PROSPECTING UNDER COVER:

Using Knowledge of Glacial Processes in Mineral Exploration

Notes to accompany CIM Short Course
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Introduction

As a glacier advances it grinds down and erodes the bedrock beneath the ice. Conceivably, ore deposits are also eroded, resulting in mineralized grains and boulders being picked up and incorporated into the advancing ice. With melting and subsequent ice retreat, all entrapped rock and mineral grains, fragments and boulders will be deposited in the form of thin or occasionally thick veneers of glacial sediment. In large areas of Newfoundland and Labrador the bedrock surface is covered with unconsolidated glacial sediment which also prevents direct observation of bedrock and therefore, of potential economic mineral deposits. A general understanding of the origin of glacial sediments provides the background necessary for use as an effective tool for mineral exploration and is necessary in order to identify economically significant minerals or geochemical anomalies within these sediments and to trace these minerals back to their source, which may be exposed bedrock or drift-covered bedrock.

Glaciers & Glaciation: General Principles

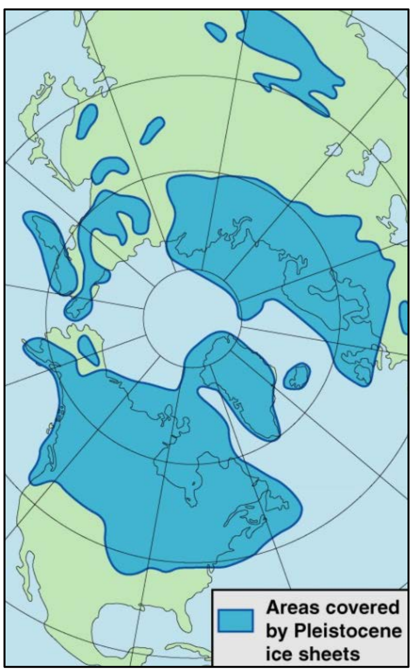

Figure 1. Areas covered by Pleistocene ice sheets during the Quaternary Period.

There were multiple glaciations in the Quaternary Period (last 1.9 Ma) during which Canada experienced at least four periods of glaciation, each followed by long interglacial periods where climatic conditions were much warmer than they are today (**Figure 1)**. The four major glacial episodes, from oldest to youngest, have been termed the Nebraskan, Kansan, Illinoian, and Wisconsinan. Some earlier glaciations (*e.g.*, Illinoian) were more extensive than the latest one (Wisconsinan).

Up to 97% of Canada was covered during these glacial extremes, whereas today only 1% of the land surface is covered by glacial ice, mainly in the Queen Elizabeth Islands, Baffin Island and mountainous regions in western Canada and the Torngat Mountains of Labrador.

The erosional and depositional effects of previous glacial events have been largely destroyed, in part by the last glacial episode, the Wisconsinan which began 70,000 years ago with final retreat occurring around 6 to 7000 years ago. Deposits sampled in mineral exploration are largely Late Wisconsinan. In Newfoundland, this glaciation ended about 10,000 years ago, and 7000 years ago in Labrador.

Glaciers range from smaller alpine glaciers that occur in mountainous terrains, *e.g.*, North and South American Cordillera, to larger Continental glaciers that occur in moderate relief terrains, *e.g.*, Central and Eastern North America.

Glaciers begin to grow in areas where annual snowfall exceeds annual melting. As snow thickens over many years, the layers near the base compact and recrystallize to eventually form dense glacial ice. The formation of glacier ice occurs in the *accumulation zone* of a glacier or ice sheet (**Figure 2**). There, the mass of ice gained each year is greater than that lost by melting. Once the ice has reached a certain thickness it begins to flow out from where the snow is accumulating, like pancake batter flows out from where it is poured in a pan. These areas of outflowing ice are referred to as *dispersal centres*. When the zone of snow accumulation shifts, there will also be a change in ice flow direction which may present as multiple striation directions (**Figure 3**).

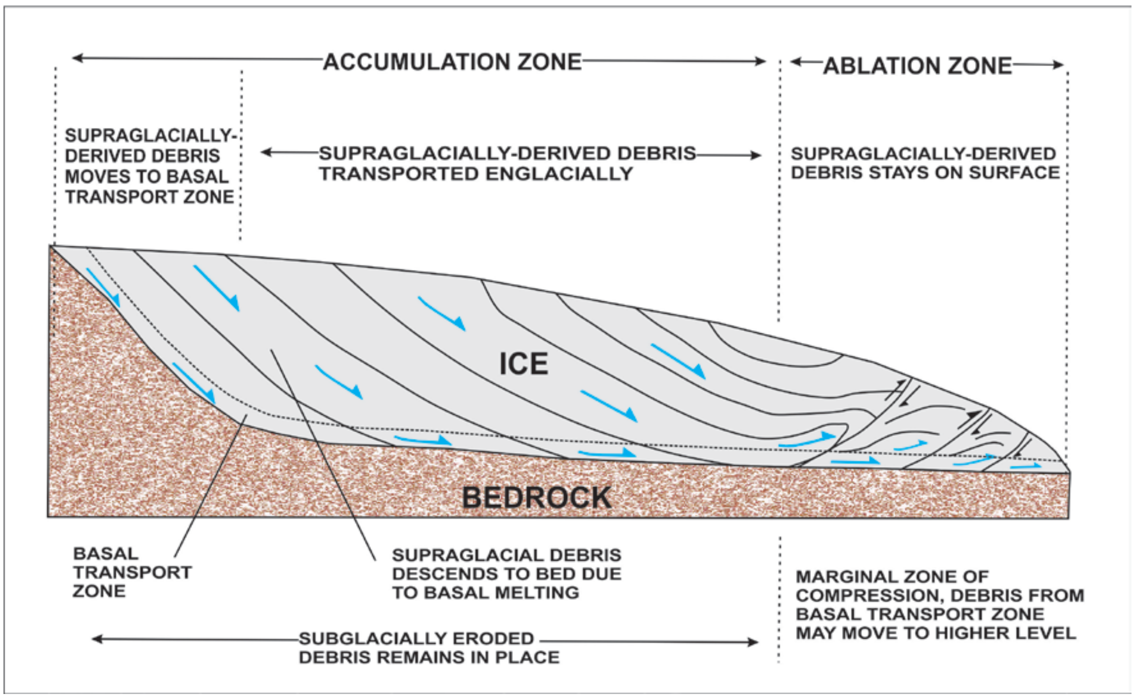

***Figure** 2.* Anatomy of a glacier. Blue arrows indicate ice-flow trajectories (modified from Boulton, 1996).

At lower elevations and under warmer temperatures, glacier ice melts at greater rates than it is formed and the glacier loses mass; this area is called the *ablation zone* (**Figure 2**). This causes the ice to thin and retreat. It thins from the snout at front of the glacier, where sediment is released in *end* and *recessional moraines*.

