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Are drumlins a product of a thermo-mechanical instability?

Roger LeB. Hooke *, Aaron Medford

School of Earth and Climate Sciences and Climate Change Institute, University of Maine, Orono, ME 04469-5790, USA

A R T I C L E I N F O

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A B S T R A C T

Of numerous theories of drumlin genesis, none has been widely accepted. It seems evident, however, that some form of positive feedback process is involved. Under certain circumstances perturbations are amplified. Herein we suggest that patchy areas of frozen bed provide the initial perturbation. Such frozen patches may occur in local areas underlain by material of lower thermal conductivity or on slight topographic highs. Drag exerted by the frozen patch deflects ice flow into its lee, dragging with it mobile till eroded from the thawed area. The energy balance is such that this till likely refreezes, either producing a topographic perturbation or amplifying an existing one. The resulting topography then deflects more of the geothermal heat away from the developing hill and into the adjacent trough, resulting in a positive feedback. Once the thermal perturbation exceeds a critical (though as yet undefined) level, melting may decouple the ice from the bed, preventing further entrainment of till from thawed areas, and thus limiting the height and length of the drumlin.
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Introduction

Previous work

Glaciers are impressive earth-moving machines so it is hardly surprising that, under suitable conditions, they should produce streamlined forms. Examples include crag-and-tail features on bedrock outcrops, roche moutonnées, rock drumlins, drumlins, flutes, and mega-scale glacier lineations. Herein, our focus is on drumlins. Drumlins are elongate streamlined hills with modest aspect ratios, horizontal dimensions of order 10^2 to 10^3 m and heights of order 10^0 to 10^1 m. They are common features of the beds of former ice sheets (e.g. Colgan and Mickelson, 1997; Hess and Briner, 2009), and have also been described from the margins of contemporary ice caps (Hart, 1995; Johnson et al., 2010). They likely form in several ways. At one end of the spectrum are those that may have arisen spontaneously from flow over a relatively flat homogeneous bed; at the other end are those formed by simple streamlining of preexisting hills. As Clark (2010) noted, many models require a process that amplifies the relief. Rates of formation must also vary. Smith et al. (2007) appear to have documented formation of one beneath an Antarctic ice stream within 7 yr.

Our interest is in drumlins formed under glaciological conditions that are tamer than those beneath ice streams. These drumlins commonly occur in fields of several tens or hundreds. We speculate that they can arise from flow over a relatively flat bed, but not a homogeneous one. The inhomogeneities, we suggest, are materials that are more resistant to deformation by the overriding ice, a possibility first proposed by Smalley and Unwin (1968).

Smalley and Unwin (1968) thought that before an ice sheet could erode a bed consisting of clastic material (e.g. till), the material would have to dilate. They reasoned that patches of coarser till would resist dilation. Erosion on either side of these patches (and likely deposition on them) would then amplify the perturbation. Baranowski (1977) suggested that the resistant patches, rather than simply being coarser, might be frozen. He viewed these frozen patches as occurring at a phase-change boundary from a thawed to a frozen bed, and believed that the patches might develop when atmospheric temperatures periodically dropped for short periods of time immediately preceding and during deglaciation.

More recently, Hindmarsh (1998, 1999), Fowler (2000, 2009, 2010), Schoof (2007), and Chapwanya et al. (2011) have explored the possibility that drumlins form from an instability arising from ice flow over deformable till. Instabilities are situations in which an *infinitesimal* perturbation sets up positive feedbacks that amplify the perturbation. Fowler (2009) compared the drumlin-forming process with the way water flowing over sand generates ripples and dunes. Hindmarsh (1999) and Fowler (2010) used a linearly viscous ice rheology in their calculations and were able to show that certain combinations of till thickness and rheology, sliding speed, and perturbation wavelength led to development of waves in the ice–till interface *normal to the direction of ice flow*. They thought ribbed moraine might be a result of this process.

