal temperature gradient. Where this gradient becomes steep enough to conduct upward into the ice more heat than is supplied by the geothermal flux, water moving outward along the ice/bed interface will refreeze. In constructing Fig. 3, conditions were selected that resulted in refreezing of basal meltwater between ~270 and 450 km from the divide, and in a frozen bed between ~450 and 840 km. The continued outward increase in strain heating, however, eventually raises the temperature above the pressure melting point again at a distance of ~840 km from the divide.

These thawed→frozen and frozen→thawed transitions, and that described next, are likely gradational, as illustrated in Fig. 4. The frozen patches in Fig. 4 may be colder because they are slightly higher (so the ice over them is thinner) or they may result from spatial differences in thermal conductivity of the substrate.

Still further downglacier, slightly down-ice from the equilibrium line, the vertical velocity becomes upward, and the temperature profile, instead of being bowed downward as in Fig. 2, is bowed upward. This dramatically reduces the amount of geothermal heat that can be conducted to the surface, and basal melt rates increase accordingly (Fig. 3). At the margin, however, where the upward vertical velocity again decreases and the ice becomes

Fig. 3. Temperature distribution along a flowline in a polar ice sheet calculated using Budd's (1969) column model. This model is crude by current standards, but it captures the essential elements of the distribution. The accumulation rate is assumed to rise from 0 at the equilibrium (Eq.) line to maximum of 0.8 m·a⁻¹ (ice equivalent) 82 km upglacier from the equilibrium line, and then decline to 0.02 m·a⁻¹ at the divide. Dashed arrow is flowline showing how cold ice from upglacier ends up beneath warmer ice downglacier, resulting in reversal of the near-surface temperature gradient. Not all of the basal regimes shown will be present along every flowline. (From Moran et al. 1980, fig. 6.) a.s.l., above sea level.

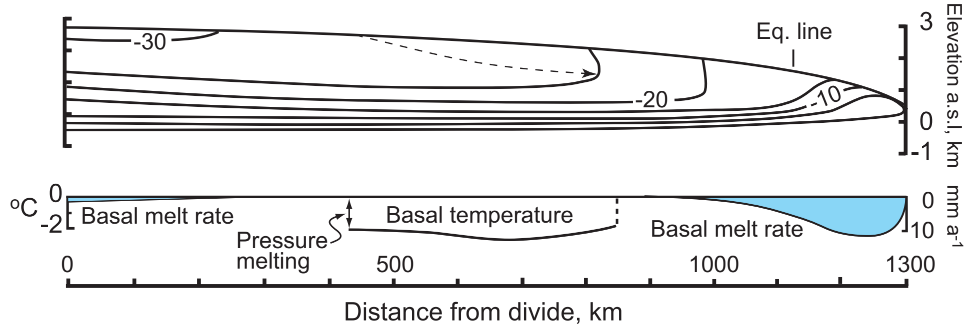
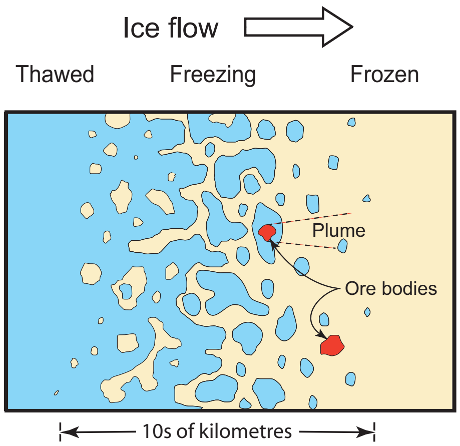


Fig. 4. Hypothetical distribution of frozen and thawed areas within a transition zone from a region of thawed bed that is undergoing refreezing to one of frozen bed. Two ore bodies are shown. The one in the region of refreezing is producing a plume. (Modified from Hughes 1992, fig. 14.)



thin, the outermost few kilometres of the ice sheet may be frozen to the bed.

Budd et al. (1971) were apparently the first to present model calculations showing this zonation of subglacial thermal regimes, but David Sugden's pioneering 1977 paper called attention to its relevance to geomorphic problems. By extending Sugden's analysis closer to the margin, Moran et al. (1980) recognized the possibility of a frozen toe there, but they assumed a steady state. Cutler et al. (2000) called attention to the fact that such frozen toes may also develop when an ice sheet advances over permafrost — a non-steady-state situation — and showed that they could have persisted for centuries.

Water movement beneath a polar ice sheet

Melt rates beneath polar ice sheets may reach several millimetres per year (Fig. 3). This water must either escape or accumulate until it can escape. Initially, it accumulates in any permeable substrate that may be present. Once the substrate becomes saturated, a water layer forms between the ice and the bed. As long as the water discharge through the substrate is less than the basal melt rate, the pressure in this layer will rise, limited only by the weight of the overlying ice. When the water pressure approaches the overburden pressure, flow is initiated between the substrate and the glacier sole.

