MELTWATER CONTROLS ON ICE-MARGINAL SEDIMENTATION

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DECLARATION

I certify that the work in this thesis is my own, unless stated otherwise. This work has not been submitted for any other degree.

Nick Spedding

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ABSTRACT

This thesis explores the influence that meltwater exerts on styles of ice-marginal sedimentation, using past and present examples from Iceland. The study glaciers display marked contrasts in form, size and composition of moraines which are unlikely to reflect differences in rates of subglacial erosion. This is because the study glaciers occupy a similar climate, show similar relief, sit above similar bedrock, and are inferred to flow at similar speeds. The observed variation in moraine properties must reflect some other process which intervenes to modify sediment transport relationships prior to the arrival of debris at the ice edge. I argue that this key factor which controls sediment transport - and, as a result, the potential to form moraines - is the behaviour of subglacial meltwater flows.

Studies of the sediment load of its outlet river show that Sólheimajökull is a highly erosive glacier, yet the quantity of debris carried by the *ice* is extremely small. Consequently, present-day moraine formation is extremely limited. This can best be explained as the product of an aggressive subglacial drainage network which captures and evacuates the bulk of debris generated by subglacial erosion. This state of high efficiency subglacial flushing is likely to dominate the sediment budget of many temperate glaciers.

Whereas the present-day margin of Sólheimajökull is debris-poor, the present-day margins of Gígjökull and Steinholtsjökull are debris-rich. This debris consists of two major populations: 1) rounded clasts set in a sorted coarse sand and gravel matrix, derived from a series of englacial debris bands, and, 2) sub-angular clasts in a poorly-sorted matrix, derived from unusually thick sequences of basal ice. Overdeepened basins lie beneath the termini of both Gígjökull and Steinholtsjökull. It seems that changes in water flow in this zone - rising water pressures associated with water flow upslope cause drainage to take up an englacial route - explain both the debris bands and the basal ice. The debris bands form as sediment-laden englacial channels close-up; simultaneously, the paucity of water at the glacier bed, in conjunction with strongly compressive ice flow, favours widespread preservation of basal ice. I extend Hooke's model of the ice-fall/overdeepening as a process system favourable to subglacial erosion to argue that it also creates conditions favourable to debris retention in ice, and so rapid accumulation of ice-marginal moraines.

The idea that contrasts in the behaviour of drainage account for contrasts in moraine development adds depth to studies of the glacial geologic record, and its interpretation in terms

of past climate change. Contrasts in the size of Sólheimajökull's Holocene moraines - those of the early Neoglacial are small, those of the Little Ice Age are big - can be explained by changes to levels of stream power/flushing efficiency as ice geometry changes in response to climate change and ice-divide migration. The restricted late Holocene advance of Gígjökull, and the size of its moraine rampart, appear to reflect the barrier action of the growing moraine. The high rates of ice-marginal deposition necessary for this to occur are consistent with past intensification of the ice-fall/overdeepening sediment transport regime. This moraine dam effect explains why the record of climate change at Gígjökull is unreliable. In contrast, the Holocene record of Steinholtsjökull - which appears to show a simple relationship with climate - implies that the past influence of the overdeepening was much diminished, with a high efficiency flushing regime installed as the ice thickened and advanced. However, an elevated area of fluted basal tills, indicative of relatively high rates of subglacial debris deposition, seems to reflect isolation from major subglacial water flows.

This thesis was written with the specific intent of linking process and form in such a way as to provide a meaningful explanation of moraine development. Different moraine forms are the emergent product of the multitude of process interactions which make up a complex causal network dispersed in time and space. The key factor which regulates this is the behaviour of meltwater. Glacier drainage systems provide the crucial contextual element which links the basic level of process (reductionist analysis) to the level of surface appearance (observations of moraine form). Drainage systems carry genuine causal powers which cannot be broken down into smaller process systems without the loss of explanatory power. This ties in with realist traditions of science, and recent ideas associated with complexity theory.

HETTY TO NANCY...

In the centre of Iceland there are only three kinds of scenery - Stones, More Stones, and All Stones. The stones are the wrong size, the wrong shape, the wrong colour, and too many of them. They are not big enough to impress, and not small enough to negotiate. Absolutely unpicturesque and absolutely non-utilitarian... Maisie was disgusted. She said it was like after a party which no one had tidied up. It's certainly hard to think how a country gets in a mess like this. A geologist would know, I suppose...

W. H. AUDEN and LOUIS MacNEICE
Letters from Iceland
1937

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I. INTRODUCTION

PREFACE

This thesis explores the ways in which meltwater controls the processes and patterns of ice-marginal sedimentation. Previous studies have tended to portray ice-marginal depositional landforms either as the product of ice action only, or as the product of a (supposedly) distinct suite of fluvio-glacial processes. Such processes are implied to give rise to (supposedly) distinct fluvio-glacial landforms. This involves an arbitrary distinction between ice- and water-related process regimes, the transition between which occurs abruptly at the edge of the ice. In this thesis I argue that it is preferable to study ice-marginal sedimentation using a wider framework which views the action of meltwater within, beneath and immediately adjacent to alpine-type glaciers as an inseparable component of a complex but coherent sediment transfer system. Recent research shows that a wide range of drainage styles can exist beneath glaciers, with important effects on ice flow behaviour and processes of subglacial erosion. It is logical to suspect that these contrasts in water flow must also create important differences in the delivery of sediment to the ice edge. What follows is my attempt to examine the geomorphic impact of meltwater on ice-marginal sedimentation at the scale of the individual glacier, using examples of contrasting Icelandic glaciers.

This thesis is inspired by a number of specific observations, some of which are substantive, others of which are conceptual. The substantive observations stem from past research, and so can be considered as 'facts', with the exception of the final point, which seems to have been overlooked in previous work (hence this thesis!), but which is little more than a statement of common sense. The conceptual points are perhaps more controversial, because they relate not to fact, but to different views as to the way in which science should be done. My argument is that meaningful study of how and why moraines form requires an approach which differs somewhat from the two traditions which currently dominate research.

SUBSTANTIVE

- Subglacial drainage occurs in several different ways (e.g. Hooke, 1989; Paterson, 1994, Ch. 6; Chapter 1.1).
- The volume and configuration of subglacial meltwater discharge at alpine-type glaciers exerts an important influence on the dynamics of ice flow (e.g. Willis, 1995; Chapter 1.1) and on processes of subglacial erosion (e.g. Iverson, 1995).

- Meltwaters draining alpine-type glaciers commonly carry large quantities of sediment picked up at the glacier bed (e.g. Maizels, 1995, pp. 369-374; Hallet *et al.*, 1996; Chapter 1.3).

- The origin of many ice-marginal deposits is not clear-cut. The term 'moraine' is often used loosely to describe any accumulation of sediment which relates to the past or present position of glacier ice, irrespective of its origin. A strict definition limits the use of 'moraine' to describe a distinct depositional landform made up of till, with till defined (*sensu stricto*) as sediment dumped directly out of ice, without significant re-working by other agents of debris transfer (Shaw, 1977). Thus many 'moraines' are strictly water-laid features such as kames or outwash fans, which contain sediments and structures indicative of debris transport by, and sedimentation out of, water (e.g. Warren and Ashley, 1994). Conversely, deposits composed of water-worked material (and sometimes given 'classic' fluvio-glacial labels such as kame or esker) can be dumped directly out of ice, and so qualify as 'genuine' moraines (e.g. Rubulis, 1983; Schlücter, 1983; Kirkbride, 1989; my observations; Chapters 4 and 5).

- Ice-marginal deposits can accumulate only if sediment enters storage faster than it is removed. It is the balance between different processes which is important, not just any single process. Thus the strength and frequency of potentially-erosive subaerial water flows must exert a major control on preservation of moraines. The rate at which sediment is delivered to the ice edge will also be important; this is the problem I examine in this thesis. Rapid delivery of large quantities of debris in ice will favour moraine build-up. This must in turn relate to the process balance which holds upglacier. The same debris cannot be carried by ice and water simultaneously. Debris tapped by the subglacial drainage system, and washed-out beyond the ice margins, cannot form moraines in the strict sense, but it is likely to contribute to some kind of proglacial river deposit. Debris retained within the ice is likely to end up as part of a moraine.

CONCEPTUAL

- Research related to glaciers tends to be split between two distinct traditions: that of glacial geology, and that of glaciology (Chapter 1.4). If anything, I think this tendency has become increasingly pronounced over the last twenty years or so. By glacial geology (Boulton, 1987) I mean work which prioritises field data, and which tends to stress the study of form, together with other 'visible' properties such as the character of sediments, but which largely avoids detailed reference to processes. In contrast, glaciology (Clarke, 1987) adopts a theoretical rather than an empirical perspective, but tends to emphasise the mechanics of process without extending the analysis to consider what are the implications for development of landforms. Work tends to be drawn to one or other of these two traditions, with the result that glacial geomorphology, which seeks to link processes and forms (see below; Chapter 1.4; Harbor, 1993), tends to be squeezed out (cf. Sugden *et al.*, 1997).

- This division, and its negative impact on studies of geomorphology, is reinforced by the tendency of scales of investigation to split between large and small. Studies by glacial geologists tend either to describe in detail the landforms and sediments of a specific locality (usually that of a former ice margin), or to describe the record of large parts of an entire ice sheet (an enterprise given a major boost by recent developments in the technology of remote sensing and GIS). In both cases, the details of processes tend to be given little attention. Similarly, studies by glaciologists tend to divide between abstract, reductionist studies - using laboratory or mathematical techniques - of individual processes, or large-scale computer simulations of past or present ice sheets (which resolve ice flow processes, but usually ignore geomorphology). Relatively little research takes place at the intermediate, catchment-scale of analysis best suited to studies which try to link processes with landform development. Successful glacial geomorphology must work from within this middle ground.

- The tendency to divide the system of interest using arbitrary and arguably inappropriate boundaries means that geomorphic studies can be unnecessarily self-limiting (e.g. Richards, 1990; Chapter 1.2, 1.3 and 1.4). Static analysis of fixed relationships highlights discontinuities created by these divisions: e.g. moraines are often seen as the tangible phenomena which define the process boundary (at the edge of the ice) between separate, and supposedly distinct, glacial and subaerial/fluvial systems. Such a view disguises the complex reality of the catchment sediment transfer system, with the result that key process relationships tend to be ignored, with accounts of landform/landscape development impoverished as a consequence.

In this thesis I set out to present a study which draws together these eight statements. My view is that glacial geomorphology must aim to put together a meaningful account of the origin of landscape which draws on a detailed understanding of the mechanisms responsible (cf. Harbor, 1993, p. 129; Chapter 1.4). It is the commitment to linking process and form which differentiates glacial geomorphology from glacial geology and/or glaciology. If we accept this, it is arguable that glacial geomorphology has a past history of relatively poor performance. Our present understanding of moraine formation is far from complete, so the role of meltwater as a control on ice-marginal sedimentation certainly merits further investigation. Current knowledge suggests that meltwater must exert a substantial influence on processes and patterns of sediment transfer, but to take the study of these further requires a perspective which avoids the conceptual deficiencies of the traditions sketched above. (It is important to stress that these traditions work perfectly well for the study of other problems; the strength or weakness of a particular approach is not fixed, but varies with the research problem.) This study explores the linked processes which act at the full-catchment-scale to transfer water, ice and debris. It is these processes which are responsible for the style and volume of moraine formation; the precise conditions of deposition at the ice margin add secondary detail, but cannot in themselves provide a full explanation. By using the perspective of process-form linkage at the catchment-scale I aim not just to shed new light on geomorphic activity at the margins of present-day glaciers, but also to bring together a body of ideas which can be used to develop better-informed accounts of the behaviour of past ice masses.

CHAPTER 1
Background

INTRODUCTION

This chapter sets out the background to this thesis. Most of what appears here builds on the Preface. I begin with a discussion of the core ideas of glacier hydrology (1.1), and move on to sketch the chief aspects of glacier sediment transport theory (1.2), including the question of its relationship with what is often thought of (if only for the sake of convenience) as the separate sub-discipline of fluvio-glacial geomorphology (1.3). I then pick up on the conceptual observations of the Preface (1.4), and enlarge upon my critique of the twin traditions of glaciology and glacial geology. I set out my own vision of what glacial geomorphology should be, and how this might apply to the study of moraines (1.5). The chapter ends by explaining why the study sites used were chosen (1.6), followed by a brief statement of the structure of this thesis (1.7).

1.1 GLACIER HYDROLOGY: A REVIEW

The style of water flow through and beneath glaciers provides the core of ideas which I use to construct this thesis. Here I introduce the basic ideas which constitute essential background material. Theoretical and empirical work of the last thirty years, starting with the 1968 International Association of Hydrological Sciences Cambridge symposium, and rapidly consolidated by the 1972 issue of the *Journal of Glaciology* which carries the seminal papers of Shreve and Röthlisberger, has greatly increased what we know (or think we know!) of the different ways in which water flows within and beneath ice (Hooke, 1989; Paterson, 1994, Ch. 6; Willis, 1995, pp. 79-83 for recent reviews). We also know much more about how water influences glacier flow, and, both directly and indirectly, processes of subglacial erosion (see below). What we know far less about is how different styles of drainage interact with subglacial debris to determine the overall balance between different styles of sediment transport, and so different styles of ice-marginal deposition. This is a problem which has largely escaped the interest of practising glacial hydrologists and geomorphologists. It is this which I address in this thesis.

SHREVE THEORY (Shreve, 1972)

This describes the distribution of hydraulic potential within and beneath ice:

$$\mathbf{HP = A + B + C + D}$$

HP hydraulic potential

A reference potential (e.g. as set by the elevation of the glacier snout)

B potential energy of water, defined with respect to A: i.e. $\rho_w \cdot g \cdot z$
- ρ_w density of water
- g gravitational acceleration
- z elevation difference between water and A

C component of pressure (i.e. > atmospheric pressure) imparted by ice overburden: i.e. $\rho_i \cdot g \cdot h$
- ρ_i density of ice
- h thickness of ice above water

D adjustment made to take account of the local pressure drop in the vicinity of a major conduit (the 'Röthlisberger' term: see below)

Shreve theory can be used to reconstruct/predict the direction of water flow. Its major points are:

- Water flows perpendicular to contours of equal hydraulic potential ('equipotentials'): i.e. it follows the steepest hydraulic gradient.
- The contribution of ice overburden to total hydraulic potential tends to dominate, so water is driven in the direction of ice surface slope.
- This means water under pressure can flow uphill...
- ...unless the uphill (or 'adverse') bed slope is more than eleven times the overlying ice surface slope (in which case term B dominates over term C, and water flow follows the slope of the subglacial topography).

Shreve theory cannot tell us how much water is in transit (this will be a function of climate - specifically the surface energy balance of the glacier - and catchment size), nor can it tell us the exact type of channel within which water flows (see below). It works well with water under high pressure beneath thick ice which sits above a gently-sloping, relatively even bed: i.e. ice-sheets (cf. Shreve, 1972; Sugden *et al.*, 1991; but see also Sharp *et al.*, 1993). In this case the final term D is relatively unimportant. The simple Shreve equation (terms A, B and C) cannot account for:

1. The local drop in channel water pressure which results from ice-wall melt induced by heat released by water flow. This factor is assessed by Röthlisberger (1972) and Hooke (1984); see below.

2. The redistribution of ice overburden pressure as ice slides across an uneven bed. The local fall in pressure in the lee of bedrock obstacles permits cavities to form (see below). Water tends to collect in, and flow (slowly) between these cavities.

STRUCTURE OF THE SUBGLACIAL DRAINAGE SYSTEM

Existing work tends to structure its discussion of subglacial drainage by use of two dichotomies: 1) 'hard' bed versus 'soft' bed, and, 2) 'distributed' versus 'discrete' drainage (Fig. 1.1). Hard bed indicates rigid, impermeable bedrock; soft bed indicates permeable, unconsolidated, potentially deformable sediment. Distributed drainage is water flow dispersed widely and more-or-less evenly over the glacier bed; discrete drainage implies that the bulk of flow occurs within a few large channels.

Weertman film

[Fig. 1.1., No. 7, hard bed, distributed drainage. E.g. Weertman and Birchfield, 1983b]

This is a film (μm thickness) or sheet (mm thickness) of water which covers large areas of the glacier bed. It is perhaps the chief style of drainage beneath hard-bed, warm-based polar outlet glaciers and ice-streams which receive limited inputs of surface meltwater. Here the bulk of meltwater is derived from basal ice melt (frictional and geothermal heat) distributed evenly across the glacier bed. At temperate glaciers film drainage is likely to be important in isolated patches only, although it is central to Weertman's basic sliding theory (e.g. Weertman, 1979). Where limestone bedrock has been exposed by glacier retreat, as at Blackfoot Glacier, USA (Hallet, 1979b; Walder and Hallet, 1979) and Glacier de Tsanfleuron, Switzerland (Hallet *et al.*, 1978; Sharp *et al.*, 1990), calcite precipitates provide indirect evidence for flow of saturated meltwaters as part of a film. Although this film perhaps covers half the bed area (as at Glacier de Tsanfleuron: Sharp *et al.*, 1989a) the volume of water it carries, and its likely impact on sediment transport, is likely to be of little significance. The bulk of meltwaters will drain as part of different drainage structures. Ideas which support this include:

- Mathematical analysis (Walder, 1982) suggests that the water film is unstable if its thickness exceeds several mm. Irregular film flow creates zones of preferential viscous dissipation/ice melt at which incipient R-channels (see below) form. This implies that film/sheet drainage is not suited to the discharge of large volumes of water. Sheets of greater thickness are stable only if bed roughness and sliding speeds are unusually high.

- Water which enters the glacier by way of crevasses or moulins reaches the bed as channel flow. Collapse of such channels to feed film flow is unlikely because it implies water flow against the expected hydraulic gradient.

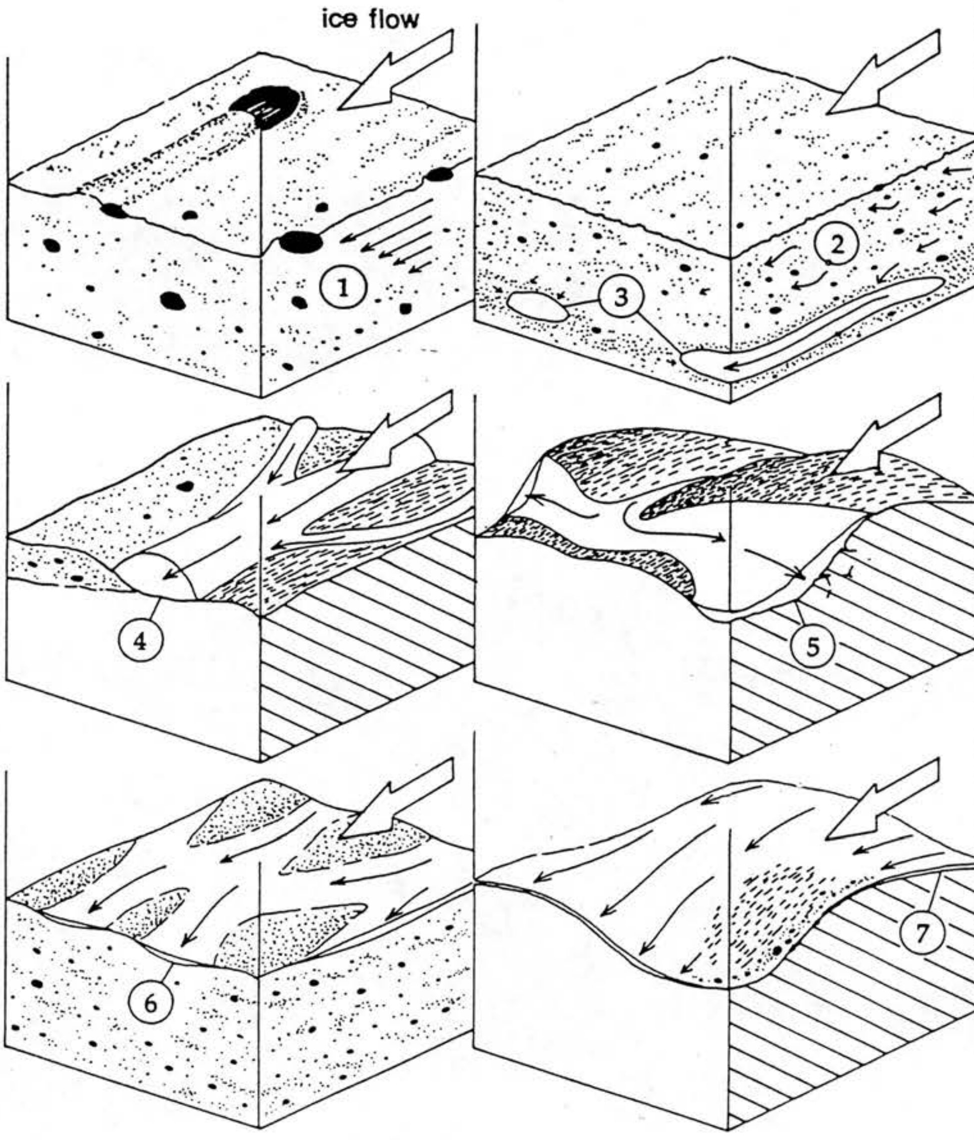

Figure 1.1

Possible styles of subglacial drainage. Graphic courtesy of D. I. Benn (forthcoming in Benn and Evans, 1998). KEY: 1) advection of pore-water; 2) Darcian flow; 3) pipe flow; 4) conduit flow; 5) linked-cavity flow; 6) canal flow; 7) film flow.

- The majority of temperate valley glaciers are drained at the snout by one or more major outlet streams.

- Pioneering studies of meltwater quality (e.g. Collins, 1979b) showed that the subglacial drainage system consists of two components, one fast and one slow.

Conduits, or R-channels

[Fig. 1.1, No. 4, hard bed, discrete drainage. E.g. Röthlisberger, 1972; Hooke, 1984]

The friction of running water generates heat, part of which is used to melt the ice of the channel's perimeter. The tendency to enlarge the channel this creates allows channel water pressures to exist at levels some way short of overburden equivalent: i.e. the local pressure drop which contributes the final term, D, to the Shreve equation. This type of channel was first analysed in detail by Röthlisberger (1972); hence **R**(öthlisberger)-channels. The ideal R-channel is thought of as a circular tube (hence the label 'conduit') cut into the ice, or, if subglacial, as a semi-circular tube, incised upwards into ice, and floored with rigid, impermeable bedrock. It is usual also to think of conduits as relatively large (scale dcm to m diameter?), so that flow is not subject to control by small bedrock roughness elements. In contrast to film flow, conduits are capable of carrying high discharges. Their efficient channel geometry - which follows from the assumptions of size and semi-circularity, both are which are more-or-less justified by observations of meltwater stream portals - allows rapid transit of meltwaters, but implies also that conduits cover only a small proportion of the glacier bed.

If steady-state conditions prevail, as Röthlisberger envisaged, the size (and so water pressure) of a conduit represents a balance between enlargement by wall melt, induced by water flow, and closure by plastic ice creep. Wall melt is a function of stream power (the product of discharge and hydraulic gradient); creep is driven by the difference between ice overburden pressure and conduit water pressure. If the conduit is assumed to be circular in cross-section, and if ice thickness, ice viscosity, bed slope, channel roughness and water discharge are known, it is possible to calculate the conduit size and water pressure for the steady-state condition in which wall melt is exactly balanced by creep closure. This problem was the subject of Röthlisberger's 1972 paper. The key finding of this analysis is that for flow in conduits water pressure falls as discharge rises: to be specific, water pressure varies as discharge$^{-1/4}$.

This inverse relationship between discharge and water pressures arises because of the (semi-) circular conduit geometry: the hydraulic radius (= channel cross-section area / wetted ice perimeter) is large. This means that any change in melt energy - from a change in discharge - is dissipated over a relatively small area, with a correspondingly large impact. Melt rate is linearly dependent on discharge, so it varies with the *square* of conduit radius (cross-section area = πr^2), whereas the creep closure rate is *linearly dependent* on conduit radius. The effect this has can be shown if we assume that discharge doubles, but (for simplicity) water flow speed stays the same. This means that channel cross-section area must double, as conduit radius rises by a factor of 1.41 (i.e. $\sqrt{2}$)[1]. Melt rates will double (roughly), but closure rates will increase by a factor of ~1.41 only. The new melt rate exceeds the new closure rate, so the channel must grow in size to restore the balance, with a commensurate fall in pressure gradient/water pressure (Hooke, 1989).

Channel stability and network structure. This inverse relationship between discharge and water pressure can be used to infer what a conduit network looks like and how it behaves. Adjacent conduits which carry different discharges are not stable with respect to each other. Water pressure in the larger conduit will be lower than that in the smaller conduit, so the pressure gradient tends to divert water from the smaller into the larger conduit. This serves to reinforce the pressure advantage and enhance the process of flow capture, because higher flow levels bring about a further fall in water pressure. As a result, a conduit network is expected to develop a dendritic form, dominated by a small number of 'trunk' channels fed by numerous small tributaries. Rising discharges downglacier tend to be routed across a progressively smaller fraction of the glacier bed (Shreve, 1972; Röthlisberger, 1972).

Open-channel flow. Röthlisberger's analysis assumes that melt and creep closure balance, so that pressurised flow takes place within a conduit completely full of water. However, this is not always the case, as Hooke (1984) shows. [Lliboutry (1983) reaches a similar conclusion, but his analysis is far less easy to follow.] If bed slope is steep relative to ice thickness - as is likely for many valley glaciers - wall melt can outstrip creep closure to such an extent that balance between discharge and water pressure becomes impossible. Excessive melt rates create free air space in conduits ('open' flow) with water at atmospheric pressure. If this happens ice exerts no influence on the direction of water flow (terms C and D of the Shreve equation disappear), so water flows directly down the bedrock slope. However, Hooke's analysis - as with Röthlisberger's - relies on several simplifying assumptions. Factors such as non-semi-circular channel geometries, enhanced conduit closure rates in zones of excess stress and/or softer ice,

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[1] This scenario implies no change in hydraulic radius, channel roughness or pressure gradient driving the flow, which is not realistic; however, the expected fall in the pressure gradient here (= flow speeds fall) will in part offset the increase in hydraulic radius and fall in channel roughness (= flow speeds rise).

or delayed transfer of heat from water to ice may mean (as, indeed, borehole measurements imply) that open-channel flow at the base of existing glaciers is less widespread than Hooke's (1984) analysis suggests (see Hooke *et al.*, 1990; also Box 7.1).

Linked-cavity system

[Fig. 1.1, No. 5, hard bed, distributed drainage. E.g. Walder, 1986; Kamb, 1987]

Ideas of flow within a conduit network correct several of the shortcomings of the water film model, but several other observations suggest that some other kind of hard-bed subglacial drainage structure also exists. These include:

- Borehole studies often indicate basal water pressures at levels too high to explain without severe modification of the conduit flow equations.
- Tracer studies indicate water flow speeds which seem rather slow for 'efficient' conduit drainage.
- Episodes of glacier uplift and enhanced sliding speed which correlate with a change in discharge and/or subglacial water pressures are difficult to explain using ideas of discrete conduit flow.
- Studies of bedrock exposed by glacier retreat confirm the existence of inter-connected cavities, formed in the lee of bedrock obstacles, which acted as an integral part of the subglacial drainage network.

These factors - particularly the evidence of exposed bedrock, which demonstrates that distributed water flow is not confined to a water film - strongly support the idea of some kind of linked-cavity system which occupies the 'middle ground' (often, it seems, literally) between film and conduit drainage. The geometry and flow behaviour of cavity networks differs from that of conduit networks because of the combined effect of bedrock topography and ice sliding, neither of which are important components of the ideal conduit model. Cavities form because of the redistribution of stress as ice presses against bed obstacles. This redistribution of stress creates perturbations in the distribution of equipotentials, and so the pattern of basal water flow, which is likely to be much more complex than the ice surface topography and overall bed slope (Shreve theory) suggest. Cavities are areas of relatively low pressure, so water tends to collect in, and flow between, cavities.

Theoretical treatments of cavities tend to take two forms. The first studies the conditions under which cavities develop, and uses water pressure as an independent variable; the second takes the existence of cavities for granted, and examines the conditions of water flow through cavities, in which case water pressure is a dependent variable. Ice sliding provides a link between these two complementary approaches: once formed, the pressure of water in cavities influences sliding speeds, but sliding speed also part-determines cavity size, connectedness and water pressure relationships.

The fundamental importance of the rock-step/cavity as a basic building block of the full subglacial process system makes it difficult to treat under separate headings. Here I digress briefly to examine the wider importance of cavities, and their formation, before returning to the specific hydrology of cavities.

Cavities: wider importance

The cavity - and the bedrock roughness feature in the lee of which it forms - forms a key component of the subglacial process regime. Cavities are important because:

- The redistribution of stresses in the immediate vicinity of the cavity/bedrock obstacle favours processes of subglacial rock fracture (Hallet, 1979a, 1981; Iverson, 1991b). This means the vicinity of the cavity is likely to be a zone of enhanced debris production.
- Cavities form an important element of subglacial drainage which covers large parts of the glacier bed (e.g. Walder and Hallet, 1979; Walder, 1986; Kamb, 1987). Water within cavities will have access to large quantities of debris, both loose, and enclosed in basal ice, which it may, or may not, be able to carry away. As the size and/or number of cavities expands (e.g. Sharp *et al.*'s 1989a study at Glacier de Tsanfleuron, Valais, Switzerland) access of subglacial waters to potentially vulnerable sediments stores will increase.
- Pressurised water which accumulates in cavities is thought to be an important influence on ice sliding speeds (see below).

Cavity formation

Cavities form in the lee of a bedrock obstacle if the local stress field prevents contact between ice and bedrock. The force which ice exerts on a rough bed is uneven: it is enhanced (normal pressure raised above the mean) as ice presses against the stoss-side of bedrock obstacles, and is reduced correspondingly (normal pressures less than the mean) in the lee of obstacles. The amplitude of this fluctuation rises with the magnitude of the basal shear stress, and the intensity of the stress concentration as determined by bedrock roughness (Weertman, 1979; Weertman and Birchfield, 1983a; Iken and Bindschadler, 1986).[2]

Cavities can form if the lee-side pressure drop exceeds the local normal force as determined by ice thickness and surface slope. This is likely to occur with steep, thin ice sliding quickly over a rough bed: a condition characteristic of the ice margin, where air-filled cavities are common. However, cavities are less likely to form like this below the thicker ice of the main body of the glacier. Here the tendency of cavities to close up under the weight of the overlying ice tends to exceed the tendency of ice to separate from bedrock associated with local pressure fluctuations. In this case, cavities can form only with the active intervention of high pressure meltwater. Water can push away the overlying ice to from a cavity if its pressure reaches a certain critical value, the *separation pressure* (Iken and Bindschadler, 1986), which is equal to the mean normal force minus the local pressure drop. Cavities form because the local force of water pushing upwards exceeds the local force of ice pushing downwards.[3] Cavities will grow as water accumulates and water pressures rise: this tends to involve uplift of the glacier, major redistribution of basal stress, major change to the behaviour of subglacial water flow, and episodes of enhanced ice motion (e.g. Röthlisberger and Iken, 1981; Iken *et al.*, 1983; Kamb *et al.*, 1985; Iken and Bindschadler, 1986). However, sustained episodes of major change - including potentially catastrophic sliding instabilities - tend not to occur because greater separation as water pressures rise tends to open up new escape routes which permit cavities to drain (e.g. Sharp *et al.*, 1989a).

Hydrology of cavities

This was first examined in detail by Walder (1986). Cavities form in the lee of bedrock obstacles, and are linked by a tortuous series of short, small channels, which can be cut in bedrock (Nye, or N-channels) or incised into ice (R-channels). This tortuosity arises because these inter-cavity links - or 'orifices' as Kamb (1987) calls them - tend to avoid the high pressure (i.e. overburden-plus) zones of obstacle stoss-sides. Here enhanced creep tends to close down small channels driven against bedrock obstacles, and water tends to be squeezed away from zones of excess pressure; only large channels, with high melt rates, are likely to survive in this kind of situation. Tortuosity of the links is reinforced by the preferred orientation of cavities perpendicular to ice flow. The structure of the ideal cavity network contrasts sharply with the structure of the ideal conduit network. The net effect is a dispersed, meandering system of

[2] The bed models of Weertman and Iken are identical, except that Weertman envisages a simple 'tombstone' bed of cubic obstacles, whereas Iken uses a sinusoidal bedrock profile.

[3] This is where the ideas of Iken and Weertman diverge: Weertman assumes that the role of water which accumulates in cavities is passive.

drainage, in which water flow is predominantly oblique to ice flow direction. The overall hydraulic gradient is gentle; overall flow speeds through these inefficient channels are slow; the water pressure tends to be high. Cavity networks are likely to be capable of carrying sizeable discharges only if the cavities are large and/or the number of channels (= cavities + orifices) is high; this implies water at relatively high pressure covers a sizeable fraction of the glacier's bed.

Cavities will be large if the rock-step which gives rise to them is high, if ice sliding is fast, and if the rate of roof closure (= ice creep rate - melt rate) is low (Walder, 1986, equations 6-8). This has two important consequences: 1) stable cavities cannot exist as effective pressure (= ice overburden - water pressure) tends to zero, and, 2) unless stream power is high (which by definition is unlikely for a linked-cavity system) sliding speeds will tend to exceed melt rates by one to four orders of magnitude (Walder, 1986, Table 1). Melt induced by running water is of negligible importance as a control on cavity size, which primarily represents a balance between creep closure and ice sliding enlargement. This means that there exists a crucial difference in the discharge-water pressure relationships of the two types of network: as discharge within a cavity network rises, so too does water pressure - the opposite of what is expected of a conduit network under steady-state conditions.

The impact of ice sliding is to create cavities which tend to be long and low in cross-section, with small hydraulic radius (i.e. a small area of flow per unit length of ice perimeter). Because of this predominant influence of sliding on cavity size and geometry, changes in roof melt rates as discharge fluctuates have little impact on cavity size and water pressures. If discharge doubles so too will the rate of ice melt, but this additional energy will be dispersed across such a large width of roof that it makes little impact relative to ice sliding, and so the melt/sliding versus creep closure balance which determines cavity size. The cavity grows, but by a factor less than two, so the water pressure rises (Hooke, 1989).

Channel stability and network structure. This positive relationship between discharge and water pressure means that unlike a conduit network, a linked-cavity network is not subject to collapse and rationalisation (unless the threshold of cavity stability is surpassed, in which case cavity networks can collapse to form conduit networks - see below). Large conduits tend to grow at the expense of smaller conduits by a process of positive feedback, but this is not the case with cavities. Large cavities which carry high flows will have higher water pressures than smaller cavities, so the smaller cavities tend to capture part of the larger cavities' flow. Water pressures and cavity size rise as flow through the once-small cavities increases; simultaneously cavity size and water pressures of the once-large cavities fall as through-flow falls. Thus cavity size and water pressures tend to equalise for all cavities. This allows the typical anastomosing network of cavities and links, with pressurised water distributed widely across the glacier bed, to remain as a stable feature of subglacial drainage.[4]

CHANGES TO THE STRUCTURE OF DRAINAGE

The presence of some threshold beyond which cavity networks are susceptible to collapse by a process of unstable orifice growth (Kamb, 1987; Willis, 1995, pp. 79-83), plus the instability associated with the inverse discharge-water pressure relationships of a conduit network create the potential for various time-dependent changes to the structure of subglacial drainage - changes which are likely to have a significant effect on rates and processes of ice flow, subglacial erosion and debris transport by both water and ice. Switches in the structure of subglacial drainage tend to be driven by fluctuations in discharge (e.g. the impact of a major storm, or seasonal variation in the volume of meltwater throughputs), although changes to glacier geometry can also be important (e.g. the build-up/purge of mass of a surge cycle, or ice mass growth/decay driven by climate change). Here I use the seasonal evolution of drainage thought typical of a valley glacier as an example of what seems to be a tendency to drainage reorganisation manifest at a range of time and space scales.

Autumn

Conduits tend to close up as meltwater production falls towards the end of summer. As discharge falls, water pressures within the smaller parts of the conduit network rise to meet the separation pressure (see above), at which point cavities open up immediately adjacent to the conduits (Sharp, 1988, p. 365). Water escapes into these cavities, and the conduits shrink, with a commensurate rise in water pressure. Loss of water to cavities further reduces the discharge through what is left of the conduit network, propagating/reinforcing the tendency for water pressures to rise to separation levels across the full system. Progressive diversion of flow into cavity storage (as well as falling system melt inputs) instigates collapse of the conduit network by a process of positive feedback. Throughout autumn, winter and spring drainage (at a much-reduced level) takes place mainly within a residual distributed cavity (plus film?) system. Pockets of water are likely to be trapped in cavities which are isolated from the wider network as discharge falls.

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[4] The exact limits of stability are set by ice and bedrock geometry, ice viscosity, and level of water discharge. Cavity/orifice geometry, channel plan-form geometry, channel roughness, hydraulic gradient (specifically local enhancement thereof) and ice sliding speeds contribute, but should reflect the influence of these first four factors (Kamb, 1987).

Spring

As surface melt rises, meltwater starts to accumulate at the glacier bed. The capacity of the linked-cavity network grows as water pressures rise: as the extent of ice-bed separation increases the size of cavities and links grows, isolated cavities join up to the wider network, the length of drainage pathways falls, and the pressure gradient rises. This process of network enlargement tends to be stable only up to a certain point: as discharge and water pressures continue to rise, the cavity network switches to a state of unstable growth. This is the product of two major factors:

1. High cavity water pressures oppose cavity closure by ice creep. If water pressures reach a sufficiently high level (effective pressures tend to zero) the net cavity closure rate becomes negative, so cavities will tend to expand continuously (Walder, 1986). This idea is closely related to that of the 'hydraulic jack' (Röthlisberger and Iken, 1981; Iken and Bindschadler, 1986).

2. As stream power rises, the rate of roof melt-back within orifices rises to levels at which it cannot be offset by creep closure and ice sliding. This instability occurs within the links, not the cavities because: a) the pressure gradient is much steeper in orifices than it is in cavities, so overall melt levels are higher; and, b) the hydraulic radius of the orifices is much larger than that of cavities, so roof melt is concentrated at levels at which it can have a major influence on channel size. Orifices subject to unstable growth are thought to evolve into fully-fledged conduits by a combined process of size increase, step-shedding and channel realignment parallel to the direction of ice flow (Kamb, 1987).

Unless offset by simultaneous episodes of faster sliding also triggered by the rise in discharge/water pressures (enhanced sliding tends to destroy incipient conduits: Walder, 1982, 1986), cavities and links subject to enlargement will tend to coalesce as nascent conduits introduce 'short-circuits' (Walder, 1986, p. 442) into the system. These conduits will exhibit the characteristic inverse relationship between discharge and water pressure, so instigating progressive growth of the conduit network at the expense of adjacent cavities: flow which converges on developing conduits will raise discharge and lower water pressures, boosting the pressure advantage of the conduit network, and enhancing the tendency to capture flow from what remains of the cavity network (i.e. the process of conduit collapse which occurs at the end of summer is reversed).

Work at the Haut Glacier d'Arolla, Valais, Switzerland shows that the key factor which controls this reorganisation of drainage is progressive retreat of the snow-line (Richards *et al.*, 1996, Fig. 8). As snow gives way to ice at the glacier surface the albedo is halved, and local meltwater inputs to the bed double (solar radiation dominates the energy balance of the Arolla Glacier). This abrupt meltwater-loading of the cavity system triggers a switch in the style of drainage as described above. Thus the conduit system expands upglacier at the expense of the distributed cavity system at a rate determined by the pace of seasonal snow-line retreat.

It is important to point out that real-world subglacial drainage structures are likely to be less clear-cut than the theoretical dichotomy of distributed versus discrete flow implies. The exact points at which a water film becomes a linked-cavity network, and at which a cavity network becomes a conduit network are far from clear. Existing glaciers are likely to show elements of all three structures simultaneously, either as part of a single network (e.g. Walder and Hallet, 1979; Sharp *et al.*, 1989a) and/or as parallel subglacial networks (see Chapter 6.4). Cavities and links originate because of sliding over a rough bed, and are likely to play some part in drainage as long as they do exist; indeed, the better-developed the conduit network, the greater is the need for some kind of distributed flow mechanism which drains areas of the bed not directly tapped by conduits. Conversely, the limited capacity of a stable cavity network implies that conduits must be a necessary and stable feature of subglacial drainage unless flow levels are extremely low (Walder, 1986, p. 443). Reorganisation of drainage is likely to be a partial process, both in space, and in time. However, it may be possible to speak of certain dominant patterns of subglacial flow, and in some cases these distinctions may be unusually clear-cut: e.g. the contrast in (overall) structure between drainage which seems to be dominated by flow in large conduits at Gornergletscher, and distributed flow (in smaller conduits and/or cavities) at its neighbour Findelengletscher (Collins, 1979b; Iken and Bindschadler, 1986), or the temporal switch between cavity and conduit drainage which controls the surge behaviour of Variegated Glacier (Kamb *et al.*, 1985; Kamb, 1987).

SUBGLACIAL WATER AND ICE SLIDING SPEEDS

It is now widely accepted that changes to subglacial hydrology influence the rate of basal sliding, although exactly what property it is which brings this about is not yet clear. The volume of water at the glacier bed, the pressure of the water, and the rate of increase of basal water storage have all been identified with faster sliding. For reviews see Weertman (1979), Paterson (1994, Ch. 7), and Willis (1995). Important field studies of water-sliding speed relationships include those conducted at Unteraargletscher, Berner Oberland, Switzerland (Iken *et al.*, 1983), Variegated Glacier, Alaska, USA (Kamb *et al.*, 1985; Kamb, 1987); Findelengletscher, Valais, Switzerland (Iken and Bindschadler, 1986), Columbia Glacier, Alaska, USA (Meier *et al.*, 1994; Kamb *et al.*, 1994), and Haut Glacier d'Arolla, Valais, Switzerland (Harbor *et al.*, 1997).

Water at the bed is thought to speed up sliding because (Iken and Bindschadler, 1986; Willis, 1995):

1. Water drowns out the smaller roughness features which make up the bed, so the frictional resistance to flow is reduced; the glacier 'sees' a bed of lesser roughness (i.e. the 'effective' bed) than that which exists in the absence of water.
2. If the water reduces or eliminates the ability of some parts of the bed to support a shear stress, bed roughness elements which remain effective must support a greater shear stress. This local intensification of shear stress at point of contact induces faster flow by enhanced enhanced plastic deformation.
3. Pressurised water acts as a 'hydraulic jack': as water pressures rise and water-filled cavities expand the ice is pushed upwards and forwards over the stoss-sides of bedrock obstacles located immediately downglacier of cavities.

These three factors imply that faster sliding in the presence of water must reflect the presence of some kind of distributed drainage network, which involves water under high pressure dispersed across wide areas of a glacier's bed. If the third factor is to be important, this must be a linked-cavity system in which transverse elements of drainage/subglacial ice surfaces on which the hydraulic jack mechanism can operate are maximised. Past work has tended to stress major changes in flow behaviour associated with widespread reorganisation of drainage: e.g. the switch from conduit to cavity drainage inferred for the 1982-1983 surge of Variegated Glacier (Kamb *et al.*, 1985). However, recent work has identified local contrasts in sliding attached to local contrasts in basal water flow: e.g. Harbor *et al.* (1997), working with borehole inclinometry at the Haut Glacier d'Arolla, identify sliding maxima tied to a narrow sub-marginal zone of variable (but not continuously high) water pressures, believed to indicate a subglacial conduit. The wider impact of such narrow zones of enhanced water and ice flow has yet to be established, but it is possible that large parts of the bed fall into their sphere of influence: e.g. stress transmission by transverse coupling changes the wider flow field, or travelling waves of water - and sediment - which originate in the conduit sweep wider areas of the glacier bed (see also Hubbard *et al.*, 1995; Chapter 2.2).

Space-time changes to subglacial water flow, and their impact on sliding speeds are important because:

- The rate at which subglacial debris is produced scales directly with sliding speeds. Faster sliding raises both clast contact force, and clast velocity: two of the four key components of Hallet's (1979a, 1981) abrasion model. Changes in the distribution of subglacial water pressures, stresses and sliding speeds are though to favour the processes of bedrock fracture and entrainment which constitute plucking (Röthlisberger and Iken, 1981; Iverson, 1991b).

- Sliding speeds control the rate at which debris is carried towards the ice edge.
- Sliding ice influences the stability of different types of subglacial drainage. Fast sliding suppresses conduit migration (Hooke, 1984) and collapse of the orifices which tie together a linked-cavity system (Kamb, 1987). This is important because big changes to the style of subglacial drainage are likely to instigate changes to the dominant subglacial sediment transport pathways.
- Spatial contrasts in water flow which change the ease with which sliding occurs (i.e. ice speed-ups and slow-downs) are likely to steepen longitudinal and/or transverse stress gradients within the basal ice. If basal ice thickens or thins its exposure to potentially destructive water flows will change (Chapters 6.3 and 7.2).

SOFT-BED HYDROLOGY

The possibility of water flow within a subglacial layer of soft, loose and more-or-less permeable sediments has been the subject of much work in recent years (e.g. Boulton and Hindmarsh, 1987; Alley, 1992; Walder and Fowler, 1994). The idea that soft-bed hydrology was likely to be important to many ice masses followed largely as a consequence of the emergence of what might be termed the 'deforming bed' paradigm (Boulton and Jones, 1979; Boulton, 1986; Alley *et al.*, 1986: Boulton and Hindmarsh, 1987): rapid deformation of subglacial till requires high pore-water pressures. However, the field of soft-bed hydrology also gathered impetus as a result of field studies which showed that, contrary to the 'hard-bed' assumptions of standard theories of hydrology, sliding and erosion, many valley glaciers rest (in part at least) on a bed of unconsolidated sediments (e.g. Engelhardt *et al.*, 1978: Hodge, 1979: Hooke *et al.*, 1997). It is evident that drainage of soft beds can occur in several ways.

Advection

[Fig. 1.1, No. 1, distributed drainage.]

Water which sits in the pore spaces of the till layer is carried forwards as the till deforms; the water itself does not flow relative to the till.

Darcian flow

[Fig. 1.1, No. 2, distributed drainage.]

Pore-water flow driven by contrasts in hydraulic potential. Discharge scales with the steepness of the hydraulic gradient, the hydraulic conductivity of the till, and the depth of the till layer. The capacity of Darcian flow is likely to be small, unless the till is unusually thick and sandy, and/or the hydraulic gradient unusually steep.

Pipe flow

[Fig. 1.1, No. 3, distributed drainage. E.g. Boulton and Hindmarsh, 1987.]

It is thought unlikely that Darcian flow on its own is capable of draining a subglacial till layer for realistic values of till conductivity and meltwater inputs. In this case, some kind of channel must form to increase the efficacy of discharge. If evacuation of water by pore flow falls short of that required for steady-state flow to prevail pore-water pressures will build up, creating the possibility of channel initiation by shear failure of the till (cf. Dietrich *et al.*, 1986: similar ideas of channel origin under subaerial conditions). Catastrophic piping failure induced by liquefaction of sediments (i.e. effective pore-water pressures fall to zero, or less) may occur. It is likely that pipe flow represents a temporary means of till drainage: pipes will tend to develop into some kind of stable channel (see below), or will collapse as pore-water pressures fall once excess water has drained away.

Sheet flow

[Strictly not shown on Fig. 1.1, but corresponds to No. 7 if bedrock is replaced by till. Distributed drainage. E.g. Walder and Fowler, 1994.]

Flow as a sheet of water sandwiched between the top of the till layer and the base of the ice is a plausible alternative to piping failure if the till layer is surcharged. This creates the possibility of (temporary, and spatially short-lived) flotation of the ice, as is inferred to happen at South Cascade Glacier (Fountain, 1992, 1994) and at Storglaciären (Iverson *et al.*, 1995). As with pipe flow, sheet flow is unlikely to be stable: depth irregularities induce channel formation by roof melt (see above).

Channel flow

Walder and Fowler (1994) provide a detailed theoretical treatment of the likely form and development of channel flow over soft beds. Till creep and/or ice creep will act to close channels down, so channels can exist as stable features only if flow levels are sufficiently high to offset creep by sediment removal and/or ice melt. Channel closure by inwards flow of ice is fastest if channel water pressures are low, whereas channel closure by inwards flow of till is fastest if channel water pressures (which set minimum values for the pore-water pressures of surrounding tills) are high. Thus Walder and Fowler propose that channels develop as one of two kinds: water flow 'attacks' either the ice above, or the till beneath, depending on which poses the greatest threat to the integrity of the channel. So:

- **Conduits.** If water pressures are typically low, a dendritic network of R-channels develops [Fig. 1.1, No. 4]. Conduits are incised upwards into ice, and sit on the surface of a stiff (i.e. weakly deforming, if at all) till layer. Conduits exhibit an inverse discharge-water pressure relationship, and behave in much the same way as hard-bed conduit networks (see above). This is the likely structure of drainage for soft-bedded valley glaciers, for which steep, thin ice tends to give rise to low basal water pressures.
- **Canals.** If water pressures are typically high, a 'canal' network develops [Fig. 1.1, No. 6]. This is made up of a distributed series of wide, shallow, braided channels, each of which is cut into (actively deforming) till, with a flat, stiff ice roof. The wide, shallow geometry expected of these canals (extending the examples of proglacial streams cut into coarse, loose sediments) produces behaviour similar to that of a hard-bed linked-cavity network: as discharge rises, channel water pressures rise also. However, Walder and Fowler arrive at this conclusion by way of algebraic methods, and fail to provide a satisfactory explanation for this in terms of physical processes. Canal geometry, unlike cavity geometry, is independent of sliding speeds, so the explanation must lie with the differential response of sediment transport and till creep as discharge rises. As flow rises, creep closure (which depends on water pressures and channel *width*: Walder and Fowler, 1994, p. 8) must outstrip sediment transport (a function of shear stress, which depends on channel *depth*) so that the increase in channel size which results is proportionately less than the increase in discharge; however, this is not clear from Walder and Fowler's paper. Canals are likely to exist beneath ice-sheets, for which thick ice and gentle surface slopes give high basal water pressures. Geological evidence (e.g. Clark and Walder, 1994) supports this conclusion.

THE RESEARCH FRONTIER

What we (think we) know about how water behaves at the glacier bed - the most important aspects of which I have just described - provides us with a powerful collection of ideas to guide our studies. However, the bulk of these ideas stem from ideal theoretical treatments which frequently rely on assumptions unlikely to hold in full for the real world. Future research is likely to try to overcome such restrictions, a process which will be helped by increasingly elaborate field techniques which permit detailed study of what in fact does go on at the ice-bed interface (e.g. Pohjola, 1993; Richards *et al.*, 1996; Stone and Clarke, 1996). Four aspects of hydrology likely to receive further attention, all of which impinge upon my arguments in this thesis, are:

1. Existing theoretical treatments of water flow tend to assume steady-state behaviour, whereas transient behaviour is likely to be important at the bed of real-world glaciers.

2. Theoretical treatments tend to work with singular subglacial drainage structures which (often implicitly) are assumed to operate in isolation, but the interaction of different types of drainage is likely to be important at the bed of real-world glaciers.

3. Theoretical treatments usually draw a distinction between hard-bed and soft-bed drainage types, whereas real glacier beds are likely to resemble a mixture of the two: i.e. the ice-bedrock interface is partly masked by discontinuous till patches of variable extent and thickness.

4. To reiterate a point I made in the introduction to this section: existing work tends to study hydrology for its own sake, or in terms of its links with ice flow. Water quality is studied because it provides clues as to what subglacial drainage looks like, and how it behaves. The effect drainage style has on catchment-wide sediment transfer, which determines what kind of depositional landforms will be built, is largely ignored. It is this aspect which I explore in this thesis.

1.2 GLACIER SEDIMENT TRANSPORT

Kirkbride (1995a) and Lawson (1993) provide comprehensive reviews. Here I sketch the core features and ideas which make up what can be called 'orthodox' sediment transport theory. Much existing work uses a basic distinction between two modes of transport (Boulton, 1978, Boulton and Eyles, 1979; Fig. 1.2):

- *Passive* transport of clasts within the *high-level* transport zone.
- *Active* transport of clasts within the *basal* transport zone.

HIGH-LEVEL TRANSPORT

High-level transport involves ice and debris which are not in contact with the glacier bed. Although basal ice can be carried into high-level transport in zones of compressive flow (e.g. at the confluence of two ice-streams, or towards the edges of the glacier) the vast bulk of ice which makes up this zone is 'englacial' - or 'meteoric' - ice, formed from snowfall, which has never been in contact with the glacier bed. Thus it is expected that debris associated with this englacial ice has never been in contact with the glacier bed (but see Chapter 5!). This lack of

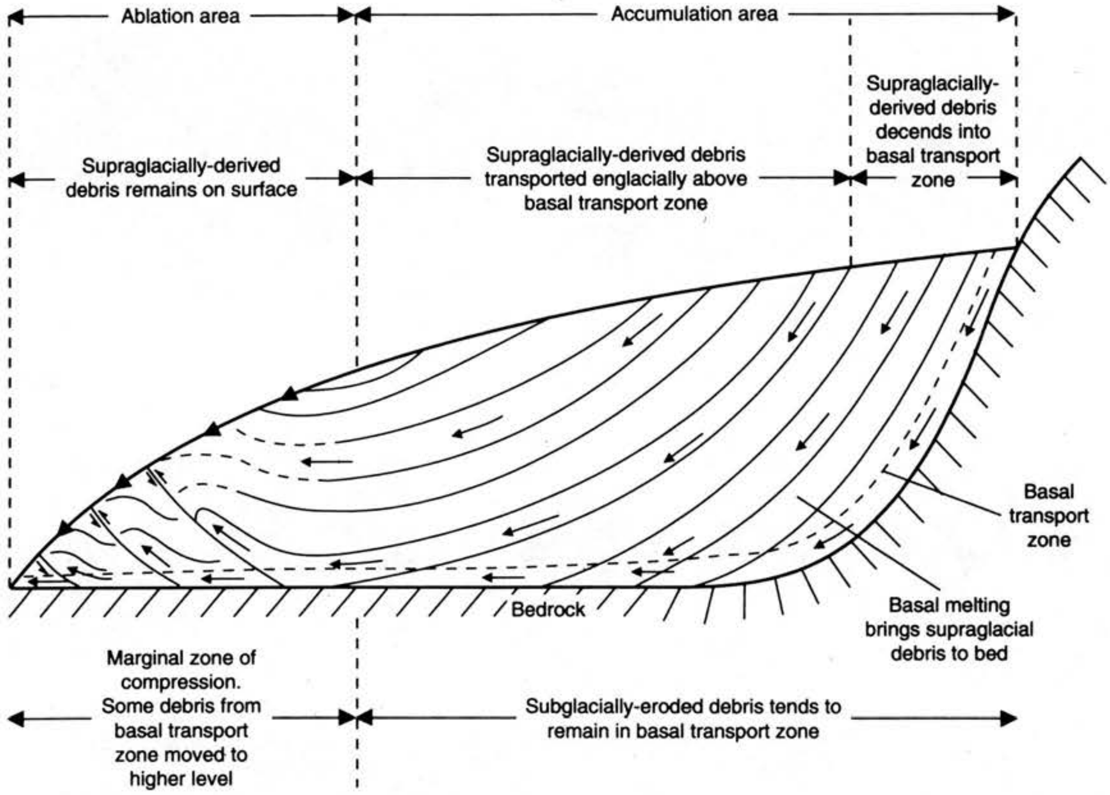

Figure 1.2

Sediment transport pathways through a valley glacier (from Boulton, 1978). See text for explanation.

contact with the bed defines passive transport. Forces which act on englacial ice, and its associated debris, are weak, so debris tends to retain its original properties (i.e. clast size, shape, angularity, texture; debris population particle size distribution).

The bulk of debris which enters directly into passive transport is derived from fracture of supraglacial rock-walls, and is delivered to the ice surface by processes of mass movement (usually rock-fall or slide) or by avalanches. Thus debris supply is controlled by a range of slope processes, although these may have some kind of indirect relationship with glacier behaviour: e.g. increased inputs of rock-fall debris may occur because ice retreat exposes a greater area of rock-wall to frost action. Debris which falls onto the ice surface of the ablation area will stay in surface transport; however, debris which falls onto a glacier's accumulation area will be buried by snowfall, and will eventually be incorporated as part of the main body of englacial ice. Such debris will eventually return to the ice surface (unless it enters the basal transport zone) with the upwards and outwards flow vectors characteristic of the ablation zone (Fig. 1.2). Debris in passive transport - whether englacial or supraglacial - can occur as individual clasts, or as distinct bands, depending on the magnitude and frequency of supply events. Compressive flow of the ablation zone encourages coalescence of debris also. The extensive spreads of supraglacial debris ('moraine') which smother the termini of many temperate alpine-type glaciers are usually the product largely of passive transport (e.g. Small, 1987a, 1987c).

BASAL TRANSPORT

Basal transport involves ice and debris immediately adjacent to the glacier bed. This ice and debris will make regular, but not necessarily continuous, contact with the bed. The ice of the basal transport zone is basal ice; this is ice which forms and reforms in contact with the glacier bed, usually under high-stress conditions. As a result, this ice tends to acquire properties (e.g. crystal size and shape, gas content, isotope composition) which render it distinct from englacial ice. It is common for basal ice to be debris-rich, but this is not a necessary property of basal ice. Some textbooks and papers incorrectly assume that basal ice is *always* debris-rich. The thickness of the basal ice layer rarely exceeds ~1.0 m (and is usually much less; indeed it is often absent altogether) because basal ice tends to be recycled: (re)formation of new ice by freezing at A is offset by melting at B. Thicker sequences of basal ice usually reflect thickening by compressive flow, or conditions of continuous freezing-on (common at the margins of polar glaciers, but sparse and short-lived at temperate glaciers). Basal ice which thickens sufficiently will increasingly escape the influence of the near-bed process regime, and so tends to enter into passive transport. However, because of the weakness of the forces acting on it (ablation of marginal exposures excluded!) such basal ice, and its constituent debris, retains its properties acquired in active transport. See Chapter 6.2 for a full review of basal ice.

Debris of the basal transport zone is produced by subglacial erosion: rock fracture and/or entrainment by ice of pre-existing debris. Pressures of space preclude a full discussion of subglacial erosion here, but I include further details as appropriate in Chapters 6 and 7. Debris of supraglacial origin can also enter the basal transport zone, by way of flow-lines or meltwater streams which intersect the glacier bed. Clast fracture occurs widely within the basal transport zone, induced by conditions of greatly enhanced stress (e.g. Hallet, 1979a, 1981; Iverson 1991a, 1991b). The progressive modification of clasts which results defines active transport. Typical changes include (Boulton, 1978; Haldorsen, 1981; Kirkbride, 1995a):

- 'Weak' elongate clasts break up into smaller, 'stronger' spherical clasts.
- Clast surfaces acquire scratches (i.e. striations).
- The angular *edges* of clasts are rounded (but rounding of clast faces usually requires water action).
- The proportion of fine-sized debris rises.
- Debris populations acquire a multi-modal, poorly-sorted particle size distribution.

Subglacial tills subject to deformation can also make up part of the basal transport zone. The flux of sediment achieved by such bed deformation can be considerable (e.g. Alley *et al.*, 1986). However, the extent and activity of mobile till layers beneath alpine-type glaciers has yet to be established with confidence, but they are likely to be of local importance at least (e.g. Benn, 1994). The distinction between basal ice and basal till is rather fuzzy in many cases: e.g. the 'active sub-sole drift' of Blue Glacier, Washington State, USA (Engelhardt *et al.*, 1978). Studies of subglacial tills tend to emphasise structures and fabrics rather than active modification of clast properties (see Benn and Evans, 1996, for a review), but forces mobilised at clast-to-clast and clast-bedrock contacts should induce some measure of crushing and fracture (e.g. Iverson, 1995; Cuffey and Alley, 1996; Iverson *et al.*, 1996). However, factors such as between-clast rearrangements (e.g. dilatation) and high pore-water pressures are likely to cushion inter-clast contacts, with a commensurate reduction in the intensity of clast modification.

EVALUATION

This distinction between high-level and basal debris transport by ice provides a simple, useful study framework. However, my view is that it is frequently given too much importance, with explanatory power mistakenly attached to what is essentially a descriptive statement of possible states of debris transport. It is easy to take the underlying processes - and process *interactions* - for granted, or to invoke the 'orthodox', ideal model to create misleading impressions as to what features are important at real glaciers. Marginal exposures of debris-rich basal ice, for example, are far less common than this model implies. A truly comprehensive model of sediment transport must take this into account. This requires a wider perspective in which greater weight is given to the full range of possible process mechanisms. Much of the knowledge necessary to do this exists, but it tends to be tucked away elsewhere (e.g. separate textbook chapters on processes of subglacial erosion or processes of basal ice formation, etc.; rock-fall processes viewed as a problem not of glacial geomorphology, but of periglacial/slope geomorphology). Perhaps the most serious deficiency is the tendency to separate transport by ice and transport by water. This makes it easy to overlook the question as to which of the two is more important, yet this is fundamental if we aspire to a satisfactory understanding of the variable character of ice-marginal sedimentation.

1.3 FLUVIO-GLACIAL GEOMORPHOLOGY

It is difficult - and perhaps pointless - to try to define what is, and what is not, fluvio-glacial geomorphology (1.4, below). Different textbooks treat the subject in different ways: recent treatments by Lawson (1993, 1995) and Bennett and Glasser (1996) opt for an integrated approach, whereas older texts (e.g. Price, 1972; Drewry, 1986) use separate chapters to cover the work of water. Sugden and John (1976) try to isolate a specific meltwater system. The label 'fluvio-glacial geomorphology' implies study of water action, and the related landforms, intimately linked with the presence of ice. However, exactly what qualifies is not clear: certain themes tend to be covered by 'mainstream' (sorry!) glaciology/glacial geomorphology (e.g. the controls subglacial drainage exerts on basal sliding and processes of subglacial erosion), whereas other themes tend to fall under 'mainstream' fluvial geomorphology (e.g. sediment transport and channel change of the proglacial zone). I find it useful to identify three broad schools of what passes as fluvio-glacial geomorphology:

1. **The subglacial school.** This has developed rapidly over the last 25 years, in step with our expanding knowledge of subglacial hydrology. Shreve's (1985) study of eskers is perhaps the outstanding example. However, work on subglacial drainage tends to emphasise the flux of water: studies of water quality which fit this framework are

designed to elucidate the style of subglacial water flow. Studies of sediment transport processes intended to improve our understanding of landform genesis are comparatively rare.

2. **The ice-marginal school.** This tends to use field data rather than the theoretical ideas central to the work of the subglacial school. The heritage of the ice-marginal school dates back to the nineteenth century, and its work falls firmly within the tradition of glacial geology (1.4, below). It centres on the description and interpretation of (mostly) ice-marginal landforms such as kames and meltwater channels, usually with a view to reconstruction of the limits and chronologies of past ice masses. Price (1973, Ch. V and Ch. VI) gives a summary of this kind of work.

3. **The proglacial school.** This concentrates on proglacial features, such as braiding and channel bar development. Two distinct sub-schools exist: a) the stratigraphic school, which studies vertical and lateral (proximal/distal) sediment sequences indicative of changing water flow regimes (e.g. Hambrey, 1994, pp. 163-172; Maizels, 1995); and, b) the process-form interaction school of the 'new fluvial geomorphology' which studies in detail the real-time dynamics of sediment transport and channel change (e.g. Lane, 1995). Sugden and John (1976) do not consider the work of the proglacial school to be part of fluvio-glacial geomorphology because it studies features which are not developed directly in contact with ice, created by processes which are not exclusive to glacial environments. This is true, but the presence of huge quantities of water-worked debris in front of many glaciers must tell us something about the behaviour of subglacial water and sediment, even if the exact details of the fluvial features are independent of ice action.

THE IMPORTANCE OF WATER ACTION

Data which suggest that typical rates of denudation of ice-covered catchments are some 10 to 1,000 times those achieved by 'standard' subaerial/fluvial processes (e.g. Hallet *et al.*, 1996) can give the impression that fluvio-glacial processes must be of minor importance relative to 'true' glacial processes (if, indeed, such things exist). Orthodox glacier sediment transport theory (see above) tends to reinforce this impression. Drewry (1986, p. 90) considers processes of plucking and abrasion to operate faster than do mechanical and chemical meltwater processes of *primary* erosion, a view which is widely shared,[5] but difficult to justify at the level of process mechanics. However, it is crucial to distinguish between production of debris by bedrock fracture and subsequent transport of debris. Subglacial water transport is likely to be unimportant in cold, fairly arid areas such as Greenland or Svalbard: the quantity of debris

[5] But see, for example, the comments of R. A. Vivian and B. Hallet in *Journal of Glaciology*, 23, 89, p. 390.

carried by water here is perhaps one-tenth that carried by ice (Hallet *et al*, 1996). As rates of precipitation and melting increase the significance of subglacial water transport rises rapidly. At many temperate glaciers (such as in Iceland) meltwater removes 10-20 times the quantity of debris discharged directly by ice. Thus the statement by Boulton (1978; Fig. 1; my Fig. 1.2) that "subglacially eroded debris tends to remain in [the] basal transport zone" will be misleading in many instances. It is the volume of debris in transport, and the *partition* of that debris between alternative transport pathways which determines the pattern of ice-marginal sedimentation (see Chapter 2.5 for a full discussion of subglacial flushing). This highlights the importance of a sediment budget perspective: i.e. the how much? where? and why? of debris viewed at the scale of the catchment sediment cascade. This is the basic theme of my thesis. This perspective also illustrates the weakness of the three-fold division of fluvio-glacial geomorphology I suggest above: the three schools are not separate, but are closely linked. The behaviour of subglacial water exerts a major influence on how much debris is produced, both by 'glacial' and 'fluvio-glacial' processes, and what happens to it thereafter: pebbles which make up part of mid-channel bar in a glacier's outlet stream (i.e. subject matter of the proglacial school) cannot simultaneously make up part of a lateral kame terrace (i.e. ice-marginal school). It is necessary to think in terms of links, not divisions, and to recognise that the suite of ice-marginal landforms encountered at any glacier arises because of some particular allocation of debris between alternative transport pathways as determined by the overall process regime extending upglacier (see 1.4, below, and Chapter 9).

1.4 A THOROUGHLY GLACIOLOGICAL GLACIAL GEOMORPHOLOGY? (a.k.a. a selective history of under-achievement)

I assume that the aim of glacial geomorphology - as with any type of geomorphology - must be to put together a meaningful account of the evolution of landscape which draws on a detailed understanding of the mechanisms responsible. My view is identical to Harbor's (1993, p. 129):

> "The primary goal of glacial geomorphology is to provide physically-based explanations of the past, present and future impacts of glaciers and ice sheets on landform and landscape development. To achieve this requires the integration of studies of landform with studies of the processes responsible for form development (over a wide range of spatial and temporal scales)."

It is not sufficient just to consider *processes*, however impressive the maths and physics used to describe these processes might be. Nor is it sufficient just to describe, and perhaps date, different *forms*. The challenge which faces the community of geomorphological researchers is how to achieve process-form *linkage*. If geomorphology is to say anything useful, this is the crucial nut we have to crack (Sugden *et al.*, 1997). Few dispute this, yet, despite remarkable progress in fields such as remote sensing, dating techniques, rock fracture mechanics, and both theoretical and empirical studies of the behaviour of water beneath glaciers, it seems that much still has to be done before the desired synthesis of process and form is achieved. If we accept that glacial geomorphology must try to explain how and why glaciers create such a diverse range of landforms, it appears we must admit that past progress has been limited.

Writing in 1972, Andrews was the first to express this sense of under-achievement:

> "In the last two decades the science of Glaciology has developed enormously in its sophistication, in its world-wide coverage of phenomena and in its potential for answering questions that have taxed glacial geologists for one hundred years or so. One of the most disappointing features of this period has, however, been the lack of the application of glacial theory to the problems confronting the glacial geologist and geographer." (Andrews, 1972, p. 2)

Andrews identifies the mismatch between theoretical and empirical/historical types of enquiries, which corresponds broadly to the dichotomy of process studies versus form studies. Sugden (1977, 1978), writing in the first two issues of *Progress in Physical Geography*, picks up on this challenge of process-form linkage, and uses it as a major theme of his reports. He writes enthusiastically of the "widening horizons opened up by the growing dialogue between glaciologists on the one hand, and geomorphologists, geologists and engineers on the other hand" (Sugden, 1977, p. 308). This implicit mixture of criticism and optimism - which clearly echoes Andrews - is also found on the first page of *Glaciers and Landscape*:

> "Insufficient modern glaciological theory has been applied to the study of glacial geomorphology and glaciologists have not been provided with reliable information on landforms. Perhaps there is a need for a more glaciological type of geomorphology and a more geomorphological type of glaciology. There is now a strong case for a realistic dialogue between those studying glacier dynamics, and those studying forms. Until this occurs, there can be few spectacular advances such as those achieved recently in fluvial and slope geomorphology." (Sugden and John, 1976, p. 1)

Although he was impressed by its scope as an introductory text, Drewry (1977) was not over-complimentary in his review of *Glaciers and Landscape*:

> "A principal aim of *Glaciers and Landscape* is to provide a link between glaciology and the study of glacial landforms - a task well overdue, but, unfortunately, not successfully accomplished. Failure stems from a deliberate attempt to treat complicated physical problems in only a highly qualitative manner, and to use 'fashionable' geographical concepts such as 'systems theory' which appear artificial and contribute little to a deeper understanding of glacial problems. The book is, as a consequence, seriously deficient in elementary physics and mathematics essential to the understanding and manipulation of glaciological concepts". (Drewry, 1977, p. 163)

His conclusion was that "[a] gap remains for a thoroughly glaciological glacial geomorphology". Drewry's contempt for the traditional geographer's cavalier approach to the 'rigorous truths' of basic physics is ill-concealed. Paterson (1981) presents similar views: "[i]n the author's opinion, a mere handful of mathematical physicists, who may seldom set foot on a glacier, have contributed far more to the understanding of the subject than have a hundred measurers of ablation stakes or recorders of advances and retreats of glacier termini" (p. 3). This remark survives in the third edition of *The Physics of Glaciers* (Paterson, 1994, p. 6).

The comments of Paterson and Drewry smack of scientism: the doctrine that the rigorous, mathematical, reductionist treatments of physics provide a superior means of study. Alternative methods of enquiry are trivial: to quote Ernest Rutherford (1871-1937), "all science is either physics or stamp collecting"! (widely quoted: e.g. Partington, 1992). However, the books by Paterson (1981, 1994) and Drewry (1986) fail to merit the accolade of a "thoroughly glaciological geomorphology". Paterson has the excuse that his book was not written as a textbook of geomorphology (he recommends Drewry's book!). Drewry's book - *Glacial Geologic Processes* - is just as much of a partial perspective as is *Glaciers and Landscape.* If Sugden and John omit too much physics to satisfy the reductionist school, Drewry makes too much of a correction to satisfy those seeking a synthesis of process and form (*Glaciers and Landscape* is the better book by far in this respect). The novelty and scope of Drewry's book - notably his use of the engineering literature to explore the details of rock fracture and sediment transport by ice and water - is impressive, but his recourse to mathematics frequently serves only to conceal the true meaning and impact of the processes discussed. The link between the process equations and the time-space development of landforms receives little attention (see Clapperton, 1987, and Dowdeswell, 1987 for reviews).

W(H)ITHER GLACIAL GEOMORPHOLOGY?

The three books by Sugden and John, Paterson, and Drewry simultaneously demonstrate the potential and the problems of a glacial geomorphology which seeks successfully to link process and form. Much of the discrepancy and discontent can be ascribed to issues of scale: as Schumm and Lichty (1965) show, geomorphic explanation is not independent of scale (see also Chapter 9). For example, Sugden's treatment of thermal regime and its influence on ice dynamics and the impact this has on landscape (Sugden and John, 1976, Ch. 10 for a summary) might fail to satisfy Drewry's image of a rigorous, mathematical science, but it does go a long way towards explaining the major features of areas such as north-east Scotland (see Hall and Sugden, 1987, and Glasser, 1995, for recent extensions of Sugden's original ideas). However, this scale problem is symptomatic of a fundamental conceptual gulf.

The picture of glacial geomorphology which emerges is of a discipline which sits awkwardly between two distinct traditions. These two traditions incorporate very different ideas and practices, so, as it drawn towards one extreme or the other, glacial geomorphology struggles to fix its identity. I call these two traditions 'glaciology' and 'glacial geology', labels taken from the historical review papers published in the special edition of the *Journal of Glaciology* published to mark the fiftieth anniversary of the International Glaciological Society (Clarke, 1987; Boulton, 1987a). These two categories are, inevitably, caricatures - but caricatures which contain sufficient truth to make a useful point. Similar dichotomies are commonly set up elsewhere: e.g. the distinction between 'timeless' and timebound' studies (e.g. Strahler, 1952) or between 'functional' and 'historical' studies (e.g. Church, 1996; Sugden *et al.*, 1997) (but see Chorley, 1978, for a detailed discussion of different types of geomorphic 'explanation'). Glaciology tends to be timeless/functional, glacial geology timebound/historical.

Glaciology: process without form

'Glaciology' incorporates process studies which fail to develop in full the implications these have for development of landforms. The majority of glaciological papers contain no reference to landforms at all. The diverse range of techniques - mathematical theories, computer simulations and laboratory experiments - which fall into this category are united by their abstract character. Glaciology ignores, or simplifies, the complexity of the real world so as to concentrate on the basic 'why' questions. Its approach is purposefully reductionist: it seeks to identify and analyse the basic process building blocks, but it does not proceed to consider how these process building blocks are put together to create real-world glacial landforms. This is its great weakness. It explores the logic of the laws of physics as we understand them, acting on specific materials, in this case ice, water and rocks, to give rise to single events. Much of glaciology deals with four basic phenomena: the failure/flow of 1) ice, and, 2) water; 3) the failure of intact rock; and, 4) the balance of forces which governs sediment transport by ice and water. It tends to assume that explanation is achieved by simple extrapolation of these basic process laws: e.g. a computer ice-sheet model which runs using Glen's flow law to describe ice mass transfer in response to imposed conditions of mass balance and basal drag (see Kerr, 1997, for a recent critique of computer models). It fails to explore the consequences of the repetitive interaction of the laws of physics: i.e. the *configurational* causal web of relationships dispersed in time and space. The reductionist trick of artificial system closure precludes satisfactory geomorphic explanation.

Glacial geology: form without process

I include in this category studies which describe surface features created by the past or present action of glaciers which tend to take the 'why' (i.e. the underlying process mechanisms) for granted. This category includes 'traditional' glacial geology (description and dating of landform sequences), plus studies of sediments and landforms/landscapes (this type of study has received a major boost in the last two decades thanks to advances in remote sensing techniques). The tradition which defines this category is that of empirical investigation, often taken to extremes in which data are given intrinsic value, rather than seen as a means to an end. This is the empiricist fallacy of positivism: the facts are believed to speak for themselves. Different approaches tend to take refuge in some form of quantification: e.g. dates assigned to different moraines, parabolic equations which describe the cross-section of a glacial trough, statistical techniques of classifying sediments. However, such an illusion of 'science' does not guarantee satisfactory explanation at the level of process.

Consequences

Glaciology and glacial geomorphology are both powerful, enduring traditions of science. It is important to stress that there is nothing necessarily wrong with these traditions; it is just that they provide an insufficient basis for glacial geomorphology which tries to explain form in terms of process. This demands elucidation of process-form linkages, a task which is beyond the scope of these traditions. If anything, these two traditions have hardened in the past two decades (in contrast to the sustained attack on reductionist and empiricist 'meta-narratives' elsewhere, notably in the social sciences, including human geography). This has inevitably damaged glacial geomorphology - the two traditions it must unite increasingly diverge - which today comes across as a deeply unfashionable discipline. Papers on glacial landforms, common in the 1960s, 1970s and 1980s, have now all but disappeared from the *Journal of Glaciology*. Annual progress reports on 'glacial geomorphology' in *Progress in Physical Geography* ended in 1985. These returned in 1992 under the significantly less distinct title of 'Progress report: Glaciers'.

It seems that the hopes voiced by Andrews and Sugden in the 1970s have yet to be met. This is not to say that no progress has been made towards a glacial geomorphology which explicitly links process (glaciology) and form (glacial geology). Excellent papers which achieve exactly this include work on striae and polished/fractured bedrock (Iverson, 1991a; Sharp *et al.*, 1989b), cirques and overdeepened basins (Hooke, 1991), drumlins (e.g. Boulton, 1987b) and eskers (Shreve, 1985). Nevertheless, the call for improved integration of process and form studies has recently resurfaced: e.g. by Harbor (1993), Knight (1992, 1995a), and, in the latest major text to be published, by Menzies (1995a):

> "[I]t is essential to comprehend the fundamentals of ice physics and dynamics coupled with an appreciation of process geomorphology and sedimentology. This conceptualisation of glacial environments within Earth Science might be termed a *glacio-sedimentological paradigm* as distinct from the past *morpho-sedimentological* methodological approach where descriptive terminology and scant regard for glaciology and glacial sedimentology merged to render narrow or self-limiting explanations...
>
> ...Perhaps central to this text and its underlying thesis is a phrase I came across many years ago, in which Yatsu incisively pointed out that 'geomorphologists had asked the questions where, when and what, but had rarely asked the questions how and why'."
>
> Preface to *Modern Glacial Environments*, pp. xi-xii. (Menzies, 1995a)

REALIST GLACIAL GEOMORPHOLOGY

The big question is: how exactly are we to bridge the gap which exists between the levels of process and form, and so successfully put glacial geomorphology back together? This is something which I return to in the final chapter. Here I wish to introduce briefly the ideas which I have used to guide my work. It is my view that the desired synthesis of process and form is not something which will occur spontaneously if we continue to persevere with the two traditions sketched above; if we are to make a determined effort to link process and form we must re-think the conceptual frameworks we use to guide our studies. In particular, we must think more carefully about exactly what are the specific conditions which enable different processes to combine and so generate specific events - in this case, construct certain types of glacial landforms. I find three inter-related study goals useful as part of this process. These are:

1. We must make a determined effort to bridge the gap which tends to divide studies of abstract theoretical structures from empirical studies of surface form. This requires that we make explicit statements about what process(es) link(s) to which form(s).

2. We must try to achieve detailed knowledge of how and why specific processes work. This is sometimes known as the recovery of 'ontological depth'.

3. We must be careful to bring our ideas together in the appropriate *context*: it is necessary to think more clearly about how things (processes and forms) relate to each other - and so set up causal webs - in time and space. Past events part-determine what happens to a system in the future; what happens elsewhere part-determines what happens at, say, place X; what happens here in turn gives rise to further changes somewhere else again.

These ideas are drawn from the perspective of the realist tradition of science, although similar ideas permeate much of post-positivist/post-structuralist work in the wider social and environmental sciences. (My introductory sources: Johnston, 1986; Gregory, 1986; Sayer, 1985; Richards, 1990. Sayer, 1992, provides a detailed survey of realist ideas, which, although written as a guide to social science research, contains much which relates directly to research in the natural and Earth sciences.)

The strength of realism is its ontology: i.e. its ideas of what exists, or what the world is like. Realism assumes that the world consists of a number of inter-related levels. What we see (form, or the empirical level of observation) occupies a different level to what makes things happen (process, or the abstract level of reductionist theorising). The mistake of empiricism/positivism is that it assumes that surface appearance (particularly regularities of surface appearance) automatically reveals cause; the mistake of reductionism is that it assumes that things happen, and so give rise to surface appearance, because of the straightforward extrapolation of basic processes. Realism assumes that the relationship between cause and effect is *not* self-evident, and tries to correct the error that empiricism and reductionism make (in different ways) by setting up as its central problem study of the various filters which intervene and mediate the relationship between cause and effect. This is what I mean by the bridge between theoretical structures and tangible surface forms.

The goal of realist research is to investigate the package of mechanisms and contingent conditions which together constitute a causal structure. This is achieved by critical thought; the process is primarily interpretative, rather than rigorously explanatory or predictive in a deterministic sense. The reasoning process simultaneously involves thinking more deeply (goal 2, above) and thinking more widely (goal 3). This is contradictory to a certain extent because it involves both abstraction (i.e. picking out individual processes to focus on their fundamental workings), and thinking in terms of wider perspectives at the same time; however, the difficulty here is reduced if we accept that different causal structures occupy different 'depths' of reality (see below).

The 'thinking more deeply' part of this process is reductionist, or what is termed the recovery of ontological depth. This means we must try to identify in detail exactly what are the processes which generate surface forms: e.g. we might identify a certain type of moraine which accumulates as debris melts out of ice at the edge of a glacier (ice which we would probably recognise as basal ice). However, it is not much of an explanation to say that moraines form because debris is released from ice at the edge of the glacier. We must think more deeply to get at the 'real' cause(s); thus here we need to know the details of debris transport by ice, and the details of rock fracture which give rise to the debris. This requires judicious reasoning: processes of rock fracture and transport are not immediately apparent in the pile of debris at the edge of the ice. Fortunately, much reductionist-type research by the glaciological school in the last twenty years has provided lots of clues as to what these individual process building blocks are, and how they might work.

However, it is clear that the reason a moraine (or, indeed, the majority of things in the real world) exists cannot be reduced to a single process. Rock fracture cannot explain moraines, but some combination of events which together constitute a complex causal network dispersed in time and space can. This is the task of adding *context* - or 'thinking more widely' - to give a full interpretation. It is just as important to understand how and why things come together as it is to understand basic processes. The idea here is that objects have certain causal powers or liabilities which are activated given suitable contingent (i.e. non-essential) conditions: e.g. a body of ice with a non-zero surface slope must impose stress (normal and shear forces) on its bed. This means that the ice has the potential power to erode its bed, but this power is realised only if certain other (contingent) conditions are met (the 'classic' example is gunpowder, which has the power to explode, but only if confined in the presence of a match). In this case, the equivalent of the match might be suitably weak bedrock (weak because the rock is made up of weak materials/bonds, or because it has a history of fatigue wear) and/or a suitable switch in the state of subglacial drainage (see Iverson, 1991b). Events are determined not by a single process alone, but by a process (or processes) which acts in a specific context. Change this context, and the event changes (see Fig. 1.3). Causal structures become yet more complex once we accept that things in the real world are rarely the product of a single event. Moraines are built up because of multiple episodes of rock fracture and debris transport which act together within a specific context which must stay favourable to the 'goal' of moraine formation over wide areas and long times. This is the idea of a complex causal chain dispersed in time and space. It is the antithesis of the classic 'closed system' laboratory experiment which consists of a simple set of objects carefully controlled to generate a single pre-determined event. Realist analysis works with the idea of choice which laboratory experiments are designed to eliminate. Ice, water and rock carry a certain range of causal powers and liabilities (e.g. the power of ice to fracture rock, the power of water to entrain loose debris, the power of rock to resist erosion), but the interaction of these materials gives rise to a number of different forms because different combinations of events within different contexts activate different potentials. Thus causality rests with the way in which things *interact* as part of specific process regimes/chains of events. This thesis explores the way in which meltwater provides different contexts which give rise to different patterns of ice-marginal sedimentation.

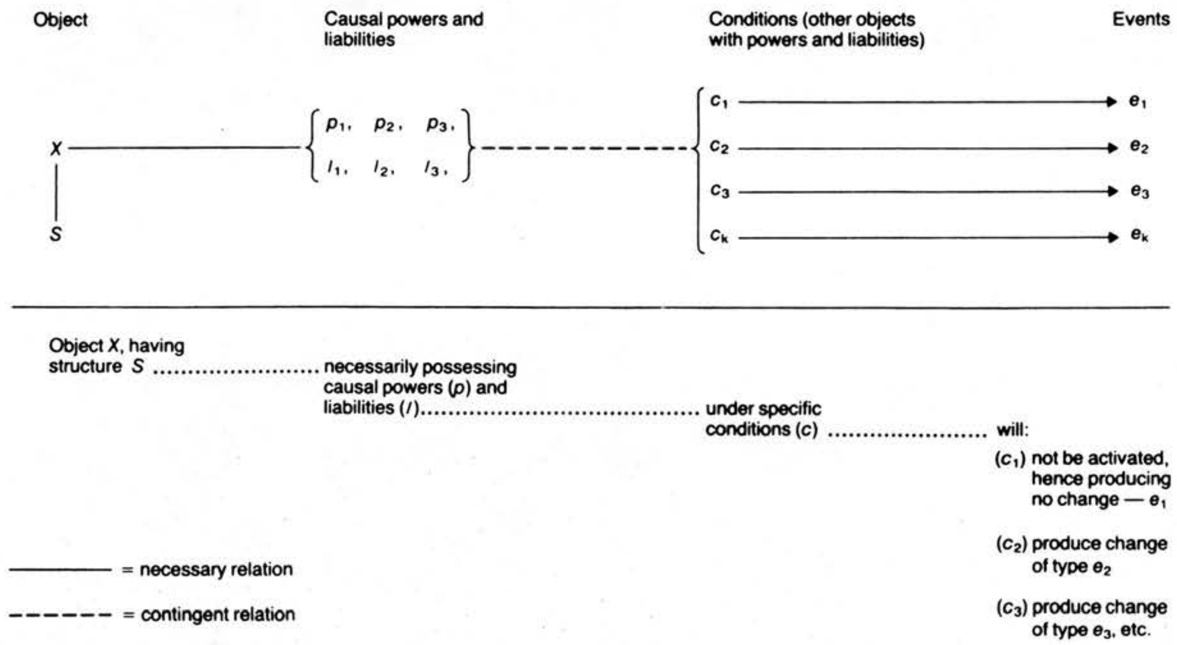

Figure 1.3

The structures of causal explanation (from Sayer, 1992, his Fig. 7). The important point here is that relationships of cause and effect are not simple and direct: a) individual processes and real-world outcomes (events) occupy different levels of reality; so, b) causality cannot be ascribed to any single process; because, c) it is *contingent* time-space *combinations* of processes (i.e. Sayer's specific conditions, c_1, c_2...) which actually make things happen.