Schoof (2007, p. 228) paraphrases the Hindmarsh-Fowler instability mechanism as follows: "When ice flows over a shallow bump in the ice–till interface, it exerts higher compressive normal stresses on the upstream side of the bump than on its downstream side. If the viscosity of ice is…much greater than that of till…more sediment flows [longitudinally] into the bump than out of it. [T]his causes the bump to grow."

Schoof extended the Hindmarsh-Fowler analysis to more plastic till rheologies, but was still unable to identify a three-dimensional instability

* Corresponding author. Fax: +1 207 581 2202.
E-mail address: rogerhooke@gmail.com (R.L. Hooke).

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that might lead to drumlins. He thought the model was likely conceptually flawed as it predicted migration of the drumlins downflow, whereas the cores of some drumlins are composed of fluvial gravels that were unlikely to have migrated. These gravels were probably deposited in a proglacial environment prior to advance of the ice. Fowler (2009), however, modified Schoof's approach and obtained waves that were stationary but still two-dimensional.

Drumlin characteristics and likely conditions of formation

In a comprehensive review of literature on 95 drumlin fields worldwide, Patterson and Hooke (1995) found that fields of distinct well-developed drumlins tended to occur within ~80 km of an associated ice margin, but separated from it by a drumlin-free zone. This is illustrated particularly well by drumlin fields left by the Green Bay lobe (Wisconsin, USA) as it retreated. The axes of drumlins in each field are normal to the recessional moraine behind which they are believed to have formed, but are not normal to earlier recessional moraines further away (Colgan and Mickelson, 1997). Some authors (e.g. Clark, 1999 and written comm., July 2012) think drumlins can form much further from the margin, but firm evidence for this appears to be lacking. This is certainly, in part, because it becomes increasingly difficult to reliably chronologically correlate a drumlin field with a margin as the distance between the two increases.

The area within several tens of kilometers of an ice sheet's margin is one (the ablation zone) in which longitudinal strain rates were likely becoming compressive, vertical velocities upward, and the ice relatively thin. Down-ice-diverging drumlin axes suggest that transverse strain rates were commonly extending. Numerical modeling (Patterson and Hooke), till deformation (Stanford and Mickelson, 1985; Menzies et al., 1997), rare fluvial interbeds (Goldstein, 1994), and other lines of evidence suggest that the basal temperature was at the pressure melting point when drumlins formed. Patterson and Hooke also noted that many well-developed drumlin fields lie upglacier from an ice margin that either ended in the sea or showed evidence of being frozen to the bed. This led them to argue that the pore water pressure was likely high, promoting bed deformation.

Patterson and Hooke found that drumlin shape and composition, substrate material and thickness, and the broad-scale topography of drumlin fields were highly variable, suggesting that none of these are crucial for drumlin formation. Stokes et al. (2011) reached the same conclusion, and suggested that a single process (or suite of processes) were likely responsible for the continuum of drumlin forms. They concluded that any proposed mechanism of drumlin formation must be able to account for the wide range of morphologies seen. The intensive search for this mechanism, starting in the late 19th century (e.g. Tarr, 1894 and references therein) serves as a reminder of our lack of understanding of a common subglacial process.

Basal thermal conditions

Parts of the beds of the Greenland and Antarctic ice sheets are at the pressure-melting temperature and others are frozen (Oswald and Gogineni, 2008; Bell et al., 2011). The same was undoubtedly true of the now-vanished ice sheets of the Pleistocene (e.g. Kleman and Glasser, 2007). A likely location for a transition from a frozen bed (up-ice) to a thawed one (down-ice) is beneath the equilibrium line. Near the equilibrium line, vertical ice velocities change from downward in the accumulation zone to upward in the ablation zone. This reduces the thermal gradient in the ice, thus "trapping" geothermal and frictional heat at the bed (Hooke, 1977). Intuitively, such transitions must occur over distances of several kilometers, with patches of frozen bed surrounding or surrounded by thawed areas (Fig. 1; Hughes, 1992). This patchiness could be associated with slight topographic irregularities (e.g. Kleman and Borgström, 1994) or with inhomogeneities in thermal conductivity of the bed material. Where conditions are right (and we do not know exactly what 'right' is in this context), erosion is likely in the thawed areas, resulting in development of some relief (or amplification of relief already present).