The water flow is driven by the gradient in the potential, Φ, given approximately by

$$\frac{\partial \Phi}{\partial s} = \rho_i g \frac{\partial h}{\partial s} + \rho_w g \frac{\partial z}{\partial s} \tag{1}$$

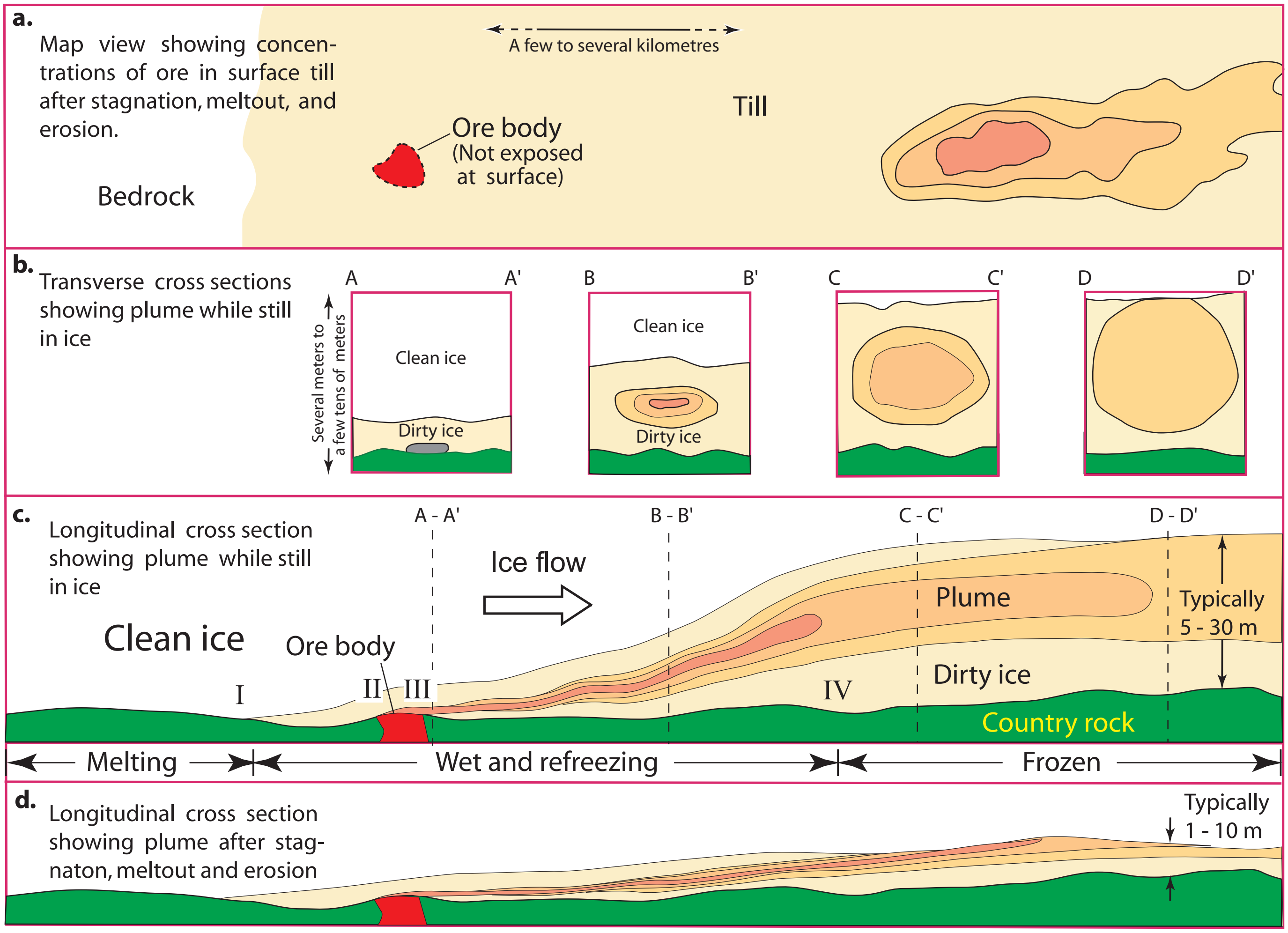
where ρ_i and ρ_w are the densities of ice and water, respectively, h is the glacier thickness, g is the acceleration due to gravity, and z is the elevation above some datum such as sea level (Shreve 1972). Water will tend to flow, either through the substrate or as a subglacial stream, in the direction, s, in which $\partial\Phi/\partial s$ is a maximum. In most cases, the bed topography, $\partial z/\partial s$, simply perturbs the flow direction without altering the overall flow, which is in the direction of decreasing ice thickness, $\partial h/\partial s$ — toward the margin.