The idea that reality is made up of a stack of different levels - or strata - is important. Reductionist analysis does not work because of this (see also Chapter 9). We cannot understand how moraines form just by breaking down the complex reality of the catchment sediment transfer system into its individual parts (i.e. separate laws of ice flow, rock fracture, debris entrainment, etc.). The power to form moraines exists at a higher, *emergent* level. Different contexts cause different causal structures to emerge, which appear as different emergent styles of ice-marginal sedimentation. This interpretation of events as part of the 'big picture' involves what Sayer (1992, p. 119) describes as a "macro-regress", as opposed to the "micro-regress" (interpretation of events in terms of the fine detail of *individual* processes) of reductionist analyses. The 'big picture' incorporates the ideas of contingent interaction, context, stratification and emergence; reductionist analysis tries to eliminate these factors, and so *undermines* the quality of explanation.

This idea of the 'big picture' requires us to make two important choices:

- How do we define the system of interest?
- What is the appropriate level of analysis/explanation to work with?

System boundaries? System closure which attempts to isolate specific strata can work to oppose satisfactory explanation, because it excludes emergent causal linkages which connect that stratum to other strata. The common practice by which geomorphology tends to be carved up into specific sub-disciplines perhaps represents such an example of erroneous system closure. Such arbitrary divisions of reality are "chaotic conceptions" (Sayer, 1992, p. 138). Fluvio-glacial geomorphology, which supposedly identifies the range of elements which relate to water action within a glaciated catchment, possibly qualifies as such a chaotic conception. The boundary between 'pure' glacial geomorphology and fluvio-glacial geomorphology is far from clear. Features which have much in common are separated, but features which have little in common are grouped together: e.g. lateral moraines (*sensu stricto*) and kame terraces are separated by classification which stresses the character of final sediment deposition (i.e. out of ice, or in water), although the constituent debris often shares a common history of erosion and transport upglacier. The similarity (of history and causal structure) between moraines and kames (traditionally a 'glacial' and a 'fluvio-glacial' feature respectively) is likely to be greater than the similarity between kames and outwash (both 'fluvio-glacial' features); see also Chapter 5. The distinction between glacial geomorphology and fluvio-glacial geomorphology creates artificial discontinuities which can mask the catchment-scale continuity essential to understanding of causal networks of sediment transport.

Level of analysis? At times it is perfectly permissible to take certain strata for granted (Sayer, 1992, p. 120). This is necessary to avoid the 'infinite regress' implicit in the reductionist method. If the rhetoric of reductionist method was to be taken to its logical conclusion, everything would have to be explained in terms of quantum mechanics! Science must make abstractions at the appropriate level if it is to work. Thus Glen's Law appears as a fundamental law of ice physics from one point of view, but from another it appears as a "convenient fiction" (Cohen and Stewart, 1994, p. 408) which provides an empirical generalisation of complex ice deformation, which works well at a suitably large scale of time and space (see Paterson, 1994, Ch. 5): i.e. bulk ice flow as described by Glen's Law represents the emergent product of the behaviour of a large number of individual ice crystals. To use Glen's Law to model some aspect of ice behaviour (e.g. closure of conduits by plastic ice flow: Röthlisberger, 1972) is just as much a 'fudge' as is the (supposedly non-rigorous) appeal to thermal regime as a basic element of causal networks which Drewry (1977) criticises as a major weakness of *Glaciers and Landscape*.

The level of entry chosen must reflect the scale of the problem: as a rule of thumb, the bigger the physical scale of the feature, and the longer its history, the greater the number of interactions which must be incorporated in some way, and the greater is the likelihood that generalisation and/or qualitative reasoning must be used. In a realist sense, explanatory rigour is judged not by the quality and quantity of the maths used, but by the adequacy of the casual framework adopted. Thermal regime represents the emergent product of the interaction of climate, ice geometry and ice dynamics. 'Warm' and 'cold' thermal regimes are not just descriptive terms; they are specific states of reality which exist at a relatively high level of abstraction, and carry genuine explanatory value because they embody genuine causal powers. The idea of thermal regime provides us with a powerful framework for causal analysis precisely because i) we can recognise the stratified nature of reality; and, ii) because studies which try to recover ontological depth (i.e. explore processes which inhabit relatively deep levels) have given us a good idea of what is entailed if a glacier is or is not frozen to its bed (e.g. Boulton, 1972a). In fact, it is likely that we have a better idea as to what is entailed at deeper levels of process for the case of warm or cold thermal regime than we do for the case of complex ice creep which Glen's Law encapsulates.

Ideas of realism applied to moraines

I have used the ideas set out in the preceding pages to guide my studies of ice-marginal sedimentation. Realist analysis is a balance between discarding too much of process, and trying to include too much of it. The trick is to define the relevant conditions which determine the events of interest - in this case, patterns of moraine formation - without falling foul of a chaotic conception. Sólheimajökull (Chapter 2) and Gígjökull (Chapters 4, 5, 6 and 7) display very different styles of ice-marginal sedimentation. I infer with confidence that both slide quickly over weak bedrock. This means it is unlikely that the contrast in moraines is explained at the (relatively reductionist) level of subglacial erosion processes. To a certain extent, therefore, I take these processes for granted, and start my search for explanation at a higher level. The explanation I choose must be consistent with likely casual chains: e.g. at the abstract level of theory, a difference in the balance between active and passive transport suggests itself as a possible cause of the difference in the moraines of the two glaciers. However, at the concrete level of surface appearance, the absence of major supraglacial rock-walls at either glacier immediately suggests that this hypothesis is redundant. To try to explain the difference observed using the active/passive dichotomy of the 'standard' glacier sediment transport model (see below) would be to invoke a chaotic conception. I prefer to invoke contrasts in subglacial drainage as the key contextual factor which determines contrasts in moraine formation. This constitutes a realist type of explanation because it draws on a complex causal structure which is not immediately evident at the level of surface appearance (what the moraines look like); nor can it be read-off as the straightforward product of individual process events.

1.5 WHY STUDY MORAINES?

> Up to the present day they [moraines] have not been paid the attention they deserve, and have been mentioned only incidentally in most of the published works... [yet] they constitute the most important feature of glaciers.
>
> Louis Agassiz, *Études sur les Glaciers*, 1843
> (Cited by Kirkbride, 1989, p. 6)

Moraines provide the ideal opportunity to study the links between process and form using the ideas I sketch above. This is because moraines form at the edge of the ice, and so must represent the totality of process interactions which make up the causal chain of sediment transfer which extends upglacier. This holds both for moraines which exist as contemporary ice-contact features, and for moraines which indicate past ice limits. Thus in this thesis I investigate the process systems which lie behind the pattern of ice-marginal sedimentation seen today at the selected study sites, and I extend the process relationships identified to elaborate the dynamics of these glaciers in the past, including their relationship with Holocene climate change. By the inclusion of these studies of past moraine formation I try to provide a specific bridge between the analytical process traditions of glaciology and the descriptive-historical reconstructions of glacial geology. These past case studies are of potential importance given the significance often attached to the glacial geologic record of Iceland as an indication of past climate change in what is thought to be the key sensitive area of the North Atlantic Polar Front.

My framework of study fills a gap in the moraine literature. Detailed studies which try to explain moraine formation as the explicit product of catchment-wide sediment transfer processes are rare. The majority of moraine studies sit firmly in the tradition of glacial geology, and stress properties of age, position (i.e. past ice extent) and sedimentology. Process relationships tend not to be explored in depth. This applies also to work which provides classification of different moraine types [e.g. Prest (1968), cited by Sugden and John (1976, Table 12.1, p. 236); Lucas and Howarth (1979)], or tries to elucidate typical ice-marginal stratigraphies (e.g. Boulton and Eyles, 1979; Bennett and Glasser, 1996, Ch. 6). Process studies which do exist tend to fall into one of two categories. The first of these investigates 'forwards' process relationships: i.e. what happens to debris, why, and with what feedback effects *after* it appears at the ice edge? (e.g. Lawson, 1982; Kirkbride, 1989). The second of these investigates 'backwards' process relationships, but works within the active/passive dichotomy of orthodox sediment transport theory (e.g. Boulton, 1978; Matthews and Petch, 1982; Small, 1987a, 1987b, 1987c; Benn, 1989: Benn and Ballantyne, 1994). These studies rarely explore processes in the kind of detail I envisage. Because it is so simple, yet widely (but not universally) applicable, this standard framework tends to obscure the need for in-depth process studies. What qualifies as explanation rests at a relatively superficial level (see above, 1.2). This is not necessarily a criticism: if used appropriately, orthodox theory works well (cf. my previous discussion of thermal regime, 1.4). If passive transport is important, explanation at the level of process tends to slip because: a) passive transport processes tend to be simple in comparison with active transport processes (by definition, very little happens to debris in passive transport!); and, b) the process of debris supply which feeds passive transport (i.e. rock-fall) falls outwith the immediate sphere of influence of glacier process studies; it is also imperfectly understood (e.g. Thorn, 1979). However, if passive transport does not figure as a major component of the catchment sediment budget then the simple contrasts of orthodox theory provide a less satisfactory explanation of moraine formation. Indeed, its usefulness collapses with respect to my study glaciers, at which limited exposures of rock-wall mean that the vast majority of debris in transport is of subglacial origin. Therefore contrasts in moraine development must be explained by contrasts in subglacial process regimes. The complexity of these merits detailed study. This task is helped by work in the 1980s and 1990s which has improved enormously our understanding of subglacial processes (for reviews see Hooke, 1989; Paterson, 1994; and Iverson, 1995). Much of this work - particularly that which relates to subglacial drainage - has yet to be incorporated into moraine studies. Indeed, it seems that study of moraines for their own sake has become deeply unfashionable. The major 1995 text edited by Menzies (see 1.4, above) includes comprehensive up-to-date reviews of the erosion and transport of sediments by ice (Iverson, 1995; Kirkbride, 1995a; Lawson, 1995), but these stop short of an account of moraine formation. Moraines fail to figure in the chapter which might be expected to discuss their formation in detail (Maizels, 1995). In my view the current status of the humble ice-edge moraine as the forgotten element of 1990s geomorphology must be remedied. This thesis is my attempt to redress this imbalance.

1.6 STUDY SITE SELECTION CRITERIA

The study sites - principally Sólheimajökull, Gígjökull and Steinholtsjökull (Fig. 1.4) - were selected because of several factors:

- Maritime (i.e. high snowfall, high melt) climate gives rapid turnover of ice, and so rapid rates of bedrock erosion and sediment transport.
- Weak volcanic bedrock of uniform character found at each site favours high rates of debris production.
- Previous work, and my study of aerial photos held at the Iceland Geodetic Survey, revealed major contrasts in past and present moraine formation.
- Reliable dates exist for much of the Neoglacial moraine record at each site (Dugmore, 1987). This gave me the chance to extend my process studies into the past with confidence.
- Proximity to roads with a bus service.

However, two factors must be considered fortuitous. The first of these is/was the paucity of supraglacial rock-walls, which enabled me to ignore supraglacial rock-fall inputs and focus my studies on subglacial processes. The significance of this became clear only as the project progressed. Secondly, sites were chosen because it was assumed that the glaciers a) lie on a steep climatic gradient which controls ice mass size and turnover, and, b) contain distinct bands of englacial tephra which can be picked up by ice radar survey. The initial working hypothesis was that larger, active glaciers (i.e. those closest to the coast) will give rise to the largest moraines (cf. Andrews, 1972). This was to be tested by using ice radar data to constrain ice flow rates and patterns, and then use this information to reconstruct sediment transport by ice. Failure of the ice radar system made this impossible. This was fortunate, because it seems that the original hypothesis qualifies as a chaotic conception: i.e. it was the product of an inadequate causal framework. The rate at which moraines accumulate *rises* with distance inland/greater continentality: i.e. Tindfjallajökull > Gígjökull/Steinholtsjökull > Sólheimajökull - exactly the reverse of what was expected by the original hypothesis!

1.7 STRUCTURE OF THIS THESIS

This thesis proceeds by means of a series of case studies, each of which examines the present or past pattern of ice-marginal sedimentation at one of the study glaciers. To a certain extent,

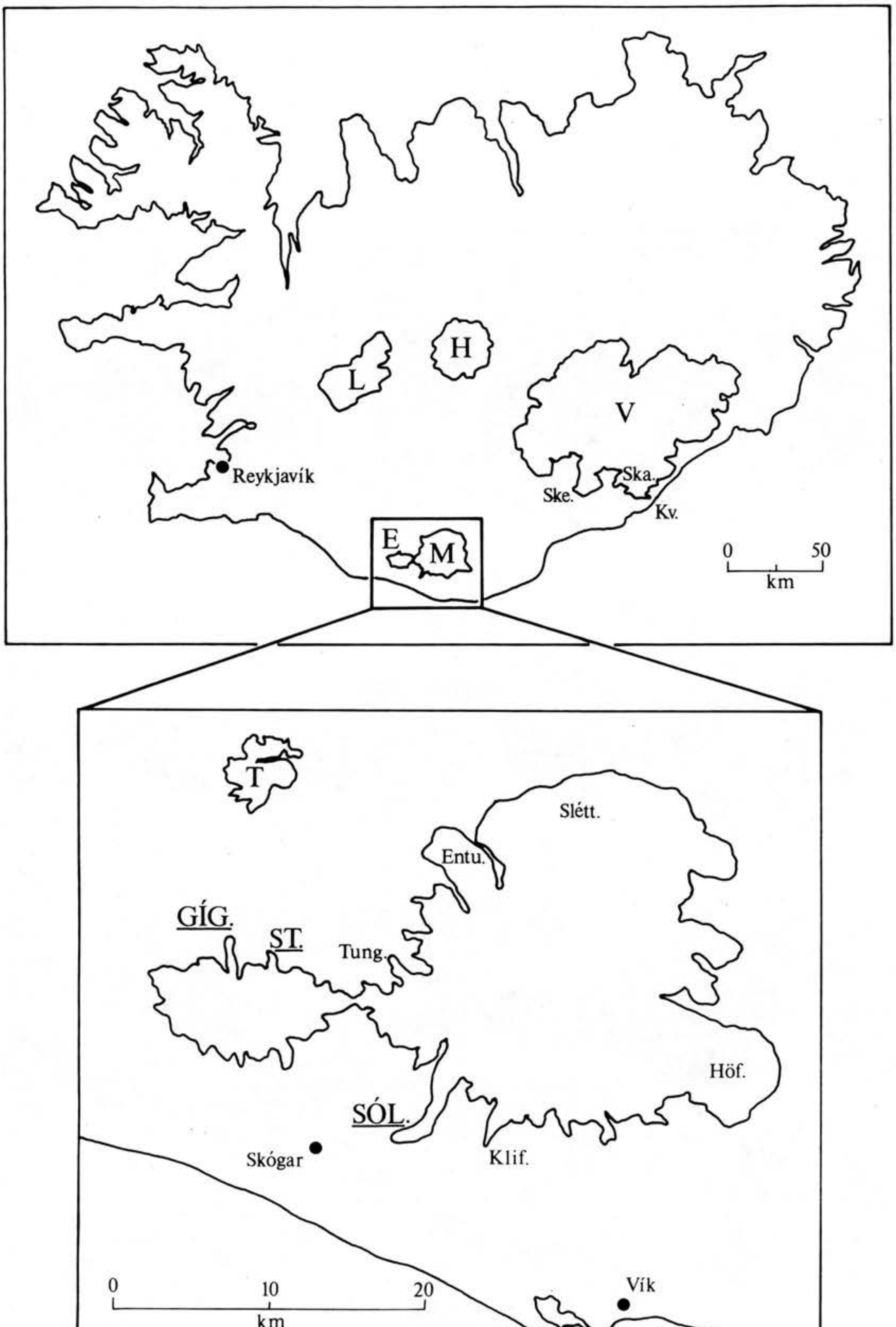

Figure 1.4

Location of study sites.
KEY:
Ice-caps: **E**yjafjallajökull, **H**ofsjökull, **L**angjökull, **M**ýrdalsjökull, **T**indfjallajökull, **V**atnajökull.
Chief study glaciers: **GÍG**jökull, **SÓL**heimajökull, **ST**einholtsjökull.
Other glaciers mentioned in the text: **Entu**jökull, **Höf**ðabrekkujökull, **Klif**urárjökull, **Kv**íárjökull, **Ska**ftafellsjökull, **Ske**iðarárjökull, **Slétt**jökull, **Tung**nakvíslajökull.

each case study stands alone, but, taken together, I believe that the case studies contribute to a single conceptual framework which we can use to analyse processes of sediment transport and moraine formation. The action of meltwater, in different ways, at different times, at different glaciers, emerges as the central feature of this framework.

The case studies complement detailed process studies, centred on the investigation of subglacial hydrology, conducted elsewhere: e.g. at the Haut Glacier d'Arolla, Valais, Switzerland (Richards *et al.*, 1996) and Variegated Glacier, Alaska, USA (Kamb *et al.*, 1985). My work builds on the ideas these studies elucidate, and explores their likely effect on the geomorphology and geology of glaciers. My approach is fundamentally inferential and interpretative (cf. Frodeman, 1995); in part, it is a deliberate attempt to push the limits to which we can use existing ideas of process behaviour with reasonable confidence. The case studies used do not qualify as rigorous hypothesis tests of the type favoured by critical rationalism (e.g. Haines-Young and Petch, 1980). The logical procedures of this philosophy work only with simple, transparent systems of self-evident cause and effect, in which the hypothesis of interest rests on the evidence of simple surface properties alone (e.g. black swans!) (Sayer, 1992). The complex, historical and heavily disguised causal structures responsible for moraine formation fail to satisfy these criteria. The success of my conjectures must be judged in terms of the practical adequacy and internal consistency of my argument. Intuition replaces logic as the touchstone of scientific assessment; the status of provisional knowledge is attached to models which seem to work, which offer a relatively simple, yet powerful explanation (i.e. Occam's Razor), which are not obviously inferior to alternative models, which tie together neatly, and which do not contain any obvious errors (e.g. process assumptions which contravene the laws of physics as applied to ice, water and rock) (Sayer, 1992). I trust that what follows meets these criteria.

II. SÓLHEIMAJÖKULL

CHAPTER 2

Sólheimajökull: Present-day sedimentation

INTRODUCTION

This chapter is inspired by my studies of present-day ice-marginal sedimentation at Sólheimajökull. The outstanding feature is the paucity of active moraine (*sensu stricto*) development, despite the fact that subglacial erosion rates are believed to be high. However, my aim here is not to establish a definitive explanation of why this is so *per se*; rather it is to use the example of Sólheimajökull to introduce the basic principle which guides this thesis: subglacial flushing of debris appears to dominate the sediment budget of perhaps the majority of valley glaciers. If a glacier is to build big moraines, then this flushing constraint must be broken. The second part of this chapter (2.5) proceeds to explore the flushing process in greater detail, using a mixture of theoretical inference, informed supposition, and data derived from detailed studies of subglacial drainage in the Swiss Alps.

2.1 SITE DESCRIPTION

Sólheimajökull (Figs 2.1 and 2.2) drains the south-western sector of the mountain ice-cap Mýrdalsjökull. It is located ~150 km ESE of Reykjavík (63° 30' N, 19° 30' W; Fig. 1.4). Mýrdalsjökull itself sits above the active volcanic massif of Katla. This last erupted in 1918, generating a major *jökulhlaup* (Tómasson, 1996). However, both the centre of volcanic activity, and the major flood route are presently displaced towards the south-east (i.e. Hofðabrekkujökull), and exert little influence on Sólheimajökull. This was different in the mid-Holocene, at which time the centre of volcanic activity was situated beneath the (enlarged? - see Chapter 3) accumulation area of Sólheimajökull. The proglacial zone of Sólheimajökull is largely made up of a stacked series of *sandar*, constructed by a number of floods, largely of volcanic origin: 85% of the thickness of *sandur* deposits contain pumice (Maizels, 1991). The last major volcanic flood occurred in the early historic period (Einarsson *et al.*, 1980). However, the Jökulsá á Sólheimasandi remains an active braided river system which continues to add to, and re-work, its *sandur*. Geothermal activity, as opposed to volcanic eruptions, is still thought to be an important influence on ice and water behaviour (Lawler *et al.*, 1996); the Jökulsá is noted for its distinct sulphurous odour!

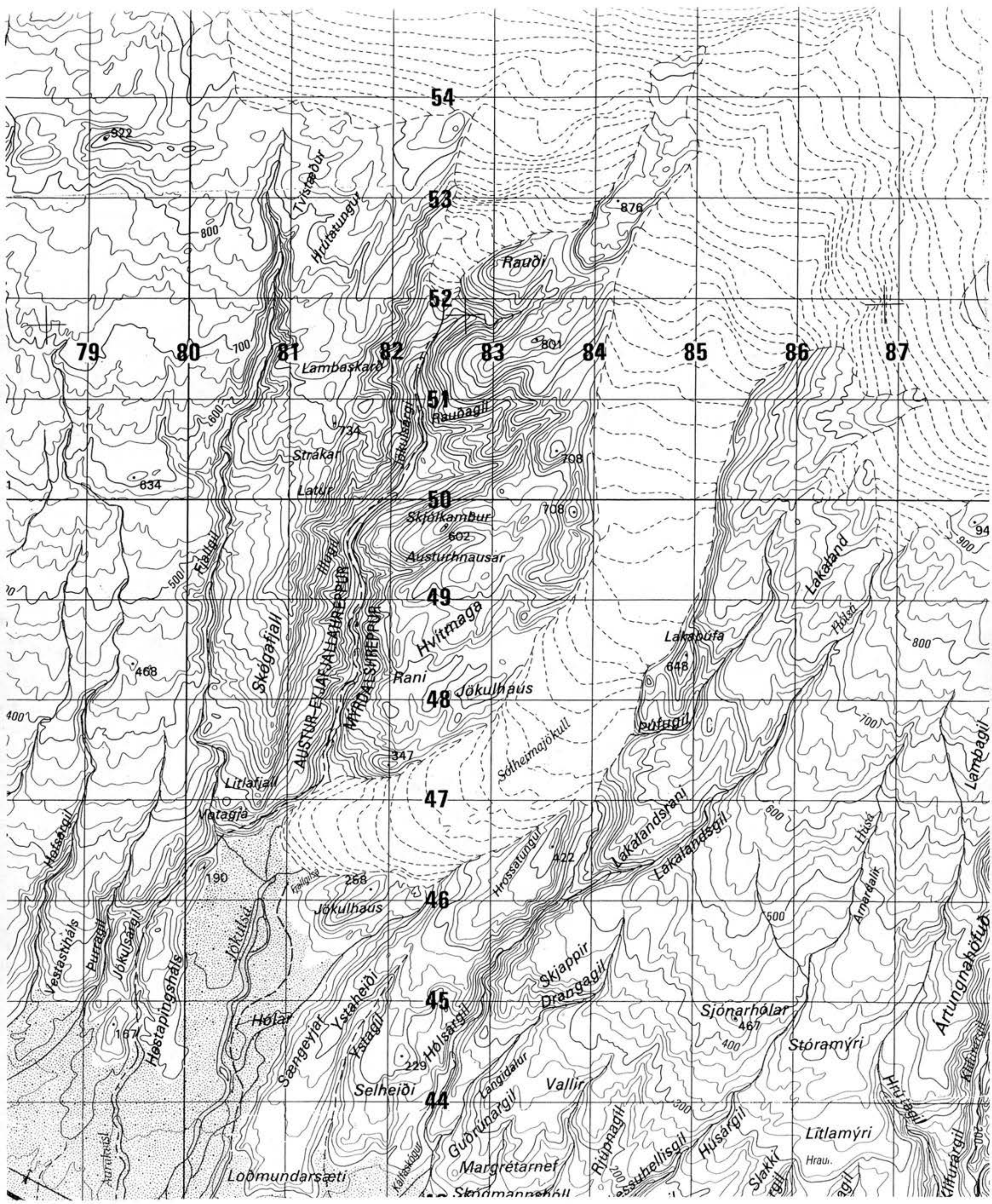

Figure 2.1

Sólheimajökull: extract from sheet 1812: II 'Mýrdalsjökull'. Scale 1:50,000. Copyright: Defense Mapping Agency Hydrographic/Topographic Centre, Washington DC, USA / Iceland Geodetic Institute.

Figure 2.2

Sólheimajökull, looking north over its *sandur* to the main snout.

Solid geology

Katla is an Upper Pleistocene massif (<0.7 Myr). The rocks surrounding Sólheimajökull consist of a series of basaltic lavas, hyaloclastites (glass and crystal mixtures) and pyroclastic rocks - particularly palagonite tuffs (palagonite is hydrated basalt) and volcanic breccias, with scattered rhyolitic intrusions (Icelandic Museum of Natural History/Iceland Geodetic Survey, 1990. Geological Map of Iceland, Sheet 6: South Iceland). Carswell (1983) describes a stacked series of six volcanic units, separated by unconformable contacts indicative of intervening episodes of erosion. These surfaces are frequently marked by striae, or a cover of conglomerate tillite/fluvio-glacial deposits. The lavas and tuffs show characteristics both of subaerial and subglacial (i.e. *móberg*: Saemundsson, 1979) emplacements. This indicates that the Pleistocene was a time of sustained volcanic activity with frequent episodes of glaciation. The key feature of the rock assemblage which affects Holocene moraine-building activity is the weak, friable nature of many of its constituents, particularly the tuffs and breccias. This helps explain the high rates of erosion inferred beneath Sólheimajökull (see below).

Climate and mass balance

The few data which are available support the inference that Sólheimajökull must be an active, maritime-type glacier, characterised by high rates of accumulation (onshore, moisture-laden winds) and ablation. Rist's data (Rist, 1957, cited by Lawler, 1991) show that mean precipitation 1931-1960 was ~1,600 mm yr^{-1} at the bridge Route 1 uses to cross the Jökulsá (~5 km from the glacier snout), rising to over 4,000 mm on the summit of the ice-cap. Maximum July radiation inputs are ~700 W m^{-2} (given likely values of albedo, this melts ~6 mm of ice and ~3mm of snow per hour); mean sea level (Vík) July temperatures are ~10°C, January 0°C. Ablation is enhanced by long summer hours of daylight (Lawler, 1991, 1994). Between 1930 and 1960, Sólheimajökull underwent retreat of 845 m; since 1960 it has advanced ~400 m (Siguðsson, 1992). This advance has speeded-up in recent years, stimulating rumours of an impending surge in 1997 (these appear to be false, however).

Glacier geometry and flow

Sólheimajökull itself covers an area of ~50 km^2. The present distance between the ice-divide (~1,400-1,500 m asl) and the snout (~100 m asl) is ~15 km (glacier centre-line distance); approximately two-thirds of this distance represents the trough outlet; the rest represents the wide and flat upper accumulation basin, much of which is underlain by parts of the Katla caldera. The caldera is ~600 m deep at its greatest, although this part is not presently drained by Sólheimajökull. Ice has cut a trough ~300 m deep in the crater lip; nevertheless, water which drains the upper part of Sólheimajökull must flow upslope before it enters the main trough. The equilibrium line (EL) is though to lie at ~1,100 m asl (Björnsson, 1979). If so, this gives an accumulation area ratio (AAR) of 2.0:1. Bedrock exposures above the ELA are absent.

The outlet - Sólheimajökull proper, effectively a valley glacier - swings in a broad arc of 10 km. Ice radar data reveal a steep-sided, regular trough. The ice is 300-400 m deep, but thins in the final 1.5 km above the snout (Andrew Mackintosh, unpublished data). The basal shear stress typical of this trough section is calculated to be 2.19 bars (2.19 x 10^5 MPa) [using the following representative figures: surface slope = 0.1, mean ice depth at centre-line = 350 m, trough width = 1,500 m, channel shape = semi-ellipse, which gives a shape factor of 0.709 (Paterson, 1994, p. 269)]. 2.19 bars is a high value for mean basal shear stress (cf. the 'typical' value of 1.0-1.5 bars); in terms of basal shear stress conditions Sólheimajökull seems similar to Jakobshavns Isbræ (basal shear stress ~2.0-3.0 bars: Clarke and Echelmeyer, 1996) or quiescent-phase Variegated Glacier (1.8 bars: Raymond, 1987), both of which flow quickly without a major contribution from basal sliding. Direct measurements of flow speed do not exist. My crude calculations suggest balance velocity at the EL is ~100 m yr^{-1} (using best-guess figures: ELA = 1,100 m; mean accumulation above the EL = 2.0 m water equivalent yr^{-1}, flow width at EL = 2,500, flow depth at EL = 300 m). Flow convergence as the trough takes shape is likely to raise ice flow speeds (by up to ~25%?) downglacier.

THE JÖKULSÁ Á SÓLHEIMASANDI

The Jökulsá á Sólheimasandi is the river which drains Sólheimajökull's terminus. Discharge and water quality data have been collected - by the Hydrological Survey of Iceland's National Energy Authority, and by Lawler and co-workers - since 1973 (Lawler, 1991, 1993, 1994; Lawler *et al.*, 1992, 1995, 1996). I use this work extensively here. Bankfull discharge at the Route 1 bridge is estimated at ~100m^3 s^{-1}; typical July discharge is 30-45 m^3 s^{-1}, with a diurnal range of 10-15 m^3 s^{-1}; higher August values (~60-70 m^3 s^{-1}) presumably reflect retreat upglacier of the snowline. A substantial part of total drainage is supplied by the tributary Jökulsárgil. This drains the gorge to the west of Sólheimajökull, at the head of which lies a small tongue of Mýrdalsjökull; it does not drain Sólheimajökull proper (a fact which Lawler's work seems to ignore). Tweed (1992) measured the discharge of this river using surface float techniques, and estimates it is ~8-12 m^3 s^{-1} in summer, the higher value following a period of rainfall and high melt.

Table 2.1 presents Lawler's revised sediment yield/erosion rate data for the Jökulsá catchment (Lawler *et al.*, 1995). These figures eliminate the substantial contribution made by tephra to the total flux of suspended sediment, but add an estimate of bed-load flux, using a Schoklitsch-type function.

Table 2.1

Sediment yield and erosion rate estimates for the Jökulsá á Sólheimasandi catchment (Lawler *et al.*, 1995). Two sets of figures are given for erosion rates: those in column 1 assume that sediment is removed evenly from the whole catchment area; those in column 2 assume that erosion takes places beneath glacier ice only.

JÖKULSÁ Á SÓLHEIMASANDI			
	Sediment yield	**Erosion rate mm yr^-1**	
	t km^-2 yr^-1	**1**	**2**
Suspended-load	9,811	3.63	5.12
Bed-load	7,100	2.63	3.71
Solutes	498	0.18	0.26
TOTAL	17,417	6.45	9.10

Lawler exaggerates the extent to which the sediment output and denudation rates of the Jökulsá catchment are exceptional.[1] Nonetheless, it is indisputable that the Jökulsá evacuates substantial quantities of sediments, much of which (although perhaps not quite as much as Lawler seems to think) must come from beneath Sólheimajökull. Even if a) the Jökulsárgil valley contributes large quantities of sediment, and, b) large quantities of sediment are entrained from loose, vulnerable bank deposits in the 5 km between the glacier snout and the Route 1 bridge, the *qualitatively* robust conclusion must be that Sólheimajökull's subglacial drainage system carries away huge quantities of debris. This is supported by the appearance and sound of the river as it emerges from the glacier snout.

2.2 STRUCTURE OF SÓLHEIMAJÖKULL'S SUBGLACIAL DRAINAGE SYSTEM

No detailed data on the configuration and behaviour of subglacial drainage exist, but it is possible to use theory (Chapter 1.1) to make certain inferences:

- Water must flow within a large conduit or conduits beneath the terminus. The major river which emerges at the snout leads me to anticipate a dendritic, or possibly braided,

[1] See Hallet *et al.* (1996) for a recent review of global contrasts in sediment yields of glacierised catchments. This includes details of calving glaciers in Alaska associated with catchment erosion rates which appear to be an order of magnitude greater than at Sólheimajökull; see also Hunter *et al.* (1996a).

drainage network, dominated by sizeable channels, which persists for several km upglacier. Flow in large channels close to the snout will occur under atmospheric pressure.

- Hooke's (1984) open-flow formula indicates that water in a semi-circular channel which carries 5 $m^3 s^{-1}$ down a 6° slope (figures thought representative of Sólheimajökull) will be pressurised if ice thickness exceeds 330 m. Thus closed flow is likely to prevail within much of the trough. The extent to which open flow persists is likely to depend on the rate at which major conduits break down into smaller conduits with distance upglacier. This is not known. The discharge figure of 5 $m^3 s^{-1}$ I use here is arbitrary, but it presumes flow in a sizeable conduit which carries something like one-sixth to one-tenth of total discharge. Higher discharges favour a greater extent of open flow, and *vice-versa*. However, I think that the conclusion that closed flow prevails is robust.[2]

- Water in the crater at the head of Sólheimajökull's catchment must flow under high pressure within some kind of distributed drainage network. This is necessary if flow is to escape *upslope* to join the main trough (cf. Björnsson, 1988; Guðmundsson *et al.*, 1997; see also Chapter 5.6).

- Basal melt induced by geothermal heat adds to total discharge. Lawler *et al.* (1996) describe a cauldron located just inside the ice divide which implies melt subsidence caused by a subglacial hot spot. The exact contribution of geothermal heat is unknown: the fact that winter flow levels of the Jökulsá fall by just two-thirds implies that it is important. However, my crude energy balance calculations suggest that summer discharge levels are roughly consistent with a solar radiation flux of 700 W m^{-2} (ignoring melt induced by sensible heat transfer, and rainfall inputs to the catchment) (cf. Grímsvötn, Vatnajökull, wherein geothermal/volcanic heat sources account for ~80% of total melt; Björnsson, 1983).

- Discharge will rise as temperatures rise, and the snowline retreats upglacier (see Chapter 1.1). The increase in discharge must be accommodated by some kind of change in the structure of subglacial drainage: an increase in flow speeds, and/or changes to channel geometries, and/or an increase or decrease in the number of active channels are all possible (cf. the cavities to conduits transition at the Haut Glacier d'Arolla, Switzerland: Richards *et al.*, 1996). However, drainage reorganisation at Sólheimajökull is likely to be less spectacular than that expected of glaciers in the Alps because winter flows here are high relative to summer flows, and the melt season in Iceland is longer.

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[2] This conclusion is supported by the observation that Hooke's formula (as I use it) is thought to over-predict the extent of open flow (Hooke, personal communication, in Menzies, 1995b, p. 222); see also Box 7.1.

- The high mean basal shear stress can be read in terms of the style of subglacial drainage. A shear stress of 2.19 bars implies a rough bed which provides substantial resistance to ice flow. This can be taken to indicate limited bed coverage by water at low pressure, which suggests a well-developed conduit network (i.e. high effective pressures across large parts of the glacier bed mobilise high resistance to sliding). This idea of (part-) sliding with a high-drag bed (i.e. sliding is driven by high basal shear stress, not by low bed roughness and/or low effective pressures) is broadly consistent with what is known of the quiescent-phase behaviour of Variegated Glacier (Raymond, 1987)[3], or the 'fast polar outlets' of the West Antarctic Ice Sheet (Cooper *et al.*, 1982; McIntyre, 1985). Sólheimajökull seems to share the characteristics of these outlets: fast sliding of thick ice starts beneath steep surface slopes as ice flow converges and accelerates at the head of a bedrock trough. However, alternative drainage styles are also (qualitatively) consistent with the calculated shear stress. Widespread water under high pressure - a linked-cavity system? - is compatible with high shear stress values if the bulk of flow resistance is supplied by large-scale bedrock roughness elements which impart form drag irrespective of the bed drainage conditions (cf. Alley, 1993). Alternatively, a sand-gravel-cobble bed (which allows for distributed pore-water drainage) is likely to provide high resistance to flow if ice can infiltrate the interstices between clasts.
- Knight and Tweed (1991) suggest that periodic drainage of a small, ice-marginal lake is caused by episodes of faster sliding which open-up escape routes for water trapped in the lake. If correct, this provides indirect evidence for some kind of distributed (linked-cavity?) system: water inputs to a cavity system are likely to enhance sliding, and some kind of marginal cavity with connections to interior drainage must open-up to allow the lake to drain. Lawler (1994), however, disputes this mechanism, arguing that lake water escapes because it is warmer than $0^{o}C$ and so melts itself a drainage passageway.
- Lawler provides evidence from direct stream flow water quality data. These clues merit further discussion.

GEOCHEMICAL TRACERS

Lawler *et al.* (1996) describe a natural tracer experiment using a parcel of geothermal fluid (largely H_2S) released to the drainage system at the head of the basin. Seismic records indicate the likely time and location of the fluid's release; its arrival at the Route 1 bridge is detected by electrical conductivity meter; and the distance travelled is taken as the valley centre-line distance. Various permutations give water flow speeds between 0.026 and 0.132 m s^{-1}, which,

[3] As against its surge behaviour: i.e. extensive, high-pressure water at the bed, with fast sliding but relatively low shear stress:

in conjunction with evidence of parcel dispersion, suggests some kind of distributed system lies beneath Sólheimajökull [cf. 0.025 m s^{-1} for the 'textbook' linked-cavity system believed to exist beneath surge-phase Variegated Glacier (Kamb *et al.*, 1985); this figure is also derived from a single experiment, but with the advantage of direct borehole dye injection].

Evaluation

Lawler *et al.* (1996) do not consider in detail what the results of this natural tracer experiment might or might not mean. As I result, I find their simple conclusion that Sólheimajökull is drained by a distributed network misleading. It is likely that a distributed network of some kind exists beneath *part* of Sólheimajökull - indeed, the mechanics of water flow upslope within the crater require this - but this does not necessarily mean that the entire glacier is underlain by distributed drainage. Three points give cause to question Lawler *et al.*'s argument:

- Lawler *et al.*'s interpretation of the water quality data ignores the possible impact of geothermal springs in the Jökulsárgil valley. These are shown on the geological map (Sheet 6: South Iceland). [Andy Dugmore, personal communication, 1997.]
- This 'experiment' - as with all simple tracer studies - is weakened by its 'black box' character. Exactly what happens to the water between its release and its detection is not known. It is perfectly possible that the water travels quickly within conduits for the best part of its journey, but spends the majority of its *time* travelling a short distance very slowly within a highly inefficient system. Slow seepage of water is a distinct possibility a) within bedrock, between the point of the fluid's origin, and the glacier bed; and, b) within the crater.
- A single tracer experiment is unlikely to characterise a glacier's drainage system adequately (cf. the 533 separate dye injections used to build up the picture of drainage at the Haut Glacier d'Arolla: Richards *et al.*, 1996). For example, parallel drainage of different types can exist (e.g. Fountain, 1993), drainage configuration can change as the summer proceeds (e.g. Richards *et al.*, 1996), or water can make surprising switches in its choice of route (e.g. Nienow *et al.*, 1996).

Verdict. The evidence of this single event provides some support for the idea that Sólheimajökull is drained by some kind of distributed network, but it is not convincing.

'HEARTBEAT EVENTS'

Lawler *et al.* (1992) and Lawler (1993) describe 'heartbeat events' in the flow records of the Jökulsá which are believed to indicate *conduit* drainage. These involve a sudden fall in discharge, followed by a rebound to higher-than-previous levels, and a return to previous flow levels. The flow pulse (drop, rebound-with-over-compensation, recovery) typically lasts ~40 minutes, and its trace is symmetrical about the prevailing level of flow: the analogy is with an ECG heartbeat trace. Suspended sediment concentrations show no response to the fall in discharge, but concentrations rise sharply with the rebound, then fall away (Lawler *et al.*, 1992, Fig. 5). 34 of these events were detected between 4 July and 10 August 1988, although none were seen to occur in 1991. Events were usually superimposed upon rising stage. These events occur too quickly to be caused by changes in surface melt or rainfall conditions, nor is it likely that subaerial processes can dam/divert flow as observed. Therefore Lawler *et al.* conclude that it is blockage of subglacial channels by ice-collapse which gives rise to heartbeat events.

Evaluation

The idea of conduit blockage and temporary diversion of flow is plausible. Ice dams on ice-marginal channels are described by Ballantyne and McCann (1980) at the Schei Glacier, Ellesmere Island; subglacial examples are described by Burkimsher (1983a) at Pasterzengletscher, Austria (this reference records similar observations at Austre Okstindbreen, Norway, and Gornergletscher, Switzerland: personal communications by W. Theakstone and D. Collins), by Hooke *et al.* (1985) at Bondhusbreen, Norway, and by Wiseman (1996) at Hintereisferner, Austria. Judging by the frequency with which the author received painful blows to the knee from ice blocks floating in its proglacial stream, ice collapse events are also common at the Haut Glacier d'Arolla! Collins (1979a) suggests that channel blockage by sediments - e.g. collapse of moraine channel banks - is also important. However, these events are likely to occur as water levels fall, and probably have the greatest impact on smaller channels.

Lawler *et al.*'s (1992) type example of a heartbeat event (i.e. their Fig. 5) shows a fall in discharge of ~10 $m^3 s^{-1}$; total discharge at the instigation of the event was ~75 $m^3 s^{-1}$. If it is assumed that 12 $m^3 s^{-1}$ was supplied by the subaerial Jökulsárgil, then the subglacial discharge fell by ~16% (unless of course it was the Jökulsárgil which was dammed!). This implies blockage of a major conduit which carries one-sixth of the total subglacial discharge. Such a conduit is likely to occur close to the glacier snout (an inference possibly consistent with the additional assumption that ice-block collapse is likely to reflect thin, brittle ice which lacks the support of pressurised water), although this will not prevent the effects propagating upstream. The interpretation must be that the ice dam raises water pressures, so forcing water out of the conduit across adjacent areas of the glacier bed, or into the main body of the ice. Once the dam clears, pressure within the conduit will fall, permitting water to return. This water is enriched with sediment which implies that a) the water is temporarily forced between ice and bed, not into the ice, and, b) these extra-conduit excursions permit water to tap fresh sediment sources. This means that these events are possibly important in sweeping the bed of a glacier free of sediment (see 2.5, below).

Hydraulic dams?

It is pertinent to note, however, that diversion of water can be caused by hydraulic, not just by mechanical (i.e. ice-block or sediment) dams. Lawler's data are not supported by observations of floating ice blocks in the river as the events subside, as are the other examples I cite above. Pressures can rise rapidly in conduits if the rate at which water is supplied exceeds the immediate capacity of the channel. Under certain conditions, pressure waves can travel downstream faster than flood waves of elevated discharge; the extent to which this happens depends on initial channel geometry (shape, size, roughness) and the rate at which discharge rises (Fountain, 1992). This means that pressure maxima are imposed upon the channel faster than flow speeds within the channel can respond. The effect is to force water out of the conduit [which requires that transient conduit pressures exceed the 'separation pressure' (Iken and Bindschadler, 1986; Chapter 1.1) at which adjacent cavities can open-up]. This can be a common occurrence: conduits are thought to adjust gradually to transmit mean discharge at weekly time-scales (Röthlisberger's steady-state theory, 1972), whereas, in the short-term, conduits behave as rigid pipes. Bindschadler (1983) notes that a 10x increase in discharge can give rise to a 100x increase in transient water pressures. If the behaviour of two conduits is out-of-phase then periodic diversion, storage and release of water at their confluence is likely to be common (e.g. conduits which drain surface meltwater from different altitudes, with different flow travel distances and times, which therefore rise and fall at different times and rates) (Smart, 1990).

Studies at the Haut Glacier d'Arolla provide field evidence of hydraulic dams. Nienow *et al.* (1996b) discuss hysteresis (i.e. non-unique discharge-velocity relationships) exhibited by multiple tracer injections into the same moulins. Negative hysteresis (i.e. through-flow velocities from a given moulin/conduit stay low, and water is held back, as bulk discharge rises) and greater dye dispersion are thought to reflect hydraulic dams, and water backing-up within moulins. This is not inconsistent with transient high pressure events which force water out of conduits; both may occur simultaneously, although it is likely that storage of excess water in moulins acts as a safety valve which dampens extra-conduit excursions. However, water bubbling-up in, or spurting out of, crevasses and moulins was/is frequently seen at the Haut Glacier d'Arolla (my observations; Peter Nienow, personal communication). These events demonstrate episodes of overburden-plus water pressure, as is (more than) required to push water out of conduits (i.e. separation pressure < ice overburden pressure). Further evidence of these events is supplied by Hubbard *et al.*'s (1995) borehole study of what they call a 'variable pressure axis' (VPA). This VPA represents a conduit which runs through a zone of distributed drainage. Within this zone of distributed drainage, water pressures remain more-or-less constant, at levels close to overburden, throughout the day; however, water pressures in the conduit display high variability. This creates a diurnally-reversing hydraulic gradient. As water throughputs rise by early afternoon, conduit water pressures rise above those of the surrounding discrete drainage, and water is forced out of the conduit: cavities progressively open-up as a wave of elevated pressure travels through and away from the conduits, so water escapes sideways. In the night conduit discharge falls, the hydraulic gradient reverses, and water can drain back to the conduit - in this case through permeable bed sediments. Turbidity measurements show that this return flow carries with it a pulse of sediment swept from the bed (cf. the suspended sediment signature of the Sólheimajökull heartbeat events). Similar diversion, storage and release of water was detected by Collins (1979b) at Gornergletscher, Switzerland. These studies imply that steady-state, single system drainage models are not necessarily best-suited to an understanding of subglacial water and sediment flows: it is the transient interactions between different elements of the total drainage structure which count.

HYDROLOGY OF SÓLHEIMAJÖKULL: SUMMARY

Without the support of a detailed programme of direct investiagtion (cf. Richards *et* al., 1996) any inferences made must be tentative. However, based on the above arguments I favour a network in which large conduits close to the snout, between which a residual distributed system exists, are progressively replaced upglacier by smaller, closed-flow conduits, and elements of distributed drainage. This is consistent with Lawler's water quality studies, what is seen at the snout, and what can be inferred from the facts that a) discharge levels will fall as catchment area falls upglacier; b) surface meltwater inputs will fall sharply above the snowline; and, c) water at the head of the basin must traverse an adverse slope. Non-steady-state behaviour is likely to be important, including both diurnal and seasonal change in drainage. Major unknowns include: a) the contrast between summer and winter drainage structures; b) the nature of the glacier bed (e.g. rough or smooth? hard or soft?); and, c) the potential impact of glacier advance, and possible change in flow speed (e.g. faster sliding suppresses cavity collapse; glacier advance across the proglacial zone increases the proportion of the bed made up of till).

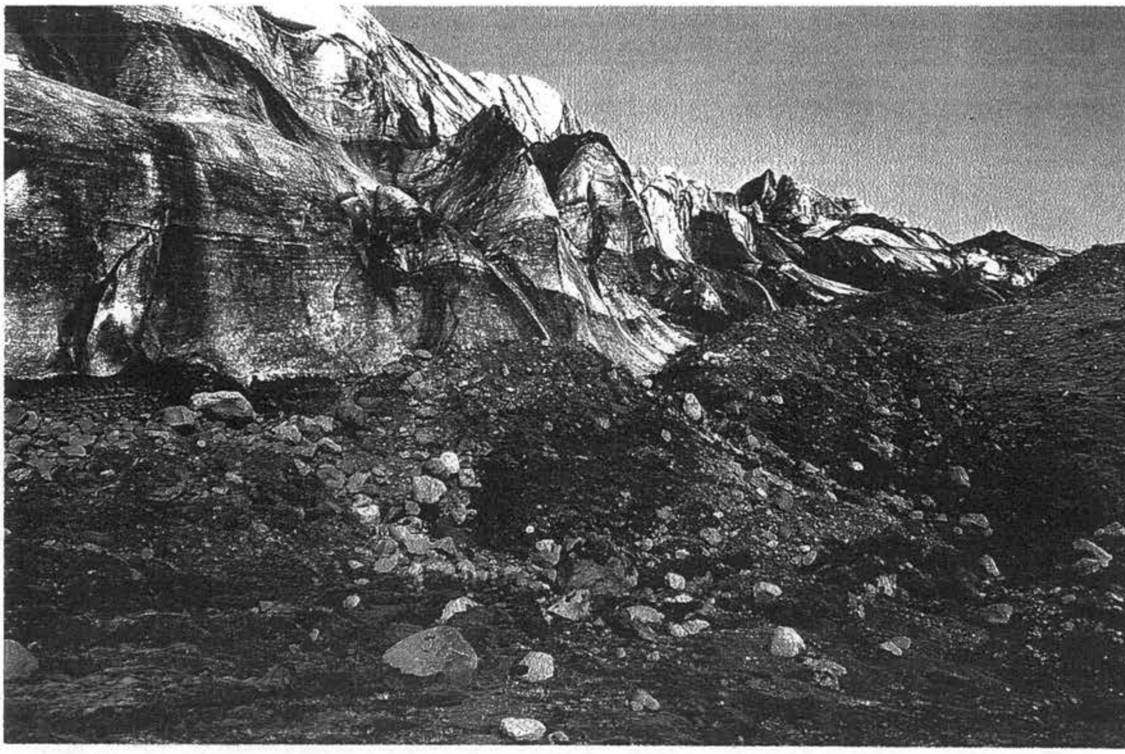


Figure 2.3 (top)

Sólheimajökull: push moraine forming at the ice edge, Jökulhaus (south). Its maximum height is ~4 m.

Figure 2.4 (bottom)

Sólheimajökull: minor lobe east of Jökulhaus (south), seen from Hrossatungur. Note: 1) the abundance of Katla 1918 ash, and, 2) the paucity of other englacial and ice-marginal debris (cf. Fig. 4.3a).

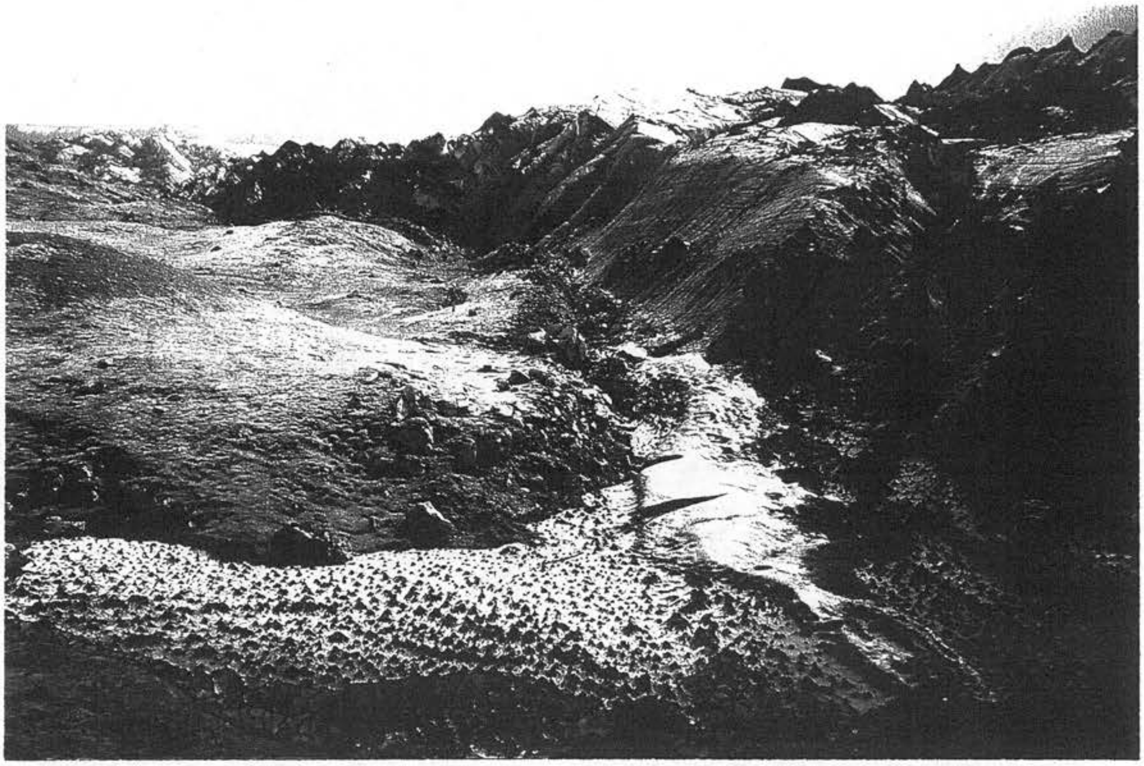
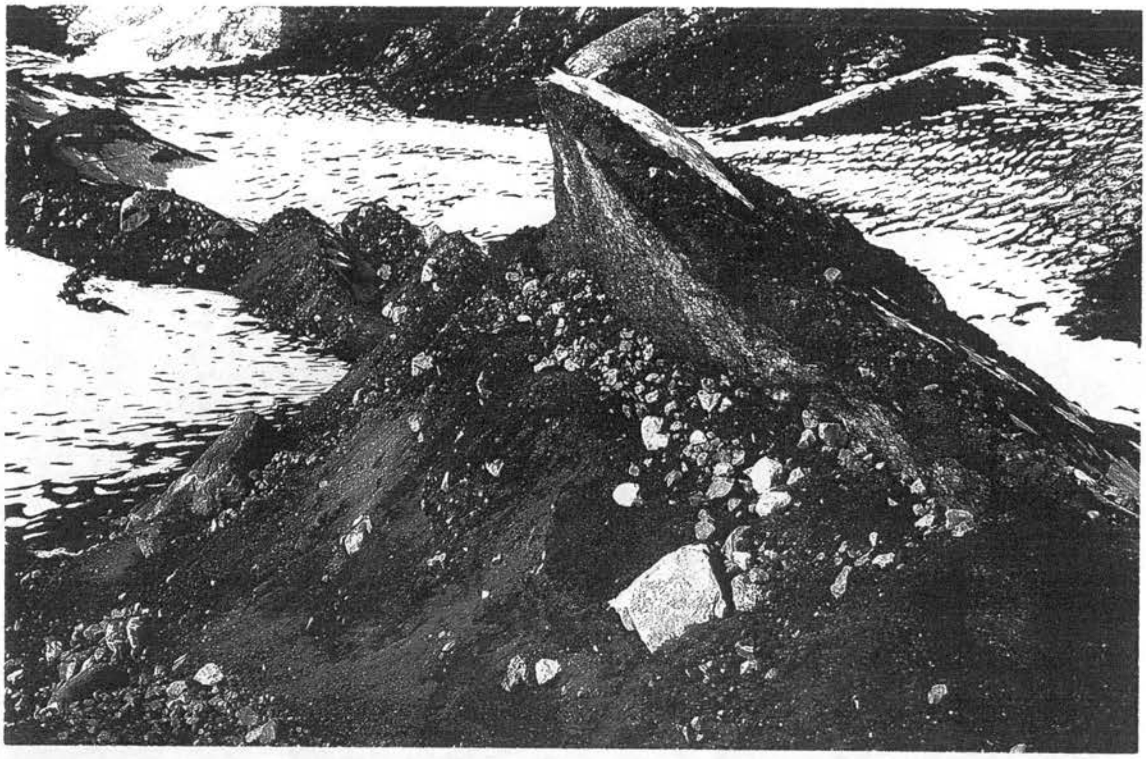


Figure 2.5 (top)

Typical view of the ice edge of Sólheimajökull (cf. Fig. 5.1), Dr A. J. Dugmore for scale. The ice is largely debris-free, with the exception of the Katla 1918 ash. The vegetated moraine to the left dates from the Little Ice Age.

Figure 2.6 (bottom)

Sólheimajökull: section through water-lain sediments which form the low, vegetated ridge immediately in front of the snout. The length of the ice hammer is 50 cm.



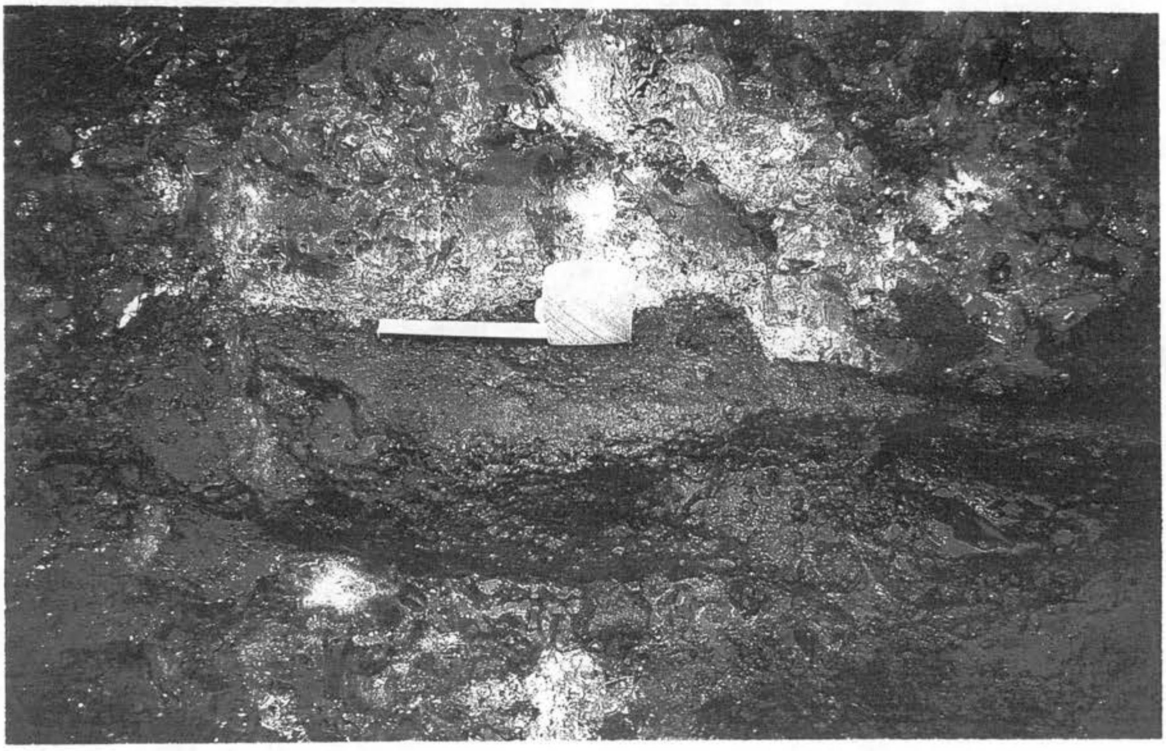


Figure 2.7 (top)
Sólheimajökull: small dump moraine (height ~0.75 m) fed by fragments of basal ice. This example is both unusually large and unusually continuous.

Figure 2.8 (bottom)

Sólheimajökull: englacial debris bands exposed at the snout. These features are similar to the englacial debris bands found at Gígjökull and Steinholtsjökull (cf. Figs 5.2 and 5.3), but are less well-developed here.

Figure 2.9

Sólheimajökull, snout: small ice-contact fan formed at the portal of an englacial conduit.

2.3 ICE-MARGINAL SEDIMENTS

I examined ~3 km of Sólheimajökull's southern margin between the Jökulsá outflow and the Hólsárgil gorge (Sheet 1812: II, GR 838 470), and ~2 km of its northern margin between the snout and the ice-dammed lake (not shown on Sheet 1812: II) at GR 834 485 (see Fig. 2.1). Difficult access prevented me from extending my search upglacier. My intention was to map the different ice-sedimentary structures, the different debris facies, and the different styles of ice-marginal moraine formation, and to estimate the volume of debris presently accumulating. The major features are as follows:

- The best-developed moraines are the push moraines in the vicinity of Jökulhaus (Fig. 2.3).
- The widespread expanse of dirty, debris-rich ice which seems to ring the terminus (Lawler, 1991) is an illusion. The vast bulk of this debris consists of a veneer of Katla 1918 tephra (Fig. 2.4).
- The bulk of the ice margin searched was free of glacial debris (Fig. 2.5).
- Active sedimentation at the snout relates largely to fluvial processes. These build up spreads of debris in ice-contact deltas and channel-bar systems where minor streams emerge from beneath the ice.
- The 'moraine' in front of the snout is a proglacial fluvial deposit. Sections reveal sub-rounded clasts, crude cross-bedding, and elements of fining-upwards structures, capped by a well-sorted, laminar series of fine sands (Figure 2.6).
- Scattered expanses of basal ice were found, but it is important to stress that these were a rarity. Two types of basal ice were identified in the field: a) clast-rich ice, with dense layers (1-2 mm thick) of silts, sands and gravels, separated by layers of clean ice and bubbles (debris concentration by mass ~20%); and, b) clast-poor ice, containing dispersed muddy-to-silty debris (debris concentration by mass ~5%). Where found, layers of basal ice were often separated by thicker layers of debris-poor, coarsely-crystalline (~2.0 cm diameter) ice.
- Exposures of stratified basal ice fed *small, broken* dump-moraine ridges (Fig. 2.7). Clast samples from these gave a mean roundness score of 2.79, s.d. ±0.05, indicative of basal transport (see Chapter 4.2 for details of clast analysis). As with the basal ice, it is crucial to emphasise the fragmentary nature of these ridges.
- Scattered englacial debris bands were found. The typical thickness of these was 1.0-2.0 cm. The contact between debris and ice was sharp. The typical interval between debris

bands was 50-60 cm; the intervening ice was clean, coarsely-crystalline, and debris-free (Fig. 2.8).

DISCUSSION

In this thesis my main concern is with debris produced by subglacial erosion, freshly delivered to the ice margin, and dumped directly out of ice to form moraines in the strict sense of the term. The bulk of ice-marginal sediments at Sólheimajökull fail to fit into this category:

- **Push moraines.** Some of these are impressive, but the majority of push moraine debris is pre-existing till which is recycled. Ice in contact with these push moraines was clean, with the exception of the Katla 1918 tephra.
- **Katla 1918 tephra.** This has a striking visual impact, but its volume is insignificant. This tephra was emplaced by volcanic eruption, not by subglacial erosion, so I do not consider it further.
- **Fluvial deposition.** Large quantities of debris are carried to the snout by small subglacial streams, thereafter to from part of proglacial water-laid deposits. The section shown in Fig. 2.6 is typical of a high-energy, proximal channel-bar complex, built up under conditions of fluctuating discharge. The sediments which make up these features were probably produced by subglacial erosion, but the final part at least of their journey to the ice margin was/is accomplished in water, so strictly they are not moraines. Note: the rate at which debris is added to these ice-proximal fluvial deposits is negligible when set alongside the rate at which debris is washed away by the Jökulsá!

Small quantities of debris only are dumped directly out of ice to form moraines:

- **Basal ice.** The two main types of basal ice observed at Sólheimajökull correspond to the two main types found (in far greater quantities!) at Gígjökull and Steinholtsjökull. These were identified as 'laminated' and 'clear' facies ice, using Hubbard and Sharp's (1995) classification (see Chapter 6.2). However, at Sólheimajökull distinct bands of clear, coarsely-crystalline ice intervene between the debris-rich layers of basal ice. This clear ice was thought to be englacial ice. If correct, this demonstrates the tectonic intercalation of basal and englacial ice to form 'stratified' facies ice. 'Solid' facies basal ice - a layer of debris stuck to the sole of the glacier - was found in small quantities at the snout. As with the push moraine debris, this solid facies ice represents recycling of pre-existing fluvio-glacial deposits and/or till, so it is not important here.

(continues after Box 2.1)

BOX 2.1
Moraine-building activity at Sólheimajökull: potential and reality

A) POTENTIAL

Here I calculate how much material would be added to moraines each year if all the debris produced beneath Sólheimajökull were to contribute to moraine accumulation. Lawler's data (Table 2.1) indicate that the mean 'catchment erosion rate' (strictly the quantity of solid material washed away) for the Jökulsá á Sólheimasandi is:

Jökulsá		
	suspended load	9,811 t km^-2 yr^-1
plus	bed-load	7,100 t km^-2 yr^-1
gives	TOTAL	16,911 t km^-2 yr^-1

The area beneath Sólheimajökull is ~50 km^2, so the total mass of material removed from the basal transport zone each year is:

$$\mathbf{16{,}911\ t\ km^{-2}\ yr^{-1} \times 50\ km^2 = 845{,}550\ t}$$

If we assume that:

- the mean density of this debris is 2.7 t m^{-3}
- the void ratio of a typical freshly-formed dump moraine is 30% (Small, 1987b)
- this debris is not washed away, but contributes to ice-marginal moraine formation

then the total volume of material added to moraines at Sólheimajökull would be:

$$\mathbf{(\ 845{,}550\ t\ /\ 2.7\ t\ m^{-3}\ )\ /\ 0.7 = 447{,}400\ m^3}$$

which, if it were evenly distributed over the full ~20 km of the ice margin below the equilibrium line, would build a continuous moraine ridge with cross-sectional area:

$$\mathbf{447{,}400\ m^3\ /\ 20{,}000\ m = 22.37\ m^2}$$

If this moraine ridge forms an isosceles triangle, with side slope angle 30° (i.e. the approximate angle of repose for dump moraine sediments), then it would be ~3.60 m high.

(PTO)

Box 2.1 (continued)

B) REALITY

Here I estimate how much of the debris produced beneath Sólheimajökull each year actually ends up in moraines. *Where found* a typical dump moraine was ~0.75 m high (see Chapter 2.3 and Fig. 2.6). If this moraine has the same isosceles triangle form as I assume above (this is a reasonable approximation), then its cross-sectional area is 0.97 m^2. If this represents an annual ridge (as seems likely), and making the extremely generous assumption that this ridge runs the full 20 km length of the below-EL ice margin without break, then the quantity of material actually added to the ice margin each year is:

$$\mathbf{(0.97\ m^2 \times 20{,}000\ m) \times 0.7 \times 2.7\ t\ m^{-3} = 36{,}666\ t}$$

This is less than 5% of the material carried away by the Jökulsá. This shows that sediment loss to water dominates the subglacial sediment budget of Sólheimajökull. Moraine forming activity is minor in comparison.

ERRORS

The data and assumptions I use here are rather crude, so some error is likely. Below I give high and low estimates of relevant factors and results using an arbitrary error band of ±25%. (I assume that the values used for subglacial area, length of ice margin below the EL, and density of sediment are fairly robust.)

Errors (±25%)			
Factor	**Units**	**HIGH**	**LOW**
A) POTENTIAL			
Solid denudation rate	t km^-2 yr^-1	21,139	12,683
Mass of debris washed away	t	1,056,950	634,150
Void ratio	none	0.875	0.525
Mean moraine XSA	m^2	37.28	16.78
Mean moraine height	m	4.65	3.10
B) REALITY			
Mean moraine height	m	0.9375	0.5625
Mean moraine XSA	m^2	1.52	0.55
Mass of debris in moraines	t	71,820	16,706

EVALUATION

If we take the low estimate of material removed by the Jökulsá and the high estimate of material added to moraines, the quantity of debris involved still differs by an order-of magnitude [i.e. (71,820 t / 634,150 t) x 100 = 11.3%]. This does not change my above conclusion. However, these calculations assume that the moraine ridges which actually do form are continuous. This is certainly not so: much of the ice margin is entirely free of debris produced by subglacial erosion and dumped out of ice. Taking this into account, my guess is that the difference rises to two orders-of-magnitude. **River sediment output is probably 100 to 200 times greater than is delivery of sediment to the ice edge within ice.**

- **Debris bands.** The origin of these is not clear. Their constituent debris was similar to that which made up the matrix of the basal ice, but other features did not obviously fit into any recognised category of basal ice (cf. Hubbard and Sharp, 1995). The intervening clear ice looks like englacial ice. One possibility is that these debris bands are small-scale examples of the features seen at Gígjökull and Steinholtsjökull which relate to flow of dirty water through englacial ice (Chapters 4 and 5). This interpretation is consistent with the ice radar evidence of a small overdeepening ~1.5 km upglacier of Sólheimajökull's snout, and/or the occurrence of high pressure water flow heartbeat events, both of which potentially switch water flow from basal to englacial routes (see above, 2.1 and 2.2). Further evidence of debris-laden englacial water flow was provided by the presence at the snout of a small ice-surface fan, containing rounded clasts, which appeared to have built-up below the mouth of a small englacial conduit (Figure 2.9). I stress that these debris band features at Sólheimajökull are small, discontinuous and infrequent; in terms of the overall sediment budget they are insignificant. This is *not* the case at Gígjökull and Steinholtsjökull.

Overview. Present-day ice-marginal sedimentation at Sólheimajökull is dominated by push-moraine activity, and by small subglacial streams exiting the ice. Delivery of debris to the ice edge within ice (the Katla 1918 tephra excluded) is non-existent over the majority of the ice-margin surveyed; elsewhere it is minor. The scattered basal ice exposures, debris bands and dump moraines which do exist represent freak, low-probability occurrences. Their presence reflects isolated events which are largely peripheral to the dominant process regime.

Wider perspectives

Superficially this presents something of a paradox. In terms of its erosivity, Sólheimajökull is supposedly Icelandic National Champion, and holds its own in the Global Premier League, yet its ice-marginal moraine-building abilities are pathetic. This imbalance can be illustrated by means of a simple calculation (Box 2.1). The solution to this apparent paradox, however, is straightforward. It is precisely because subglacial water transport dominates the sediment budget of the Jökulsá catchment that Lawler a) is able to use gauging techniques with confidence to reconstruct catchment erosion rates, and, b) arrives at such high figures for these. My guess is that ice accounts for just ~1% (if that) of total sediment output at Sólheimajökull. Table 2.2. presents comparable estimates of the proportion of total debris transport accomplished by ice for other glaciers.

Table 2.2

Some estimates of the proportion of of total glacier debris flux discharged by ice.

Sediment output in ice			
Glacier	**Location**	**% output in ice**	**Source**
Breiðamerkurjökull (1732 to 1890)	South Iceland	~3.0	Björnsson (1996)
Bondhusbreen	Norway	~5.0	Hooke et al. (1985)
Grand Pacific Glacier	Glacier Bay, Alaska, USA	1.3	Hunter et al. (1996b)
Margerie Glacier	Glacier Bay, Alaska, USA	4.1	Hunter et al. (1996b)
Muir Glacier	Glacier Bay, Alaska, USA	22.8	Hunter et al. (1996b)
Bas Glacier d'Arolla	Valais Alps, Switzerland	~24	Warburton and Beecroft (1993)
Glacier de Tsidjiore Nouve	Valais Alps, Switzerland	~40-60	Small (1987b)
Sólheimajökull	**South Iceland**	**<1 (?)**	**This study**

Sediment can have a complex transport history, but sediment particles (unlike quantum particles!) cannot be in two places at the same time. Alternative transport regimes determine different sediment budgets, and so different styles of ice marginal sedimentation. This brings me to the key idea of my thesis:

EROSIVE GLACIERS CANNOT BUILD LARGE MORAINES IF WATER SWEEPS CLEAR THE BULK OF THEIR DEBRIS.

This simple idea is almost entirely ignored by the established literature, yet it is the intensity of the flushing process which I believe exerts the primary control on the mixture/sequence of sediment transport pathways which determine the pattern of ice-marginal sedimentation. For a given rate of debris supply, the rate at which ice-marginal moraines accumulate will be inversely proportional to the efficiency with which water can capture and evacuate debris. As the example of Sólheimajökull shows, debris production by ice and debris retention within the ice are two very different things. The early 1970s *Journal of Glaciology* debate between Boulton and Andrews over which type of glacier carries most debris seems to have arisen because both (at least initially) seem not to have grasped this distinction, and what it implies for the way in which debris is partitioned between alternative transport pathways. Boulton argues that sub-polar glaciers carry the greater debris load, because the change from warm to cold thermal regime favours debris entrainment, but suppresses meltwater activity; Andrews argues that greater quantities of debris - and so larger moraines - are associated with active temperate glaciers. As Kirkbride points out (in what is the only comment on flushing I can find in published syntheses of glacial geomorphology):

> [T]his debate... [i.e. Boulton vs. Andrews] raises questions about the distinction between erosion and entrainment capability. Warm-based glaciers appear to lose much of their potential debris load to flushing from the glacier sole by meltwater and derived rainfall, giving the appearance of being relatively free of debris even though they are highly erosive.
>
> (Kirkbride, 1995a, p. 284)

[Boulton and Andrews set out their key ideas in Boulton (1972a) and Andrews (1972a). The debate itself can be followed in Boulton (1970, 1971, 1972b) and Andrews (1971, 1972b).]

2.4 SÓLHEIMAJÖKULL AND GLACIER DE TSIDJIORE NOUVE: TYPE EXAMPLES (?) COMPARED

(A wider assessment of the role of subglacial flushing)

Glacier de Tsidjiore Nouve (Fig. 2.10) is a small valley glacier, located at the head of the Val d'Hérens, Canton Valais, Switzerland (catchment area 4.15 km^2, length 5 km, relief 1,600 m). It starts at a wide, high-altitude (~3,500-3,796 m) snow saddle, narrows and plunges through a steep ice-fall (~0.75 km at ~30^{o}) to end as a narrow tongue of ice confined between large moraine ramparts (distal slopes to 60 m in height). The ice-fall, and the upper part of the tongue are enclosed by steep rock walls. Detailed studies of a) its ice facies and patterns of sediment transport; b) the form, sediments and chronology of its lateral moraines; and, c) the water quality of its proglacial streams make Tsidjiore Nouve perhaps the type example of an Alpine glacier (see Small, 1987a, 1987b; Souchez and Lorrain, 1987; and Gurnell, 1994 for summaries of this work). Lawler (1991) notes the resemblance between Sólheimajökull and Tsidjiore Nouve: the snouts of both are shot-through with englacial debris bands (Lawler's source is Gurnell, 1987, Figure 12.10, p. 337). These debris bands have very different origins, however: those at Sólheimajökull consist of tephra, whereas those at Tsidjiore Nouve reflect rock-fall inputs (Small and Gomez, 1981); neither set reflects subglacial erosion. Gurnell's picture shows an unusually large amount of *clean* ice exposed; above its terminal ice cliff, the tongue of Tsidjiore Nouve is smothered by a thick layer of angular debris. The comparison between Sólheimajökull and Tsidjiore Nouve is useful not because of any similarity (as Lawler implies), but because of the *contrast* between the two glaciers. Important differences exist between their debris transport systems and patterns of moraine-forming activity. Sediment budget data collected at Tsidjiore Nouve by Small and co-workers (1987b) permit direct

Figure 2.10

Glacier de Tsidjiore Nouve (centre-to-right of picture), seen across the Val d'Arolla. Note: 1) the large areas of supraglacial rock-wall (including the north face of the Pigne d'Arolla (centre-top); and, 2) its debris-choked snout, flanked by large lateral moraine ridges.

comparison with Sólheimajökull. These data illustrate the impact that subglacial flushing can have. The figures I use here are the mean of Small's (1987b) maximum and minimum estimates for catchment erosion rate at Tsidjiore Nouve. Where appropriate I make use of an arbitary ±25% error band. N.B. The values for moraine volume given here exclude corrections for void space.

Moraine accumulation

As a first approximation, if flushing is negligible, I expect:

$$\mathbf{MAMA = MCER \times (CA / P{<}EL)} \qquad (1)$$

MAMA:	mean annual moraine accumulation, $m^3\ m^{-1}\ yr^{-1}$
MCER:	mean catchment erosion rate, $m\ yr^{-1}$
CA:	catchment area, m^2
P<EL:	glacier perimeter below the equilibrium line, m: i.e. theoretically, that part of the ice-margin along which sediments can accumulate

This gives:

Sólheimajökull: $0.00625 \times 50 \times 10^6 / 20 \times 10^3 = \mathbf{15.625\ m^3\ m^{-1}\ yr^{-1}}$

(LOW = $11.725\ m^3\ m^{-1}\ yr^{-1}$; HIGH = $19.525\ m^3\ m^{-1}\ yr^{-1}$)

Tsidjiore Nouve: $0.00193 \times 4.15 \times 10^6 / 4 \times 10^3 = \mathbf{2.002\ m^3\ m^{-1}\ yr^{-1}}$

(LOW = $1.502\ m^3\ m^{-1}\ yr^{-1}$; HIGH = $2.509\ m^3\ m^{-1}\ yr^{-1}$)

which suggests that Sólheimajökull adds at least ~5 times as much debris to each metre of its margin per year than does Tsidjiore Nouve. This clearly is not so: field observations of moraine accumulation under what seem to be the *most favourable* conditions show that Sólheimajökull adds ~1 $m^3\ m^{-1}\ yr^{-1}$ (Box 2.1), but Tsidjiore Nouve typically adds ~2-8 $m^3\ m^{-1}\ yr^{-1}$. This mismatch between expectation and observation is resolved if we add a suitable 'flushing factor' to equation 1, which then becomes:

$$\mathbf{MAMA = MCER \times (CA / P{<}EL) \times (1 - FF)} \qquad (2)$$

FF: flushing factor = proportion of debris removed by water

Equation 2 then gives:

Sólheimajökull: FF = 0.995-0.95; **MAMA = 0.08-0.8** $\mathbf{m^3\ m^{-1}\ yr^{-1}}$

(LOW = $0.06\ m^3\ m^{-1}\ yr^{-1}$; HIGH = $0.98\ m^3\ m^{-1}\ yr^{-1}$)

Tsidjiore Nouve: FF = 0.54 (±25%: FF = 0.405-0.675); **MAMA = 0.92** $\mathbf{m^3\ m^{-1}\ yr^{-1}}$

(LOW = $0.49\ m^3\ m^{-1}\ yr^{-1}$; HIGH = $1.49\ m^3\ m^{-1}\ yr^{-1}$)

The true value of **MAMA** for Sólheimajökull probably falls close to the LOW estimate (i.e. ~ 0.1 $m^3\ m^{-1}\ yr^{-1}$?), whereas the mean value of **MAMA** calculated by me for Tsidjiore Nouve using Small's (1987b) data (i.e. ~0.9 $m^3\ m^{-1}\ yr^{-1}$) is likely to be fairly reliable. This implies that an order-of-magnitude contrast in rates of moraine growth exists between the two glaciers: i.e. **MAMA** Tsidjiore Nouve >> **MAMA** Sólheimajökull. This overturns the discrepancies in mean catchment erosion rates (i.e. Sólheimajökull > Tsidjiore Nouve), which illustrates the crucial role of flushing. Even if the HIGH estimate applies to Sólheimajökull, and the LOW estimate to Tsidjiore Nouve (which I think is highly unlikely) we still have a situation whereby a 3x advantage in mean erosion rate is reduced by water action to just a 2x advantage in mean moraine accumulation.

Subaerial debris inputs

We can expand equation 2 to give:

$$\mathbf{MAMA} = \{ [\mathbf{MSER \times GBA \times (1 - FF_A)}] + [\mathbf{MRRR \times RA \times (1 - FF_P)}] \} / \mathbf{P{<}EL} \qquad (3)$$

MSER:	mean subglacial erosion rate, $m\ yr^{-1}$
GBA:	glacier bed area, m^2
FF_A:	flushing factor for debris in active transport
MRRR:	mean rock-wall retreat rate, $m\ yr^{-1}$
RA:	rock-wall area, m^2
FF_P:	flushing factor for debris in passive transport

Equation 3 assumes that all debris of subaerial origin enters passive transport - a necessary simplification here.

The second term in equation 3, which describes debris inputs of subaerial origin, is of negligible importance to Sólheimajökull. Tsidjiore Nouve, however, is very different. If, in the absence of any relevant data, FF_P is set to zero, and following Small's estimate that 20% of moraine inputs are from active transport, then:

MAMA = $\{ [0.0014 \times 3.6 \times 10^6 \times (1 - 0.85)] + [0.0019 \times 1.55 \times 10^6] \} / 4{,}000$

$= 0.19\ m^3\ m^{-1}\ yr^{-1} + 0.73\ m^3\ m^{-1}\ yr^{-1}$

$= \mathbf{0.92\ m^3\ m^{-1}\ yr^{-1}}$

This illustrates the 'override' effect major rock-fall inputs exert on moraine formation. Because the bulk of this debris never makes it to the glacier bed it stays separate from the major drainage routes, and so is immune to flushing. (Supraglacial or englacial streams can remove sediment, but supra/englacial drainage tends to be much less widespread than is drainage at the glacier bed.) If subaerial debris is excluded, **MAMA** at Tsidjiore Nouve falls to 0.19 $m^3\ m^{-1}\ yr^{-1}$

(LOW = 0.0 m^3 m^{-1} yr^{-1}; HIGH = 0.57 m^3 m^{-1} yr^{-1}), a value similar to that expected at Sólheimajökull.

Evaluation

It seems fair to conclude with confidence that subglacial drainage at Sólheimajökull seems to be substantially more aggressive than it is at Tsidjiore Nouve - **(1 - FF_A)** at Sólheimajökull is possibly an order-of-magnitude higher - but flushing of basal debris is still important at Tsidjiore Nouve. Subglacial debris transport by water dominates subglacial debris transport by ice. This inference matches the evidence of Hubbard and Sharp's (1995) basal ice studies: debris-rich laminated facies ice was absent at Tsidjiore Nouve, as it was at 7 out of 11 Alpine glaciers included in their study. The process of Weertman-type regelation implies that formation of laminated basal ice must be widespread beneath temperate sliding glaciers, so its absence at Tsidjiore Nouve means that it is destroyed, or that it exists, but it is not identified because it contains no debris (see Chapter 6.2 for a full discussion of basal ice processes; see also Box 2.2). Either way, it highlights the importance of subglacial debris flushing.

2.5 TOWARDS A THEORY OF SUBGLACIAL FLUSHING

INTRODUCTION

What it is exactly which controls subglacial flushing is a major puzzle. It is clear that a full understanding of ice-marginal moraine development requires detailed knowledge of the ways in which water and sediment interact at the bed of a glacier. However, acquisition of this knowledge is likely to be a hugely complex problem, in part because:

1. It is extremely difficult to make direct observations of subglacial sediment transport [see Pohjola (1993) and Stone *et al.* (1993) for examples of state-of-the-art techniques: respectively miniature video cameras and miniature turbidity and EC meters, both for use within boreholes]. The limited techniques and observations which are available are unlikely to characterise the full range of behaviour.

2. The range of factors likely to influence flushing - e.g. the magnitude-frequency distribution of run-off; the time-space distribution of sediment supply; possible time-space combinations of different types of drainage - is likely to be enormous.

3. Given identical boundary and initial conditions, it is possible that there is more than one stable means by which evacuation of water and sediment can be achieved.

It seems to me that the phenomenon (phenomena?) of subglacial flushing possibly qualifies as an example of 'trans-science' (cf. Douglas, 1988): i.e. we can ask the question "what is it that controls subglacial flushing?", but - such is the complexity of real-world subglacial behaviour - a satisfactory answer is beyond the immediate capabilities of present-day science. This outlook is perhaps too bleak, but it is certain that flushing, viewed in its true context as a dispersed process which acts with different intensities across the entire glacier bed, is poorly-suited to traditional reductionist models of scientific explanation; nor is the accumulation of field data particularly helpful. Empirical studies of sediment yield (Hallet *et al.*, 1996 for a review) usually take the underlying processes for granted; indeed, when converting sediment flux to rates of erosion many rely - often implicitly - on the assumption that flushing is 100% efficient (e.g. Lawler, 1991; Lawler *et al.*, 1992, 1995). In this section I take a step back from the sediments found at Sólheimajökull to consider the different factors which are likely to combine to control subglacial flushing. No comprehensive study of this problem presently exists, although several studies - notably the work of D. N. Collins (see below) - provide important clues. What follows is a mixture of deductive thinking which builds on established elements of water sediment transport theory, critical review of others' work, and my own ideas.

CONTROLS ON SEDIMENT TRANSPORT

With few exceptions (e.g. the possibility of full pipe-flow sliding bed transport: Saunderson, 1977) the mechanics of subglacial sediment transport by water must mirror the mechanics of subaerial water transport. Here I introduce some of the key building blocks of process fluvial geomorphology, and explore their likely impact on subglacial sediment transport by water. Processes unique to the subglacial environment can change the boundary conditions which influence sediment transport (e.g. channel migration by differential melt of ice-walled channels, or deformation of soft till which carries sediment to the channel), and key features of the subglacial regime will have different characteristic properties (e.g. closed-flow within part-ice-walled channels imparts different roughness values; driving stresses enhanced by the extra component of pressure head; change in the magnitude-frequency distribution of run-off).

Sediment transport by rivers is a complex process. The flux of water and sediment imposed from upstream interacts with channel materials within a given reach to create specific patterns of water flow, sediment transfer and channel morphology. A delicate balance of forces relates channel form, flow characteristics, sediment transfer and channel morphology (e.g. Ashworth and Ferguson, 1986; Lane, 1995). Flux, not stability, is likely to prevail in high-energy rivers which carry large quantities of sediment. In subaerial rivers, changes in flow conditions are accommodated by differential transport of sediments; however, subglacial waterways are likely to possess additional degrees of freedom because channel geometry also responds to changes in ice flow, water pressures and ice-wall melt. Even so, these extra complications do not alter the basic premise which must be central to a theory of subglacial flushing, namely that sediment transport reflects the balance between the competence and capacity of water flow, and the quality and quantity of sediment supply. If flushing is to be 100% efficient:

- Water must access all sediments produced by subglacial erosion.
- Water must flow with sufficient force (speed and volume) to remove these sediments.
- Transport must be sustained between the point of sediment origin and the edge of the ice (in reality, transport of coarse clasts tends to proceed as a series of jumps between pockets of temporary storage; sustained transport is achieved if the number of clasts which enter into storage equals the number of clasts released from storage).