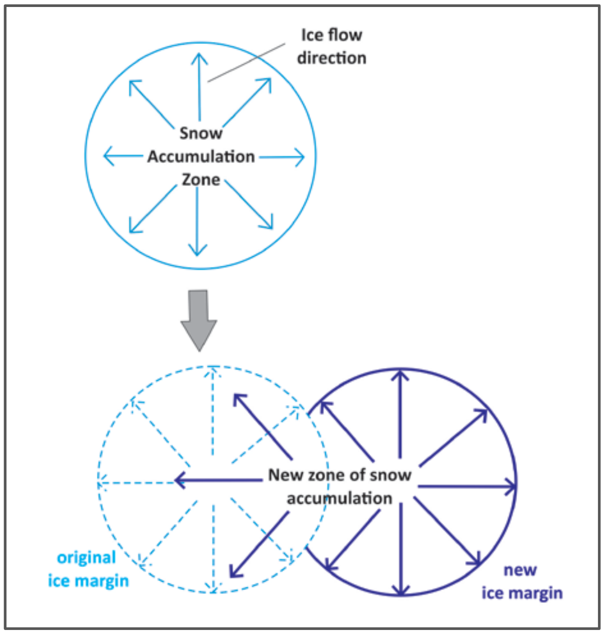

Figure 3. *Shifting ice dispersal centre. As the zone of snow accumulation shifts, there is a change in ice-flow direction which may be recorded as multiple striation directions and results in a new ice margin.*

Transfer of ice between the accumulation and ablation zones occurs through the process of *creep* or deformation. Glacier ice moves under the influence of gravity in response to both vertical (compressive) and shear stresses. The rate of glacier movement is mostly dependent on the surface slope of the glacier, the thickness of the ice (shear stresses and rates of ice movement increase as ice thickness increases), and ice temperature ('warm' ice close to the melting point can deform and move much more rapidly than 'cold' ice).

The thermal regime of a glacier is a description of the temperature of the ice, which affects not only the rate of movement but also its capacity to erode, transport and deposit sediments. *Cold-based* glaciers are typical of cold, high-latitude regions (*e.g.*, present-day Antarctica), where the temperature at the base of the ice is well below the temperature at which melting occurs, at the pressure present at the base of the glacier, and there is no liquid water present. These glaciers typically move very slowly by internal deformation (creep) at rates of only a few metres per year and are ineffective at eroding bedrock. As a consequence, cold-based glaciers cannot create or move much sediment and are ineffective geomorphic agents.

In warm and moist climates, such as those found in Alaska or the Canadian Rockies, ice is close to the pressure melting point (just below 0°C) and moves by a combination of creep and by sliding over films of water at the base of the ice. Some refreezing of this water occurs (regelation) in the lee of bedrock irregularities creating an effective method for incorporating (‘*freezing on*’) debris into the ice base. This debris is carried within a thin basal debris layer (usually < 1m thick) that consists of irregular layers of ice and sediment. Rapidly flowing, warm-based ice carries a lot of debris, is highly abrasive, and can readily carve into bedrock and transport large amounts large amounts of glacial debris. Glacial material may be carried away by subglacial rivers or may be transported within the ice as *englacial load*, or at the base of the ice.

Glaciated vs nonglaciated terrains

There are important differences between glaciated and nonglaciated areas, summarized in **Table 1.**

Glaciated terrain	Nonglaciated terrain
Soil development generally shallow (~1 m)	Soils can be developed to depth > 100 m (regolith)
Material deposited by glacial processes	Material in situ or remobilized by fluvial, eolian, or chemical processes
Dispersal patterns not confined to a drainage basin except in mountainous regions with valleys	Dispersal confined to drainage basins
Minerals of different bedrock sources may be present in glacial sediments (*e.g.*, uranium, zinc, and chromium enriched together)	More likely to have minerals with a single bedrock source
Sediments still contain minerals that are usually broken down in the first stages of weathering (*e.g.*, carbonates, sulphides, olivine, pyroxene)	Many of these minerals have been destroyed by soil-forming processes and weathering in oxidizing environments

***Table 1*. Glaciated vs nonglaciated terrains**

Glacial History of Newfoundland and Labrador

The island of Newfoundland and Labrador have very different glacial histories and styles of glaciation. Labrador was glaciated by the continental-scale Laurentide Ice Sheet (LIS), whereas the island of Newfoundland supported numerous independent ice-caps, on a much smaller scale. Evidence of pre-late Wisconsinan glaciation is rare.

Newfoundland

The tip of the Northern Peninsula was the only area covered by the LIS during the last glaciation. The rest of the island was covered by the Appalachian ice complex, characterized by smaller ice caps independent of the LIS. Separate centres of glacier accumulation existed on the Avalon Peninsula, in central Newfoundland, and on the Long Range Mountains. Complicated ice-flow patterns resulted where an area was, at different times, covered by ice from more than one ice centre. As the ice melted, these accumulation areas became isolated from each other, and as many as 15 smaller ice caps probably existed for a short time (**Figure 4**). The interaction of numerous small ice caps resulted in a complicated ice-flow history with further local topographic affects.

The farthest limit of glacial advance in many areas was onto the continental shelf, around 18,000 years ago. As the glaciers melted, coastal areas became ice-free between 11,000 to 14,000 years ago. Once the ice was restricted to land, ice retreat was slower and mostly through ablation. The climate cooled again about 11,000 to 10,000 years ago, and some glaciers readvanced for a short time.

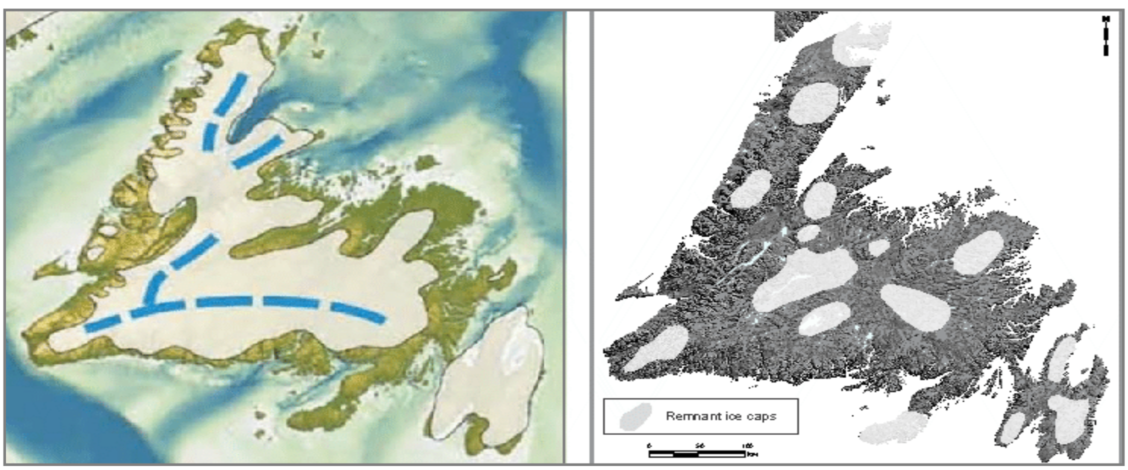

Figure 4. *Pattern of glaciation on the Island of Newfoundland. A) Retreat of ice on to land and the location of ice divides at 12 000 years ago) (after Shaw et al., 2006). B) Map of Newfoundland showing the approximate location of remnant ice caps as the Newfoundland Ice Cap disintegrated (modified after Grant, 1974).*

Labrador

The LIS covered most of Canada during the late Wisconsinan glaciation. All of Labrador was covered by the eastern sector of the LIS (except for the highest peaks of the northern Torngat Mountains, and Mealy Mountains south of Lake Melville) which reached its maximum approximately 20 000 years ago. Ice flowed out from the centre of the Labrador Peninsula in all directions. In northern Labrador, the Torngats presented a barrier to eastward ice movement. Glaciers reached the coast via the major valleys through the mountain range. The farthest extent of the ice sheet is marked by end moraines and *kames*, although in places the ice sheet terminated in the sea.

The major valleys leading to the coast contain thick sequences of glaciofluvial outwash, deposited as the ice sheet retreated. These are commonly overlain by marine sediments that were deposited during periods of higher sea level following deglaciation. On the uplands, recessional moraines and eskers mark the pattern of retreat. As the ice melted, lakes formed where river drainage was blocked by the ice sheet. The largest of these (glacial lakes Naskaupi and McLean) were trapped between the westward- retreating ice margin and the drainage divide between the Atlantic Ocean and Ungava Bay. The main LIS finally melted in the Schefferville area of western Labrador about 6,500 years ago. Small *cirque glaciers* are still found in the Torngat Mountains today.