The idea that frozen areas might stimulate drumlin growth is over 60 yr old; Armstrong and Tipper (1948, p. 293) suggested that 'a knob of frozen till' might do the trick. Other glacial landforms have also been attributed to processes in the vicinity of a water/ice phase change boundary at the bed. Most notably: (*i*) Hättestrand and Kleman (1999) have attributed some ribbed moraine to tensional fracturing of a meters-thick layer of frozen till, (*ii*) Kleman and Borgström (1994) describe landforms developed around frozen patches in areas of likely *extensional* ice flow, and (*iii*) Jansson and Kleman (1999) describe bedrock knobs with paired horns of till extending down-ice. Jansson and Kleman attribute these latter landforms to flow on a thawed bed at lower elevations around a hilltop that is frozen. Preservation of the landform, they posit, requires that totally frozen bed conditions resume later.

Our conceptual model

If a frozen patch is strong enough to resist entrainment by the glacier, it increases drag on the glacier sole relative to that in adjacent thawed zones. Conservation of mass then requires that a slight dimple develop in the glacier surface. (Depressions in the lees nunataks are an extreme form of this dimpling.) Surface slopes into the dimple from either side drive a herringbone flow at the bed in the lee of the up-ice edge of the frozen patch (Fig. 2). Such flow is also theoretically possible in a purely 2-D situation, as on the flanks of a flute of uniform height (Schoof and Clarke, 2008). Evidence for such flow is seen in the herringbone patterns found in till fabric studies on some drumlins (Andrews and King, 1968; Embleton and King, 1975, p. 412) [although not on others (e.g. Walker, 1973)] and on flutes (Shaw and Freschauf, 1973; Rose, 1989; Benn, 1994). [Most till fabric analyses on drumlins are done in places, as along the drumlin axis or at the stoss end, where one would not expect to see the herringbone pattern. Others are at cross sections exposed by erosion an unknown distance from the stoss end (e.g. Andrews and King, 1968; Embleton and King, 1975). As a herringbone pattern is expected to be best develop midway along the drumlin (Fig. 2), these studies are inconclusive.] Under suitable conditions, the ice sliding diagonally into the lee of such a perturbation will drag wet till with it.

Both topographic and thermal inhomogeneities affect heat flow to Earth's surface. Isotherms deep in the earth are roughly spherical, parallel to Earth's mean surface. Just beneath the surface, however, isotherms must warp to conform to the topography (Fig. 3) or to the presence of a cold patch. Heat flow is normal to isotherms, so near the surface beneath a topographic irregularity geothermal heat moves towards valleys and away from ridges (Fig. 3; Lees, 1910; Lachenbruch, 1968; Turcotte and Schubert, 2002). A frozen patch resulting from a conductivity contrast would have the same effect, albeit with a more complicated temperature distribution than that shown in Figure 3.

(For convenience in the following discussion, we refer only to topographic perturbations, but the principles apply equally well to perturbations resulting from conductivity contrasts. We also refer to the "ridge" and "trough" of the perturbation. These regions are identified in Figure 3. The ridge is assumed to be frozen and the trough thawed.)

In the ice immediately above the bed, the curvature of the heat flux lines is reversed. Thus, more heat is conducted upward into the ice from ridges than from troughs (Fig. 3).

Thus, the energy balance in the troughs is positive (more heat reaching the interface from below than is withdrawn upward), leading to melting. Conversely, the energy balance over ridges is negative, resulting in freezing.

The negative heat balance on the ridges should freeze at least some of any till dragged to this location by the converging ice flow, increasing the height of the perturbation. At the same time, the trough is being deepened by erosion. This increase in relief increases the difference in heat flux, and growth of the perturbation proceeds in a positive

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