Glacier erosion and entrainment

To form a dispersal plume, a glacier must entrain material from the bed and transport it downflow. Some of this material was likely saprolite formed by pre- or inter-glacial weathering (e.g., Feininger 1971). Most known dispersal trains, however, are in basal till deposited by the Laurentide and Fennoscandian ice sheets that reached their maximum extent about 20 ka. While these probably derived much of their debris load from earlier tills, they must also have eroded unweathered crystalline rock by abrasion and quarrying. In general, abrasion produces silt and clay-sized particles, while quarrying results in macroscopic fragments. Rock strength, however, also affects particle size; weaker rocks are less likely to yield pebble and coarser debris, so studies focusing on only one size fraction, particularly the coarser fraction, may overlook significant anomalies (Dilabio and Shilts 1979; Shilts 1993).

The question of how glaciers entrain sediment was debated in the 1950s and 1960s and is still not well understood (Alley et al. 1997). Goldthwait (1951) believed that debris was sheared into the ice and actually carried to the glacier surface along the shear planes. This hypothesis was initially widely accepted. Weertman (1961), however, perhaps inspired by Drygalski (1897), suggested instead that material was entrained at places where meltwater, moving outward from a zone of net melting, was refreezing to the glacier sole. Radar soundings in Antarctica have revealed a remarkable example of this process (Bell et al. 2011): basal ice layers with a mean thickness of ~490 m are interpreted as originating from basal refreezing over 30–60 kiloyears, suggesting a freeze-on rate of 5–15 mm/a. Christoffersen et al. (2010) argue that a layer of debris bearing ice at the base of Kamb ice stream in Antarctica originated in this way. Alternative explanations for several of the features Goldthwait (1951) attributed to shear were provided by

Fig. 5. Genesis of a 3-D plume (modified from Miller 1984; used with permission of Maney Publishing, London). (*a*) Map view of the plume as exposed at the surface. (*b*) Transverse cross sections through plume while still in ice. (*c*) Longitudinal section showing (*i*) initiation of entrainment at I where water begins to refreeze to the glacier sole, (*ii*) initiation of dispersal train at II, and (*iii*) cessation of entrainment at IV where water supply is exhausted. (*d*) Plume after till is deposited. This plume geometry was obtained by compressing the geometry in panel *c* vertically and shearing it horizontally. Compare with panel *a*.



measurements near the margin of Barnes Ice Cap on Baffin Island (Hooke 1973). Alley et al. (1997) discussed other processes for entraining debris such as shearing in zones of high longitudinal compression; injection into basal crevasses (see also Ensminger et al. 2001); and regelation normal to the bed. Bed-normal regelation forces ice down into permeable sediment (e.g., Drygalski 1897; Iverson and Slemmons 1995; Iverson 2000) that then may be sheared loose from the bed. Observations beneath 213 m of ice at Engabreen, Norway, provide intriguing evidence for this process (Iverson et al. 2007). However, it, alone, cannot raise debris further off the bed as it is transported. Alley et al. (1997) also discussed a variation of Weertman's refreezing mechanism involving formation of frazil ice in water moving up an adverse bed slope and trapping fine sediment particles in the flow.

Herein, we focus on the refreezing mechanisms, recognizing, however, that other processes may also entrain material.

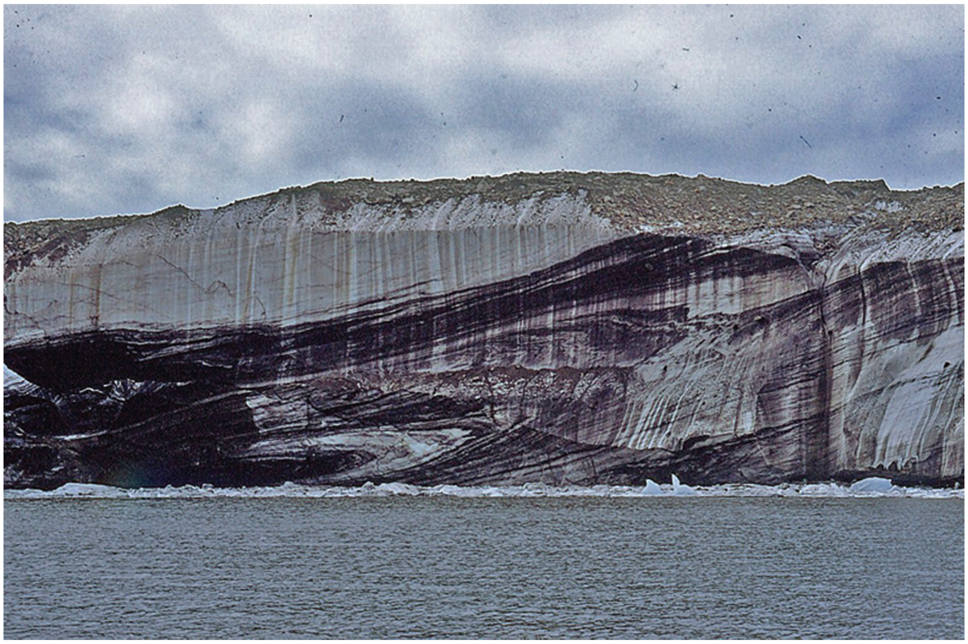
A point to emphasize from this discussion is that debris-rich basal ice accumulates through time and, contrary to normal sedimentary successions, it accumulates from the bottom up so the last material entrained is at the bottom. In many situations this stratigraphy is likely to persist as the debris is deposited as till. The implications of these facts were not fully appreciated by Drake (1983) and Miller (1984).