In the central part of the Labrador Peninsula, ice flow was complex and at least four ice-flow directions have been identified from striations, streamlined landforms and distribution of clasts and erratics (Klassen and Thompson, 1989; **Figure 5**). This variation in ice flow is a result of the area's proximity to the centre of the Labrador sector of the Laurentide Ice Sheet, including one or more of its ice divides (Prest, 1984). In coastal areas, the ice-flow history is generally simpler (Klassen and Thompson, 1993).

Glacial Deposition

Glacial deposition is discussed in terms of the form of the deposits (geomorphology) and the type of sediment. Geomorphology can often be used to interpret sediment genesis. The position of glacial transport can be used to help interpret the distance of transport of debris.

The most common sediment type deposited by glaciers is a poorly sorted deposit composed of clay, silt, sand, gravel, and boulders referred to as glacial *diamicton* or till. Ice is able to transport and deposit till in a number of ways. Most till is transported by a glacier at or near the base of the ice; this subglacially-derived debris is referred to as *basal* or *lodgement* till and is commonly *short-travelled*, that is, relatively close to its bedrock source. Till may also be transported above the basal transport zone and carried *englacially* within the ice itself or on its surface as *supraglacial* debris. It rises to this position from the ice base as it is carried in the ice or it can be eroded at this position from terrain that extends above the basal transport zone (**Figure 2**).

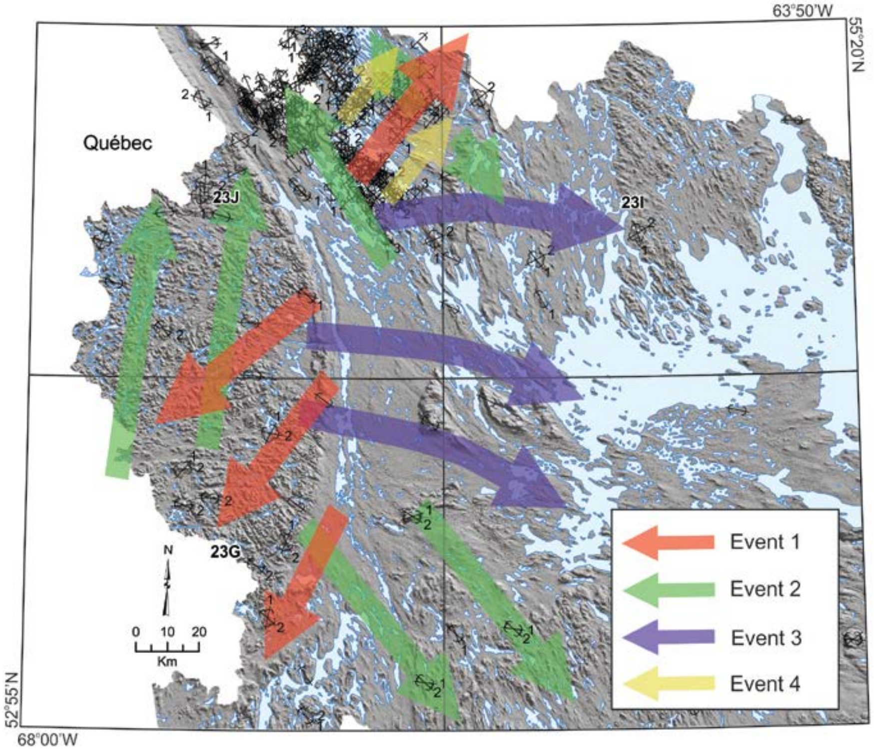

Figure 5. *Ice-flow patterns in western Labrador; numbers indicate the relative age of events with one being the oldest (modified from Klassen and Thompson, 1989).*

Diamicton is a textural term for poorly- to non-sorted sediments that occur in a variety of depositional settings (*e.g.*, debris-flow deposits from landslides). Careful observation and interpretation of its characteristics are required to determine its origin. Glacial diamictons are deposited directly by ice and are commonly associated with highly complex ice-contact facies deformed by ice melt, collapse, and bulldozing. Associated landforms provide additional clues as to depositional origin.

Basal (or lodgement) till

Warm-based ice slides on a thin film of water produced from the melting of the underlying ice. Debris within the ice tends to work its way through the glacier towards the base and this subglacial sediment is typically short-travelled. Recognizing this type of sediment is important in drift-prospecting programs.

Till may be produced from the melt-out of sediment from the base of moving ice and plastering of this material onto the substrate (lodgement till). Lodgment of debris onto a rigid substrate (bedrock) produces dense, compact, over-consolidated diamicton with few sedimentary structures. It may also contain sub-horizontal shear planes and slickensided surfaces, in places where shear stresses within the accumulating till exceed the strength of the material and slippage occurs. These shear stresses also cause the preferential alignment of the long axes of clasts in a direction parallel to ice flow (referred to as a *clast fabric*). Bullet-shaped clasts are common in lodgement tills and provide evidence of subglacial erosion (**Figure 6)**. They are oriented with the streamlined, pointed end up-glacier (or up-ice). Measurement of the long-axis orientation of clasts embedded in subglacial tills can provide data on ice-flow directions. Striated clasts are also common and are also useful in determining ice-flow directions. A basal till containing abundant very angular rocks is likely to be near its source.

Figure 6. *Bullet-shaped clast in lodgement till with its streamlined, pointed end up-ice. Striations are also present on this clast (traced in yellow) and can be used in determining ice-flow direction.*

Melt-out till

It is common to find a veneer of englacially-transported material overlying basal till. This material may be produced by the passive melt-out of basal and englacial debris under stagnant debris-rich ice that is *downwasting* in situ without subsequent transport or deformation (*melt-out till*). Melt-out till forms as debris is released from the ice either subglacially or supraglacially and the characteristics of the till will be mostly inherited from the ice from which the debris is released. These tills are typically less compact and contain more sand and gravel lenses and/or foliation than basal till, although they commonly show a strong fabric. The properties of melt-out till can be modified during or after deposition, particularly when downslope remobilization occurs and they often contain further-travelled glacial sediment.

Figure 7. Hummocky terrain, central Labrador.

Areas of stagnating ice often present as irregular hummocky terrain with boulder-strewn surfaces (**Figure 7)**. Similar to snow melt in the spring, ice does not melt uniformly and leaves irregular mounds and hollows as it melts in situ.

Deformation till

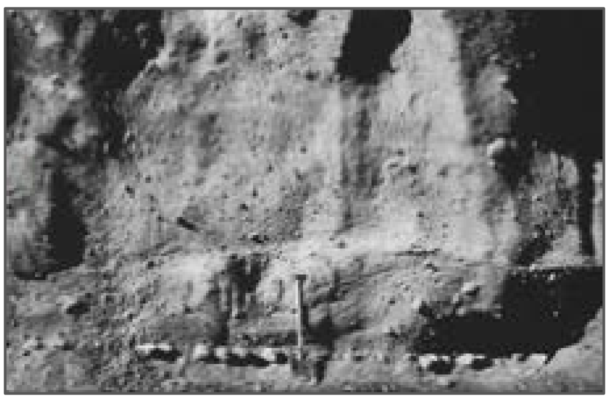

Figure 8. *Boulder pavement in deformation till (at base of shovel blade).*

Deformation till is formed from the subglacial mixing of pre-existing sediment that is moved within a subglacial *traction layer*, comprising water-saturated debris, similar to wet concrete. This deforming layer allows sediment to move along under the ice and also helps the ice to flow. The glacier almost floats across the bed which results in high pore-water pressures, and a subsequent stiffening of this layer by dewatering leaves overconsolidated deformation till. The characteristics of deformation tills vary according to the texture and composition of the pre-existing sediment incorporated into the deforming layer, their permeability and drainage characteristics, and the amount and type of strain the material has undergone. They can range from structureless to stratified with distinct textural banding, and can show evidence of faulting or folding of incorporated sediment layers. Distinctive horizons of clasts (*boulder pavements*) are one of the most common indicators of subglacial deformation and form during episodic erosional phases.