These conditions need not hold instantaneously, but must be met over a given interval (several years to several decades?) appropriate to the process(es) and time-scale(s) of moraine formation.

Competence

This is strictly defined as the largest clast a given water flow is able to carry. It is frequently defined as the largest clast in transport, but because of supply limitation effects this measure can underestimate a flow's true competence. Competence rises with some index of flow strength: strictly this is flow bed shear stress (which is a force), but flow velocity or discharge are often used for practical purposes. The balance of forces which controls transport (i.e. the transport/deposition threshold) is difficult to assess, so competence is often replaced with ideas of the critical flow required to entrain sediment of a given calibre [i.e. the storage/erosion threshold (Richards, 1990)]. Two end-member models are defined: *size-selective entrainment* (transport) and *equal mobility* (entrainment/transport). The first incorporates the intuitive view that stronger flows are necessary to move larger clasts, because resistance to transport scales with clast weight (e.g. Naden, 1987); the second assumes that clast hiding and protrusion effects largely cancel out the effect of clast weight, so a wide range of sediment sizes starts to move at a given shear stress. This is likely to occur if the channel bed is made up of a mixture of sediment sizes (as is likely with subglacial tills; see Haldorsen, 1981, on the particle size distribution of subglacial wear products), whereupon it is the median clast size which tends to determine the threshold of entrainment (e.g. Andrews, 1983; Brayshaw, 1985). Whereas much work on fluvio-glacial sediment transport uses size-selective transport as its interpretative framework - Saunderson (1977) invokes the sliding bed to explain the apparent *anomaly* of poorly-sorted sands, gravels and cobbles in the Guelph esker - it is possible that equal mobility provides a better framework of analysis (e.g. Walder and Fowler, 1994). Entrainment functions are not necessarily a reliable guide to sediment transport, however. Less energy is needed to keep a clast in transport than is required to start it moving (i.e. the deposition process is not the simple reverse of the entrainment process), so over-passing of coarse sediments can occur even if the critical flow for entrainment is not attained at that spot. Hiding and protrusion effects act with much less strength on clasts in transport, so that entrainment with equal mobility is not incompatible with size selective deposition. Yet more complexities arise because the flow shear stress responsible for entrainment and transport of sediments is itself a part-function of channel perimeter sediments (i.e. the product of past transport events) which impart flow resistance (e.g. Ferguson *et al.*, 1989).

Issues of competence are important because subglacial erosion - particularly plucking - is likely to generate large quantities of coarse sediment. A complete theory of subglacial flushing must take this into account. Whereas fast flows in major conduits are likely to be competent to carry all but the largest boulders, these cover only a small fraction of the glacier bed. The bulk of erosion will take place well away from the major axes of discrete flow: step (i.e. bedrock obstacle) cavity systems are thought particularly conducive to both plucking and abrasion (Iverson, 1995). If flow strength is low, as is likely in many kinds of distributed networks, then transport of coarse clasts is likely to be episodic, perhaps even non-existent. However, it is difficult to make predictions with any kind of confidence because lots of different types of distributed flow exist: e.g. a quasi-stable regelation water film within which flow competence rarely exceeds 20 microns (Hallet, 1979b), the inefficient, tortuous, low water velocity linked-cavity network envisaged by Kamb (1987), and the relatively efficient, high-speed, low sinuosity, low water pressure network inferred for Glacier de Tsanfleuron, Switzerland by Sharp *et al.* (1989a) all qualify as distributed drainage.

Cavities. Throughflow of water in cavities is sluggish (Chapter 1.1), so under a wide range of flow conditions I imagine coarse clasts to remain immobile. It is possible that this debris accumulates as a coarse lag deposit, as observed, for instance, at the Ghiacciaio di Pré de Bar, Courmayeur, Italy. Here the immediate forefield, exposed by recent rapid retreat of the glacier, consists of a matrix-poor spread of cobbles and boulders. The existence of widely-incompetent distributed drainage, in conjunction with the protection coarse clasts give to fine sediments (i.e. conditions of equal mobilty delay/prevent removal of fine debris) helps to explain the widespread occurrence of soft beds ('sub-sole drift') beneath many glaciers (e.g. Bas Glacier d'Arolla, Gornergletscher, Haut Glacier d'Arolla and Glacier de Tsidjiore Nouve, all in Canton

Valais, Switzerland). If water cannot shift coarse clasts, perhaps ice can. A second plausible scenario involves intervals in which coarse clasts remain immobile within water-filled cavities, which alternate with episodes of cavity drainage, cavity closure, clast incorporation into, and transport by, basal ice. In this case, coarse sediments - or, indeed, clusters of mixed-size sediments held together by a central large clast - are likely to move towards the ice-margin as part of a parcel of basal ice (even if the ice of this parcel episodically melts and re-forms: see Chapter 6.2 and 6.3). This should improve moraine-building potential. If flushing of this debris is to occur, melt of this basal ice must coincide with contact between the debris which is freed and suitably competent elements of subglacial drainage. Progressive comminution of debris in basal ice, which should increase the ease of entrainment of individual clasts, is possibly important to this process.

Orifices. The orifices which link cavities[4] are likely to create a second kind of competence barrier which limits flushing efficiency. Water flow speeds and shear stresses in orifices are likely to be sufficiently high to carry pebbles and cobbles (Kamb, 1987, envisages jets of water - $\sim$0.25-0.5 m s^{-1} - squirting between cavities), but the size of coarse clasts is likely to exceed the size of these orifices (i.e. sub-dcm scale). This must mean that large clasts which make it into orifices regularly get stuck. Several different responses are likely once this happens. If sediment clogs the channel, rising water pressures possibly cause flow to divert elsewhere, to leave a relict, debris-choked orifice which closes down and becomes part of the basal ice (cf. the *englacial* relict conduit debris bands found at Gígjökull and Steinholtsjökull: see Chapter 5). Alternatively, it is feasible that a local steepening of the hydraulic gradient in the vicinity of the clast jam enhances local melt rates so that the channel is enlarged just enough to permit the clast to pass through. In effect, a high-pressure water jet pile-drives the clast(s) through the orifice. A plausible middle-way scenario is that of a part-debris-choked orifice which simultaneously functions both as a sediment trap and a relatively efficient hydraulic connection (i.e. sediment accumulates, but water continues to pass through). If this orifice is later shed by the cavity system, to be incorporated into the process of collective, co-operative orifice growth by which a conduit system is established at the expense of cavities (Kamb, 1987), then this sediment is likely to be released. In this event the pulse of suspended sediment thought to indicate catastrophic reorganisation of subglacial drainage (Collins, 1989; Sharp, 1991) is likely to be accompanied by a bed-load pulse as well. However, if such a bed-load pulse exists, the chances of detecting it at the glacier snout are slim: a) unlike with suspended sediment transport, competence and/or capacity constraints are likely to dampen the passage of bed-load

4 And perhaps also exceptionally wide but shallow (possibly collapsing) conduits: see Shreve, 1985; Röthlisberger and Lang, 1987, pp. 275-276; Chapter 5.6.

waves; and, b) high-resolution monitoring of bed-load outputs from stream portals tends to be difficult, unreliable, and rarely used as a field technique.

Capacity

Capacity is defined as the total quantity of sediment which a given flow is able to carry. This is usually expressed as the sediment flux, mass per unit width per unit time. This intuitively relates to the excess flow strength above that required for sediment entrainment. This can be expressed by a number of empirical-functional sediment transport relationships, such as the Meyer-Peter and Müller tractive force equation, which predicts bed-load flux to be proportional to (τ_0 - τ_{crit}), for which τ_{crit} is defined by a suitable entrainment function (e.g. Richards, 1982, pp. 110-113; Church, 1996, pp. 150-153; Walder and Fowler, 1994).

It is self-evident that the volume of sediment which can be carried by water has a major influence on subglacial sediment budgets. If sediment inputs consistently exceed the transport capacity of local drainage - even if that drainage is technically competent to move sediment of the size involved - debris will accumulate. This perhaps favours its incorporation into, and subsequent transport by, basal ice, or it may encourage the growth of a deforming till patch. If these coalesce, the result will be a deforming bed proper. (The potential relationships between debris transport by basal ice, by subglacial till deformation, and by subglacial water flow, and the impact these have on styles of ice-marginal sedimentation, have yet to be explored in full; see Chapter 8 for some of my ideas.) Transport of suspended sediment is much less vulnerable to capacity-control than is bed-load transport, which is why suspended sediment studies can be used with some success to reconstruct subglacial drainage: i.e. sediment supply factors, not flow properties, act as the key control of sediment flux.

In practice it is difficult to differentiate between sediment transport and channel change controlled by changes in flow competence, and sediment transport/channel change controlled by changes in flow capacity (e.g. see Ashmore, 1991). Current work in proglacial rivers (which are likely to resemble major sediment-rich, open-flow conduits typical of glacier termini) seems to favour size-selective transport under the influence of competence relationships at low flows, but as flow levels rise, equal mobility transport controlled by capacity relationships tend to take over. Conditions of equal mobility imply the abrupt release into transport of large slugs of sediment; once this has happened subsequent small changes in flow strength - e.g. as channel geometry changes - which reduce flow capacity are likely to induce deposition of much of this slug. If these relationships hold across wide areas of a glacier's bed then I expect winnowing of fine sediments in relatively isolated parts of the drainage network at low flow levels (e.g.

Hubbard *et al.*, 1995), with large-scale flushing of sediments in bulk restricted to major channels and/or subglacial flood events (see below).

Sediment supply relationships

These are important because it is common for sediment transport by water to be supply-limited. This can reflect lack of flow competence, and/or exhaustion of sediment supplies (which implies water runs over solid bedrock and/or clean ice). Efficient flushing relies on sustained contact between water and sediment. Sediment delivery to channels involves debris transport within ice, by till deformation, or within tributary channels (i.e. sediment is carried to the channel). Alternatively, the channel can go to the sediment, by means of channel migration, or excursions of water beyond the confines of the channel (i.e. subglacial flood events, which can occur at a wide range of magnitudes and time and space scales; cf. my discussion of 'heartbeat events', 2.2, above). If contact between water and sediment is limited - i.e. a high proportion of sediment flow paths fail to intersect competent/capable drainage elements - then flushing will be suppressed, with moraine-building potential increased correspondingly.

BASAL ICE

Sediment incorporated into basal ice is not immediately available to subglacial water flow. Unless it is released by clast expulsion (e.g. into cavities) in the presence of a sharp pressure gradient, or ice fracture induced by water-imposed tractive force, sediment must first be freed by melting of ice. Basal ice melts regularly as pressures rise against the stoss faces of bed obstacles, but these same high pressures tend to divert water flow away from such areas. Net melting of ice is induced by geothermal heat, by heat from sliding - both expected to be widely dispersed across the glacier bed, so sediment so released will not necessarily coincide with competent/capable water flows - and by frictional heat from water flow (viscous dissipation): see Box 2.2. If water flow is uniform, the melt rate of ice by running water is determined by unit stream power (= energy loss per unit channel width per unit channel length per second) divided by the latent heat of fusion. I anticipate that, on average, a clast 2.0 cm in diameter embedded in temperate basal ice will fall out once 1.0 cm of surrounding ice melts. Table 2.3 (overleaf) shows how long this will take for a realistic range of channel flows.

Table 2.3

Likely clast release from basal ice. This shows **A**) time taken to melt out a 2.0 cm diameter spherical clast, and, **B**) time taken for basal ice to cross channel, calculated for a realistic range of channel flows and sliding speeds. Clast loss is expected if **A** < **B** (i.e. shaded cells). These calculations assume:

- Semi-circular conduit, bedrock floor.
- Hydraulic gradient = 0.1.
- Water flow-lines intersect ice flow-lines at a right-angle (i.e. basal ice crosses water by the shortest route).

Channel diameter and melt rate calculations follow Hooke (1984, equations 3 and 6).

Clast release from basal ice				
		A	**B** Time taken for basal ice to traverse channel, hrs	
Discharge, m^3 s^-1	**Channel diameter, m**	**Time required for melt out, hrs**	**Sliding speed 0.1 m d^-1**	**Sliding speed 1.0 m d^-1**
0.01	0.12	~120	28.6	2.86
0.10	0.35	~12	**83.3**	8.33
1.00	0.87	~1.2	**207**	**20.7**

The time for which ice and water are in contact relative to channel stream power (which determines melt) must exert an important influence on flushing (depending also on the ratio of debris entrained in basal ice to loose debris, which is a major unknown). Repeated contacts between basal ice and a number of small channels are necessary to match the melt impact of contact between basal ice and a large channel with higher stream power (but this is likely to be offset in part by the fact that heat advection, which transfers the melt impact downstream, is likely to be more pronounced in large channels). Duration of contact reflects not just the proportion of bed taken up by drainage elements with different specific stream powers, but also the angle of contact between ice/debris and water flow. If the flow paths of basal ice and water are broadly parallel, as is expected of Röthlisberger-channels (Chapter 1.1), contact between ice and water will be prolonged, irrespective of the relative speeds of ice and water flow. However, if ice and water flow paths meet at an angle, and the channel location is fixed (e.g. by bedrock as a Nye-channel, cavity or orifice) ice will move over and across the channels, in which case the melt/flushing potential is controlled by the balance between channel stream power, channel width, and ice sliding speed (Table 2.3). The 1.0 $m^3 s^{-1}$ channel introduced above is 0.87 m wide (Hooke, 1984, equation 3), which means that ice sliding at 1 m d^{-1} (0.042 m hr^{-1}) will take ~20 hours to cross it if ice and water flow paths meet at right angles. Clast capture is expected, because it requires just 1.2 hours to melt out the 2.0 cm clast. However, with smaller channels/lower stream powers, clast survival is predicted. The width of the 0.1 $m^3 s^{-1}$ channel is 0.35 m; perpendicular ice sliding at 1 m d^{-1} will cross this in ~8.3

BOX 2.2
Annual destruction of basal ice at Sólheimajökull

Here I use simple calculations - which rely on crude approximations - to estimate:

1. The total quantity of basal ice at Sólheimajökull which is destroyed each year.
2. The quantity of basal ice which is destroyed by geothermal heat and heat from sliding.
3. The quantity of basal ice which is destroyed by heat released by water running at the glacier bed.

My 'best-guess' calculations imply that the quantity of heat supplied to the base of the glacier matches that necessary to melt all the basal ice likely to be present. If this inference is anything like correct, it explains why, on average, so little basal ice is found at the margins of Sólheimajökull today. However, these calculations cannot predict the distribution of basal ice at the glacier bed, nor the distribution in time and space of basal heat supply. Thus isolated pockets of basal ice which are in contact with areas of below-average heat supply can survive, and so make it to the ice edge to contribute to moraine accumulation (see Chapter 2.3).

BEST-GUESS CALCULATIONS

Thickness of basal ice destroyed

Lawler's data (Table 2.1) show that ~850,000 t of solid debris are produced by subglacial erosion below Sólheimajökull each year. I assume that half of this debris is immediately washed away by basal drainage, leaving the other half to be incorporated into basal ice. This debris will survive in basal ice for some time, but eventually it is released by melt-out and flushed away. I take the typical debris content of basal ice to be 13% by mass, equivalent to 4.7% by volume. [This figure is the mean of 42 samples of basal ice taken from Sólheimajökull, Gígjökull and Steinholtsjökull (see Chapters 4 and 6). I use the data from Gígjökull and Steinholtsjökull as well because the quantity and quality of data from these two glaciers are better. Basal ice from Gígjökull and Steinholtsjökull is likely to be broadly representative of Sólheimajökull also, because all three slide at similar speeds over similar bedrock.]

Mass of basal ice destroyed each year:

$$\mathbf{4.25 \times 10^8\ kg \times (100 / 13) = 3.27 \times 10^9\ kg}$$

If a) debris concentration in basal ice by mass = 13%, b) density of ice = 900 kg m^{-3}, and, c) density of sediment = 2,700 kg m^{-3}, then the density of this basal ice = 984.6 kg m^{-3}. This then makes the *volume* of basal ice destroyed:

$$\mathbf{3.27 \times 10^9\ kg / 984.6\ kg\ m^{-3} = \sim 3.3 \times 10^6\ m^3}$$

~95% of this mixture is clean ice (i.e. get rid of the volume of basal ice taken up by debris) = $3.135 \times 10^6\ m^3$. If this ice is evenly distributed across the glacier bed, the *thickness* of basal ice destroyed each year is:

$$\mathbf{3.135 \times 10^6\ m^3 / 50 \times 10^6\ m^2 = \sim 0.06 m}$$

(PTO)

Box 2.2 (continued)

This figure makes sense: Hubbard and Sharp (1993) simulate basal ice formation and reformation by way of Weertman-type regelation (see Chapter 6.2), and find that only in exceptional circumstances does the basal ice layer exceed 0.1 m in thickness.

Net basal melting

This refers to melt of basal ice brought about by heat other than that supplied by pressure melting. Net basal melting is caused by:

1. Geothermal heat.
2. Heat released by the friction of basal sliding.
3. Heat released by the friction of water flow (Chapter 1.1).

Hubbard and Sharp (1993) assume that a) geothermal heat melts 6 mm of basal ice per year, and, b) sliding at 20 m yr^{-1} generates heat which melts 6 mm of ice per year also. If these figures were to apply to Sólheimajökull's subglacial regime, then running water must melt

$$(\sim 6 \text{ cm} - \sim 1 \text{ cm}) = \sim 5 \text{ cm}$$

to make up the shortfall in basal ice melt that the quantity of debris carried by the Jókulsá implies. However, the values used by Hubbard and Sharp are probably conservative in the case of Sólheimajökull. If a) the geothermal heat flux is twice the typical value (this is not unlikely), and, b) sliding takes place at ~100 m yr^{-1} (Chapter 2.1), then the thickness of basal ice which running water must melt falls to:

$$[\sim 6 \text{ cm} - (\sim 1 \text{ cm} + \sim 3 \text{ cm})] = \sim 2 \text{ cm}$$

Thus I estimate that water flow must melt something like **1×10^6 m^3 to 2.5×10^6 m^3** of basal ice every year.

Energy released by water flow

Here I calculate an independent estimate of the likely quantity of basal ice which running water melts each year. For any glacier in equilibrium, water equivalent inputs must equal water equivalent outputs, so I estimate annual run-off at Sólheimajökull to be:

$$[(1.6 \text{ m} + 4.0 \text{ m}) / 2] \times 50 \times 10^6 \text{ m}^2 = 140 \times 10^6 \text{ m}^3$$

[1.6 m yr^{-1} and 4.0 m yr^{-1} are Rist's minimum and maximum annual precipitation totals for the Jókulsá catchment (Rist, 1957, cited by Lawler, 1991).]

The accumulation area occupies 33 km^2; if mean annual *net* accumulation is 2.0 m water equivalent, then total annual runoff from the accumulation area is:

$$33 \times 10^6 \text{ m}^2 \times (2.8 - 2.0 \text{ m}) = 26.4 \times 10^6 \text{ m}^3 = 2.64 \times 10^{10} \text{ kg}$$

which must make the annual run-off from the ablation area:

$$(140 \times 10^6 \text{ m}^3 - 26.4 \times 10^6 \text{ m}^3) = 113.6 \times 10^6 \text{ m}^3 = 1.136 \times 10^{11} \text{ kg}$$

(PTO)

Box 2.2 (continued)

I estimate the mean height lost by run-off from the accumulation area to be:

$$[(1{,}500 \text{ m} + 1{,}100 \text{ m}) / 2] - 100\text{m} = 1{,}200 \text{ m}$$

and the mean height lost by run-off from the ablation area to be:

$$[(1{,}100 \text{ m} + 100 \text{ m}) / 2] - 100 \text{ m} = 500 \text{ m}$$

(glacier summit = ~1,500 m asl; ELA = ~1,100m asl; glacier snout = ~100 m asl)

Thus the total annual energy released by running water works out as:

$$9.81 \text{ m s}^{-2} \times [(2.64 \times 10^{10} \text{ kg} \times 1{,}200 \text{ m}) + (1.136 \times 10^{11} \text{ kg} \times 500 \text{ m})]$$

$$= 8.68 \times 10^{14} \text{ J}$$

Part of this energy will be taken up by sediment transport; perhaps as much as 20% of the total [cf. Bagnold (1966), cited by Richards (1982, p. 116)]. (If flow occurs under closed conditions, then some energy must also be used to warm up water as its pressure melting point falls, although this quantity is likely to be small if the overall hydraulic gradient is steep.) What energy is left melts ice:

$$6.94 \times 10^{14} \text{ J} / 3.35 \times 10^{5} \text{ J} \text{ (i.e. latent heat of fusion)}$$

$$= \sim 2.1 \times 10^{9} \text{ kg} = 2.3 \times 10^{6} \text{ m}^{3}$$

However, ice melt will not be confined to the glacier bed: perhaps as much as half of total melt induced by water consumes (debris-free) surficial and englacial ice. Thus by this method I estimate that running water melts something like **1.15×10^{6} m^{3} to 2.3×10^{6} m^{3}** of basal (i.e. debris-rich) ice per year at Sólheimajökull.

EVALUATION

These calculations provide two independent estimates of the quantity of basal ice which melts below Sólheimajökull every year:

1) Derived from Lawler's measure of sediment output: **1.00 to 2.5×10^{6} m^{3} yr^{-1}**

2) Derived from estimates of annual run-off: **1.15 to 2.3×10^{6} m^{3} yr^{-1}**

These correspond to destruction of a basal ice layer with a mean thickness of 0.02 to 0.05 m. Hubbard and Sharp suggest that basal ice layers formed and reformed beneath the middle of glaciers which exceed 0.05 m in thickness are rare. Taken with this point, I feel that these figures support the inference (made initially on the basis of field observations of the glacier margin) that that vast majority of basal ice which forms below Sólheimajökull is destroyed by net basal melting, and its debris washed away.

ERRORS?

The reliability of these 'best-guess' calculations is not known. Below I present data which work with an arbitrary ±25% margin of error (which I trust is sufficiently wide to incorporate the true field values). This error band is applied to each variable in turn, so calculations incorporate 'carry-over' error. This provides maximum (HIGH) and minimum (LOW) estimates for likely values of basal ice melt calculated by the two different methods.

(PTO)

Box 2.2 (continued)

SÓLHEIMAJÖKULL: ANNUAL DESTRUCTION OF BASAL ICE Error evaluation (±25%)				
Variable	**Units**	**HIGH**	**BEST-GUESS**	**LOW**
Quantity of basal ice destroyed				
Debris output of Jökulsá	t	1,050,000	850,000	640,000
Proportion of debris from basal ice	%	75	50	25
Mass of debris from basal ice	kg	788,000	425,000	160,000
Debris content of basal ice by mass	%	9.75	13.00	16.25
Debris content of basal ice by volume	%	3.5	4.7	6.1
Density of basal ice	kg m^-3	962	984	1010
Volume of basal ice destroyed	m^3	8.4 x 10exp6	3.3 x 10exp6	9.8 x 10exp5
Of which actually ice (i.e. debris excluded)	m^3	8.1 x 10exp6	3.1 x 10exp6	9.2 x 10exp6
Total thickness of ice destroyed	m	0.16	0.06	0.02
Thickness destroyed by friction + geothermal heat	m	0.01	0.01-0.04	0.02
Thickness which running water must melt	m	0.15	0.02-0.05	0.00
Volume which running water must melt	**m^3**	**7.5 x 10exp6**	**1-2.5 x 10exp6**	**0.00**
Melt energy released by running water				
Annual runoff	m^3	175 x 10exp6	140 x 10exp6	105 x 10exp6
Runoff from accumulation area	kg	6.6 x 10exp10	2.6 x 10exp10	0
Runoff from ablation area	kg	1.09 x 10exp11	1.14 x 10exp11	1.05 x 10exp11
Total energy release	J	1.31 x 10exp15	8.68 x 10exp14	5.15 x10exp14
Energy used for sediment transport	%	10	20	20
Energy used for ice melt	J	1.18 x 10exp15	6.94 x 10exp14	4.12 x 10exp14
Propotion of ice melt at bed	%	100	50-100	50
Volume of ice melt at bed	**m^3**	**3.9 x 10 exp6**	**1.15-2.30 x 10exp6**	**6.8 x 10exp5**

The 'worst-case' scenario (i.e. that which is least consistent with my hypothesis!) implies that running water must melt $\mathbf{7.5 \times 10^6\ m^3}$ of basal ice, whereas it is, in fact, capable of melting only $\mathbf{6.8 \times 10^5\ m^3}$ of basal ice: i.e. a order-of-magnitude discrepancy exists (see shaded cells in the table above). However, I do not think that this provides sufficient evidence to refute the idea that most basal ice which forms at Sólheimajökull is destroyed, much of it by water action:

- This *is* the 'worst-case' scenario, which involves extreme estimates.
- This 'worst-case' scenario involves a basal ice layer which *on average* is 16 cm thick. The work of Hubbard and Sharp (1993) shows that this is unlikely.
- Within the glaciological tradition it is common practice to consider calculations which match to an order-of-magnitude to represent a good fit!

hours, whereas ~12 hours are required for melt-out of a 2.0 cm clast (smaller clasts - <~1.40 cm - should be lost, however).

These calculations are crude (for one thing, the channel geometries, roughnesses and discharges used are idealised), but they do cover most possibilities, and so provide some insight. Perhaps of most significance is what they imply about the relative strength of basal-ice-melt flushing in conduit and cavity networks. I expect this to be high, but spatially-restricted, in conduits, but low with respect to the *single* cavity, which carries low power flow. This discrepancy is likely to be enhanced if cavities encourage fast sliding. However, this low intensity flushing tendency must be offset against:

1. The fact that hydraulic gradients are high in orifices (but see my comments on sediment in orifices above).
2. The total number of contacts between water and basal ice in a linked-cavity system is likely to be high because of a) the large number of individual passageways, and, b) the fact that cavities and orifices, unlike conduits, tend to be aligned transverse to ice flow.

Some indication of the flushing potential likely to be associated with cavity networks/slow sliding speeds is given by the contrasts in basal ice seen at Glacier de Tsanfleuron and Glacier de Ferpècle, Switzerland. No laminated facies basal ice is found at Tsanfleuron, which has a (relatively non-sinuous, efficient) cavity network (but also inferred low water pressures, with relatively slow sliding: Sharp *et al.*, 1989a), whereas Glacier de Ferpècle has well developed marginal exposures of laminated basal ice, and an impressive network of deep, straight Nye-channels (Hubbard and Sharp, 1995; my observations, summer 1990). The rate at which partial melt-out of clasts is recovered between channel intersections by the bed-normal component of ice flow (cf. Hallet, 1979a) is likely to be important, because this will influence a clast's *cumulative* vulnerability to flushing. Full clast enclosure is impaired by high levels of basal debris (clasts impede regelation and plastic deformation: Iverson, 1993). This illustrates just how complex the phenomenon of subglacial flushing is: i.e. susceptibilty of basal debris to flushing depends upon the concentration of debris in basal ice, but this itself is determined in part by the efficiency with which subglacial flushing proceeds!

EXISTING THEORY: THE WORK OF D. N. COLLINS

Work by Collins (most of it at Gornergletscher, Kanton Wallis, Switzerland) on the subglacial interaction of water, ice and sediment, as revealed by the changing pattern of suspended sediment deliveries to the ice front, provides the most important collection of ideas in print relevant to a comprehensive theory of subglacial flushing (Collins, 1979a, 1988, 1989, 1995b, 1996). I use these ideas as a basis for my own work (see below). The five key features of Collins' work are:

1. Glacier bed model: more-or-less continuous layer of potentially deformable till. At Gornergletscher, this layer of 'ground moraine' seems to vary between zero and several metres thick (Röthlisberger, 1972; Iken *et al.*, 1996).

2. Drainage model: the location of the major arterial channels (conduits) is pretty-much fixed, although channel size, perhaps also channel sinuosity, fluctuates; these larger channels are possibly incised into bedrock. Seepage through ground moraine, and water flow through tributary cavities and smaller channels feeds water to these arterial conduits. Tributaries are mobile, and cut upwards into ice, and downwards into sediments (cf. Knight, 1992, p. 88; Walder and Fowler, 1994). This picture matches the largely independent conclusions of Iken *et al.* (1996) regarding the drainage structure beneath Gornergletscher:

   > "... [A] poorly interconnected subglacial drainage system near the centre line. Probably a few large drainage channels divert meltwater from moulins, and collect water from widespread sediment layers and temporarily shrinking cavities. [However] near the margin... drainage courses were discrete passageways, presumably linked cavities on clean bedrock, and were numerous and durable."
   >
   > Iken *et al.* (1996, pp. 244-245)

   The extent to which the subglacial drainage network of Sólheimajökull resembles this description is unknown.

3. Migration of smaller channels is temporally frequent, and spatially widespread. Channel migration occurs: a) as water melts channel walls; b) as bank sediments erode; and, c) as sliding ice carries channels over the glacier bed. It is difficult to differentiate between channel migration, and reorganisation of drainage in terms of a change in the number, size or shape (cross-section or plan-form) of channels. This occurs at a wide range of scales, usually related to channel blockage by ice or sediments, or in response to diurnal or seasonal fluctuations in discharge.

4. Flood events at a wide range of scales are important elements of drainage reorganisation. Enhanced episodes of melt, rainstorms and lake drainage all generate flood events.

5. Collapse of moraine-channel banks, and deformation of sediment into conduits account for a large part of sediment delivery to channels. These mechanisms are believed to be particularly important in-between flood events.

COLLINS' PURGE-RECHARGE MODEL

Collins brings the core of these ideas together in his simple conceptual model of the interaction of subglacial sediments and meltwaters (Collins, 1995b, 1996). This represents the first serious attempt to develop a theory of subglacial flushing: i.e. it specifies the events and processes responsible for the evacuation of *suspended* sediments. It is a purge-recharge model, intended to replicate what happens at Gornergletscher. Suspended sediment delivery to the glacier snout is primarily controlled by drainage development/reorganisation associated with subglacial flood events. The time series of suspended sediment flux reflects the changing percentage area of the bed accessed by meltwaters (this rises quickly under flood conditions), and exhaustion of sediment supplies in the aftermath of such events.

Suspended sediment flux at Gornergletscher.

This can be split into two components (Figure 2.11):

1. Background flux, sustained throughout the melt season (late May to late September). This background flux rises sharply in the aftermath of the 'spring event', and thereafter tends to decline gradually.

2. Major spikes of suspended sediment delivery - associated with a rise in discharge - are thought to relate to subglacial flood events during which competent/capable waters spread out to cover an enlarged proportion of the glacier bed, so gaining access to fresh stores of sediment which are swept away. The first of these flood events is the 'spring event' (Röthlisberger and Lang, 1987, pp. 261-266), which represents the 'breakout' of waters stored in subglacial pockets once melt rates rise. Water sweeps across wide areas of the glacier bed as it escapes, initiating an integrated, efficient drainage network in the process. Later flood events reflect rainstorms, episodes of enhanced surface melt, and drainage of the Gornersee (an ice-marginal lake trapped at the confluence of Gornergletscher and Grenzgletscher, its chief tributary). The suspended sediment spikes are short-lived: sediment flux falls sharply, despite sustained levels of high discharge in many cases. This implies widespread exhaustion of sediment supplies.

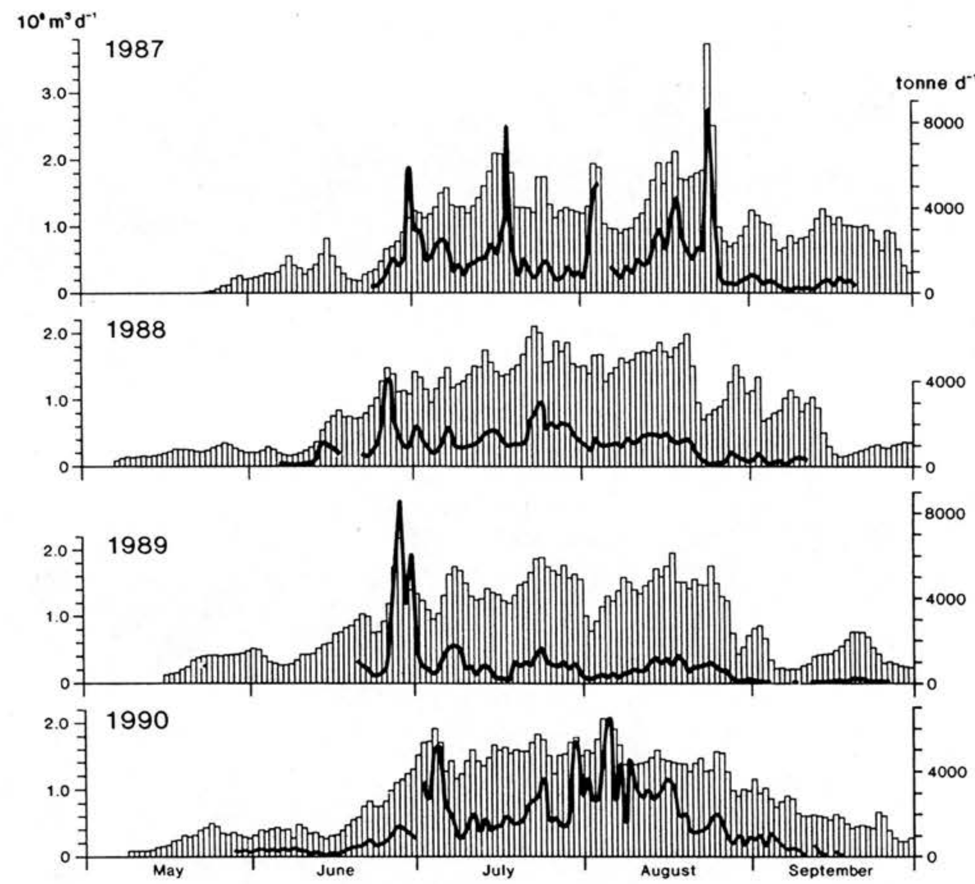


Figure 2.11

Daily total discharge (columns) and measured daily total suspended sediment flux (curves) in the Gornera, Valais, Switzerland, in the months May to September, 1987-1990 (from Collins, 1996).

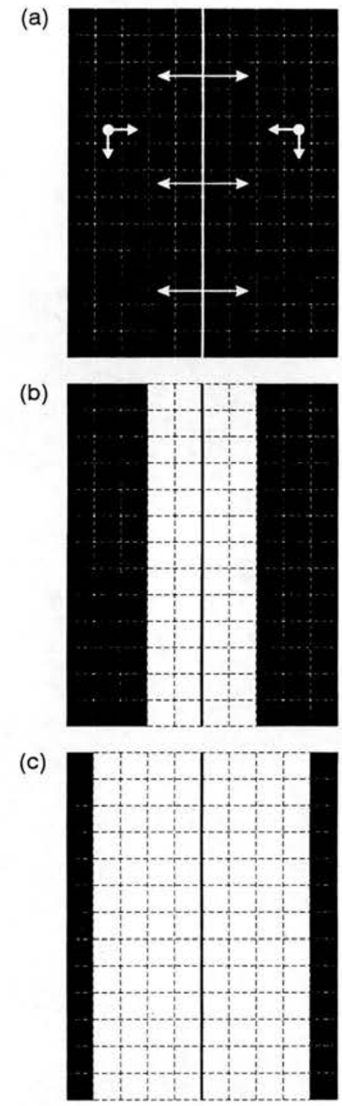


Fig. 3. Plan view of grid-square lattice representing the subsole of a valley glacier, showing the direction in which the total wetted area expands equally on both sides of the thalweg with increasing discharge (long arrows), and longitudinal and transverse directions in which stored sediment is deformed (short arrows) in the model. Shaded areas indicate presence of sediment storage at the subsole, the area retaining sediment declining as increases in discharge expand the wetted area between the end of winter (a), through spring (b) to maximum flow in summer (c).

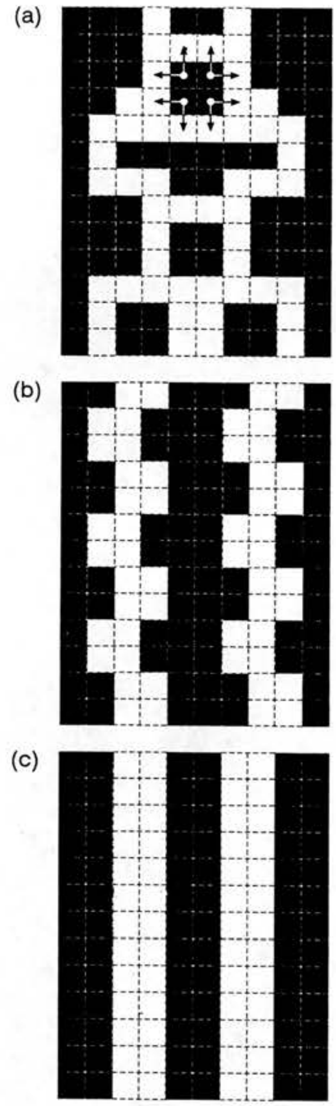


Fig. 7. Various plan configurations in which water covers 40% of the area of grid lattice. Shaded areas indicate presence of sediment storage at the subsole. Total length of channel margin decreases from anastomosing (a) through crenulated or meandering (b) to straight (c) channel patterns. The arrows indicate the directions in which sediment deformation into water occurs if squeezing occurs across all boundaries of cells adjacent to flowing water.

Figure 2.12 (left)

The Collins model: glacier bed and basic scheme of drainage reorganisation (from Collins, 1996).

Figure 2.13 (right)

Different styles of subglacial drainage, as incorporated into the Collins model (from Collins, 1996). The switch from a) to c) by way of b) simulates seasonal reorganisation of drainage.

Features of Collins' model

- The glacier bed consists of a simple rectangular grid. Cells which make up this grid are designated 'wet' or 'dry' (my terms). Wet cells are occupied by filaments of discrete drainage, competent and capable of sediment evacuation. Dry cells contain residual drainage, which is incompetent/incapable. The exact type of drainage is not specified, but pore-water, and/or film flow is implied/expected. Within dry cells: a) abrasion continuously creates fresh supplies of sediment, and, b) continuous deformation of sediments occurs. This carries sediment i) towards the glacier snout, and, ii) towards the drainage filaments (i.e. low pressure axes) which occupy the wet cells.

- All sediment is instantaneously purged from cells which are designated wet.

- Sediment recharge occurs in dry cells. Recharge is equal to [(daily rate of abrasion + daily inputs of deforming sediments) - daily loss of deforming sediments] x days which cell stays dry. Continuous accumulation is expected if the cell stays dry because it is assumed that abrasion rates >> deformation rates.

- The background flux of suspended sediments is sustained by deliveries by deformation to wet cells which evacuate sediments. The quantity of sediment supplied to wet cells is proportional to the total perimeter of wet cells: i.e. the total sinuosity of filaments (Richards, 1982, Fig. 1.4, p. 10, and p. 182).

- The input which runs the model is the discharge time series of the Gornera (i.e. the outlet of Gornergletscher). Winter discharge is negligible, so all cells are designated dry, and sediment accumulates throughout late autumn, winter and spring over all the glacier bed. Wet cells first appear, and flushing begins, as discharge rises to significant levels in late spring.

- Sediment spikes of flood events represent lateral reorganisation of drainage in response to elevated water inputs (Figure 2.12). Flow expands sideways from its central axis to cover an increased area of the glacier bed, whereupon dry cells are converted to wet cells. The number of cell conversions/additional area of the bed incorporated by the flood event is calculated by a discharge-bed coverage rating curve which assumes 90% bed coverage at maximum discharge. The suspended sediment flux at each time step represents the number of wet cells x the quantity of sediment accumulated in each cell.

- The size of each suspended sediment spike reflects bed coverage, and the time elapsed since each cell tapped was last occupied by filament drainage. This incorporates supply limitation/sediment exhaustion effects. Successively higher flood levels, expected to cover successively greater areas of the glacier bed, produce major sediment spikes, most obviously the spring event. (Thus the efficiency of flushing varies considerably over the melt season. However, this within-season variability is unlikely to have a major impact on 'big picture' debris transport, sediment budgets and moraine building activity, which relate to several seasons' cumulative behaviour: debris transport by ice is ~4 orders-of-magnitude slower than debris transport by water. The interval between major subglacial flood events as envisaged by Collins is unlikely to allow sufficient time for most parcels of subglacial debris to escape to the ice edge. See below.)

- The basic model uses flow expansion of a single, straight, central filament (i.e. a master conduit located at the base of the subglacial trough). It can be modified to include some element of seasonal drainage reorganisation/rationalisation by allowing the total sinuosity (i.e. number of channels x channel 'wiggliness') to vary as the melt season proceeds. Progressive rationalisation of drainage (e.g. Richards *et al.*, 1996) can be simulated by starting with a wiggly, anatomising network, which eventually is replaced by a single straight conduit. The effect of this is to increase the background sediment flux in the early part of the melt season (high total length channel perimeter) and let it fall thereafter: see Figure 2.13.

Model performance

The model successfully reproduces the *timing* of major sediment spikes recorded in the Gornera in summer 1987, although it is less accurate in its prediction of the absolute and relative *magnitudes* of those peaks. Nevertheless, this seems to confirm the importance of a) time elapsed since last flood of a given magnitude, and, b) flood level excess over the previous event as major controls on the (within-melt-season) space-time impact of flushing. However, the concept of the background sediment flux upheld by sediment deformation is not reproduced; Collins believes that this reflects the difficulties of choosing appropriate parameters (e.g. deformation rate, length of channel perimeter which receives sediment) to represent a complex three-dimensional process which is highly variable in space and time.

A CRITICAL ASSESSMENT OF COLLINS' MODEL

Collins' conceptual model seems to be fundamentally sound. It illustrates the basic relationship central to my thesis, namely that an aggressive subglacial drainage network which has repeated access to the majority of the glacier bed throughout the melt season will tend to be extremely efficient at removing sediment of subglacial origin which accumulates at the bed. The grid cell model structure implies a simple conceptual distinction between 'wet' and 'dry' bedded glaciers. If over the relevant time-scale the majority of a glacier's bed at some time is 'wet', it is likely that the bulk of sediments are flushed, with moraine-building potential diminished correspondingly. The relevant time-scale is crudely defined by:

$$\mathbf{MCTPD / MSDT = MCRT}$$

MCTPD: mean clast transport path distance between points of origin and the edge of the ice

MSDT: mean speed of debris transport by ice and/or bed deformation

MCRT: mean clast residence time if flushing does not take place

E.g. if the mean clast flow-path length is 2.0 km, and mean clast speed is 50 m yr^{-1}, then the relevant property which determines bulk flushing efficiency is likely to be the proportion of the total bed area which is 'wet' at some time in every **(2,000 m / 50 m yr^{-1}) = 40 years.** This kind of crude calculation, which relies on mean values, cannot exclude 'extreme' behaviour: in this case, the fact that clasts which originate close to the ice edge stand a much greater chance of survival - particularly if drainage favours the centre of the glacier. However, this does provide valuable insight because (as I infer in sections 2.3 and 2.4) it implies that flushing is likely to dominate the subglacial sediment budget: e.g. if Collins' guess that the one-year flood typically covers 90% of the bed with 'wet' drainage is correct, then the probability that this 90% p.a. coverage will include ~100% of the bed area if repeated every year for 40 years is high. Debris which originates close to the glacier centre is likely to be flushed away with a much greater frequency.

By necessity, Collins' model makes several simplifications/omissions, as he admits (1996, p. 231). A full description of the subglacial sediment budget requires detailed knowledge of the magnitude-frequency characteristics, and spatial distribution of subglacial erosion processes (which create debris), and sediment deformation processes (which redistribute debris). These must be balanced against the space-time entrainment capabilities of changing styles of subglacial drainage. All these are (inevitably) subject to severe parameterisation in Collins' model. Plucking, for instance, is excluded, as is debris entrainment into, and transport within, basal ice. Instantaneous flushing of a wet cell is a reasonable assumption for fine sediments (but this ignores factors such as bed armouring), but it is is unrealistic for coarse debris, as is likely to be produced by plucking. These are minor deficiencies, not serious objections to the basic model form. Three inter-related points, all central to a full theory of flushing, merit further discussion. These are:

A. To what extent is sediment carried to wet cells, as opposed to changes in the distribution of wet cells tapping fresh sources of sediment? This raises questions about the general relevance of Collins' bed model, and the potential role of subglacial sediment deformation.

B. To what extent is the event-based nature of Collins' model applicable to other glaciers?

C. What exactly is involved when a cell changes from being 'dry' to being 'wet'?

I take each of these points in turn, and evaluate it as a step towards a refined model of subglacial flushing (implicitly a model which works well for the case of Sólheimajökull).

A) COLLINS' BED MODEL AND PROCESSES OF SEDIMENT DELIVERY

Storage and deformation of sediments across wide areas of a glacier's bed will not necessarily be a suitable bed model for all glaciers: it may fit Gornergletscher, but this does not mean it is appropriate for, say, Sólheimajökull. Of potentially greater importance, however, is the objection that Collins' view of sediment delivery processes is possibly inappropriate for all valley glaciers, Gornergletscher included. The failure of Collins' model to reproduce satisfactorily the background sediment flux perhaps relates not to poor choice of relevant parameters, but to deficiencies in model conceptualisation. The role of till deformation as a major mechanism by which sediment is supplied to channels (it must account for half or more of the sediment output of the Gornera if Collins is correct) has yet to be confirmed (see Ó Cofaigh, 1996, for a similarly sceptical analysis of the sediment deformation and removal hypothesis as explanation for the origin of tunnel valleys). Collins justifies his ideas by appeal to Walder and Fowler's (1994) theory of channelised subglacial drainage with a deformable bed. This justifies the principle of sediment delivery to channels, but - if my reading of Walder and Fowler is correct - it is by no means safe to conclude that it must be important at Gornergletscher and similar valley glaciers, of which Sólheimajökull is one.

Fowler and Walder's ideas

Fowler and Walder's (1994) theory stems from the inference that channel closure by ice creep is likely to be fastest if water pressures are low, whereas channel closure by till creep is likely to be fastest if water pressures are high. This implies the existence of a critical effective pressure (at which the creep closure rates of ice and till are equal) which determines the choice of drainage style (see Chapter 1.1). If effective pressure exceeds this critical pressure (i.e. low water pressures) a Röthlisberger-type conduit, incised upwards into ice, with a stiff till floor, is expected. This does not favour till deformation. Fowler and Walder estimate that this critical pressure is $\sim 8 \times 10^5$ Pa (it will be lower if till is stiff). If ice is ~300-400 m thick, as it is at both Gornergletscher and Sólheimajökull, then this represents water pressures equal to ~70-80% of overburden. Such pressures are expected, but are not likely to exist as a stable feature in the vicinity of large conduits capable of transporting large quantities of sediment (cf. Röthlisberger *et al.*, 1979; Hubbard *et al.*, 1995). If the hydraulic gradient/ice surface slope is ~0.1 (again, as is typical of Gornergletscher and Sólheimajökull) both canals (into which till deforms readily) and conduits are a stable choice of drainage *individually*, but canals are not expected to be stable in the presence of conduits. This follows from the inference that a) the discharge-water pressure relationship for canals is positive, but it is inverse for conduits; and, b) the critical pressure is met at lower discharge in conduits (Walder and Fowler, 1994, their Figs 2 and 4). If a nascent channel - which originates as a film instability - starts out cut equally into ice and till, water pressures will fall below the critical pressure as it grows in size, instigating fully-fledged conduit development, with its progressive fall in water pressures. Many other conduits below valley glaciers originate at the surface, and so are fully established by the time their water reaches the glacier bed. If conduits (low pressure) and canals (high pressure) exist side-by-side, conduits should gather water by seepage through till at the expense of neighbouring canals. As a result, conduit water pressures fall, which reinforces their tendency to capture flow; till pore-water pressures also fall, but this works to hinder till deformation, and so favours further conduit drainage. These theoretical arguments imply that sediment delivery to channels by till deformation is likely to be a short-lived, spatially-restricted phenomenon only - perhaps important as water pressures rise widely across the bed prior to the spring event (or within overdeepenings: see Chapters 5.6 and 6.2). This suggests that sediment delivery to channels by till deformation fails to make an important contribution to flushing.

Alley's ideas

If a canal-type network exists, whether as a stable or as a transient feature, its ability to capture sediment is questionable. Alley (1992) uses a simple Coulomb-type force-balance analysis in which till behaves as a perfectly plastic material to argue that the till catchment zone surrounding channels is restricted to perhaps 5-10 times the channel depth: i.e. a channel 1.0 m deep (which is large!) is likely to capture till only from the adjacent 10 m of bed at best; beyond this distance, till strength exceeds the stress (ice pressure minus channel water pressure) driving deformation. This conclusion is robust if alternative models of till rheology - Bingham fluid or viscous fluid - are used. Effective pressures at the channel margin are zero (i.e. till pore-water pressure = channel water pressure) so till deformation here is quickest. Pinching-out of till, and clast interaction factors (i.e. clasts are large relative to the thickness of the till layer) create a barrier to till deformation which tends rapidly to isolate channels. Alley concludes that pervasive till flux to channels requires spacing of channels at intervals $<\sim 10$ m, and/or high channel mobility. He implies that the first condition is unlikely - particularly if these channels are to be competent and capable. Migration of channels is conceptually a different process - i.e. in this event, the channel goes to the sediment - but it is worth remembering that migration of channels cut into sediment requires differential bank erosion. If Walder and Fowler's conclusion regarding the likely geometry (i.e. incised upwards into ice) and behaviour of till-floored channels beneath valley glaciers is correct, then the area of channel bank sediments exposed to erosion is extremely limited. Migration of 'classic' R-channels must proceed largely by differential ice melt (see below). Unless this ice is dirty, migration will release limited quantities of debris only.

Deformation of till into high pressure channels?

Channel water pressure relationships also imply that till deformation into channels is limited. High channel water pressures suppress till deformation. High pressures are expected as a steady-state feature of small channels (Röthlisberger, 1972), which are likely to occupy a larger bed area, with longer total channel perimeter, than larger channels. However, high water pressures are expected also as transient features of larger channels (cf. Hubbard *et al.*, 1995). In this case channel water pressures can exceed the pressures acting on the adjacent bed area, so that water is pumped out of the channel, rather than sediment being pushed into it (2.2, above). High water pressure events can decouple ice from till, causing a sharp reduction in rates of till deformation (Iverson *et al.*, 1995; Ó Cofaigh, 1996). As Hubbard *et al.*'s study also illustrates, return flow in response to a subsequent drop in conduit water pressures can winnow fine sediment from till in the zone immediately adjacent to the conduit (i.e. the zone in which Alley calculates till deformation is a realistic prospect). Selective removal of fines will create a coarse residual till, which is likely to have greater shear strength, and so reduced propensity to deform.

Deformation of till into low pressure channels?

There are reasons to suspect that low pressure channels work against till deformation as well. The potential impact of the Weertman pressure-dam/ice bridging effect has yet to be resolved. If channel water pressures are low, the overlying ice is not fully-supported, and so the excess weight is transferred to the glacier bed to the immediate sides of the channel (Weertman and Birchfield, 1983; see Chapter 6.4). This implies that the zone in which the bed pressure gradient forces till and water towards channels is severely restricted; Weertman argues that bed pressures in excess of overburden reverse the pressure gradient a short distance (~one channel radius) beyond the channel, thereby isolating it from the rest of the bed. Walder and Fowler dismiss this effect because strictly it applies to parallel, straight channels on a rigid bed. However, it is difficult to see why they consider a porous bed to be less vulnerable to pressure bridging effects. The transfer of weight - even if less marked than in the ideal case[5] - should increase the strength of till close to low pressure channels, and so *suppress* deformation, even if it fails to reverse the direction in which till deforms. Alley (1992) excludes the pressure dam effect from his analysis, but he does acknowledge its potential importance. Support for its impact is given by Boulton's (1976) study of flutes (i.e. till squeezed into a tunnel-cavity). The zone which feeds sediment into the cavity to form the flute is restricted by pressure dams flanking the cavity. Troughs either side of the flute indicate where sediment was squeezed out, and into the cavity, from immediately beneath the zone of maximum pressure; however, these troughs are flanked by subsidiary anticlines which indicate squeezing of till *away* from the cavity in response to local reversal of the pressure gradient.

Verdict

The contribution of soft-bed hydrology/till deformation to the overall flushing process cannot be confirmed. Although deformation of sediments beneath valley glaciers is a proven fact (thanks to borehole instruments deployed at Trapridge Glacier, Yukon Territory, Canada, and Storglaciären, Sweden), its interaction with subglacial channels has yet to be demonstrated satisfactorily. Because of this, plus the many theoretical objections and uncertainties I have just reviewed, and the basic point that subglacial tills are not necessarily a ubiquitous feature of all valley glaciers, I think it is best to explore alternative processes likely to contribute to flushing.

B) ADEQUACY OF A FLOOD EVENT MODEL?

Collins' model is strongly event-based. Subglacial floods of relatively high-magnitude but low frequency are thought to account for a large part of the total sediment flux. The basic premise, namely that unusually high flows permit water to access fresh sediment stores, is sound: e.g. Lawler (1991) reports that a doubling of discharge in the Jökulsá (August 1988) was associated with a twenty-five fold increase in suspended sediment concentrations. However, the exact balance between flood events and 'normal' flows/extended periods of supply limitation, is not clear. This in part reflects the fact that exactly what is entailed by the discharge peak/flood event is unclear; indeed, it is unknown! (see below). Sólheimajökull for example, is not presently subject to regular floods as ice-dammed lakes drain: the events described by Knight and Tweed (1991) are insignificant (Lawler, 1994), whereas the Jökulsárgil valley last saw the presence/drainage of a sizeable lake in the 1930s (Tweed, 1992). Subglacial drainage of the Gornersee usually takes place once every year. Discharge and sediment flux at Gornergletscher

[5] Hubbard *et al.*'s 1995 study indicates flow of *water* into low pressure conduits, which implies that the pressure dam seal is far from perfect.

respond to a specific Alpine-type climatic regime (run-off between late May and late September; distinct spring event, etc.) which contrasts sharply with the regime at Sólheimajökull. Features of this regime which differ from those expected of a typical Alpine glacier include (Lawler, 1991):

- 38% of total flow occurs between October and March.
- There is no clear spring event.
- Really high flows are restricted to late summer and early autumn.
- The second highest suspended sediment concentration recorded between 1973 and 1988 occurred in October, the fifth highest in November, and the sixth highest in February.

Different glaciers, with different regimes, are likely to have different propensities for flood events (in part because the run-off regime will determine what qualifies as an unusually high flow), and are likely to have different drainage structures, likely to respond to high discharge conditions in different ways, with different implications for flushing of sediments. Similarly, factors such as basin geometry, relief/hypsometry, percentage area bare rock and percentage area snow cover are likely to influence the impact rainstorm events have on the style and flushing impact of subglacial drainage. Steep bare-rock slopes and small percentage snow cover (i.e. as typical of an Alpine glacier basin) are likely to encourage rapid flow peaks (cf. Collins, 1995a), whereas limited rock slopes and large, high-elevation areas of relatively gentle snow-pack, which can store water like a sponge (as at Sólheimajökull?), will tend to dampen rainfall-induced flood flows.

Haut Glacier d'Arolla and Gornergletscher: a comparison of flow regime and sediment output

Studies of suspended sediment delivery at the Haut Glacier d'Arolla (Richards *et al.*, 1996; Clifford *et al.*, 1995; Gurnell, 1994) provide an illuminating comparison with studies at Gornergletscher. Whereas the two glaciers share many characteristics[6] the Arolla data seem to display features which differ from the water/sediment behaviour at Gorner in important respects. These data suggest alternative process regimes which favour efficient flushing, regimes which perhaps apply also to Sólheimajökull and other glaciers.

As with Gornergletscher, changes in suspended sediment concentration and flux at Arolla tend to reflect drainage reorganisation. Arolla too has its spring event (25 June in 1990);

[6] This is to be expected because - despite the fact that Gorner is much bigger - the two occupy similar Alpine catchments, separated by just 20 km.

thereafter drainage evolves largely by headwards extension of a conduit system as the snow-line retreats upglacier, and melt-induced discharge rises. This mode of drainage reorganisation (i.e. headwards expansion) is different to that envisaged by Collins, in which progressive rationalisation of drainage plays a relatively minor role, with *lateral* reorganisation (i.e. flood flow expansion) identified as the factor which determines sediment outputs. This raises questions as to whether or not Collins' view of drainage reorganisation is correct/sufficient: unlike Arolla, little direct evidence is available of drainage system change at Gornergletscher. However, this is not my main point here: characteristics such as rising discharge with temporary flow peaks, negative water balance, changing area of the bed tapped by competent/capable drainage, and evidence of supply exhaustion are common to interpretations of sediment flux at both glaciers. It is their late-season behaviour which seems to differentiate between the two. Whether or not this is a genuine difference in behaviour is unclear because of ambiguities in the data, differences in the quality of data, and differences in the presentation of data, but the interpretations of water sediment interactions attached are different, which in turn implies alternative flushing regimes.

Maximum suspended sediment concentrations at Arolla occur in mid-to-late August (Fig. 2.14), associated with the maximum area of ice exposed, and the fullest extent of the developed conduit network (a residual distributed system is believed to exist between conduits, and at the head of the basin beneath snow and firn). Suspended sediment concentrations here scale with consistently high discharge: the discharge-suspended sediment concentration regression relationship gives a relatively high R^2 of 78%, which implies overall limited scatter/hysteresis, and a relative absence of sediment exhaustion. Consistently high, positive residuals in August show that suspended sediment concentrations are higher than might be expected for all levels of discharge. If data for mid-to-late August only are examined, suspended sediment peaks are found to match closely diurnal discharge, and the fit of the linear regression relationship improves further. It is not clear if this increase in sediment concentration translates into an increase in sediment flux, because discharge tends to fall in August as solar radiation receipts fall. However, the discharge data in Richards *et al.* (1996) and the suspended sediment data in Clifford *et al.* (1995) seem to show that late season sediment flux stays high even as bulk discharge levels start to decline, an inference supported by Gurnell (1994): in 1989, mean daily sediment yields in August were 156% of July yields, whereas in August mean daily discharge was just 86% of its July equivalent.

The contrast here with Gornergletscher is interesting. August sediment flux at Gornergletscher tends to tail off, reflecting both widespread sediment exhaustion and falling

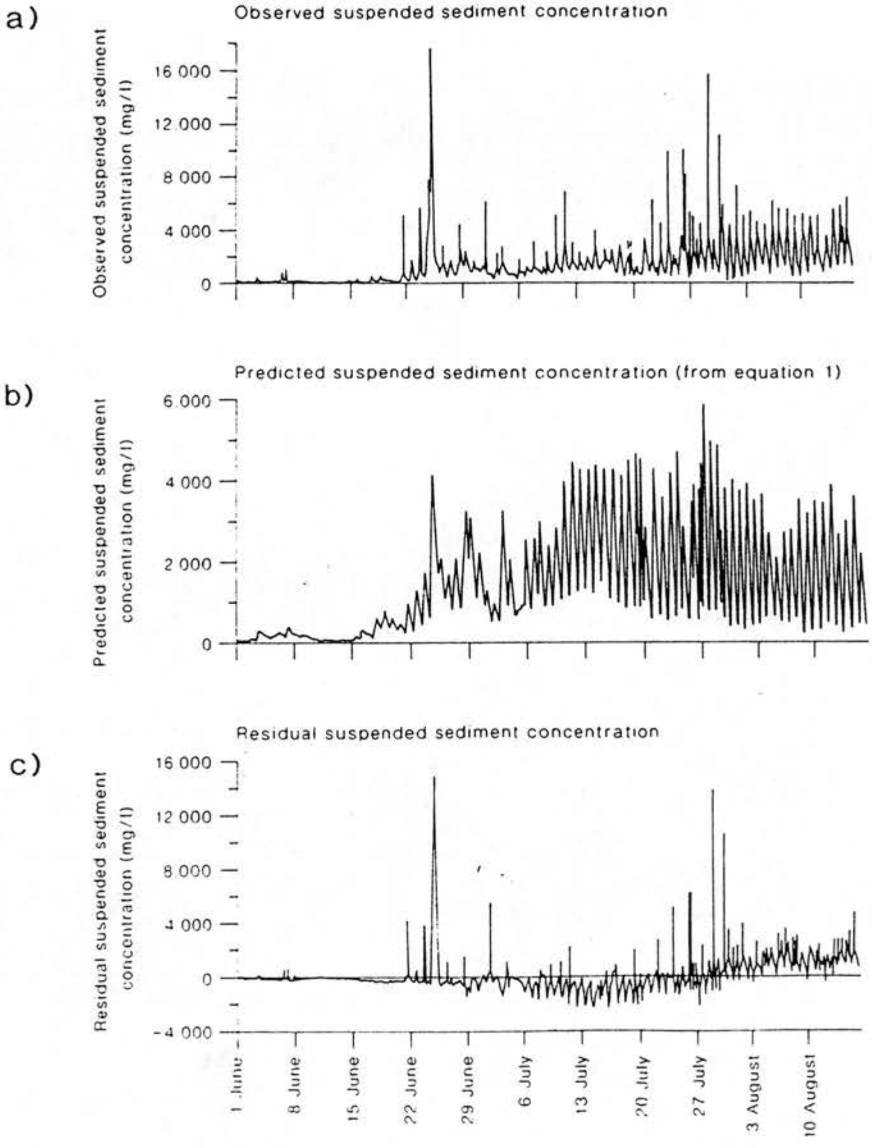


Figure 2.14

Suspended sediment outputs from the Haut Glacier d'Arolla, summer 1990: a) observed (i.e. logged at hourly intervals); b) as predicted by linear regression; c) residuals (= observed minus expected). From Clifford *et al.* (1995).

discharges. Scattered flood peaks are required to raise (briefly) sediment yields: notably in 1987 a rainstorm (23-24 August) with a recurrence interval of 10-30 years, which gave rise to the year's largest suspended sediment spike (Collins 1995a, 1995b). In contrast, at the Haut Glacier d'Arolla, suspended sediment yields stay consistently high: neither 'Collins-type' flood events, nor sediment exhaustion seem to exert a major influence. It is the presence of a widespread, highly-integrated conduit network which appears to exert the primary control on sediment evacuation. This implies that - at the within-melt-season-scale - flushing efficiency is not so much event-dependent, but reflects also the end (quasi-stable?) state of the drainage evolution process. Why this is so is not known, but the following phenomena provide possible explanations:

1. **Late-season production of subglacial debris is higher than in early-season.** This negates any tendency towards supply exhaustion. This is unlikely: maximum surface flow speeds/fastest sliding episodes at Arolla, as with many Alpine glaciers, are observed in early summer (Peter Nienow, personal communication), so maximum debris production is expected at this time also - unless widespread clogging of the ice-bedrock interface by sediment yet to be flushed occurs! Were this to happen, then greater late-season production of debris is feasible. However, the Haut Glacier d'Arolla is relatively sluggish - mean sliding speeds at the equilibrium line of ~12 m yr^{-1} (Ian Willis, personal communication) - so it is my guess that absolute levels of debris production are low. This undermines the idea of high rates of debris production in late season as a possible explanation for sustained sediment outputs.

2. **The extended length and bed coverage of the established, integrated conduit network breaks sediment supply barriers associated with lack of channel competence and/or capacity.**

3. **Enhanced mobility - i.e. channel migration - of the established, integrated conduit network breaks sediment supply barriers associated with spatial isolation of debris.**

4. ***Diurnal-scale* flood events are important.** Regular extra-conduit flow excursions involve water sweeping across wide areas of the glacier bed to tap fresh sediment stores (see my discussion of heartbeat events, 2.2, above). This final part of the melt season at Arolla is a time of rapid melt and rapid delivery of waters, characterised by a positive water balance (i.e. englacial and subglacial stores of water fill up) and high pressure surcharging of conduits. Hubbard *et al.*'s (1995) borehole study of the variable pressure axis (see above) was carried out during this final stage of melt season activity (16-22 August 1993).

Evaluation

The last three of these mechanisms taken together suggest a process regime conducive to flushing which invokes concepts not specifically considered in Collins' model, although channel migration and flow forced out of conduits feature strongly in Collins' other work, which avoids the constraints of abstraction inevitably attached to model construction. It is highly unlikely that Collins' model is entirely incorrect; if so, the fit to spikes of suspended sediment observed at Gornergletscher is remarkably fortuitous! I repeat: the basic principle seems perfectly sound. However, it seems likely that Collins' model disguises the real-world impact of certain processes, incorporated implicitly, or even by chance (e.g. lateral drainage reorganisation covers much of the impact of headwards drainage reorganisation, or sediment delivery by till deformation part-incorporates the impact of channel migration). Collins' model is highly unlikely to apply equally well to all glaciers. Flushing is not likely to be a process which can be reduced to one or two key mechanisms of universal applicability: till deformation and flood events will be far more significant at some glaciers than others. Gornergletscher is likely to differ from Haut Glacier d'Arolla is likely to differ from Sólheimajökull, although flushing at all three is thought to be highly effective (basal ice development is weak at Arolla; laminated facies ice is absent: Hubbard and Sharp, 1995). I find the integrated conduit system scenario as inferred to apply at Arolla attractive as a possible explanation of flushing at Sólheimajökull. This alternative body of ideas makes an important contribution to flushing theory which, when added to Collins' work, can claim to be reasonably comprehensive. This scenario is consistent with what little can be inferred from available data, and perhaps provides a better fit to the less-seasonal behaviour of subglacial drainage at Sólheimajökull than does Collins' purge-recharge model - especially so if the reduced seasonal contrast and higher overall levels of discharge permit an established conduit network to survive as the major means of drainage beneath much of the ice for much of the year.

C) WHAT EXACTLY DOES THE 'DRY' TO 'WET' CELL TRANSITION INVOLVE?

The difficulties involved in specifying key processes inject much of the ambiguity into Collins' model (and in part create the event vs. background process debate I have just discussed). The simple concept of flow expansion as discharge rises provides the strength of the model - it identifies the fundamental principle that 'wet' beds facilitate removal of debris - but it includes its chief weakness also: i.e. what exactly defines a 'wet' bed? Three issues raised by Collins are important:

1. **What exactly is the nature of the discharge-bed coverage relationship?** It is feasible that a given flood event can be routed with equal ease through either a widely-distributed system of small cavities and channels, or through a system of large, hydraulically-efficient conduits (or, indeed, by a process which involves interaction of the two systems). In the first case, bed coverage and flushing potential are expected to be high, whereas in the second case they are expected to be much lower. This uncertainty over the style of flood discharge applies at a wide range of scales, from small alpine-type glaciers to what happened beneath the last Laurentide Ice-Sheet (e.g. Shoemaker, 1992; Shaw, 1994). Increasingly it is evident that subglacial hydrology cannot be defined in simple terms: ubiquitous drainage structures, both in space and time, do not exist, and abstract, steady-state drainage models are likely to provide inadequate representations of flow behaviour. This carries clear - and, it has to be said, slightly depressing! - implications for any theory of flushing. However, factors such as glacier turnover (fast sliding inhibits conduit development), bedrock control of drainage routes, and tendency to seasonal reorganisation of drainage (e.g. the different climates of the Alps vs. Iceland vs. the Himalayas) are likely to exert an important influence on both overall drainage system structure and overall flushing efficiency.

2. **What is the nature of channel growth as discharge rises?** Some potentially important considerations are:

   - Any type of drainage in which flow is open has spare capacity: extra flow can be taken up by increasing the cross-sectional area of flow within the existing channel form. The extent to which spare capacity prevails is likely to reflect:

     a) Substrate control of channel form.

     b) The 'typical' level of discharge to which channel geometry tries to adjust.

     c) Antecedent flow conditions, especially the time elapsed since, and magnitude of, the last flood event.

     d) Glacier geometry, which controls channel recovery rate by ice-creep closure.

     These factors are probably impossible to quantify with accuracy. However, field studies show that open flow is relatively uncommon, so the bulk of channels must adjust to increased flows by other means.

   - Channels confined by bedrock are likely to respond first by an increase in stage; if this is impossible, by an increase in water pressure and flow velocity. If roof contacts with ice are disregarded, such behaviour is unlikely to raise levels of sediment in transport (although if conditions are exceptional, catastrophic enlargement of bedrock channels by cavitation can occur: Drewry, 1986, pp. 68-72). Sediment yields at Glacier de Ferpècle are low by standards of the Valais: equal to only ~20% of yields at the Haut Glacier d'Arolla (Gurnell, 1994). This reflects the fact that Ferpècle has a hard rock bed, unusual for this area; however, given that Ferpècle also displays extensive exposures of basal ice, it is likely also to reflect that it is part-drained by a network of deep Nye-channels.

- Channels confined by ice can adjust to a rise in discharge by ice-melt, but this response is sluggish (see above). The immediate response (i.e. as channel pressures rise to meet separation pressures) is often to expel water into adjacent areas of bed storage/fresh sediment stores. Cavity growth as water volumes/pressures rise represents a similar process. This raises issues of channel integrity: channel breakdown, even if it is only temporary or partial, is likely to raise flushing efficiency.
- Channels cut into sediment enlarge themselves by the excess of bank erosion over till deformation (Walder and Fowler, 1994). Loose, cohesionless sediments favour bank erosion. Subaerial streams cut into fluvio-glacial sediments tend to have high width-to-depth ratios, with mobile bank sediments contributing to a marked propensity for channels to widen - so tapping sediment - as discharge rises (statistically, width rises as the cube-root of discharge: Richards, 1982, Fig. 6.1b). However, this process usually tends to be self-limiting: velocity and bank shear stress should fall as channels widen; if these fall below the critical levels necessary for sediment entrainment, channel form will stabilise (or shrink if deposition occurs), and so release of sediments will end.
- Many channels are likely to have hybrid forms: i.e. their perimeters consist of a mixture of clean ice, dirty ice, bedrock and sediment. If so, the direction in which channels enlarge will be important: lateral growth is likely to access greater quantities of debris than vertical growth (Fig. 2.15). However, the style of channel change is likely to be difficult to predict: e.g. 1) enlargement by bank erosion is flow threshold-dependent, enlargement by ice-melt is not, something which implies cross-over behaviour - from ice-melt to sediment removal - as discharge rises; e.g. 2) realistic ice-channel geometry under steady flow conditions is probably broad and shallow[7], but under flood conditions energy loss is greatest where flow is deepest, so roof-melt then favours a semi-circular or arch-shaped conduit (Shreve, 1985).

[7] In this case roof creep closure and lateral melt-induced growth dominate (Hooke *et al.*, 1990).

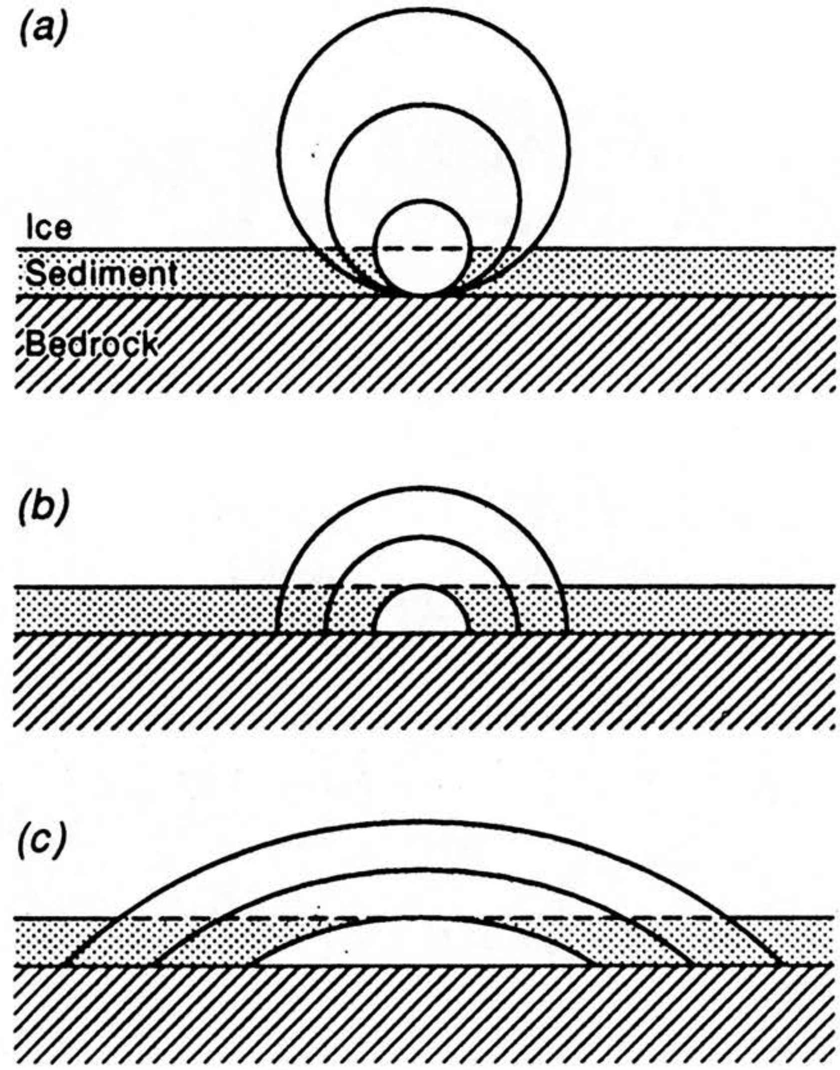


Figure 2.15

Channel geometry, channel growth, and potential release of sediments (from Collins, 1995b). 'Sediment' in this figure indicates either soft, loose basal debris and/or debris incorporated into basal ice. As channels enlarge, the proportion of their perimeter made up of sediments (whether loose, or in basal ice) tends to fall. Sediment release as channels enlarge is likely to be proportionately greater for the case of predominantly lateral growth (c) than it is for the case of predominantly vertical growth (a).

3. **What is the extent of, and the controls upon, channel mobility?** Switches in channel position offer an alternative to channel growth in response to changing discharge - and, unlike channel enlargement, are also likely to occur at steady discharges. Greater shifting of subglacial channels is thought to explain elevated yields/reduced evidence of sediment exhaustion at Glacier de Tsidjiore Nouve in comparison with its neighbour, Bas Glacier d'Arolla (Gurnell *et al.*, 1988), although independent evidence for this different behaviour relies purely on anecdotal and photographic evidence of portal variability (i.e. the outlet of Tsidjiore Nouve is highly mobile, whereas the outlet of Bas Arolla is fixed). Very little is known of subglacial channel mobility, however. The fact that Nye-channels are not found with greater frequency suggests that flow migration must be considerable - unless, of course, one wishes to argue that sediment supply to fixed channels is so low that bedrock erosion is next-to-impossible!

ADDITIONAL FACTORS (Beyond the Collins model...)

Box 2.3 summarises the processes likely to contribute to subglacial flushing, and tries to identify the factors which determine the efficacy of each process. It indicates the range of possibilities we can draw upon to inform investigations of ice-marginal sediment accumulation; however, although reasonably comprehensive, it stops well-short of a rigorous theory of flushing. This requires *full* knowledge of subglacial conditions specific to the glacier under study, whereas at present we have reasonable knowledge of changing subglacial drainage structures and bed properties at just a handful of valley glaciers (notably Haut Glacier d'Arolla, South Cascade Glacier, Storglaciären, Trapridge Glacier and Variegated Glacier). These empirical findings demonstrate the diversity of subglacial regimes, but, although informative at a qualitative level (cf. Box 2.3), they cannot be used to build a deterministic theory capable of detailed quantitative predictions.

Much of the difficulty associated with any attempt to formulate a comprehensive theory lies with the fact that certain properties simultaneously seem both to encourage and to inhibit subglacial flushing. Three of these are:

1. Contact between water and basal sediment is maximised if the area of the bed covered by active drainage is maximised, but this implies some kind of distributed system in which flow volume, stream power per unit bed area, and so flow competence and capacity are likely to be relatively low.

2. The transport capability of larger conduits is likely to be greater, but - unless basal ice is exceptionally thick - local debris supply is likely to be limited because a high proportion of the channel perimeter is likely to be cut into clean ice (see Box 2.3).

3. Both unstable drainage networks prone to episodes of reorganisation (e.g. as at Gornergletscher?), and established, integrated networks (e.g. the late-season drainage of the Haut Glacier d'Arolla?) seem to favour high flushing efficiency.

Box 2.3 introduces two processes which so far have received little attention. The first of these is sediment delivery to conduits by deformation and melt of debris-rich basal ice (cf. Box 2.2). This follows from Röthlisberger's (1972) steady-state analysis of conduits; before the 'soft-bed revolution' plastic creep of dirty ice was the process by which sediment was assumed to travel to channels, as used notably by Shreve (1985) to account for the preponderance of relatively coarse, angular, poorly-sorted debris of local origin in the sharp-crested eskers of the Katahdin system, Maine, USA. This process is likely to be important if large parts of a glacier's bed consist of rigid bedrock, or if Fowler and Walder's (1994) conclusions as to the likely geometry and behaviour of drainage below a valley glacier are correct; however, its importance will also rely on the proportion of debris produced which is actually entrained into basal ice. Although processes by which debris entrainment occurs are reasonably well-understood, little, if anything, can be said about the absolute and relative quantities of debris incorporated into the basal ice of temperate glaciers (see Chapter 6 for more on basal ice).

The second issue is that of channel sinuosity and migration. Classic steady-state theory assumes that melt and closure are balanced evenly across the full perimeter of the conduit which makes differential channel change, and development of conduit sinuosity impossible. Thus channel sinuosity is restricted to that introduced by bed roughness perturbations of the hydraulic potential field: the obvious example is Kamb's (1987) 'textbook' linked-cavity system. However, dye-tracing studies and the evidence of eskers clearly indicate that conduits are commonly sinuous - but sinuosities are rarely >2.0, and so match values common for subaerial channels. (The sinuosity of a channel made up of a succession of inflected semi-circles is 1.57.) It is expected that subglacial channels meander: non-uniformity - which creates the tendency to differential ice-melt, or, indeed, differential sediment transport, by way of an uneven distribution of channel perimeter shear stress - seems to be the natural state of fluid flow (e.g. Gorycki, 1973), and, once established, tends to exaggerate itself by a process of positive feedback (e.g. Lewin, 1975 - on meander development in the River Ystwyth, Wales). Supraglacial ice channels commonly display meanders which fit the 'excess stream power' hypothesis: i.e. sinuosity is a function of initial power expenditure per unit bed area [Knighton

(1972) Østerdalsisen, Svartisen, Norway; Ferguson (1973): Bas Glacier d'Arolla, Switzerland]. Similarly, differential ice melt is expected to give rise to subglacial meanders. In some places of above-average flow energy expenditure channel wall retreat from ice melt will exceed replacement of ice by inwards ice flow (ice melt > ice flow = 'erosion'); conversely, in areas of lower wall melt, inwards ice flow will dominate (ice flow > ice melt = 'deposition'). This will create a sinuous channel. Factors likely to set subglacial meanders apart from supraglacial meanders include:

- Supra- and subglacial flows are likely to occur under much the same hydraulic gradient (conveniently approximated by ice surface slope), but subglacial discharges are likely to be higher because catchments are larger (supraglacial catchments tend to be truncated by crevasses or moulins). This suggests that subglacial stream powers are higher, which in turn implies development of more sinuous meanders.
- However, bedrock and sediments, not just ice, are likely to provide perimeter resistance to subglacial meander development. This will part-determine channel plan-form regardless of stream power.
- Surface meanders are not affected by ice deformation. However, meanders which cut across subglacial ice-flow paths must offset the impact of ice advection. Fast sliding ice tends to suppress meander development (Hooke, 1984). See Chapter 7.1 for some thoughts on how this process is likely to affect conduit flow and debris flushing in ice-falls.

Subglacial meanders are potentially important because:

- Sinuous channels increase the length of channel perimeter in contact with potential sediment stores.
- Flow inertia causes overshoot (or undershoot) of the high velocity filament in sinuous channels, so creating preferential zones of melt (or scour of bank sediments) which encourages delayed (or premature) inflection of the channel bend, and so channel migration.
- If channel banks are undercut, bank collapse possibly: a) raises sediment supply; and, b) promotes channel blockage (by sediments or ice) which leads to channel diversion.
- Sinuous channels increase the likelihood of closed flow, and tend to raise water pressures in closed conduits. This reduces spare capacity in the drainage system, and increases the probability of extra-conduit flow excursions.

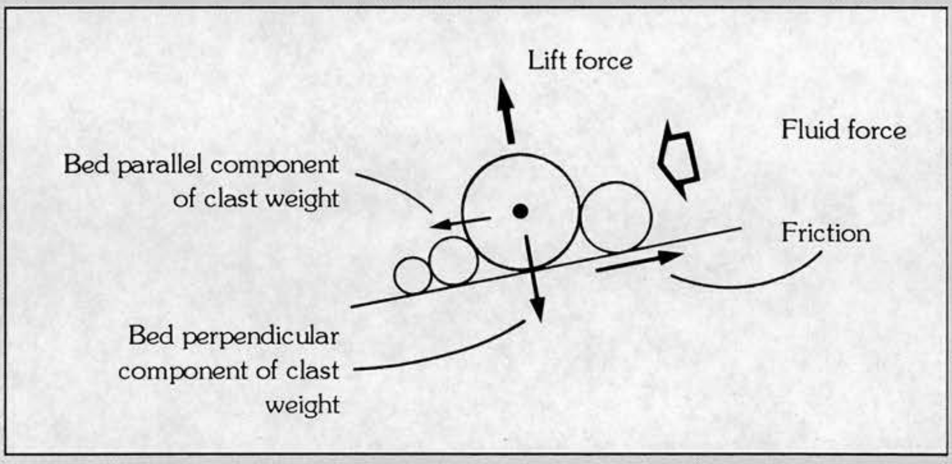
BOX 2.3
Subglacial flushing: a conceptual framework

In this box I summarise the key ideas and factors relevant to a theory of subglacial flushing.

BASIC PREMISE:
Flushing efficiency rises directly as bed coverage by competent and capable water flows rises.

A) DIRECT ENTRAINMENT OF LOOSE SEDIMENTS

e.g. at channel margins or in cavities



Entrainment criterion

TRACTIVE FORCES: $\mathbf{W \cdot \sin\beta + \tau_0 \cdot A + LF}$ **must exceed**

RESISTING FORCES: $\mathbf{W \cdot \cos\beta \cdot \mu}$

W	buoyant weight of clast
β	slope of stream bed
τ_0	bed shear stress (fluid force)
A	exposed area of clast
LF	lift force
μ	friction coefficient - incorporates packing, pivot, hiding and protrusion effects

Enhanced by:

- High power flows
- Fine, spherical, cohesionless sediments
- Fine sediments entrained most easily if bed sediments are uniform (hiding effects minimised)
- Coarse sediments entrained most easily if bed consists of mixed-size clasts (protrusion effects maximised)
- Low rates of sediment supply from upstream: i.e. excess capacity of flow at any spot is high

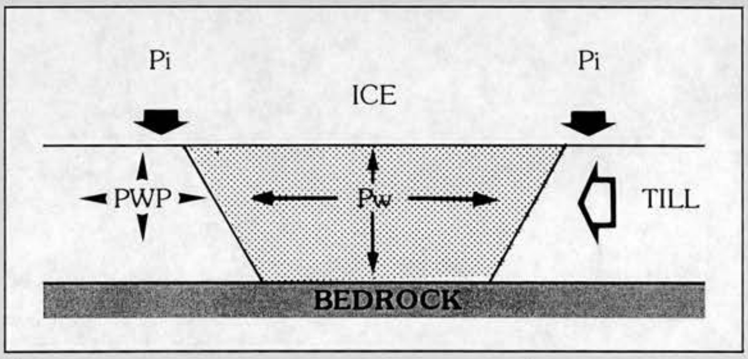
(PTO)

Box 2.3 (continued)

B) SEDIMENTS GO TO THE CHANNEL

1. TILL DEFORMATION INTO CHANNELS

Key references: Boulton and Hindmarsh, 1987; Alley, 1992; Fowler and Walder, 1994; Collins, 1995b



Pi ice pressure
Pw channel water pressure
PWP pore-water pressure

STRENGTH OF TILL:

$$\mathbf{s = c + N \cdot \tan \phi}$$

c cohesion
N effective pressure: channel margin = (Pw - PWP)
at a distance = (Pi - PWP)
ϕ friction angle of till

RATE OF TILL DEFORMATION proportional to:

$$\mathbf{(Pi - Pw)^a \cdot N^{-b}}$$

Enhanced by:

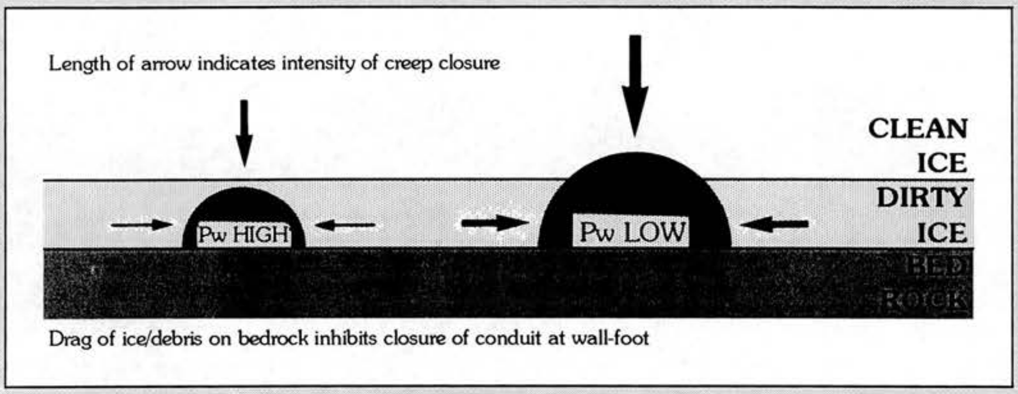
- Thick ice, gentle ice surface slopes
- Low channel water pressures
- High till pore-water pressures
- Fines content of till high (gives low till shear strength/viscosity)
- High bank shear stress/stream power (favours rapid entrainment and removal of sediments)
- Wiggly, anastomosing canal network with irregular bed topography (disrupts Weertman pressure-dam effect)

(PTO)

Box 2.3 (continued)

2. DEFORMATION OF DIRTY BASAL ICE INTO CONDUITS

Key references: Röthlisberger, 1972; Hooke, 1984; Shreve, 1985; Hooke *et al.*, 1990



Ruling equations

RATE OF CLOSURE proportional to:

$$\mathbf{A\ (\ Pi\ -\ Pw\ )^{n}}$$

A coefficient from Glen's Law (ice softness)
n exponent from Glen's Law (commonly taken to be 3)

MELT RATE proportional to:

$$(\rho_w \cdot g \cdot Q \cdot s) / L = \tau_0 \cdot u_w$$

ρ_w density of water
g gravitational acceleration
Q water discharge
s hydraulic gradient
L perimeter of channel made up of ice
u_w water flow speed

STEADY STATE, CLOSED FLOW: melt rate = closure rate

OPEN FLOW: melt rate > closure rate

Enhanced by:

- Thick ice (high Pi)
- High discharge, with flow under steep hydraulic gradient (i.e. high stream power)
- Straight, semi-circular, low roughness channels (low Pw)

(PTO)

Box 2.3 (continued)

Additional factors:

- Clean ice is likely to make up a large part of the perimeter of large conduits: i.e. conduit radius > thickness of basal ice layer. With flow under a hydraulic gradient of 0.1, through a channel of semi-circular cross-section, with Manning roughness, $n = 0.05\ m^{-1/3}\ s^{-1}$ (using Hooke, 1984, equation 3):

* If the basal ice layer is 0.1 m thick, it will enclose fully only those flows $<0.025 m^3\ s^{-1}$.