A model for formation of dispersal plumes

Our model for formation of dispersal plumes depends, globally, on the overall glacier regime, and locally on the thermal state at the bed. We couch the discussion in terms of an ore body, although a local bedrock source of any distinctive lithology can form a plume. We also couch it in terms of a steady state, as if the ice sheet were plunked down on the landscape fully formed and at thermal and kinematic equilibrium with the prevailing conditions. Non-steady-state complications are discussed later.

Based on the foregoing discussion, we think plumes are probably initiated in areas like that between 270 and 450 km in Fig. 3 where water is refreezing to the glacier sole in a wide diffuse boundary zone like that illustrated in Fig. 4. Significant incorporation of country rock begins where refreezing starts (at I in Fig. 5*c*). In the absence of an effective mechanism for incorporating large amounts of debris, and consistent with our steady-state approximation, the ice upflow from this location is relatively clean. The plume is initiated at II, where the refreezing occurs over the ore body. Downflow from point III ore fragments are elevated into the glacier by continued refreezing, incorporating more country rock, and dispersed by diffusion processes (discussed later in the text). We view debris entrainment as a contin-

Fig. 6. Recumbent fold in basal ice of Barnes Ice Cap, Baffin Island, Canada. The fold is believed to have formed as a result of a slight change in flowlines in basal ice following a climatic perturbation (Hudleston 1976) (photo: R.LeB.H., summer 1969).



uous process between I and IV so a rather thick basal layer of dirty ice accumulates. Particles higher in this layer will have come from further upglacier, consistent with the observations of Sauramo (1924) and others. The thickness of the layer will depend on the difference between the heat flux to the bed from below and that upward from the bed into the ice, and on the sliding speed of the glacier. The concentration of debris will depend on the erodibility of the bed.

The moving dirty basal ice can experience one of two fates: (*i*) it can freeze to the bed once the supply of liquid water from upglacier is exhausted (at ~500 km in Fig. 3 or IV in Fig. 5), or (*ii*) it can reach a place where there is net basal melting again without passing over a zone of frozen bed. In the former case there will be increased compressional strain, folding, and possibly some minor thrust faulting. These processes will thicken the layer of dirty ice, but the concentration of particles in the ice will not change. Debris carried in ice above the bed will continue to move downglacier. With the basal ice now frozen to the bed, shear strain rates are likely to increase, further smearing out the feature. In the latter case, the dispersal feature will be gradually let down and particles will be lodged in the substrate. Because the basal melt rate is small compared with the flow rate, this will also smear out the feature over a considerable distance downflow.

Panel *d* of Fig. 5 shows the plume once the till has been deposited by lodgment or meltout. The plume has been reduced in thickness and sheared horizontally. This was the state sketched by Miller (1984) based on two well-documented examples (see Fig. 8), and is the state depicted in panel *a* and in Fig. 1. (As sketched, the transition from panel *c* to panel *d* requires a change in regime, a non-steady-state.).

Precisely the same processes could occur in a transition to a frozen toe in a marginal zone.

A noteworthy characteristic of plumes is that an ore body in one location may produce one while other(s) in the same area do not (Grip 1953, p. 721; McClenaghan et al. 2002, p. 317). This could well be because refreezing of basal meltwater was occurring over the ore body that produced a plume, but not over the others, and is consistent with the model of basal temperature distribution in Fig. 4.

Vertical and lateral dispersion of indicators

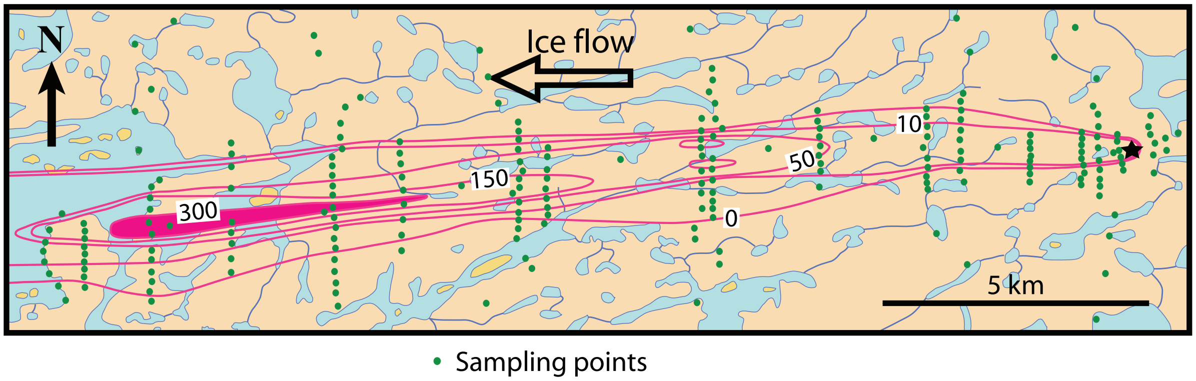
The concentration of indicators is 0 outside a plume and rises to a maximum in its center. Over an ore body, the concentration of ore clasts in the last ice to have been accreted approaches 100%. Further downflow, particles are affected by vertical shear in the ice ($\partial u/\partial z$ where u is the ice velocity in the downflow or x direction, and y and z are the axes in the transverse and vertical directions); by horizontal shear ($\partial u/\partial y$) as the ice passes frozen patches, coarser particles ploughing into the bed, or zones of higher bed roughness; by local changes in ice flow direction in response to bed topography; and by collisions between particles as they shear past one another (Weertman 1968). These processes are diffusive; they move ore particles preferentially outward toward ice in which their concentration is low, and particles of country rock inward, into the dispersal train, where their concentration is also low. The dispersal feature is thus broadened, thickened, and diluted.