Ice-flow directional indicators

As a glacier flows, it picks up material by freezing it to its base. Rock debris along the base of the ice acts much like sandpaper as it is dragged across the bedrock surface. The resulting bedrock surface is scratched (striated), grooved and polished. This process produces both small-scale linear scratches and grooves, and large-scale bedrock erosional features like fjords, *rôches moutonnées*, and *crag-and-tail* hills. All of these features are used in determining ice-flow directions. Depositional landforms also provide indicators of ice-flow direction; those composed of till are likely oriented parallel or perpendicular to ice flow, whereas those that contain farther-travelled glacial sediment are more likely to be unoriented.

Subglacial landforms

1. Crag-and-tail hills

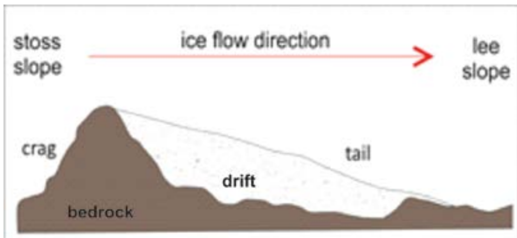

Figure 9. *Crag-and-tail hill (Ryder, 1995).*

Crags are formed when a glacier passes over a resistant rock formation. The ice erodes the surrounding softer material, leaving the rocky crag protruding from the surrounding terrain which shelters the softer material in its wake forming a tapered ridge or ramp (the tail) on its leeward side. Crag-and-tail hills form parallel to ice flow and the sediment in the tail consists of till and as such, is suitable for sampling.

2. Drumlins

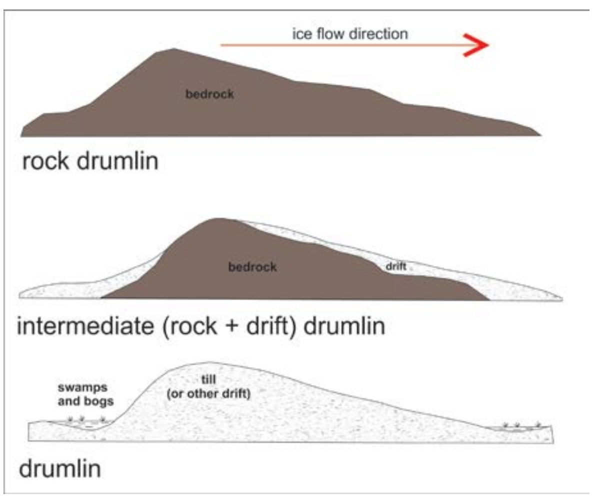

Figure 10. *Drumlins (Ryder, 1995).*

Drumlins are elongated hills or ridges formed by glacial ice acting on underlying till. They may be composed of similar material to the till of the surrounding area or be composed almost entirely of bedrock, and occur in different forms ranging from wide parabolic ridges to elongated spindle forms. Drumlins are typically 1–2 km long, less than 50 m high and between 300 to 600 m wide. Drumlins are often found in "fields" of similarly shaped and sized hills oriented parallel to the direction of ice flow.

3. Flutes

Flutes are narrow, elongated, straight, parallel ridges generally consisting of till, but sometimes composed of sand or silt/clay. Flutes typically reach a height of only a few metres above the surrounding terrain and may extend up to several kilometres in length. Flutes are oriented parallel to the direction of ice movement, and are likely formed when boulders become lodged at the base of the ice. As the ice flows around these boulders it creates elongated cavities on the down slope side of the boulder parallel to the ice flow. These cavities are then filled with glacial sediments (till and outwash). As the ice recedes, it exposes these long, low ridges of till.

4. Moraines

Deposition of subglacial till creates low-relief till plains (*ground moraine*), often forming gently rolling hills or plains that may have a variety of different types of landforms on their surface. Moraines have numerous different forms, described below.

End moraines are till ridges deposited along the front edge (snout) of the glacier (perpendicular to ice flow) (**Figure 11a**). End moraine size and shape are determined by whether the glacier is advancing, receding or at equilibrium. The longer the terminus of the glacier stays in one place, the more debris accumulates in the moraine. *Terminal moraines* mark the maximum advance of the glacier. *Recessional moraines* are smaller ridges (usually occurring as a series of ridges) left as a glacier pauses during its retreat, often called *cross-valley moraines* when they occur in valleys in mountainous areas (**Figure 11b**). Other small transverse moraines include *deGeer moraines* which form parallel to an ice front grounded in shallow water. DeGeer moraines form as till is ploughed to the glacial snout during minor *stillstands* or glacial readvances.

Lateral moraines are till ridges deposited along the sides of a glacier (**Figure 11c**). The unconsolidated debris can be deposited on top of the glacier by frost shattering of the valley walls and/or from tributary streams flowing into the valley. This material is carried along the glacial margin until the glacier melts. Because lateral moraines are deposited on top of the glacier, they do not experience the postglacial erosion of the valley floor and as the glacier melts, lateral moraines are usually preserved as high ridges.

Rögen (or *ribbed*) moraines typically occur as a series of curved to sinuous ridges perpendicular to ice flow that are often closely and regularly spaced (**Figure 11d**). The depressions between the ribs are sometimes filled with water. Drumlins and flutes are often found in close proximity of Rögen moraines, and are often interpreted to be formed at the same time.

Hummocky moraine (or *hummocky terrain*) is a strongly undulating surface of ground moraine showing steep slopes, deep, enclosed depressions and meltwater channels. It has long been regarded as a supraglacial deposit formed from the downwasting (*i.e.*, thinning) of stagnant ice. However, hummocky moraine may also be composed of the same till found in nearby streamlined landforms and in these instances, it is recognized as the product of pressing and squeezing of soft till substrate below stagnant portions of the ice margin.