* If the basal ice layer is 0.5 m thick (unrealistic!?) it will enclose fully only those flows $<1.42\ m^3\ s^{-1}$.

  Unless exceptional tectonic thickening occurs, basal ice layers >0.1 m thick are rare.

- Small conduits are more likely to be fully enclosed in basal ice, but melt and closure rates will be low.

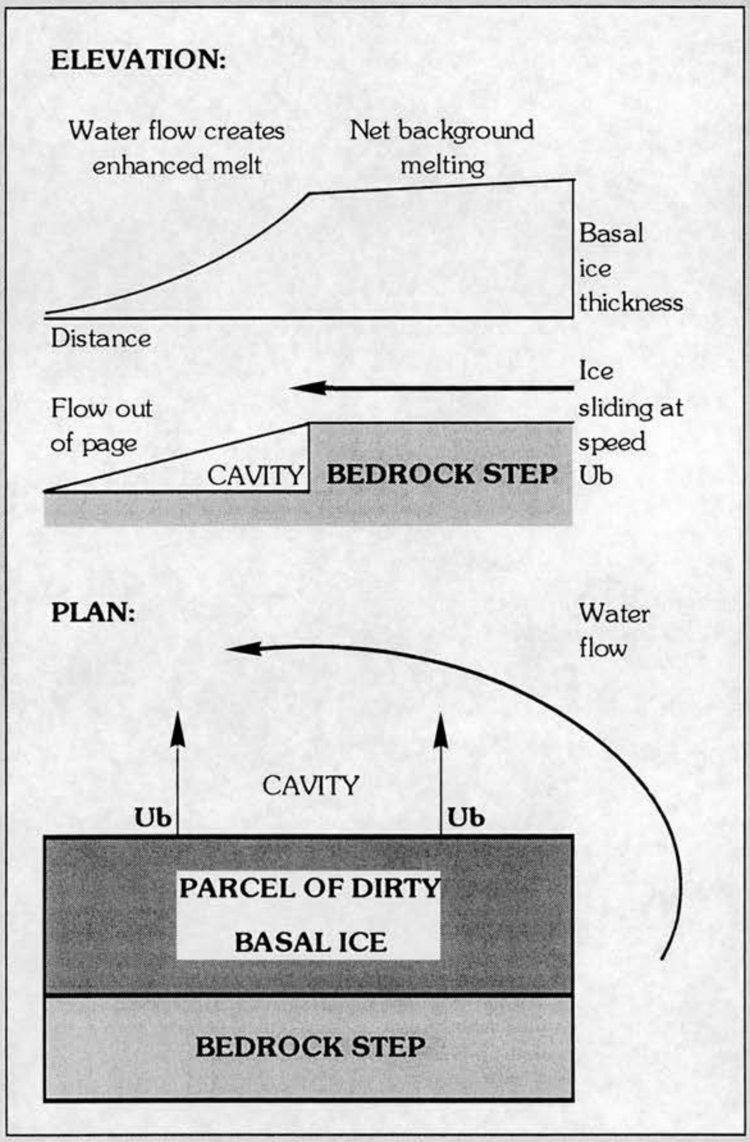
- Conduits roofs - likely to be made up of clean ice - tend to close faster than conduit wall-feet (ice-debris drag here inhibits creep closure) - particularly if conduits deviate from the ideal semi-circular shape.

- Impact of Weertman pressure-dam effect? If marked (= large, straight, parallel channels with flow at low pressure over an even, rigid bed) it will restrict the lateral zone from within which basal ice can converge on conduit flow axis; implies deformation of clean roof ice dominates.

(PTO)

Box 2.3 (continued)

3. ICE SLIDING CARRIES DEBRIS TO CHANNELS



Enhanced by:

- Preponderance of transverse channel elements
- Individual transverse channel elements are wide, with high stream power
- Sliding speeds are slow relative to stream power
- Limited recovery of melt losses by bed-normal component of ice deformation between channels (favoured by low bedrock roughness; close channel spacing)

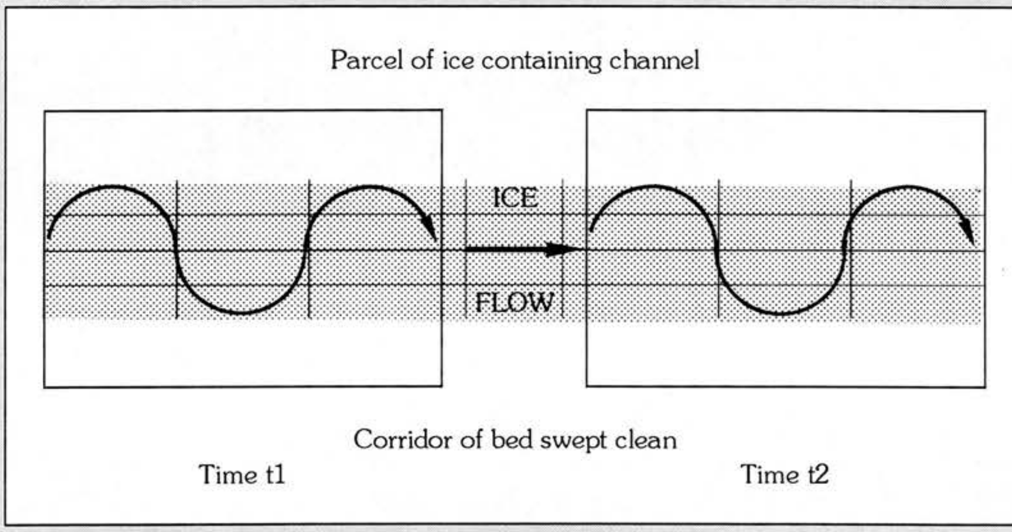
(PTO)

Box 2.3 (continued)

C) CHANNELS GO TO THE SEDIMENTS

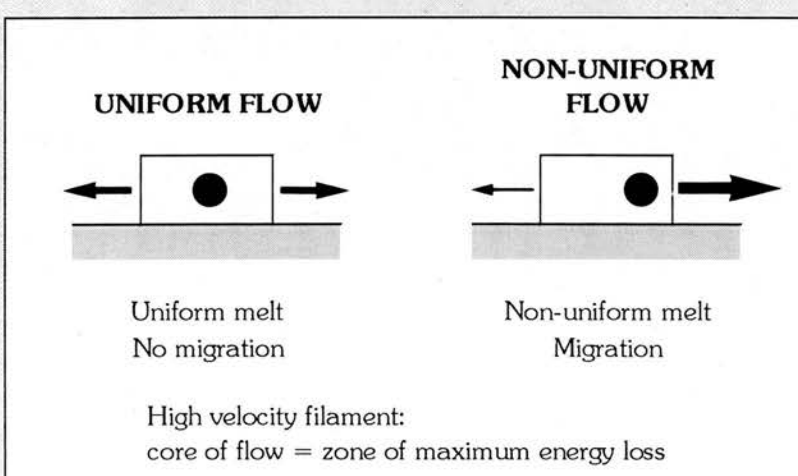
Key references: Collins, 1979a; Hooke, 1984

1. SINUOUS CHANNEL IS CARRIED FORWARDS BY ICE



2. CHANNEL MIGRATES THROUGH ICE

Sinuosity develops if the high velocity filament is displaced:



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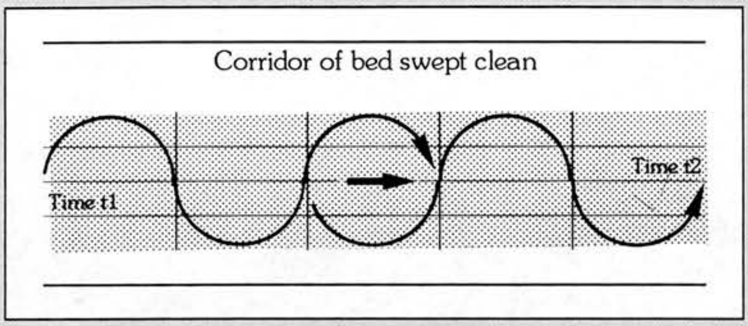
Box 2.3 (continued)

Impact:



Enhanced by:

- Optimum combination = high sinuosity channels migrating rapidly within fast sliding ice (maximises area of bed swept clean per unit time) - but this is not a likely combination!
- High ('excess') stream power - creates high sinuosities and high migration potential (cf. 'classical' meander theory)
- Highly non-uniform flow: distinct high-velocity filament and so distinct maxima and minima of energy dissipation
- Bed slope *downglacier* relatively *gentle*: a) this raises excess levels of stream power; b) *steep* slopes direct flow directly downslope and so restrict channel sinuosity
- Relatively steep *transverse* bed slopes: channels tend to side-slip downslope
- Sliding speeds relative to melt rates low (fast sliding opposes differential melt and so suppresses channel migration

N.B. Meander development by differential sediment transport is similar to meander development by differential ice-wall melt (the difference lies with the nature of the channel perimeter resistance) so I do not consider it in detail here.

4a. DRAINAGE REORGANISATION

Key references: Collins, 1989; Kamb, 1987; Humphrey and Raymond, 1994

Triggers:

1. Fast sliding over rough beds carries conduits into contact with bed obstacles, so conduits are squeezed shut
2. Channel blockage - by ice or sediments - forces flow diversion
3. Rapidly imposed additions to volume of flow and/or rising water pressures

(PTO)

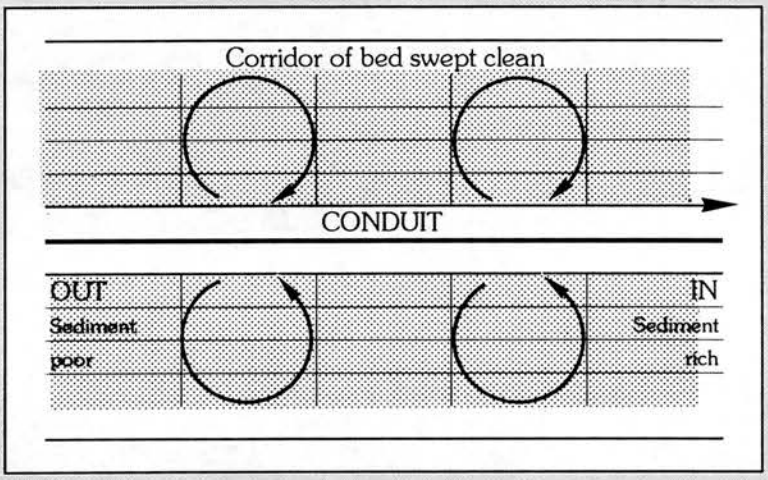
Box 2.3 (continued)

Similar to:

4b. EXTRA-CONDUIT FLOW EXCURSIONS

Key references: Collins, 1979b; Hubbard *et al.*, 1995

If Pw > Ps (Ps = separation pressure):



Enhanced by:

- Low separation pressures (= rough bed, low debris content of basal ice: Iken and Bindschadler, 1986; Schweizer and Iken, 1992)
- Limited spare capacity in channel (closed, high-pressure flow, not open flow)
- Rapid variability of Pw (= high melt rates, fast delivery from glacier surface, distinct diurnal melt cycle)
- Channel blockage (see above)

(PTO)

Box 2.3 (continued)

D) WHAT IS THE OPTIMUM NUMBER OF CHANNELS?

- The larger the number of channels, the greater will be the bed coverage. As channel numbers rise, and channels size falls, drainage network structure switches from discrete to fully-developed distributed status. Likely relationship:

  Bed coverage α (number of channels)^x

  with x >1.0 (power relationship reflects reduced hydraulic efficiency of smaller channels).

- Numerous, smaller channels likely to be fully-enclosed in basal ice (see above).

- Numerous, smaller channels, closely-spaced, are less susceptible to isolation from adjacent water, basal till and basal ice by the Weertman pressure-dam effect: a) Pi - Pw tends to be smaller, and, b) bed roughness elements exert a greater influence on smaller channels, and so permit breach of the pressure seal.

- Smaller conduits achieve greater sideways shift for a given energy loss per unit time because of their smaller cross-sectional area:channel perimeter ratio. E.g. 10 semi-circular conduits which each carry 0.1 $m^3 s^{-1}$ down a slope of 0.1 potentially shift ~50 m in 100 days; a single conduit which carries 1.0 m s^{-1} down a 0.1 slope expends an equal quantity of melt energy, but migrates just ~20 m (Hooke, 1984, equation 6). Coupled with closer spacing of smaller conduits, this implies a greater bed area is swept clear as the value of **(number of conduits / given discharge)** rises. However, smaller conduits with reduced melt rates are less able to oppose fast sliding.

- **BUT** the sediment transport capability of smaller conduits is limited!

- The probability of clast evacuation is likely to be maximised if the maximum distance between any point on the glacier bed and a major drainage channel is minimised. This is because

  1. clast entrainment and transport
  2. extent of bed separation (both vertical and lateral)
  3. extent of flow excursions
  4. extent of channel migration

  can all be conceptualised as *probabilistic* phenomena, subject to *distance decay*: e.g. as a negative exponential function. Thus the probability that these processes (individually or collectively) operate over distance x (and so tap debris) is high if x is small, but falls rapidly as x rises (if x is the distance between a major channel element and a given point on the glacier bed).

EVALUATION

So: what is the optimum drainage structure for maximum efficiency flushing?

- **BEST-GUESS: an efficient, integrated network of major trunk conduits, fed by a dense network of cavities and (mobile?) smaller passageways, with isolated areas of the bed (film or pore-water drainage only) limited.**

- **EVIDENCE FOR THIS KIND OF DRAINAGE STRUCTURE BENEATH MANY VALLEY GLACIERS IS COMMON. THIS SUPPORTS MY INFERENCE THAT FLUSHING DOMINATES THE SUBGLACIAL SEDIMENT TRANSPORT REGIME OF MOST VALLEY GLACIERS.**

2.6 SYNTHESIS:
SUBGLACIAL FLUSHING OF ICELANDIC GLACIERS

Because of the number of different possibilities (i.e. individual processes and combinations of processes) it is impossible to specify exactly what conditions determine the efficiency of subglacial flushing for any given glacier. Nevertheless, the basic premise is straightforward: **high power flows in contact with debris across wide areas of a glacier's bed will tend to sweep that bed clear of that debris**. Subglacial drainage systems appear to be immensely versatile in this respect: a wide range of behaviour seems to satisfy this basic flushing condition (Box 2.3). Superficially, episodes of major drainage reorganisation (e.g. the spring event, or progressive headwards collapse of the cavity system: Collins, 1989; Richards *et al.*, 1996) and periods of apparent network stability (e.g. late season conduit drainage at Haut Glacier d'Arolla: Clifford *et al.*, 1995; Hubbard *et al.*, 1995) seem to be very different, but in fact they do share key common characteristics: i.e. alternate storage and release of water; short-lived high-pressure triggers; and extensive bed separation, flow diversion and reorganisation.

My view is that flushing dominates the sediment budgets of perhaps the majority of temperate glaciers, to such an extent that it should be regarded as the typical or default process regime. It is easy to specify qualitatively the conditions which break this behaviour, and so carry the ice-water-debris system into an alternative process regime: **widespread separation of water and debris is required**. This is achieved, for example, if large quantities of rock-fall debris travel to the ice edge by passive transport, or if pervasive marginal freezing occurs. Neither of these is particularly effective in the case of many outlet glaciers in Iceland, so that extensive moraine development at certain glaciers presents itself as a major puzzle. Chapters 4 to 7 examine the case of Gígjökull and Steinholtsjökull, which display rapid rates of moraine accumulation, but without any immediately obvious explanation for why the flushing constraint is by-passed here. In contrast, I put forward Sólheimajökull as the type example of a glacier for which subglacial flushing dominates the overall pattern of sediment transport.

Stream power

As a first approximation I presume that stream power (Ω) acts as the fundamental control of flushing efficiency: i.e.

$$\Omega = \rho_w \cdot g \cdot Q \cdot s$$

ρ_w	density of water
g	gravitational acceleration
Q	discharge
s	hydraulic gradient

This is a simple, but potentially powerful, relationship, which encapsulates many existing ideas. Stream power represents the rate at which potential energy is converted into some kind of geomorphic work. Because the flushing process is so complex I think it is extremely difficult to be more specific than this - but then again, perhaps it is not necessary to look for greater detail. Stream power, as with thermal regime for instance (Boulton, 1972a; Sugden and John, 1976; see Chapter 1.4), provides a certain type of explanation at a particular level of reality. At the scale of the full glacier (which will incorporate a wide range of flushing behaviour) a simple concept such as stream power *emerges* as a satisfactory and potentially powerful explanatory tool. Although use of a stream power index will disguises many important factors, such as the exact configuration of the drainage network, time-dependent changes of drainage, or the possibility of major zones of englacial flow, it does carry within it a strong measure of process input, because it relates directly to process mechanisms which exist at a deeper level of reality (Chapter 1.4), notably:

1. Rates of ice melt induced by running water (important in itself, but indirectly also as half of the process balance which determines water pressure in conduits).

2. Rates of sediment transport: stream power is the rate at which bed shear stress is renewed, i.e.

$$(\rho_w \cdot g . Q . s / w) = \tau_0 \cdot u$$

w	channel width
τ_0	bed shear stress (the force which determines sediment entrainment and transport)
u	mean water flow speed

3. Propensity for meanders to develop.

All of these factors must be central to any full theory of flushing. Stream power is also useful because it tends to scale with catchment area and mean catchment slope. These variables are quick to calculate, and provide a convenient surrogate measure for use in geomorphic studies; see Chapter 3 for an example.

Basin topography and ice geometry

It is possible that use of the stream power index can be extended to incorporate other potentially important factors. The first of these is basin shape, which will affect *specific* stream power: i.e. stream power per unit 'flow width', with 'flow width' here defined as width of the trough floor, rather than, as is usual in fluvial geomorphology, the width of individual stream channels. Dispersed, low power flows are likely to present less of a threat to the survival of subglacial debris. Thus topographic convergence - i.e. a wide basin feeds a narrow trough, as at Sólheimajökull, for example - is likely to concentrate water action at the base of the ice. In contrast, Hofðabrekkujökull, the relatively unconfined, piedmont-type south-eastern outlet of Mýrdalsjökull, has no major outlet, but is drained by a large number of smaller streams, and has major exposures of basal ice. Absence of flow convergence here, however, is likely to be assisted also by drainage through a porous tephra layer (the product of successive *jökulhlaups* from Katla, the last of which was in 1918).

Ice surface slope, trough size and ice mass turnover are also likely to be important. *Glacier* power, defined by

$$\rho_{ice} \cdot g \cdot Q_{ice} \cdot \sin \alpha$$

ρ_{ice} density of ice
Q_{ice} ice discharge
α ice surface slope

controls the energy expended on subglacial erosion and transport of debris (Andrews, 1972). If glacier power is high relative to subglacial stream power, then this will favour construction of large moraines (i.e. lots of debris is produced, but little of it will be washed away). This kind of condition possibly applies to many sub-polar glaciers which have warm-bed interiors, but frozen margins (cf. the Boulton vs. Andrews debate, 2.3, above). Conversely, if glacier power is low relative to subglacial stream power, then small, if any, moraines only are likely to be built (i.e. little debris is generated; that which is is readily washed away). Sluggish Alpine glaciers, subject to high melt rates, which are presently in retreat (e.g. the Haut Glacier d'Arolla) possibly fit this category (although debris of supraglacial origin in passive transport is likely to contribute significantly to moraine growth at these kind of glaciers). Because of its maritime location both glacier power and stream power are high at Sólheimajökull. The evidence of present-day moraine formation (or the lack of it!) indicates that in this case the balance between the two favours stream power; however, a small change to this balance to the advantage of glacier power (as I infer to have occurred at Sólheimajökull in the Little Ice Age: Chapter 3) would be likely to reduce the relative impact of flushing, and so boost rates of moraine formation.

CHAPTER 2: SUMMARY

- Subglacial erosion rates at Sólheimajökull are thought to be high, yet present-day moraine accumulation is extremely limited. This must mean that the bulk of the debris produced is removed by the work of subglacial water flow.
- This simple observation defines the basic idea which guides my thesis: subglacial meltwater activity exerts a major - if not the dominant - influence on the distribution of debris between alternative transport pathways, and so the catchment sediment budget and styles of ice-marginal sedimentation. Previous work has tended to overlook this crucial principle.
- Preservation of debris requires widespread separation between it and competent/capable water flows. This is unlikely to occur at many temperate glaciers for the case of debris which travels as part of the basal transport zone.
- Factors which favour efficient flushing seem to include:
  1. Generally high, but variable, throughputs of water under the influence of a steep hydraulic gradient (i.e. high stream power).
  2. A subglacial drainage network which combines episodes of reorganisation with integration of its different elements (i.e. water - and debris - across wide areas of the glacier bed is fed rapidly into relatively large channels). It is likely that episodes of reorganisation are a necessary part of the process by which drainage becomes integrated.
  3. High water pressure events - at a wide range of scales - are likely to be important. These cause enhanced bed separation, episodes of channel reorganisation/migration, and increased contact between debris and elements of competent and capable flows. Steady-state drainage theories are likely seriously to underestimate the frequency with which such high pressure events occur.
- Very little is known of the subglacial drainage at Sólheimajökull, but what can be inferred is consistent with the prevalence of these three factors.
- Because meltwater activity exerts a major influence on styles of ice-marginal sedimentation, changing drainage regimes must be taken into account in studies of the glacial geologic record. In Chapters 3 and 8 I try to show how the past impact of subglacial hydrology can be used to improve the process input into studies of past moraine development.

CHAPTER 3
Sólheimajökull: Neoglacial moraine accumulation

INTRODUCTION

This chapter explores the possible ways in which climate change, variable ice dynamics, and time-dependent changes to subglacial hydrology interact, and how these changes are likely to influence subglacial flushing efficiency. Specifically, it examines Holocene contrasts in moraine development by Sólheimajökull in the context of Dugmore and Sugden's (1991) ice-divide migration hypothesis which purports to explain the anomalous fluctuations of Sólheimajökull in this time. I argue that a meaningful account of ice dynamics influenced by subglacial drainage adds necessary explanatory detail missing from the original ice-divide migration hypothesis. Thus simultaneously I try to refine the ice-divide migration hypothesis and explain the contrasts in flushing efficiency inferred.

The discussion I present here fails to match 'traditional' views of science as a process of empirical inductive reasoning, or deductive hypothesis testing. It is a bold idea which cannot be justified by appeal to observation, nor is it subject to rigorous test intended to falsify the hypothesis (cf. Alley *et al.*, 1994). This makes me both a bad positivist and a bad Popperian! However, it is consistent with the outlook which views geology as a necessarily interpretative science (Frodeman, 1995), and it satisfies my initial vision of a theoretically-informed glacial geomorphology applied to problems of Quaternary geological reconstructions (Chapter 1.4 and 1.5). My story draws its strength from its appeal to current theory (why have theory if we cannot put it to use?) and the internally-consistent nature of the process behaviour I invoke. It is *just* an idea, which, even if incorrect, illustrates the *potential* of using coupled accounts of variable drainage and flushing efficiency to support glacial geologic studies. However, I think it is a plausible idea which deserves to be taken seriously in the absence of an alternative hypothesis of obviously superior merit.

3.1 SÓLHEIMAJÖKULL'S HOLOCENE GEOMORPHIC RECORD

Studies by Dugmore, using historical records and a mixture of lichen, tephra and ^{14}C dating, provide a detailed picture of Holocene ice advances and flood activity at Sólheimajökull (Maizels and Dugmore, 1985; Dugmore, 1987, 1989; Dugmore and Spedding, unpublished data, 1995). This record is the best yet obtained of Holocene glacier activity in Iceland (Guðmundsson, 1997), a fact which reflects in part the unusual behaviour of Sólheimajökull in this time (see below). Three aspects of this behaviour require an explanatory process input beyond the traditional concern of the glacial geologist with the dating of events. These are:

1. What explains the history, changing frequency and changing type/trigger mechanism of floods?
2. What explains the anomalous fluctuations of Sólheimajökull?: i.e. why did episodes of advance and retreat occur out-of-phase with its neighbours?
3. What explains the change in moraine volumes associated with different episodes of ice advance?

In this chapter I concentrate on the third of these points. It should, I hope, be clear from Chapter 2 that I expect variations in moraine size and composition to reflect not just changes in glacier size, turnover and erosivity associated with episodes of climate change, but to reflect changes in catchment sediment budget associated with changes in flushing efficiency also. Previous work has considered the relationship between climate variability and the style of debris deposition associated with changing abundance of meltwater (e.g. Shaw, 1977; Ruszczynska-Szenajch, 1981), although this often takes availability of debris for granted; much less seems to have been written on potential changes to subglacial process regimes, and how these are likely to affect the partition of debris between alternative transport pathways which control debris delivery to the ice edge.

Here I consider the extent to which changes in flushing regime can be used to solve the puzzle presented by the record of Holocene moraine construction at Sólheimajökull. The largest moraines date from the period which leads up to the Little Ice Age, at which time Sólheimajökull was little further forward than it is at present; at the time of its maximum Holocene advance, moraine development was meagre. Although the process justification is far from clear (e.g. Harbor and Warburton, 1992, 1993), both conventional wisdom and the bulk of published data (e.g. Hallet *et al.*, 1996) support the idea that, in the case of temperate ice, subglacial erosion rates tend to rise as glacier area, length and thickness increase. Thus we have an apparent paradox: the volume of moraines built by successive episodes of advance varies inversely with the inferred level of erosive activity. Contrasts in flushing efficiency offer a possible solution. It is impossible to demonstrate that this did indeed occur; however, below I argue that there are strong reasons to infer that important changes to subglacial hydrology and the overall level of stream power did take place beneath Sólheimajökull in the late Holocene. This argument involves reappraisal and refinement of the ice-divide migration hypothesis which Dugmore and Sugden (1991) use to account for the anomalous fluctuations of Sólheimajökull; this implies a direct link between stages of ice advance, changes to ice dynamics and subglacial hydrology, and contrasting episodes of ice-marginal sedimentation (i.e. points 2 and 3, above, are linked).[1]

Little Ice Age moraines

The best-developed moraines at Sólheimajökull date from the period leading up to, and including the Little Ice Age: ~1,400 BP to 50 BP (Dugmore, 1987; Fig. 3.1). These consist of a regular, ribbed, 'wash-board' series of continuous, parallel ridges, which extend for ~1.5 km along the flanks of Hrossatungur (southern margin: Fig. 3.2a) and Hvítmaga/Jökulhaus (northern margin: Fig. 3.2 b): see Fig. 2.1. These moraines are dump ridges fed by debris in active transport. Clasts are predominantly sub-angular: mean roundness of a sample taken at Hrossatungur = 2.92. s.d. ±0.06, n = 63). This matches closely with the roundness scores of clasts derived from existing basal ice exposures sampled adjacent to the youngest (i.e. ice-proximal) of these ridges.

I calculate the mean size of these moraines using a simple geometric model representative of the Hrossatungur ridges: ten identical ridges with scalene triangle cross-section are assumed to fill completely a slope of 15° inclination and length 800 m. I use two values for the angle of the ridge sides: a) 20°, which fits measurements of Little Ice Age ridges in front of the snout (Sharp and Spenceley, 1993), and b) 30°, which seems close to the angle of repose of tills, and fits better my impression of these ridges in the field. This makes the total moraine volume between 372.0 and 976.5 m^3 m^{-1} ice edge, which, given that these ridges represent ~1,400 years' accumulation makes the mean rate 0.27-0.69 m^3 m^{-1} yr^{-1}. This compares with my estimate of ~0.1 m^3 m^{-1} yr^{-1} for ice-edge/slope-foot accumulation at present (Box 2.1), and

[1] It is not impossible that changes in subglacial flushing efficiency also influence the sequence of flood events (i.e. a link to point 1, above). Flushing affects the quantity of debris retained in ice; debris content a) influences sliding speeds (Schweizer and Iken, 1992), and so closure of potential lake outlets; and, b) the flotation condition which determines lake drainage (Tweed, 1992, 1995).

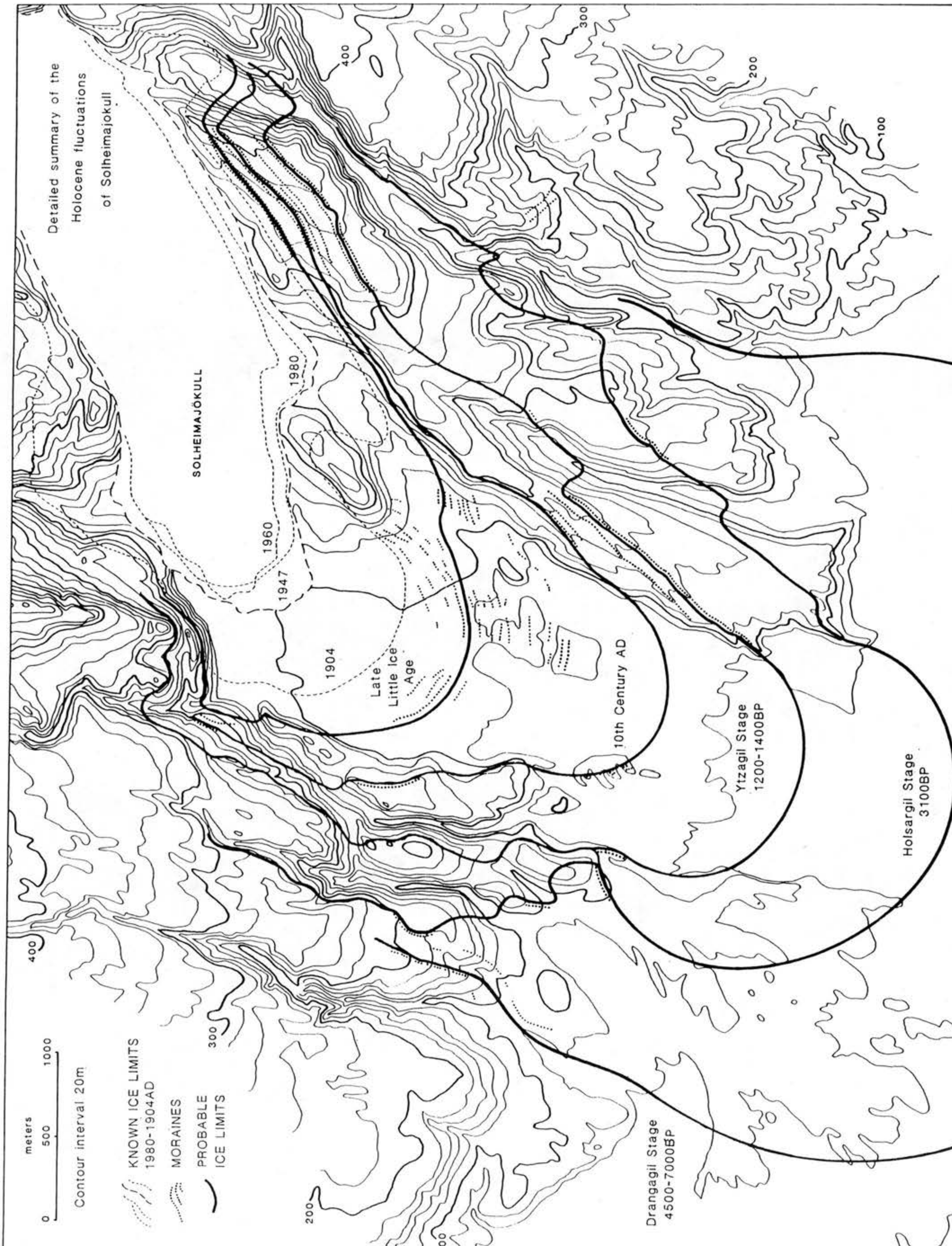


Figure 3.1

Moraine limits and Neoglacial fluctuations of Sólheimajökull (from Dugmore, 1987).

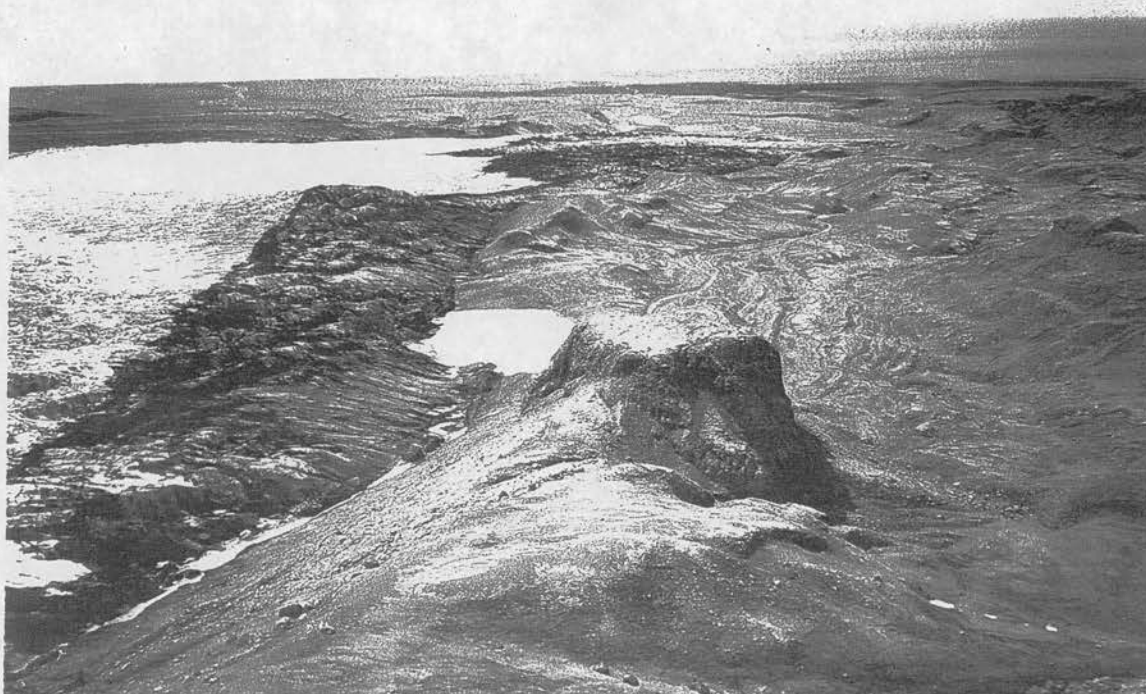


Figure 3.2

Sólheimajökull: Ystagil/Little Ice Age moraine ridges. Top (a): Hrossatungur, southern margin. Bottom (b): Hvítmaga/Jokulhaus, northern margin.

so implies much more pronounced moraine construction in the past, and - by inference - a different subglacial flushing regime.

Early Neoglacial moraines - or the lack of them!

No estimate of rates of moraine accumulation in the early part of the Neoglacial is available because so little geomorphic evidence survives. Prior to *c.* 1500 BP moraine ridge construction by Sólheimajökull was extremely limited (Fig. 3.1). I believe this to indicate a past level of moraine formation - and so high efficiency of flushing - similar to that which exists at Sólheimajökull today. It is unlikely that the (much larger - see below) early Neoglacial glacier was much less erosive than it is at present. It is possible that it did build large moraines, but that these have failed to survive: the potential impact of frequent *jökulhlaups* suggests itself as a potent force for destruction of moraines. If any large moraines were built in front of the snout it is highly likely that they were wiped out by floods. However, elsewhere (i.e. either side of the Jökulsá valley) there are 'safe', elevated areas, located well away from flood drainage routes which potentially should have favoured moraine construction and survival - but the moraines which are actually found in these areas are small and fragmented (Dugmore, 1987; Fig. 3.1). This paucity of moraine development in what should have been favourable areas implies low levels of debris delivery to the past ice margin. The difference in size between these small early Neoglacial moraines, and the large Little Ice Age ridges is thought be a genuine reflection of flushing-controlled contrasts in moraine-building potential

Past changes to subglacial hydrology?

It is impossible to obtain direct independent evidence of a possible change in flushing regime. Few successful studies of past subglacial drainage networks exist; those that do rely on direct erosional or depositional evidence of former channels. Such quality evidence is highly unusual. However, I think it is possible to draw support for a change in subglacial hydrology by means of constrained theoretical inferences as to the changing geometry of Sólheimajökull over the Holocene, and its interaction with bedrock topography and changing geothermal/volcanic influences. This stems directly from Dugmore and Sugden's ice-divide migration hypothesis. In the rest of this chapter, I set out the argument for important changes to the subglacial hydrology of Sólheimajökull over the late-Holocene, changes which potentially explain both the observed contrasts in moraine size (my way into this problem), and the changing location of moraine formation (i.e. fluctuations of the glacier margin). The exact trajectory of the argument is not immediately obvious, so I think it is useful here to summarise my two main ideas:

1. Contrasts in Neoglacial moraine formation at Sólheimajökull arise because of a change in the efficiency of flushing. Large Little Ice Age moraines indicate a relatively 'dry' bed condition. Small moraines of the early Neoglacial [Dugmore's (1987) Drangagil and Hólsárgil stages] indicate a 'wet' bed condition, perhaps similar to that believed to exist below Sólheimajökull today.

Independent support for past changes to the subglacial process regime:

2. The anomalous late-Holocene fluctuations of Sólheimajökull represent a response to episodes of climatic deterioration controlled by ice-divide migration. However, the relationship between climate, glacier catchment area and ice advance intrinsic to the ice-divide migration hypothesis itself reflects the interaction of ice geometry, subglacial geothermal activity and bedrock topography which regulates the basal meltwater flux. Changes to this change the intensity of sliding. The limited Little Ice Age extent of Sólheimajökull part-relates to suppressed sliding with a 'dry' bed, whereas its unusually large mid-Holocene advance part-relates to enhanced sliding with a 'wet' bed.

3.2 THE ICE-DIVIDE MIGRATION HYPOTHESIS

Recent work has revised the idea that, in terms of environmental change, the Holocene in Iceland was a relatively uneventful epoch. Increasingly it is evident that (Guðmundsson, 1997):

- Significant environmental change was widespread.
- The widespread evidence of ice throughout the Holocene implies that loss of ice cover between Preboreal/Boreal times and the beginnings of the Neoglacial was less extensive than previously thought.[2] It is unlikely that rapid recovery of large ice masses recorded soon after the inferred start of the Neoglacial occurred without the pre-existence of significant ice cover able to 'cold-start' the process of glacier expansion as climate began to deteriorate. This idea of relatively persistent ice cover is supported by modelling studies of ice mass accumulation elsewhere (e.g. Payne and Sugden, 1990 - see below) which imply that major expansion of ice takes $\sim 10^3$ years.
- Climate started to get colder ~6,000-5,000 BP: i.e. start of the Neoglacial.
- At least five warm/cold cycles occurred in the last 5,000 years.
- These are superimposed on a secular trend of progressive cooling: i.e. each cold cycle tends to be colder than its predecessor.

[2] Björnsson (1979) believed that in the Altithermal ice disappeared from all but the highest mountain peaks, such as Öræfi, ~ 2,000 m asl, an opinion which ws previously accepted as the orthodox view.

- The Little Ice Age seems to represent the coldest/snowiest time of the late-Holocene in much of Iceland; temperatures seem to have been consistently cold since *c.* 1500 AD, with the maximum extent of ice masses attained *c.* 1700-1900 AD in most areas.

The Neoglacial at Sólheimajökull

Dugmore (1987) identifies three distinct prehistoric still-stands of Sólheimajökull (Fig. 3.1):

1. **Drangagil** stage, 4,500-7,00 BP.
2. **Hólsárgil** stage, 3,100 BP.
3. **Ystagil** stage, 1,200-1,400 BP.

Since settlement, distinct ridges record ice advance episodes in the 10th century AD, and in the late Little Ice Age. The wash-board ridges of Hrossatungur imply that regular episodes of advance and still-stand occurred between the Ystagil stage and the end of the Little Ice Age at the start of this century: the succession of ridges built up on this flank shows no obvious breaks.

The key point here is that successive episodes of late-Holocene ice advance attained progressively smaller extents. The terminal moraine of Drangagil stage Sólheimajökull is located ~6 km forward of the present terminus, whereas the Little Ice Age moraines are just 1 km or less in advance of the present terminus. This is a major puzzle, because the overall trend was for climate to get colder as the Neoglacial progressed. The history of neighbouring outlets matches this trend: Klifurárjökull, Seljavallajökull and Steinholtsjökull all reached their late-Holocene maxima in the Little Ice Age at the time greatest cooling is inferred; this corresponds to a fall in regional ELA of ~400 m. (Gígjökull behaves differently again: see Chapter 8.) It is extremely unlikely that this anomalous out-of-phase behaviour of Sólheimajökull can be explained by regional contrasts in climate. Indeed, it is important to point out that the timing of episodes of ice advance identified at Sólheimajökull by Dugmore reflect an Iceland-wide climate signal, even if their extent does not: the Drangagil, Hólsárgil and Ystagil stages correlate with episodes of advance by small corrie glaciers of the Tröllskagi Peninsula, west of Akureyri in north Iceland (Stötter, 1991, cited by Guðmundsson, 1997). This rules out interpretations of ice advance at Sólheimajökull which rely entirely on non-climatic factors: e.g. the possibility of major surges associated with eruptions of Katla. It is clear that climate change must trigger the unusual advances of Sólheimajökull, but some other factor(s) must intervene to regulate the extent of these advances.

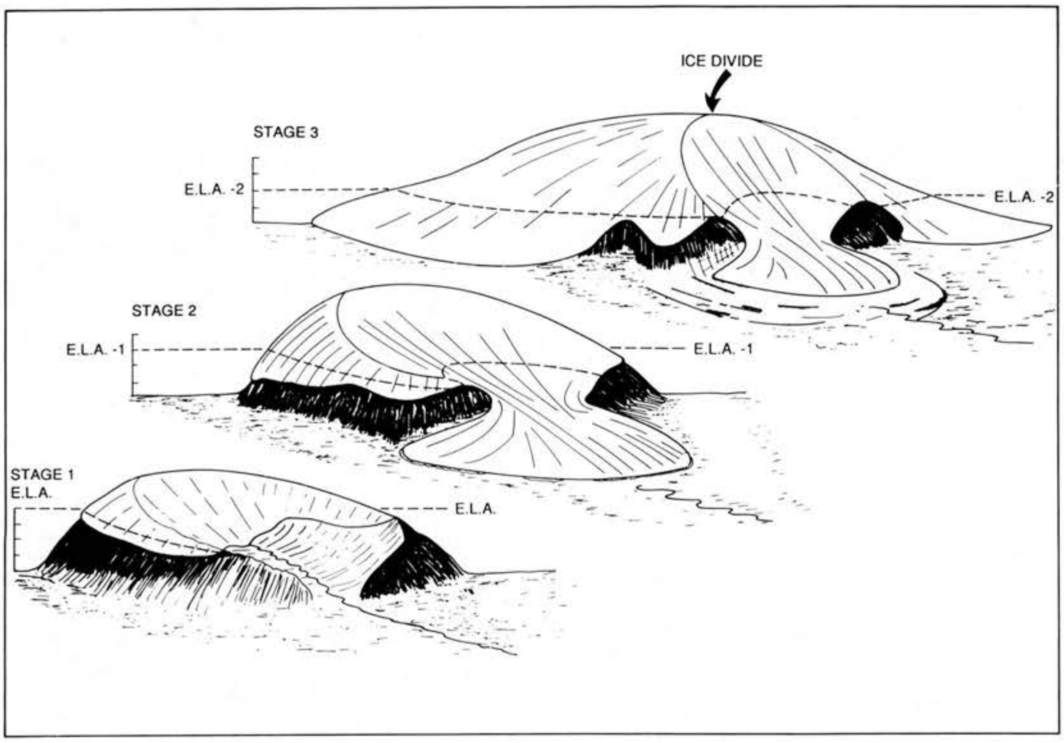


Figure 3.3

Idealised model of the effects of ice-divide migration (from Dugmore and Sugden, 1991). The glacier forms in a basin in Stage 1. With a falling ELA, ice from the whole basin flows strongly downvalley (Stage 2). In Stage 3 the ice-cap divide has migrated over the basin and diverted some ice; although the ELA is lower, there is less ice flowing downvalley.

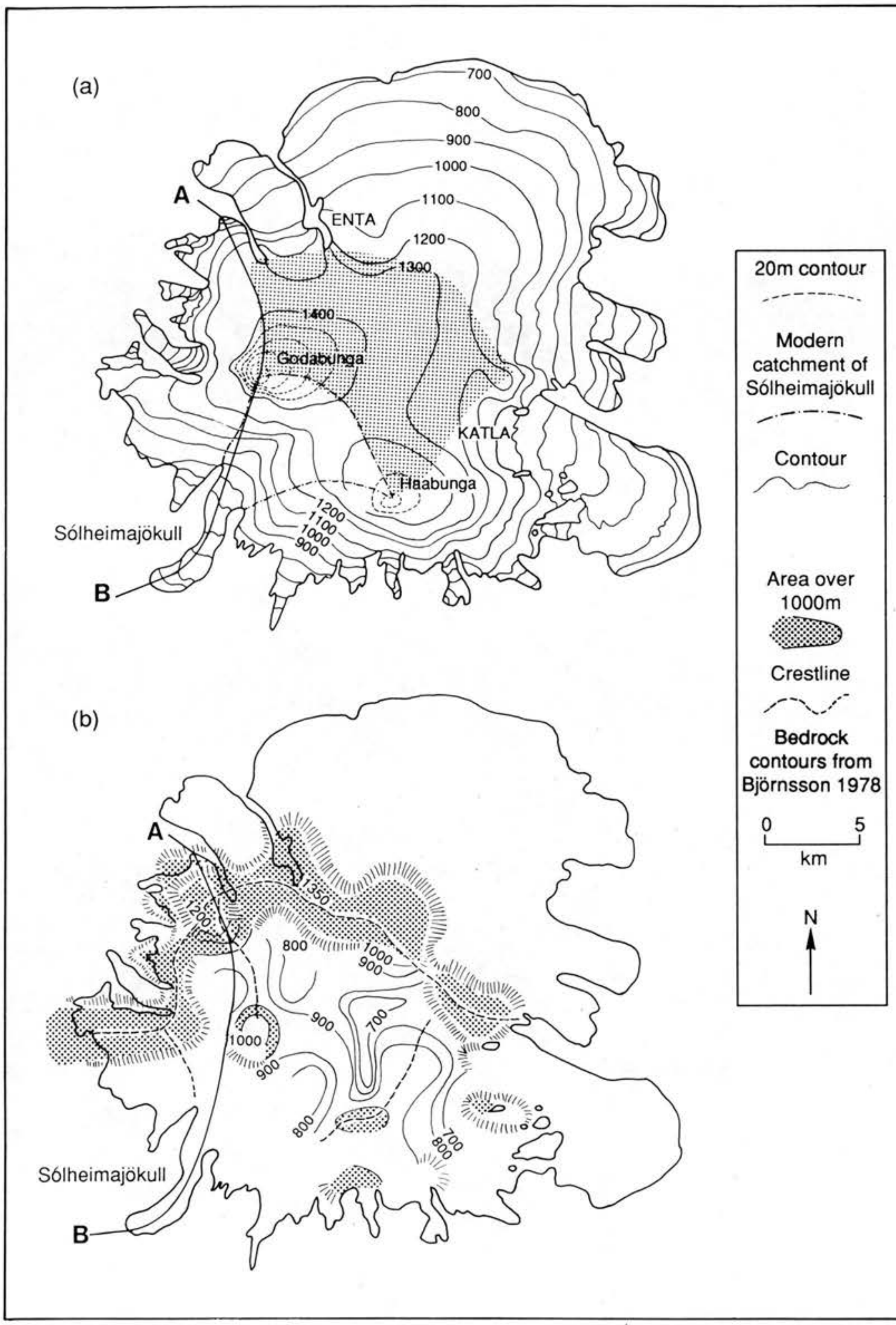


Figure 3.4

Ice surface and bedrock topography of Mýrdalsjökull (from Dugmore and Sugden, 1991). Top (a): surface contours, showing that the present ice-divide of Sólheimajökull is located towards the south-west of the ice-cap. The area above the crater in which the ice-divide is potentially free to migrate is shaded. Bottom (b): bedrock topography, showing rim and floor of the crater. N.B. the bedrock topography depicted here for the southern rim is too low (Lawler *et al.*, 1996: new ice-radar survey). The bedrock bump beneath Goðabunga is ~1,400 m high, that beneath Haabunga is ~1,200 m high.

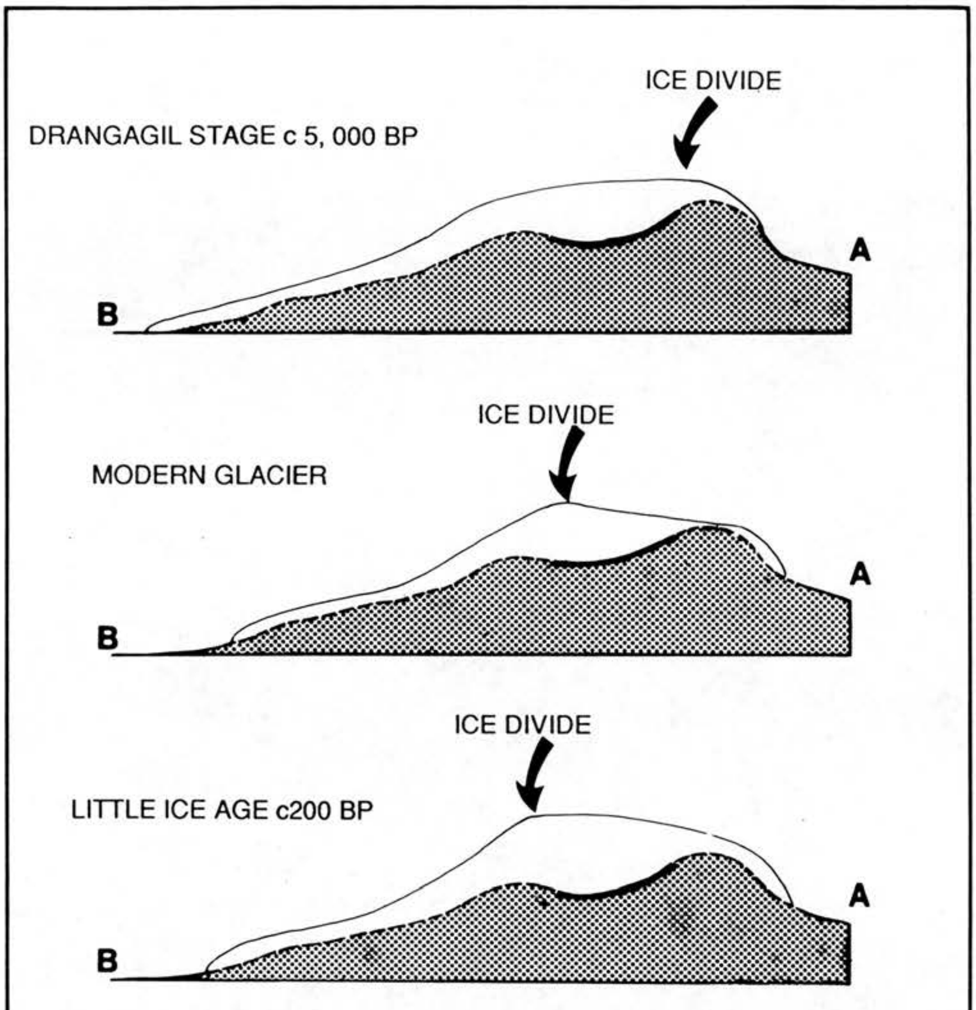


Figure 3.5

Schematic longitudinal profiles of Sólheimajökull at different times (from Dugmore and Sugden, 1991). Line A-B is shown on Fig. 3.4. Top: Drangagil stage (mid-Holocene); middle: present-day; bottom: Little Ice Age. See text for explanation.

Dugmore and Sudgen (1991) argue that ice-divide migration is this missing factor. Their hypothesis is simple: in the early part of the Neoglacial, the catchment of Sólheimajökull was much larger than it is at present, so cold episodes give rise to major ice advances; progressive reduction of catchment size as the Neoglacial proceeds accounts for the tendency for successive episodes of ice advance to fall in extent (Fig. 3.3). The hypothesis of ice-divide migration is realistic because the upper edge of Sólheimajökull's catchment is not constrained by topography (Figs 3.4 and 3.5). The wide summit dome of Mýrdalsjökull sits above the Katla caldera; the area inside the crater rim represents a 'battleground' within which several different outlets compete to capture snow-fall and so ice flow. Sólheimajökull's chief opponents in this respect are Entujökull (north), Sandfellsjökull (east) and Hofðabrekkujökull (south-east). The exact location of the ice-divides and the partition of flow between the different outlets depends upon interaction between the distribution of snow-fall (mass gain) and the different levels of activity (mass loss) of competing outlets.

It is possible that in the early Neoglacial Sölheimajökull was fed by a catchment much bigger than today's if its ice-divide was located above the major bedrock divide of the crater's northern rim (ice radar surveys by Björnsson show that this is generally higher and more continuous than the crater's southern rim). The extent of the Drangagil advance is consistent with this enlarged catchment coincident with a relatively mild climate similar to that which exists at present. Dugmore and Sugden (1991) suggest that as cooling and ice growth continued, progressive shift of the ice-divide towards the south-west occurred because the enlarged ice-cap became increasingly proficient in its capture of snow-fall from the heavily moisture-laden, onshore south-westerly winds. In this way, as the total ice volume of Mýrdalsjökull rose, the catchment area of Sólheimajökull fell, so explaining why the Little Ice Age maximum cold episode coincides with such a restricted advance by Sólheimajökull. Rudimentary reconstructions of ice limits using best-guess values for catchment sizes, AARs and ELAs give firm support to this hypothesis. Ice-divide migration affects only those glaciers which drain the central crater: topographically-constrained outlets such as Klifurárjökull which drain the outer slopes of the volcano are not affected, and so respond to climate change in a 'normal' fashion.

3.3 ICE-DIVIDE MIGRATION: REAPPRAISAL AND REFINEMENT

It seems to me that the extent of the Drangagil advance means that Dugmore and Sugden's ice-divide migration hypothesis must be broadly correct: unless we permit the catchment area to vary substantially, it is difficult to imagine how, given the likely climate ~5,000 BP (i.e. not dissimilar to today's), Sólheimajökull can have advanced so far. Here, however, I aim to 'get behind' the simple, functionalist nature of the AAR/ELA calculations used to reconstruct the likely past states of Sólheimajökull, and try to think through by what processes such changes in glacier geometry might have been achieved. It is clearly false to claim that this reasoning is independent of my hunch that some kind of change in basal hydrology must have occurred. Nevertheless, I believe that the story which follows is consistent with existing theory, and provides a plausible account of Sólheimajökull's Holocene activity which simultaneously strengthens Dugmore and Sudgen's ice-divide migration hypothesis, and supplies insight into how possible changes to the subglacial flushing regime might explain the observed contrasts in moraine development.

ACCUMULATION AREA RATIO/EQUILIBRIUM LINE ALTITUDE (AAR/ELA) TECHNIQUES

The strength of the AAR/ELA technique of past glacier reconstruction lies with its simplicity: namely the principle that the ratio of the area of snow accumulation necessary to balance a given area of ice melt is fixed in time and space. The convenience and popularity of this simple method is supported by its proven performance (see Torsnes *et al.*, 1993, for a recent evaluation of different geometric methods of glacier reconstruction in which the AAR/ELA technique works best). However, the heavy use and success of the technique in part reflects the fact that it tends to be applied to glaciers which match closely the conditions on which the technique implicitly relies, in situations which prioritise the 'what' happened?, but have little need to ask 'why?' and 'how?'. The current example of Sólheimajökull does not meet these criteria. AAR/ELA techniques tend to work best *if*:

- The glacier's upper limit is fixed.
- The ice mass is rectangular, and underlain by even bedrock topography.
- The mass balance and area-altitude relationships are uniform and linear.
- Flow dynamics are uniform.
- All these relations stay fixed in relation to each other as climate changes.

It is likely that many alpine-type valley glaciers resemble such an ideal ice mass reasonably well; however, it is highly unlikely that an outlet glacier with a history and situation as complex as Sólheimajökull fits this picture. In this kind of 'deviant' situation, AAR/ELA techniques can provide important clues (in this case on how the catchment area of Sólheimajökull must have changed), but they cannot give much insight into processes which lie at a deeper explanatory level (see Chapter 1.4).

Sólheimajökull: complicating factors

Bed topography. Recent ice radar surveys confirm that the summit zone of Mýrdalsjökull lies above a deep, asymmetrical and topographically complex caldera, with a volume of 110 km^3, associated with the Katla volcanic system (Björnsson *et al.*, 1995; Lawler *et al.*, 1996). The picture of subglacial topography used by Dugmore and Sugden - reproduced in Fig. 3.4b - is broadly correct, although the bedrock bumps beneath Goðabunga and Haabunga are higher than indicated: ~1,400 m and ~1,100 m respectively. This type of topography - notably the absence of a central bedrock divide able to fix the position of the ice-divide - permits significant variations in flow behaviour to occur, which creates difficulties for the use of AAR/ELA techniques to study outlet fluctuations. Computer models by Payne and Sugden (1990) show how large topographic basins similar to the Katla caldera control the time-space pattern of ice-cap growth in a way which gives rise to non-linear response of outlet glaciers, such as those draining Mýrdalsjökull (see below).

Basal boundary conditions. A second factor likely to disrupt the simple relationship between climate and ice extent which must be considered is the possibility that a change in basal conditions alters the ease with which ice flows. This is currently an important issue in process-based studies of glacier fluctuations at a wide range of scales. Surge activity of valley glaciers is an obvious example (e.g. Sharp, 1988); a second is MacAyeal's (1993) binge-purge model of the Hudson Bay lobe of the Laurentide Ice-Sheet, associated with cyclical thaw of a soft bed which generates surge episodes which discharge the icebergs of Heinrich events. The point I wish to make here is not that ideas such as this are correct, but that they exist, are plausible, and must be taken seriously. Indeed, MacAyeal's model seems to be deficient because it underestimates the importance of climate as a trigger of Heinrich events; evidence of synchronous episodes of ice-rafting from the Laurentide, Greenland and Icelandic Ice-Sheets (Bond and Lotti, 1995) implies a climatic trigger which affects the entire North Atlantic area, even if the exact response reflects the internal workings of the ice-cap system. This coupling of climatic trigger with response controlled by the internal workings of the ice-cap system (plus the possibility of feedback effects) is exactly what is inferred/required to explain the Holocene behaviour of Sólheimajökull.

Factors such as iceberg calving or abrupt activation of deformable beds are not thought to have affected Sólheimajökull in the Holocene. However, changes to the volume, routing or flow configuration of subglacial meltwater likely to bring about time- and space-dependent changes in ice dynamics are a realistic possibility. The AAR/ELA technique cannot resolve satisfactorily this kind of change. Changes in ice geometry, or in the level of the geothermal heat flux, can induce a switch in meltwater activity (which is likely to influence flushing of debris as well as ice flow). Björnsson (1988) discusses the relative control ice surface and bedrock topography exert on subglacial drainage in areas of elevated geothermal heat flux, such as the Katla massif. Blankenship *et al.* (1993) consider how enhanced basal meltwater production in the volcanic West Antarctic rift system helps to sustain ice-stream activity; Sturm *et al.* (1991) argue that post-1965 advance of glaciers in the Wrangell Mountains, Alaska, USA relates not to climate change, but to the impact of greater volcanic heat flux on flow processes. In my view, ideas such as these justify the following analysis. It seems crucial to consider the potential impact of changing ice geometry, subglacial geothermal activity and bedrock topography on sliding intensity. I explore this possibility below, and conclude that there exists a strong case to infer that changes in sliding intensity regulated the response of Sólheimajökull to simultaneous change of climate and catchment area. The expected change in flushing efficiency appears as a by-product of this.

REAPPRAISAL OF THE AAR/ELA CALCULATIONS

Here I re-work the AAR/ELA calculations. My intention is not to endorse the technique, but to illustrate and explore possible past states of Sólheimajökull. Different combinations of values suggest alternative mass balance regimes which I use to structure my discussion on possible past process behaviour.

Dugmore and Sugden (1991) used Björnsson's (1979) benchmark figures for southern Iceland to explore their hypothesis of ice-divide migration: they assumed that 1) present-day ELA is 1,100 m, and, b) this corresponds to an AAR of 1.7:1, which stays fixed as climate changes. These figures imply that the extent of the Drangagil advance is consistent with a limited fall in ELA relative to the present (minus 100 m or less) if the ice-divide at the time was located above the major bedrock divide of the crater's northern rim. If this idea is correct, it explains the paradox of the Drangagil advance at a time of supposedly mild climate.

My calculations (using the Landmælingar Íslands 1:100,000 special sheet: 'Þórsmörk and Landmannalaugar') fail to reconcile the values of ELA and AAR Björnsson suggested. If the present-day regional ELA is 1,100 m, then I calculate the AAR of Sólheimajökull to be ~2.1:1; conversely, if I accept that the AAR of 1.7:1 is correct, I get an ELA of ~1,200 m. This raises the important issue of time-space deviations from supposedly characteristic parameters. Indeed, if such deviations can be inferred with a reasonable degree of confidence - as I try to do here - they provide a valuable guide as to what processes underlie episodes of climate-glacier change. Here, I prefer to retain the regional ELA of 1,100 m, and take the AAR of 2.1:1 this implies as an artefact of glacier hypsometry. This is consistent with the ideas of Furbish and Andrews (1984). The gentle gradients of the wide summit dome of Mýrdalsjökull (i.e. above ~1,300 m) create a large increase in area for each unit increment of altitude: a significant deviation from the uniform area-altitude distribution the use of simple AAR/ELA techniques implicitly assumes. In turn, this suggests that a unit increase in glacier catchment area provides an increment of snow-fall smaller than that which might be expected from 'standard' altitude-area-accumulation relationships (i.e. the glacier adds area quickly, but snow-fall slowly, as altitude increases, because quantity of snow-fall usually scales directly with altitude). It seems reasonable to anticipate that an ice mass such as Mýrdalsjökull which has a wide, flat and unconfined accumulation area will need a relatively high catchment area to sustain an outlet glacier of a given extent: i.e. it will have a high AAR.

Mýrdalsjökull and Sólheimajökull: mid-Holocene

Plausible catchment size/mass balance states which give rise to the Drangagil advance include (Fig. 3.6):

a. **Sólheimajökull's ice-divide coincides with the crater's northern rim.** If the best-guess present-day ELA of 1,100 m is used, this makes the AAR 2.44:1. This is high, but it is within the range thought consistent with steady-state conditions (between 4:1 and 1:1: Torsnes *et al.*, 1993). Given the likely glacier hypsometry at the time (i.e. area rises rapidly with height), plus the possibility that snow-fall diminished with height and distance inland (i.e. exhaustion effects) it is not grossly unrealistic.

b. **The ice-divide of Drangagil stage Sólheimajökull did not coincide with the crater rim.** If the AAR is set at 2.1:1, with the ELA at 1,100 m the ice-divide shifts south-westwards. If this shift is uniform, the ice-divide swings in a wide arc from north to east, and lies within ~800 m of the bedrock divide.

c. **The ice-divide coincides in part with the crater's northern rim, but sits above the bedrock ridge at 850-900 m which divides the crater into west and east basins.** Relative to scenario a), 44 km^2 of Sólheimajökull's catchment are lost to Hofðabrekkujökull. Repeat calculations with this new ice-divide give: i) AAR = 1.7:1,

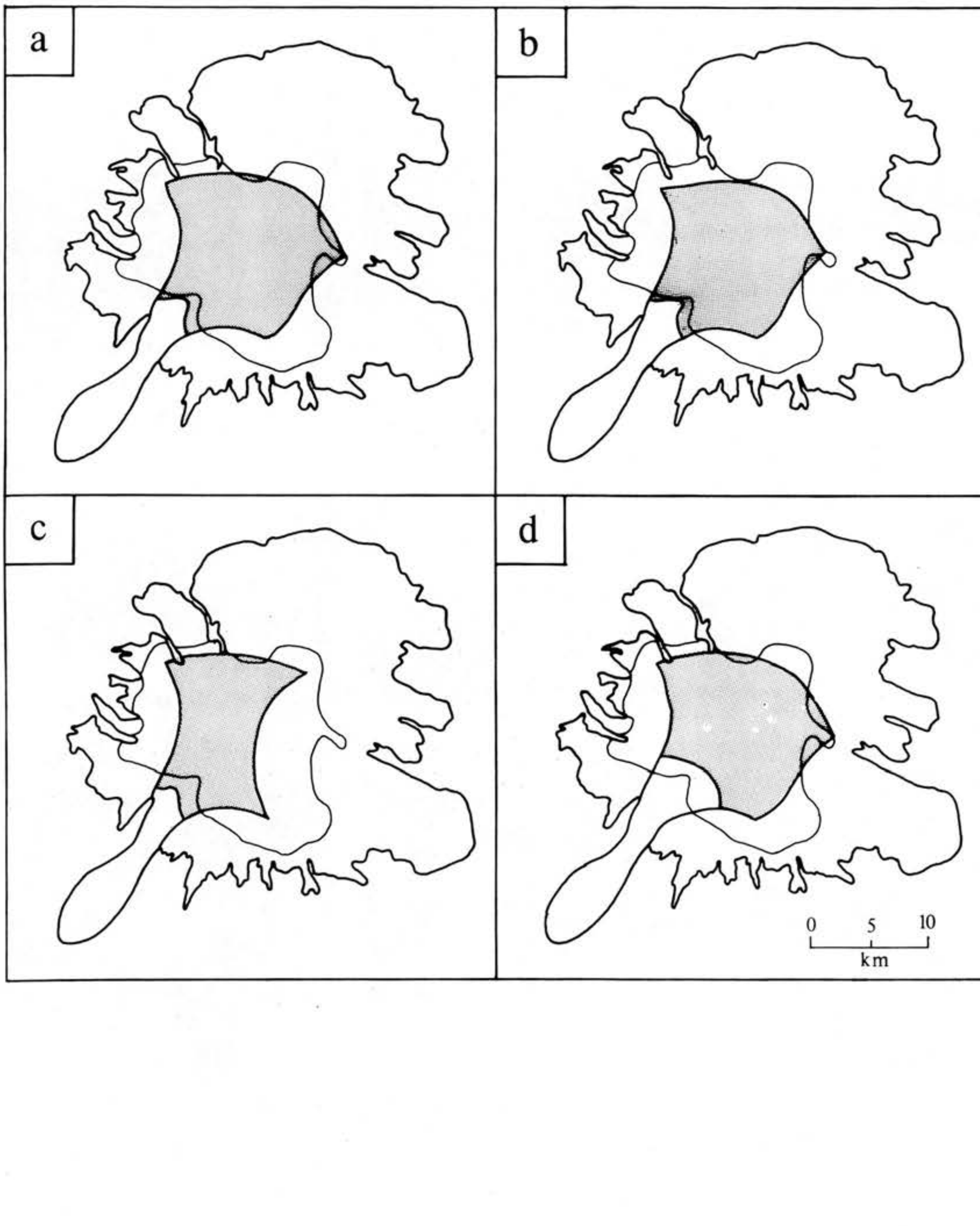


Figure 3.6

Plausible catchments of Drangagil stage Sólheimajökull. Shading indicates extent of the hypothetical accumulation area; the thin line is the present-day 1,200 m contour, shown for reference.

a. ELA 1,100 m, AAR 2.44:1.
b. ELA 1,100 m, AAR 2.1:1.
c. ELA 1,000 m, AAR 1.7:1.
d. ELA > 1,300 m, AAR 1.7:1.

See text (pp. 132-133) and Fig. 3.4 for further details.

ELA = 1,000 m, or, ii) AAR = 2.1:1, ELA = 800 m. The first is consistent with the magnitude of climate change Dugmore and Sugden anticipate, but implies a much-reduced degree of ice-divide migration; the second implies that the speed and intensity of climate change at the start of the Neoglacial was far greater than thought: i.e. ELA falls by 300 m relative to present, whereas the Little Ice Age peak fall is 400 m.

d. **The ice-divide coincides with the crater's northern rim, but with an AAR of 1.7:1.** This gives an ELA in excess of 1,300 m. This scenario implies that the crater must have been full of ice at the start of the Neoglacial. If Katla was ice-free at the time climate started to deteriorate then the limited areas of bedrock above 1,300 m must first have nourished relatively small glaciers which would not fill a large crater sufficiently quickly to feed major outlet glaciers.

These examples illustrate two things:

1. Some measure of ice-divide migration (i.e. change in catchment size) must be invoked as part of any likely explanation for fluctuations of Sólheimajökull over the last ~5,000 years.

BUT

2. Although it illustrates a *qualitative* principle, the AAR/ELA technique cannot be used with confidence to constrain *quantitatively* the likely range of ice-divide migration.

It is possible that the change in climate was greater, and the extent of ice-divide migration less pronounced, than Dugmore and Sugden assume/imply. It is for this reason that I choose to use a parallel approach to investigate further the likely history of glacier activity: one which draws on process-reasoning to make up for the limitations of the AAR/ELA technique. The AAR/ELA technique implicitly assumes the existence of spatially and temporally invariant relationships between catchment area, altitude and distribution of snow-fall and melt, derived from an arbitrary empirical standard. Its use compares specific instances of glacier extent and associated climate (as measured by the ELA), separated by a given time interval: i.e. it compares states A and B, usually with the AAR taken as fixed, with the ELA as the target output. The technique ignores the time-dependent sequence of changing forms and processes, modified by feedback loops, by which the system is transformed: i.e. what happens between states A and B, and how and why this happens. It is this behaviour which I consider below.

GROWTH OF MÝRDALSJÖKULL

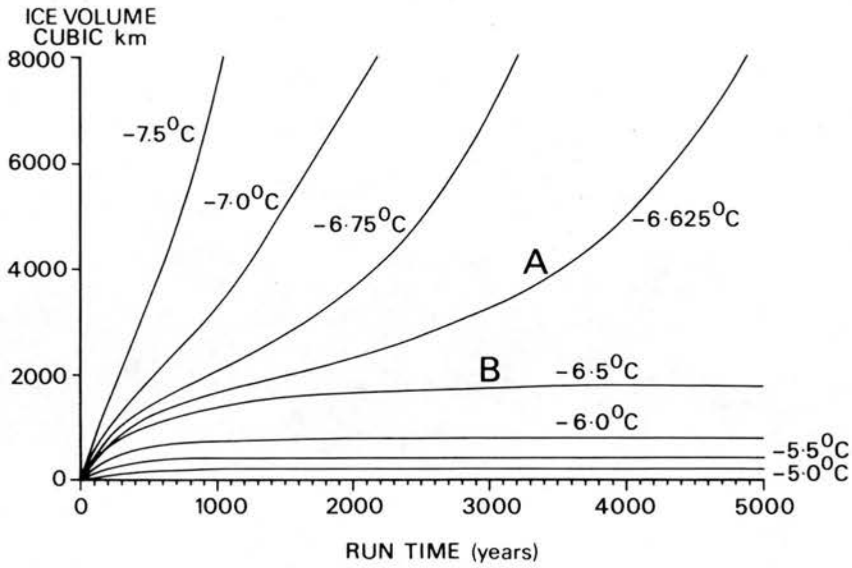
Dugmore and Sugden's simple model of ice-divide migration (Fig. 3.3) seems to imply that ice-divide migration occurs *after* the ice-cap is established as a major feature which sits above the major bedrock divide on which it grew up *de novo.* At the time the paper was written, such a scenario was justified by prevailing knowledge: a) it was widely thought that deglaciation of Iceland was pretty-much complete, so Katla was expected to be largely, if not completely, ice-free; and, b) ice radar surveys implied that the Katla basin was relatively open to the south-west, and backed by a major bedrock ridge to the north. New data have revised these points: a) it is now thought that significant ice cover has persisted throughout the Holocene (Guðmundsson, 1997), so that small glaciers did exist on the crater rim - possibly even a full-scale, if small, ice-cap - at the start of the Neoglacial; and, b) new ice radar data (Björnsson *et al.*, 1995; Lawler *et al.*, 1996) show large areas of high ground exist in the south-western part of the Katla caldera. These data imply that 1) the process of Neoglacial ice growth began at a more advanced stage; and, 2) the processes of ice growth and ice-divide migration cannot be thought of as essentially separate, but must be considered together as a single phenomenon.

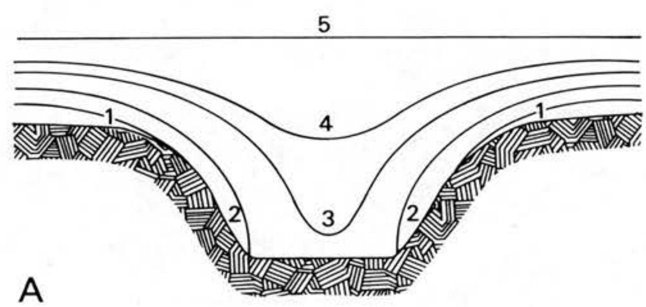
Before a major outlet can develop, ice must fill the crater. Payne and Sugden (1990) study a similar problem of basin infill by ice with a climate-ice flow model of the Loch Lomond Stadial in Scotland which reconstructs the pattern of ice growth at Rannoch Moor. The initial impact of a deep topographic basin is to delay ice growth: the basin acts as a heat sink which consumes ice faster than ice can fill it. If, however, temperatures fall below a certain threshold, positive feedback (associated with increased altitude, extent and albedo of the accumulation area surface) induces a switch from localised mountain glaciation to development of a large ice-cap with major outlets. Payne and Sugden's model highlights two pertinent factors (Fig. 3.7):

1. It takes a long time to fill up a deep basin with its floor at relatively low altitude. (It takes >2,000 years with a temperature depression relative to today of nearly 7°C (equivalent to a fall in ELA of ~1,000 m) to fill Rannoch Moor to the point at which it feeds major outlets.)

2. Interaction of the convergent ice masses which fill the basin builds up an ice dome which has its ice-divide *midway* between the basin's bedrock divides.

It seems safe to expect that the late-Holocene growth of Mýrdalsjökull was *qualitatively* similar to the pattern Payne and Sugden's model predicts. Rannoch Moor and the Katla caldera are of similar size, both display relief of 600-700 m, and both occupy windward, maritime locations. However, Mýrdalsjökull was probably part-developed as an ice-cap when the Neoglacial began, whereas Payne and Sugden's model simulates ice accumulation and basin infill from scratch.





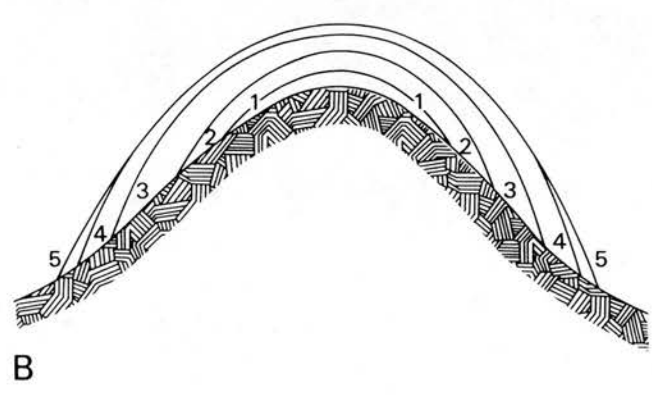


Figure 3.7

Ice mass growth: A) in an enclosed basin; and, B) above a ridge (from Payne and Sugden, 1990). Two points are important: 1) although it is possible to build-up thick ice caps in deep basins, it takes a long time to do this (~1,000-2,000+ years in this example); and, 2) once it takes shape, the ice-divide of this ice-cap will tend to stand in the middle of the basin.

This is likely to be part-offset by the fact the Younger Dryas cooling of the North Atlantic was several times that expected of the late-Holocene in Iceland.

I expect that growth of Mýrdalsjökull in the mid-to-late-Holocene was roughly symmetrical, much as Payne and Sugden's model envisages. This is important, because marked ice-divide migration, as Dugmore and Sugden envisaged, requires the initial presence of an *asymmetrical* ice-cap. I infer that uneven accumulation is restricted just to the early stages of ice growth: glaciers first appear on the continuous and consistently high (~1,350 m) bedrock ridge which forms the crater's northern rim. Glaciers which develop to the south of the nascent ice-divide drain to the deepest part of the crater (floor below 700-800 m), which acts as a heat sink with relatively high ablation.[3] This heat sink limits the speed with which these glaciers are able to expand and start to fill up the crater, so asymmetrical ice-cap growth is suppressed. However, it is feasible that these early stages of ice-cap growth are not relevant to Neoglacial ice expansion if a sizeable ice-cap was in place at the start of cooling.

Although its highest parts match the height of the northern rim, the crater's southern rim is more broken, so I anticipate limited glacier development as the ELA first falls to meet bedrock. However, as the ELA falls further - say, below ~1,200 m - I expect that ice growth to the south catches up quickly. This is because of the irregular topography of the southern part of the crater; relative to the northern part it has reduced relief and comparatively high mean altitude favourable to ice growth. I infer that substantial quantities of southern ice, pushing north, must develop before ice emanating from the north can overwhelm the entire crater to push large outlets south. This means that infill of the crater is relatively symmetrical, so interaction between convergent ice produces an ice-divide situated above the middle of the crater, which reduces the catchment area available to feed Drangagil stage Sólheimajökull. Two further factors support this inference:

- The possibility that ice 'leaks' out of the crater to escape south-eastwards in the direction taken today by Hofðabrekkujökull.
- The possibility that the topographically-induced change in the distribution of snow-fall/mass balance which drives ice-divide migration starts before inundation of the crater is complete, so enhancing accumulation to the south of the crater at the expense of the north.

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[3] Two further enticing possibilities are ablation enhanced by a) an elevated geothermal heat flux and/or volcanic eruptions (cf. Björnsson, 1983), and, b) mass loss by calving into a crater lake.

Evaluation

I think it is reasonable to conclude that the ice-divide of Drangagil stage Sólheimajökull did not coincide with the major bedrock divide of the crater's northern rim. My best-guess scenario closely resembles scenario c), above. If we allow i) Sólheimajökull to drain half the crater (currently it drains approximately one-fifth), and, ii) ELA to fall to 1,000 m, which preserves the best-guess present-day AAR of ~2.1:1 then it transpires that this hypothetical Sólheimajökull falls ~3 km/7.5 km^2 short of the extent indicated by Dugmore's (1987) reconstruction of the Drangagil limits. This mis-match can be made up by changes to the assumed catchment area or AAR, or by keeping the same figures, but letting ELA fall to 800 m. If climate deterioration was faster, and of larger magnitude than anticipated, it is not necessary to try anything too 'fancy' to explain the Drangagil advance (even so, the scenario here still gives Sólheimajökull a much bigger catchment area than it has today). However, I wish to continue by exploring the potential impact at the time of subglacial drainage on basal boundary conditions and ice flow. These process considerations are not tied to the catchment area/ice extent calculations: I expect these processes to affect Sólheimajökull as long as it drains part of the crater, although the *intensity* of these processes is likely to alter with shifts in the location of the ice-divide and the centre of volcanic activity. However, it is possible that the impact of these processes contributes to variations in catchment size which explain both the extent of the Drangagil advance, and changes in the intensity of flushing.

3.4 GEOTHERMAL ACTIVITY, TOPOGRAPHY AND SUBGLACIAL HYDROLOGY

The crater is an area of elevated geothermal activity (Lawler *et al.*, 1996) which escalates at intervals to give volcanic eruptions and/or *jökulhlaups* (e.g. Katla, 1918: Tómasson, 1996). The geological record confirms that this activity has been a persistent feature throughout the Holocene. I expect high, geothermally-enhanced rates of basal melting within the crater beneath the catchment of Drangagil stage Sólheimajökull (cf. Björnsson, 1983), the more so because it is thought that the centre of activity, presently located to the east underneath Katla (i.e. at the head of Hofðabrekkujökull), was found further to the south-west (i.e. towards Sólheimajökull) in the mid-Holocene (Einarsson *et al.*, 1980). Meltwater which collects at the base of the ice will be driven upslope and out of the crater if 'Shreve's criterion' is met: water under pressure flows upslope if the backwards slope of the bedrock does not exceed eleven times the forwards slope of the ice surface (Shreve, 1972; Björnsson, 1988; Chapter 1.1). The summit dome of present-day Mýrdalsjökull is characterised by extremely gentle slopes ($<2^{\circ}$), but it is likely that the Shreve criterion was met in the mid-Holocene: mean bedrock slopes in the south of the crater are less steep than those of the north, and the overlying ice surface sloping south was likely to be steeper than it is today if the ice-divide was farther to the north (cf. Fig. 3.5). This geometry is inferred to pump substantial quantities of crater meltwater towards and below mid-Holocene Sólheimajökull, most likely within some kind of distributed drainage network. This is because 1) water produced by dispersed basal melting will begin its journey as part of a distributed network; and, 2) high water pressures tend to render conduit drainage upslope unstable. This instability (collapse of conduits) takes effect if adverse bedrock slopes exceed ice surface slopes by a factor of 1.3-2.0 (Shreve, 1985; Röthlisberger and Lang, 1987), as is likely with water which drains a gentle ice dome which covers a deep crater (see Chapter 5.6 for more details of water flow in overdeepenings, of which this crater is a special instance). This implies that 1) water which starts at the glacier bed as part of a distributed system cannot join up into conduits; and, 2) if water from the surface reaches the crater bed in conduits, these conduits will tend to collapse and feed water into the adjacent distributed system.

ENHANCED SLIDING AND GLACIER ADVANCE?

Some kind of distributed network preserved beyond the crater rim beneath the trough-part of Sólheimajökull perhaps explains the extent of its Drangagil stage advance. Large volumes of water under high pressure widely spread over a glacier's bed can speed up basal sliding for a given level of basal shear stress (e.g. Iken and Bindschadler, 1986; Blankenship *et al.*, 1993); if sustained, faster sliding can cause ice to advance (e.g. Kamb *et al.*, 1985; Shoemaker, 1992). It is possible that the Drangagil advance represents a permanent 'mini-surge' of Sólheimajökull driven by favourable subglacial drainage which pushes the overlying ice forwards. Simultaneously, the wide reach of subglacial waters implicated in this scenario explains the increase in flushing efficiency and smaller moraines.

If such enhanced sliding is to be sustained, the integrity of the distributed network must be preserved (low water pressures in conduits acting over a small area of the bed are unlikely to favour sliding). Bedrock of the trough floor which slopes with the ice surface offers little protection to distributed drainage. Ice melt by running water and/or the leverage exerted by high water pressures tends to destabilise distributed flow and force its collapse into conduits, unless forces tending to enlarge channels are balanced by forces tending to close them down. The likely geometry inferred for Drangagil stage Sólheimajökull suggests that this balance of forces favours (quasi?) stability of distributed drainage:

- Advance of Sólheimajökull by ~6 km relative to today reduces the mean ice surface slope/mean hydraulic gradient by approximately one-third (see below). This lessens the overall rate at which water flow releases heat to enlarge ice-walled channels.
- High water pressures, coupled with marked convergence of ice flow as it exits the crater to enter the outlet trough, favour high sliding speeds. Fast sliding helps to preserve distributed drainage because:
  a) Advection of ice into potentially unstable channels/orifices counteracts the tendency for enlargement by roof-melt.
  b) Rapid sliding drives incipient conduits into contact with the stoss faces of bedrock obstacles, whereupon enhanced plastic deformation in response to elevated stresses tends to close conduits up.