Global changes in flow direction in response to climate change or other perturbations (e.g., Flint 1971, pp. 176–177; Shilts, 1993, p. 347) are also especially important in broadening a plume. Viewed at this scale, they are advective rather than diffusive. Equally important, but greatly underappreciated, is the fact that changes in glacier regime can result in isoclinal recumbent folds in basal ice (Fig. 6) (Hudleston 1976). Such folding is responsible for a lack of stratigraphic correlation in the basal 200 m of the Greenland Ice Sheet in the ice cores GRIP and GISP II (Meese et al. 1997, p. 26 420) which are only 28 km apart. Folding may well be a major contributor to mixing of indicators. As amplitude/wavelength ratios in such folds commonly exceed 10^2, this process may appear to be diffusive.

Erosion of till surfaces

Where entrainment of debris is initiated upflow from the ore body, the proximal part of a plume will be overlain by ice that contains debris but is free of the indicator lithology. More distally, a dilute part of the plume may reach the surface by diffusive processes (Fig. 5*b*, Section D–D′). However, to expose the high-concentration core, as shown in Fig. 5*a*, it would seem that some special circumstance is required.

Fig. 7. The Ranch Lake kimberlite train. Indicator mineral is Cr–diopside in the 0.42–0.84 mm size range. Contours are number of grains in a 20 kg sample. Kimberlite source is at ⋆. Modified from McClenaghan and Kjarsgaard (2007).



The core of a plume may become exposed if the indicator-poor cover is removed, either by subglacial or subaerial erosion. As till is gradually lodged, ice flow continues. Thus, the core of the plume may be deposited, and then the overlying, not-yet-deposited, indicator-poor cover sheared away and replaced by clean ice. Subglacial meltwater may also erode the top of till; eskers on the Canadian Shield commonly lie in meltwater corridors, 1–5 km wide, that are typically interpreted to have been eroded subglacially (Craig 1964; Rampton 2000). Alternatively, erosion may occur post-glacially by subaerial processes.

Folding in the ice (Fig. 6) may also leave the core of the plume exposed at the ground surface after deglaciation.

Time scale for formation of a dispersal plume

Consider an ore body 400 km from the center of an ice dome and under 3 km of ice flowing radially. The mean accumulation rate between the dome and the ore body is 0.2 m/a (ice equivalent). These are roughly the conditions at the transition from thawed to frozen bed in Fig. 3. To transfer all of that ice downglacier, and thus preserve a steady-state ice sheet profile, the depth-averaged ice velocity (commonly called the balance velocity) over the ore body would have to be ~13 m/a. The velocity near the bed would depend on the fraction of the bed that was frozen, but would be significantly less than this (Hughes 1992). Thus, a dispersal train 10 km long might take well over 1000 years to form. A thousand years is a substantial length of time, but it is only a small fraction of the time during which many areas were covered by ice in the last 100 000 years.

Longitudinal variation in indicator concentration

Shilts (1976, pp. 209–210; 1993, p. 344) noted that, in some indicator trains, concentrations of indicators peaked "in till at or close to their source." He thought that, in an "ideal" situation, the concentration, c, of indicators would decrease with distance, x, from the source following a "negative exponential" pattern. Larson and Mooers (2004) showed analytically that the *depth-averaged* concentration of indicators should fall off hyperbolically with distance. Their relation can be written as

$$c = \frac{1}{a + bx} \tag{2}$$

where, for any given situation, a and b are constants that depend on the mass of till in transport upflow from the ore body, the breadth of the ore body in the direction of flow, and the erosivities of the ore and country rock. In a till sheet only a metre or so thick, the variation in concentration with depth may be minimal and, barring non-steady-state phenomena, the variation in indicator concentration in surface samples might follow this relation, at least approximately. However, the model illustrated in Fig. 5 suggests that, in thicker till sheets, one might expect a gap between the source and the maximum concentration in surface samples.