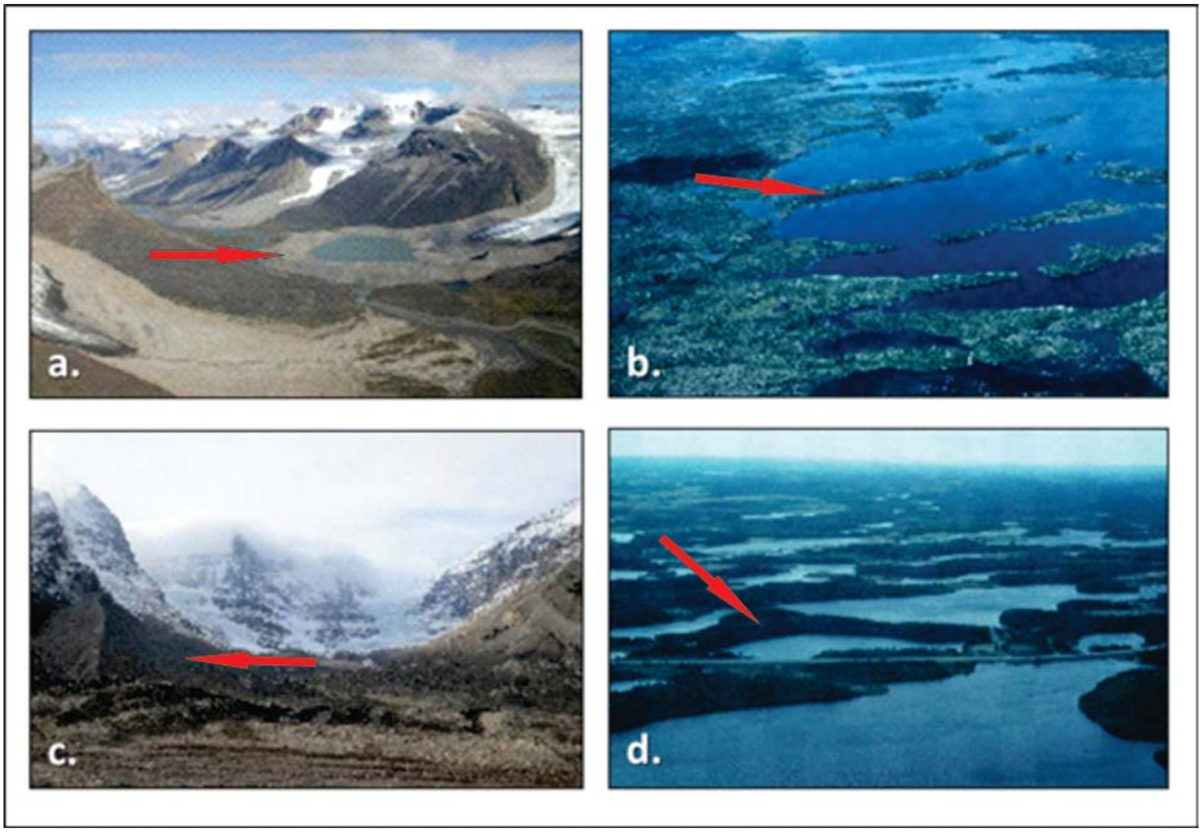

Figure 11. *Moraines. A) end moraine at snout of Barnes Ice Cap, NU. B) cross-valley moraines in Melody Lake area, LAB. C) lateral moraines along the sides of Athabasca Glacier, AB. D) Rögen moraines in Ocean Pond area, Avalon Peninsula, NL.*

How to determine ice-flow directions

Large-scale bedrock features

Large-scale bedrock features are commonly tens to hundreds of metres log, tens of metres wide and a few to tens of metres high. These features can be used to determine ice-flow direction.

1. Rôches moutonnées

Figure 12. Rôches moutonnée, Strange Lake, LAB.

Rôches moutonnées, asymmetric rock outcrops or ridges that have been scraped, smoothed and polished on one side (*up-ice* or *stoss* side) by abrasion of advancing ice, whereas the other side (*down-ice* or *lee* side) is rough and jagged due to plucking of rock fragments as the ice moves over and onward (**Figure 12**).

2. Lee-side plucking

Lee-side plucking is an erosional process that takes place as glaciers advance over rock obstructions. The end result will be a rough, steep-sloped, lee-side outcrop. Lee-side plucking features may range in size from hills (rôches moutonnées) to 1 mm high irregularities on a bedrock surface.

3. Stoss and Lee shapes

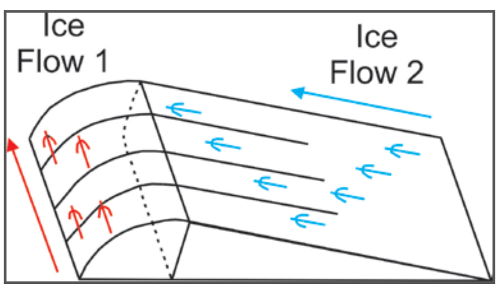

Figure 13*. Stoss and lee relationships. Ice Flow 1 (older) is preserved in the lee of Ice flow 2*

Stoss and lee shapes develop as a glacier slides along a horizontal surface or up a small slope. The ice will create striations, but where the rock surface slopes into a depression, especially if there is a relatively steep down-slope to the depression, the ice will leave no striations in the lee of the high area up-ice of it. This is the most common indicator of ice-flow direction (**Figure 13**).

Small-scale linear bedrock features

Small-scale linear bedrock features such as striations (striae) are orientated parallel to the direction of ice flow. They are best observed on newly exposed rock surfaces. Unless the rock is extremely resistant to weathering, these features will not be preserved.

1. Striations are grooves or scratches etched into the underlying bedrock by sand or rock particles embedded in a basal ice layer moving under considerable pressure (**Figure 14a**).

Striations and grooves are abundant on many rock surfaces. The Newfoundland provincial striation database has over 12000 striations recorded.

2. Nailhead and wedge striations are striations that deepen and widen in the direction of ice flow, often terminating in a pit or gouge.

3. Chatter marks, gouges, and fractures are fracture marks or cracks in bedrock that record the removal of rock flakes by subglacial quarrying. Chattermarks are usually only a few centimetres wide, and often occur at the base of shallow grooves with their open or concave sides facing down-ice. They also commonly occur as a series of closely spaced fractures nested one inside the other, resulting from repeated fracture events beneath a single overpassing clast. Crescentic gouges range from a few centimetres to more than a metre across and the horns of the gouges point up-ice. Crescentic fractures may have their horns pointing up- or down-ice. Fractures with up-ice pointing horns commonly form in isolation, whereas those with down-ice pointing hors generally occur in a series with the width of individual fractures decreasing down-ice.

4. Rat tail features are linear features that are essentially small-scale equivalents of crag-and-tails. They exhibit a resistant rock knob on the up-ice end and are created by the removal of less resistant material on either side.

Striation Mapping

Ice-flow patterns are not based on individual measurements but from numerous measurements both on individual outcrops and over an area. To accurately measure striations in the field, the following steps should be followed:

Once a polished outcrop is located, careful examination and cleaning of the surface is required to reveal all the striations. A shovel, scrubbing brush, and spray bottle are useful tools. The visibility of striations is highly dependent on lighting conditions, and bright sunlight is preferable to overcast conditions. Shading of the surface, or examination from different angles assists in identification. Select a horizontal flow surface wherever possible. Results on sloping or rough surfaces can be highly variable because of the local effect of the bedrock surface on ice flow. Using a compass, measure the orientation of striations. The ice-flow direction(s) must also be determined. For example, you may determine that the orientation of striations is 090-270 (east-west). Did the ice flow from west to east or east to west? It is necessary to examine the outcrop for stoss and lee relationships or crag-and-tail features to determine the up-ice or down-ice direction, as well as the surrounding area for any larger-scale directional features.

Multiple ice-flow directions

During the last stages of glaciation in Newfoundland there were a number of ice dispersal centres whose effects we see preserved as multiple striation directions in some areas. Earlier ice-flow directions are preserved mainly in the lee of the later ice-flow, allowing their relative ages to be determined (**Figure 14b**). Most work on glacial transport in Newfoundland indicates that the latest ice flow is responsible for most of the movement of sediment.

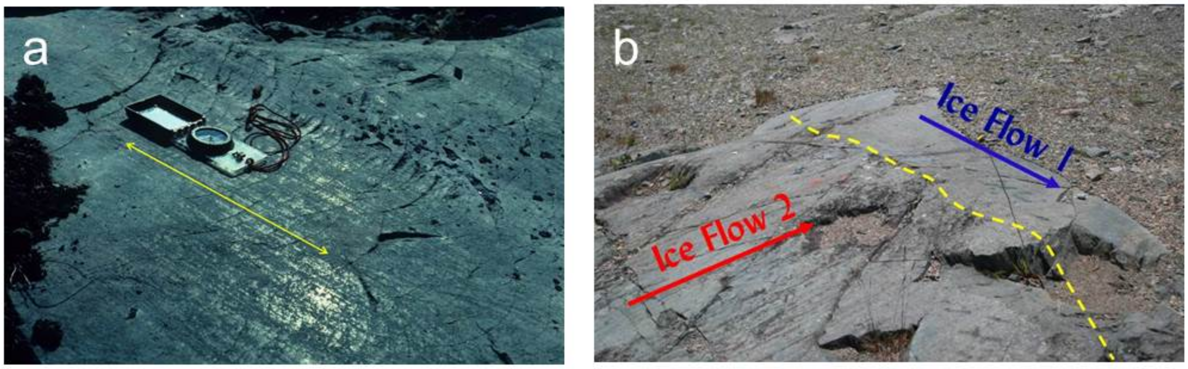

Figure 14. *A). Striations on a flat rock surface, Central Mineral Belt, Labrador. B) Multiple ice-flow directions; striations from Ice-flow1 are preserved in the lee of Ice-flow 2. Dashed yellow line marks sharp slope break resulting from glacial sculpturing.*

Sampling strategies

A variety of sampling media can be used in glaciated terrain depending on the scale of the sampling survey; these include lake sediment and water sampling (reconnaissance to regional scale), stream sediment sampling (reconnaissance to local scale), soil sampling (dependent on the size of the target), and till sampling (reconnaissance to local scale; discussed below). The effects of glacial transport will be manifest in all these media; therefore, an understanding of glacial history will be just as important as when sampling till.