Kamb's (1987) stability parameter (Ξ), which describes the threshold at which a linked cavity system tends to collapse (i.e. Ξ >1.0) rises directly with the 1.5 power of hydraulic gradient (increased instability), and falls (increased stability) as a function of the square root of sliding velocity.

GLACIER DOWN-DRAW AND BACK-WASTAGE?

This enhanced sliding hypothesis is not inconsistent with the ice-divide migration hypothesis: an enlarged catchment which taps water and ice from the crater is required to start the process. However, it does raise the possibility that a much-enlarged catchment (as in Dugmore and Sugden's original hypothesis) appears as effect, not cause, of ice advance, with northwards retreat, not southwards advance, of the ice-divide with time! My argument for growth of a relatively symmetrical ice cap and reduced catchment area of Sólheimajökull (which possibly 'under-explains' the extent of the Drangagil advance) rests on the assumption that ice flow either side of the ice-divide is even; an imbalance in ice flow will carry ice away faster in one direction than in the other, and so modify the position of the ice divide. This is what I infer to have happened here.

I envisage a process of fast sliding and extended ice advance sustained by a progressive increase in glacier catchment area achieved by down-draw of ice and active back-wastage of the ice-divide. It is feasible that this process eventually reaches relative stability to give a glacier still-stand - as indicated by the moraines of the Drangagil stage? Ice advance by enhanced sliding can be sustained only if Sólheimajökull is able to tap fresh stores of ice to compensate for the increased losses in its enlarged ablation area: i.e. it must enlarge its catchment area. If a glacier is confined by bedrock at its upper limit this is impossible, but if - as with Mýrdalsjökull - catchment limits are defined by ice-divides, catchment area is in part subject to definition according to the relative activity levels of neighbouring glaciers: in this case Sólheimajökull, Hofðabrekkujökull, Entujökull and Sléttjökull. Faster sliding carries mass downglacier, the likely effect of which is to cause the ice surface slope behind the zone of enhanced sliding to steepen. Steeper ice raises the basal shear stress and promotes faster flow; longitudinal stress transmissions pulling ice forwards are also possible. This creates the potential of a positive feedback loop, whereby a zone of steeper ice surface slopes and faster flow works backwards, propagating a process of surface down-draw upglacier. This down-draw progressively cuts back into the ice-cap, pushing back the ice-divide towards the north. Back-wastage of Sólheimajökull's ice-divide adds water and ice from the crater, likely both to sustain the enhanced sliding/down-draw loop, and to balance mass loss in the enlarged ablation area (cf. Alley *et al.*, 1994, who invoke a similar scenario of "water piracy" by Ice Stream B to explain the stagnation of Ice Stream C). If correct, it is possible that a small episode of advance with faster sliding, triggered by climate change, or perhaps volcanic activity, can give rise to a much-enlarged catchment with a big ice advance as the end product.

This idea of surface down-draw and back-wastage of the ice-divide is not just idle speculation on my part. It does have a respectable academic pedigree: e.g. Eyles and McCabe (1989) invoke it to explain the complex geological record, indicative of rapid ice mass loss, fast ice flow and rapid delivery of debris, of the Irish Sea Basin; Bindschadler *et al.* (1996) and Bindschadler (1997) argue that reduced basal friction and faster flow of the West Antarctic Ice-Streams can instigate headwards migration, capture of ice-stream 'headwaters' (headices?) by neighbouring ice-streams of greater aggression, and - possibly - catastrophic collapse of the interior of the West Antarctic Ice-Sheet! (See also Alley *et* al., 1994.) Thus it is conceivable that this process of enhanced sliding explains the combination of limited climatic change, marked advance of Sólheimajökull, and enlarged catchment area in a way which refines and strengthens Dugmore and Sugden's hypothesis of ice-divide migration.

THE LITTLE ICE AGE

The challenge of the Drangagil stage is to provide a convincing explanation of anomalously large ice advance, at a time of relatively mild climate, which is consistent also with high efficiency flushing of debris beneath an erosive glacier. The Little Ice Age presents the reverse problem: i.e. how to explain an anomalously small ice advance, at a time of relatively harsh climate, which accounts also for relatively pronounced moraine development thanks to relatively inefficient flushing of debris beneath a glacier inferred to be less erosive? Not surprisingly, therefore, my preferred solution largely involves reversing the logic of the early Neoglacial scenario.

The moraine record indicates that at its greatest Little Ice Age extent (*c.* 1750 AD: Andy Dugmore, personal communication) Sólheimajökull was ~3 km^2 further advanced than it is today. Dugmore and Sugden (1991) argue that at this time the regional ELA fell by ~400 m relative to its present-day value of 1,100 m (i.e. to 700 m). This fall of the ELA, taken with the limited ice advance, suggests several possibilities:

1. **The ice-divide coincides with its present position, with an AAR of 2.2:1.**
2. **The ice-divide shifts ~200 m to the south-west, with an AAR of 2.1:1** (i.e. best-guess present-day value).
3. **The ice-divide shifts ~1 km to the south-west, with an AAR of 1.7:1** (as used by Dugmore and Sugden). This scenario best fits Dugmore and Sugden's reconstruction of the glacier profile on the basis of the available geomorphic evidence (Fig. 3.5).

Ice-divide migration involves a change in the ice-cap's centre of mass; in this instance it shifts to the south-west. This requires that the south-western parts of Mýrdalsjökull gain mass relative to its northern and eastern parts. If this is to happen, mass gain in the south-west must increase, and/or mass loss fall; conversely, the north and east must experience a net loss of mass. Two things are likely to accomplish this:

- Change to the distribution of snow-fall: the south-west captures increased quantities of snow at the expense of the north, as Dugmore and Sudden infer.
- Reduction in the quantity of mass carried away from the ice-divide towards the south-west: i.e. the discharge of upper Sólheimajökull falls relative to outlets which share its ice-divide. It is possible that, as the centre of volcanic activity shifted towards the east, part of the catchment of Sólheimajökull was captured by active back-wasting of Hofðabrekkujökull stimulated by enhanced sliding, much the reverse of what I envisage was the case with the Drangagil stage Sólheimajökull (above). Such a 'see-saw' of mass between the two outlets is feasible: today Sólheimajökull and Hofðabrekkujökull share a ~12 km boundary which forms the major ice-divide of Mýrdalsjökull.

Beyond the AAR/ELA method

Simple use of AAR/ELA techniques implicitly assumes that the distribution of altitude/area/snow-fall remains fixed over time, in which case the AAR is unchanged also, as Dugmore and Sugden assume. If ice-divide migration occurs, in part driven by a change in the distribution of snow-fall, this is highly unlikely. If the centre of mass shifts south-westwards because the developing ice dome becomes increasingly proficient at scavenging snow-fall fom the south-westerly winds, as the total accumulation area of Sólheimajökull falls, each unit area which remains receives an extra increment of snow-fall. This implies that the total mass loss and terminus retreat which actually occur for a given reduction in accumulation area must be less than expected from use of an invariant AAR. Loss of area on its own does not provide a reliable guide to changes in glacier mass balance (Furbish and Andrews, 1984). This suggests that the neat fit to the Little Ice Age moraine record which Dugmore and Sugden arrive at using a fixed AAR appears to be fortuitous (although, to be fair, their calculations were designed to illustrate the idea of ice-divide migration; they were not intended to provide a quantitatively reliable description of Sólheimajökull's behaviour). The fit is correct, but the methods used to achieve this fail to account for the fall in AAR with time that the process of ice-divide migration seems to require. The simplest way to put this right is to take the changes to hypsometry and snow-fall distribution expected between mid- and late-Holocene into account, and allow the AAR to fall to reach 1.7:1 by the Little Ice Age, as I do: i.e. I prefer to use a higher value (2.1:1) both for the mid-Holocene and the present-day (see above), at which times I infer snow-fall distribution and ice dynamics to differ from those of the Little Ice Age.

A further factor disguised by the AAR/ELA technique is that changes to mass balance need not necessarily be taken up by changes to glacier plan-form area (i.e. advance or retreat). Ice can thicken or thin to satisfy mass balance requirements. Fig. 3.5, bottom, shows Dugmore and Sugden's reconstruction of Sólheimajökull's Little Ice Age profile, which I believe to be robust. In comparison with its mid-Holocene state there is little change to glacier relief, but the effect of ice-divide migration (summit dome shifts south) with northwards retreat of the terminus is to shorten Sólheimajökull considerably. As a result it is simultaneously both steeper and thicker. This suggests that part of the mass 'gain' (strictly reduced mass loss) associated with enhanced snow-fall is taken up not by amelioration of ice retreat, but by building up the glacier thickness. Steeper, thicker ice raises the basal shear stress. In turn this should give rise to faster flow, and a corrective tendency for Sólheimajökull to thin-out and push its terminus forward. Its Little Ice Age profile demands that the bed of Sólheimajökull supports a shear stress higher than that of the mid-Holocene: i.e. it requires a change in the basal boundary condition of ice flow. In effect, if it is useful to envisage Drangagil stage Sólheimajökull as a glacier in a state of permanent 'mini-surge', Little Ice Age Sólheimajökull represents the complementary quiescent phase in which mass builds up at the head of the glacier. To extend this parallel with surging glaciers: a change in subglacial drainage, which reduces the volume and/or pressure of water in contact with large areas of the glacier bed, offers itself as a potential explanation for diminished flow activity. Several factors support this:

- A colder climate with smaller catchment area reduces surface inputs of meltwater to the glacier bed.
- As the ice surface steepens, an increase in the hydraulic gradient driving water flow increases the probability that orifices linking cavities will collapse to form conduits.
- Input of basal meltwater from the crater progressively falls. This is likely if:
    a) The centre of maximum melt shifts eastwards (with possible impact on the activity of Hofðabrekkujökull - see above).
    b) The position of subglacial drainage divides moves. Water flow follows the ice surface slope (Shreve, 1972), so as ice-divide migration progressively reduces the area of the crater overlain by ice which slopes towards Sólheimajökull, its subglacial water catchment shrinks. If Dugmore and Sugden's reconstruction of the Little Ice Age profile is correct (Fig. 3.5), the ice-divide eventually passes beyond the crater rim, at which point the crater ceases to supply meltwater to Sólheimajökull, full stop.
- If sliding is suppressed, the likelihood of widespread conduit drainage increases (fast sliding ice tends to stabilise distributed drainage).
- If flushing is suppressed - as I infer both from the moraine record and my predictions of change in drainage - it is possible that debris drag raises the level of shear stress the bed supports, and suppresses ice sliding (Schweizer and Iken, 1992).

Changes to ice geometry weaken sliding; in turn, weaker sliding impacts upon ice geometry. If sliding intensity falls, then ice is expected to build up at the head of the glacier. This helps explain: 1) the extent and stability of ice-divide migration inferred; 2) Sólheimajökull's limited Little Ice Age advance; 3) its steeper, thicker profile; and 4) the contrasts in the size and continuity of moraines formed at different times. If this idea is correct, the crucial regulatory factor here is changing subglacial hydrology, which also affects flushing efficiency. A 'dry' high shear stress bed condition favours moraine construction because the fall in the intensity of flushing is inferred to exceed the fall in the intensity of sliding/erosion (cf. Chapter 2.6, on the relative balance between glacier power and subglacial stream power). It is possible that the late-Little Ice Age geometry of Mýrdalsjökull/Sólheimajökull represents the maximum extent of ice-divide migration which can be achieved: a) ice cannot pile-up indefinitely at the head of Sólheimajökull, because eventually rising shear stress should force a corrective wave of ice downglacier (cf. glacier surges); and, b) it is likely that the bedrock bumps of Goðabunga and Haabunga eventually bring back-wastage of Hofðabrekkujökull to a halt.

Jökulhlaups

The record of *jökulhlaup* activity lends support to this hypothesis of ice-divide migration coupled with shift eastwards of the centre of volcanic activity. Abundant pumice-rich sediments indicate that frequent volcanic floods of high-magnitude issued from beneath Sólheimajökull in pre-settlement times (Maizels and Dugmore, 1985; Dugmore, 1987). However, since settlement the incidence of volcanic floods has declined: the last volcanic flood at Sólheimajökull was in 1360 (Einarsson *et al.*, 1980). This fall in flood frequency does not relate to diminished activity of Katla, which has erupted on average every 50 years for the last 1,000 years. These floods have escaped beneath Hofðabrekkujökull. Direction of flood flow follows ice surface slope (e.g Guðmundsson *et al.*, 1997), so the switch in routing implies a change in the location of the eruption/melt centre relative to the ice-divide (as I suggest above, the potential link between flow routes and ice geometry provided by differential ice flow makes it possible that these are related). The last big Sólheimajökull flood in the 10th or 11th century AD possibly indicates the last major eruption of the main volcanic crater before the ice-divide moved outside of the crater rim. Since this time, Hofðabrekkujökull has acted as the escape route for crater meltwaters.

3.5 DISCUSSION: FLUSHING CONTROLS - FURTHER CONSIDERATIONS

In Chapter 2 I argued that crude calculations of surrogate stream power are likely to give a rough indication of subglacial flushing efficiency. Ice-divide migration changes the glacier catchment area, its length, and its relief ratio/mean slope, so it should also change the overall magnitude of subglacial stream power (unless reduced area and steeper slopes cancel each other out). Here I use this idea to calculate surrogate subglacial stream powers for Drangagil stage and Little Ice Age Sólheimajökull. The contrasts in moraine size lead me to expect that the 'stream power' of the Drangagil stage must be much greater. For the calculations which follow (Table 3.1) I take the mean surface slope *beyond* the crater rim as the relevant slope which drives water flow. Water which travels uphill to reach the crater rim *gains* energy, and is unlikely to contribute to much evacuation of debris; it is the expenditure of energy as subglacial water flows downhill which is important. Table 3.1 shows my 'best-guess' calculations of surrogate subglacial stream power.

Table 3.1

Sólheimajökull: surrogate subglacial stream powers calculated using best-guess Holocene geometries.

SÓLHEIMAJÖKULL: SURROGATE SUBGLACIAL STREAM POWERS			
	Property	**Drangagil Stage**	**Little Ice Age**
a)	ELA	1,000 m	700 m
b)	AAR	2.1:1	1.7:1
c)	Total catchment area	~135 km^2	~35 km^2
d)	Glacier length beyond crater rim	~17 km	~12 km
e)	Glacier relief	~1.3 km	~1.3 km
f)	Mean ice surface slope: i.e. (e) / (d)	0.076	0.108
g)	Stream power/flushing efficiency: i.e. (c) x (f)	**10.26**	**3.67**

Errors and evaluation

The error potential here is high. In particular, values for catchment area used are poorly constrained by best-guess assumptions regarding what happens to the ELA and AAR of Sólheimajökull over the Holocene. However, the area of Drangagil stage Sólheimajökull has to fall by ~50%, and the area of Little Ice Age Sólheimajökull has to rise by ~50% before the calculated values of surrogate subglacial stream power equalise. Other errors lie not with the data (the values used for glacier length and glacier relief are reasonably certain), but with the method. Although often used for this purpose, catchment area is a poor surrogate for total discharge: e.g. run-off per unit area in the Little Ice Age is likely to be less than it was for the Drangagil stage if the climate was colder. Similarly, the use of the mean slope-'discharge' product is not likely to be that reliable: e.g. if a greater proportion of Drangagil stage Sólheimajökull occupies high elevations, more water falls a greater height, and so raises total power expenditure. (N.B. Both these possible errors imply that the given figure *underestimates* the total stream power of Drangagil stage Sólheimajökull relative to its Little Ice Age equivalent, whereas a challenge to my hypothesis requires errors which incorrectly elevate Drangagil stage stream power, and/or reduce Little Ice Age stream power.) Despite the shaky foundations on which these calculations are based, the difference in surrogate 'stream power' - i.e. that calculated for the Drangagil stage is 2.8 times greater than that for the Little Ice Age - seems sufficiently large to be able to accommodate major errors and still give support to my inference that the flushing efficiency of Drangagil stage Sólheimajökull was much higher.

Past state of the drainage network

The parallel I draw between Holocene Sólheimajökull and surging glaciers is useful, but it is important not to read too much into it. Surge theory - notably that part attached to the 1982-1983 surge of Variegated Glacier (Kamb *et al.*, 1985; Kamb, 1987) - establishes the idea of a change in the configuration of subglacial drainage which simultaneously brings about acceleration and advance of ice, and an increased efficiency of subglacial flushing (Humphrey and Raymond, 1994; Sharp *et al.*, 1994; see Chapter 6.3). However, surge changes to water flow, ice dynamics and terminus position occur far more rapidly than the changes I discuss above, for which the time-scale (100s to 1,000s of years) incorporates the impact of climate change. Presumably, the contrast in basal conditions which regulates surges is much greater than the contrast I invoke as part of this gradual change to mass balance, glacier geometry, and ice, water and sediment dynamics. This means that the instance of Variegated Glacier cannot necessarily be used to justify a full-scale, high pressure, large water storage linked-cavity network beneath Drangagil stage Sólheimajökull. Although some measure of increased bed coverage by water is required to speed up sliding (increased water volume and higher water pressures), it is probably not necessary to have a linked-cavity network permanently established. The moraine evidence implies that the flushing efficiency of the Drangagil stage was similar to that of Sólheimajökull today, for which a full linked-cavity network is not inferred (the high values of basal shear stress suggest otherwise for one thing; see Chapter 2.2). A hundred-fold increase in sliding speed (as at Variegated Glacier) is not required to explain Sólhaimajökull's behaviour in the Holocene. The overall increase in sliding speeds I infer for the Drangagil stage is possibly consistent with a semi-permanent, relatively widespread distributed network which activates temporary pulses of faster flow: e.g. the repetitive sequences of short-lived, high pressure bed separation events and part-complete drainage reorganisation which Collins (1989) believes to occur at Gornergletscher (see also Walder, 1986). Such a mixture of (non-steady-state) discrete/distributed drainage interaction is consistent with the case of present-day Sólheimajökull, and - by extension - Drangagil stage Sólheimajökull also. There are good reasons to believe that drainage of this kind favours high efficiency flushing (Chapter 2.5). If this is so, then reduced flushing of Little Ice Age Sólheimajökull probably ties in with a high shear stress, low bed coverage drainage state, possibly associated with reduced discharge through a relatively stable conduit network.

3.6 CONCLUSIONS

The ice-divide migration hypothesis is improved if the potential impact of changing ice dynamics and changing subglacial hydrology are included. I believe that my ideas, in conjunction with those advanced by Dugmore and Sugden (1991), provide a plausible, internally-consistent account of the response of Mýrdalsjökull and Sólheimajökull to fluctuations of climate and a shift in the distribution of volcanic activity. Existing theory supports the three chief ideas on which this revised ice-divide migration hypothesis is founded, specifically that:

1. Changes to the way in which climate, ice geometry and bedrock topography interact are likely to alter subglacial drainage.

2. Changes to subglacial drainage can influence flow speeds and ice advance.

3. Changes to ice dynamics and glacier geometry can induce further changes by positive feedback response, and so establish a self-sustaining regime of ice advance or retreat.

The inferred switches to subglacial hydrology which regulate this behaviour also involve a change in flushing efficiency, as the moraine record shows; conversely, the contrasts of the moraine record can be read as support for the idea that important changes to subglacial drainage did indeed take place. Thus we have a geomorphic record of glacier fluctuations 1) which represents a complex response to climate change, and, 2) varies in its quality because of hydrological factors intimately associated with this complex response. Areas of poorly-constrained ice topography which occur in conjunction with volcanic activity which might display similar behaviour - at present, or in the past - include other parts of Iceland, parts of Alaska, South America, and - on a grand scale - Antarctica.

CHAPTER 3: SUMMARY

- Contrasts in the size and continuity of moraines developed by Sólheimajökull at different times in the Holocene are most likely explained by past fluctuations of the intensity of subglacial flushing. This is thought to have been low in the Little Ice Age ('dry' bed, large moraines), but high at the start of the Neoglacial ('wet' bed, small moraines, as for today). These contrasts in flushing must have been the product of past changes to subglacial hydrology.

- Any changes to subglacial drainage invoked must tie in with Sólheimajökull's anomalous response to late-Holocene climate change. Dugmore and Sugden's ice-divide migration hypothesis is likely best to explain this.

- Changes to ice flow controlled by variable subglacial drainage conditions fit the ice-divide migration hypothesis. The added element of process realism seems to correct minor deficiencies of this hypothesis in its basic form, associated with a) the anticipated pattern of ice-cap development, b) the unrealistic assumption of time-invariant AAR, and, c) the likelihood of time-dependent changes in the basal shear stress condition of Sólheimajökull.

- Changes in Sólheimajökull's catchment area and relief associated with ice-divide migration support the idea that subglacial hydrology must have varied over the Holocene. Drangagil stage Sólheimajökull and Little Ice Age Sólheimajökull reflect emergent, quasi-robust, self-organising, self-sustaining process domains (see Chapter 9), established because of the interdependent interaction of several factors (mass balance change, ice-divide migration, changes to ice flow and changes to water flow). The fact that these reinforce each other results in a complex response to initial perturbation of the system from outside. This external trigger is climate change, possibly reinforced by changes in volcanic activity (although big changes in ice geometry are believed to affect volcanic activity; if so, this internalises the volcanic factor). The possibility of such complex response carries important implications for reconstruction of past climates using the geomorphic record (see Dugmore and Sugden, 1991).

- Changes to flushing efficiency and moraine development emerge as a by-product of this complex response/switch between dominant process domains. This inference is supported by crude calculations of surrogate subglacial stream power.

GÍGJÖKULL AND STEINHOLTSJÖKULL

Current moraine accumulation at Gígjökull and Steinholtsjökull provides a stark contrast to what is seen at Sólheimajökull. Whereas Sólheimajökull is notable for the paucity of its ice-marginal sediments, Gígjökull and Steinholtsjökull display unusually large quantities of marginal debris for glaciers which receive little in the way of supraglacial debris inputs. This contrast suggests that important differences must exist between Gígjökull/Steinholtsjökull and Sólheimajökull in the way in which ice and water behave. I address two distinct styles of present-day sedimentation at Gígjökull and Steinholtsjökull: a) stacked moraine ridges made up of water-worked debris, and, b) moraines fed by unusually thick sequences of debris-rich basal ice, and move on to argue that meltwater drainage is the key factor which links the two together. I use this idea to develop a conceptual model of sediment transfer patterns characteristic of terminal overdeepenings, which can be used to explain the past record of moraine accumulation at glaciers - such as Gígjökull and Steinholtsjökull - which have these.

My study of Gígjökull and Steinholtsjökull divides as follows:

Present-day sedimentation:

Chapter 4: BACKGROUND, METHODS AND DATA

Chapter 5: WATER-WORKED ENGLACIAL DEBRIS

Chapter 6: BASAL ICE

Chapter 7: SEDIMENT TRANSFER IN ICE-FALLS AND OVERDEEPENINGS

Past sedimentation:

Chapter 8: NEOGLACIAL MORAINE ACCUMULATION

CHAPTER 4
Gígjökull and Steinholtsjökull: Background, methods and data

4.1 SITE DESCRIPTION

Gígjökull and Steinholtsjökull (Figs 4.1 and 4.2) are the two largest outlets of the small mountain ice-cap Eyjafjallajökull, southern Iceland (63° 40' N, 19° 40' W), located approximately 150 km ESE of Reykjavík (Fig. 1.4).

Solid geology

Eyjafjöll is a Pleistocene volcanic massif which last erupted in 1821-23, the second of only two historical eruptions (Dugmore, 1989). This last eruption gave rise to a flood, evidence of which is abundant beyond Gígjökull's large lateral-terminal moraine. Bedrock is similar to that found at Sólheimajökull: a mixture of basaltic lavas and palagonite tuffs and breccias (Chapter 2.1). Sections exposed in the high cliffs flanking Gígjökull's ice-fall show a complex stacked sequence of lavas and pyroclastic flows. Much of this material is thought to be highly susceptible to both subaerial and subglacial erosion.

Climate and mass balance

Comprehensive data are not available. However, because of the maritime location of Eyjafjöll - directly exposed to onshore moisture-bearing winds - accumulation, ablation and glacier turnover are all expected to be high. Rist (cited by Lawler, 1991; see Chapter 2.1) states that the 1954-55 accumulation at the summit of the neighbouring Mýrdalsjökull was 5,800 mm water equivalent (w.e.), so similarly high snowfall on the top of Eyjafjallajökull - which is ~200 m higher than Mýrdalsjökull - seems reasonable. This is supported by my observations in late August 1993: crevasses exposed at the head of Gígjökull revealed several metres (~5 m plus?) of snow lying above firn. Ablation (strictly, surface lowering) of Gígjökull's terminal lobe was measured at 14 stakes for ~30 days in August and September 1993, and averaged ~6 cm d^{-1}. Ablation at a further stake located higher up in the bottom part of the ice-fall was ~20% lower. The equilibrium line on Eyjafjallajökull is believed to lie at ~1,100 m a.s.l. (Björnsson, 1979). This is broadly consistent with crude AAR calculations. With the ELA at 1,100 m, the AAR for

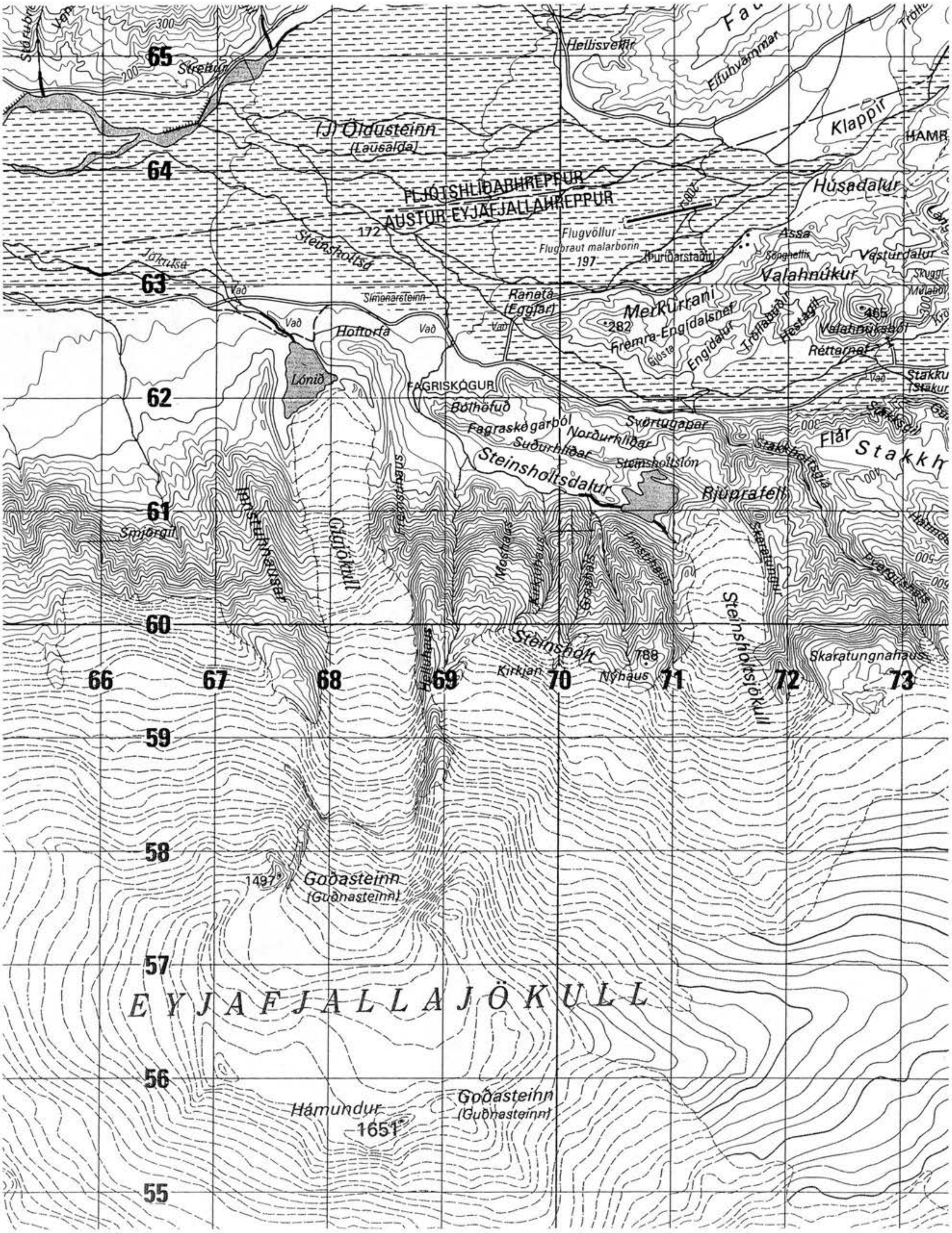


Figure 4.1

Gígjökull and Steinholtsjökull: extract from sheet 1812: III 'Eyjafjallajökull'. Scale 1:50,000. Copyright: Defense Mapping Agency Hydrographic/Topographic Center, Washington DC, USA / Iceland Geodetic Survey.

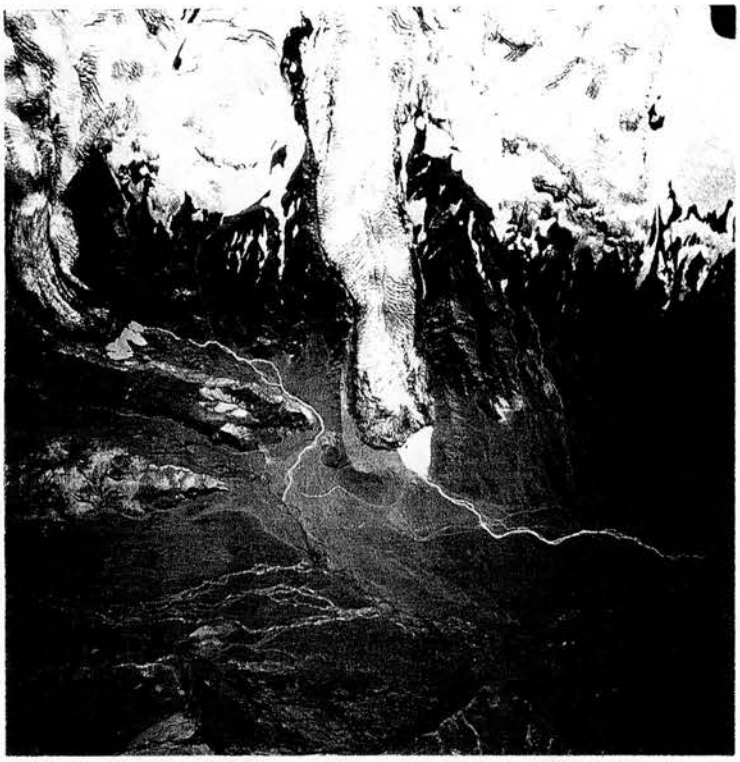
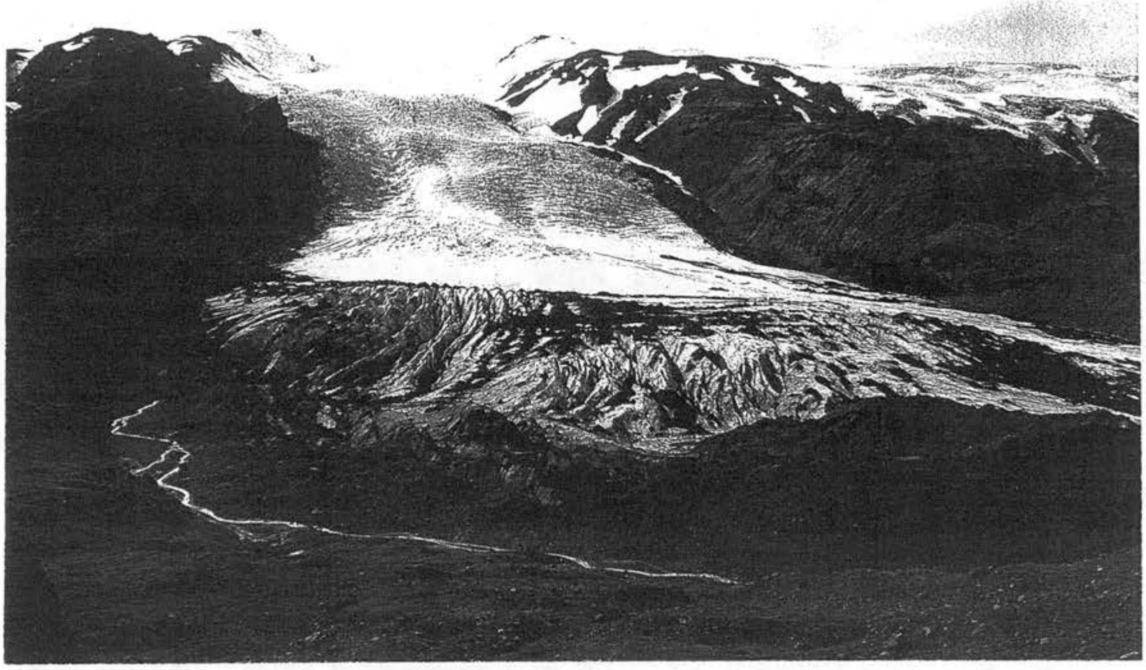


Figure 4.2

Aerial view of Steinholtsjökull (L) and Gígökull (R) in 1987. Photo: Iceland Geodetic Survey.



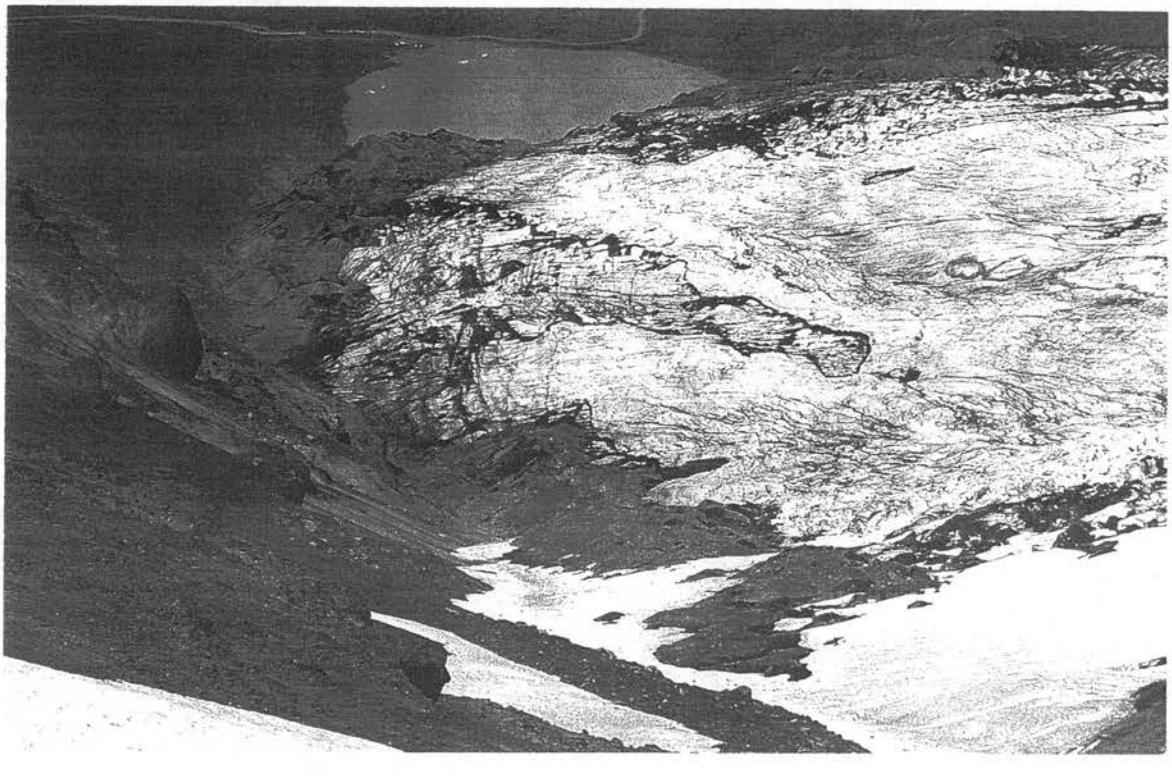


Figure 4.3

Two views of Gígjökull. Top (Fig. 4.3a): seen from its eastern Neoglacial rampart. Thick basal ice exposures (dark) lie above the stream, with relict conduit debris above this (lower-centre-right of picture). Bottom (Fig. 4.3b): the terminus seen from ~700 m asl, above the left bank of the ice-fall. Note: 1) the large spreads of relict conduit debris which choke the left margin of the terminal lobe; and, 2) the tephra loop.

Gígjökull works out at ~2.0:1. However, if Björnsson's AAR of 1.7:1 is used, the ELA works out at ~1,150-1,200 m.

GÍGJÖKULL

Gígjökull (Fig. 4.3) drains the central crater of the dormant Eyjafjöll volcano. It occupies an area of 8.4 km^2, and descends 1,450 m in its length of 6.5 km. Its middle section, beyond the crater lip, consists of a large ice-fall (~3,000 m in length, height drop ~1,000 m, mean angle 16^o, perhaps 25^o+ at its steepest), in front of which a gentle terminal lobe ends in a small lake trapped in front of the ice. This lake indicates the existence of a terminal overdeepening (see below), partially exposed by Gígjökull's retreat from its Neoglacial maximum, reached during the nineteenth century. This century, Gígjökull underwent a marked phase of retreat prior to *c.* 1960, replaced by advance up until *c.* 1990, after which the ice front has remained relatively stable (Sigurðsson, 1992).

Glacier geometry and ice flow field

Ice depth data obtained in 1993 suggest that beneath the terminal lobe the ice is ~100 to 140 m thick. The cross-section seems to be asymmetrical, with the zone of maximum depth located ~250 m east of the 'central' (i.e. fastest) flow-line (picked out by a ramp of clean ice which descends to the lake, enclosed within the marked displacement downglacier of the 1947 tephra outcrop - see Figs 4.2 and 4.4). This is supported by a similar reconnaissance ice-radar survey conducted in 1994 (Rogers *et al.*, 1994), which gave near-identical results. The ice is believed to thin rapidly towards the snout: the 1994 ice radar survey data strongly suggest the presence of an overdeepening, although the number and quality of the bed returns are not sufficient to confirm this with confidence. However, 1996 ice radar data, collected using an improved system, provide a reliable centre-line profile ('centre-line' defined by the line of inferred fastest flow: Fig. 4.4) which clearly shows a pronounced overdeepening, with maximum ice thickness of 167 m (unpublished data courtesy of Andy McBain, Kate Seal, Penny Thomson and Emma Walters). Fig. 4.5 shows this profile. At its steepest, the adverse slope of the terminal *riegel* is of the order of 60% (30^0); the ice surface above this slopes at ~20% (12^0). Thus the bedrock slope dips backwards approximately three times as steeply as the ice surface slope dips forwards. This has important implications for glacier drainage (Chapter 5.6).

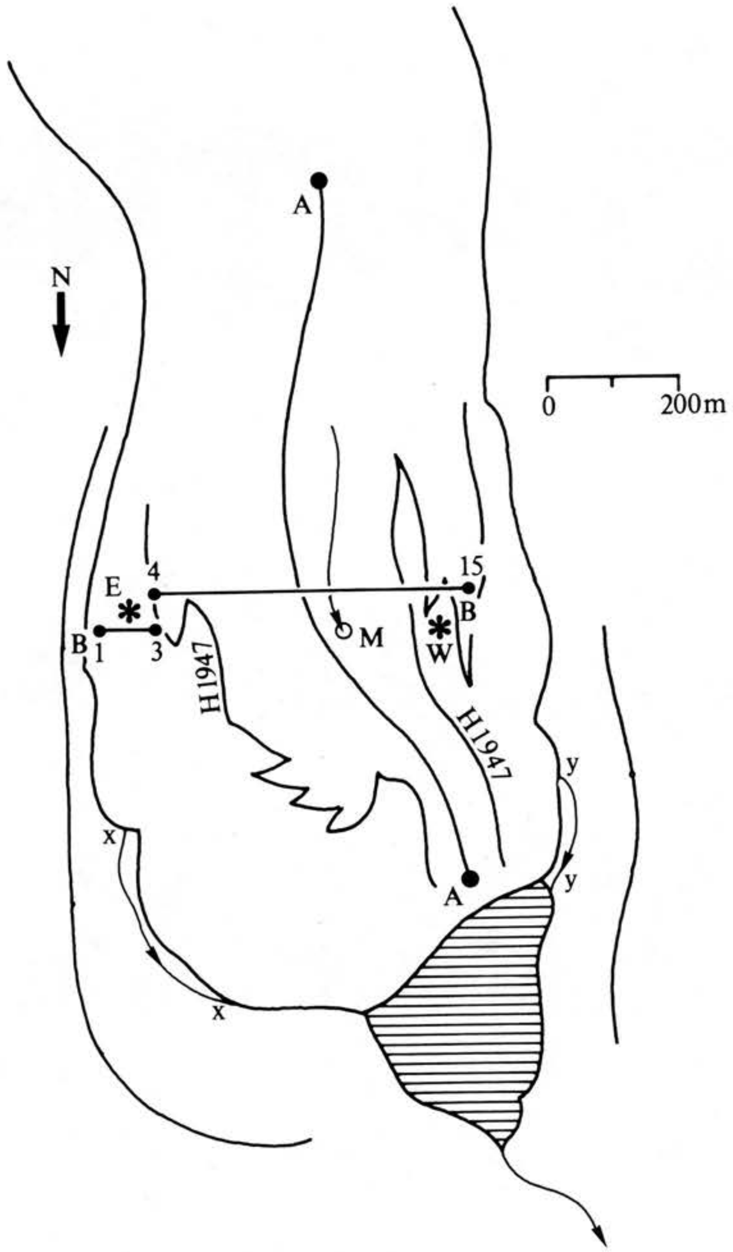


Figure 4.4

Gígjökull: approximate location of ice surface velocity measurements, ice radar profiles, and strain nets. KEY: **A** to **A** = longitudinal radar profile (Fig. 4.5) / surface flow speed survey (Fig. 4.7); **B** to **B** = lateral radar profile (see text) / surface flow speed survey (Fig. 4.6: numbers indicate marker stones); **W** = western strain net; **E** = eastern strain net (Table 4.1); **M** = moulin at foot of surface valley; **H1947** = outcrop of Hekla 1947 tephra; **x-x** = eastern marginal stream; **y-y** = western marginal stream (both as in 1994).

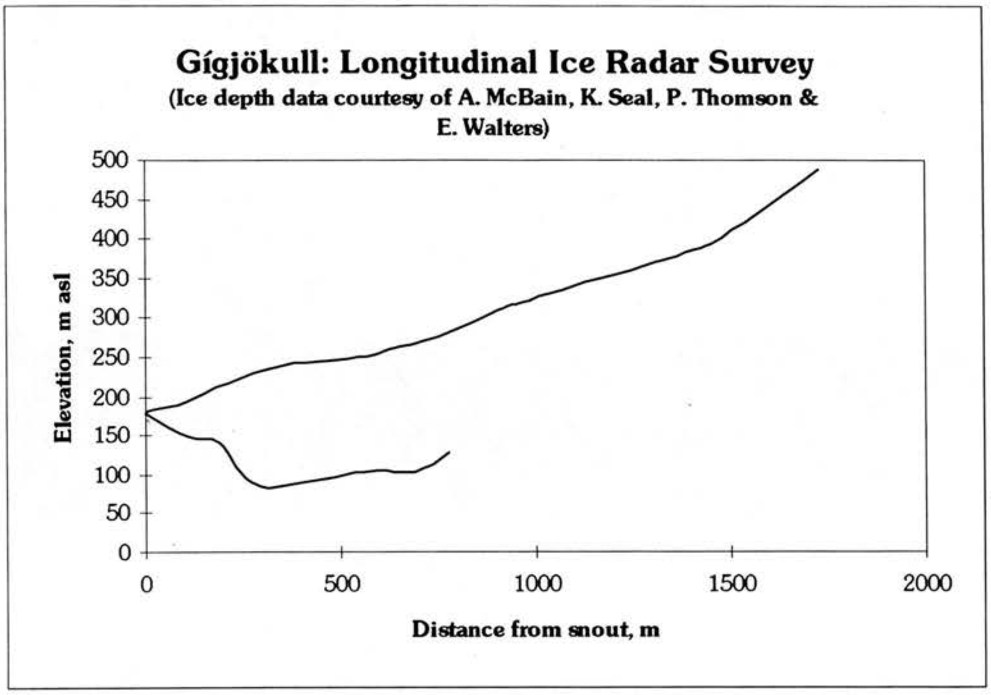


Figure 4.5

Gígjökull: longitudinal ice-radar profile, 1996. See Fig. 4.4 for location of survey line.

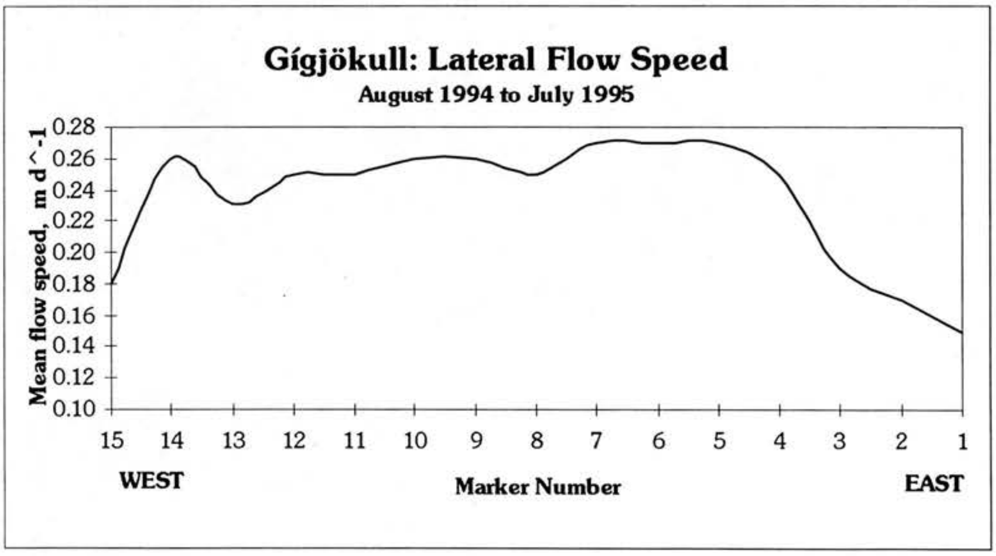


Figure 4.6

Gígjökull: mean lateral ice surface velocities, August 1994 to July 1995. See Fig. 4.4 for location of survey markers.

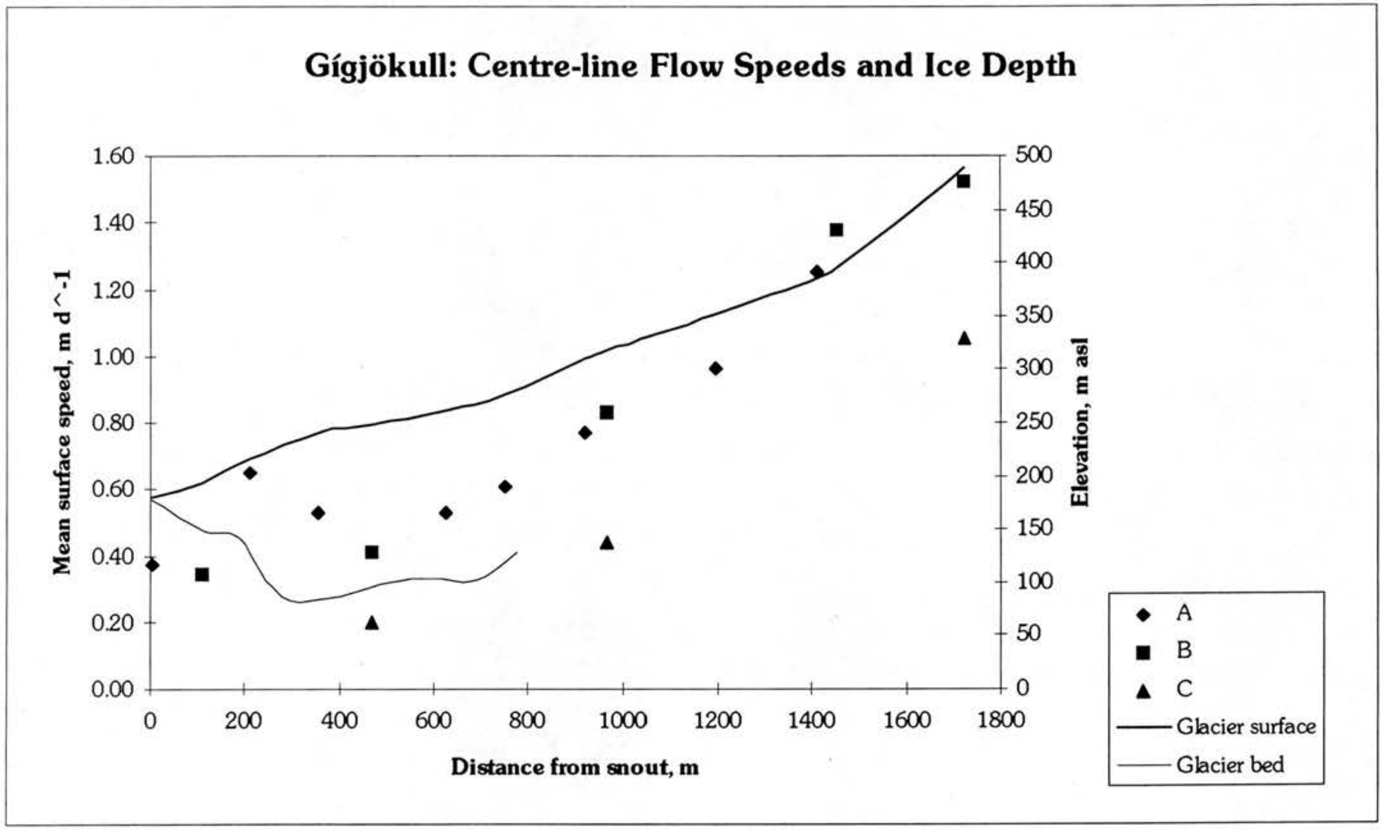


Figure 4.7

Gígjökull: longitudinal ice geometry and flow speeds. Left axis: crude surface velocity data. Right axis: ice surface and bed elevation data. See Fig. 4.4 for location of profile.

VELOCITY DATA SERIES:
A: R. Rogers, A. Schweizer, K. Simpson, C. Symes (Rogers *et al.*, 1994): 4 days, July 1994
B: My data: 26 days, August-September 1993
C: My data: 300-350 days, September 1993 to June/August 1994
Ice depth data courtesy of A. McBain, K. Seal, P. Thomson and E. Walters

Ice flow speeds. Using 'best-guess' figures (ELA = 1,100 m; mean net accumulation above EL = 2.0 m w.e. yr^{-1}; flow width at EL = 800 m; flow depth at EL = 150 m) gives an estimated balance velocity at the EL of ~100 m yr^{-1}. Surface flow velocities of the terminal lobe (~600 m from the snout) measured by the change in position of marked rocks (Fig. 4.4) between August 1994 and July 1995 average ~0.26 m d^{-1}/~95 m yr^{-1} (Fig. 4.6). Mean flow speeds are pretty much uniform across much of the glacier's width, but fall sharply towards the ice margins. This indicates some measure of 'plug flow', which implies that basal sliding makes a significant contribution to overall ice flow at this location.

Table 4.1

Gígjökull: strain calculations.

Gígjökull: Strain Calculations			
Strain Axis	**Strain, d^-1**	**% Strain in x**	**Cumulative Strain**
Western strain net, 26 days, August to September 1993			
Flow towards ice margin (x)	-0.00057		-0.0150
Lateral stretching (z)	0.00048	84	0.0125
Vertical thickening (y)	0.00009	16	0.0025
Eastern strain net, 26 days, August to September 1993			
Flow towards ice margin (x)	-0.00105		-0.027
Lateral stretching (z)	0.00064	61	0.017
Vertical thickening (y)	0.00041	39	0.010
Eastern strain net, 696 days, August 1993 to July 1995			
Flow towards ice margin (x)	-0.00063		-0.44
Lateral stretching (z)	0.00035	56	0.24
Vertical thickening (y)	0.00028	44	0.20

Ice flow pattern. The lower part of the glacier exhibits a strongly compressive flow regime: centre-line surface velocities fall by a factor of ~5 between the lower parts of the ice-fall and the snout, a distance of just over 1.5 km (Fig. 4.7). Long-term (i.e. over-winter) data show substantially slower flow speeds than do short-term (i.e. summer) data. This implies the existence at Gígjökull of a marked seasonal contrast in velocity (i.e. flow is faster in summer), as is common at alpine-type glaciers. Strain-net data also show that flow is compressive. Two strain nets - one eastern and one western - were set up as close to the ice edge as crevasses allowed (Fig. 4.4). Each was located just upglacier/inside of the major zones of ice-marginal sedimentation. Strain rates were calculated using Nye's method, as given by Dackombe and Gardiner (1983, pp. 163-165): Table 4.1. These data show substantial longitudinal shortening (the *x* axis here is directed towards the ice margin at an angle) taken up by lateral stretching (i.e. in *z*) and vertical thickening (i.e. in *y*). These strain calculations agree with what can be inferred from ice structures. Zones of intense semi-radial crevassing immediately adjacent to both strain nets indicate lateral stretching of the ice; tight folding of the ice foliation and the

Figure 4.8

Steinholtsjökull.

1947 tephra band (both exposed in the walls of crevasses adjacent to the eastern strain net) indicate longitudinal compression and vertical thickening. When compared with late summer data, 23 months of data from the eastern strain net suggest that strain, like surface velocity, shows marked seasonal variations, with summer maxima. (Unfortunately, marker stones for the western strain net left on the ice surface in September 1993 were not recovered the following summer.) Two further points are worthy of note: a) strain rates appear to be greater at the eastern margin than at the western margin; and, b) a greater proportion of longitudinal compression (i.e. 44% vs. 16%) is taken up by vertical thickening at the eastern margin than at the western margin.

STEINHOLTSJÖKULL

Steinholtsjökull (Fig. 4.8) drains the flank of the ice-cap east of Gígjökull, and is similar to its neighbour in most respects, although smaller (area 6.92 km^2, length 4.8 km, altitudinal range 1,320 m) and, by inference, less active (no flow data are available for Steinholtsjökull). Steinholtsjökull too has advanced rapidly in recent decades (as is shown by the sequence of aerial photos taken between 1978 and 1989), as a result of which the size of its proglacial lake has progressively fallen. This lake - as with Gígjökull - is thought to indicate a terminal overdeepening. If the evidence of the 1990 1:50,000 map sheet '1812 III: Eyjafjallajökull' - which shows Steinholtsjökull >500 m behind its present snout position - is correct, an adverse slope must lie beneath the front 250 m or so of Steinholtsjökull's terminal lobe (*c.* 1995) (see Fig. 4.1).

4.2 FIELD METHODS

Fieldwork was designed 1) to identify the major styles of moraine-forming activity at each glacier, and, 2) to relate these to sediment transport pathways. Initial interpretations were made on the basis of the debris' wider relationships with ice structure and inferred flow pattern, supported by studies of a) clast form, and, b) the debris content of ice samples from which the sediment was derived.

CLAST FORM

Clast form reflects both the properties of the source bedrock, and the subsequent modification of the clast by various processes experienced whilst in transport or storage. Distinct modes of sediment transport supposedly impart upon clasts a characteristic form, so measures of clast form can be used to reconstruct sediment transport pathways. Boulton (1978) used these ideas (in conjunction with particle size analysis) to differentiate between two major styles of glacial debris transport: *passive* supra- or englacial transport, in which debris tends to retain the characteristics of the process (e.g. rock-fall of frost-shattered bedrock) by which clasts are delivered to the glacier, and *active* subglacial transport, wherein processes of fracture and crushing bring about progressive, and frequently substantial, modifications to the character of the debris. Passive transport tends to give rise to clasts which are angular and elongate, whereas active transport tends to produce clasts which are edge-rounded and equant. Boulton's paper popularised the use of clast form indices in studies of glacial geomorphology; it also did much to create a theory of sediment transport by glaciers which is relatively simplistic and inflexible (Chapter 1.2; Kirkbride and Spedding, 1996, p. 160).

Three aspects of clast form are usually identified: 1) *shape* (which pertains to the relative lengths of a clast's *a*, *b*, and *c* axes); 2) *roundness* (the extent to which the edges/faces of a clast appear round); and, 3) *texture* (surface roughness). The last of these is difficult to quantify (although textural features such as striations can give crucial clues to the style of transport) so most authors rely on study of the co-variance of shape and roundness (e.g. Benn and Ballantyne, 1994).

Sample analysis

Clasts with *a* axes between 30 and 120 mm were selected at random from the different debris types (see Fig. 4.9 for location of sample sites). Individual sample size was determined by the quantity of data which filled a single page of my field notebook: typically 63 clasts. Clast roundness was assessed using the six point scale of Powers (1953). This relies on subjective and qualitative description, whereby clasts are assigned to one of six roundness categories according to their appearance and 'feel' (Table 4.2). Each roundness class was assigned a score between 1 (very angular) and 6 (well-rounded) which allows a mean value of roundness to be assigned to each sample. Benn and Ballantyne (1994) object to this technique of deriving a 'mean roundness' statistic because a) clast roundness data frequently do not form a normal distribution (which means outliers of a skewed distribution can exert a substantial influence on the mean roundness value), and, b) use of discrete classes (i.e. ordinal, not interval/ratio data) introduces an element of bias. However, I suggest 1) the arithmetic mean is a valuable intuitive concept which is not tied to the normal distribution; 2) considerations of the (population) normal distribution and interval/ratio data become crucial only if heavy reliance is to be placed upon the use of inferential parametric statistics; 3) in this case at least, the use of mean roundness scores to discriminate between different types of debris works! 'Practical adequacy' justifies the technique, irrespective of technical objections (Sayer, 1992, Ch. 2 and Ch. 6).

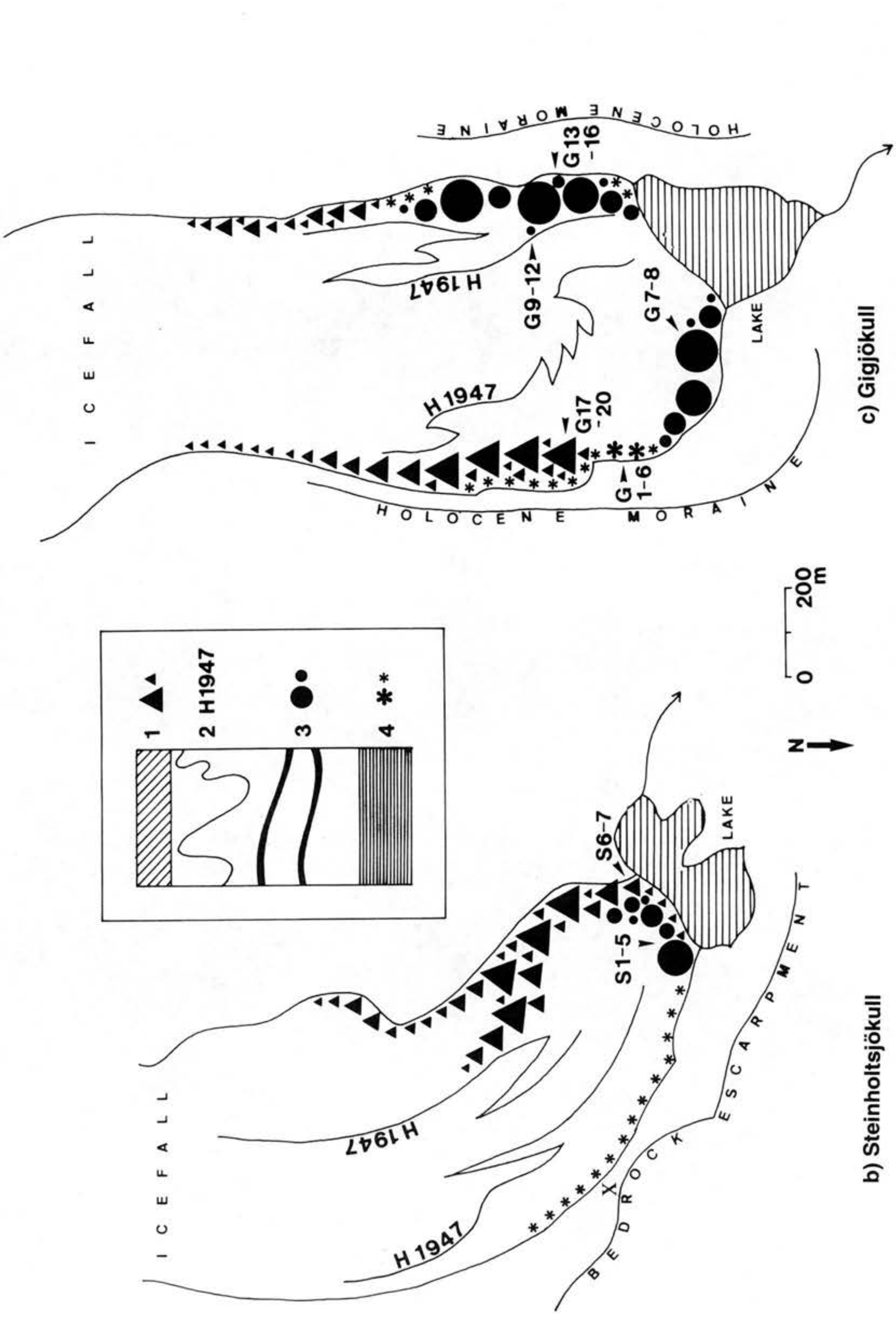


Figure 4.9

Termini of Steinholtsjökull and Gígjökull, showing the distribution of debris types. Sample numbers (arrows) refer to samples in Table 4.3. Inset: stratigraphic relationships of ice and debris. Map symbols: 1) supraglacial rock-fall debris; 2) Hekla 1947 ash; 3) englacial debris bands/water-worked supraglacial debris; 4) basal ice and debris.

Table 4.2

Classification of clast roundness. After Powers (1953) and Benn and Ballantyne (1994).

Classification of clast roundness		
Roundness Class	**Score**	**Description**
Very angular	1	Very acute edges and/or sharp protuberances
Angular	2	Acute edges with no evidence of rounding
Sub-angular	3	Rounding confined to edges, faces intact
Sub-rounded	4	Rounding of edges and faces; often faceted
Rounded	5	Marked rounding both of edges and faces; merging of edges and faces
Well-rounded	6	Faces and edges indistinguishable

Measures of clast shape did not turn out to be helpful. Two indices were tried: maximum projection sphericity [= $(c^2/ab)^{1/3}$ = 1.0 for a perfect sphere] and the C_{40} flatness index [% of clasts in sample with (*c* axis/*a* axis) ≤ 0.4 (Benn and Ballantyne, 1994)]. Debris samples which seemed to relate to different transport pathways on the evidence of their wider field relationships scored similarly on measures of sphericity and C_{40} flatness. Mean roundness scores alone proved to be perfectly adequate at distinguishing between different debris types. Benn and Ballantyne suggest that poor performance of clast shape indices in previous studies reflects a poor choice of specific shape index. Here, however, the poor performance of the shape indices appears to be a genuine reflection of the range of debris types/transport styles found at Gígjökull and Steinholtsjökull, and not an artefact of poor choice of statistical indicator. This is because:

1. Initial breakdown of bedrock produces relatively strong, equant forms to begin with. This was the impression created by car-sized blocks of bedrock stranded in front of Steinholtsjökull (presumably by the 1967 *jökulhlaup*) which had subsequently disintegrated *in situ*.

2. The utility of different indices of clast form depends upon the stage of the debris cascade from which the sample is taken. Processes of crushing and fracture which distinguish between active and passive transport tend to convert 'immature' weak (elongate) shapes to 'mature' strong (equant) shapes, without a similarly large change in clast roundness. Once formed, strong forms tend to be stable, and further evolution of 'mature' clasts tends to involve changes to clast roundness (Kirkbride, 1989, 1995a, Fig. 8.14a). The key contrast at Gígjökull and Steinholtsjökull is not that of the traditional active-passive transport dichotomy but (to anticipate the argument) that created by the modification of mature basal transport zone debris by fluvial processes. Fracture is unlikely to feature strongly at this stage of clast evolution, which tends to be dominated by rounding of clast edges and faces.

Table 4.3 summarises the clast form data.

Table 4.3

Gígjökull and Steinholtsjökull: clast form data.

Gígjökull					
Sample ID	**No. of Clasts**	**Mean Roundness**	**Roundness SD**	**Mean Sphericity**	**Sphericity SD**
Debris from basal ice					
G1	63	2.69	0.50	0.79	0.07
G2	63	2.67	0.62	0.78	0.08
G3	72	2.50	0.53	*	*
G4	63	2.71	0.52	*	*
G5	72	2.53	0.50	*	*
G6	63	2.54	0.62	*	*
G1-G6	**396**	**2.60**	**0.54**	*	*
Water-worked debris: East					
G7	63	3.50	0.82	0.81	0.08
G8	63	3.43	0.69	0.80	0.08
G7-G8	**126**	**3.47**	**0.72**	*	*
Water-worked debris: West					
G9	59	3.85	0.76	0.81	0.07
G10	63	3.90	0.64	0.79	0.07
G11	63	3.92	0.75	0.79	0.07
G12	63	3.76	0.69	0.81	0.08
G13	52	3.75	0.63	0.84	0.08
G14	63	3.75	0.74	0.81	0.09
G15	61	3.92	0.90	0.78	0.10
G16	63	3.79	0.83	0.80	0.06
G9-G16	**487**	**3.83**	**0.75**	*	*
Rock-fall debris					
G17	59	2.24	0.62	*	*
G18	63	2.32	0.62	*	*
G19	63	2.33	0.60	*	*
G20	63	1.92	0.69	*	*
G17-G20	**248**	**2.19**	**0.62**	*	*

Steinholtsjökull					
Sample ID	**No. of Clasts**	**Mean Roundness**	**Roundness SD**	**Mean Sphericity**	**Sphericity SD**
Water-worked debris					
S1	78	3.38	0.71	0.79	0.10
S2	76	3.30	0.59	0.79	0.07
S3	76	3.46	0.54	0.80	0.08
S4	74	3.40	0.68	0.78	0.06
S5	66	3.33	0.75	0.82	0.09
S1-S5	**370**	**3.38**	**0.68**	*	*
Rock-fall debris					
S6	75	2.33	0.58	*	*
S7	79	2.35	0.60	*	*
S6-S7	**154**	**2.34**	**0.60**	*	*

PARTICLE SIZE ANALYSIS

This gives a second measure by which debris can be classified (e.g. Boulton, 1978; Haldorsen, 1981). Forty samples of frozen ice/sediment mixtures were chipped out of the ice margins for particle size analysis. Six samples of frozen debris band material (see below, and Chapter 5) were obtained (with difficulty!) at Gígjökull from exposures adjacent to G13 (four samples) and G9 (two samples). Thirty-four samples of basal ice (see below, and Chapter 6) were taken: twenty-four from Gígjökull (adjacent to site G1) and ten from Steinholtsjökull (**X** marks the spot: Fig. 4.9). The number of samples taken was limited by time, what I was able to carry, and the excess baggage regulations of *Icelandair*. I decided that particle size data were likely to be of greater value for detailed interpretation of what was identified in the field simply as 'basal ice'. Particle size data were not a critical factor used to identify the water-worked debris of the englacial debris bands. This explains the sampling bias in favour of basal ice.

Electrical conductivity

Meltwater samples were obtained as a by-product of the particle size analysis sample collection procedure (see below). The electrical conductivity (EC) of these filtrates was measured in the field. This was intended as a crude test of whether or not ice was of meteoric (englacial) or basal origin (see Chapter 1.2). EC gives a crude indication of the concentration of solutes in ice/meltwater. Snow is subject to leaching prior to its transformation into glacier ice, so englacial ice usually gives rise to chemically pure meltwaters. Chemical weathering processes occur widely at the glacier bed because meltwater here is in contact with rock material, so basal ice tends to be enriched in solutes relative to englacial ice.

Debris concentration

Debris concentration by mass of each sample was obtained as a second by-product of the particle size analysis procedure (see below). This is a useful measure because it can be used:

- To identify basal ice.
- To differentiate between different types of basal ice (e.g. Hubbard and Sharp, 1995).
- To differentiate between basal ice and debris-rich ice of different origin.

Table 4.4a

Gígjökull: ice/debris analysis.

Gígjökull				
Sample ID	**Total Mass g**	**Mass of Debris g**	**Concentration %**	**EC mS cm^-1**
Englacial ice				
71	*	0	0	1.4
72	*	0	0	1.7
73	*	0	0	2.1
74	*	0	0	1.6
Ice and debris mixture, relict conduit debris band				
31	304	239	78.6	63.3
32	256	191.1	74.6	51.6
37	149	102.8	69	9.8
38	199	143.4	72.1	15.2
39	169	125.4	74.2	14.1
40	*	179.9	*	21.1
Basal ice				
33	89	0.6	0.7	18
34	74	7.6	10.2	15.5
35	294	18.4	6.3	14.5
36	229	13.5	5.9	20.5
51	144	11.2	7.8	28.7
52	144	16.7	11.6	22.1
53	94	6.1	6.5	24.3
54	159	13.6	8.6	17.8
55	179	14.6	8.2	16.4
56	174	13.4	7.7	14.7
57	129	51.5	39.9	24.8
58	129	14.6	11.3	19.8
59	134	12.6	9.4	24.2
60	149	16.7	11.2	22.8
61	124	*	*	21.6
62	104	44.9	43.2	28.2
63	144	1.4	0.9	16.7
64	184	4.4	1.8	14.3
65	124	4.2	3.4	17.4
66	209	7.1	3.4	19.1
67	114	2.8	2.5	25.4
68	154	36.3	23.6	24.7
69	134	29.9	22.3	31.9
70	139	21.8	15.7	32.4

Table 4.4b

Steinholtsjökull: ice/debris analysis.

Steinholtsjökull				
Sample ID	**Total Mass g**	**Mass of Debris g**	**Concentration %**	**EC mS cm^-1**
Basal ice				
41	74	21.1	28.6	46
42	129	5.3	4.1	24.1
43	104	1.2	1.2	22
44	144	45.2	31.4	32.2
45	154	45.5	29.6	38.5
46	149	1.1	0.7	17.8
47	74	24.6	33.2	31.2
48	144	32.5	22.6	34.1
49	109	3.5	3.2	23.7
50	144	25.8	17.9	25.2

CORRECTIONS (these apply to Tables 4.a and 4.b)

TOTAL MASS:

1. Whatman filter paper, 24.0 cm, Grade 1 Qualitative - 4.00 g
2. Heavy duty plastic bag used to weigh samples - 11.96 g
3. Plastic bag used to collect samples - 4.83 g

OVERALL CORRECTION:

Field mass - (1 + 2 + 3 + 3) - 25.62 g

MASS OF DEBRIS:

4. Individual aluminium pie case: - 1.03 g

Sample collection and processing

Blocks of ice were extracted from the ice face using an ice hammer (Mountain Technology) and a suitable level of violence. The blocks were washed free of surface debris, and weighed to the nearest 5 g with a 1 kg tubular spring balance (Salter). Each sample was then double-bagged and sealed in domestic freezer bags, labelled, and taken back to the tent in a 'cool bag' (i.e. a large, insulated plastic bag designed to carry home frozen foods from the supermarket). This was used to minimise melting and the chance of sample leakage. Ice samples were set out in filter funnels, and left to melt at air temperature. 24.0 cm diameter Whatman Qualitative Grade 1 filter papers, which retain particles in excess of 11 μm, were used. The filtrate was collected, and its electrical conductivity measured with a Jenway hand-held conductivity/temperature probe, using the 0-200 μm cm^{-1} range (resolution 0.1 μm cm^{-1}) with automatic temperature compensation of the readings (standardised to 25 °C).

Filter papers and trapped debris were taken back to the laboratory, dried thoroughly, and weighed to an accuracy of 0.1 g by digital balance. With suitable corrections (see notes with

Table 4.4) this gave the weight of debris collected, and the concentration by mass of debris in the ice for each of the samples [i.e. (mass of debris / total mass of ice plus debris) x 100%]. Table 4.4 shows these results.

The majority of individual samples were too small to make sieving of each practical or reliable, so debris of the same type was lumped together to create three bulk samples:

Gígjökull: debris band sediments (971 g);

Gígjökull: basal ice (395 g);

Steinholtsjökull: basal ice (184 g).

Pooling of samples in this way is unlikely to prejudice the results because:

1. The possibility that individual samples of different debris types might be mixed together was not supported by my observations of the debris' field relationships.
2. My reading of the relevant literature did not lead me to believe that I was dealing with more than one type of debris band or more than one type of basal ice.
3. It is the overall pattern of the debris flux within ice, not small-scale variations in this, which is important for this thesis.

The samples were shaken through a full-phi interval sieve-stack using the range -4ϕ (16.0 mm) to 4ϕ (0.0625 mm). Each size fraction was weighed to the nearest 0.1 g, and the mass calculated as a percentage of the total mass. Frequency and cumulative frequency distributions of particle size were drawn up for each of the three samples. It was evident in several cases that a few large clasts >16.0 mm diameter exerted an undue influence on the particle size distributions. This class was therefore excluded from the analysis, a measure which can be further justified because clasts of this size are included in the clast form analysis of the adjacent ice-contact debris (see above). Fig. 4.10 and Fig. 4.11 show these results; see also Table 4.5.

Table 4.5 (overleaf)

Gígjökull and Steinholtsjökull: ice and debris analysis, summary statistics.

KEY:

Gígj.	Gígjökull
RCDB	Relict conduit debris bands debris
Stj.	Steinholtsjökull
Pdb/T B-L	Pré de Bar/Triolet bed-load

Summary Statistics, Ice + Debris Analysis				
	Gígj. RCDB	Gígj. Basal Ice	Stj. Basal Ice	PdB/T B-L
Mean debris conc., %	73.7	11.4	17.3	N/A
St. Dev., %	3.5	11.2	13.6	N/A
Mean EC, μS cm^-1	29.2	21.5	29.5	N/A
St. Dev., μS cm^-1	22.5	5.4	8.6	N/A
Mean particle size, mm	0.90	0.97	0.88	1.59
Minus 1 s.d., mm	3.36	5.28	4.40	4.89
Plus 1 s.d., mm	0.24	0.18	0.17	0.52

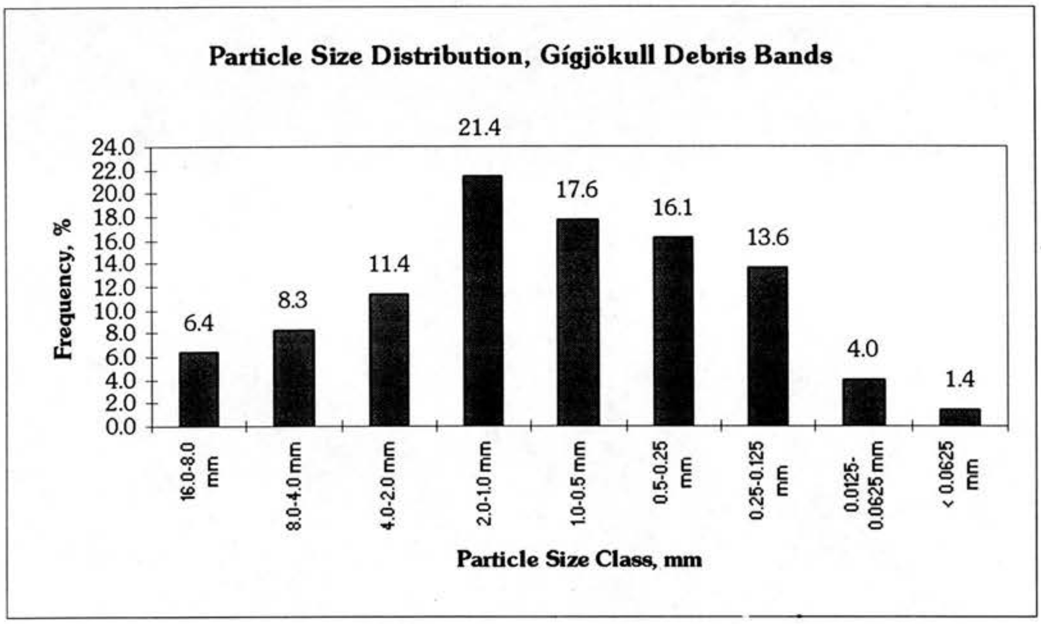


Figure 4.10a

Particle size distribution, Gígjökull relict conduit debris bands sediment.

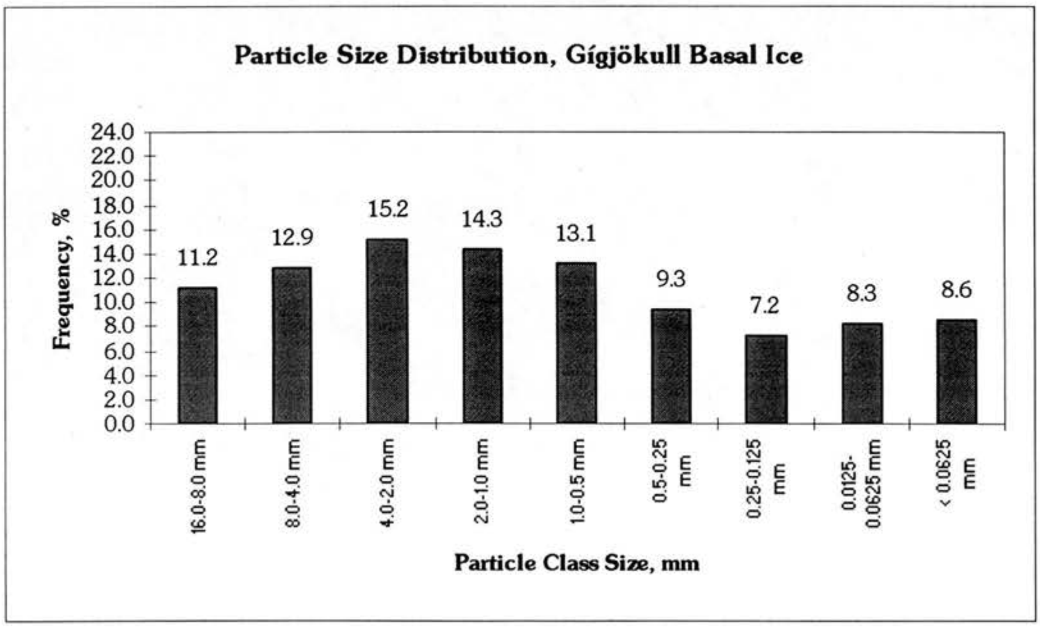


Figure 4.10b

Particle size distribution, Gígjökull basal ice debris.

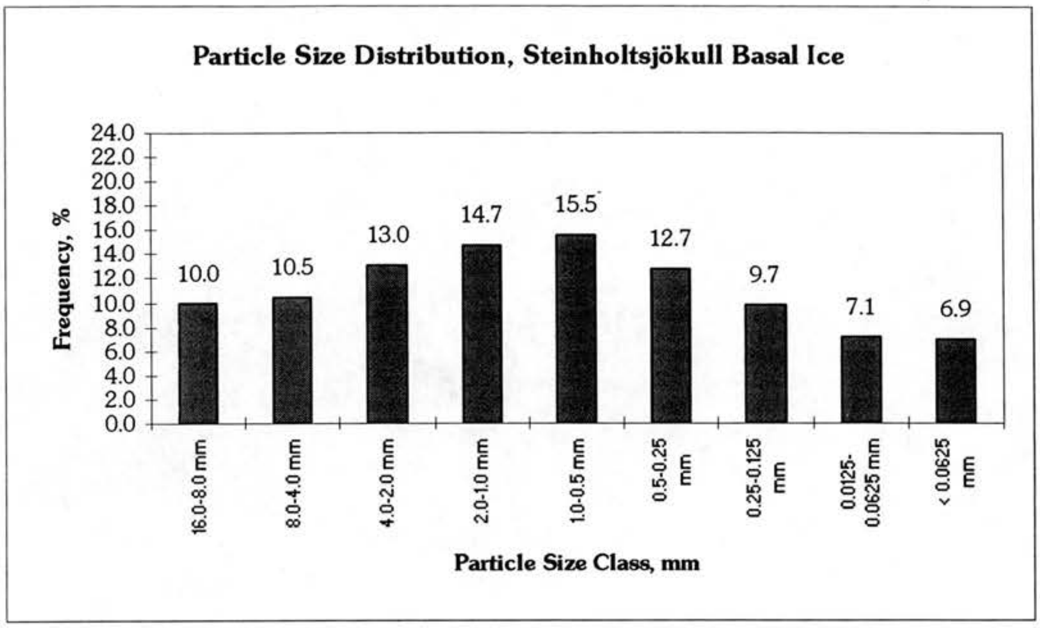


Figure 4.10c

Particle size distribution, Steinholtsjökull basal ice debris.

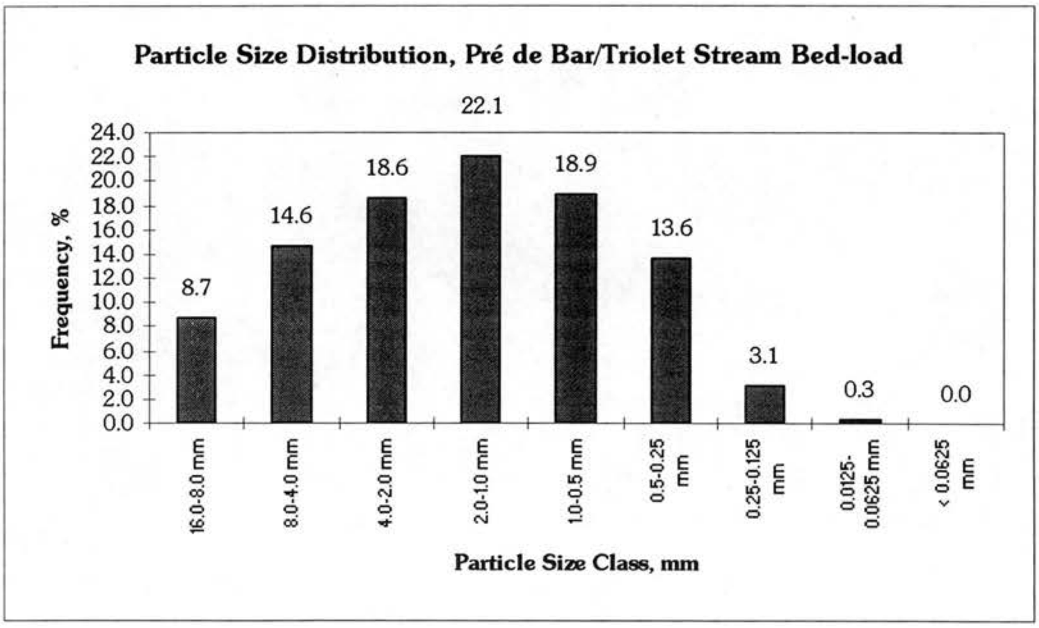


Figure 4.10d

Particle size distribution, Ghiacciao di Pré de Bar/Ghiacciao di Triolet outlet stream bed-load.

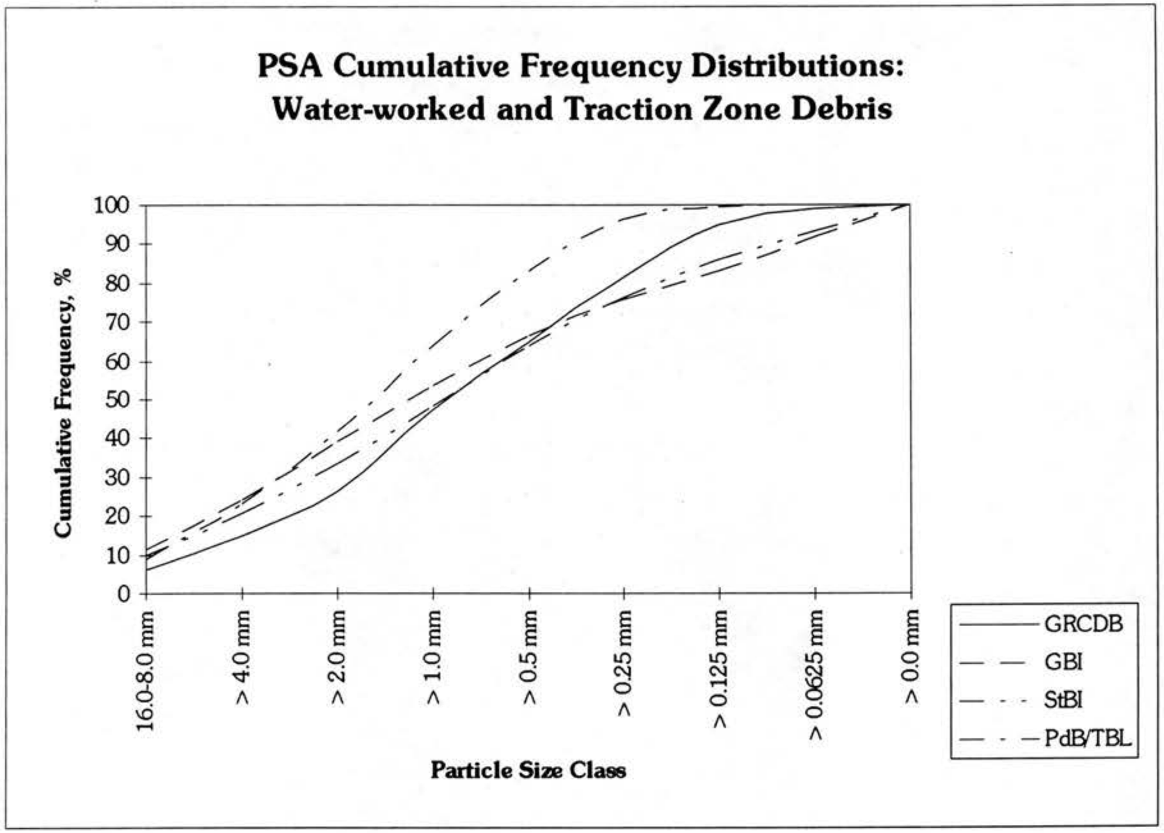


Figure 4.11

Particle size analysis: cumulative frequency distributions, clasts <16.0 mm *b* axis.