Partitioning

Dilabio and Shilts (1979) and Shilts (1993) emphasized that the composition of *unweathered* till from a single site commonly varies with grain size. The latter notes, for example, that "the presence of abundant local pebbles and cobbles should not be taken to mean that the finer fractions of the till are likewise of local origin, nor should the opposite be assumed" (p. 344). This is a very interesting observation for which we suggest three possible explanations. (*i*) Particles of different sizes may be deposited at different rates. Finer particles may settle out though a millimetre-scale subglacial water layer while coarser ones are dragged further by the moving ice. (*ii*) Different rock types, or different minerals in a rock (Shilts 1993, pp. 338–339), likely produce particles of different sizes. These may then become mixed by the diffusive processes described earlier in the text. (*iii*) Fine particles are commonly produced by abrasion while coarser ones by plucking. On a relatively smooth bed, abrasion can occur without plucking. In any case, this observation suggests that laboratory analysis of samples should not be confined to a small range of grain sizes, and that, at least during reconnaissance work, analysis of separate size fractions could be useful.

The non-steady-state

Many idiosyncrasies of plumes can be attributed to non-steady-state conditions. We have already alluded to some high-frequency non-steady-state processes like collisions between particles or ploughing of coarser particles into the bed. Lower-frequency changes could involve the distribution of frozen and thawed patches on the bed, the erosivity of the bed (perhaps as a surface weathered zone is removed), and paths of meltwater movement along the bed. These latter three are probably the most important for explaining characteristics of plumes. At still lower frequency is migration, either upglacier or downglacier, of the transition zone between thawed and frozen zones, and climate changes affecting the overall flow field. We illustrate these with some case studies and a hypothetical example.

Ranch Lake kimberlite train

The Ranch Lake kimberlite train (McClenaghan et al. 2002) extends westward from a kimberlite pipe beneath Ranch Lake in Canada's Northwest Territories (Fig. 7). The train can be traced for 70 km, but the segment of most concern to us is the first 20 km in which the train widens from 0.5 to 2 km. Contrary to eq. (2), the concentration of the indicator minerals, Cr–diopside (Fig. 7) and pyrope in the 0.42–0.84 mm size range, increases downflow, reaching maxima between 15 and 20 km from the subcrop (the "exposure" of the source beneath the till). Most samples were collected from shallow (0.2–0.8 m) hand-dug pits in mudboils in places where the till sheet was 2–10 m thick.

The failure of the indicator-mineral concentration to decrease as predicted by eq. (2) could be due to the following:

(*i*) Differential rates of comminution in which coarser particles in the till are comminuted to the 0.42–0.84 mm range faster than existing particles in that range are reduced to finer sizes. Although some authors suggest that comminution is of minor importance in tills (Dredge et al. 1997; McClenaghan and Kjarsgaard 2001), in some cases it may be a factor (Dreimanis and Vagners 1971, 1972). In this case it cannot be ruled out with available data.

(*ii*) A period during the time of westerly ice flow in this area when the kimberlite eroded faster than either earlier or later, or during which the country rock eroded faster, thus diluting the concentration of indicator minerals. Changes in erosion rates, are likely causes of gaps between subcrops and indicator maxima as Puranen (1990, p. 29) noted. Such changes could be caused by changes in erosivity of the rock as deeper layers become exposed, in the distribution of frozen and thawed patches, or in water movement along the ice-bed interface. McClenaghan et al. (2002) suggested that the erosion rate might also decrease as differential erosion increased the depth of the kimberlite below the average elevation of the bedrock, but one can question whether enough erosion occurred during formation of the train to have had such an effect.

(*iii*) The 3-D character of the plume. Although not clearly defined due to lack of subsurface data, this possibility should not be dismissed. As noted, samples were collected from mudboils which likely homogenize the upper part of the till. Tills are typically <2 m thick in this area. However, near the kimberlite, which lies in a depression, they may be thicker. Thus, the mudboils may not have been deep enough to sample high concentrations of indicator phases nearer the bed.

Lough Derg plume

The Lough Derg plume (Fig. 8) in central Ireland is one of two studied by Miller (1984). Till in the area ranges from 4 to 15 m in thickness. The indicator element is Pb derived from the Old Red Sandstone. Miller (p. 72) notes that the Old Red "… commonly contains elevated Pb values near the formation top" and he frequently refers to an "assumed source" of an anomaly. We interpret this as indicating that the source of any given anomaly is not known, but that numerous sources are likely to have existed. Miller, as noted, attributes rising plumes to low-angle thrust planes in the till, and thinks the bed topography may deflect a plume upward.

There are three anomalies (A–C in Fig. 8) in the distribution for which we will offer possible explanations:

(*i*) The high concentration at the surface at A may have been sourced further upflow at a time when the bedrock there was producing till with high concentrations of Pb or was eroding more rapidly than it was later. Some post-depositional erosion is likely to have occurred here.

(*ii*) The >400 ppm Pb zone at the surface at B may reflect a period of higher erosion rate, or erosion of bedrock with a higher Pb concentration, near the top of the bedrock high at D, the rest of the plume having come from lower on this hill. On such low hills, changes from freeze-on to frozen are likely common, as geothermal heat is deflected away from highs toward lows by such topography (Hooke and Medford 2013).