Till sampling

Many glacial features may be difficult to see on the ground, and often they are first identified and mapped from air photos that are later ground-truthed (economics/logistics are also a factor). This generally guides sampling programs and helps interpret the geochemistry data.

Ideally, the shortest-travelled sediment is sampled so that the bedrock source of any anomalies can be found quickly. An understanding of geomorphology is necessary for these interpretations (*e.g.*, landforms oriented parallel or perpendicular to ice flow likely contain basal till, which is typically short travelled and hummocky terrain typically contains farther traveled material). In many places where basal till was deposited, a thin veneer of supraglacial sediment was laid down over it. Where possible samples should be taken below this supraglacially-transported sediment, within what is often basal till. The general characteristics of basal and englacial till should be borne in mind when interpreting geochemical, indicator or boulder-tracing results.

Sampling is most commonly done from hand dug pits, usually 50-80 cm in depth. Samples can also be obtained from mud boils (restricted occurrence at lower latitudes), road cuts, stream cuts, freshly-dug basements, and trenches. Sampling is usually conducted on a grid (less true now than it used to be, with ready availability of GPS) and sample density is dependent on the survey type (reconnaissance versus detailed) and size and shape of the mineralized target. The orientation of the grid and the spacing of samples with respect to ice-flow direction are very important. For example, if ice flow was to the east, sample spacing should be tighter in a north-south direction to maximize the probability of detecting an anomalous geochemical response. If sampling mineralized boulders, their distribution can be used to estimate the size of the source target and the grid sample spacing may be adjusted to reflect the anticipated dimensions of the source. It is also important to consistently sample only one soil horizon, ideally the C horizon or where this is not possible, to take notes on the sampled horizon (discussed in following section).

Soil Profiles

In glaciated areas like Newfoundland and Labrador, a soil is the upper weathered part of the unconsolidated sediment at the ground surface. Most soils in Newfoundland are classified as *podzols* (although other soil types, such as regosols and gleysols are also found). A podzol generally consists of three horizons: A, B and C. The Ah horizon is generally dark brown to black, composed mainly of organic remains (mixed with mineral material to a greater or lesser extent), and ranges from about 5 to 20 cm thick (**Figure 15)**. There are limited prospecting applications for the humus layer (some success stories from Ontario). In forested areas of Newfoundland the base of the A horizon is marked by a thin white to beige-coloured horizon referred to as the Ae horizon, a zone of intense leaching by acids from plant roots, organic decay and slightly acidic rain and snow. All minerals and elements or interest to mineral exploration are removed or severely depleted in this horizon. Geochemical analyses from samples in this horizon are not normally useful (an exception would be if prospecting for resistant minerals like beryl).

The B horizon is generally a rusty brown and is enriched by the elements leached from the A horizon. It is also enriched in metallic elements that are contained in the decaying organic matter derived from overlying plants and trees. This horizon ranges from about 10 to 50 cm thick and is a good sample source for geochemical analysis.

The C horizon is primarily an unweathered sediment found directly on top of the bedrock. Most C-horizon sediments are grey or dark brown in colour. Geochemical anomalies in the C-horizon are commonly lower in overall concentration level than in the B horizon but their concentration is closer to that of the material from which the soil is derived, and has not been enriched by weathering. Till samples collected routinely by the Geological Survey are from the C-horizon (ideally) or BC-horizon.

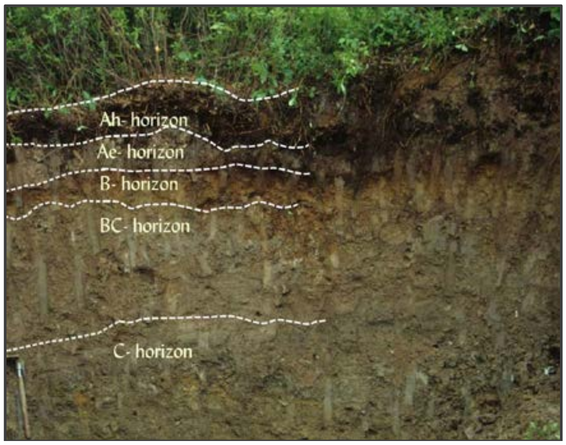

Figure 15. *Profile through podzol showing soil horizons.*

Glaciofluvial and glaciolacustrine deposits

Glaciofluvial and *glaciolacustrine* deposits are laid down by rivers and in lakes from the meltwater of receding glaciers. Sand and gravel are deposited by flowing water either under or in front of a glacier as it drains away, or by rivers and streams that flowed long after the glaciers had melted. Sediment deposited by this flowing water cannot be easily traced to its source because it was derived by erosion of other unconsolidated, transported material such as till. The creation of this sediment is a least one step removed from relatively simple glacial transport. The source of silt and clay is even more obscure, because it is likely washed out of till, carried in suspension by flowing water and then deposited in a standing body of water. If a geochemical anomaly is discovered in this sediment then it must be traced back to its till source and then to its glacial source.

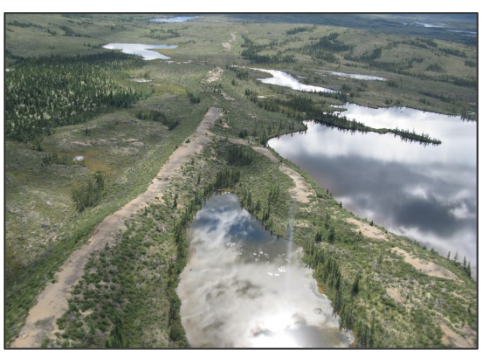

Figure 16. *Eskers, western LAB.*

Eskers form as ice is melting and consist of sinuous sand and gravel and cobble ridges that have been subjected to the washing of finer material, which also helps to concentrate heavy minerals (**Figure 16).** Their orientations generally follow the retreat phase of ice flow. The transportation distance of the sediment can vary from <1 km to >200 km. The uppermost coarse *lag* is the "last gasp" of meltwater in the glaciofluvial system and likely derived from washing of local (<25 km) tills or previously deposited *ice-contact sediments*.

Glaciomarine deposits

Deltas are generally composed of fluvial sands and gravel, overlying marine muds, formed at the front edge of ice where it meets the sea or a lake. They are often marked by a flat top that was once a flood plain, with coarse gravelly deposits on its surface. They help constrain the *marine limit* for an area which marks the highest elevation that sea level reached at some point in the past. Caution must be exercised when sampling below this elevation as sediment may have become reworked through wave processes. There may also be glaciolacustrine delta deposits around lakes which tell us about previous higher lake levels, and while not suitable for samples may be potential sources for aggregate. They are often recognized as terraces with breaks in slope.