KEY:

GRCDB	Gígjökull relict conduit debris band sediments
GBI	Gígjökull basal ice debris
StBI	Steinholtsjökull basal ice debris
PdB/TBL	Ghiacciao di Pré deBar/Ghiacciao di Triolet outlet stream bed-load

4.3 DEBRIS TRANSPORT AT GÍGJÖKULL AND STEINHOLTSJÖKULL: OVERVIEW

The major features of the ice-marginal geomorphology and sedimentology are near-identical. The outstanding feature of both glaciers, in contrast to Sólheimajökull, is the large quantities of moraine actively forming. Fig. 4.9 illustrates the distribution of debris types; see also Table 4.5. Four distinct categories are identified:

1. **Rock-fall directly onto the ice surface.** This is important below the equilibrium line only; within the accumulation areas of the two glaciers (i.e. > ~1,100 m) the area of exposed bedrock is negligible. Rock-fall debris is everywhere distinct because of its high angularity (characteristic of passive transport), its colour (commonly brown rather than grey) and its position on the ice surface at the highest transport level (i.e. supraglacial transport below the equilibrium line). Although its volume is significant, it is not considered further here.

2. **Tephra from the 1947 Hekla eruption.** This occurs as a thin (< 2.0 cm) , tightly folded but coherent layer of gravel-sized clasts within the englacial ice. Only its upglacier outcrop is shown in Figs 4.4 and 4.9, below which the ice is covered in a patchy spread of tephra. However, although visually impressive, the contribution of the tephra to the total volume of sediment is negligible.

3. **Debris derived from a multiple series of englacial debris bands.**

4. **Debris derived directly from extensive sequences of dirty, crudely stratified ice.** This was identified immediately as 'basal ice'.

Together 3) and 4) account for the bulk of sediment currently contributing to moraine formation. Both debris bands and basal ice which appear similar were identified at Sólheimajökull (albeit with far less certainty); however, the frequency with which they appear there is negligible in comparison with the abundance of ice-marginal debris at Gígjökull and Steinholtsjökull. The next two chapters investigate the nature of the debris bands (Chapter 5) and basal ice (Chapter 6) in detail. Chapter 7 develops a single model of sediment transfer which ties these together in a way which accounts for this puzzling abundance of moraine forming activity.

CHAPTER 5
Gígjökull and Steinholtsjökull: Water-worked englacial debris

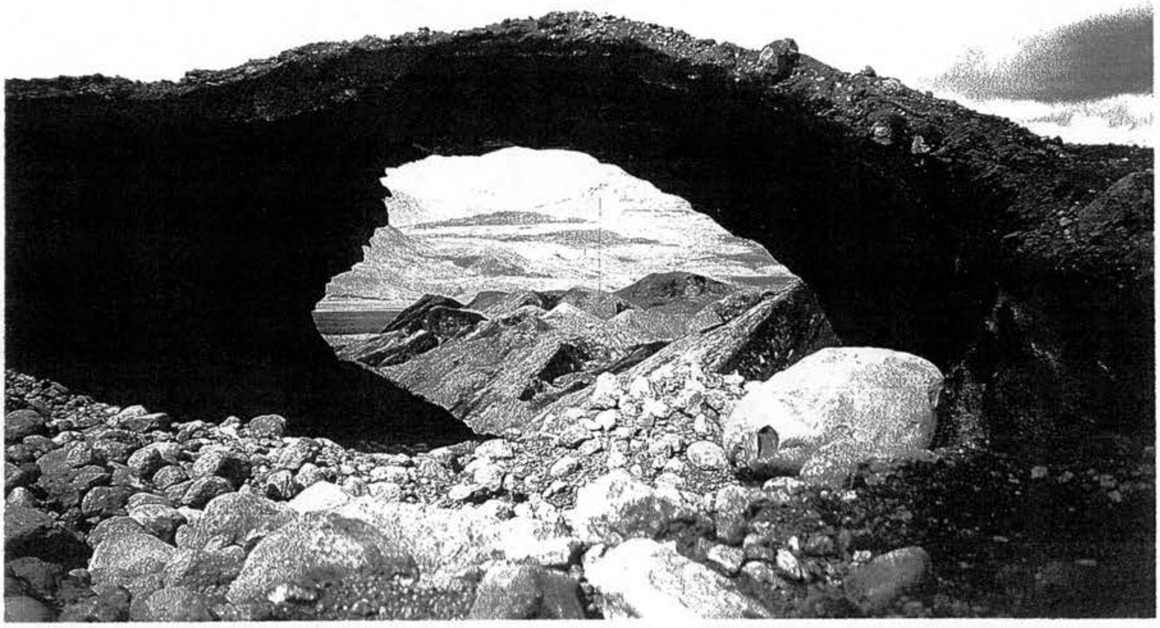
INTRODUCTION

This chapter examines the origin of the conspicuous englacial debris bands which feed much of the present-day moraine accumulation at Gígjökull and Steinholtsjökull. The clast and matrix properties of the sediments which make up these debris bands indicate water transport, so it seems that these debris bands must originate as sediments trapped within englacial waterways. These must once have flowed subglacially. It is unusual a) for water which runs at the glacier bed subsequently to take up an englacial flow route, and, b) for debris once entrained by water to pass back into transport by ice. However, both of these phenomena are consistent with our current understanding of the hydrology of overdeepenings. Meltwater drainage at Gígjökull and Steinholtsjökull acts directly to enhance moraine formation - the exact opposite of its impact at Sólheimajökull. This style of sediment transport and moraine formation does not fit the orthodox sediment transfer model. [Note: much of this chapter is published as part of Kirkbride and Spedding (1996).]

5.1 ENGLACIAL DEBRIS BANDS

Large areas of the termini of Gígjökull and Steinholtsjökull are buried beneath a stacked series of overlapping moraine ridges (Fig. 4.9). Typically these are several metres in height and asymmetrical in cross section, with a relatively gentle proximal (upglacier) slope and a steep distal (downglacier) face. The contact between the innermost of these moraine ridges and clean glacier ice is sharp: the debris which feeds these moraines is derived from a discrete source, rather than dispersed throughout the ice (Figs 5.1 and 5.2).

Sections exposed in moulins, and ice cliffs which form parts of the ice margins on which debris cannot rest, together with judicious excavation using an ice hammer, reveal a stacked series of debris bands (Figs 5.2 and 5.3) and channel fills (Fig 5.4) which feed the moraine ridges. The debris bands lie below the Hekla 1947 tephra layer, and above the basal ice, although at one small section exposed at the north-western margin of Gígjökull in 1994 debris



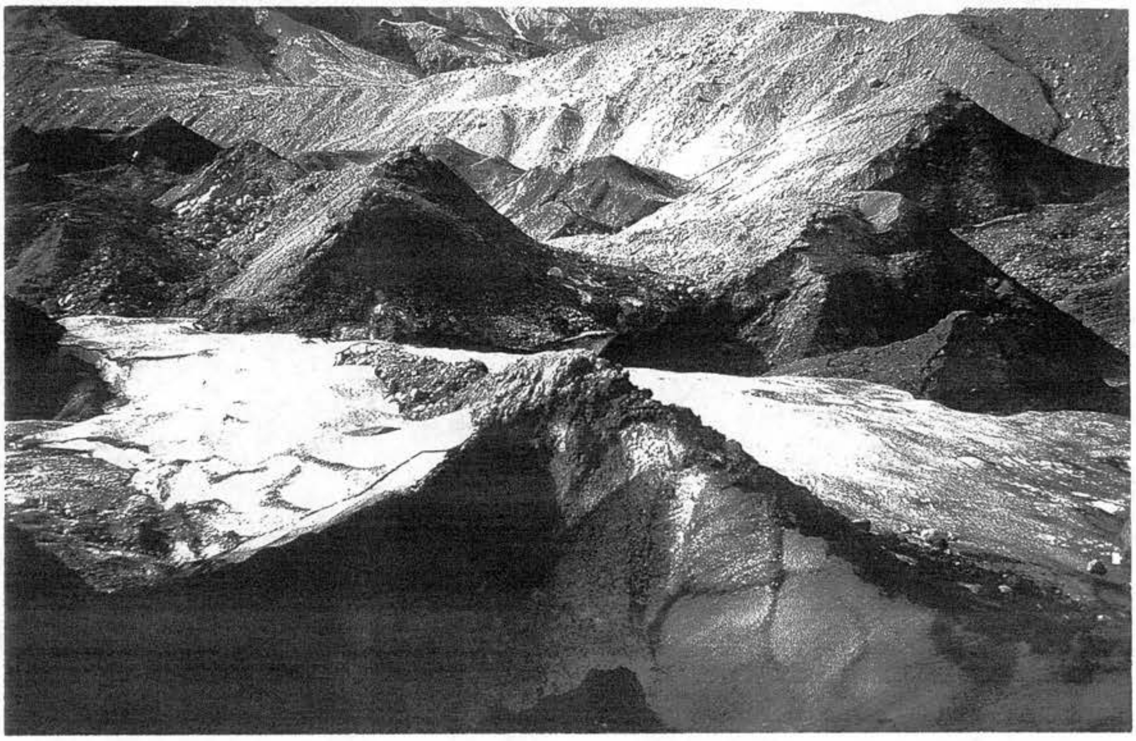


Figure 5.1 (top)

The debris-choked western margin of Gígjökull. Debris in this picture is entirely derived from englacial debris bands. One such debris bands provides the insulation which allows the roof of the arch to survive.

Figure 5.2 (bottom)

Gígjökull, western margin, sites G9-G12. Englacial debris bands (foreground) feed a stacked series of moraine ridges.

Figure 5.3 (top)

Steinholtsjökull, above sites S1-S5. Close-up view of englacial debris band. Note: 1) the rounded character of the large clast; 2) the sharp contact between englacial ice and debris band; and, 3) the dense concentration of debris, and absence of clear ice pockets, in the debris band (cf. Fig. 6.2).

Figure 5.4 (bottom)

Close-up view of large pocket of water-worked sediments, Gígjökull. Note the crude sorting and stratification. This debris sits on the surface several metres above the englacial/basal ice contact, and is preserved largely intact after the melt of interstitial ice.

bands were observed to cut through the basal ice. Patchy exposures - largely because so much debris chokes the ice margins - make it difficult to make accurate statements on debris band frequency; the best single exposure, in a moulin at the inward edge of Gígjökull's western moraine ridge (adjacent to sites G9-G12), revealed a series of four bands spaced roughly at 50 cm intervals. Typically the debris bands dip upglacier, are 5-20 cm thick (but channel fills are substantially thicker), and are laterally continuous for several tens of metres. Whereas crevasse exposures show the tephra layer to be highly convoluted in cross-section, the debris bands appear to be gently warped. I assume that the tephra is folded during its passage through the extensive ice-fall of each glacier, so the fact that the debris bands are not tightly folded in this way indicates an origin below the ice-fall.

The character of the debris, and its relationship with the surrounding ice, is distinctly different from that of the debris derived from the basal ice (see Chapter 6 for a full discussion of this basal ice). Key features which distinguish the debris bands are:

- **High debris content** (see Tables 4.4 and 4.5). The debris bands are made up of sediment which has ice within the interstices, rather than ice which contains debris. (Hitting these debris bands full on with the ice hammer felt not unlike striking solid rock!) Debris concentrations are much higher than in the basal ice. The mean debris concentration by mass of five samples taken from the debris bands at Gígjökull was 73.7%, s.d. ±3.5%, whereas the mean debris concentration of 34 basal ice samples taken from both glaciers was 13.1%, s.d. ±11.9%.

- **Particle size distribution** (see Figs 4.10 and 4.11). Although the mean particle size is similar, relative to basal ice, debris from the debris bands is better sorted, with a rather more peaked, unimodal distribution. The paucity of fine particles (< 0.125 mm: i.e. fine sands and smaller, which are common in basal ice) in the matrix material of these debris bands is notable. The particle size distribution of debris band matrix debris is similar in *shape* to that of bed-load/bed-material samples taken in proglacial streams: i.e. it reflects transport by water. To illustrate this I use a sample obtained (June 1996) from the Dora di Ferret, the outlet stream draining the Ghiaccaio di Triolet and Ghiaccaio di Pré de Bar, Courmayeur, Italy, which displays a similar degree of sorting. The comparison, however, is instructive rather than convincing: the Dora di Ferret has different energy levels and sediment supply relationships, plus possible bias because of fine sediment passing through the mesh of the Helly-Smith sample bag is unknown. Together, these factors are likely to explain the coarser mean particle size of the Dora di Ferret sample.

- **Sharp contact between debris and surrounding englacial ice.** The bands are enclosed by clear, coarsely crystalline ice in which bubbles are common, spherical, and evenly distributed: properties which suggest that the ice has not been subject to the intense stress-strain characteristic of the near-bed zone. Dirty basal ice (i.e. debris dispersed widely throughout the bulk of the ice) carries a characteristic signature of deformation and/or (partial) recrystallisation (e.g. Hubbard and Sharp, 1995; see Chapter 6). Where the debris bands were seen to cut through basal ice, the contact between the two was equally sharp, supporting the distinct identity of the debris bands. Electrical conductivity (EC) measurements further differentiate between the two types of ice (Table 4.4): the mean EC of melted samples of basal ice was 29 $\mu S\ cm^{-1}$ ($n = 34$, s.d. $\pm 16.4\ \mu S\ cm^{-1}$) whereas samples of melted englacial ice taken immediately above and below two debris bands exposed adjacent to sites G9-G12 registered 1.4, 1.6, 1.7 and 2.1 $\mu S\ cm^{-1}$. Although it is possible that this clear ice has been in contact with the bed without major modification, the contrast with the thick (>10 m) and well-developed sequences of what I identify with confidence as basal ice suggests otherwise. If the 'englacial' ice is really 'basal', a high degree of differential basal ice formation is required. Furthermore, sediment from these 'englacial' debris bands is found on the surface of Gígjökull >10 m above the sharp contact between 'englacial' and debris-rich ('basal') ice. The simplest explanation is that the contrast between 'englacial' ice and 'basal' ice is genuine.
- **Clast roundness** (Table 4.3). Clasts derived from the debris bands are significantly more rounded (mean roundness score between 3.30 and 3.92) than clasts from basal ice (mean roundness score = 2.60). The clast samples from debris bands fall into two groups, the second being the more rounded: S1-S5 plus G7 and G8 (Steinholtsjökull plus Gígjökull east) and G9-G16 (Gígjökull west). For details of statistical analysis see Box 5.1.

INTERPRETATION

The sub-rounded/rounded nature of the larger clasts, the crude level of matrix sorting and the relative paucity of fine particles within the matrix 1) distinguish the englacial debris from debris derived directly from the basal traction zone, and, 2) indicate it has been subject to the action of running water. The field relationships of the debris bands (i.e. within clean englacial ice) imply that at least the final part of this water-working takes place within an *englacial* conduit (s). I interpret these debris bands to represent relict englacial conduits. The relationship of the

BOX 5.1
Statistical testing of clast roundness differences

KOLMOGOROV-SMIRNOV (K-S) TEST

(e.g. Blalock, 1981, pp. 266-269.)

This is the non-parametric alternative to the two-sample **t**-test for difference in means. The K-S test requires: 1) independent, random samples, and, 2) ordinal data; it does not require continuous, interval-level data, nor does it require that parent populations follow normal distributions. The K-S test offers superior performance to the Mann-Whitney test if the ranked data include large numbers of tied scores. This makes the K-S test particularly suitable for samples - such as clast roundness counts - which consist of ordinal data split between several categories. It tests for significant difference between two samples; implicitly a significant difference in central tendency:

- H_0: the two samples are drawn from a single population.
- H_1: the two samples are drawn from separate populations - in this case, populations which differ significantly in terms of clast roundness.

The K-S test evaluates the difference between the cumulative frequency distributions of the two samples. The test statistic used (**D**) is the maximum difference between the two cumulative frequency distributions. The test assesses how likely it is that this difference can arise by chance (i.e. it uses a null hypothesis of 'no difference'). If 1) the combined sample size is large, and, 2) the direction of difference can be predicted (i.e. inspection of data predicts that sample B consists of clasts which are less angular than those of sample A) the sampling distribution of the test statistic **D** closely follows the χ^2 probability distribution. This permits rapid evaluation of the null hypothesis using χ^2 tables. Full details are given in Blalock (1981).

TEST SAMPLES

Field inspection and preliminary data survey suggest the presence of three distinct debris populations (ignoring rock-fall debris) if clast roundness is used as a discriminatory property (see Table 4.3):

1. Debris from basal ice: samples taken at Gígjökull only (G1-G6): **GBI**
2. Debris from englacial debris bands: Steinholtsjökull plus Gígjökull east (S1-S5 and G7-G8): **StRCD** and **GERCD**
3. Debris from englacial debris bands: Gígjökull west (G9-G16): **GWRCD**

This gives 'lumped' random samples of 396, 498 and 487 clasts respectively for use with the K-S test.

(PTO)

Box 5.1 (continued)

TEST A:

K-S Test, Clast Roundness: GBI vs. St/GE RCD					
Roundness Score	**BI**	**Proportion**	**RCD**	**Proportion**	**Difference**
1	6	0.015	2	0.004	0.011
Up to & including 2	158	0.399	25	0.050	0.349
Up to & including 3	391	0.987	300	0.602	**0.385**
Up to & including 4	395	0.997	471	0.946	0.051
Up to & including 5	396	1.000	495	0.994	0.006
Up to & including 6	396	1.000	498	1.000	0.000

Maximum difference **D** = 0.385

$\chi^2 = 4\ (0.385)^2\ [(396 \times 498) / (396 + 498)] =$ **130.8** (2 d.f.)

Significance > 99.9% (p = 0.001)

REJECT H_0: the difference between the two samples is genuine. Debris from the debris bands is significantly different to debris from basal ice.

TEST B:

K-S Test, Clast Roundness: St/GE RCD vs. GW RCD					
Roundness Score	**St/GE RCD**	**Proportion**	**GW RCD**	**Proportion**	**Difference**
1	2	0.004	0	0.000	0.004
Up to & including 2	25	0.050	5	0.010	0.040
Up to & including 3	300	0.602	163	0.335	**0.267**
Up to & including 4	471	0.946	390	0.801	0.145
Up to & including 5	495	0.994	485	0.996	0.002
Up to & including 6	498	1.000	487	1.000	0.000

Maximum difference **D** = 0.267

$\chi^2 = 4\ (0.267)^2\ [(498 \times 487) / (498 + 487)] =$ **70.2** (2 d.f.)

Significance > 99.9% (p = 0.001)

REJECT H_0: the difference between the two samples is genuine. Debris band debris from the western margin of Gígjökull is significantly different to debris band debris from Steinholtsjökull and the eastern margin of Gígjökull.

TWO-SAMPLE t/z-TEST

The advantage of the K-S test is that its use does not rely on rigorous assumptions about the quality of data used. This is offset by the fact that non-parametric tests tend to be less powerful than their parametric equivalents: i.e. the chance of a Type II error (i.e. accepting a false null hypothesis) is higher. This is not important here because of the high level of confidence with

(PTO)

Box 5.1 (continued)

which I *reject* the null hypotheses. However, there is a further disadvantage to use of the K-S test. If the K-S test enables us to accept a significant difference between the two samples, it is not clear exactly where this difference lies: within central tendency, form or dispersion? In contrast, successful rejection of the null hypothesis using a difference-of-means test specifies the quality wherein lies the significant difference. Quantitative discrepancies in the mean roundness scores of different samples/populations intuitively carry an advantage over rather vague, qualitative notions of 'difference', so here I assess briefly the use/adequacy of the parametric difference-of-means test:

1) Large sample size negates the requirement that population roundness scores follow a normal distribution. If $n \geq 120$ the **t**-distribution follows the **z**-distribution.

2) These are high quality ordinal data. Although the assessment procedure used is crude, the ideal roundness categories into which clasts are placed stand in fixed relation to each other: Powers' original scale uses mean radius of curvature (i.e. interval data) to specify categories, using a log transform to enhance its utility with respect to less angular clasts (Powers, 1953).

3) 1 and 2 remove two major objections to use of the **t**/**z**-test, but the absence of continuous data stands. However, the **t**-test is 'robust' under a wide range of conditions (i.e. it works, despite the fact that test conditions violate the assumptions of the ideal statistical model). Given the extra information on the nature of difference attached to the **t**-test, it makes sense to use this in parallel with its non-parametric alternative (Blalock, 1981, pp. 274-275).

Z-test, Clast Roundness Data			
Sample	**Size (n)**	**Mean (x)**	**S.D. (s)**
Gígjökull Basal Ice	396	2.60	0.54
RCD: Steinholtsj. + Gígj. E	498	3.40	0.50
RCD: Gígj. W	487	3.83	0.75

Test statistic: $\mathbf{z} = (x_1 - x_2) - 0 / \sqrt{[(s_1^2 / n_1) + (s_2^2 / n_2)]}$

Test A: GBI vs. St/GE RCD

H_1: Mean roundness of Steinholtsjökull and Gígjökull east debris band debris > mean roundness of Gígjökull basal ice debris.

z = -22.85; vanishing probability that difference in means is zero.

Test B: St/GE RCD vs. GW RCD

H_1: Mean roundness of Gígjökull debris band debris > mean roundness of Steinholtsjökull and Gígjökull east debris band debris.

z = -10.75; vanishing probability that difference in means is zero.

VERDICT

These tests mirror the results of the K-S tests, and support use of mean roundness scores to separate different debris populations.

debris bands to the tephra suggests that the process by which this debris is re-entrained by ice to form the debris bands occurs in the region beneath the ice-falls. In turn this implies that at least part of the internal drainage of the terminal lobes of Gígjökull and Steinholtsjökull is through englacial rather than subglacial routeways. The fact that the debris from the western sample sites at Gígjökull is significantly more rounded suggests that conduit drainage is better developed here than beneath the eastern side of Gígjökull and beneath Steinholtsjökull in general.

5.2 NOVELTY?

Here I discuss the novelty of the debris bands and the associated moraines found at Gígjökull and Steinholtsjökull in terms of two questions:

1. Have these kind of water-worked debris band features been described before?
2. Is my interpretation of the debris bands' origin (i.e. within englacial conduits) robust?

PREVIOUS STUDIES

Similar moraine deposits composed of water-worked debris emerging from englacial ice are found at the Mueller and Tasman Glaciers, New Zealand (Kirkbride, 1989, 1993; Kirkbride and Spedding, 1996), and at Hofðabrekkujökull, Iceland (Näslund and Hassinen, 1996). These papers acknowledge the resemblance of these sediments to 'traditional' fluvio-glacial esker deposits, but talk firmly in terms of moraines (*sensu stricto*) fed by debris bands. Röthlisberger (1979), Mathews (1979), Schlücter (1983) and Rubulis (1983) also describe water-worked debris carried into ice, subsequently to be abandoned, by englacial streams. Price was first to discuss in detail ice-cored spreads of water-worked debris, as found by him at Casement Glacier, Alaska, USA (Price, 1966) and Breiðamerkurjökull, Iceland (Price, 1969); see also Boulton (1967). These sediments appeared to have been laid down within anastomosing channel networks, so Price described them not as moraines, but as 'englacial or supraglacial esker complexes'. This perhaps calls into question the common textbook distinction between features of 'pure' glacial origin and features of fluvio-glacial origin (Chapter 1.3): it is clear that the sediments featured in all these studies, whether designated as moraines or eskers, share a common origin related to sediment transport in englacial channels. Warren and Ashley (1994) discuss the (mis)use of the term 'esker', which they argue is best restricted to elongate and sinuous ridges of water-worked sediment which represent a *direct* record of a glacier's drainage system. This definition includes both the textbook subglacial channel-fill esker, and Price's en/supraglacial eskers (which Warren and Ashley term fluvial ice-channel-fill eskers). However, it excludes the features at Gígjökull and Steinholtsjökull, which seem to represent an *indirect* record of glacier drainage: it is difficult to find clear evidence of past drainage structures in the appearance of the moraine ridges. Three factors are likely to explain this. Working backwards, these are:

- Re-mobilisation of debris after its appearance at the ice surface.
- Deformation of channel-fill deposits by a subsequent phase of transport within ice. The greater the distance of ice transport and/or the greater the intensity of deformation of the enclosing ice the less likely it is that the channel-fill deposits will preserve their original identity: e.g. Knudsen (1995) describes Price-type sinuous englacial eskers transformed into 'concertina' eskers by the 1964 surge of Brúarjökull, Iceland.
- Structure of the original drainage network. If the effect of these first two factors is weak, the eventual deposit is likely to match closely the original pattern of drainage. 'Typical' eskers - whether sub-, en- or supraglacial - reflect discrete, sizeable channels. Whereas the large channel-fill deposits found at Gígjökull seem to match this image (Fig. 5.4), it is likely that the debris bands relate to some kind of *dispersed* englacial drainage network: i.e. a large number of small, interconnected channels and water pockets (see Hooke and Pohjola, 1994; and below, 5.6). Fossil evidence of major channels cannot exist if such channels are missing!

So: what is new about this case study? I take the work of Price as the benchmark by which to judge mine because: a) it set a precedent (at least in the English language) for detailed studies of water-worked *englacial* ice contact debris; and, b) much of it stems from work in Iceland, and so invites direct comparison with my work here. The features Price describes are certainly similar to those found at Gígjökull and Steinholtsjökull, even if he and I find different answers to the question "when does an esker cease to be an esker?" However, three important factors differentiate my work from his:

- **Ice dynamics.** Price describes esker complexes found in channels which drain thin, sluggish or stagnant ice, whereas the ice of the termini of Gígjökull and Steinholtsjökull is relatively thick and fast-flowing.
- **Source of debris.** Price's eskers form in channels which originate at the glacier surface, and gather their sediment load from abundant spreads of supraglacial moraine. This is not the case at Gígjökull and Steinholtsjökull: sediments which make up the debris bands here must be picked up by *subglacial* water flows (see below, 5.4).

- **Specific context.** The debris bands at Gígjökull and Steinholtsjökull seem to be the specific products of water flows which must traverse terminal overdeepenings. This is not the case with the examples Price describes.

ALTERNATIVE INTERPRETATIONS

My tactic here is to review alternative hypotheses which possibly explain how these debris bands form. I try to show that these ideas are less likely to be correct than is the relict englacial conduit hypothesis. The three hypotheses I consider are:

1. Passive and/or active debris transport origin, as described by orthodox sediment transport theory.
2. Thrust origin.
3. Entrainment of debris by ice override, and subsequent folding of basal ice.

'Orthodox' sediment transport theory (see Chapter 1.2)

'Standard' accounts of high-level debris bands tend to invoke either the passive englacial transport of debris incorporated into the ice from supraglacial sources above the firn line, or basal debris elevated from the bed at the meeting place of two ice streams, or, indeed, both together (see Small, 1987a and Kirkbride, 1995a, pp. 273-275 for reviews). The sharply-defined debris bands which feed the medial moraines of the Glacier de Tsidjiore Nouve (Small and Gomez, 1981; Figs 3, 4 and 5) and the Haut Glacier d'Arolla (Gomez and Small, 1985; Figs 3 and 4) (both in Canton Valais, Switzerland) certainly appear similar to the debris bands at Gígjökull and Steinholtsjökull. However, these Swiss examples are made up of coarse (>2.0 mm), angular debris, characteristic of passive transport.[1] The debris signature of the englacial debris bands is clearly different to that of debris associated with other transport pathways which operate at Gígjökull and Steinholtsjökull (i.e. rock-fall debris in supraglacial transport, or debris contained in basal ice: Tables 4.3 and 4.5; Figs 4.9, 4.10 and 4.11). This leads me to reject conventional passive or active transport pathways as possible explanations for the origin of these englacial debris bands.

[1] The debris band which feeds the Glacier de la Mitre moraine (which forms at the confluence of the Mitre and Haut Arolla Glaciers), is an exception; this contains more rounded clasts and a finer matrix characteristic of active transport within the basal traction zone.

Thrust origin

Recent work has revived the possibility that debris may be carried from the basal transport zone into high-level transport by thrusting (e.g. Hambrey *et al.*, 1996; Bennett *et al.*, 1996; cf. Boulton, 1967). Debris so entrained seems to form discrete, continuous bands superficially similar to those at Gígjökull and Steinholtsjökull, which feed similarly asymmetrical moraine ridges. The debris bands and ridges at Sørbreen, Spitsbergen which Boulton (1967) describes are, in fact, made up of rounded, sorted, stratified sediments, elevated by thrust action as ice-advance overrides proglacial outwash. However, it is unlikely that water-worked debris represents the only source of basal sediments available to feed hypothetical thrusts at Gígjökull and Steinholtsjökull; the abundance of traction zone debris present both in basal ice (Chapter 6) and the Neoglacial moraines (Chapter 8) suggests otherwise. Thrust action is likely to preserve the clast properties of active transport, in which case why is it that debris band sediments are clearly different to basal traction zone sediments? The simple answer is that the debris bands studied are not of thrust origin. Additional evidence supports this conclusion: the close spacing, wavy appearance, shallow dip, limited modification of surrounding ice (cf. the mylonitisation of thrust-zone ice identified by Tison *et al.*, 1993), and lack of clear connection to the glacier bed typical of the debris bands observed fail to match published accounts of thrust-origin debris bands. However, the possibility that thrust action contributes to redistribution of debris cannot be ruled out (although no evidence of fresh thrust activity was seen). The compressive flow regime of the two glaciers, enhanced as ice encounters the adverse slope at the exit of their terminal overdeepenings, certainly creates the possibility of thrust development, so it is not implausible that the debris bands act as *pre-existing* planes of weakness which permit some degree of thrust displacement.

Ice override and folding

Sugden *et al.* (1987) describe debris bands at Jakobshavns Isbræ, Greenland which originate as basal ice. Sediment is entrained by subglacial freezing-on, and subsequent tectonic deformation thickens, folds and elevates the debris bands so formed. Widespread freezing-on of debris is unlikely to occur at Gígjökull and Steinholtsjökull because it is too warm, but debris entrainment by ice override (e.g. Evans, 1989; Sharp *et al.*, 1994) is a possibility given their history of advance since *c.* 1960. This makes it important to consider the hypothesis that the debris bands form by folding of debris entrained at the glacier bed. My view is that this is unlikely:

- Subglacial entrainment of debris as ice advances implies the formation of solid facies basal ice (Chapter 6). The continuously sharp, curvilinear contact which exists between debris band sediments and enclosing ice is not characteristic of solid facies basal ice, which usually displays a highly irregular, discontinuous ice-debris contact.

- If these debris bands represent some kind of deformed basal ice (which I doubt), how is it that it sits on top of a second type of basal ice with very different properties? This returns us to the problem of differential basal ice formation (see above, 5.1). Two distinct types of basal ice imply two distinct subglacial regimes. It is difficult to explain the process switch in simple terms. The fact that the debris bands sit above the (genuine?) basal ice implies that they must be older; however, debris bands were also found to cross-cut basal ice, which indicates that the debris bands are younger than the basal ice. This difficulty disappears if we accept that the debris bands are not some kind of basal ice feature, but form in once-subglacial channels which leave the glacier bed.
- Folding of debris bands together with adjacent debris-poor ice (as this hypothesis requires) implies a fairly close match of bulk ice structures. This is not seen. The debris bands fail to conform with other structures found in the ice (the H1947 tephra layer, lineations of basal ice, foliation of englacial ice); in fact, the debris bands commonly cross-cut other structures. Such divergence is difficult to reconcile with widespread tectonic deformation of ice as seen at Jakobshavns Isbræ (Sugden *et al.*, 1987, Fig. 2a).

Verdict

These alternative hypotheses fail to provide a satisfactory explanation of debris band origins. At best the ideas I consider above demand what seems to be unnecessarily complicated behaviour of ice and debris if they are to work, and so fail to pass the test of Occam's Razor. The most straightforward explanation which best fits the observations - notably the facts that the debris bands a) are enclosed in englacial ice, b) form a sharp contact with this surrounding ice, and, c) contain a distinct sediment population consistent with water transport - is that the debris bands form in some kind of englacial channel. In the rest of this chapter I examine this idea in depth.

DISCUSSION

5.3 ORIGIN OF THE DEBRIS BANDS

Much of the moraine presently accumulating at Gígjökull and Steinholtsjökull appears to have reached the ice margin by a mechanism which is not easily explained by a straightforward transport pathway model (Chapter 1.2). A large proportion of the debris reflects transport by running water, with clast modification induced by the tumbling action of turbulent flow, yet it is deposited directly out of englacial ice; hence my interpretation of these debris bands as relict conduit fills.

Standard accounts tend to distinguish between distinct and separate transport pathways associated either with ice itself, or with water within or beneath ice. Each involves distinct processes of clast modification, and ultimately produces a distinct landform, either glacial (moraines) or fluvio-glacial (eskers, etc.) (Chapter 1.3). Sediment may be entrained by ice initially, but once it reaches the glacial drainage network it is assumed 'lost' to water transport, whereupon it is either carried away beyond the ice margin, or accumulates within the conduit to form an esker. However, in this case sediment characteristic of water transport is deposited directly out of ice. Thus debris initially transported by ice (whatever its source and level of transport) is subsequently entrained by water, and carried some distance sufficient to permit modification to clast shape, before passing back to high-level transport [i.e. above the basal transport zone (Boulton, 1978)] within the ice, and eventual deposition at the ice margin. Transport pathways within many valley glaciers may therefore be more complex than is often assumed, and the configuration and behaviour of the glacial drainage network may also act as a control on ice-margin geomorphology and sedimentology as well as on subglacial processes. The importance of these wider, catchment-scale links tends not to figure as a major item in published discussions of the controls on ice-marginal sedimentation, however: in large part it seems precisely because these wider links transcend the established, reductionist sub-disciplines of traditional glacial and fluvio-glacial studies (Chapter 1.4).

This relict conduit-fill interpretation of active ice-marginal sedimentation raises three further questions:

1) How does the debris get into the conduits?

2) Why are the conduits located within the main body of the ice, rather than running at its bed as is commonly assumed/observed?

3) Why is the debris returned from water transport to ice transport?

5.4 HOW DOES DEBRIS GET INTO THE CONDUITS?

Ice flow which converges on a conduit (Chapter 1.1) will carry with it a steady flux of debris if the ice is dirty. Alternatively, a conduit which migrates through debris-rich ice can be expected to entrain large quantities of sediment (see Chapters 2.5 and 7.1 for a full discussion). This debris can be picked up from the ice surface by a supraglacial stream which later becomes englacial, as Price (1966, 1969) suggests for Casement Glacier and Breiðamerkurjökull. Streams within the ice can tap englacial debris dispersed throughout the ice. This is likely to happen at many alpine-type glaciers which receive large supraglacial inputs of debris above the EL. As basic transport pathway theory describes, this debris tends to be incorporated into the ice, travels along high-level englacial pathways, and returns to the surface in the ablation zone where ice flow trajectories are directed upwards. This makes it available to englacial streams. With reference to the Mueller and Tasman Glaciers, Kirkbride and Spedding (1996) anticipate a process whereby conduits running at high-level within such debris-rich ice 'hijack' its debris load.

With small outlet glaciers such as Gígjökull and Steinholtsjökull, however, this kind of mechanism cannot apply. With the exception of the sporadic airfall input of tephra (Hekla 1947 being the most recent example) supraglacial debris sources above the EL are negligible here. Debris is not dispersed throughout the ice of the terminal lobes in the same way that it is, at, say Mueller Glacier, or Glacier de Tsidjiore Nouve. Debris at high-level below the EL must be derived from *subglacial* sources, whereafter it is carried from the basal transport zone into an englacial position. I infer that conduits collect debris, presumably from basal ice or loose debris whilst running at, or close to, the glacier bed: see 5.5, below.

5.5 WHY ARE THE CONDUITS ENGLACIAL?

ENGLACIAL DRAINAGE IS UNSTABLE...

The debris band evidence of high-level drainage presents something of a puzzle; the more so if it is required - as here - that water which runs through these englacial conduits must carry a large debris load. The physics of ice-and-water interactions implies that water tends to find its way towards the bed of the glacier, at which point flow becomes stable (Röthlisberger and Lang, 1987; Röthlisberger, 1993; Shreve, 1972):

The density of water exceeds that of ice, so:

1. Water which part-fills a crack/tube of ice will tend to work its way downwards. This is because the stress difference between ice and water is greatest at the water surface; the rate of closure here exceeds the rate of closure at the base of the crack, so driving the water towards the glacier bed.

2. Large cracks/tubes in ice entirely filled with water will tend to enlarge themselves and propagate towards the glacier bed by the process of hydraulic fracturing. This is because the excess pressure of the water column relative to the surrounding ice rises as water depth increases.

3. The distribution of hydraulic potential within an ice mass [defined by Shreve (1972) as the sum of the elevation head and the pressure head: Chapter 1.1] will direct water towards the glacier bed. This is because water density imparts elevation head, whereas ice density imparts pressure head; replacing x metres elevation with x metres of ice overburden as ice depth rises involves a fall in hydraulic potential.

Running water generates frictional heat which melts the surrounding ice, so:

4. If the tube/channel is only part-filled with water, wall melt will be directed preferentially against the ice at its base, creating a tendency for the tube/channel to sink within the ice. The fall in the pressure melting point of water as pressure rises also supplies heat to reinforce this tendency for downwards melt.

1, 2 and 4 are thought to explain why crevasses/moulins extend into major subglacial passageways, whereas 3 is invoked to justify widespread flow of water towards the bed by seepage within the vein network.

...SO WHY DO WE GET ENGLACIAL DRAINAGE?

Theory implies it should be otherwise, but reports of englacial drainage at wholly temperate glaciers are common (ice below the pressure melting point is not a factor here). These reports are usually justified by appeal to one of two ideas: the gradient conduit (e.g. Burkimsher, 1983b) or what I call the 'transient' conduit (e.g. Stenborg, 1973).

Gradient conduit. Röthlisberger (1972) defines the gradient conduit as a channel which sits within the ice at its hydraulic grade line; it occupies the highest level possible so that, although flow occurs under closed conditions (i.e. the conduit is completely full), water pressure is atmospheric. However, it seems that the gradient conduit may exist as a theoretical curiosity only (Hooke, 1984, pp. 183-184; Röthlisberger and Lang. 1987, p. 259): the idea of the gradient conduit sits somewhat uncomfortably alongside arguments 1 to 4 given above, so that, despite the fact that theoretically it can persist as a stable feature of a glacier's drainage system, it is difficult to see why it should ever establish itself in the first place.

Transient conduit. Channels which start at the ice surface can reach the edge of the glacier (e.g. if near-surface drainage is controlled by crevasses) or dissipate before they sink as deep as the glacier bed. I use the term 'transient' to indicate that these fail to reach a stable position with respect to ice-and-water physics. It is feasible that ice flow directed towards the glacier surface, and/or advection of heat released by running water extend the distance a conduit survives at high-level (the first of these counteracts the tendency of the channel to sink by whatever means, the second delays channel-bottom melting; both might be important in the case of sizeable channels which run through a zone of strongly compressive flow). This type of channel appears to give rise to Price's englacial eskers, but is less likely to figure at Gígjökull or Steinholtsjökull.

Debris supply

The issue of debris supply provides the crucial clue which demonstrates that neither the gradient conduit nor the transient conduit can give a satisfactory explanation of debris band origin at Gígjökull and Steinholtsjökull. Both ideas support the presence of englacial channels within glacier termini, but neither can explain the fact that these englacial channels must carry a sizeable sediment load if they are to give rise to debris bands of the type found at Gígjökull and Steinholtsjökull. If englacial streams originate at the ice surface, but never run at its bed (as is the case with both gradient and transient conduits), how is it that they carry large quantities of sediment if that sediment must come from the glacier bed (see above, 5.4)? Three possibilities suggest themselves:

1. Thrusting carries subglacial debris into high-level transport, so it is made available to englacial streams. This is plausible, although there is no field evidence to support this idea.

2. Elevation of basal ice carries debris into high-level transport so it is made available to englacial streams. The stratigraphic relationships make this unlikely: the (high-level) basal ice exposures at Gígjökull and Steinholtsjökull are found below, outside and downglacier of the relict conduit debris bands (Fig. 4.9). This is not what is required if high-level basal ice is to supply large quantities of debris to englacial streams: a suitable debris source must be located below, *inside*, and *upglacier* of the debris bands (ice and water rarely flow upglacier!).

3. **Water (not ice) carries the debris into high-level transport.** This interpretation best matches the field evidence. I infer that conduits collect debris - loose debris, or debris from basal ice - whilst running at the glacier bed. These conduits subsequently leave the bed to take up an englacial route, so carrying their debris into high-level transport. The section of basal ice in which debris bands were seen to cut layers of basal ice at a high angle (above, 5.1) lends empirical support to this inference. This scenario best explains the level of sediment sorting and clast rounding typical of the debris band material, which is likely to require transport of sediments by water for a considerable distance. This is not a problem if debris-laden conduits flow *subglacially* for perhaps two

or three kilometres (Gígjökull is 6.5 km long), but the first two hypotheses, which consider only englacial water flow, allow for just several hundred metres water transport at best (i.e. following elevation of subglacial debris towards the terminus).

Englacial drainage: overdeepening override

The field evidence fails to support the first two hypotheses, which carries the important implication that neither the concept of the gradient conduit nor that of the transient conduit can help explain why the debris bands form. However, field evidence does provide firm support for the third hypothesis. This redefines the key problem: i.e. **why is it that debris-laden streams leave the glacier bed to take up an englacial route?** Current thinking suggests this requires high water pressures - necessary to drive water *away* from the glacier bed, against what is otherwise the most favourable flow path. Flood events commonly give rise to suitably high water pressures (e.g. Warburton and Fenn, 1994), but these are short-lived. Persistent high water pressures and englacial drainage are thought to relate to situations in which water must traverse an overdeepening. Given the evidence for overdeepenings beneath the termini of both Gígjökull and Steinholtsjökull (Chapter 4.1: Fig. 4.5), this idea clearly merits further investigation.

5.6 HYDROLOGY OF OVERDEEPENINGS

High water pressures are required to force water through an overdeepening. Water backs up within a conduit, causing water pressures to rise, and flow within the conduit tends towards instability as its low pressure advantage relative to adjacent areas of distributed drainage falls away. Four different ideas exist as to what happens next (Fig 5.5); two of these invoke a switch from subglacial to englacial drainage.

WATER FLOW IN CONDUITS: BEDROCK TOPOGRAPHY AND THE MELTING-FREEZING TRANSITION

Water flows from high to low pressure, so, if the water is not to freeze, as the water pressure falls, heat must be supplied to the running water to raise its temperature to the higher pressure melting point. Usually this heat is supplied by viscous dissipation (friction) within the running water: approximately one-third of the heat generated by flow through a horizontal conduit is needed to warm water to the rising pressure melting point as the water pressure falls (Röthlisberger, 1972, p. 179). See Box 5.2a.

BOX 5.2

Simple pressure and temperature relationships within conduits (after Röthlisberger, 1972)

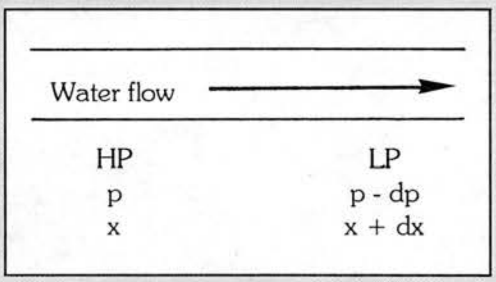
These simple equations show that, given a sufficiently steep adverse bed slope, basal conduits will tend to freeze up. This conclusion follows from the possibility that the positive **dz/dx** term may exceed the negative **dp/dx** term (see below for explanation). It is independent of the specific values of water pressure and discharge involved.

5.2a LEVEL CONDUIT

In this case, the change in elevation head (= height of conduit above a given datum) is zero. Total water pressure (strictly, hydraulic potential) = pressure head (local water pressure). Assumptions:

1. Steady flow, without acceleration.
2. Instantaneous transfer of heat generated by water friction.
3. No external sources of heat.

Water flows from high pressure (HP) to low pressure (LP):



p — pressure head, or local water pressure within conduit; equivalent to height of hydraulic grade line above glacier bed if conduit flows at glacier bed; independent of conduit elevation

x — distance downstream

d — denotes 'change in'

Total energy loss per unit time per unit length of conduit (**-dE**) = stream power = discharge-slope product. Note: minus terms indicate energy loss!

$$\mathbf{-dE = -Q\,dp}$$

Q — water discharge through conduit

(PTO)

Box 5.2a (continued)

Total energy loss is partitioned between:

1. heat required to adjust water temperature to the falling pressure melting point (**dE_t**), and
2. heat available to melt ice of conduit walls (**dE_m**):

$$-dE = -dE_t + -dE_m$$

$$-dE_m = -dE - -dE_t$$

Proportion of heat taken up by **dE_t**:

$$-dE_t = - 0.316\ Q\ dp$$

The constant 0.316 includes terms for: 1) the change in the pressure melting point of H_2O with a unit change in pressure; 2) specific heat capacity of water; and, 3) density of water. The constant 0.316 applies to air-free water; the appropriate constant for air-saturated water is 0.413. So:

$$-dE_m = - 0.684\ Q\ dp = - 0.684\ dE$$

IMPLICATION

- Within a level conduit approximately two-thirds of heat released by water goes to melt the ice walls.

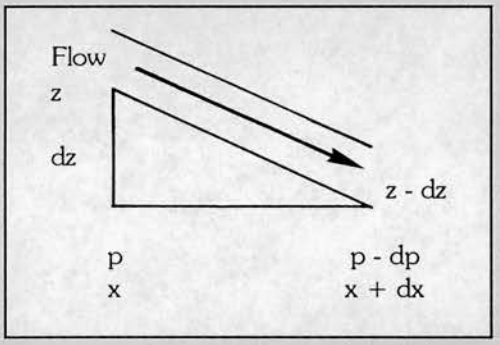
(PTO)

5.2b INCLINED CONDUIT

In this case, the downstream change in elevation of the conduit introduces an additional term: elevation head:

elevation head = ρ_w g dz

ρ_w density of water
g gravitational acceleration
z height of conduit/glacier bed above given datum



Total water pressure (hydraulic head = **f**) = pressure head + elevation head:

$$\mathbf{f = p + \rho_w g z}$$

$$\mathbf{-df = -dp + \rho_w g dz}$$

Water flow downhill: ρ_w **g dz** is negative (i.e. **-dz/dx**)

Water flow uphill: ρ_w **g dz** is positive (i.e. **+dz/dx**)

$$\mathbf{-dE = - Q df}$$

$$\mathbf{-dE_t = - 0.316 Q dp}$$

(From above; pressure melting point is independent of elevation head.)

So:

$$\mathbf{-dE_m = Q (-df - - 0.316 dp)}$$
$$\mathbf{= Q (-dp + \rho_w g dz - - 0.316 dp)}$$
$$\mathbf{= Q (- 0.684 dp + \rho_w g dz)}$$

IMPLICATIONS

- **$-dE_m$ = 0** (no ice-wall melting) when **- 0.684 dp + ρ_w g dz = 0**: i.e. a critical value of uphill slope **+dz/dx** exists whereupon all flow energy not used to warm up water is used to raise the water's elevation head, so no energy is left for melting.
- When **ρ_w g dz > - 0.684 dp**, **dE_m** is positive: water gains heat (required to raise its temperature to the higher pressure melting point). Negative **dE_m** = wall melt; positive **dE_m** = wall *freezing* - water gains heat from latent heat of fusion as new ice forms.

In the case of an inclined conduit, the energy released by a drop in local water pressure is supplemented by the change in energy (loss or gain) related to the downstream change in elevation (fall or rise) of the conduit: the total hydraulic gradient is the sum of the fall in pressure head and the change in elevation head (Röthlisberger, 1972, equation 14, p. 184). If the conduit is located at the base of the glacier, the downstream gradient of elevation head is negative where the glacier bed slopes downwards, representing an additional release of energy to melt conduit walls. However, if the downglacier bed slope is upwards - an 'adverse' slope, as found at the exit of an overdeepening - work will have to be performed to raise the elevation of the water, so that the change (i.e. gain) in elevation head acts against the fall in pressure head. If the adverse slope is sufficiently steep, the energy loss term becomes positive: i.e. the water gains heat energy, and 'positive' melt - i.e. freezing - occurs. See Box 5.2b.

Röthlisberger (1972) anticipates this situation, which is discussed in greater depth by Shreve (1985) and Röthlisberger and Lang (1987, pp. 244-245). As ice surface slope and bed surface slope converge above the exit of the overdeepening the glacier thins rapidly, and the pressure melting point rises. It turns out that if the adverse bed slope is ~1.3-2.0 times the ice surface slope the energy released by the pressure head drop, after accounting for that part required to drive the water upslope, is only just sufficient to warm the water to the rising pressure melting point.[2] No surplus energy is available to melt the walls of the conduit.[3] If the adverse bed slope is steeper than this critical factor, the available melt energy becomes negative, and the direction of heat flux is reversed. The shortfall in the energy required to keep the water at the pressure melting point is provided by the release of latent heat of fusion as water within the conduit freezes. Röthlisberger (1972) and Shreve (1985) envisage that this freezing takes effect as ice accretes on the conduit walls. This reduces the cross-sectional area of the conduit, and enhances the tendency for water pressures to rise (see Chapter 1.1 on conduit flow and water pressures). Field observations at Matanuska Glacier, Alaska, which has a suitably deep overdeepening beneath its terminus, led Lawson and his co-workers to suggest that the growth of frazil ice within supercooled water can also be important. Channels can close down - and water pressures rise - because of the combined effect of ice growth on the conduit

[2] The uncertainty in the factor 1.3-2.0 reflects the difference between water which is saturated with air, and water which is air-free. A unit fall in water pressure produces a rise in the pressure melting point of air-saturated water which is one-third greater than the rise in the pressure melting point of pure water, with the result that the critical condition which defines the melting-to-freezing transition is reached at gentler values of adverse bed slope relative to surface slope.

[3] This ignores potential inputs of heat from water with its temperature above the pressure melting point, from ice-flow friction, or from geothermal sources.

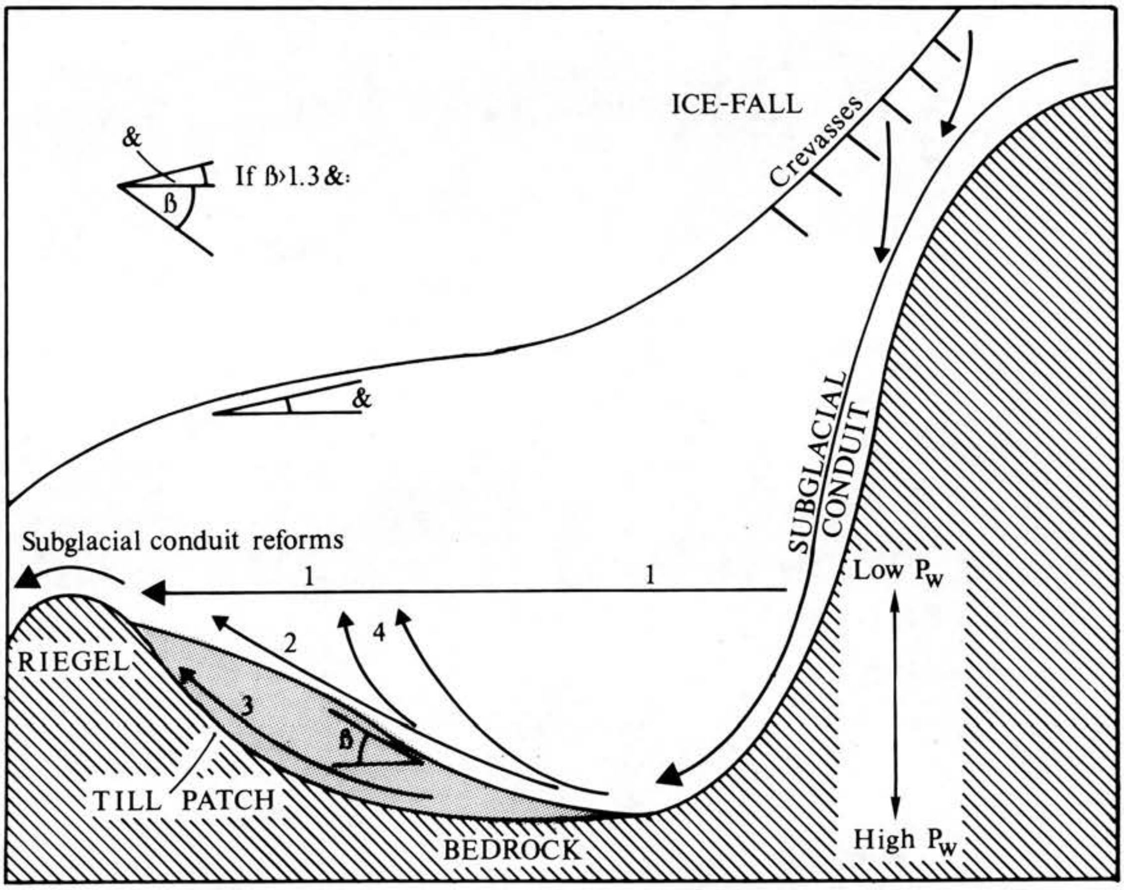


Figure 5.5

Postulated styles of drainage through an overdeepening.
KEY:

1) Lliboutry gradient conduit.
2) Röthlisberger distributed drainage at ice-bedrock (ice-sediment?) interface.
3) Walder-Fowler canals cut into deforming till.
4) Hooke-Pohjola englacial channels.

walls and clogging-up by frazil ice crystals and sediments deposited within the flow (Lawson *et al.*, 1993). See Chapter 6.2.

MODELS OF WATER FLOW IN OVERDEEPENINGS (Fig.. 5.5)

1. Lliboutry's gradient conduit

Lliboutry (1983) was the first to put forward a specific hypothesis to explain water flow through an overdeepening. Water within sizeable conduits which slope downhill is expected to flow at low pressures; often at atmospheric pressure (see also Hooke, 1984; Chapter 1.1). Overdeepenings represent a special case in which conduit water pressures are likely to rise. This Lliboutry attributes to a) the gentle hydraulic gradient - and so low rates of wall melt - present across overdeepenings, and, b) enhanced rates of tunnel closure - perhaps by a factor of four - experienced at the glacier bed. (Recent work adds the impact of rising pressure melting point, as in the previous section.) This central subglacial channel represents the high pressure 'choice', which Lliboutry contrasts with the (theoretically) plausible alternative (after Röthlisberger, 1972) of a high-level, near-ice-marginal, gradient conduit, which runs at the level of the *riegel*. Creep closure of gradient conduits is expected to be slow because of a) the reduced ice overburden pressure acting on high-level channels, and, b) the absence of the enhanced plastic deformation characteristic of the near-bed zone. As a result, water pressures here will be low, tending to atmospheric (i.e. as expected of the ideal gradient conduit; water in a gradient conduit need not back up to force itself over the *riegel*). Water flows from high pressure to low pressure, so if the two exist side-by-side the gradient conduit will capture water from its subglacial neighbour. Thus the gradient conduit tied to the level of the *riegel* represents the stable (i.e. path of least resistance) solution to water flow through an overdeepening. The exact location of the conduit is expected to shift as discharge fluctuates. Water which previously runs subglacially leaves the bed to take up this gradient route, so creating the potential for water-worked debris of basal origin to be carried into englacial ice.

The Lliboutry-type gradient conduit carries similar status to its Röthlisberger-type parent: a theoretical possibility which lacks conclusive field verification. Nevertheless, detailed hydrological studies of three glaciers with known overdeepenings - Glacier d'Argentière, France, Storglaciären, Sweden, and South Cascade Glacier, USA - show some evidence in support of marginal drainage by means of such a gradient conduit. Borehole investigations at the Glacier d'Argentière show a steep cross-glacier hydraulic gradient which drives water towards a distinct zone of near-atmospheric water pressures, perched adjacent to the glacier's right bank, 25 m higher than the low point of the *riegel* (Hantz and Lliboutry, 1983). Similar axes of discrete, low pressure flow, displaced towards the edge of the ice and running at high levels which roughly match the elevations of the *riegels* are inferred at Storglaciären (Hooke and Pohjola, 1994) and South Cascade Glacier (Fountain, 1994). However, if it does exist, the Lliboutry gradient conduit need not exclude alternative styles of drainage. Hooke and Pohjola argue that lateral drainage at Storglaciären represents just part of the total meltwater flux through the overdeepening (see below), whereas much of the water which runs through the 'gradient' conduit at South Cascade Glacier appears never to have been at the glacier bed (Hodge, 1976; Fountain, 1992, 1993; Fountain and Walder, 1993).

2. Röthlisberger's distributed drainage

Whereas Lliboutry suggests a switch within overdeepenings from discrete subglacial to discrete englacial/marginal drainage, Röthlisberger favours a switch to *distributed* subglacial drainage, implicitly as water flows between ice and bedrock (Röthlisberger and Lang, 1987, pp. 244-245 and 275-276). His 'valve effect' (Röthlisberger, 1972, pp. 188-192), caused by pressure-dependence of the melting point, makes it far more difficult for water to flow from low to high elevations within the ice than *vice-versa*. Water which reaches the bed tends to stay there (see 5.5, above). As conduit water pressures rise to meet ice overburden, hydraulic gradients in the vicinity equalise, or reverse, and water is forced out of the conduit to flow over the glacier bed within some kind of distributed system: sheet flow, or possibly linked cavities. When this happens, the ability of subglacial water flow to carry sediment plummets.

3. Walder-Fowler canals

These represent the soft-bed alternative to Röthlisberger's hard-bed distributed drainage; till is believed to accumulate within overdeepenings because the competence/capacity of water flow (i.e. flushing of sediment) falls (Hooke, 1991). Sheet flow, whether between ice and bedrock, or between ice and till exists as a transient phenomenon; in the presence of fine-grained sediments, relatively low hydraulic gradients, and high water pressures, the stable solution for basal drainage is an arborescent system of wide, shallow channels (canals) cut into deforming till (Fountain and Walder, 1993; Walder and Fowler, 1994). Sluggish water flow under high pressures within some kind of distributed drainage system which runs through and/or over subglacial till is inferred at South Cascade Glacier (Hodge, 1979; Fountain, 1992, 1993, 1994).

4. Hooke and Pohjola's englacial conduits

This style of drainage was identified by detailed studies of Storglaciären using tracer, borehole and video techniques (Hooke *et al.*, 1988; Pohjola, 1994; Hooke and Pohjola, 1994). Hooke and Pohjola describe what seems to be a dynamic of network numerous small, interconnected englacial conduits, which drain the bulk of the overdeepening at relatively high pressures. Nevertheless, these small englacial conduits enjoy a pressure advantage over larger subglacial conduits. This is believed to reflect different conduit geometry (circular englacial conduits close less easily than lenticular subglacial conduits) and reduced channel roughness. These differences permit the existence of a hydraulic gradient sufficiently steep for water to flow from high pressure areas at the glacier bed up into englacial conduits, without the connections freezing shut as the pressure melting point rises (i.e. the barrier to upwards flow of Röthlisberger's valve effect is overcome). Active upwelling of water and sediment was observed in boreholes below the points at which these intersected englacial voids. This supports the idea that, if subglacial water pressures are suitably high, water (and sediment) can be driven away from the bed into englacial ice.

GÍGJÖKULL: FIELD EVIDENCE OF HIGH WATER PRESSURES AND ENGLACIAL DRAINAGE

The englacial debris bands indicate fossil channels, and so give us some idea of what the drainage of Gígjökull's terminal lobe looks like, and how it behaves. The following observations also provide evidence - independent of the relict conduit debris bands - that high water pressures and/or englacial drainage occur at Gígjökull. These observations are all consistent with some kind of overdeepening drainage scenario, although some fit certain of the above models better than others.

1. No upwelling into the lake was seen. This implies that Gígjökull is not drained at its terminus by a large, central, subglacial conduit, as is typical of many alpine-type glaciers: e.g. Sólheimajökull. Water which enters the lake directly out of ice must flow within dispersed channels.

2. Summers 1994 and 1995. Turbid streams flowed adjacent to both the left (W) and right (E) ice margins. Both streams emerged from beneath the ice several metres in elevation above, and several hundred metres upglacier of, the terminal lake. Flow out of the lake exceeds the total discharge of these two ice-marginal streams: salt/current meter measurements suggest that peak discharge of the lake outflow stream is ~4.0-8.0 $m^3\ s^{-1}$ (unpublished data, cited by Rogers *et al.*, 1994); my crude estimates for the left and right marginal streams are 2.0 $m^3\ s^{-1}$ and 0.5 $m^3\ s^{-1}$ respectively at peak flow. These figures

support the inference that widespread englacial and/or subglacial seepage contributes substantial volumes of water to the lake.

3. Summers 1993 and 1997. The left-bank ice-marginal stream was replaced by an englacial conduit which emerged from the frontal ice cliff several metres above the level of the lake. The lake was fed by a small waterfall. In July 1997 a small delta was seen to be building-up at the foot of this waterfall (Andy Dugmore, personal communication).

4. July 1994. Widespread upwelling of turbid water was observed in cracks on the terminal ice ramp located several metres upglacier of the lake shoreline.

5. July 1994. Weathering-out of ice at the foot of the ice-fall exposed two features indicative of englacial drainage. The first was a large, sub-horizontal tube, circular in cross-section (~1.5 m diameter), identified as a remnant of an englacial conduit; the second was an exfoliating dome of ice (height ~2.0 m), thought originally to have been pushed-up by a pressurised englacial water pocket (cf. Echelmeyer *et al.*, 1991).

6. July 1996. Turbid geysers were seen to shoot several metres into the air on several occasions. These were also located close to the foot of the ice-fall (Andy Kerr, personal communication). If the water came from the glacier bed (as its sediment load implied) these geysers are examples of artesian fountains: i.e. water pressure (if only temporarily) exceeds ice overburden pressure.

EVALUATION: WHICH MODEL BEST FITS THE FIELD EVIDENCE?

It is not impossible that all four of these models (or some hybrid/variation thereof) operate simultaneously to drain the overdeepenings of Gígjökull and Steinholtsjökull. However, neither Röthlisberger-type nor Walder-Fowler-type drainage can give satisfactory accounts of the origin of the englacial debris bands because they do not provide for englacial drainage. This does not mean such styles of drainage are not important at Gígjökull and Steinholtsjökull: it is possible that both contribute to basal ice development (Chapter 6).

Observations 1 (no major channel enters lake) and 2 (elevated ice-marginal streams) provide some support for Lliboutry's gradient conduit hypothesis. Some kind of sediment-laden gradient conduits at Gígjökull and Steinholtsjökull could provide the key factors which lie behind debris band formation. However, observation 3 (subaerial lake inputs < lake outputs) strongly suggests that *if* it exists, gradient conduit flow is not solely responsible for drainage of Gígjökull's overdeepening. Some water will reach the lake by way of the ice-bed interface (i.e. the Röthlisberger and/or the Walder-Fowler models), but it is likely that some water, and its sediment, will take up an englacial route under the influence of high basal pressures, as observations 4 (upwelling of turbid waters at snout), 5 (englacial conduit/ice blister) and 6 (turbid geysers) demonstrate. These observations tie in with the basic idea of the Hooke-Pohjola model. Additional features of the drainage at Storglaciären which fit what is observed/inferred at Gígjökull and Steinholtsjökull include:

- **Large number of channels.**

  Storglaciären. Hooke and Pohjola estimate that the cross-section surveyed contained ~4,800 englacial conduits, of which ~600 were active at any time.

  Gígjökull. The limited number of suitable exposures indicates that a large number of debris bands exist within a few metres vertical of the basal ice. Each debris band represents a former channel. The relative age of the debris bands (i.e. the time-space pattern of englacial sedimentation events) is not known, but the visual evidence is consistent with Hooke and Pohjola's picture of a complex, three-dimensional network of dispersed englacial channels, located within the lower layers of the englacial ice (Roger Hooke, personal communication, Reykjavík, 1995).

- **Size and shape of channels.**

  Storglaciären. Video shows englacial conduits to be oval-to-circular in shape, with typical dimensions of 0.1-0.4 m diameter. Other drainage routes exploited sub-horizontal cracks within the ice. Several channels were seen to contain sand and gravel.

  Gígjökull. Some larger pockets of water-worked sediment indicate sizeable water pockets and channels, but the majority of debris bands found were less than 20 cm thick. The extent to which debris band dimensions match the size of the englacial channels in which they formed is not known. If this match is close - which implies that a) sediment must choke the original channel, and, b) post-deposition deformation is limited - then the englacial channels at Gígjökull must be of similar size and shape to those at Storglaciären. (Note: not all channels are likely to give rise to debris bands.)

- **Inferred water flow speeds.**

  Storglaciären. Calculations by Hooke and Pohjola using realistic estimates of discharge, hydraulic gradient, channel geometry and channel roughness suggest that *mean* flow speeds may reach ~0.3 m s^{-1} or so within parts of the englacial conduit network.

Gígjökull. The size, geometry and melt discharge here are of similar magnitude to Storglaciären, so it is reasonable to anticipate similar flow speeds if channel geometries are similar. The evidence of pebbles, which are common in the debris bands, supports this: the Hjulström Curve shows that transport of pebbles requires flow speeds of ~0.3 m s^{-1}, although transport of larger cobbles (i.e. >64 mm diameter) clearly requires larger discharge events/higher flow speeds.[4]

- **Switching of active channels**

  Storglaciären. This is believed to be common, because of the impact channel constrictions exert on water flow (5.7, below).

  Gígjökull. Observations of sporadic surface upwelling possibly indicate and coincide with episodes of channel switching. Channel switching is likely to induce deposition of sediment, so the presence of the debris bands can be read as evidence of widespread drainage reorganisation (5.7, below).

Verdict

It is clear that any argument must be tentative! Direct data to describe subglacial and englacial drainage at Gígjökull and Steinholtsjökull do not exist. Costly, cumbersome techniques of direct investigation restrict such studies to a handful of glaciers; even then there are many uncertainties (see Hooke, 1989, p. 221). Field data have yet to replace theoretical structures as a means of study, in part because - as here - field data frequently cannot be used with confidence to choose between different drainage models. Several alternatives are plausible. These have important characteristics in common; indeed, different styles of drainage are not necessarily mutually exclusive. It seems reasonably certain that drainage through overdeepenings represents a special case in which rising water pressures render subglacial conduits unstable. This will change the pattern of water and sediment transport, and so is likely to affect moraine formation. Any or all of the drainage models discussed above may operate at Gígjökull and Steinholtsjökull; however, only the Lliboutry and Hooke-Pohjola models allow for water to carry sediment into the main body of the glacier, as debris band formation requires. Field data (or inferences from these) give some support for both the Lliboutry and the Hooke-

[4] I admit that this is a simplistic argument, but there is no better alternative. The Hjulström Curve is crude: e.g. strictly it does not apply to mixtures of sediment; it ignores the possibility of pipe flow/sliding beds; it ignores clast weight acting to push it against water flow when transport is uphill, etc... Nevertheless, the evidence of the debris band sediments does indicate that flow speeds in the englacial channels are probably of the same order of magnitude as Hooke and Pohjola infer for Storglaciären.

Pohjola models; the latter perhaps offers superior explanatory power as far as debris band formation goes (see also 5.7, below), but this is not proof that Hooke-Pohjola englacial drainage actually exists at Gígjökull and Steinholtsjökull. It is important to remember that these four models all describe ideal styles of overdeepening drainage. Considerable variation on these themes is likely.

5.7 WHY DOES DEBRIS RETURN TO ICE TRANSPORT?

Deposition in englacial channels

Sediment abandoned within englacial waterways is likely to be incorporated rapidly into ice, thereafter to appear at the ice margins as debris bands. It is often assumed that deposition within ice-walled channels within which water flows under pressure is self-defeating: the sediment blockage creates a local steepening of the hydraulic gradient, so raising the water flow speed and the sediment transport rate. Because of this, major events such as large-scale roof collapse are invoked to explain esker formation (e.g. Bennett and Glasser, 1996, pp. 273-274). However, this seems unnecessary; it is likely that sediment deposition is a widespread feature of a typical valley glacier's drainage network. Deposition represents a complex and continuously shifting balance between flow properties (discharge, flow speed, bed shear stress), local channel geometry and sediment supply relationships. Studies of active *proglacial* streams over the last ten years show that major within-reach episodes of deposition can be expected even under high-flow conditions [see Lane (1995) for a review of this 'new fluvial geomorphology'; Ashworth and Ferguson (1986) and Lane *et al.* (1996) for typical case studies of active channel change in high energy rivers]. Warren and Ashley (1994) observe that the behaviour of high-energy, sediment-rich subglacial drainage is unlikely to differ hugely from that of subaerial rivers, so features such as convergent/divergent flow structures are likely to contribute to esker formation.

The analogue of dynamic alluvial channels leads me to expect widespread reworking of sediments if the basic conditions of abundant sediment supply coupled with spatially and temporally variable high-energy flows are met with sufficient frequency. Alluvial channels are self-forming by differential sediment transport (erosion predominates in some parts, deposition in others). Within ice-walled channels, further possibilities for sedimentation are introduced by channel migration induced by differential wall melting, or channel switching in response to changes in the hydraulic potential field. Channel switching within complex braided networks is likely to be enhanced by the three-dimensional nature of the glacier drainage network [see T. J. Hughes, personal communication in Seaberg *et al.* (1988, p. 225), plus other Storglaciären references as given above]. Tracer studies demonstrate that en- and subglacial drainage replicates features of typical subaerial streams conducive to sedimentation; e.g. sinuous and braided channels, pool-riffle structures, side-spill backwaters. Two examples illustrate the possibilities. Röthlisberger (1993) lists meandering channels heavily laden with sediment within gently-sloping terminal lobes as one of three typical styles of subglacial drainage (e.g. at Aletschgletscher, Switzerland). Hooke (1984) associates multiple eskers in Norway with channel migration. He argues that as within-channel bars build up under open flow conditions, the thread of fastest flow is pushed down the bar flanks (topographically-induced acceleration to use the terminology of fluvial geomorphology), boosting the tendency for the conduit to switch position. Ultimately this constructs a complex braided network of channels/eskers. It is possible that the multiple debris bands at Gígjökull and Steinholtsjökull reflect a similar process at a reduced scale.

Deposition induced by drainage reorganisation

Features specific to glacier drainage networks may be important also. Towards the end of the summer/beginning of the autumn the conduits of a typical valley glacier tend to close down as meltwater throughput falls, leaving a residual distributed drainage system (Chapter 1.1). It is common for discharge to fall by two orders of magnitude [e.g. Bondhusbreen, Norway: Hooke *et al.* (1985)], which, in conjunction with the much reduced efficiency of distributed drainage, will produce a major fall in the sediment transport competence and capacity of the drainage network. This is likely to strand sediment within the ice, to be incorporated as the process of channel closure progresses. Thus it is possible that the debris bands represent a fossilised 'snapshot' of the end-of-summer state of the drainage system, preserving the pockets of sediment storage surviving the last major episode of englacial channel change. However, it is worth noting that winter discharge of the Jökulsá á Sólheimasandi falls to just one-third of its summer levels (Lawler, 1991). Delayed release of stored water, groundwater inputs, and an elevated geothermal heat flux within the crater (Lawler *et al.*, 1996) are likely to sustain this relatively high level of winter flow. No winter flow data for Gígjökull or Steinholtsjökull are available, but my guess is that flow falls by a factor of ~10. Eyjafjöll remains active - enhanced seismic activity led to the issue of possible eruption warnings in June 1994 - but less so than Katla, which implies a geothermal heat flux of lesser magnitude. An aerial photograph of Gígjökull, taken by Ragnar Sigurðsson (Guðmundsson and Sigurðsson, 1995) shows no ice-marginal drainage at Gígjökull, but the Steinholtsá is still flowing. The outflow of Gígjökull's lake cannot be seen clearly on this picture. The example of Sólheimajökull, in contrast to typical Alpine or Norwegian examples, implies that autumn closure of channels, so trapping sediment, is perhaps not as important a factor in Iceland as it might be elsewhere.

Shifting channels during the summer represents an alternative scenario favourable to sediment deposition and its entrainment by ice. Both the Lliboutry and the Hooke-Pohjola models of overdeepening drainage make provision for change in the location of the active channel. Lliboutry argues that the level of the *riegel* defines the likely elevation of the gradient conduit only approximately; its exact position will change as the hydraulic grade line fluctuates with changing discharge. This creates a picture of an ice-marginal channel which intermittently switches its line within a zone of ice perhaps several tens of metres thick. This in turn defines an ice-marginal zone within which sediment is likely to be abandoned, and returned to ice transport.

Hooke and Pohjola's scenario is equally, if not more, attractive. They stress the dynamic nature of the drainage network at Storglaciären, with just one-sixth of englacial conduits active at any time. Switching between active conduits appears to be controlled by constrictions which represent the vulnerable parts of the drainage network: the englacial conduits are thought to resemble 'strings of sausages' (my term!) rather than continuous tubes, with each sausage (i.e. the interval between constrictions) estimated to be 70 m in length. Frequent disruption of drainage as constrictions close up is supported by delayed return and exaggerated dispersion of dye. If pockets of dye can be trapped, so too can pockets of sediment. Switching of water between conduits possibly reflects what was once an active conduit becoming blocked with sediments. Hooke and Pohjola point out that the sediment-poor drainage of Nordjokk, the outlet stream fed by these conduits, is strongly suggestive of englacial drainage. This is true, but as much of the water within these feeder streams has probably run at the glacier bed immediately beneath and downglacier of Storglaciären's ice-fall, it can be argued that its lack of sediment load reflects the fact that much of it has been lost in transit. Given that 1) channels are likely to divide, so lowering their sediment transport capability, as part of the subglacial-to-englacial drainage transition, and, 2) the match inferred between the pattern of the Gígjökull/Steinholtsjökull debris bands and the ideal Hooke-Pohjola model, this offers a plausible mechanism for debris return to the ice of potentially high explanatory value.

CHAPTER 5: SUMMARY

Large quantities of sediment presently accumulating at Gígjökull and Steinholtsjökull show clear signs of water transport. This sediment is derived from distinct englacial debris bands, so these ridges do not fall into any of the usual categories of fluvial-glacial landforms; nor do these debris bands match with conventional explanations of debris in high-level transport. Here I have explored what seems to be the most satisfactory mechanism - namely that the debris bands represent the fossil trace of former englacial conduits - by building on previous observations, and piecing together what seems to be the relevant hydrological theory. The presence of a terminal overdeepening at both Gígjökull and Steinholtsökull seems to provide the key structural control. Current thinking on the behaviour of meltwater within overdeepened basins is consistent with this relict conduit hypothesis, although it is impossible to be specific as to exactly what happens to construct the debris bands.

Pronounced moraines of the type described here can be expected if:

1. Large quantities of debris are carried within the *englacial* part of a glacier's drainage network.
2. Large quantities of this debris tend to be abandoned within the ice, so forming the relict conduit debris bands.
3. Ice-flow trajectories and ablation ensure that this debris congregates at the glacier surface.

Several possible scenarios exist in which these criteria are likely to be met. In the case of Gígjökull and Steinholtsjökull the following seems to fit best. Numerous conduits start in the ice-fall, and, guided initially by crevasses, quickly find their way through relatively thin ice to the glacier bed. Fast sliding with cavitation over weak bedrock (see Chapter 6) generates abundant quantities of subglacial debris, much of which enters these subglacial conduits. Water pressures rise as these channels try to exit the overdeepening developed at the foot of the ice-fall, forcing drainage to take up an 'easier' englacial route, carrying at least part of its debris load with it. Disruption of drainage and active channel switching within the englacial drainage network causes pockets of sediment to be abandoned within the ice. Plastic ice flow closes down these abandoned channels, trapping the sediment as debris bands. These are carried rapidly to the ice margin under the influence of the strongly compressive terminal flow regime, there to build up the overlapping moraine ridges.

IMPLICATIONS

- This part of the study suggests that traditional frameworks of analysis are not fully sufficient. The traditional distinctions between 1) 'active' and 'passive' transport, and 2) 'glacial' and 'fluvio-glacial' processes and landforms both fail here (which is not to say that they do not work perfectly well in other situations). It seems necessary to widen conventional sediment transport pathway theory to facilitate a more flexible framework which allows for the full complexity of possible transport relationships (Kirkbride, 1995a; Kirkbride and Spedding, 1996). This must take full account of the underlying process mechanics, including, as illustrated by this case study, the wider impact of catchment-scale ice-water interactions.

Following on from this:

- Unduly narrow explanatory frameworks potentially can lead to incorrect interpretations of both past and present ice-marginal facies. To take a hypothetical example: material erroneously identified as outwash which was in fact dumped directly out of ice. To give a concrete example: Krüger (cited by Näslund and Hassinen, 1996) believes that thrust action is responsible for extensive accumulations of moraine at Hofðabrekkujökull, whereas Näslund and Hassinen follow similar reasoning to that of Martin Kirkbride and myself to favour an origin related to englacial drainage (Jens-Ove Näslund, personal communication, Reykjavík, 1995; Näslund and Hassinen, 1996).

- Work is needed to elucidate the impacts of glacier hydrology on glacial geomorphology. The bulk of hydrological studies are undertaken to study drainage for its own sake, or perhaps as part of ice flow studies; possible wider implications are usually restricted to passing remarks. Exceptions which convincingly link hydrology and geomorphology include Walder and Hallet (1979), Shreve (1985), Sharp *et al.* (1989a), Sugden *et al.* (1991), Syverson *et al.* (1994) and Clark and Walder (1994). These papers all use direct evidence of past drainage structures: Nye-channels, eskers or canals. This study starts to explore wider possibilities using indirect evidence of past drainage structures. I suggest that this style of moraine development is characteristic of (but not exclusively related to) englacial drainage through an overdeepening. This raises the enticing possibility of using the geomorphological record to make intelligent guesses as to glacier palaeo-hydrology, even if no direct evidence of a channel exists.

- Certain styles of drainage seem to give rise to certain wider patterns of ice-marginal sedimentation. Sólheimajökull has few debris bands, and little basal ice, as is the case also with Tungnakvíslajökull, whereas Gígjökull, Steinholtsjökull, Hofðabrekkujökull and Kvíárjökull - all of which are believed to exhibit high pressure englacial drainage - all have large quantities of debris band deposits and extensive basal ice sequences (my observations, summers 1995 and 1996). This possible link forms the subject of the next two chapters.

CHAPTER 6
Gígjökull and Steinholtsjökull: Basal ice

INTRODUCTION

Exposures of basal ice found at the margins of Gígjökull and Steinholtsjökull are unusually thick for temperate valley glaciers; the volume of basal ice here is reminiscent of sub-polar (i.e. mixed thermal regime) or surging glaciers. This chapter tries to makes sense of this puzzle by matching field evidence with theoretical and conceptual inferences. Existing work on basal ice tends to concentrate on the precise mechanisms by which it forms: a body of work which sits firmly within the reductionist traditions of glaciology (Chapter 1.4). Less tends to be said about the wider relationships which govern debris supply at point of origin, and subsequent transport and preservation of basal ice once it has (re)formed. These wider relationships are crucial to a full understanding of why it is that large quantities of debris-rich basal ice appear at the margins of Gígjökull and Steinholtsjökull. My tactic here is to review both the mechanisms by which basal ice forms (6.2), and the wider perspectives which apply (6.3), as a prelude to identification of the type of basal ice found at Gígjökull and Steinholtsjökull. Two plausible models of basal ice development fit the field evidence (6.4), both of which stem directly from what is known, or can be inferred with confidence, of the catchment-scale behaviour of ice and water at the two study sites.

BASAL ICE: DEFINITION

It is usual to identify two (supposedly) distinct types of ice: 1) *englacial* or 'meteoric' ice, which makes up the major part of the ice body, and is derived from snow by firnification processes which operate at the ice surface, and within the main body of the ice; and, 2) *basal* ice which forms in contact with the glacier bed when pre-existing ice is modified (involving partial/complete recrystallisation and/or partial/complete melting and refreezing), or when water at the bed (derived from whatever sources) refreezes. Basal ice acquires characteristics representative of its interaction with the glacier bed which render it distinct from englacial ice. Typically these characteristics include: a higher debris content; different crystal properties; a distinct, and variable, chemical and isotopic signature; and a highly non-uniform structure, comprising a number of discrete layers and lenses of variable extent.

6.1 FIELD OBSERVATIONS

Gígjökull

Basal ice exposures appear as a steep cliff face of debris-rich ice (Figs 4.3, 4.9 and 6.1). This face is continuous for over 500 m at the eastern margin, and exposures reach heights in excess of 10 m. To the south, basal ice appears beneath angular rockfall debris; to the north (i.e. towards the snout) the basal ice is progressively buried beneath increasing quantities of material derived from the relict conduit debris bands discussed in the previous chapter. Exposures of basal ice at the western margin are scarce, which, in part, may reflect burial beneath large quantities of relict conduit and rock-fall debris. However, it is possible that development of basal ice is highly asymmetrical, so that basal ice is not created/does not survive in such large quantities close to the western margin of the glacier.

Steinholtsjökull

Difficult access prevented a full survey of the northern margin, but at two locations where access to the ice was possible a debris-rich ice cliff similar to that at Gígjökull was found. My impression was that basal ice exposures like this are largely continuous along the length of the northern margin between the foot of the ice-fall and the terminus (Fig. 4.9). Nothing is known about basal ice development at the southern side of the terminus (i.e. true left bank) because this area is swamped by rock-fall debris, much of it associated with the 1967 cliff collapse.

SEDIMENT ACCUMULATION FED BY BASAL ICE

The flux of dirty ice carries a constant supply of debris to the ice margins. Here it is released by melt, to fall or slide down the ice cliff to accumulate at its base. No attempt was made to measure ablation of the ice cliff, but a relatively high-melt rate is inferred because:

1. The angle at which incoming solar radiation meets the basal ice cliff face is high. This is because of a) the steep angle of the ice cliffs ($>45^\circ$, tending to 90° in many places), and, b) the low elevation of the sun in the sky (summer maximum $\sim 50^\circ$). This means that short-wave radiation inputs are relatively intense for much of the day in summer (e.g. between ~2 a.m. and ~10 a.m. for ice cliffs facing north-east, as at Gígjökull).

2. The veneer of dark debris plastered against the face reduces the albedo of the surface relative to clean ice, and so enhances absorption of incoming radiation. The steep angle of the ice cliff face prevents debris building up to thicknesses sufficient to retard melt.

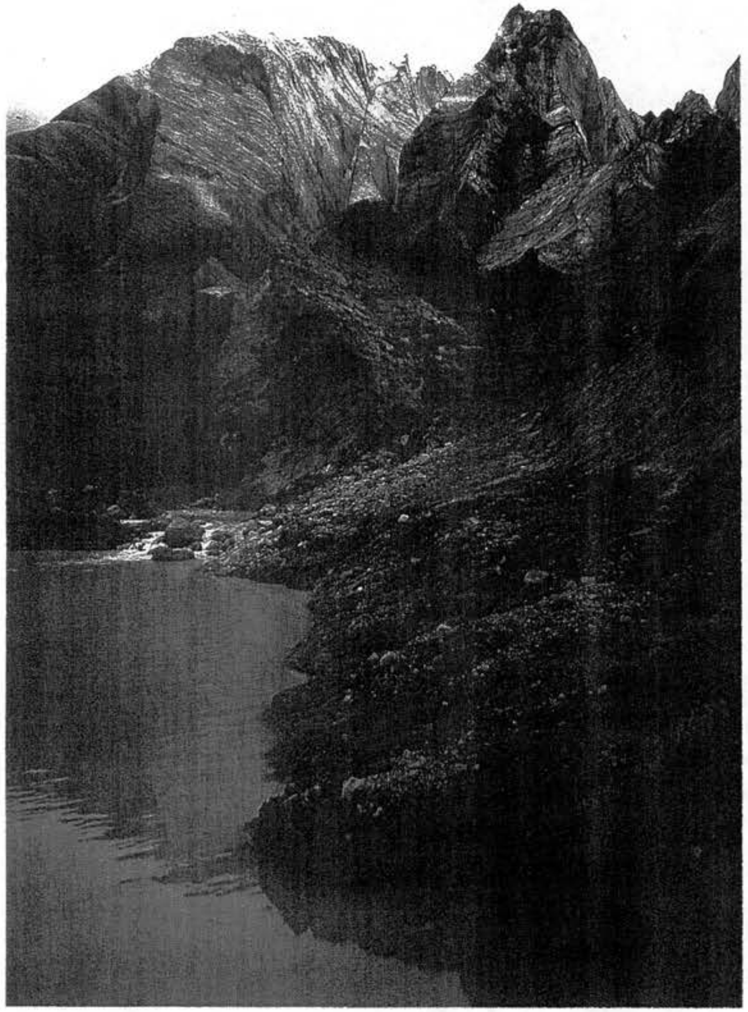


Figure 6.1

Basal ice cliffs at Gígjökull, site G1. Note the sharp contact with clean (white) englacial ice.