(*iii*) The separation of the >800 ppm Pb zone at C from the tip of the buried zone at E likely reflects a period of reduced erosion rate on the lee flank of the hill at 225 m or a period of higher flow speed (or both). A period of reduced refreezing on the lee side of that hill could reduce erosion and increase sliding speed.

The Red Hill syenite boulder train, New Hampshire (USA)

The Red Hill syenite boulder train (Fig. 9*a*) was described by Goldthwait (1968) and is one of those discussed by Drake (1983). It extends ~35 km downflow from a 340 m high hill composed of a distinctive syenite, reaches a maximum width of ~14 km, and has several indicator maxima. The indicators are syenite boulders in stone walls. Goldthwait (1968, p. 25) notes that, "Stone counts of smaller pebbles in till suggest much the same pattern…but with slightly higher percentages". There is a strong indicator maximum at the base of Red Hill and then what Drake calls a "shadow zone" between Red Hill and what Drake considered to be the principal maximum along the northern side of the train. Drake attributes this shadow zone to the distance required for a plume to reach the surface.

When ice covered Red Hill and was actively eroding it, the hill was likely overlain by syenite-bearing dirty ice, which probably in turn was overlain by syenite-free dirty ice from upflow (Fig. 9*b*). Later, downflow from the hill, subglacial melting deposited this entrained debris as lodgement till. There may also have been a till cover on the hill that was later removed by subaerial erosion. Without 3-D data, we cannot know whether the syenite-bearing till is underlain by syenite-free till.

Given that Red Hill projects well above the surrounding till blanket, the barren till in the shadow zone is noteworthy. It may have come from up-ice of Red Hill as we have shown Fig. 9*b*. However, equally likely is the possibility that non-steady-state processes were again involved. Owing to changes in water flow or thermal regime, erosion of the syenite could have ceased well before deglaciation, so ice with the highest concentration of syenite was moved away from the hill.

Similarly, the other two areas of high syenite concentration, further downflow and along apparently different flowlines, likely reflect an earlier period of erosion, perhaps concentrated on opposite flanks of the hill leading to two peaks rather than only one. This may have occurred at a time when the top of the hill was frozen while the bottom was thawed, a likely common occurrence as noted earlier.

Complete detachment of a train from its source

When an ice sheet becomes frozen to its bed, the pressure melting isotherm no longer lies at the glacier sole; it is deeper in the substrate. When it is only a few (or a very few tens) of metres below the sole, and when the water pressure beneath it is high, failure may take place along the isothermal surface. Substantial packages of subglacial sediment or bedrock have thus become attached to an ice sheet and transported, *en masse*, tens to hundreds or thousands of metres (Liszkowski 1987; Aber 1989). A layer of till, up to some metres in thickness and containing a plume, could thus be detached from its source. Such displacement would be difficult to recognize in the field, and would likely prove confusing and frustrating to the prospector following a plume up-ice in hopes of finding its source.

Conclusions

Entrainment of substantial volumes of subglacial material likely occurs in areas where the sole of a continental ice sheet is at the pressure melting temperature and meltwater from upglacier is refreezing to it, incorporating particles from the bed. An ore body in such an area would constitute a localized source for clasts with a distinctive lithology. Extending an analysis by Puranen (1988, 1990), we argue that as ice moves away from this source, more water is frozen to the glacier sole, entraining more country rock and raising the dispersal train off the bed (Fig. 5*c*). This results in a rather thick layer of dirty ice of which the dispersal train constitutes only a part. The dispersal train is preserved when the dirty ice becomes frozen to the bed and remains so as the ice retreats, or when basal melting begins to occur and the entrained

Fig. 8. Cross section of the Loch Derg plume studied by Miller (1984). Contours at 200, 400, and 800 ppm Pb. Modified from Miller (1984, fig. 6; used with permission of Maney Publishing, London).