Sea Level Change

Understanding the distribution of marine sediments is important in planning geochemical sampling, because, marine deposits are more difficult to link to bedrock source than primary glacigenic sediments. The Newfoundland and Labrador coastline shows numerous raised beaches and deltas, which mark the position of the coast as sea-level fell during deglaciation. Today, much of the Newfoundland coast is sinking as a result of continued settling of the crust, although the coast of Labrador continues to rise as a result of isostatic adjustment.

In Newfoundland, landforms that mark the highest level of the sea are found at higher elevations towards the northwest, with the highest marine limits found on the tip of the Northern Peninsula. Most of the island shows raised beaches, apart from small areas on the Avalon Peninsula. In Labrador, the highest raised beaches are in the southeast where beaches occur up to about 150 m above current sea level. The marine limit decreases northward, to about 55m in the Torngat Mountains, and 17 m at Cape Chidley.

Glacial Dispersal

The ice-flow patterns from striations indicate the general direction of ice-flow, but commonly do not provide data on the specific location of glacial dispersal or distances of glacial transport. Dispersal includes the entrainment, deposition and transport of glacial debris. If the bedrock source is visually or chemically distinctive, the path of ice movement may be recorded as a dispersal train. Dispersal can be mapped using the following clast-size fractions (> 5mm: pebbles, cobbles, boulders), till (< 0.063 mm (silt/clay; most commonly used) or < 0.002 mm (clay; uranium exploration): geochemistry for diagnostic elements), and heavy minerals (0.25 to 2.0 mm). There are several key influences on dispersal patterns:

1. Lithology of the bedrock source. The amount of material dispersed by ice is mainly a function of the nature of the source of the material. The amount of bedrock eroded is dependent on its nature (hard versus soft, permeable versus impermeable), and the bedrock topography and structure (jointed versus massive). Hard, massive, impermeable rocks (*e.g*., granite, rhyolite) will provide relatively little material in contrast to softer bedrock such as limestone or shale. The texture of the till is also determined in part by the nature of the bedrock.

2. Topography of dispersal area (flat vs irregular). A bedrock obstruction may cause higher subglacial pressure on the up-ice side of the obstruction and a rough topography may cause pressure variations which may enhance erosion or create topographically-controlled ice flow. The angle between ice-movement direction and the strike of the rocks is also important. For example, background gold content in the Le Ronge Belt in Saskatchewan, where ice movement was sub-parallel to the strike of the greenstone belt, is much higher than in the Abitibi, where they are mutually sub-perpendicular, despite the Abitibi being much more extensive and productive of gold mines.

3. Deposition and transportation of glacial debris. Debris is entrained in the basal zone of the glacier in the *tractive layer* between the ice and the immobile bed. The densest concentration of debris occurs near the base of the ice in basal or lodgement till and this material has the shortest transport path. Debris concentration is less scattered throughout or on the surface of the ice and this supraglacial material is more far travelled.

4. Changing patterns of deposition and erosion. Near ice divides there are likely to be more complicated vertical and lateral changes in dispersal due to changing patterns of erosion and deposition. This will be reflected in the shape of the dispersal train, discussed in the following section.

5. Concentration of component decreases down-ice until it merges with natural background. The more distinctive the component, the further it can be traced. Boulders are more distinct and related directly to bedrock, making them easier to track than geochemical signatures.

The pattern of glacial dispersal may be determined from simply plotting the distribution of clasts on a map to show the general shape of the distribution pattern. Quantitative approaches to clast dispersal can be done using several methods. Average transport distances can be estimated from plots of clast lithology against distance from source. This transport-distance distribution method identifies the rock type and probable source of a number of clasts or boulders at a given site. The range of distances to source area for a given rock unit is estimated by plotting the ice-flow direction onto a bedrock map, and measuring the distance up-ice to the proximal and distal contacts of the unit. The distance is plotted against cumulative-percentage composition on log-probability axes. A straight line can usually be fitted to the distribution and the mean transportation distance estimated by graphical methods.

The half-distance method can also be used to estimate transport distance from clast lithology. The half distance is the distance at which the frequency falls to half its original value. Another method is the transport distance distribution method which plots the log-normal distribution of each rock type found at a site against the distance to the nearest source. Half-distances are calculated using linear regression and provide an indication of the transport distance of indicator clasts.

Dispersal Trains

A mineralized dispersal train of distinct boulders or geochemical anomalies is an elongate lens of till, oriented parallel to ice flow, which is normally hundreds or thousands of times larger in area than its bedrock source, making it easier to detect than the bedrock mineralization from which it is derived, especially in drift-covered areas. Near the source, at the 'head' of the train, concentration is greatest; further down-ice, it declines gradually to slightly above normal levels. The down-ice end of the dispersal train is referred to as the 'tail'. A dispersal train is not, however, a two-dimensional feature. The third dimension, vertically through a till sheet, is critical to the understanding of the dispersal train and the focusing of follow-up trenching or drilling. The ability to follow the dispersal train vertically down towards bedrock will help to focus exploration, especially if overburden is too thick for trenching alone (**Figure 17**).

Dispersal trains vary in shape depending on their ice-flow influences and may also be affected by post-depositional processes. They may occur as elongate straight to slightly sinuous lines (ribbon-shaped) as the result of ice flow in only one direction (*e.g.*, Strange Lake; **Figure 18**). Dispersal patterns that widen laterally in the down-ice direction (fan-shaped) are the result of ice flow in two or more directions (*e.g.*, Central Mineral Belt; **Figure 19**) and amoeboid-shaped trains form from complicated ice flow in multiple directions.

Clasts or boulders may also be sampled for rock-type analyses. Depending on the mineral being sought, all that is required is an easy-to-recognize rock type that is significantly different from other rocks found on the ground surface and has some geologic relationship to the target mineral or commodity. These are then compared with bedrock geology maps to provide information on distances and directions of dispersal and are dependent on accurate bedrock mapping.

When sampling clasts, it is recommended that at least 100 clasts are collected and any rare exotic clasts are noted. While this may over-represent the exotic component, it has the advantage of providing a gross estimation of clast proportions while allowing definition of the dispersal of exotic clasts. The pattern of glacial dispersal can then be determined by plotting the distribution of clasts on a map such that the gross shape of glacial dispersal trains may be shown.

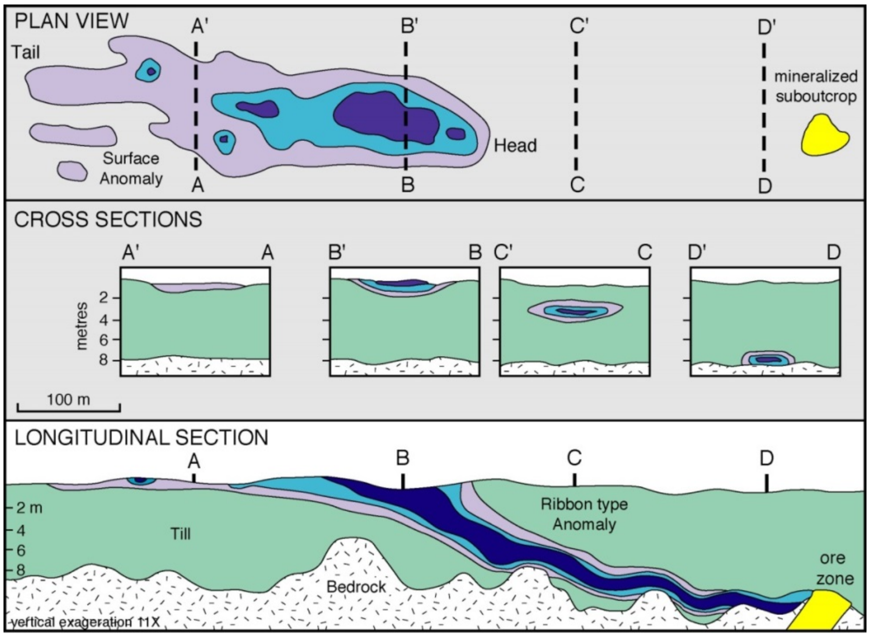

Figure 17. *Schematic sections and plan of glacial dispersal from an ore zone. Assuming a single till-depositing event, the distance 'up-ice' to the bedrock source of a geochemical anomaly or mineralized boulder, detected at surface is dependent on till thickness (Miller, 1984).*