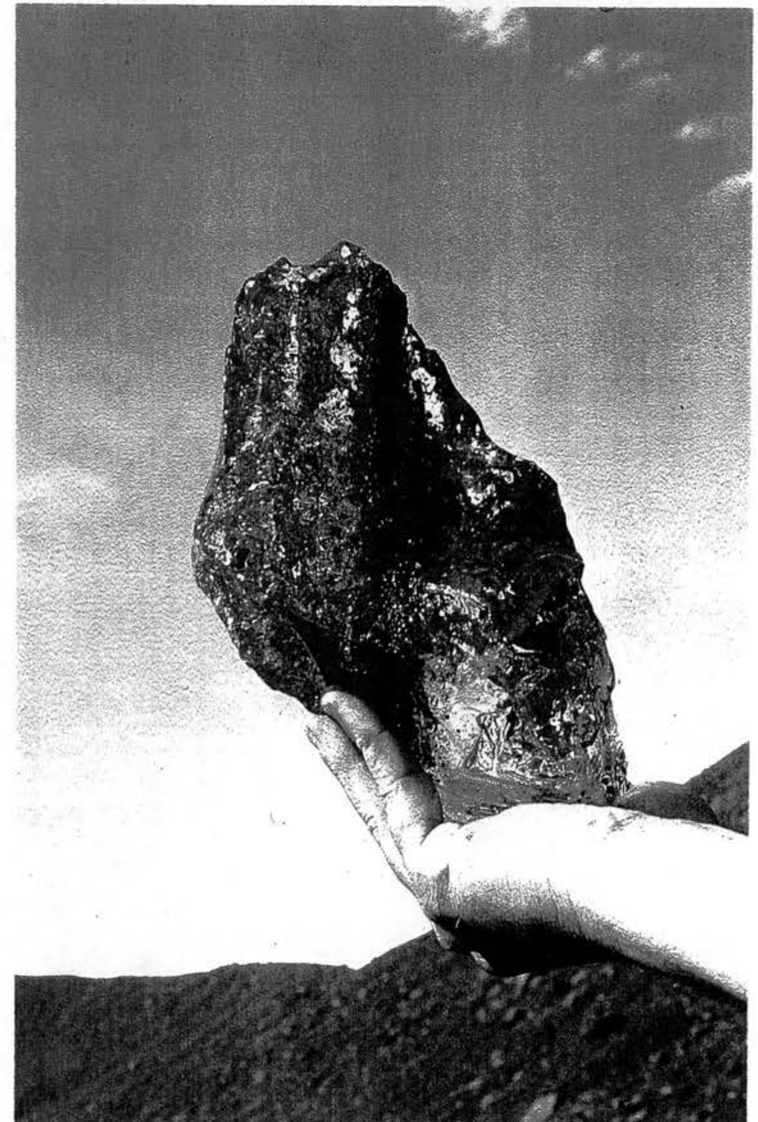


Figure 6.2

Typical block of basal ice. Dirt-rich laminae alternate with bands and pockets of clean, bubble-free ice.

3. Heat trapped between the ice cliff face and its enclosing moraine wall (Gígjökull) or bedrock wall (Steinholtsjökull) will enhance melt by means of a) reflection of incoming short-wave radiation, b) absorption of short-wave and emission of extra long wave radiation, and, c) elevation of the sensible heat flux (e.g. Chinn, 1987).

At Steinholtsjökull, debris released from the basal ice chokes the narrow gap between the northern ice margin and the bedrock escarpment which confines it. Beneath the more extensive (eastern) exposures of basal ice at Gígjökull debris accumulates as a moraine pediment progressively built up by parallel retreat of the ice cliff face (Fig. 6.1). This pediment is underlain by a sloping platform of ice protected from melting by the insulation of the debris cover built up above it. Sediment transfer across this pediment occurs by way of individual stone fall/slide events, and by bulk redistribution of sediment as 'flow till', which in places imparts a crude layered structure to the debris. The pediment is drained at its base by a shallow (~10 cm deep) marginal meltwater stream. Although this was turbid, it appeared to entrain little in the way of bed-load, suggesting that, under usual stream flow conditions, larger clasts are not removed from the immediate vicinity of the ice margin. Elsewhere, beneath the less extensive exposures, debris from basal ice has accumulated in discrete, sharp-crested moraine ridges, typically 0.5-1.0 m in height. As of summer 1993, up to three parallel ridges were observed together. I take these to represent a recessional sequence of annual moraines, preserved intact because debris accumulation and reworking is limited by the smaller contributing area for debris and meltwater.

Character of the debris

Coarse clasts are plentiful. The vast bulk are angular or sub-angular (mean roundness score of 396 clasts = 2.60, s.d. ± 0.54). This clearly differentiates the basal debris from the relict conduit debris (see Table 4.3 and Box 5.1). Only 5 from 396 clasts - 1.3% - counted were classified as sub-rounded (4) or rounded (1), whereas 52% of clasts from samples of relict conduit debris fell into one of the three rounded categories (sub-rounded = 40%). Particle size analysis of the debris matrix was not carried out, but the poorly-drained nature of the pediment, plus widespread incidence of thin, clast-free muddy swathes (sheet structures produced by de-watering associated with flow till activity?) indicated a relatively high proportion of fines.

Character of the ice

The dirty basal ice lies beneath clean englacial ice, with a sharp, discordant contact between the two (Fig. 6.1). The basal ice seems to be made up of a single facies, or perhaps a mixture of two ice types, but problems with access to steep ice cliffs, the dangers of frequent stone-fall, and the dirty nature of the ice surface made a comprehensive survey of the exposures near-impossible. Inspection of clean sections and samples removed with an ice hammer and washed shows that debris within the ice exists largely as a series of more-or-less continuous silty-sandy-gritty laminae, which alternate with layers of clear ice (Fig. 6.2). Typically these laminae are 1-2 mm thick, and are occasionally cut by the inclusion of larger clasts. This ice is bubble-poor; where present, bubbles formed distinct layers which parallel the debris laminae. The density of the debris laminae varies widely: frequently several cm of tightly-packed alternate debris-rich and clear layers are broken by layers of clear, but bubble-poor, ice several cm thick. Pockets of this clear ice include clouds or clots of debris, with muddy inclusions which look like small, elongate (<2 mm) pockets of muddy water trapped within the ice. My field notes make a tentative distinction between debris-layered and debris-dispersed ice, although no sharp distinction between the two types is evident. From field study it was not clear whether this was a real distinction between two different types of basal ice which occur in juxtaposition (perhaps due to tectonic deformation of the ice), or just the outcome of a single process of basal ice formation acting with different intensities. Single blocks of ice sample (<10 cm diameter) often show evidence of both layered and dispersed characteristics. However, classification of basal ice types by debris content alone is dangerous because it is the process of re-freezing or metamorphism which defines basal ice type; the inclusion of debris in whatever quantity is usually a secondary characteristic contingent upon the availability of debris.

Structures. It was far from clear whether or not the basal ice exposures display evidence of substantial tectonic deformation. The ice cliff appears as a dirty, stratified sequence of sub-horizontal, gently-warped bands with no distinct evidence of folding or faulting: the outstanding feature of the ice is its dirt-covering, which, in conjunction with access difficulties, makes close study highly problematic. Nevertheless, it is plausible, even likely, that the basal ice is highly deformed: both the sheer thickness of the exposures and the highly compressive flow inferred for Gígjökull (Chapter 4.1; Table 4.1) favour this. Clear fold and fault structures are best preserved at medium intensities of tectonic deformation; very intense shear can deform and draw-out tectonic features to such an extent that the end result appears as a series of sub-horizontal discontinuities which (particularly when smothered with debris) may not be obviously distinct from the original sub-horizontally-stratified state of the ice [see below: Variegated Glacier case study (6.3)].

Sample collection and processing

Samples of basal ice were taken, and analysed for electrical conductivity of meltwater, debris concentration by mass, and debris distribution. See Chapter 4.2 for details. For convenience, I reproduce Figs 4.10b and 4.10c here.

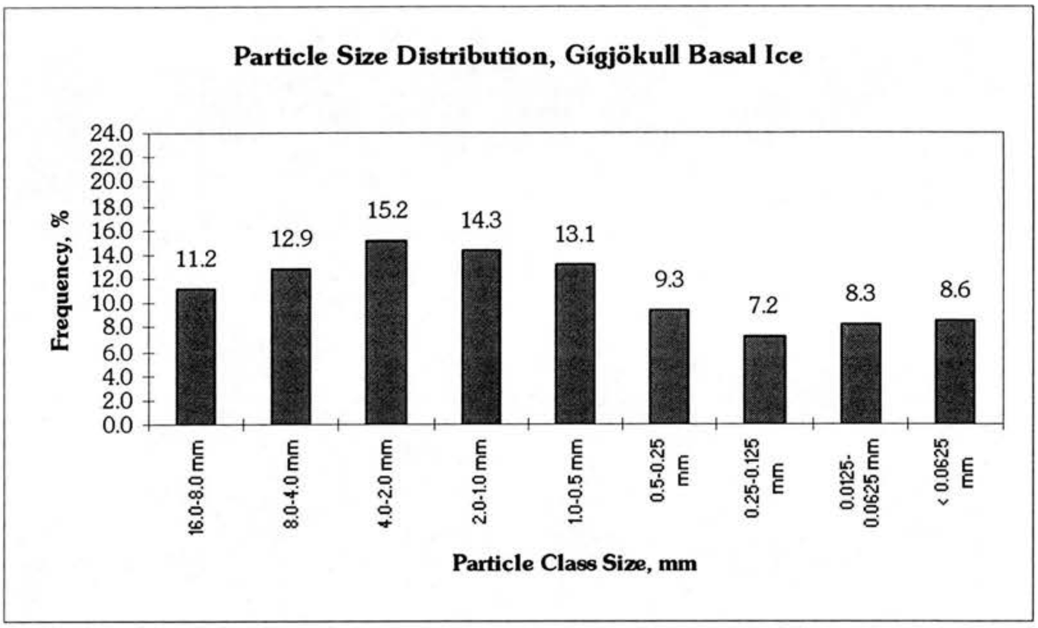


Figure 6.3a

Particle size distribution, Gígjökull basal ice debris.

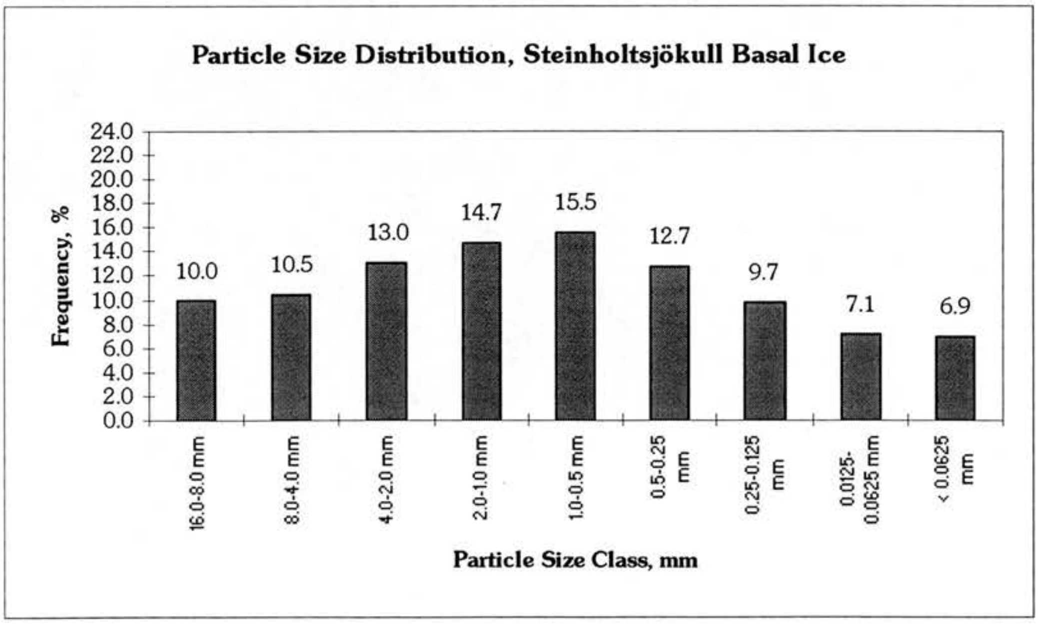


Figure 6.3b

Particle size distribution, Steinholtsjökull basal ice debris.

INTERPRETATION

The detailed appearance of the ice, and study of its debris content confirm the initial 'working' label (attached purely on the basis of its dirty appearance) of 'basal ice'. The sub-angular nature of the clasts reflects the edge-rounding produced by the clast-clast and clast-bed contracts characteristic of the basal traction zone. The poor level of sorting is also diagnostic of a subglacial debris population (Boulton, 1978). The Gígjökull (Fig. 6.3a) sample strongly suggests the stretched multi-modal distribution typical of basal ice debris. Haldorsen (1981) considers that this spread of particle size represents the sum of individual uni-modal populations, each of which relates to a specific wear process of the subglacial environment: pebbles and gravel represent the fracture of intact bedrock to give lithic fragments; sand-sized fragments represent the breakdown of bedrock/fragments into its constituent mineral grains by crushing failure; and silts and clays represent the product of the abrasive wear breakdown of individual mineral grains. Relative to the Gígjökull sample, debris from the Steinholtsjökull basal ice (Fig. 6.3b) seems to be depleted slightly in the coarser gravel fractions. This may reflect less efficient bedrock fracture beneath the smaller, thinner glacier, which, on the basis of geometry, is inferred to be less active than its neighbour,[1] although it is perfectly possible that this is an artefact of the smaller sample size. Both samples contain a high proportion of fine sand and silt-sized particles relative to the water-worked debris of the relict conduit debris bands (Fig. 4.9a), suggestive of restricted efficiency of flushing across substantial areas of the two glaciers' beds.

DISCUSSION

It is evident that large quantities of debris currently accumulating at the margins of Gígjökull and Steinholtsjökull are derived from what I identify with confidence as basal ice. The key question is: why does so much debris-rich basal ice reach the ice margins? This is an *upglacier* problem which exists at the catchment scale, the answer(s) to which rests not with any single mechanism, nor any simple pattern, but with a number of inter-related sediment transfer processes.

Whereas the incidence of basal ice exposures and associated moraines was found to be extremely sparse at Sólheimajökull, basal ice is extremely well-developed along much of the margins of Steinholtsjökull, and at Gígjökull especially. The size of these exposures exceeds by some way the thickness of the basal ice layer commonly reported for temperate glaciers (usually

[1] Sharp *et al.* (1994) find a similar coarse fraction deficiency in basal ice produced during the quiescent phase of Variegated Glacier relative to its surge phase.

<1.0 m) and gives rise to a correspondingly rapid rate of ice-marginal sediment accumulation. The puzzle is to account for this unusual thickness of basal ice at the glacier margins: i.e.

1. By what means did the debris-rich ice originate and subsequently arrive at the ice margin?

2. Why did the debris take - and remain within - that particular 'choice' of transport pathway, instead of taking an alternative transport pathway with different geomorphic consequences?

My answer to the first question takes up the rest of this chapter; my answer to the second question starts with this chapter, but spills over into Chapter 7, which examines wider sediment transport relationships in more detail. Ultimately I wish to argue that the intensity and stability of the basal ice pathway inferred at Gígjökull and Steinholtsjökull relates to the interaction of bedrock topography, meltwater drainage within and beneath the ice, and ice flow dynamics. This three-way interaction suggests a relationship between the presence of the relict conduit debris bands discussed in Chapter 5 and the quantity of basal debris observed at the ice margins (see Chapter 7).

To identify the central problem as a problem of basal ice formation and preservation is perhaps to muddle rather than to clarify the issue. Several different types of basal ice exist, within widely different contexts, not all of which are equally consistent with the examples of Gígjökull and Steinholtsjökull. Indeed, much work on basal ice studies basal ice for its own sake, and fails to explore the crucial wider relationships. The discussion which follows is my attempt to make sense of our current level of understanding, and to produce a coherent synthesis of those aspects of the 'basal ice problem' most relevant to moraine-forming activity at Gígjökull and Steinholtsjökull, and temperate alpine glaciers generally.

6.2 FORMATION OF BASAL ICE

The majority of work has viewed the basal ice problem as a problem of the mechanism by which basal ice forms, and - to a lesser extent - deforms. This is indisputably important, but it tends to place basal ice in an abstract context which obscures the relationship between the subglacial environment, sediment transport processes and pathways, and ice-marginal sediment accumulation. What is required - which is what I try to construct here - is a framework of study which treats basal ice as an integrated problem with major ramifications for glacier sediment budgets and ice-marginal geomorphology. This breaks down the problem into three stages:

1. **Availability of debris at the ice-bedrock interface.** Although basal ice is strictly ice which has undergone substantial alteration at the glacier bed, basal ice is more commonly defined as ice produced subglacially which carries large quantities of debris. Basal ice must be dirty if it is to build up large ice-marginal moraines. This is not a trivial point: clean basal ice can, and does, occur. Lliboutry (1994) states that dirty basal ice is not found at the three largest glaciers in the French Alps : the Mer de Glace and the Glacier d'Argentière (Chamonix) and the Glacier Blanc (Haut Dauphiné). Studies by students of Aberdeen University (June 1996) found four different types of basal ice at the snout of the Ghiacciaio di Pré de Bar, Courmayeur, Italian Alps (which is adjacent to the Glacier d'Argentière), but only the lowermost of these, corresponding to Hubbard and Sharp's (1995) 'solid' facies, contained anything other than negligible quantities of debris (see below).

   Availability of debris prior to entrainment is determined by two broad factors: a) production of debris at the ice-rock interface [see Iverson (1995) for a review of subglacial erosion processes], and, b) retention of debris at the ice-rock interface. Lliboutry implies that, in the case of the French glaciers, it is the weakness of subglacial erosion which accounts for the debris-free character of the basal ice, but the possibility that debris is removed before it is entrained is also important. Debris which becomes part of a deforming bed, or is flushed away by meltwater (Chapter 2; see below) cannot be incorporated into basal ice.

2. **Initial entrainment of debris.** This can happen in two broad ways: a) by the freezing of water to create or re-form ice, or, b) by traction. I consider these processes in greater detail below.

3a. **Redistribution of debris by ice flow/tectonic processes.** Because it forms under high stress conditions, basal ice is frequently subject to substantial deformation. Indeed, it is thought that the presence of large quantities of debris can weaken basal ice relative to clean ice (Hubbard and Sharp, 1989, pp. 550-552). Deformation of basal ice has been described at scales which range from that of the single cavity-bedrock obstacle (e.g. Rea and Whalley, 1994) to the wider flow field of substantial parts of the Greenland Ice Sheet (Knight *et al.*, 1994). Whatever the scale of deformation, basal ice can either be thickened or thinned; ice can also move relative to the debris it contains. Processes such as these can change the appearance of the basal ice layer, with important effects on the flux of debris-rich ice to the glacier margins

3b. ***Preservation of basal ice.*** Basal ice must reach the ice margins if it is to form moraines of the type described at Gígjökull and Steinholtsjökull. Reductionist studies of basal ice mechanics usually exclude this crucial factor. By its very nature, basal ice is highly vulnerable. The tendency for basal ice to be destroyed can be enhanced or diminished by the pattern of deformation and/or the wider flow field. The extent to which basal ice is in contact with meltwater draining at the bed is also likely to determine its preservation potential (Hubbard and Sharp, 1993, 1995), but this factor has received very little attention.

MECHANISMS OF BASAL ICE FORMATION (Table 6.1)

Studies reveal several different mechanisms by which basal ice can form. These different mechanisms operate at a wide range of scales, are not all equally effective at producing thick sequences of basal ice, and are not necessarily mutually compatible. Different styles of basal ice reflect different subglacial contexts (Hubbard and Sharp, 1995), and it is this subglacial context which is the key to sediment transfer processes and moraine formation. Here, as a first step towards an explanation of the thick marginal basal ice exposures at Gígjökull and Steinholtsjökull, I review the range of mechanisms by which basal ice is believed to form, prior to selecting those which best apply here. Much of this section follows Hubbard and Sharp (1989, 1995).

A) DEBRIS ENTRAINMENT BY REFREEZING

1. Weertman regelation

This mechanism of basal ice formation is derived directly from Weertman sliding theory (e.g. Weertman, 1979; Paterson, 1994, pp. 135-140), which considers how temperate ice (i.e. ice at the pressure melting point) can bypass bed obstacles which provide resistance to ice flow. Weertman assumes that water flow at the glacier bed takes place within a film of microscopic (micron-scale) thickness, sandwiched between impermeable ice above and impermeable bedrock below. Enhanced stress as ice encounters the stoss face of a bedrock obstacle produces a fall in the pressure melting point: ice melts, and the water passes around the bump to refreeze in its lee. Here, stress reaches a local minimum, with a corresponding rise in the pressure melting point. Water within the film collects in this zone of low pressure and refreezes; this releases latent heat of fusion, which flows back upglacier through the bedrock bump along the thermal gradient created by the difference in pressure melting point, and so sustains the regelation process. Thus refreezing in the lee of bedrock obstacles incorporates any fine debris present.

CLASSIFICATION OF BASAL ICE FACIES					
ICE TYPE/FACIES	TYPICAL THICKNESS (m)	APPEARANCE	DEBRIS CONTENT (1)	FORMATIVE MECHANISM	WIDER CONTEXT
Englacial (meteoric) *	10^1 - 10^2	Clean ice, bubble-foliated	Negligible	Firnification	Near-surface metamorphic processes
Clear *	10^-1 - 10^0	Translucent, contains debris smears and flattened bubbles. May form matrix for other basal ice facies.	Low	Deformation/metamorphism + Lliboutry-type regelation	Hard-rock bed, interface at PMP. 'Dry' bed. Bedrock roughness scale dcm to m.
Laminated *	10^-2 - 10^-1	Closely-space debris laminae, separated by clean, bubble-free ice.	High	Weertman-type regelation.	Hard-rock bed, interface at PMP. 'Dry' bed, low basal melt rates. Bedrock roughness scale cm to dcm.
Interfacial layered *	10^-2 - 10^-1	Stacked series of ice coatings; bed-parallel layering.	Middling?	Cold-induced freezing.	Sheet flow within near-marginal bedrock cavity. Interior: ice/bedrock interface below PMP.
Interfacial continuous *	10^1	Massive ice; sub-vertical debris/bubble lineations.	Middling?	Cold-induced freezing.	Freezing front penetrates standing water within near-marginal bedrock cavity.
Dispersed *	10^-1	Bubble-free; dispersed debris; crudely-layered.	Middling	Entrainment + metamorphism of interfacial ice.	Variable water pressures and sliding rate. Cavity closure.
Solid *	10^-2 - 10^-1	Clast-supported pods or layers of frozen sediment. Ice interstitial.	Very high	Freezing-on of loose sediment.	Marginal ice, unconsolidated substrate.
Stratified *	10^1	Layers (10^-1 - 10^0 m) of debris-rich ice alternate with layers of englacial ice, similar thickness.	Middling	Tectonic inter-stratification of lowermost ice layers.	Compressive flow field.
Individual clasts within ice matrix	N/A	Clast enclosed within basal ice; matrix can be any of basal ice facies. Clear or laminated facies most likely.	Variable	Robin heat-pump/hydraulic jack. Regelation and/or plastic deformation.	Hard rock. Rough bed/cavities. Variable water pressure + relatively high sliding speeds. Low basal melt rates.
W/K-'stratified' (2)	10^0 - 10^1	Debris-rich laminae alternate with clear ice layers. Pods + layers of dense (solid facies) sediment concentration.	High to very high	Pervasive open-system freezing-on; unlimited heat sink.	Hard and/or soft rock beds. Marginal zone of transition, warm to cold thermal regime.
L-'stratified' (3)	10^0 - 10^1	Debris-rich laminae alternate with clear ice layers. Occasional clasts + sediment clots.	High	Pervasive open-system freezing-on within subglacial drainage network; unlimited heat sink.	Hard rock. Soft rock? Channels freeze/clog-up as water tries to exit marginal overdeepening.

Table 6.1
Classification of basal ice facies. See next page for notes.

Table 6.1

Classification of basal ice facies. Largely after Hubbard and Sharp (1995). * denotes ice facies taken directly from Hubbard and Sharp.

NOTES:

1. Debris content: qualitative impression relative to other types of basal ice; exact debris content depends not just on mechanism and immediate context of formation, but on wider availability of debris also.

2. W/K- 'stratified' = Weertman/Knight-type stratified ice.

3. L- 'stratified' = Lawson-type stratified ice.

N.B. Lawson's use of 'stratified' (as followed by Knight and co-workers) is not exactly equivalent to Hubbard and Sharp's use of 'stratified'. Tectonic deformation is a necessary factor for Hubbard and Sharp's stratified facies; tectonic deformation is contingent, but not necessary, for W/K- and L-type stratified ice. The context within which freezing occurs, not appearance, largely differentiates W/K- and L-type 'stratified' facies.

Ice created by Weertman-type regelation tends to consist of a stacked series of closely-spaced layers of clear, refrozen ice, with the upper surface of each picked out by fine, quasi-continuous debris laminae. These layers are broadly parallel to the bed, and extend downglacier from the crest of the bedrock obstacle which gave rise to them (Hubbard and Sharp, 1993). Hubbard and Sharp (1995) term this the 'laminated' facies.

Because Weertman-type regelation is believed to act as an essential component of basal sliding by temperate glaciers, basal ice formation by this process must also be ubiquitous beneath sliding temperate glaciers. However, the thickness of regelation facies rarely exceeds a few cm because ice which forms in the lee of one obstacle tends to be destroyed by pressure melting when it is driven against the stoss face of the next bedrock obstacle downglacier. Weertman's original bed model consists of uniform cubic obstacles which interrupt what is otherwise a plane bed; as much ice melts against the stoss face of any obstacle as refreezes in its lee, so basal ice is continuously recycled within a layer equal in thickness to the height of the cubic obstacles. Larger obstacles (a rougher bed) should therefore allow for a thicker basal ice layer to (re)develop, but there is a limit to this because regelation becomes relatively ineffective once rising obstacle size limits conduction of heat upglacier. The typical value quoted both for Weertman's 'controlling obstacle' and the upper limit for the basal ice layer beneath temperate glaciers is ~0.5 m. Hubbard and Sharp (1993) have re-worked Weertman's analysis using real-world, asymmetrical bed profiles, plus an additional component of net basal melting related to geothermal and frictional heat sources. Model runs which study the evolution of the basal ice layer across ~25 m of glacier bed suggest that only "exceptionally" does regelation-type basal ice reach thicknesses >10 cm; frequently it is destroyed altogether (cf. Box 2.2). The maximum *at-a-point* thickness calculated for the regelation layer was 14.9 cm, for a phyllite bedrock profile substantially rougher than the other three bedrock profiles used.

2. Lliboutry-type regelation

Lliboutry questions Weertman's model (Lliboutry, 1993, 1994). The standard (i.e. Weertman-type) theory of regelation/sliding assumes the presence of a continuous water film trapped between temperate, dry, impermeable ice and bedrock. Pressure melting of ice is assumed to take place directly against the bed, in response to stress fluctuations associated with the redistribution of ice overburden pressure. Lliboutry argues:

1. The Weertman film is a "myth" which has yet to be verified directly in the field, whereas empirical studies demonstrate convincingly that 'blue' basal, bubble-free ice is permeable. Distributed water flow at the glacier bed occurs by means of the capillary network which exists between individual ice crystals.

2. The pressure-temperature field which controls melting and freezing is determined by local stress concentrations acting at the scale of the individual ice crystal, not by the redistribution of normal pressure in the vicinity of a bedrock bump cm to m in extent.

Lliboutry's new theory dismisses the previous emphasis on the lateral transition between stoss-side melting and lee-side freezing in favour of a distinction with height above the glacier bed. Water migrates from the stoss- to lee-side of bumps *within* a bottom layer of ice he estimates to be ~20 cm thick in the case of valley glaciers. Beneath this, immediately adjacent to the bed, is a layer of regelation ice ~3.5 cm thick, which continuously gains ice by freezing on. Energy released by freezing here is balanced by melting within the layer above, and by the drainage of excess water squeezed out of the pores within the ice.

Lliboutry argues that debris is entrained by the regelation process to create a thin, silt-rich layer of basal ice, as indeed is commonly observed beneath valley glaciers. Thus, although the processes involved are subtly different, the end-result is effectively identical to that predicted by the Weertman-type mechanism: a basal layer of refrozen ice, rich in debris laminae. As with the Weertman-type mechanism, in its simple form Lliboutry's mechanism involves the recycling of a finite quantity of heat and mass within a closed system, and so is self-limiting: the transition with height above the bed from freezing to melting prevents the continuous accretion of ice.

Clotted ice. This type of ice was identified by Knight (1987) in marginal exposures on outlets of the West Greenland Ice Sheet, and was largely attributed to small-scale regelation processes (Sugden *et al.*, 1987; Knight, 1989). Knight and Knight (1994) suggest specifically that it is this 'clotted' ice which is characteristic of Lliboutry's pore-water flow mechanism. Clotted ice consists of basal ice containing dispersed aggregates (i.e. clots) of fine debris which collect at the three-grain intersections of the vein network. The speckled appearance of this ice contrasts with the (implicitly) linear debris stratification of the Weertman-type regelation ice. Nevertheless, differences in the debris signature of regelation ice are likely to reflect not just the exact mechanisms of water flow and melting/refreezing, but also factors such as the availability of debris within the freezing zone, and the post-entrainment history of the ice.

Hubbard and Sharp's (1995) 'clear' facies - identified as a ubiquitous feature of the 11 Alpine glaciers studied - corresponds broadly to Knight's clotted ice. Clear facies ice is translucent in appearance, and contains debris smears and clouds of flattened bubbles. Its lack of layering, lower debris content, co-isotopic signature and thickness (up to several metres thick) distinguish it from the laminated facies associated with Weertman regelation. Hubbard and Sharp acknowledge that micro-scale melting and freezing at crystal boundaries and water flow through the inter-granular vein network is probably central to the formation of clear facies ice, but argue that the Lliboutry mechanism is necessary rather than sufficient in this respect. Debris too coarse to be carried within the vein network, plus the observed thickness of the clear facies ice suggest that some additional process must be invoked. Hubbard and Sharp suggest strain-induced metamorphism of englacial ice carried into contact with bedrock obstacles. Intense deformation explains: a) partial recrystallisation, and the distortion/loss of bubbles; b) entrainment and smearing-out of coarse debris; and, c) thickening of the basal ice layer.

3. Robin heat-pump effect

[Robin, 1976; Goodman *et al.*, 1979; Röthlisberger and Iken, 1981]

This means by which basal ice can form also invokes pressure-induced phase changes, but, unlike the Weertman and Lliboutry regelation mechanisms, it is not associated with any specific type of basal ice. It is likely that ice formed by the Robin heat-pump effect resembles either laminated or clear facies basal ice, depending on the water flow processes involved; however, unlike the Weertman or Lliboutry mechanisms, the Robin heat-pump effect can be used to explain the presence of large clasts in basal ice.

The differential loading of bedrock obstacles creates zones of pressure-melting (e.g. stoss-sides of bedrock obstacles, cavity lips) which release meltwater. Whereas Weertman's mechanism involves refreezing of this meltwater in the immediate vicinty, and the return flux of heat upglacier (scale $<\sim 1$ m), Robin envisages that the water from pressure melting escapes along a pressure gradient as flow within either the inter-granular vein network, or at the ice-bedrock interface. This represents a loss of latent heat. With the return to lower pressures/higher pressure melting point - which occurs as ice is carried downglacier, or with a change in the proportion of the weight carried by water collected in the subglacial film (Robin, 1976) or the cavity network (Röthlisberger and Iken, 1981) - ice cannot adjust immediately to the higher pressure melting point because of this deficit of latent heat. Heat must be drawn from elsewhere, which creates a basal zone of freezing up to several metres in diameter. The necessary heat is first derived by freezing of any meltwater, of whatever source (e.g. squeezed from basal ice, or surface inputs). This freezing-on of film water to the base of the ice traps fine debris as with the Weertman regelation mechanism - the key difference lies with the direction of the latent heat flux and the source of the water which freezes. However, if the available quantity of meltwater which freezes liberates insufficient latent heat to raise the ice temperature to its new, higher pressure melting point, the ice will freeze to its bed: the heat required to raise the pressure melting point of the ice is drawn from the bedrock by conduction (i.e. heat is lost at the ice-rock interface), and a local patch of cold ice forms.

The hydraulic jack. Röthlisberger and Iken (1981) extend Robin's analysis to suggest that, when ice freezes to bedrock in this way, fracture can occur within the bedrock, rather than within the ice, or between ice and bedrock, if the bonds within the bedrock are weaker than those involving ice. Change in water pressures within the wider cavity network increase the forwards/upwards velocity of the basal ice at the single cavity scale; as this ice is carried forward, fragments frozen to is base are torn away from the bedrock step. Coupling the Robin heat-pump effect, which creates cold patches at the edge of rock steps upglacier of cavities, to this 'hydraulic jack' provides a partial explanation of plucking [i.e. entrainment of (usually) sizeable rock fragments by a wear process which involves contact between ice and bedrock only]. Röthlisberger and Iken's mechanism accounts for the detachment of rock fragments when the internal cohesion of bedrock has been weakened by some preparatory failure process. Bonds of freezing are rarely of sufficient strength to fracture intact bedrock. Iverson (1991b) suggests that fatigue wear of bedrock by the mechanism of Griffith crack propagation under the influence of repeated fluctuations of cavity water pressure and cycles of loading and unloading explains this preparatory fracture process, and completes the analysis of plucking. Iverson's hypothesis, however, is disputed by Lliboutry (1994) who questions the reality of plucking on the grounds that the near-instantaneous drainage of cavities/loading of the rock step envisaged by Iverson's computer simulation has yet to be verified by field measurement.

Röthlisberger and Iken's analysis suggests how larger clasts can be incorporated into basal ice alongside finer debris trapped by freezing of the water film. Rock fragments frozen to the base of the ice are carried away with the forward motion of the glacier. Clasts are subsequently enclosed by ice as the cavity - within which pressure is now much reduced, and tending towards atmospheric - is closed by the influx of ice (Röthlisberger and Iken, 1981, Fig. 3a, p. 59). Hallet (1979a, 1981) discusses the flow of ice around single clasts within a relatively debris-poor layer of ice immediately adjacent to a bedrock wall (see also 'B) ENTRAINMENT BY TRACTION', below).

As with the regelation mechanisms, there tends to be a limit to the thickness of the basal ice layers produced by the Robin heat-pump effect. The process operates within a closed system which involves the *redistribution* of pressure, heat and mass (i.e. ice/water), rather than any net change in the quantity of heat or mass present. Loss of latent heat is balanced by the release of heat from elsewhere (i.e. latent heat if water is available for freezing, or heat from bedrock if the supply of water is limited), so that the creation of basal ice will tend to be offset by its destruction in zones of enhanced pressure elsewhere.

4. Seasonal freezing-on

In winter, temperatures beneath temperate glaciers can fall below $0^{o}C$, whereupon water and debris in contact with the base of the ice can freeze-on to create basal ice. Basal temperatures can fall below freezing by two means:

1. Cold air penetrates to the glacier bed by way of marginal cavities, or by the subglacial drainage network.
2. The winter cold wave penetrates thin ice to reach the glacier bed.

Hubbard and Sharp (1995) identify two basal ice types which form in this way: a relatively debris-poor 'interfacial' facies, and a debris-rich 'solid' facies. The distinction between the two rests largely with the character of the substrate: interfacial facies form above bedrock, solid facies above unconsolidated sediments.

Solid facies ice - described as "clast-supported pods or layers of frozen sediment; ice is interstitial" - potentially forms a highly effective source of moraine accumulation because of its high debris content. For instance, Krüger (1993, 1995, 1996) interprets the moraine ridge complex developed in the immediate proglacial zone of Mýrdalsjökull's wide and flat northern lobe as the product of seasonal freeze-on. The glacier margin here consists of a feather-edge of ice, which permits an extensive marginal zone of basal freezing to develop each winter. Subglacial till which freezes to the base of the ice shears above the 0°C isotherm, and is carried forward as a layer of basal ice by the winter advance. Debris is released as the ice melts and retreats in the following summer. As the ice margin has been roughly stationary in the 1980s and 1990s, the winter advance has tended to re-occupy the same position each year. In this way, a composite moraine ridge has formed by this cyclical process of winter freeze-on, advance, and summer melt-out. Each cycle adds a new slab of re-worked material to the proximal face of the ridge.

5. Pervasive freezing-on: the importance of thermal regime

The temperature of glacier ice depends upon heat inputs from the glacier surface, heat generated by ice flow, and the geothermal heat flux. Under certain conditions of heat loss, large patches of ice are created within which the basal temperature is permanently below the pressure melting point of ice (Weertman, 1961; Boulton, 1972). This is common towards the edge of glaciers in particularly cold climates (i.e. climates in which summer warming cannot destroy entirely the previous winter's cold wave). As the ice thins, the englacial temperature gradient rises, with the result that heat is lost to the surface faster than it can be replenished from basal and internal sources. This creates a widespread zone of marginal basal freezing. Water draining to the ice margin can then freeze to the base of the glacier, incorporating debris in the process. The continuous nature of the freezing process, together with an abundant supply of meltwater means substantial thicknesses of basal ice can build-up (e.g. Knight, 1995b, especially his Fig. 1). Alternatively, slabs of sediment can freeze-on to the base of the ice as discussed above, but in this case the tendency to freeze-on is more permanent.

6. Freeze-on within subglacial drainage passageways

Whereas the Robin heat-pump mechanism describes freezing-on necessary to maintain ice at its pressure melting point, freezing-on can also be induced by the requirement to keep *water* at the pressure melting point. I discuss above (Chapter 5.6 and Box 5.2) why it is believed that a channel which traverses an overdeepening can be choked/frozen shut because of a shortfall in heat energy released by meltwater flow. Lawson and his co-workers argue that this represents an important mechanism of basal ice formation (Lawson, 1993, p. 55; Lawson *et al.*, 1995; Strasser *et al.*, 1996). Sediment is trapped within the ice as channels close down:

1) individual particles can act as nuclei for the growth of frazil ice crystals; 2) sediment can be trapped within anchor ice which develops on the walls of the channel; and, 3) the open structure of platy ice crystals acts as an efficient trap for sediment in transport. In this way, thick debris-rich sequences of ice can accumulate rapidly. Subsequent metamorphic changes create a sediment-rich stratified basal ice facies.[2] Exposures of this ice in excess of 10 m thick are found at Matanuska Glacier. Analysis of the tritium (^3H) content of the ice reveals that 1.4 m of this ice has accumulated since 1952.

The study at Matanuska Glacier by Lawson and his colleagues is important here because:

1. It provides field evidence in support of a 'new' mechanism of debris-rich basal ice formation.
2. This mechanism is specifically related to the presence of ice-marginal overdeepenings (ice formed in this way at overdeepenings upglacier would tend to be destroyed by subsequent melt).
3. Within open-system flow (i.e. external inputs of meltwater for freezing) the impact of the overdeepening on drainage behaviour creates the fixed tendency for progressive accumulation of ice by freezing necessary to build up major sequences of basal ice.

This mechanism describes a direct link between subglacial water flow, basal ice formation and the delivery of debris-rich ice to the glacier margins. Given the theme of this thesis it is significant that Lawson *et al.* (1995) suggest that large moraines which mark former ice margins at the exit of many glacial troughs reflect this process of basal ice formation.

B) ENTRAINMENT BY TRACTION

1. Entrainment of loose clasts

Accounts of basal ice formation tend to highlight mechanisms whereby debris is incorporated into ice as the ice refreezes. However, a loose clast at the glacier bed will also be removed if the tractive force acting on it (the drag force imposed by the component of ice flow parallel to the glacier bed) exceeds the contact force (a function of the normal force exerted by the overlying ice). The balance of forces involved here is similar to that which controls entrainment of bed-load within a river. For larger clasts ($>\sim 1$ m) the normal force and bed-parallel forces related to the clast's body weight (buoyant weight if the clast is enclosed by the ice) can be important. Clasts can be dragged or rolled along between the base of the ice and bedrock, or

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[2] 'Stratified' in the sense of Lawson, 1979, as cited by Hubbard and Sharp (1989) - see their Fig. 2; Hubbard and Sharp (1995) use 'stratified' to describe basal ice thickened by marginal compression, usage not directly equivalent to Lawson's use of 'stratified'.

can be enclosed and entrained by regelation and/or plastic deformation of ice around the clast. The theoretical distinction here between ice flow around clasts by regelation/plastic deformation and entrainment by Weertman-type refreezing may be somewhat unrealistic; however, the entrainment of clasts of pebble size and larger, which are commonly found enclosed in basal ice, and which cross-cut linear features related to refreezing, has been ascribed to plastic deformation. Experimental work by Iverson (1993) shows regelation of ice into a loose sediment layer can entrain clasts. Hallet (1979a, 1981) provides perhaps the most comprehensive analysis of debris transport by traction, which suggests that, for the simple case of a single, spherical clast above a plane bed of hard rock, clasts will be entrained and remain in motion if the basal velocity of the ice exceeds ~1 cm yr^{-1}. Lodgement can take place at elevated basal velocities if other factors are involved: e.g. body weight of large clasts, zones of unusually high basal melting relative to forward ice velocity, or clast-clast interaction. The thickness of basal ice developed by tractive entrainment of clasts will be defined by the diameter of the clasts affected. Large clasts entrained within basal ice act as quasi-mobile bed obstacles, and so themselves give rise to laminated and/or clear types of basal ice by means of their impact on basal ice deformation and water flow (i.e. by way of Weertman- and Lliboutry-type regelation).

2. Incorporation of pre-existing ice or snow

Loose masses of ice or snow, and any associated debris, can also be entrained by the tractive force of ice flow, and incorporated within the basal ice of the glacier. 'Apron over-ride' is believed to be an important influence on the distribution of debris at the margins of polar and sub-polar glaciers : e.g. Evans (1989) describes how glaciers draining to Phillips Inlet, north-west Ellesmere Island incorporate a mixture of collapsed ice, thrust blocks, previously detached ice-cored moraine and fluvio-glacial sediments as they advance. Apron over-ride has also been observed at several surging glaciers, but tends not to be important at non-surging temperate glaciers because melt of snow and ice blocks tends to be much quicker than glacier advance. However, detached ice which forms in marginal cavities [Hubbard and Sharp's (1995) interfacial ice] can survive for a time sufficient to enable its incorporation into the main body of the glacier. Tison and Lorrain (1987) suggest that entrainment of ice which forms on the floor of marginal cavities accounts for a significant part of the total basal ice facies observed at the Glacier de Tsanfleuron, Valais, Switzerland. These floor coatings form - often overnight - when water draining into cavities encounters subglacial temperatures. As the glacier ice which forms the cavity regains contact with the cavity floor, the drag force imposed detaches - and subsequently deforms - the coatings to create a distinct element of basal ice. Hubbard and

Sharp (1995) use the term 'dispersed' facies to describe the metamorphic product of this interfacial ice.

POST-ENTRAINMENT HISTORY OF BASAL ICE

The mechanism(s) by which basal ice forms and reforms exerts the primary control on its characteristics, but ice characteristics will also reflect factors which act on it after its (re)formation. These factors largely relate to deformation of the ice, which tends to bring about a redistribution of debris (i.e. differential thickening and thinning of basal ice), and/or the destruction of basal ice.

Deformation of basal ice has been studied or inferred at a number of scales, ranging from the metre-scale of single bedrock obstacles (e.g. Boulton, 1979; see below), through the wider pattern of deformation observed in large (10-100 m) cavities (e.g. Rea and Whalley, 1994) to the several-tens-of-kms scale of large parts of the West Greenland Ice Sheet margin (Knight *et al.*, 1994; see below). Hubbard and Sharp (1989, pp. 534-542 and pp. 550-553) provide a comprehensive review; here I touch on those ideas which relate most directly to the links between basal ice formation/preservation and marginal moraine formation.

Debris will tend to congregate at the glacier bed in those areas in which basal melting induces a component of ice flow directed against the bed (Röthlisberger, 1968, cited by Hubbard and Sharp, 1989; Hallet, 1979a). Straining of basal ice will also tend to drive debris towards the bed. Boulton (1979) describes the behaviour of debris-rich basal ice as it encounters a bedrock obstacle at Breiðamerkurjökull, Iceland. Flow separation produces streaming of basal ice into the troughs flanking the bedrock hummock, whereas the basal ice moving across the hummock's summit (i.e. zone of divergence) thins. Basal ice in the troughs thickens because of this streaming, and is enriched in debris because of enhanced basal melt within the troughs, which reflects higher levels of frictional heat generated by the preferred flow routing both of ice and meltwater. It is difficult to be specific as to whether or not this behaviour improves the chances of large quantities of debris-rich basal ice reaching the ice margins; in part, this is likely to depend on where within the glacier debris convergence takes place. Debris convergence thickens basal ice and/or raises its debris concentration, but it also carries debris towards the ice-rock interface, at which it is likely to be most vulnerable to melt-out (and flushing) by heat released by pressure-induced melting, friction from ice flow, geothermal heat, or friction from water flow.

Knight *et al.* (1994) use similar ideas to explain the distribution of clot size within the 'clotted' (clear) ice facies exposed at the edge of the West Greenland Ice Sheet. Irregular bedrock topography with relief of ~400 m gives rise to flow divergence/convergence at a scale of several kms. Sluggish ice, frozen to its bed, occupies the bedrock highs; ice converges on the troughs to create fast-flowing, warm-based lobes. Within clotted ice before it undergoes large scale deformation, clot size tends to decay with distance above the bed. Further downglacier, small clots are found above the bedrock highs and at the termini of the lobes; larger clots are found in greatest profusion midway along the lobe sides. This pattern reflects the two effects just discussed: divergence of ice at the bedrock highs carries larger clots into the troughs, but progressive basal melting at the lobe centres destroys the lowermost layers of the basal ice populated by large clots. Knight *et al.* point out that basal melting appears to be greatest along the axes of major subglacial meltwater routes (presumably this reflects both direct melt of ice in contact with running water plus the indirect impact of water raising sliding speeds and so the quantity of frictional heat generated by ice movement?), and note also that contrasts in moraine composition and size can be expected to relate to the wider ice and water flow pattern. Far-travelled lobate debris is expected to be finer and to exist in greater volume (presumably debris production relates positively both to length of flow line and basal ice velocities) but finer debris which co-exists with zones of maximum meltwater activity is likely to be flushed away.

This study by Knight *et al.* represents one of the few published papers which 1) tries to make a link between *interior* formation of basal ice and the character of ice-marginal sedimentation, and, 2) identifies the crucial difference which exists at the transport pathway/sediment budget level of analysis between basal ice production and basal ice preservation. However, the scale of analysis which includes regional contrasts both in bedrock relief and of thermal regime is somewhat removed from the scale of debris transport systems such as Gígjökull, Steinholtsjökull or Sólheimajökull.

Whereas flow divergence/flow extension works to thin out ice, flow convergence and flow compression tend to thicken it. Compressive flow takes place where the ice velocity parallel to the glacier centre-line (i.e. x direction) falls downglacier. Thus it tends to be especially marked towards glacier termini (Reid, 1896; Paterson, 1994, pp. 253-254), although it can also relate to changes in bedrock topography or changes in the character of the bed. Hubbard and Sharp's (1995) study identified a thick (>10 m) 'stratified' facies of inter-folded debris-rich basal ice and clean englacial ice at Glacier de Tsidjiore Nouve and Glacier de Giétro related to pronounced marginal zones of compression (see also Tison *et al.*, 1989, especially their Fig. 1). It appears that, given a suitable flow field, tectonic deformation can stack basal ice to thicknesses which exceed those typical of temperate alpine glaciers by an order of magnitude. Such tectonic stacking carries basal ice away from the 'danger zone' of potential destruction adjacent to the glacier bed.

SUMMARY

Table 6.1 sets out the major types of basal ice. In effect, this represents a list of 'multiple working hypotheses': each basal ice type represents a possible candidate for the type(s) of basal ice which we see at Gígjökull and Steinholtsjökull. Firm identification of basal ice types usually requires direct access to the subglacial environment and/or complex chemical and/or co-isotopic analysis. Without these, any interpretation attached to basal ice must be tentative; however, given careful reasoning certain processes/basal ice facies can be inferred with reasonable certainty to be less likely than others, or, indeed, an impossibility. Before I proceed with this reasoning, however, I wish to step back to examine the wider contexts within which basal ice links to ice-marginal sedimentation, and to support this with the case study of basal ice at Variegated Glacier, Alaska (Sharp *et al.*, 1994), which, in certain key aspects, appears to be strikingly similar to Gígjökull.

6.3 WIDER PERSPECTIVES

In this section I move beyond the mechanisms by which basal ice forms to consider the wider relationships which control the transport of debris within basal ice at the scale of the full glacier. Two questions are of special importance here:

1. What governs availability of debris for incorporation into basal ice?
2. What determines whether debris within basal ice survives its journey to the ice margin?

These two questions change the study framework from one which concentrates on single process events to one which emphasises sequences of different process events, linked in time and space, which together constitute a causal chain (cf. Chapter 1.4 and 1.5). In this way I recast the basal ice 'problem' at a larger scale, in terms which relate directly to the issue of moraine formation. I use Sharp *et al.*'s (1994) study of Variegated Glacier, USA to illustrate some of the important points, and proceed to argue that similar things apply to Gígjökull and Steinholtsjökull also (6.4, and Chapter 7).

Although the distinction is far from perfect, there seems to be a tendency to approach the study of basal ice at three levels:

1. **Mechanics of basal ice formation.** Usually theoretical or experimental, this reductionist approach isolates individual processes without reference to the wider context. What context is included usually represents the abstract, ideal conditions which define the (semi-realistic at best: e.g. uniform cubic bed obstacles) parameters within which the process in question is believed to operate. Such analysis provides a fine example of system closure to define an artificial system (see Chapter 1.4).

2. **The single rock-step/cavity scale.** This studies the interaction of processes of erosion and entrainment of debris as ice moves across a 'rough' bed.

3. **The (long) flow-line scale.** This takes a wider catchment perspective, within which marginal exposures of basal ice are taken to reflect the sequence of events which occurs upglacier.

1 feeds into 2 and 3: 2 tends to relate to smaller glaciers, particularly temperate alpine-type glaciers (e.g. Glacier de Tsanfleuron and Glacier de Tsidjiore Nouve, Valais, Switzerland; Souchez and Lorrain, 1987). Field studies often rely on chance exposures of basal ice in marginal cavities, or on access to the bed by subglacial tunnels excavated for engineering, usually HEP, schemes (e.g. Souchez *et al.*, 1973; Jansson *et al.*, 1996). Studies at this scale usually try to resolve in some detail fundamental processes such as ice flow around single bedrock obstacles, meltwater flow through individual cavities, and the associated mechanisms of subglacial erosion. Studies which try to extend the analysis beyond the immediate vicinity of the bed access point are far from common. In contrast, the flow-line scale tends to emphasise larger glaciers, particularly ice sheets which have a marginal zone of 'cold' ice. At this larger scale (several kms) the importance of individual process mechanisms tends to be reduced, and processes are subsumed within accounts of wider patterns such as downglacier transitions in thermal regime, or the switch from extending to compressive flow. Knight *et al.* (1994), discussed above, provides an excellent example of such work. Without question this kind of large-scale pattern is important: explanatory schemes such as that of Knight *et al. work* - to follow Sayer (1992), studies like this merit the label 'good science' because of their 'practical adequacy'. However, it is often necessary to read 'between the lines' to identify the relevant process links, and the large-scale patterns which inform these studies tend to work less satisfactorily at the smaller scale of the typical valley glacier. Different scales of system require different levels of analysis (Schumm and Lichty, 1965; Chapter 1.4 and Chapter 9). In my view, with respect to our attempts to understand moraine formation at valley glaciers, studies of basal ice have largely missed the point.

Much of what we understand of basal ice development stems from theories of glacier sliding and basal hydrology (regelation, water films, inter-granular vein networks, enhanced plastic deformation, etc.). Such treatments purport to explain the transfer of ice and water, yet it is this key feature of *flux* which is frequently ignored by isolated studies of basal ice. Given that its high debris content is usually identified as the outstanding feature of basal ice, it seems to me to be a yet greater paradox that studies of basal ice - particularly theoretical and experimental studies - usually take the presence of debris for granted. Recent developments in the study of subglacial erosion (Iverson, 1995) tend not to make much impact on studies of basal ice; studies of the dispersion of debris by different sediment transport pathways (by ice, water, or deforming sediment) figure less highly still. However, the production of basal debris, and its allocation between different transport pathways, must be controlled by wider ice and water dynamics, which, in turn, is to stress process linkages. This is why I break down the basal ice 'problem' into three stages (see above). What seems to be required is some kind of compromise between 2 and 3: analysis of basal ice at the level of the *linked* steps and cavities which are found along a valley glacier flow-line. Marginal exposures of basal ice reflect the cumulative impact of the diverse space-time interactions of ice, debris and water upglacier. Hubbard and Sharp (1995), Sugden *et al.* (1987) and Knight (1995b) put forward similar arguments: by piecing together different bits of knowledge pertaining to the subglacial environment, we can use basal ice exposures to reconstruct in some detail what happens in the subglacial interior. To reverse this logic: if we can infer what happens in the interior of the glacier, we can begin to understand why, in certain cases, we find abundant sediment accumulation fed by basal ice. Such accounts are likely to rely on inference and interpretation rather than 'hard facts' (Frodeman, 1995; Chapter 9). It is reductionist, 'rigorous', often quantitative treatments at the theoretical and empirical levels which provide the basic building blocks, but sediment accumulates at the edge of glaciers because of process *linkages*, within which certain quasi-stable patterns emerge. The complex nature of these links tends to defy rigorous treatment. The level at which individual processes operate (e.g. single fracture episode) is not identical to the level at which the major patterns emerge (e.g. progressive erosion of a bedrock trough): e.g. see the different - if complementary - analyses of plucking by Iverson (1991b) and Hallet (1996).

LINKED STEPS AND CAVITIES

Availability of debris

Fast sliding over a rough bed seems to be the fundamental requirement for high rates of subglacial erosion. Coupled analyses of plucking and abrasion which work across the entire glacier bed do not exist; however, it seems widely accepted that rates of bedrock erosion scale with some power of sliding velocity ≥ 1.0. Hallet's (1979a, 1981) abrasion model suggests that abrasion rates scale with the square of sliding velocity, but the relationship between sliding speeds and plucking has a less rigorous theoretical grounding. Shoemaker (1986, cited by Iverson, 1995) suggests that the relationship is linear, but alternative reasoning [e.g. widespread survival of *roches moutonnées*, or the prevalence of coarse clasts in ice-marginal deposits (General Discussion, *Journal of Glaciology*, 23, 89, pp. 385-392)] indicates that rates of plucking and abrasion must be similar. This in turn implies that abrasion and plucking rates relate to sliding speeds in much the same way (i.e. both proportional to the square of the sliding speed?). However, analysis of the sediment load carried by the stream draining Variegated Glacier during its 1982-1983 surge event implied that a linear relationship exists between sliding speed and bedrock erosion (Humphrey and Raymond, 1994). This ignores potential gauging errors (e.g. quantification of the bed-load flux?) or the possibility of subglacial sediment storage. Humphrey and Raymond point out that drilling turned up no evidence for the accumulation of a deforming bed during the surge event, but seem not to have considered the possibility that large quantities of sediment may have been trapped within basal ice (see below).

Climate (mass balance) and topography (which influences glacier geometry) exert the primary controls on sliding speeds (Andrews, 1972; Sugden and John, 1976, Ch. 3). However, recent research has focused on the relationships between cavity formation, subglacial water pressures, sliding speeds and rates of subglacial erosion. These operate within the wider constraints climate and topography impose on ice behaviour, but the details of these processes are not self-evident in simple power law sliding speed/erosion rate relationships. Water pressure within cavities is believed to influence sliding speeds (rising and/or high water pressures enhance sliding speeds). Change in sliding speeds relates to change in water pressures, and - by inference - change in the glacier-wide extent of cavitation (see Chapter 1.1). Tendency to high subglacial water pressures will enhance basal debris production. However, if erosion rates do scale with some power of sliding speed > 1.0, non-steady ice flow will further elevate erosion rates, an effect which is potentially important, not just at surging glaciers, but at any glacier which experiences time-dependent changes in its basal drainage (Sharp, 1988, No Date). The propensity for cavities to form will reflect the balance between basal shear stress and effective normal pressure (Bindschadler, 1983; Hallet, 1996). Ice thickness and ice surface slope are seen as key controls on cavity formation (separation is favoured by steep, thick ice) but the character of subglacial drainage and the bedrock geometry (mean bed slope, dip upglacier of the stoss faces of bedrock obstacles) are important also (Röthlisberger and Iken, 1981; Iken and Bindschadler, 1986).

Taken together, the analyses of Robin (1976), Hallet (1979a, 1981) Röthlisberger and Iken (1981) and Iverson (1991b) suggest that rapid ice flow across a rough bed creates conditions particularly conducive to the formation of debris-rich basal ice (i.e. high rates of both debris production and debris entrainment). The widespread formation of cavities, within which water pressure fluctuates, encourages 1) bedrock fracture, 2) debris entrainment by freeze-on because of the extensive development of cold patches by the heat-pump mechanism, and, 3) debris entrainment by traction as the cavities close down. These three processes are all suppressed where ice/bed separation does not occur widely, or where water pressures stay consistently high. Debris which accumulates within basal ice in the lee of rock steps promotes active abrasion (which releases further debris to basal ice) of the stoss face and top of bedrock bumps downglacier. Abrasive wear of a rough bed is further intensified by the relatively high clast-bed contact forces and high bed-tangential velocities produced by the enhanced strain as ice flows around obstacles.

Preservation of basal ice

Active erosion coupled with active basal ice formation/debris entrainment is just part of the picture, however. The flux of debris-rich basal ice must be preserved. Debris can be lost after initial erosion but prior to initial entrainment, or it can be lost by release from the basal ice which encloses it. However, as much basal ice continuously melts and re-forms, this distinction is not particularly meaningful. Debris can be lost to a subglacial till layer (which can be deforming or lodged) or melted out and flushed by meltwater. The processes by which debris passes from basal ice to subglacial till are not well understood. Hallet's abstract analysis of contact force and particle velocity makes it difficult to explain 'lodgement' under conditions likely to prevail beneath typical temperate valley glaciers, although such till layers are believed to exist (e.g. at South Cascade Glacier and Storglaciären). As with basal ice preservation, accumulation of a subglacial till layer also implies inefficient meltwater flushing, so the basal ice and subglacial till layer transport pathways/storages share similar relationships to meltwater activity - or the lack of it. See Chapter 2.5 for a full discussion of meltwater flushing.

Two points are worth noting here. The first is that destruction of basal ice is not in itself damaging to the basal ice flux - indeed, large parts of basal ice must melt as part of the regelation/deformation processes by which sliding, and so the flux of basal ice towards the ice margins, is achieved. A steady flux of debris takes place at the same rate as the recycling of the basal ice layer. In this way, a 'parcel' of ice/water and debris is progressively 'shunted' downglacier, although its exact physical identity is continuously redefined. It is the destruction of basal ice which permits debris to be removed from any given parcel which detracts from the basal debris flux. Beneath valley glaciers I suggest that it is meltwater action which is most likely to intervene in this way, but this aspect of the relationship between meltwater flow and basal ice development has received little attention. However, work by Hubbard and Sharp (1993, 1995) at the Glacier de Tsanfleuron - which is believed to have a well-developed linked cavity drainage network which covers approximately half the glacier bed (Sharp *et al.*, 1989a) - identifies meltwater flushing as a critical process. Their computer simulation, which assumes the presence of a Weertman film only, predicts the build-up of a pretty-much stable layer of regelation ice several cm thick. However, laminated facies regelation ice is not found at Tsanfleuron. Hubbard and Sharp attribute its absence to the distinctive effect of contact between the waters of the linked cavity system and the basal ice layer. Flow within the linked-cavity network captures water from the film (with any fine debris also in transport) and removes it before it can refreeze; plus it melts ice and removes debris with which it is in contact. The widespread calcite precipitates found at Tsanfleuron give further support for the reality of this flushing effect. The isotopic signature of the precipitates indicates formation with a locally-derived (i.e. from melt of basal ice) meltwater film, enriched in heavy isotopes. This enrichment seems to require removal of light isotopes, which can best be explained by melting and flushing of the isotopically-light lower parts of the regelation ice layer.

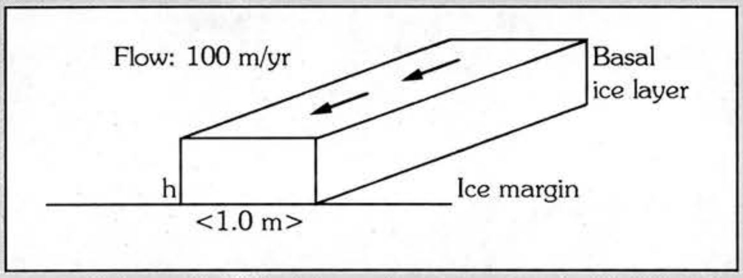
My second point is that basal ice layers need not be over-thick to be capable of building up substantial moraines. The fact that basal ice layers rarely exceed 0.5 m thickness (e.g. Sugden and John, 1976, p. 161; Lawson, 1993, p. 53; Kirkbride, 1995a, Fig 8.3, p. 268) itself is not sufficient reason to question the efficacy of subglacial debris transport pathways as the source of large moraines: e.g. as do Matthews and Petch (1982) with respect to the moraines of the Jotunheim area of Norway. Given suitable time, a basal ice layer which is 0.5 m thick or less, and which moves at reasonable speeds, can build up sizeable moraines (see Box 6.1) - but only if each ice/debris parcel stays together. Matthews and Petch's scepticism perhaps relates to this 'lack of thickness' trick; it is not the thickness of basal ice so much as its debris content and spatial continuity which is crucial to the debris flux. I suggest that the increased thickness of basal ice does not directly raise the debris flux because, at a rough guess, thick exposures of basal ice result largely from tectonic thickening associated with flow deceleration. It is the indirect effect of basal ice thickening (i.e. ice transfer in the vertical plane, not the horizontal) which seems to be important: as the basal ice thickens, debris is increasingly separated from the destructive action of meltwaters running at the glacier bed. I suspect that it was the patchy *extent* of marginal exposures of basal ice which was the true source of Matthew and Petch's scepticism.

BOX 6.1
Simple reconstruction of the basal debris flux

These simple calculations suggest that, *if* the identity of basal ice parcels is preserved, a constant flux of debris within a relatively thin (but coherent) basal ice layer can give rise to substantial moraine accumulations. The calculations 1) predict the flux of debris through a unit length of ice margin; and, 2) convert this flux to the height/cross-sectional area of a single, symmetrical moraine ridge. I assume here that net loss of debris whilst basal ice is in transit is negligible.

The inputs used are thought to be representative of Gígjökull:

- Basal ice velocity = interior glacier sliding speed = 100 m yr^{-1}.
- Basal ice debris content = 10% by mass = 4% by volume.
- Angle of repose for basal till = 30°.
- Void ratio of moraine ridge = 30% (after Small, 1987b).
- Basal flux per unit width towards the ice margin = 100 h m^3 yr^{-1}, if h is the thickness of the basal ice layer.



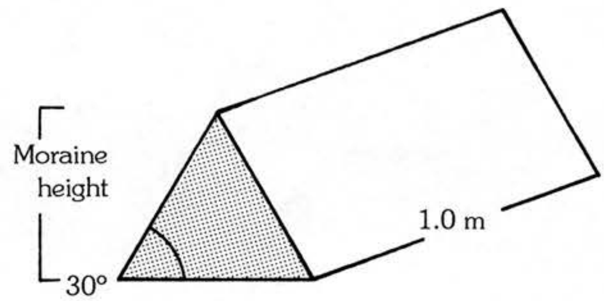
The yearly debris flux is equal to the basal ice flux multiplied by volumetric debris concentration: i.e. (100 h m^3 yr^{-1}) x 0.04. The high sliding speed representative of the glacier centre (Chapter 4.1) is used here because I assume:

1. That the bulk of debris is derived from the interior parts of the glacier bed;
2. That flux continuity is preserved because the sharp fall in velocity towards the ice edge is compensated for by a thickening of the basal ice layer. E.g. it is likely that ice flow near to the ice margin at Gígjökull is reduced to 20 m yr^{-1} or less; however, I assume that this five-fold reduction in velocity is accompanied also by a ~five-fold increase in the thickness of the basal ice layer. The thickness of basal ice exposures seen at the edge of Gígjökull suggests that this assumption is not implausible.

(PTO)

Box 6.1 (continued)

The debris flux is assumed to build a moraine ridge in the from of an isosceles triangle, with sides sloping at 30°:



MORAINE ACCUMULATION: RESULTS			
Basal ice thickness (h), m	**Moraine volume, 1 yr, m^3**	**Moraine height, 1 yr, m**	**Moraine height, 100 yrs, m**
0.10	~0.57	~0.58	~5.80
0.25	~1.42	~0.91	~9.10
0.5	~2.86	~1.28	~12.80

EVALUATION

These calculations are very simple, but seem to fit field observations from Gígjökull quite well. Individual ridges adjacent to basal ice, each believed to represent a single year's accumulation of debris were 0.5-1.0 m in height. However, it is important to stress that this model assumes continuity of the debris flux within basal ice. I believe this to be that case beneath large parts of Gígjökull's terminus, but beneath other glaciers (e.g. Sólheimajökull: Boxes 2.1 and 2.2) basal ice destruction is likely to be more pervasive.

CASE STUDY: VARIEGATED GLACIER, ALASKA, USA (Sharp *et al.*, 1994)

Variegated Glacier is a surging glacier in the St Elias Range, Alaska, USA. The 1982-1983 surge was studied in depth to give a detailed picture of the relationships between basal drainage, ice flow dynamics and debris production (Kamb *et al.*, 1985; Sharp, 1988; Humphrey and Raymond, 1994). Sharp *et al.* (1994) discuss the basal ice facies produced prior to and during the surge event. The features derived seem strikingly similar the basal ice at Gígjökull, which suggests similar origins. Two key features characterise the basal ice exposures at Variegated Glacier: 1) the thickness of the exposures, and, 2) the contrast in the styles of basal ice development associated with quiescent and surge phases.

Surge phase ice consists largely of debris-rich laminated facies (mean debris content = 10.9% by volume = ~25% by mass). This reflects rapid rates of debris production associated with fast sliding with cavitation during the surge event. This ice forms by open-system regelation refreezing. The upper layers of this laminated ice also formed by freezing but are believed to relate to the quiescent phase because of contrasts in the particle size distribution. Relative to the (lower) surge-phase ice, the upper laminated ice is depleted in coarse gravel (bedrock fracture is suppressed by less rapid sliding) but enriched in fine debris (the quiescent phase conduit system covers much less of the glacier bed than did the distributed linked cavity system of the surge, which means much less effective flushing of fine debris). Large parts of the upper basal ice, however, consist of relatively debris-poor clear facies ice, produced by metamorphism of englacial ice. This too is likely to reflect contrasts in the style of basal drainage. Hubbard and Sharp (1995) suggest that clear ice reflects basal ice flow controlled by large bed roughness elements, whereas laminated ice reflects basal ice flow largely controlled by small roughness elements. This switch in the scale of dominant roughness is consistent with the 'drowning' of large parts of the bed with the switch from discrete to distributed drainage.

The total thickness of the basal ice exposures rises from 0.5 m to ~13 m over a distance of <1 km. This reflects the intense horizontal compression experienced immediately in advance of the surge front: horizontal shortening measured at the glacier centre-line rose from 10% shortening to 55%, with a corresponding increase in the vertical extent of ice, over the same distance. Thus the basal ice which starts off as a typically thin (i.e. <0.5 m) layer thickens downglacier with progressively more intense fabric deformation. Upglacier, major structures within the basal ice consist of rooted, open, upright folds; downglacier folds are rootless, recumbent, and highly skewed in the direction of ice flow. Low-angle thrust faults cut the overturned limbs, with the fold hinge and upper limbs tending to be boudinaged such that fold structures lose their identity. This shows how initial failure with simple folding of a weak basal ice layer is supplemented by rising strain imposed upon the developing folds. The end result is pervasive deformation with a tendency to rotate structures to parallel the major (i.e. bed-parallel) plane of shear (Sharp *et al.*, 1994, Fig. 7). Initial (i.e. low-intensity) deformation structures tend to be destroyed as the basal ice is increasingly homogenised (see also Hart, 1995c).

This study is important because it shows:

1. That the character and debris content of basal ice depends on high rates of debris production.
2. That switches in the style of subglacial drainage can change the style of basal ice development, both by their effect on ice flow and erosion processes, but also by changing the efficiency of subglacial flushing.
3. How tectonic deformation associated with intense compressive flow can produce thick sequences of basal ice of extent and debris content, and which give rise to a style of ice-marginal sedimentation, usually associated with sub-polar glaciers.
4. How tectonic thickening helps preserve basal ice against the destructive effects of basal melting and debris flushing.

These factors all relate to contrasts in ice dynamics and basal drainage particularly characteristic of hard-bed, surge-type glaciers. However, similar, if less marked time-space contrasts are also found at many non-surging glaciers. I wish to suggest that different patterns of basal ice development and ice-marginal sedimentation at non-surging glaciers reflect these contrasts also. The case studies of Gígjökull and Steinholtsjökull certainly seem to support this.

6.4 GÍGJÖKULL AND STEINHOLTSJÖKULL: TWO MODELS OF BASAL ICE DEVELOPMENT

The arguments set out in the preceding pages provide the detailed framework which is required to interpret the high rates of basal ice-fed sediment accumulation at Gígjökull and Steinholtsjökull. To reiterate: this requires a catchment-scale perspective which examines the wider processes of ice, water and sediment transfer. In this respect, I suggest that hydrology is a critical factor, both through its direct influence on subglacial debris transport, and its indirect influence on ice dynamics and subglacial processes of erosion. The model I introduce here might well apply to other glaciers with similar geometry. By way of contrast with the previous case study of Sólheimajökull, I believe it supports the hypothesis of debris evacuation developed in Chapter 2.

MECHANICS OF BASAL ICE FORMATION: EVALUATION

It is possible to rule out several possible mechanisms of basal ice formation with a firm degree of confidence. Pervasive freezing-on because of a change in thermal regime suggests a simple explanation for the thickness of basal ice, but it is assumed that Gígjökull and Steinholtsjökull are wholly temperate ice masses, as are the vast majority of glaciers in Iceland. This hypothesis is thus disqualified. Similarly, I reject the hypothesis that the basal ice is the product of repeated seasonal freezing. This has been identified as an important factor which gives rise to significant ice-marginal sedimentation elsewhere in Iceland (Krüger's work - see above), but at an interior, highland (~700 m asl) location. The steep, thick ice fronts, maritime location, and relatively low snout altitude (180 m Gígjökull; 250 m Steinholtsjökull) are likely to work against major episodes of winter freezing-on. Were winter freezing-on to occur, given the unconsolidated nature of the substrate at the ice margins, widespread development of solid facies ice is expected, but no true solid facies ice was found. Lack of marginal cavities and drainage passageways implies that development of ice within such environments is minimal, and no suitable candidates for interfacial basal ice facies were identified. Within this temperate environment melt of marginal snow or ice is likely to occur at rates faster than ice advance, so incorporation of pre-existing ice or snow in large quantities is not thought important (although *isolated* patches of rotting, debris-laden snow stuck to the base of the ice were found). This leaves just two likely candidates: Lawson-type 'stratified' ice, or the bundle of processes/ice facies which relate to sliding over a rough bed: clast-rich clear and laminated ice (Table 6.1).

Lawson-type 'stratified' ice

Lawson's mechanism of basal ice formation by freezing-up of subglacial drainage passageways is particularly attractive given the evidence that the termini of both glaciers lie above overdeepenings. This mechanism accounts for the accumulation of large thicknesses of debris-rich basal ice. However, although the topographic context seems suitable, it is not possible to confirm that this type of ice exists at Gígjökull and Steinholtsjökull. Distinct visual criteria which firmly identify this refrozen ice have yet to identified - in part because of post-freezing metamorphism. The thick stacks of crudely stratified basal ice at Gígjökull and Steinholtsjökull are broadly consistent with Lawson's descriptions of the refrozen facies at Matanuska Glacier, but can be ascribed to alternative mechanisms with equal confidence. It is the tritium analyses of the Matanuska ice which confirms its drainage system freezing origin (Strasser *et al.*, 1996);

without this kind of analysis, such an origin cannot be confirmed at Gígjökull and Steinholtsjökull.

Clast roundness clues. The contrast in the character of coarse clasts (i.e. >3.0 cm diameter) released from a) the basal ice, and, b) the relict conduit debris bands raises questions about the origin of the basal ice. Clasts from the basal ice are predominantly angular to sub-angular; clasts from the debris bands predominantly sub-angular to sub-rounded (Table 4.3; Box 5.1). Röthlisberger/Lawson and Hooke/Pohjola present different pictures of what happens to basal drainage as water pressures rise. Hooke and Pohjola argue that water is squeezed into englacial passages as water pressures rise, carrying water-worked debris with it, as seems to occur at Gígjökull and Steinholtsjökull. The alternative Röthlisberger/Lawson scenario has water squeezed out into a distributed basal drainage network, where it freezes. If this happens here as well, why then does the basal ice so formed not contain rounded debris? It is highly unlikely that the rounding occurs only within the englacial part of the drainage system, whereas preferential selection of rounded clasts for englacial routing seems impossible. Widespread 'unrounding' of clasts re-entrained within basal ice seems highly implausible. The contrast in clast character seems therefore to indicate that the debris bands and basal ice do not both relate to a single integrated drainage network which divides, but originate within distinctly different transport pathways. Two possibilities exist:

1. Two separate drainage networks exist: one in which rounding of sediment occurs before basal drainage becomes englacial drainage, and one within which rounding does not occur before the network freezes shut to form the basal ice.
2. The basal ice does not form by freezing-shut of a subglacial drainage network.

MODEL 1

2 favours the 'rough bed sliding' hypothesis for the origin of the basal ice; first, however, I wish to explore the plausibility of option 1. This requires twin drainage networks, between which there is limited interchange of water and debris. The term 'parallel' (as in parallel electrical circuits - as opposed to 'in series') is often used to describe this kind of situation. Debris rounds within network A, but not within network B. If there is no difference in clast lithology this implies that water and debris within network A is farther-travelled, and/or that flow power within network A is higher. To extend this logic: contrasts in discharge carried and stream power can be expected to give rise to differences in typical water pressures (Chapter 1.1). The co-existence of two separate networks with different pressure characteristics (relating to number of channels, channel geometry, discharge, etc.) possibly explains why basal water flow is

partitioned between two distinct pathways as pressures rise: i.e. the 'choice' of flow pathway used to escape the overdeepening depends upon initial flow conditions. Network A, and its rounded debris, becomes englacial; network B, and its angular debris, stays at the bed and freezes. What the key contrasts of initial flow which determine this choice might be is not known.

The difficulty here is how to account for the stable co-existence of two drainage networks with different pressure characteristics: lower pressure flows capture water - and debris? - from higher pressure flows, to create a single, integrated network (Shreve, 1972; Röthlisberger, 1972). This implicitly assumes that different parts of drainage are given sufficient time/space to unite. It is conceivable that the transience argument applies with respect to Gígjökull and Steinholtsjökull: i.e. the drainage networks are not given the chance to unite as a single stable feature because the perturbation introduced by the overdeepening intervenes. However, several different ideas exist which purport to explain the presence of drainage divides within/beneath a single glacier:

1. **Reversal of the hydraulic potential gradient.** This can happen given suitable interaction of ice surface slope, ice thickness, bed topography, and water flow characteristics (Shreve, 1972; Bjornsson, 1988; Sharp *et al.*, 1993).

2. **Weertman's 'pressure dam'** (Weertman and Birchfield, 1983b; Menzies 1995b). This involves reversal of the pressure gradient in the immediate vicinity of subglacial conduits which prevents water from draining into these conduits. The presence of major axes of discrete drainage (e.g. network A, within which rounding occurs?) is believed to split the glacier bed into several distinct drainage basins; large areas of supposedly stable distributed drainage (e.g. network B - no rounding?) exist in the areas between conduits. This reflects the bridging effect associated with low pressure conduit flow. Large thicknesses of ice bearing down on the conduit roof are not fully-supported by the pressure of water within the conduit. This imbalance must be compensated by a zone of excess pressure: the ice which makes up the conduit walls takes the extra weight of the conduit roof (tens or hundreds of metres of ice). The transfer of load creates pressure *levées* which drive the water at the bed *away* from the conduit.

3. **Contrasts in bed make-up.** If the glacier bed consists of a mixture of hard rock ridges and soft, permeable till patches this can produce a complex pattern of water pressure relationships which create different subglacial drainage basins. Fountain (1993) believes this may be important in the case of South Cascade Glacier.

4. **The valve effect** (see Chapter 5.6). Flow within ice which descends gains energy relative to the changing pressure melting point; flow which ascends loses energy relative to the changing pressure melting point. This makes it difficult for water to flow from low to high elevations, to the extent that transfer of water between drainage networks at different levels can become impossible. Röthlisberger (1972) uses the valve effect to argue for the stable co-existence of subglacial and marginal gradient conduits.

These theoretical arguments are backed-up by ample field evidence of parallel drainage structures of some kind, including: Pasterzengletscher (Burkimsher, 1983b), Midtdalsbreen (Willis *et al.*, 1990), Haut Glacier d'Arolla (Sharp *et al.*, 1993), Storglaciären (Seaberg *et al.*, 1988; Hooke *et al.*, 1988), South Cascade Glacier (Fountain, 1992, 1993, 1994) and Gornergletscher (Iken *et al.*, 1996). The last two examples exhibit parallel drainage and overdeepened basins, but whether the relationship is causal or coincidental is not clear.

Hypothesis

It is clear that the possibility of two parallel drainage systems enjoys widespread support. Perhaps the best clue is the fact that this hypothesis requires that the two systems are sufficiently different to give rise to significant differences in clast roundness. What follows represents a plausible scenario which merits presentation. I cannot claim it is correct, nor can I disprove it. Network A consists of relatively large, debris-laden conduits within which flow energy is sufficient to round clasts rapidly. These conduits drain the crater and the major part of Gígjökull's ice-fall, so discharge within each conduit is expected to be relatively high, and water/debris relatively far-travelled. Network B consists of some mixture of small conduits, cavities and films which drain the exit of the ice-fall and the main part of the overdeepening. Relative to Network A, discharges and travel distances are small, sediment load is angular, flow is sluggish and water pressures high. However, the Weertman pressure dam effect protects the water and debris of B from capture by A. Water pressures rise in both networks within the overdeepening. In response, the conduits of Network A take up a relatively low pressure englacial route after the Lliboutry or Hooke-Pohjola models (or indeed some hybrid of the two). Thereafter englacial debris bands form (see Chapter 5). The water and debris of Network B, however, cannot follow suit: insufficient energy is available to overcome the valve effect, so the water is trapped at the bed within some kind of collapsing distributed system (Röthlisberger and/or Walder-Fowler scenarios). Widespread freezing occurs, trapping the angular, poorly-sorted debris load as Lawson-type basal ice.

QED - or perhaps not! How likely is it that a relatively restricted drainage network such as this hypothetical Network B can access the large quantities of debris, or transport the coarse clasts, as found in the basal ice? This is unresolved.[3] These doubts favour the alternative hypothesis, that the basal ice is *not* Lawson-type 'stratified'. This implies that the vast bulk of basal drainage takes a single route (although there seems to be no overwhelming reason why this should be so). At Gígjökull and Steinholtsjökull the presence of the relict conduit debris bands must mean that, if this is the case, the englacial route is the single pathway of choice. Chapter 7 explores the possible consequences of this. This single choice scenario presumably excludes residual drainage of local meltwater which drains at the bed - as a Weertman film, within the vein network, or within a till layer? This scenario seems to apply at Storglaciären, wherein all but this residual meltwater leaves the bed: when asked how much drainage within the Storglaciären overdeepening stays at the bed, Roger Hooke replied "absolutely zip!" (personal communication, Reykjavík, 1995).[4] Lawson does not mention evidence of englacial drainage at Matanuska Glacier, so it is possible that all drainage stays at the bed there - but Hart (1995) and Richard Waller (personal communication, Leeds, 1995) report the presence of englacial debris bands. Hart believes these to be thrust features, but without firm evidence (soft beds absorb stress and suppress thrust activity?)...

MODEL 2

This assumes that freezing-induced shutdown of subglacial channels has little direct effect on basal ice formation (this does not exclude possible freezing-shut of the residual basal drainage). The appearance of the basal ice at Gígjökull and Steinholtsjökull supports an alternative origin:

1. Its laminar character suggests Weertman-type regelation.

2. The presence of clear pockets with debris clots suggests Lliboutry-type regelation, plastic deformation and metamorphism.

3. The presence of coarse clasts suggests bedrock fracture and entrainment by traction.

These processes/basal ice types are all consistent with rapid interior sliding over a hard bed (Hubbard and Sharp, 1995; 6.2, above; Table 6.1), which is exactly what I infer for Gígjökull and Steinholtsjökull (Chapter 4.1). Hubbard and Sharp argue that both clear and/or laminated

[3] It can be argued that debris within the basal ice has not been carried within Network B, but is incorporated from the store of till plastered upon the adverse slope as water refreezes around it. This presents two obstacles, however: 1) basal ice so formed should resemble solid facies ice, not some mixture of clear and laminated facies; and, 2) if debris is not being brought in by the water, the store of till must eventually be exhausted.

[4] Water flow within/above Storglaciären's till layer appears to be coupled *in series* with the englacial drainage network: i.e. hydraulic connections exist between the till and the conduits (Hooke *et al.*, 1997).

facies ice present in abundance represents isolation of basal ice from all drainage types bar the basal film and vein network. Water transport of large clasts requires a high-competence conduit or cavity network (prior to freeze-induced shutdown?), but the presence of such a network is thought to destroy clear and laminated facies basal ice. This lends support to this second model, which assumes that channelised drainage takes up an englacial route, leaving little but the residual film, vein network or till pore-water flow at the bed. Large chunks of basal ice - within which transport of coarse, angular clasts occurs - must exist apart from major drainage routes for some distance upglacier of the ice margins. However, bedrock fracture and tractive entrainment responsible for the inclusion into basal ice of abundant large clasts implies that cavity drainage is important at some stage.

The appearance of the basal ice matches closely what is expected of a regelation-metamorphic origin, but this alone cannot explain the basal ice thickness. The closed system recycling of regelation processes prevents thick sequences of basal ice building up, unless tectonic thickening occurs. However, this is to be expected given the inferred ice flow fields at the glacier termini (working backwards, tectonic thickening suggests a widespread absence of meltwater at the bed - see next chapter - and supports my tentative conclusions about which style of drainage at the termini is likely). Extensive deformation with intense bed-parallel shear explains the absence of clear deformation structures, as at Variegated Glacier (6.3, above); it also accounts for what appears to be a thorough mixing of clear and laminated facies so that the two ice types cannot easily be separated. High rates of erosion provide large quantities of debris for incorporation into basal ice as ice slides through the ice-fall; downglacier, within the overdeepening, rapid tectonic thickening with little contact between meltwater and basal ice ensures the preservation of a large proportion of this basal ice, thereby sustaining the high rate of debris flux *within ice* to the ice margins. The full details of this scheme, which hinges on the presence of the pronounced ice-fall and overdeepening, are given in the next chapter.

CHAPTER 6: SUMMARY

- Thick exposures of debris-rich basal ice are found in large quantities at Gígjökull and Steinholtsjökull. These give rise to extensive accumulations of ice-marginal sediments.

- The majority of published basal ice studies present insufficient information to provide a satisfactory account of ice-marginal sedimentation out of basal ice. Debris transfer takes place at the catchment-scale, so a catchment-scale perspective is needed. Basal ice sediment transfer must be viewed as a series of process linkages which explain 1) the availability of debris at the ice-bedrock interface; 2) the initial entrainment of debris; 3a) the redistribution of debris; and, 3b) the preservation of basal ice. The behaviour of water at the glacier bed is likely to exert a major influence on all links in this chain. Few published studies recognise the crucial importance of hydrology. The work by a) Hubbard and Sharp, and, b) Lawson and co-workers, represents the major exception. I use this as the foundations on which to build my analysis of basal ice sediment transfer at Gígjökull and Steinholtsjökull.

- The basal ice exposures at Gígjökull and Steinholtsjökull are consistent with two origins: 1) refreezing of water within the debris-laden channels of a collapsing distributed drainage network to give Lawson-type 'stratified' basal ice; or, 2) tectonic thickening of a stacked sequence of clear and laminated ice facies, initially formed by basal sliding over the rough beds of the glaciers' interiors. The quantity of coarse, angular debris contained in the basal ice possibly favours this second interpretation.