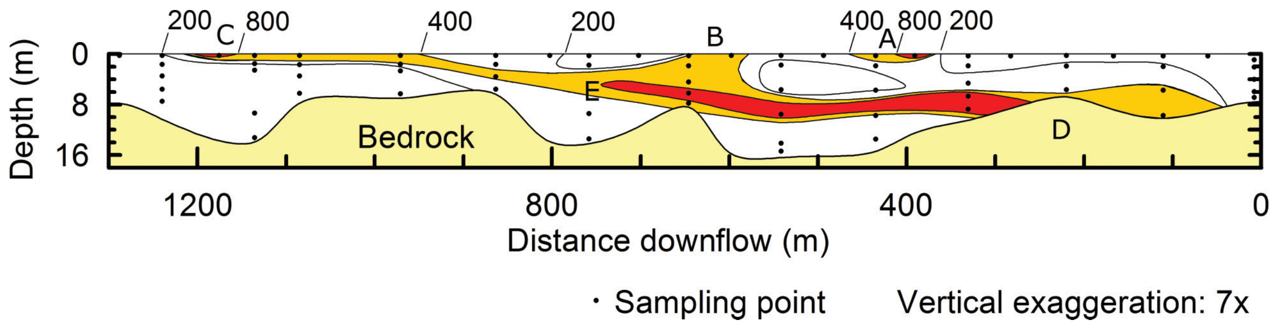
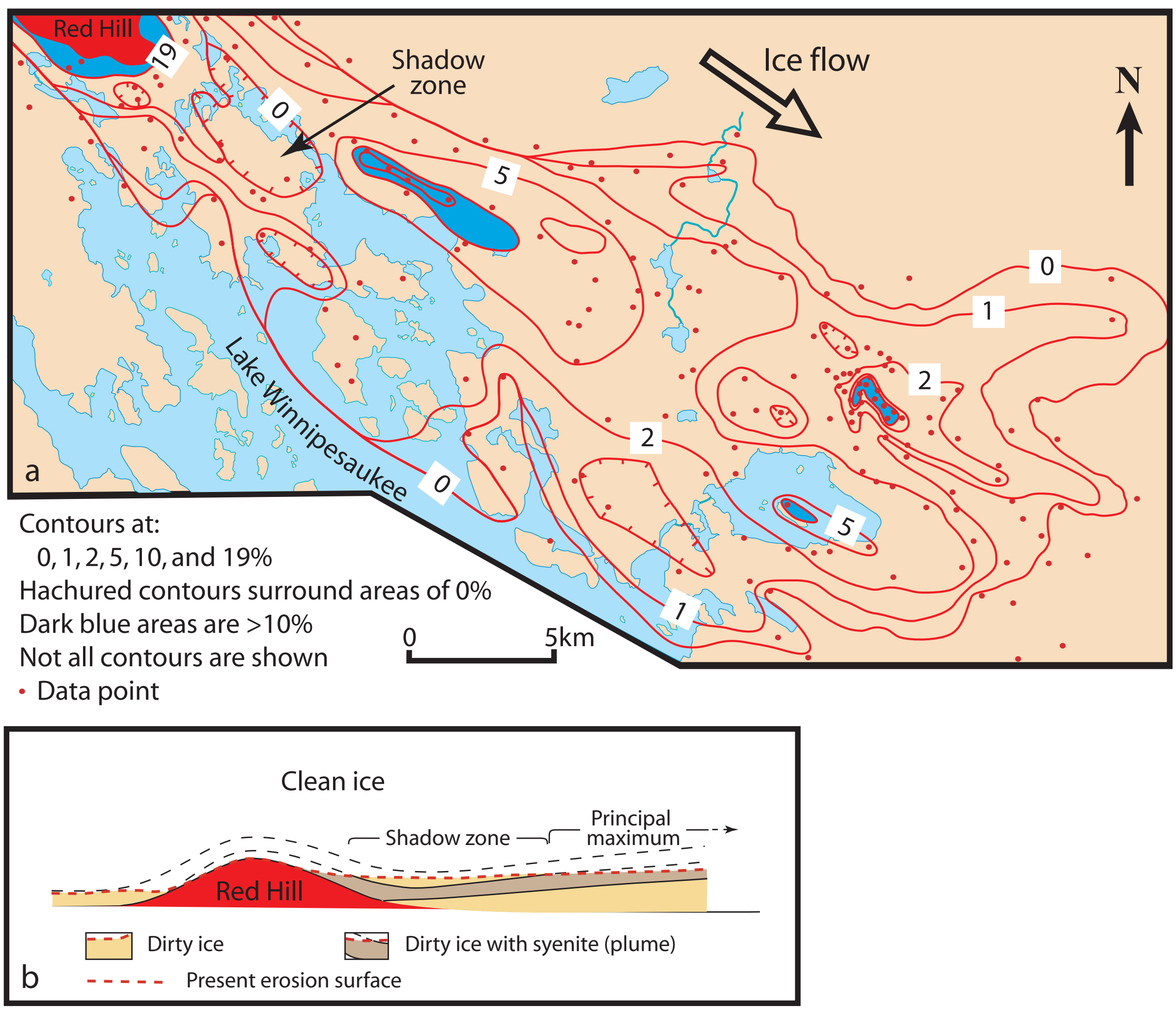


Fig. 9. (*a*) Map of Red Hill indicator fan discussed by Drake (1983). (Modified from Goldthwait 1968, fig. 7A; used with permission of the New Hampshire Geological Survey.) (*b*) Schematic longitudinal cross section showing geometry while ice was present. During deposition, thicknesses of the dirty ice layers would be reduced and sheared, but otherwise their relative positions would not be changed. (Colours in figure refer to online version.)



debris is deposited as lodgment till. Either process would likely smear out the dispersal train substantially.

Non-steady-state changes offer many possibilities for perturbing this basic pattern. Over the several centuries that it likely takes to generate a dispersal train, changes in meltwater supply to the area of eroding bed or in basal thermal regime are to be expected, and will alter both the erosion rate and the ice flow speed, leading to longitudinal variations in indicator concentration. Changes in flow direction are also likely on these time scales, and will lead to broadening of the dispersal train.

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