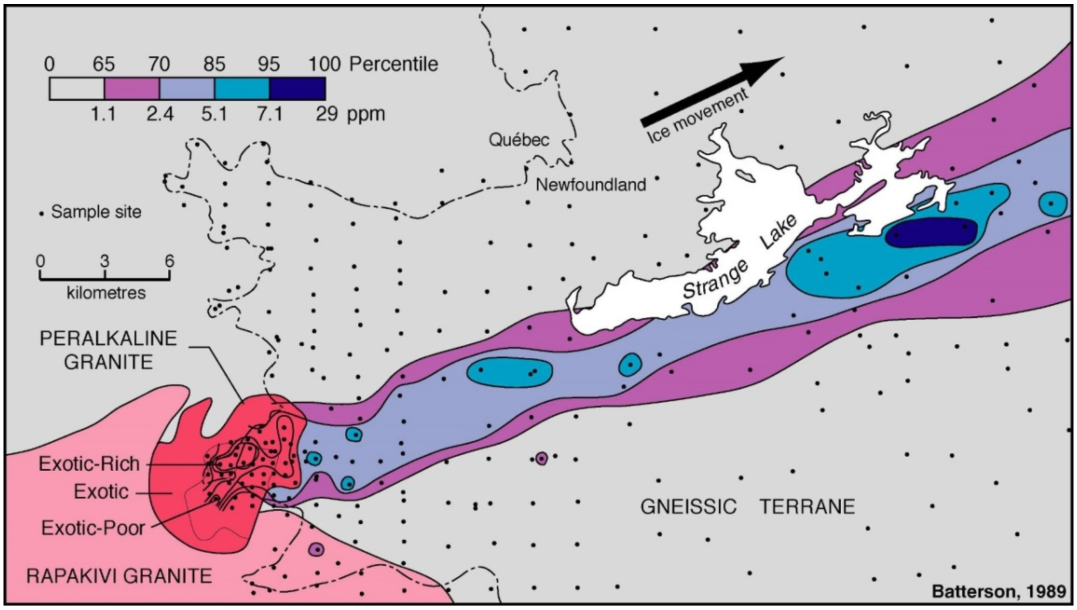

Figure 18. *Ribbon-shaped dispersal train showing dispersal of Beryllium (Be) from the Strange Lake rare earth-rare metal deposit, Labrador (Batterson, 1989).*

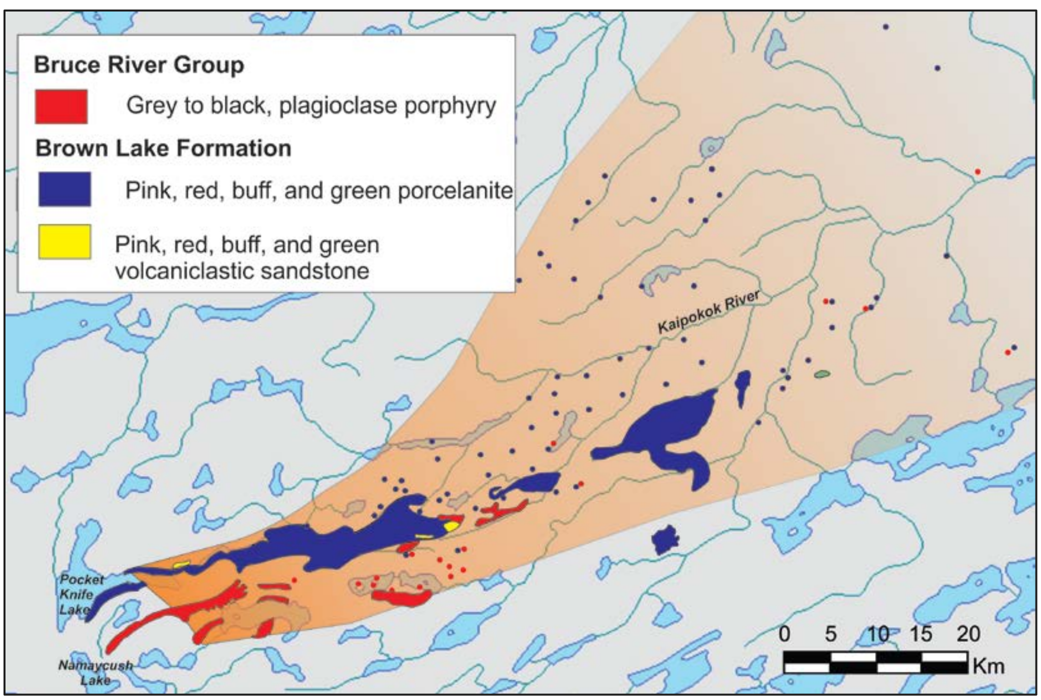

Figure 19. *Fan-shaped dispersal train (shown in orange and resulting from 2 ice-flow directions) of indicator erratics that broadens down-ice in the Central Mineral Belt, Labrador (Batterson and Liverman, 2000).*

The differences in glacial histories between Newfoundland and Labrador are reflected in their glacial dispersal patterns. The Island was covered by multiple coalescing ice caps which are represented by short, diffuse dispersal patterns. Dispersal trains are commonly less well-defined, especially geochemically, due to compositional similarities in adjacent rock units. Labrador, on the other hand, was covered by the Labrador Sector of the Laurentide Ice Sheet with a dispersal centre in western Labrador. Because of this, dispersal patterns have an amoeboid pattern in the Labrador Trough, for example, shown by a series of dispersal trains in several directions, commonly of limited extent. Dispersal trains away from the ice dispersal centres are commonly longer and ribbon-shaped (*e.g.*, Strange Lake) or fan-shaped (*e.g.*, Central Mineral Belt).

Gold dispersal

Glacially transported gold grains commonly show evidence of their glacial and postglacial history in their shapes and surface textures. These features can aid in locating the gold source area by indicating the relative distance of transport, the possibility of recycling and postglacial weathering effects.

In the classification scheme developed by Overburden Drilling Management, sand and silt-sized gold grains can be classified into *pristine*, *reshaped* or *modified* grains. Pristine grains include those that have not travelled far from source, or gold grains that have weathered postglacially out of unstable host rocks found within the till. Reshaped grains consist of recycled fluvial and/or farther-travelled gold. They are typically modified in shape by rounding, flattening, etc. Reshaped grains are difficult to follow back to a bedrock source.

Summary

- Glacial erosion, deposition, and transport reflect glacial history and ice-flow dynamics.
- A multi-faceted approach, including striation mapping, surficial mapping, geochemical sampling, and clast-provenance analysis, should be employed in any drift exploration program.
- Mineral abundance, chemistry, shape and surface features may provide important information about bedrock source, including style of mineralization, host lithology, alteration, or grade, as well as distance of glacial transport.
- Glacial dispersal can be mapped using different size fractions: till (silt to clay fraction for geochemistry), indicator minerals, and mineralized boulders.
- Glacial dispersal is a result of all ice-flow events; the size, shape, orientation and composition of glacial dispersal trains reflect: change in ice-flow direction and provenance, and the distance of glacial transport related to either change in flow velocity or to debris position within the ice.

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