- Whichever model is the best, the control exerted by the ice-fall/overdeepening on the inter-related processes of ice, water and debris transfer is crucial. Chapter 7 develops this idea of the ice-fall and overdeepening as a specific kind of hydrological-geomorphological system particularly conducive to high rates of ice-marginal sedimentation.

CHAPTER 7

Gígjökull and Steinholtsjökull: Sediment transfer in ice-falls and overdeepenings

INTRODUCTION

In Chapter 2 I used the example of Sólheimajökull to develop the idea that water tends to control the subglacial sediment transfer system of a typical valley glacier. Flushing of debris seems to represent the most powerful of alternative transport pathways, with the effect that - especially in the absence of major supraglacial debris inputs - erosive glaciers need not build up major ice-marginal moraines. However, major proglacial spreads of water-laid deposits can be expected. Gígjökull and Steinholtsökull are an exception. Although the exact quantities of debris evacuated by meltwater at these glaciers are unknown, it is clear that an unusually high proportion of the total debris flux is retained within the ice, and so adds to ice-marginal moraines. These moraines are fed largely from englacial debris bands and basal ice, neither of which are abundant at Sólheimajökull. Some factor(s) must exist which switches the ice-water-debris transfer system at Gígjökull and Steinholtsjökull into a different state to that at Sólheimajökull. This factor must permit subglacial erosion to provide abundant quantities of debris (as is inferred for Sólheimajökull) but it must also break the flushing constraint which limits debris retention by ice. It is this combination of high rates of debris production with high rates of debris retention (which I stress once again are two very different things) which is likely to create near-optimum conditions for moraine formation.

In Chapter 5 I argued specifically that the englacial debris bands relate to the special case of drainage through an overdeepening. Below I wish to extend the logic of this to consider the possible impact of bedrock topography and drainage behaviour on the pattern of basal ice transfer. I enlarge on basal ice Model 2 (Chapter 6.4) to show that this works also largely because of the presence of the overdeepenings at Gígjökull and Steinholtsjökull. Indeed, it seems that the relationship between the debris bands and the basal ice is causal, rather than coincidental. This highlights the importance of the ice-fall/terminal overdeepening as a distinct coupled sediment transfer system likely to favour ice-marginal sedimentation.[1] I use basal ice

[1] Sólheimajökull has no real ice-fall, but ice radar survey shows evidence of an overdeepening (Andrew Mackintosh, unpublished data). However, this seems to be less pronounced than the

(continues over page)

Model 2 here because of a combination of doubts over the relationship between water transport and large angular clasts, and gut instinct. It makes little difference to my ideas if basal ice Model 1 is, in fact, the better match with reality: Lawson-type 'stratified' ice depends even more strongly on the existence of a terminal overdeepening (Chapter 6.2 and 6.4).

The first part of this chapter (7.1) largely follows Hooke (1991). This paper describes what is likely to happen, and why, at an ice-fall/entrance to an overdeepening. In the second part (7.2) I extend this argument to consider what kind of behaviour is to be expected at the exit of a terminal overdeepening, and how this relates to the pattern of sedimentation observed at Gígjökull and Steinholtsjökull.

My approach

As with Hooke's paper, my argument rests on what I trust is intelligent guesswork, constrained by field observations; to paraphrase Richards (1996), my key procedure is explanation achieved by extrapolation of the understanding of process. Confidence in the quality of the explanation lies not in any formal hypothesis test, or 'verification' by computer model, but with the internal consistency of the theoretical package, or, "explanation confirmed by coincidence" (Richards, 1996, p. 180). If several different elements tie together without contradiction, it increases the chance that the reasoning is correct. This strategy represents a pragmatic response to the different conditions under which geological science - as opposed to laboratory science - operates (Frodeman, 1995; see Chapter 9); it also helps to emphasise the underlying level of process (see Chapter 1.4). Richards uses the example of the Haut Glacier d'Arolla hydrology project: process inferences drawn from a range of complementary techniques converge on a single time-space picture of the glacier's drainage system. The quality and quantity of data produced by field studies of glacial geomorphology is different to that produced by hydrology studies, however; here I try to build an interpretative framework which combines observations of form, rather than measurements which relate directly to process.

It is important to stress that - as the catchment-scale of analysis requires - it is not any single process which provides the core of the explanation, but the combination of processes, and the particular way in which they come together, which is crucial. This is a move away from

overdeepenings at Gígjökull (certainly) and Steinholtsjökull (possibly); the drainage system is likely to be different when it encounters the adverse slope; and - of greatest importance - the overdeepening ends 1 km upglacier of the snout, so that its impact on water and debris transfer (if any) can be undone.

reductionist to configurational investigation. This is a scientific tactic which assumes that we do have adequate knowledge of the immanent (the basic process building blocks), but this alone is unsatisfactory: successful interpretation/explanation requires we investigate how these building blocks interact in time and space (Lane and Richards, 1997; see also Kerr, 1997 and Chapter 9 for further thoughts on the relationship between individual processes and trajectories of landform development). Specific circumstances give rise to certain quasi-stable combinations of process, which give rise to specific - and often self-reinforcing - styles of behaviour: i.e. in this case the particular pattern of sedimentation which seems to be favoured by the existence of an ice-fall and terminal overdeepening.

7.1 THE ICE-FALL

RAPID BEDROCK EROSION

Hooke's analysis starts with the inference that the steep trough floor which underlies an ice-fall (equivalent to the headwall of a corrie) is likely to be subject to intense attack by plucking. This enhanced rate of subglacial erosion excavates the overdeepening. Iverson's work (1991b) suggests that effective plucking involves cyclical loading and unloading of rock steps associated with water pressure fluctuations. Cavity formation is favoured by fast sliding over a rough bed; this is expected at Gígjökull as ice leaves the crater to enter the ice-fall, converges as it is squeezed between the trough walls, thins, and accelerates. Measurements imply summer sliding velocities in excess of 1.5 m d^{-1} (Chapter 4.1). The stepped geology of Eyjafjöll possibly also favours cavity formation. Calculations of basal water storage, released by floods, supported by observations of bedrock exposed by ice retreat, suggest that South Tahoma Glacier, Mount Rainier, Washington State, USA is underlain by an unusually extensive number of cavities. This is thought to reflect the tendency of the inter-stratified lavas and pyroclastic rocks which make up the steep slopes of Mount Rainier (20^{o} to 30^{o}) to erode into ledge-like structures (Walder and Driedger, 1995). This description seems to fit Eyjafjöll also. Weak lithology is further likely to enhance plucking. Whereas the tensile strength of basalt lavas is relatively high (~150-350 MPa) the tensile strength of pyroclastic rocks - which make up the zone cut by Gígjökull and Steinholtsjökull's ice-falls (field observations; Geological Map of Iceland, Sheet 6: South Iceland, 1:250,000, 1990) ranks among the lowest of any consolidated rock (<~75 MPa; Drewry, 1986, p. 34).

Crevasse water input to cavities. Surface meltwater is rapidly channelled to basal cavities beneath ice-falls. Surface melt cannot travel far before it is captured by the dense network of crevasses formed by surface stretching as the ice accelerates. These crevasses enter into moulins which rapidly connect to the glacier bed. Cavities tend to collect water which drains towards the bed because they usually represent local zones of low pressure; indeed Röthlisberger and Lang (1987, p. 254) suggest that a moulin/crack which propagates towards the glacier bed will be drawn towards bed irregularities such as a rock step/cavity because of the local perturbation of the stress field. Cavity water pressure fluctuations replicate the diurnal melt cycle because rapid delivery provides little opportunity for dampening-out of the meltwater input signal. Changes to water inputs/water pressures will change sliding velocities, cavity interconnection and bed drainage conditions (Iken and Bindschadler, 1986). Changes in sliding (i.e. the extent and intensity of basal decoupling) driven by active cavities will be transmitted unevenly across large parts of the ice-fall (Röthlisberger and Iken, 1981; Hanson and Hooke, 1994). These things mean that the time-space pattern of meltwater inputs typical of ice-falls is likely to produce just the kind of step loading-unloading cycles which accelerate both rock fracture and removal. Such variable conditions are likely to be repeated regularly throughout the summer, year after year.

Basal ice formation

Cyclical cavity growth and closure favours entrainment of plucked debris both by freezing-on and tractive entrainment by plastic flow. Elevated levels of basal debris in turn are likely to raise debris production by abrasion as basal ice moves across the series of bedrock obstacles and cavities which constitute the rough floor of the ice-fall (Iverson, 1995). The wider flow field (ice mass continuity constraints), variable water pressures, and longitudinal stress transmissions will all tend to act to speed up sliding. Enhanced plastic deformation and rapid regelation are required in part to achieve these higher sliding speeds; the likely effect is both to raise the rate of abrasion further, and to increase the rate at which clear and laminated facies basal ice forms and reforms. The overdeepening owes its origin to sustained erosive activity: beneath the ice-fall parcels of basal ice are rapidly and continuously created and evacuated.

BASAL ICE-WATER FLUX RELATIONSHIPS

Established theory offers little firm guidance here, beyond the suite of features - film, cavity, orifice, small conduit, large conduit... - likely to cooperate to drain the ice-fall. Exactly what elements interact, in what way, is entirely conjectural. However, the pattern of ice-marginal sedimentation seen at Gígjökull and Steinholtsjökull provides two important clues:

- A substantial fraction of debris must enter relatively efficient, high-energy components of the drainage system, thereafter to form the relict conduit debris bands.

- A substantial fraction of debris must survive its journey through the ice-fall and overdeepening as parcels of basal ice.

This means that certain ice-debris flow paths exist which are likely to encounter destructive elements of the basal drainage system, whereas contact between other ice-debris flow paths and major elements of basal drainage must be limited.

The floor of the overdeepening represents 'safety' for parcels of basal ice (7.2, below). If a parcel of debris-rich basal ice makes it this far without capture, I expect it to contribute to ice-marginal sedimentation. Remember: the flux of basal ice must be sustained, but the basal ice layer need not be excessively thick (Box 6.1), nor must it be perfectly continuous. The loss of the odd parcel of basal ice, or part of the odd parcel makes little difference: compressive flow processes will thicken up the basal ice layer and erase any gaps. The analogy of a race in which basal ice parcels must evade capture by water is useful, because it provides some sort of explanation without needing to specify anything in detail about the nature of drainage beneath the ice-fall. If the 'predatory capacity' of basal drainage is assumed fixed and equal everywhere (it need not be known!), the probability that any given parcel of basal ice will make it to the safety of the 'finish line' (i.e. the floor of the overdeepening) will be a function of:

1. The volume of debris-rich parcels in transit.
2. The distance that the parcel must travel to reach the finish line.
3. The speed at which the parcel can travel this distance.

Flushing is likely to dominate if ice is debris-poor, and travelling slowly over a long distance; however, if the situation involves large quantities of debris-rich ice making a short, fast dash for safety some high proportion can be expected to survive. These are the conditions I expect to apply at Gígjökull and Steinholtsjökull: e.g. the distance between the middle of Gígjökull's ice-fall and the floor of its overdeepening is just ~1.5 km, sliding speeds are rapid, and basal ice debris-content is high.

Ice-falls, drainage and sediment yields

Gurnell (1995) presents Swiss data which give a useful insight into the relationship between ice-fall drainage and debris flushing. She contrasts the suspended sediment yield of the Haut Glacier d'Arolla with that of the neighbouring Bas Glacier d'Arolla and Glacier de Tsidjiore Nouve. Whereas early summer suspended sediment discharge is roughly similar for all three glaciers, late summer yields for the Haut Glacier d'Arolla are substantially higher than those of the other two. This difference cannot be explained by contrast in glacier size, aspect, altitude or bed material: all are similar. The significant factor seems to be that, whereas the Bas Glacier d'Arolla and Glacier de Tsidjiore Nouve have large ice-falls, the Haut Glacier d'Arolla does not. Intensive tracer, chemical and water balance investigations provide a detailed picture of the seasonal evolution of drainage (driven by upglacier retreat of its snowline) at the Haut Glacier d'Arolla (Richards *et al.*, 1996; Peter Nienow, unpublished data); high late-summer sediment flux appears to reflect the establishment glacier-wide of an integrated trunk conduit system (Clifford *et al.*, 1995; see also Chapter 2.5). Gurnell infers that such an integrated conduit network is not established at the Bas Glacier d'Arolla and Glacier de Tsidjiore Nouve because of the disruptive effect their ice-falls exert on drainage. She does not enlarge as to what exactly this disruption is; the implication seems to be that drainage occurs largely within a poorly-connected series of linked cavities and smaller conduits. Field data to confirm this are not available, although fast sliding in ice-falls will tend to suppress enlargement of incipient conduits, and so protect the cavity system against collapse (Walder, 1986; Kamb, 1987; Chapter 1.1).

This picture of drainage achieved by a mixture of linked cavities and small conduits is similar to that envisioned at Findelengletscher (Collins, 1979b; Iken and Bindschadler, 1986), South Tahoma Glacier (Walder and Driedger, 1995) and Storglaciären (Hooke, 1991, and personal communication, Reykjavík, 1995). It seems to fit Gígjökull and Steinholtsjökull also. Several crevasses/moulins are likely to connect to each large cavity; in turn, several cavities are expected to feed each small conduit. The exact relationship between cavity and conduit drainage is unclear, in part because of the tendency to set up cavities versus conduits as a distinct dichotomy. There seems to be no firm answer to the question: when does an orifice draining a cavity qualify for conduit status? However, because the number of subglacial conduits is likely to be tied to the number of feeder crevasses (i.e. lots!), the total number of conduits expected is large, with correspondingly small discharges.

Cavity drainage

The cavity-rock step system is where basal ice parcels are likely first to form, and thereafter to reform, as ice moves downglacier. Where and when active, drainage through cavities themselves is likely to be sluggish; it is the rapid dissipation of hydraulic head within the orifices which connect cavities which is likely to capture the debris of basal ice (Chapter 2.5). Because the orifices tend to be orientated transverse to ice flow, the frequency with which orifices are likely to intersect basal ice flow is high. The example of Glacier de Tsanfleuron (Sharp *et al.*, 1989a; Hubbard and Sharp, 1995; Chapter 6.3) shows that the linked cavity network can be extremely effective at destruction of basal ice; elsewhere, however, it is evident that cavity drainage presents less of a threat to the survival of basal ice: e.g. Variegated Glacier (Sharp *et al.*, 1994). Fast sliding and high levels of debris production (see above) will tend to neutralise the destructive potential of the cavity network (Tsanfleuron is thought to slide at ~0.2 m d^{-1}, whereas Gígjökull slides >1 m d^{-1} in summer). If cavities and orifices cover only a fraction of the glacier bed; if the quantity of debris present is high; if sliding is fast; if the volume and flow speed of water within the system fluctuates (at low levels, a significant proportion will be confined by rock, not ice); and if the number of active cavities/degree of connectivity changes regularly, then the probability that a large proportion of basal ice will survive transit through the cavity zone is high. All these conditions are thought to hold beneath the ice-falls of Gígjökull and Steinholtsjökull.

Conduit drainage

Several points can be made which suggest that debris capture by conduit flow is limited (see also Chapter 2.5 and Box 2.3):

- Discharge carried within a relatively large number of conduits (100? 1,000?) occupies only a small percentage of the glacier bed (see Box 7.1). Wide zones of distributed drainage are likely to exist between conduits, within which basal ice is relatively safe (a linked-cavity network may occupy some of this inter-conduit area, however).
- Thin ice means conduit closure rates are low, so the flux of debris-rich basal ice into conduits is likely to be low also.
- Wall melt rates are calculated to exceed closure rates by two orders of magnitude (using equations 6 and 8 of Hooke, 1984) (see below). This implies open flow conditions, which in turn implies that surcharging of conduits, with flood excursions sweeping across large areas of the glacier bed, (cf. Collins, 1996; Chapter 2.5) are likely to be rare.
- The steeply-sloping bed, with fast sliding, is likely to discourage conduits from swinging across basal ice flow lines. This preserves the wide inter-conduit zones in which basal ice is safe. This point merits further discussion.

Migration of conduits

Water flow under atmospheric pressure only is controlled by bedrock topography, so open conduits are expected to migrate across the glacier bed as they melt themselves a pathway downslope (Hooke, 1984). This tendency will not threaten basal ice preservation if the sub-ice-fall topography is an inclined plane, sloping evenly and directly towards the glacier snout. If so, conduits will run directly towards the snout, and so preserve the inter-conduits areas in which destructive water flow is limited. Water flow which cuts across the trend of ice flow, however, threatens the survival of basal ice. This will be the case if the bedrock falls towards the glacier centre. If it exists, conduits will tend to slip into this central valley. Calculations of excess melt rates over closure rates [after Sharp *et al.* (1993), but using values thought typical of Gígjökull (Box 7.1)] indicate that, *ceteris paribus*, conduit migration below Gígjökull's ice-fall is likely to be pronounced: e.g. conduits are calculated to shift their location four times faster than at the Haut Glacier d'Arolla; the ice at Gígjökull is thinner, and the bedrock slopes which drive stream flow are steeper. Given 100 days' activity, a conduit which typically carries 0.06 m^3 s^{-1} is calculated to shift 16 m; 100 of these conduits (see Box 7.1.) will therefore shift a total of 1,600 m, and so sweep the entire width of the bed (the initial distance between 100 conduits which begin evenly-spaced is just 7.78 m). Six large conduits, each carrying 1.0 m^3 s^{-1} are less effective, but in total are still expected to sweep over half the width of the bed (i.e. 528 m / 800 m).

However, this tendency must be offset against the sliding action of ice. Water can cut across the direction of ice flow only if it can melt ice faster than sliding carries ice into the conduit (wall melt rates must balance both conduit closure rate *and* ice advection: Hooke, 1984). The process envisaged by Hooke is similar to that suggested by Kamb (1987), whereby advection of ice counteracts the tendency for melt to enlarge orifices. Rapid sliding will tend to turn any transverse conduit parallel to the direction of ice flow. If sliding is rapid relative to excess melt rates, conduits will tend to swing to be pinned in the direction of sliding, and migration is severely curtailed. Sliding speeds beneath Gígjökull's ice-fall are likely to exceed excess melt rates by some significant factor: ~10 in the case of the hypothetical 0.06 m^3 s^{-1} conduit. The effect this has on conduits' ability to shift position is likely to preserve large parcels of basal ice which exist between conduits.

[N.B. The calculations of conduit migration used here must represent conservative estimates. Hooke rewrites the basic wall melt formula, which predicts *mean* melt rates, for the arc (i.e. ice perimeter) of a semi-circular, subglacial conduit, in a way which solves simultaneously for suitable values of discharge, slope and channel dimensions (i.e. his equation 6, as used by Sharp *et al.*, 1993). However, as Hooke recognises, wall melt need not be distributed equally across the channel ice perimeter. Preferential melt will give rise to enhanced channel migration; at the limit, Hooke suggests wall melt/migration rates can be twice those expected if wall melt is evenly distributed. E.g. if this is true, the 0.06 m^3 s^{-1} will migrate 32 m in 100 days (ignoring the appropriate correction for sliding).]

Box 7.1

Likely water flow conditions in Gígjökull's lower ice-fall

This box uses the work of Hooke (1984) to estimate certain properties of conduit flow beneath Gígjökull's lower ice-fall:

1. Whether flow within conduits is likely to be open or closed?
2. Likely size of conduits?
3. Likely number, spacing and bed coverage of conduits?

These calculations are crude approximations at best. Hooke's method works correctly only under ideal conditions. Factors which these calculations ignore include (Hooke, 1984; Hooke *et al.*, 1990):

- Conduit shapes which deviate from the ideal semi-circular cross-section (broad, low conduits close by plastic creep more easily).
- Non-uniform flow (energy released is used to speed flow up, not to melt conduit walls).
- Non-instantaneous transfer of frictional heat (heat is carried downstream before it is released to melt walls).
- Heat loss to ice walls by conduction because of discrepancy in the local pressure melting point of ice and water (water pressure < ice pressure).
- Changes in ice viscosity because of changes in the basal stress field.
- Enhanced closure of tunnels in the vicinity of bedrock obstacles.
- Channel roughness incorrectly specified (Manning's n = 0.05 $m^{-0.33}$ s^{-1} is used here).
- These relationships strictly apply to ice-walled tunnels only. Very small channels (elements of distributed drainage), and/or drainage controlled in part by bedrock topography are not likely to be described adequately by the Röthlisberger-Hooke equations.

Nevertheless, I believe that these calculations provide important clues as to what happens with drainage beneath Gígjökull's ice-fall.

OPEN CHANNEL FLOW

This is expected if the mean rate of conduit wall melt exceeds the mean rate of conduit closure (see Chapter 1.1). The critical discharge at which this occurs (**Q_{crit}**) is given by (Hooke, 1984, equation 9):

$$\mathbf{Q_{crit} = [\, h^3 \,/\, (\, c \sin^{7/5} b \,)\,]^5}$$

h ice thickness
b glacier bed slope
c constant = 6.55×10^8 $m^{12/5}$ $s^{1/5}$

(PTO)

Box 7.1 (continued)

I use the following values, thought to be representative of Gígjökull's lower ice-fall:

- Ice thickness (h) = 100 m
- Mean bed slope (b) = 20°
- Typical total daily peak flow through lower ice-fall = 6 $m^3 s^{-1}$
- Width of lower ice-fall = 600 m

These values give $\mathbf{Q_{crit} = \sim 1.5 \times 10^{-11}\ m^3\ s^{-1}}$

If flow occurs under pressure, the number of channels required to drain the lower ice-fall is:

$6\ m^3 s^{-1} / 1.5 \times 10^{-11}\ m^3 s^{-1}$ (i.e. Q_{crit}) = $\mathbf{4 \times 10^{11}}$

The diameter of a single, semi-circular channel which carries $1.5 \times 10^{-11}\ m^3\ s^{-1}$ down a 20° incline is 0.000032 m [from Hooke's modification of the Darcy-Weisbach pipe-flow equation (Hooke, 1984, equations 1, 2 and 3)]. 4×10^{11} of these channels placed side-by-side occupy 12,800 km - over 20,000 times the space available! (i.e. width of ice-fall = 600m). This implies two important things:

1. **The bulk of flow beneath the lower ice-fall is likely to be under atmospheric pressure (i.e. open flow).**
2. **The bulk of flow beneath the lower ice-fall is likely to run in a relatively small number of relatively large conduits.**

Despite the many uncertainties (see above), these conclusions are likely to be robust.

PLAUSIBLE FLOW SCENARIOS?

The following give some indication of the channel spacing and bed coverage likely to prevail beneath Gígjökull's ice-fall.

- **A: 100 channels, each carries 0.06 m^3 s^{-1}**

  This gives channel diameter = 0.22 m; drainage occupies (0.22 m x 100) = 22 m of given bed cross-section = 3.67% of available bed space.

- **B: 1,000 channels, each carries 0.006 m^3 s^{-1}**

  This gives channel diameter = 0.088m; drainage occupies (0.088 m x 1,000) = 88 m of given bed cross-section = 14.67% of available bed space.

N.B. Approximately 25,000 channels, each of which is 0.024 m in diameter, and carries $2.4 \times 10^{-4}\ m^3\ s^{-1}$ are required to cover the entire bed.

IMPLICATION

- Significant elements of drainage likely to pose a major threat to the integrity of basal ice cover only a limited fraction of the glacier bed.

Under closed flow conditions, drainage integration is promoted by the presence of a master conduit at lowest pressure, upon which others converge (Chapter 1.1); under open flow conditions, a dendritic network is likely to form as flows slip downslope to converge on a central conduit located at the base of the subglacial trough. Beneath ice-falls closed flow is unlikely (Box 7.1), and fast sliding counteracts the tendency of channels to shift down bedrock slopes. These factors are likely to inhibit drainage integration. This is potentially important if Gurnell's suggestion that drainage integration promotes sediment evacuation is correct.

Evaluation

The above reasoning suggests that both cavity and conduit drainage are likely to be relatively inefficient at capturing and flushing debris beneath ice-falls such as at Gígjökull and Steinholtsjökull: large quantities of basal ice are expected to reach the safety of the overdeepenings' floors. However, it is important to reiterate: large quantities of debris *will* be captured by drainage, as the debris bands show. Large volumes of debris-rich basal ice created by fast sliding and rapid erosion permit relatively high losses to be sustained.

7.2 SEDIMENT TRANSPORT IN A TERMINAL OVERDEEPENING

Basic idea

Between the entrance to the overdeepening (i.e. its ice-fall) and its exit a major change in ice and water flow takes place: fast sliding with copious subglacial drainage is replaced by subdued sliding, intense thickening of ice, diminished basal drainage, and enhanced englacial drainage. These changes have a large influence on sediment transfer, as the pattern of moraine formation at Gígjökull and Steinholtsjökull shows. I infer that the behaviour of ice and water in overdeepenings is not conducive to debris flushing. Two factors seem to be of special importance: a) the switch from subglacial to englacial drainage; and, b) tectonic thickening of the ice, notably its debris-rich basal layers. These two factors bring about separation of water and sediments: debris-rich basal ice survives intact in large quantities, whereas sediment carried by water into high-level transport is subsequently abandoned within englacial ice to form relict conduit debris bands. These conditions take effect within a few hundred metres of the ice margin, which means that the system cannot return to its previous state (= aggressive subglacial drainage?). This is important: englacial drainage and compressive flow are stable features of glacier ice behind an adverse slope, but are likely to give way once more to subglacial drainage and extending flow once ice clears the *riegel*. This will not happen if the *riegel* coincides with the ice margin, in which case the change in sediment transport pathways is likely to have a pronounced impact on the build-up of moraines, as here at Gígjökull and Steinholtsjökull. However, if the overdeepening lies some distance upglacier of the snout (e.g. as at Sólheimajökull?), the effects discussed below are likely to be undone. In this section, I explore the processes which are likely to explain the distinct pattern of sedimentation associated with the terminal overdeepenings at Gígjökull and Steinholtsjökull. Similar behaviour can be expected at other glaciers characterised by comparable inputs of water and debris to terminal overdeepenings; Kvíárjökull, Öræfi, provides once such example (see Chapter 8.2). In particular I wish to stress the strong inter-relationships which exist between different elements of this transport regime, including the possibility that the disposition of water implied by the debris bands controls the disposition of basal ice. What follows is plausible, and internally consistent given the current state of knowledge; it has not been - and perhaps cannot be - verified/falsified in any formal sense.

ENGLACIAL DEBRIS BANDS: HYDROLOGICAL IMPLICATIONS

The relict conduit debris bands (Chapter 5) show that: 1) as water pressures rise, a large proportion of drainage must take up an englacial route; 2) that large quantities of debris must be carried into high-level transport by this switch; and, 3) that a large part of this debris is subsequently dumped out of water and entrained by ice. The debris bands so formed make a key contribution to high levels of ice-marginal sedimentation. However, the presence of these debris bands implies also that perhaps the bulk of drainage in the overdeepenings occurs some distance above the glacier beds. If so, subglacial flushing potential should fall sharply: debris and basal ice at the glacier beds are not in contact with this now englacial water. What water does remain at the beds (i.e. residual drainage) will be less likely/less able to attack and evacuate basal sediments. The inference that the bulk of water leaves the bed is supported by the example of Storglaciären (see Chapter 5.6), although it cannot be confirmed for Gígjökull and Steinholtsjökull by simple field observations, nor by appeal to theory (other plausible alternatives exist). However, the following points provide support for the idea that drainage becomes predominantly englacial:

- If large quantities of water stay at the bed (within a drainage network which theory predicts must begin to collapse: Chapter 5.6), why does basal ice contain no rounded debris? (See Chapter 6.2 and 6.4)

- Is a cavity network capable of transmitting discharges of several cumecs a feasible prospect at the exit to an overdeepening? Reduced ice flow rates and enhanced back-stress as ice presses against the *riegel* are expected to diminish cavity formation, and cavity drainage also (Fountain, 1994; Hooke and Pohjola, 1994).

- The survival of large quantities of basal ice supports the presence of residual subglacial drainage only (cf. Chapter 2.5; Boxes 2.2 and 2.3).

TECTONIC THICKENING OF ICE

This is the second key factor which combines with paucity of subglacial water flow to reduce flushing and enhance moraine accumulation. Tectonic thickening builds up the thick sequences of debris-rich basal ice exposed at the ice margins of Gígjökull and Steinholtsjökull. Reduced marginal ice flow speeds (e.g. Fig. 4.6) mean that this will not necessarily raise the debris flux; tectonic thickening is important because it improves the survival chances of basal ice by carrying it away from the 'danger zone' at the bed, where destructive stresses and contact with meltwater are highest (Chapter 6.3). Thickening of basal ice reflects longitudinal compression of ice because of:

1. The downglacier fall in flow speeds with distance towards the snout (Figure 4.7).
2. The wider flow field of the ice-fall and overdeepening. Ice dives out of the ice-fall, and piles up behind the *riegel*. This effect is similar - if less spectacular - to what happened immediately behind the surge front of Variegated Glacier in 1982-1983 (Sharp *et al.*, 1994).
3. Change in the quantity and configuration of basal drainage.
4. Change in the flow resistance imparted by basal debris drag (this clearly presupposes the survival of extensive patches of basal ice).

3 and 4 are speculative, but are consistent with current theory. Both involve a change in basal boundary conditions, in this case from 'easy', low flow resistance to 'hard', high flow resistance conditions. As the ability of the glacier bed to support a basal shear stress rises, basal sliding is suppressed, and ice deformation enhanced. The junction between cavity and conduit systems inferred at Variegated Glacier, which fixed the position of the surge front (Kamb *et al.*, 1985) is a large-scale example of this. Tison *et al.* (1989) report a similar example at a smaller scale from Glacier de Tsidjiore Nouve: an abrupt increase in tectonic deformation of ice was believed to reflect a switch from relatively impermeable sediments which favour film drainage to porous, better-drained sediments which increase bed roughness.

Change in basal drainage

The quantity of water at the glacier bed must fall, perhaps to a fraction of what it was at the entrance to the overdeepening (see above). Simultaneously, the propensity for cavities to form is expected to be suppressed. If - as is widely held - either the volume of water at the bed (e.g.

Weertman and Birchfield, 1983a; Humphrey and Raymond, 1994) and/or the water pressure within widespread cavities (e.g. Iken and Bindschadler, 1986) leads to faster ice sliding (Chapter 1.1), a sharp fall in bed 'lubrication' (for want of a better word) is likely to suppress sliding, steepen the negative downglacier longitudinal stress gradient, and so increase the intensity of ice compression.[2]

Greater debris drag

Hallet (1981) suggests that the extra component of flow resistance associated with contact between debris in basal ice and the glacier bed will reduce ice flow speeds. Schweizer and Iken (1992) investigate this effect with a computer model of debris-laden basal ice sliding over a rough bed. Basal ice concentrations measured at Gígjökull and Steinholtsjökull (Table 4.4) indicate that it is the Hallet friction model which is appropriate (i.e. debris concentrations are high, but not so high that ice cannot flow freely around clasts). Hallet friction (Hallet, 1979a) states that friction will rise with a) greater clast concentration, and, b) greater bed-normal ice velocity. It is independent of basal water pressures.[3] Schweizer and Iken (1992, p. 84) suggest that intense strain such as that associated with ice flow through ice-falls will increase the concentration of clasts at the glacier bed (providing of course that they are not flushed out!). The bed-normal component of ice velocity varies with the sine of obstacle stoss-face angle, so the adverse slope of the *riegel* acts as a 'mega-obstacle' which creates a widespread zone of high clast contact force/debris drag.[4] Schweizer and Iken's model shows that high basal debris contact forces a) reduce sliding velocities, and, b) push back the point at which water pressures exert an important influence on sliding (i.e. the 'separation pressure', at which cavities can form, rises). Thus it is not implausible that greater debris drag contributes to the tectonic thickening process.

[2] Pressurised (i.e. near-overburden?) water flow in a film represents a plausible drainage scenario, but, if so, its impact on sliding is less likely to be important, because it is assumed that the film pinches out in the vicinity of the bedrock obstacles which control resistance to sliding (e.g. Walder, 1986). However, the possible impact of pore-water/sheet flow drainage within/above a hypothetical till layer (see below) is entirely unknown.

[3] This assumption is suspect, but any exact solution to the problem of sub-clast water pressures is likely to be appallingly complex (Hindmarsh, 1996).

[4] If this effect is to be sustained, there must be a rapid input of clasts to balance those removed by thickening of basal ice (Schweizer and Iken, 1992, pp. 90-91), which ties in with wider ideas of debris-rich basal ice formation, preservation and debris content increase as it passes through the ice-fall (see also 7.1, above).

Additional factors

Although I identify englacial drainage and tectonic thickening of basal ice as the two key processes which define the overall sediment transport regime of a terminal overdeepening, the following factors may also be important.

Reduced migration of conduits. Within the overdeepening itself, conduit migration is likely to be suppressed, a factor which further favours preservation of basal ice (Chapter 2.5, and 7.1 above). Lower rates of conduit migration obviously follow from the facts that conduits are likely to collapse and/or take up an englacial route(s), but the expected change in hydraulic gradient and water pressures within the overdeepening will reinforce this tendency. Thin ice, steep hydraulic gradients, high conduit discharges and open flow conditions elevate levels of conduit migration, whereas relatively thick ice, gentle hydraulic gradients, low conduit discharges and closed flow conditions will characterise the floor of an overdeepening. Back-pressure effects are likely to extend this zone of curtailed flow migration towards the ice-fall (Hooke, 1984; Sharp *et al.*, 1993).

Subglacial till layer? The possibility that a basal till layer exists presents something of a hypothetical 'spanner in the works' (if only because implicitly I have been using a hard-bed model until now). Hooke (1991) predicts that a subglacial till layer will develop against the adverse slope of the *riegel* as the sediment transport capability of water flow collapses. Such a till layer is inferred to exist at South Cascade Glacier (Fountain, 1992, 1993, 1994); its presence at Storglaciären has been confirmed by borehole investigations (Iverson *et al.*, 1995; Hooke *et al.*, 1997). No evidence, either for or against, exists at Gígjökull or Steinholtsjökull. *If* a till layer exists, any, or all of the following effects may operate:

- Deformation of the till layer absorbs stress, and lessens the intensity of ice deformation/thickening of basal ice (cf. Hubbard and Sharp, 1995). However, if till is largely made up of sand and gravel[5] it will be relatively porous, and so less readily deformed.
- Thick, porous till favours relatively high capacity Darcian flow. Water which trickles through sediments is not in contact with basal ice. Dispersed, sluggish flows generate little heat, whilst the sediment barrier is likely to reduce the efficiency of heat transfer to basal ice. This favours its preservation.

[5] As is inferred at South Cascade Glacier - reflecting flushing of fines at the bed? (e.g. Hubbard *et al.*, 1995) or selective removal of fines into englacial flows?

- Under high water input conditions the till layer fills up with water, and high pressure sheet flow in contact with basal ice results. The regularity with which this occurs is likely to reflect till thickness and permeability (as well as climate, weather, and the configuration of the wider drainage system). At Storglaciären, these sheet flow events simultaneously decouple basal ice from till, but give rise to ice surface velocity peaks (Iverson *et al.*, 1995). This implies that sliding ice takes up stress previously borne by subglacial sediments, which creates the possibility that near-bed strain/basal ice thickening is (temporarily) reduced, at the same time as sheet flow-induced basal melt rises.

EVALUATION. The evidence of the ice-marginal sediments at Gígjökull and Steinholtsjökull strongly suggests that the impact of basal till, if relevant, is not a key factor: it must reinforce, or dampen (or both, at different times and places), the influence of the wider factors I discuss above, but it cannot override them. It is important to remember also that the very existence of a subglacial till layer demonstrates that subglacial flushing must be relatively weak (see my discussion of the Suðurhlíðar Neoglacial moraine complex at Steinholtsjökull, Chapter 8.5).

ASYMMETRY OF SEDIMENTATION AT GÍGJÖKULL: A SPATIAL TEST

The asymmetry of ice-marginal sedimentation observed at Gígjökull supports the big picture I sketch above. This asymmetry provides a spatial test which boosts my confidence in my model of sediment transport relationships. The main feature of this asymmetry involves the lateral balance between basal ice and debris band sediments. Basal ice is developed in greater quantities at Gígjökull's eastern side than it is at its western side, whereas the converse is true with the relict conduit debris bands (Fig. 4.9 and Chapter 6.1). A possible explanation lies with lateral asymmetries in both the volume of subglacial water flow, and the degree of ice thickening (i.e. the two key factors I identify above). The bulk of water seems to be captured by the western side of the glacier: the surface valley here (Fig. 4.4) implies the presence of a large englacial or (more likely?) subglacial conduit, which seems to be confined to the west by a longitudinal bedrock ridge. [This ridge pins back the tephra outcrop to the foot of the ice-fall: Jeremy Everest (1994), ice-radar data.] The imbalance between the flow levels of the two ice-marginal streams (i.e. west >> east; Chapter 5.6) supports this also. Strain net data (Table 4.1) show that longitudinal shortening close to the western margin is half that experienced at the eastern margin (probably because ice flow feeds into a belt of relatively fast flow beneath the terminal ramp); lateral stretching takes up 84% of that shortening which does occur. My ideas predict that at this 'wet' side of Gígjökull flushing will be high, and debris preservation low. The limited quantities of basal ice here support this. Conversely, at the 'dry' eastern edge, ice flow deceleration, longitudinal shortening and vertical thickening are intense, and marginal basal ice exposures are extensive. The larger quantities (Fig. 4.9), and greater roundness (Box 5.1) of western debris band sediments relative to eastern debris band sediments arguably reflects this inferred asymmetry in water flow as well. [I find it tempting to conclude that the observed asymmetry in the flow field of the terminal lobe also reflects contrasts in drainage at the glacier bed (cf. Harbor *et al.*, 1997).]

CHAPTER 7: SUMMARY

- As with Sólheimajökull, Gígjökull and Steinholtsjökull are believed to generate large quantities of debris by subglacial erosion. However, unlike Sólheimajökull, at Gígjökull and Steinholtsjökull an unusually high proportion of debris is retained within the ice - as dirty basal ice, or within relict conduit debris bands - eventually to reach the ice margins and so contribute to moraine accumulation.

- These high rates of debris retention indicate that some factor must exist at Gígjökull and Steinholtsjökull which part-neutralises the aggressive tendencies of subglacial drainage. This factor is the presence of ice-falls which feed into terminal overdeepenings.

- Fast sliding through the ice-fall creates debris in quantities sufficiently great that - despite moderate losses to water transport - basal ice carries abundant debris into the overdeepening. Rising water pressures here give rise to debris-laden englacial drainage, which forms the englacial debris bands (Chapter 5). Simultaneously, the loss of basal water, coupled with intense compressive stresses, generates thick sequences of basal ice which ensure high rates of debris delivery to the ice margins. Thus the adverse slope of the terminal *riegel* emerges as the key factor which controls the behaviour of ice and water - and so sediments also - at Gígjökull and Steinholtsjökull.

- The ice-fall/overdeepening gives rise to a robust arrangement of sediment transport pathways, fixed by the interaction of bed topography, ice flow dynamics and changing styles of glacier drainage. This arrangement seems to be particularly conducive to moraine formation. If this transport pathway configuration/entrenched tendency to dump large quantities of debris at the ice edge is sustained, it is likely to have a major influence on the glacial geologic record. I investigate this possibility in the next chapter.

CHAPTER 8
Gígjökull and Steinholtsjökull: Neoglacial moraine accumulation

INTRODUCTION

The relationship between this chapter and Chapters 4, 5, 6 and 7 is much the same as that between Chapters 2 and 3. I take the process model I advance to explain present-day behaviour, and consider the extent to which it applies to the Neoglacial moraine record. However, here I deal with the moraine records of two separate glaciers. These records are very different. This is something of a puzzle, because the present-day pattern of sedimentation at Gígjökull and Steinholtsjökull is pretty much identical. I argue that the divergence of the Neoglacial record indicates a divergence of characteristic flushing efficiency: it is reduced further at Gígjökull, but it is increased at Steinholtsjökull. However, I also identify a local sub-system of limited flushing capability which existed beneath the Suðurhliðar lobe of Little Ice Age Steinholtsjökull. This was associated with a stable process domain which favoured *subglacial* deposition and deformation of till.

8.1 GÍGJÖKULL'S NEOGLACIAL MORAINE RECORD

The immediate forefield of Gígjökull is dominated by a single, large moraine ridge. This is up to 60 m high, and extends beyond the bedrock escarpment for ~1 km, enclosing the terminal lobe and the proglacial lake (Figs 8.1 and 8.2; see also Figs 4.1 and 4.2). The ridge is broken at the lake outflow, but is otherwise continuous. The two outstanding features of this rampart moraine which require explanation are:

- Its sheer size.
- Its relationship with episodes of climate change.

I explore these issues in the context of Dugmore's (1987, 1989) topographic pinning point hypothesis. I draw on the process model advanced in the Chapter 7 to expand, and in part, revise this. I argue that the two features are intimately connected: the size of the moraine reflects the relatively 'restricted' response of Gígjökull to Neoglacial climate change, which means it tends to reoccupy similar positions for substantial intervals of time, whereas this restricted response to climate change reflects the presence of such a big moraine ridge. This

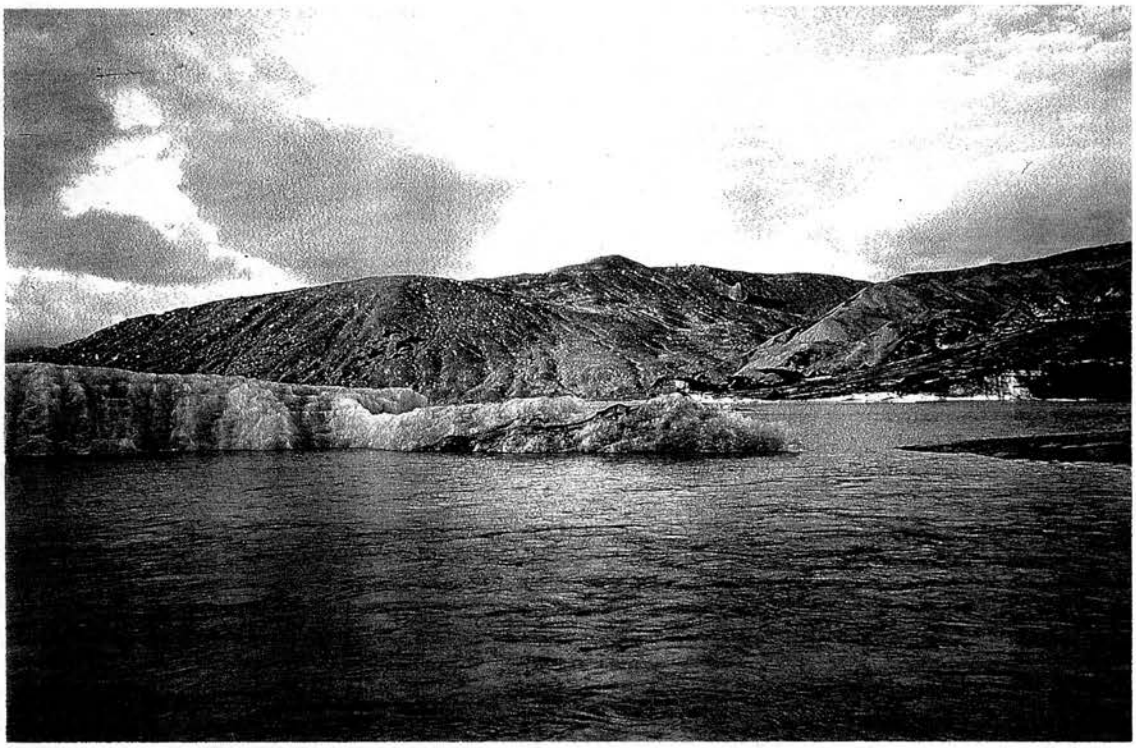


Figure 8.1

View of Gígjökull's eastern moraine rampart.

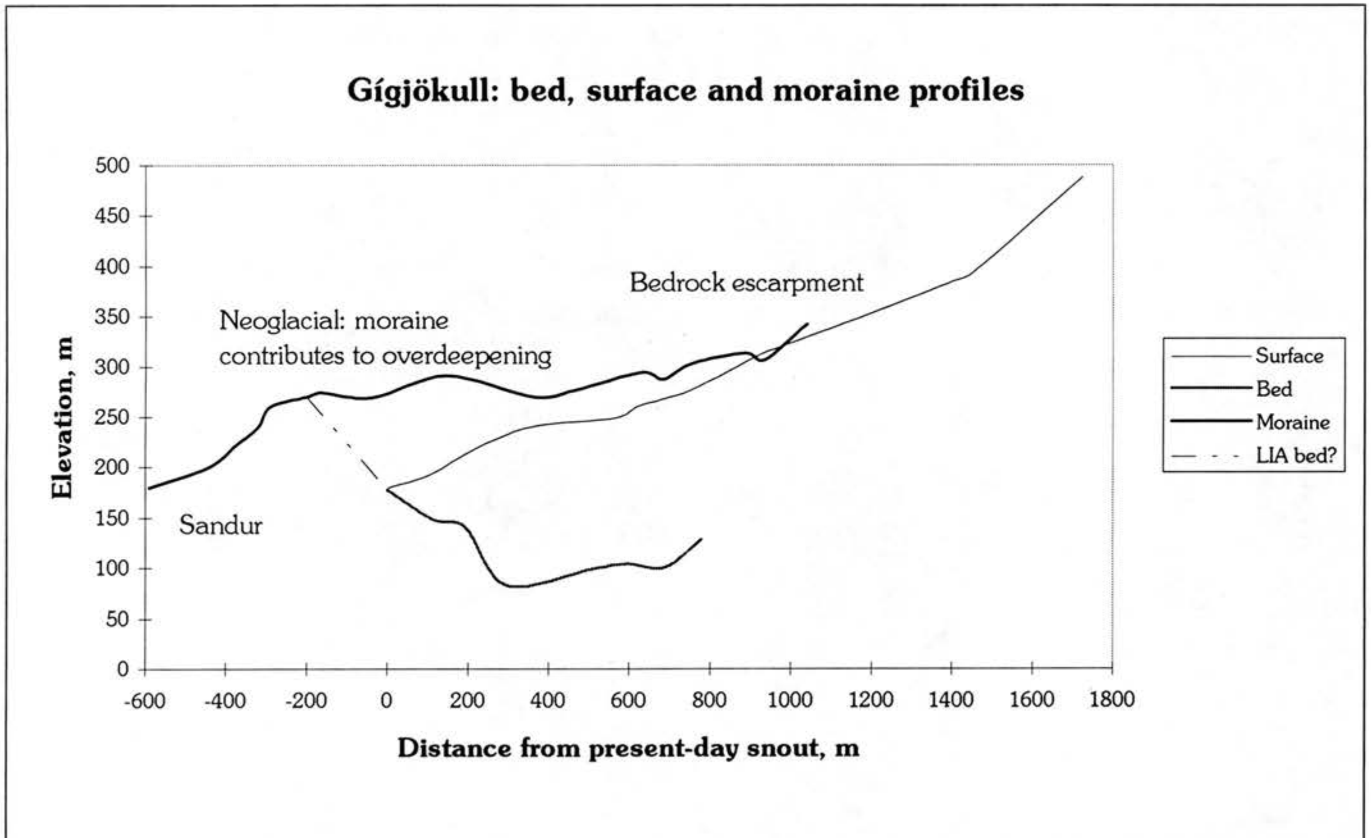


Figure 8.2

Gígjökull: Present-day centre-line glacier geometry set against the profile of the eastern Neoglacial moraine (i.e. ~ice-margin profile). Data from EDM and ice-radar survey. This implies that in the Neoglacial: a) the inner slope of the rampart moraine enhanced the overdeepening, and, b) the terminal lobe was flatter than it is today. See text for discussion.

implies the operation of a positive feedback which links evolving moraine morphology to ice flow and advance behaviour.

MORAINE MORPHOLOGY

It is clear that the rampart moraine is a composite feature; built-up by several episodes of ice-advance and still-stand: several smaller, 'parasitic' ridges superimposed upon the major ridge structure can be identified. This composite origin is confirmed by tephra and lichen dating: i.e. different ridge elements are given different dates (see below). However, as a manifestation of process geomorphology I prefer to treat the ridge as a single monster feature: it is its bulk properties which I think are important here, not the fine detail of individual climate-induced fluctuations.

Moraine size

I estimate the total volume of the moraine with a simple geometric model which represents the ridge as a scalene triangular prism of length 2,000 m, mean height 50 m, inner slope angle 30° (sediment lies at the angle of repose), outer slope angle 20° (slope part-degraded). This makes the total ridge volume (2,000 m x 5,600 m^2) = 1.12 x 10^7 m^3. If it is assumed that a) the contribution to total volume of bedrock or ice-core is negligible (this is thought likely); and, b) the void ratio is 20% (cf. Small *et al.*, 1984), then this makes the total volume of sediment 8.96 x 10^6 m^3. There is no obvious evidence to indicate that the moraine was built up by push-induced recycling of pre-existing proglacial sediments, so I assume that the moraine volume can be read-off in terms of direct subglacial erosion. Thus moraine volume equates to net bedrock erosion of ~1.0 m across the entire catchment of Gígjökull (i.e. roughly 9 x 10^6 m^3 / 9 x 10^6 m^2). If a) subglacial erosion rates at Gígjökull are similar to those at Sólheimajökull, as is likely because the two glaciers are thought to slide at similar speeds over similar bedrock, and, b) Neoglacial rates were similar to present-day rates (~6 mm yr^{-1}?), then this means that the Gígjökull moraine contains the product of ~170 years' erosion. This enables me to calculate a crude estimate of mean Neoglacial flushing efficiency at Gígjökull. The ridge is perhaps as much as 5,000 years old (see below) which implies that it contains ~3.5% of all debris eroded in this time: i.e. mean Neoglacial flushing efficiency is 96.5%. Flushing efficiency falls rapidly if it is assumed that the moraine is younger, and was built up by discontinuous occupation during episodes of ice advance: e.g. it falls to 86% if it is assumed that the moraine is 2,500 years old, and was built up by ice over half this interval.[1] Table 8.1 presents my best-guess estimates, alongside high and low estimates calculated using an arbitrary, cumulative 25% error band.

Table 8.1

Estimates of rates of moraine accumulation and flushing efficiency for Gígjökull's Neoglacial moraine rampart.

NEOGLACIAL MORAINE ACCUMULATION AT GÍGJÖKULL: Error evaluation (±25%)						
Property	**Units**	**LOW**	**A**	**BEST GUESS**	**B**	**HIGH**
Moraine mean height	m	37.5	50	50	50	62.5
Inner slope angle	degrees	37.5	30	30	30	22.5
Outer slope angle	degrees	25	20	20	20	15
Moraine cross-sectional area	m^2	2420	5600	5600	5600	12000
Volume of sediment	m^3	4.8 x 10exp6	8.96 x 10exp7	8.96 x 10exp7	8.96 x 10exp7	1.92 x 10exp7
Depth of erosion equivalent	m	0.43	1.00	1.00	1.00	2.13
Subglacial erosion rate	mm yr^-1	8	8	6	6	4
Moraine volume erosion equivalent	yrs	54	125	170	250	532
Time taken for active deposition to build up moraine	yrs	5000	5000	2500	1000	1000
Mean annual rate of moraine accumulation (MAMA)	m^3 m^-1 yr^-1	0.48	1.12	2.24	5.6	12.0
Flushing efficiency	%	99.0	97.5	93.0	75.0	47.0

It is difficult to provide reliable estimates of rates of moraine formation and flushing efficiency: hence my use of here a generous error margin, which gives rise to the wide range of values set out in Table 8.1 Much of this uncertainty arises because the past rates of subglacial erosion and time taken to build up the moraine are unknown (Table 8.1, columns A and B, which assume that the best-guess estimate of moraine volume is reasonable). However, I am fairly confident that the best-guess scenario is a better guide to true values than are the extreme HIGH and LOW estimates, so I think it is reasonable to conclude that the striking contrast in flushing efficiency and moraine development which exists between Gígjökull and Sólheimajökull today held in the past as well. I calculate mean (= Neoglacial optimum) accumulation rates at Ystagil/Little Ice Age Sólheimajökull to have been 0.27-0.63 m^3 m^{-1} yr^{-1} (Chapter 2); the equivalent MAMA at Gígjökull is likely to have been ~1.0-2.0 m^3 m^{-1} yr^{-1}, sustained for a much longer time (2,500-5,000 years). The 'worst-case' LOW scenario for Gígjökull matches the mid-range of my estimate for MAMA under *most-favourable* Neoglacial moraine-forming conditions at Sólheimajökull (Chapter 3), so I am confident that this conclusion is robust.

[1] This figure applies only to those parts of the Neoglacial for which ice was actually in contact with the moraine rampart.

Moraine structures

Few sections give little clue as to the moraine's internal structure. My overall impression is of a chaotic pile of largely structureless debris. Parts of the moraine are composed of indurated sediments, but most parts of the non-vegetated inner flanks seem to exist in a just-stable state at the sediment angle of repose. However, small sections at the base of the inner slope of the western ridge reveal a crudely laminar debris structure which mirrors surface slopes, and includes elements of graded bedding with cobble and boulder beds. This is consistent with partial reworking by flow till activity (Lawson, 1982). Similar structures were seen to form as the product of active flow-till redistribution of present-day ice contact debris.

The outer flanks of the ridges are semi- to fully-vegetated, and seem to have been subject to widespread degradation: i.e. outer slopes stand well-below the sediment angle of repose. This is likely to reflect both 'standard' subaerial mass movement and slope wash processes, and the impact of *jökulhlaups*: the outer flanks of both ridges are cut by a number of distinct flood channels (e.g. the prominent notches in the moraine profile at ~700 m and ~900 m: Fig. 8.2). The contrasts in slope angle, stability, level of soil development and vegetation cover probably reflect near-continuous exposure of the outer slopes to subaerial processes over hundreds to thousands of years, but support and protection of the inner slopes from sustained occupation by glacier ice.

Moraine sediments

The contribution of rock-fall debris to the moraine ramparts is negligible. The western moraine (the safer to work on) was studied in detail: noticeable concentrations of angular clasts consistent with rock-fall inputs were found only on the proximal parts of the ridge (cf. Matthews and Petch, 1982; Matthews, 1987). The bulk of the ridge contains clasts which are sub-angular to sub-rounded, perhaps with a rise in the frequency of sub-rounded clasts towards the distal parts of the ridge. However, I am not particularly concerned here with the possible existence and significance of clast roundness gradients; it is the overall distribution which I think is important (Fig. 8.3). The variance of clast roundness displayed by debris of the Neoglacial rampart moraine is much greater than the variance of individual debris populations identified today at Gígjökull (Chapter 4). Indeed, it seems likely that the rampart moraine is constructed largely of a mix of two debris populations: its clast roundness distribution is remarkably similar to the *lumped* distribution of the relict conduit debris and the basal ice debris (although the *exact* proportion contributed by each of the two debris types is unknown). I use this fact to propose that the styles of debris delivery which operate at Gígjökull today were also responsible

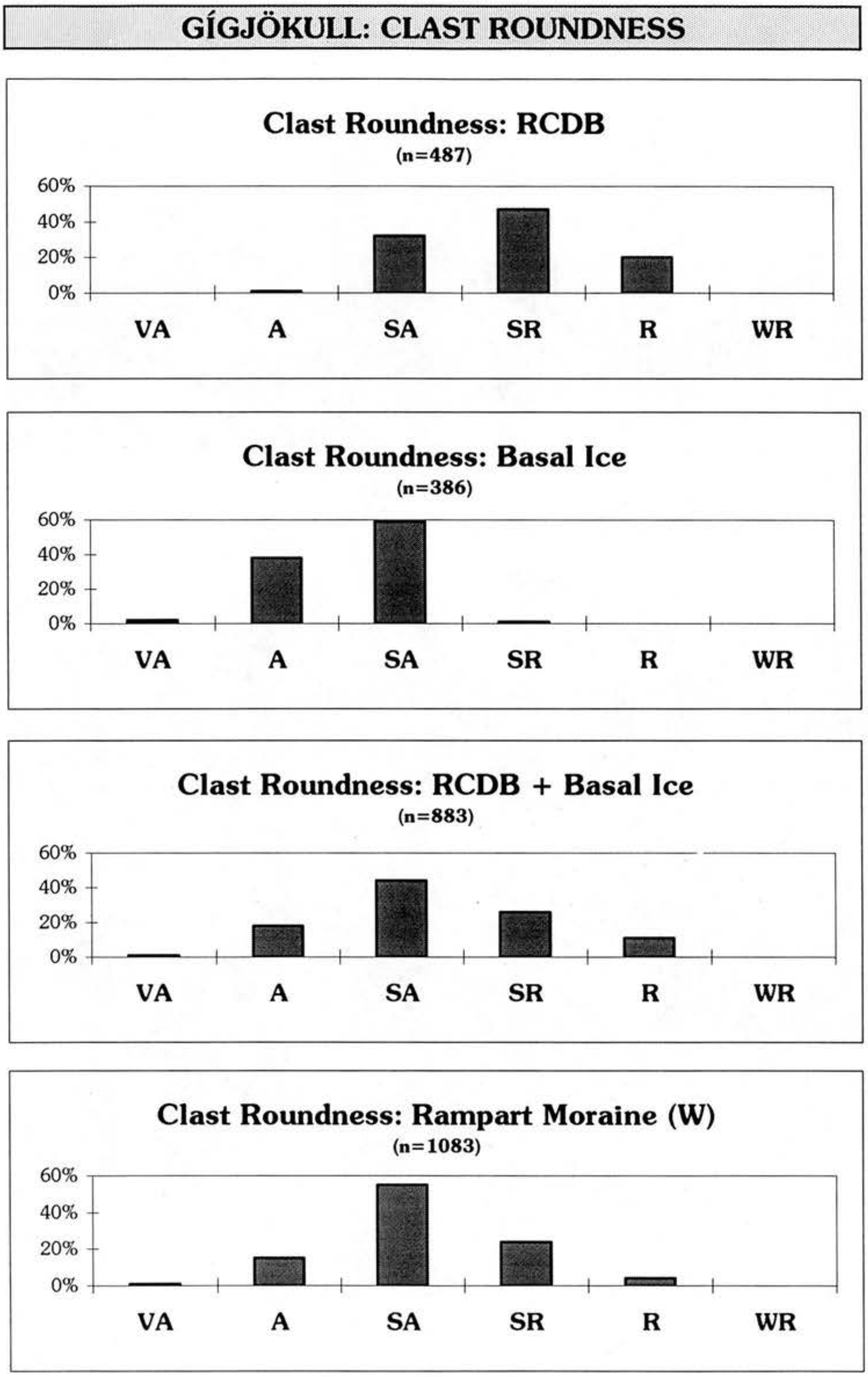


Figure 8.3
Gígjökull: clast roundness distributions. RCDB = relict conduit debris band sediments.

for building its Neoglacial moraine. This seems reasonable; if anything I expect the processes to have worked with greater intensity in the past (see below). Contrary to Matthew's (1987) expectations, I anticipate that the contribution of rock-fall debris drops as climate cools, because the area of exposed rock-wall falls rapidly as the glacier expands and thickens.

MORAINE CHRONOLOGY

It is clear that the mid-to-late-Holocene behaviour of Sólheimajökull and Gígjökull was very different, although both may be considered to be equally anomalous. Sólheimajökull displays an unusually detailed record of ice advances/climatic fluctuations which is extended in space, thanks to the probable impact of ice-divide migration, which eliminates the tendency for successive episodes of successively greater ice advance to override and destroy moraines of previous advances (Chapter 3). Gígjökull displays a record of similar detail: Dugmore and Kirkbride (draft MS) detect eleven episodes of ice advance in the past 2,000 years. The record of Gígjökull is *compressed* in space, however. This record was put together by painstaking tephra and lichen dating of individual fragments which make up the composite moraine. This shows that, despite multiple re-occupations by ice, the moraine has never been fully overridden and destroyed, as is common elsewhere.

Historical records show that Gígjökull occupied much of the proximal slopes of its Neoglacial moraine early in this century (Eythórsson, 1931, cited by Dugmore, 1989). This implies ice occupied the crest of the moraine ridge late in the Little Ice Age. Till plugs which occupy spill-way breaches of the crest of the western ridge show that ice must have made contact with the ridge crest after the 1821 eruption and flood. However, it is clear that much of the moraine is much older than the Little Ice Age (Dugmore, 1989; Dugmore and Kirkbride, draft MS; Mulligan, 1994). Extensive, little-disturbed soils have developed since the Hekla 1341 eruption. The maximum age of the moraine must pre-date 250 AD: a soil pit dug on the outer flank of the eastern moraine revealed 'Layer Y', separated from basal till by ~40 cm of soil and unidentified tephra bands (Mulligan, 1994; see also Dugmore, 1987). Extrapolation of soil accumulation rates implies that this basal till is ~2,200 years old. The location of this soil pit a short distance from the ridge crest implies that large parts of the moraine must be older still. This soil pit evidence suggests more-or-less continuous construction of the ridge over the entire span of the Neoglacial.

This dating carries two important implications:

- Despite the odd ice excursion, the composite ridge seems to have been built upwards and backwards by the 'retro-stacking' of slabs of sediment. This fits the accretionary method of lateral moraine construction envisaged by Röthlisberger and Schneebeli

(1979). It does not fit Small's (1983) 'superposition' model of dump moraine growth, whereby debris is bulldozed over or dumped at the ridge crest to cascade down its outer flanks, so the ridge builds upwards and outwards.

- At the time of coldest/snowiest climate in the Little Ice Age, Gígjökull occupied a position at the moraine crest which, if anything, was behind the extent it reached earlier in the Neoglacial under conditions apparently less favourable to ice advance.

Rudimentary AAR calculations support the idea that Gígjökull was seriously 'under-advanced' in the Little Ice Age. I use Dugmore and Sugden's (1991) figure of maximum regional ELA depression of 400 m, and calculate the AAR assuming that the edge of Gígjökull coincides with the crest of the rampart moraines. Taking uncertainty as to the present-day ELA into account, this suggests a late Little Ice Age AAR of between 2.86:1 and 3.60:1. These values are high in comparison with 'conventional' values of ~2.0:1. I infer from these calculations that Gígjökull was held back in some way at the time of the Little Ice Age: i.e. it should have advanced substantially further, to a position well-beyond the rampart moraine. It is unlikely that the Little Ice Age catchment of Gígjökull was smaller than it is today. It is far less good a candidate for ice-divide migration than is Sólheimajökull: a) Gígjökull's catchment - the summit crater of Eyjafjöll - faces north, *away* from prevailing winds; and, b) the bedrock topography of Eyjafjöll is steeper, and so less suited to ice-divide migration, than the bedrock topography beneath Mýrdalsjökull.

8.2 TOPOGRAPHIC PINNING POINTS?

Dugmore (1987, 1989) first identified the 'under-advance' of Gígjökull. He believed this reflects:

1. The interaction of climate, ELA and topography. Gígjökull is expected to advance rapidly as climate first starts to deteriorate because of the relatively wide and flat summit zone of Eyjafjöll. Thus a small fall of the ELA adds a large expanse of accumulation area. However, as climate continues to deteriorate (i.e. in the Little Ice Age) ELA falls across the steep escarpment into which Gígjökull has cut a deep trough. This confines the glacier. Thus the same small fall in ELA adds little extra to the accumulation area, and restricts ice advance (see Kerr, 1993, for more on this kind of idea).
2. As Gígjökull tries to advance across its *sandur*, its advance is pinned back because it spreads out sideways, and so rapidly increases its accumulation area. Thus the *sandur*

edge represents a particularly stable location. Gígjökull can reach it quickly as climate deteriorates, but finds it difficult to advance further as climate continues to deteriorate.

Collectively, Dugmore compared these two factors to topographic pinning phenomena characteristic of calving glaciers. I am sure that this idea is correct in spirit, but it seems deficient as it stands:

- The contrast in mountain/glacier topography explains why Gígjökull becomes progressively less responsive the greater the deterioration in climate. However, it fails to explain why Gígjökull gets trapped within its previous moraine limits with such a high AAR.
- I think the reasoning of 2) is incorrect. If Gígjökull spreads sideways as it hits the *sandur*, its area continues to increase, so strictly it is 'advancing'. The large lateral ramparts suggest that this sideways spread effect must have been limited: i.e. the glacier is *confined* between its lateral moraines. Indeed, it is likely that the topographic effect of the *sandur* is to *enhance*, not restrict, ice advance. Because the *sandur* is relatively flat, as the glacier moves across it, it finds it difficult to raise total melt to balance accumulation by descending to lower, warmer altitudes. Instead, it must *increase* the extent of its accumulation area to compensate (Furbish and Andrews, 1984). The contrast in bed condition perhaps provides a better explanation of why *sandar* can pin-back advancing glaciers. Ice can slide easily over relatively smooth, impermeable bedrock, but a gravel-cobble bed, with low water pressures, into which ice can infiltrate, provides a rough bed, likely to offer greater flow resistance, and restrict glacier advance. [See Bentley (1996) for a firm (if not exactly analogous) example of the impact a switch from an 'easy flow' to a 'hard flow' bed can have on moraine formation.] However, although I imagine this effect is important in certain cases (e.g. piedmont lobes such as Hofðabrekkujökull or Skeiðarájökull) I do not think it is the key factor here.

    It is important not to take the calving glacier analogy too far:

- The idea is important: i.e. whereas the front of 'normal' glaciers conceivably can occupy an infinite number of positions given appropriate climate/mass balance conditions, glaciers with termini subject to pinning point control can find stability at just a few field locations (see Warren, 1992, for a full review of calving glacier behaviour). However, in the case of calving glaciers, pinning points represent topography/water depth controls which determine mass loss by calving. This means that calving glaciers are decoupled from climate at the process level, because calving is not closely related to climate, as ice melt is. Calving is not likely to have been an important process of mass loss at Gígjökull: as it advances, any proglacial lake will tend to shrink. Ice melt is still likely to control its mass balance. Thus Gígjökull is decoupled from climate only in the sense that its exact extent is not likely to be a reliable indication of climate change; it is not decoupled at the level of process.

- Secondly, the asymmetry of pinning point control is likely to operate in reverse at Gígjökull. Topographic influences on calving glaciers tend to induce gradual advance, but rapid, often catastrophic, retreat. However, the effect of topographic pinning points as I infer at Gígjökull exerts the greatest impact on glacier *advance*. Gígjökull is brought to an abrupt halt as it tries to move forward, but is (relatively) free to retreat in an orderly fashion. This asymmetry of behaviour in advance and retreat is likely to have important implications for the style of moraine development (see below).

My view is that the key topographic contrast is not that between the steep escarpment and the *sandur* as such; it is the topography which develops by subglacial erosion and ice-marginal deposition as a result of this 'parent' contrast. This relates directly to the ice, water and sediment dynamics I believe to be characteristic of the ice-fall/terminal overdeepening system. My interpretation of the topographic pinning point effect represents a direct, time-dependent extrapolation of the process domain I describe in Chapter 7. Dugmore (1989) suggests that topographic pinning points control the activity of other Icelandic glaciers, notably those which drain Öræfajökull. My ideas support this: the example of Kvíárjökull is entirely consistent with what follows here. Kvíárjökull is surrounded by the largest Neoglacial moraine rampart in Iceland (Thórarinsson, 1956), developed at the junction of volcanic escarpment and *sandur*. Its ice-fall drains directly to a pronounced terminal overdeepening. Present-day moraine accumulation is impressive, and reflects extensive exposures of debris-rich basal ice, and large numbers of englacial debris bands which contain water-worked debris (my observations, September, 1996).

THE MORAINE DAM EFFECT

Two things are required to build a large moraine (cf. Boulton *et al.*, 1985):

1. Rapid delivery of large quantities of debris to the ice margin. This requires: a) fast ice flow/high rates of ice melt; b) thick debris-laden horizons within the ice; and, c) high debris concentrations in these horizons.

2. The ice margin must occupy the same position for long intervals.

The ice-fall/overdeepening sediment transfer model I describe in the previous chapter meets the first of these criteria. The make-up of the rampart moraine seems perfectly consistent with a mixture of basal traction zone debris and relict conduit debris. The exact mechanism by which debris is added to the moraine (e.g. dump, push, squeeze, lodgement) is not the key factor here; it is the processes which deliver the debris to the near-vicinity of the developing moraine.

It is clear that for Gígjökull the second requirement cannot be met by an appeal to climate-induced ice-margin stability. Frequent re-occupation of the moraine by the ice-margin is required over perhaps a 5,000 years period of highly variable climate (Guðmundsson, 1997). For much of this time, ice is expected to advance beyond the moraine, but it did not. This seems to indicate the operation of some topographic barrier, which the ice hit, and occupied several times. I believe that this topographic barrier was the terminal overdeepening itself. This consists of two parts: 1) *bedrock*: the adverse slope of the *riegel*; and 2) *sediment*: the adverse slopes of the developing moraine. If the maximum height of the moraine is added to the maximum depth of the overdeepening then this represents a barrier to ice advance 210 m high sustained at an angle of 25-30° (Fig. 8.2). I expect that it is the impact of the *riegel* which first restricts ice advance, and, because of the related sediment delivery processes, first initiates localised deposition of large quantities of sediment: the situation I envisage here is similar to that at Gígjökull today. However, this in itself is not sufficient to pin back the glacier: as climate continues to deteriorate, thickening ice will overwhelm the *riegel*, and tend to push Gígjökull forward once more. At this stage the moraine dam effect must take effect in order to localise ice advance. This means that the force with which the moraine pushes back against the ice must counter the tendency of advancing ice to push the sediments forward, or to override the sediments. Intuitively, this requires that moraine growth must be rapid relative to the rate of ice advance and the forwards component of ice terminus velocity. Initial construction of this barrier possibly occurs by push moraine development: advancing ice progressively shovels sediment into a pile which eventually reaches sufficient size to restrict severely ice advance (in effect, the bulldozer is brought to a halt). Ice must then thicken to override the barrier, which is likely further to restrict ice advance - forwards ice flow components give way to upwards ice flow components - and so increase the rate at which debris is added to the moraine. I infer that this must instigate a process of positive feedback: as debris piles up at the ice edge it holds the ice back; the pinned ice thickens, and adds further debris to the moraine dam; this reinforces the tendency to suppress ice advance, so the ice thickens further, and adds more debris to the moraine, etc... This scenario is supported by the dating evidence which implies that the moraine was built upwards and backwards. This shows that ice was rarely, if ever, able to override the moraine. Its attempts to thicken sufficiently to overwhelm the moraine barrier must have succeeded only in building up the barrier further. If correct, this version of the topographic pinning point hypothesis simultaneously explains the size of Gígjökull's Neoglacial moraine, and the restricted extent of its Neoglacial advances.

Previous work

I can find few specific references to this moraine dam effect [a BSc thesis by Allathan (1995), which evaluates the response of Kvíárjökull (see above) to past climate change, provides an exception]. The idea that a large moraine rampart can constrain the advance of its glacier is implicit in Forbes' 1846 account of the Brenva Glacier, Courmayeur, Italy (Forbes, 1900). Its neighbour, the Miage Glacier, also appears to provide a fine example of the moraine dam effect in action [my observations; see Forbes (1900) and Whymper (1981) for descriptions of the Miage moraines]. So too do many glaciers in the Himalayas (Doug Benn, personal communication). However, perhaps the closest to a systematic statement of the moraine dam effect is the body of work on Neoglacial moraine forms and chronologies in the Cordillera Blanca of Peru.

The Cordillera Blanca, Peru. Major moraine ramparts frequently block valleys of the Cordillera Blanca, notably in the Huandoy-Huascaran massif (Lliboutry *et al.*, 1977a; Clapperton, 1993, pp. 317-319). These reflect major inputs of rock-fall debris associated with the steep alpine relief, active tectonic uplift and weak, widely-shattered granodiorite rocks. It is clear that these moraine ramparts restricted past ice expansion, and so stopped advancing glaciers occupying their valley floors in full. It is a short step from this clear evidence of the lateral confinement of ice to infer that moraine back-stress effects also restricted advance of the glacier front. This is certainly probable in the case of Glaciar Hatunraju, which presently sits within the confines of a monster moraine 300 m high (Lliboutry, 1977). The glacier and its moraine execute a curious 90° turn, which cannot be explained by valley topography. Lliboutry argues it reflects breach of the moraine dam, which makes possible ice advance, whereas before it was confined:

1. Advance of Glaciar Hatunraju out of its tributary valley builds up the huge moraine midway across the floor of the main valley.
2. Retreat of the ice allows a proglacial lake to form behind the moraine dam.
3. A major *aluvión* event (Lliboutry *et al.*, 1977a) creates a breach in the side of this moraine dam (an *aluvión* is a catastrophic flood of liquid mud).
4. When it next advances, Glaciar Hatunraju exploits this lateral breach, and so turns through 90° to flow down the axis of the main valley. It continues to build up its moraine. This change in direction implies that the glacier is turned by back-stress effects as it meets the old moraine front.

Further parallels between Glaciar Hatunraju and Gígjökull support the idea of the moraine dam effect. Lliboutry *et al.* (1977b) identify six advance episodes marked by six stacked moraines at the nearby Glaciar Safuna. However, just the single large ridge is found at Hatunraju. This implies that the Hatunraju moraine is a composite feature, built up by successive re-occupation - in this case by six advance episodes if the correlation Lliboutry *et al.* make is correct. This further suggests that initial growth of the moraine must have been sufficient to restrict the first, and largest, of these advances (equivalent to Safuna Stage I); if not, successively smaller advance episodes should have failed to reach the first ridge, with subsequent ridges built inside it, not on top of it.

Lliboutry's work at Hatunraju and Safuna casts light on the process of sediment redistribution by which such large moraines are constructed. He describes the big ridges of the Cordillera Blanca as push moraines, but these are not simple push moraines in the sense of (small) bulldozer or squeeze ridges. Lliboutry *et al.* (1977b) report "sheared strata of ice and earth" at the base of present-day Glaciar Safuna. At Hatunraju, Lliboutry (1977) envisages a process of debris creep induced by ice drag, largely without the medium of interstitial ice (additional transport by debris drag in this way seems necessary to explain the huge size of the moraine). The picture of debris transport immediately adjacent to the inner slopes of these moraines is of a hybrid zone of traction which sits somewhere between the ideal categories of transport by basal ice and transport as part of a true deforming bed. This inferred mixture of drag, creep and tectonic processes fits descriptions of 'active subsole drift' (e.g. Engelhardt *et al*, 1978; Echelmeyer and Zhongxiang, 1987). Redistribution of debris in the immediate vicinity of the moraine as part of this kind of process is consistent with my image of moraine accretion by way of successive slabs of sediment added to its inner slopes (cf. 'tectonic stacking': Boulton *et al.*, 1985, Fig. 4). However, at Gígjökull I think this process is likely to take effect immediately adjacent to its moraine ridge only, because of the special conditions of stress, debris availability/concentration, and ice flow found there. I do not envisage that it replaces other processes - debris transport as part of basal ice or englacial debris bands - which account for wider-scale debris delivery to the ice-marginal zone.

Columbia Glacier, Alaska, USA. The concept of the moraine dam effect fits with ideas of glacier force balance (e.g. Van der Veen and Whillans, 1993). The gravitational stress which drives ice must be balanced by basal drag, side-wall drag and tension or compression (pulls or pushes) within the ice. Obstructions to ice flow such as a transverse bedrock ridge (*riegel*) or moraine dam create back-stress effects (i.e. enhanced flow resistance) which are transmitted upglacier by means of stress gradients. Such obstructions commonly act as 'sticky spots' which exert a major influence on overall ice dynamics. Thus a sticky spot located at the terminus of a glacier is likely to control ice flow of the terminal lobe, and so influence patterns of sedimentation. Few glaciers have been studied in sufficient detail to document this effect precisely: Columbia Glacier, Alaska, USA, perhaps provides the best example here (Krimmel and Vaughn, 1987). The final section of Columbia Glacier is overdeepened, and it terminates at a moraine shoal (or at least it did prior to 1983). For its final 4 km Columbia Glacier must flow uphill, climbing 300 m at a mean angle of ~5° (the adverse slope is concave upwards, but it is still much less steep than that of the moraine/overdeepening at Gígjökull). The back-stress effect of this moraine shoal is/was pronounced. Between 1977 and 1986, terminal velocities varied inversely with glacier length: i.e. the further forward Columbia Glacier got, the greater the force with which its moraine shoal pushed back. This overturned the velocity-length relationship observed upglacier. Overall retreat of Columbia Glacier in recent decades has progressively reduced the impact of this back-stress effect (successive seasonal advances reached less far forward).[2]

Significant differences exist between Columbia Glacier and Gígjökull, notably the fact that Columbia Glacier has a tidewater terminus. Calving changes the frontal dynamics at Columbia Glacier in a way which cannot have applied to Neoglacial Gígjökull (but see below): notably the absence at Columbia Glacier of compressive ice flow at the terminus (Venteris *et al.*, 1997). However, the inferred impact of back-stress effects is the same: abrupt change to flow behaviour as fast sliding ice builds up behind a terminal barrier. Van der Veen and Whillans (1993) identify wave-like flow perturbations in the terminal zone of Columbia Glacier, thought to relate to the disruptive impact of the moraine shoal and the adverse slope. The image of waves of ice and debris piling up behind Gígjökull's moraine for much of the Neoglacial provides a potentially attractive explanation of moraine construction: i.e. the arrival of each wave instigates an enhanced episode of debris override, tractive entrainment and emplacement of a fresh slab of subsole drift to build-up the moraine.

[2] N.B. Venteris *et al.* (1997) dispute the importance of the back-stress effect at Columbia Glacier. However, it is not clear that the two papers are contradictory. Venteris *et al.* evaluate the hypothesis that back-stress controls terminus position, and conclude that it is incorrect; however, Krimmel and Vaughn assume that terminus position controls back-stress. This does not necessarily exclude the possibility that thinning cycles, driven by changes in subglacial hydrology, control calving and so terminus position, as Venteris *et al.* argue. It is true that amplitude of the annual speed cycle rises sharply after Columbia Glacier retreats from it moraine shoal (which is not what would be expected if back-stress governs the flow speed of the terminus), but it is not impossible that this reflects a change in ice geometry and/or subglacial hydrology conditioned by disappearance of the adverse bed slope.

GÍGJÖKULL: PROCESS CONSIDERATIONS

It is likely that the processes I sketch in Chapter 7 operated with greater intensity at times of increased Neoglacial ice extent:

- The size of the overdeepening grows as two positive feedbacks combine. Enhanced erosion at the foot of the ice-fall progressively deepens the entrance (Hooke, 1991) at the same time as enhanced deposition at the exit progressively builds up the adverse slope which closes the overdeepening (see Fig 8.2).
- The crests of both the east and west ridges run at a mean angle of $\sim 3^{\circ}$ over a distance of ~ 1 km (EDM data: Fig. 8.2). *If* these moraines provide a reliable guide to past ice-margin geometry (the possibility that proximal sediments of a steeper moraine/ice margin have been removed by flood flows cannot be discounted) then this implies that at times of maximum extent, the surface slope of Neoglacial Gígjökull was much less steep than it is today (the mean slope of the final 1 km of present-day Gígjökull is $\sim 9^{\circ}$). However, it is likely that the mean angle of the adverse slope at the exit to the overdeepening remained much as it is today (i.e. $>20^{\circ}$). This implies that 1) the disparity between surface and bed slopes rises; and, 2) the zone across which this disparity exceeds the critical factor of 1.3-2.0 (see Chapters 4.1 and 5.6) increases in extent. This suggests that the tendency for conduit drainage to break down and take an englacial route is strengthened, which further implies 1) increased formation of englacial debris bands, 2) improved preservation of basal ice, and, 3) reduced flushing efficiency relative to the present day. These factors favour rapid moraine growth (see Chapter 7).
- At times of ice advance, sediments of the moraine walls will form large parts of the glacier bed. If these are permeable (the Hatunraju moraine is permeable at the level of its glacier: Lliboutry, 1977), water can escape by pore flow. This further reduces the quantity of water at the bed, which directly suppresses flushing, and, by raising the bed friction, enhances compressive flow and thickening of marginal debris-rich horizons.
- It is possible that, as ice piles up behind the moraine barrier it steepens to form a central bulge. If so, this suggests that surface slopes are steeper than the moraine crests indicate, and so the intensity and extent of changes to drainage should be reduced. However, if this happens, the cross-glacier surface slope should steepen, which will strengthen the lateral components of flow, and raise the intensity with which debris-laden ice is driven towards the glacier sides.

A big problem with this idea of the moraine dam is how to satisfy mass balance requirements with such a large AAR: i.e. melt per unit area of the proportionately small ablation area must be unusually high. Perhaps the simplest way to explain this is to assume that Neoglacial Gígjökull never did reach equilibrium; despite the fact that it was seemingly stationary, it continued to 'advance' at times of favourable climate by thickening within its moraine barrier. However, the following factors possibly explain enhanced ablation rates:

- Dark, debris-rich ice melts more quickly than clean ice.
- Compressive flow creates a surface bulge which has a high surface area to plan-form area ratio.[3] Small areas of relatively steep-sided ice - i.e. the walls of this hypothetical bulge - are likely to provide a substantial increase in melt at this latitude because of the low angle of the sun in the sky (i.e. steeper ice slopes raise the angle of incidence).
- 'Calving' of ice blocks occurs as ice builds up at the moraine crest. This was observed at Kvíárjökull in the late nineteenth century (Thórarinsson, 1956). If this process operates widely it simultaneously provides an efficient means of mass loss (calving of a 1 m^3 ice block is equivalent to 17 days' ablation of 1 m^2 of ice surface at present rates) and helps to explain why the glacier cannot over-top and advance beyond its moraine: i.e. the 'calving' process must operate with an intensity similar to that of Alpine cirque glaciers perched at the edge of rock steps. This is perhaps unlikely.

8.3 CLIMATE CHANGE, GLACIER BEHAVIOUR AND STYLE OF MORAINES

(a.k.a. the hummocky moraine debate revisited)

I do not wish to rehearse in full here the debate over the relationship between climate change and the style of moraine formation. I do wish to point out that the example of Gígjökull adds to the body of work which indicates that moraine formation cannot be read-off as a simple consequence of climate; it must be studied as a problem which depends upon the style and quantity of debris delivery to the ice edge: i.e. at the level of geomorphological process. If the moraine dam hypothesis is correct, it is clear that the impact of debris on the ice dynamics of the terminus is to disguise heavily the relationship between the climate of the Neoglacial and the Gígjökull moraine. It is impossible to read the moraine as any kind of climate signal without prior interpretation of it as a process geomorphology signal.

I suspect also that retreat of Gígjökull is likely to create a complex moraine sequence that must be read first in terms of debris delivery. This impinges upon the 'hummocky moraine'

[3] To take an extreme example: the surface area of a hemisphere is twice that of the flat disc it covers.

debate (for a summary see Benn, 1992). Response of a glacier trapped behind a moraine dam to conditions of negative mass balance is likely to be a two-stage process:

1. Initial response involves ice thinning, rather than retreat as such. It is possible that insulation provided by an extensive debris cover further delays recession of the ice margin. In this time, large quantities of debris are likely to collect adjacent to the inner flanks of the moraine. However, once retreat occurs, this debris is likely to be re-worked by slope processes, so the form of the moraine reflects 'subaerial' factors such as sediment angle of repose or the intensity of slope wash/debris flow activity which cuts gullies into the moraine slope.

2. As the ice retreats from the moraine walls more space becomes available, so deposition of debris is less confined, and debris is dumped onto less steep slopes which limits re-working of debris by large-scale slope processes (small-scale slope processes associated with ice-cores are likely still to be important however). Active ice retreat is likely to give rise to extensive deposits of hummocky moraine, because of the process of 'incremental marginal stagnation' induced by a thick debris cover (Eyles, 1983). (N.B. It is not necessarily accurate to assume that the debris-covered ice is stagnant/decoupled from active ice upglacier: Kirkbride, 1995b).

If its present retreat continues, I expect Gígjökull to enter this second phase of retreat moraine development (subsequent ice advance has destroyed/submerged evidence of more extensive retreat earlier in this century). Kvíárjökull currently is at this stage. Today it stands ~500 m inside the foot of its eastern moraine rampart. Much of the area between the ridge foot and the present-day ice front is infilled with an expanse of hummocky moraine (my observations, September 1996). Eyles includes Kvíárjökull as one of his examples of incremental ice-marginal stagnation (Eyles, 1983, Fig. 5a). However, he attributes the thick debris cover responsible for this local ice 'stagnation' to supraglacial rock-fall inputs, but this is not the case for much of the moraine at Kvíárjökull, or at Gígjökull.

Moraine form reflects the local processes which build moraines. These in turn depend upon the catchment-scale processes which deliver debris to the ice edge. As Eyles suggests, local 'stagnation' of debris-rich ice margins is entirely compatible with active ice retreat. The cryptic order which Bennett and Boulton (1993) identify in the hummocky moraines of the Scottish Highlands does not necessarily indicate a nested recessional sequence in which each single ridge represents a temporary pause during ice retreat. Multiple englacial debris bands at Gígjökull today give rise to a stacked, quasi-concentric series of supraglacial ridges. If these survived (hypothetical) *in situ* melt of the ice (i.e. ice stagnation!), then the result would be a nested ridge sequence associated with a *single* terminus position which does not reflect active retreat![4]

The example of Gígjökull entirely supports Benn's (1992) conclusions: 'hummocky' is useful simply as an adjective which describes accurately the appearance of many moraine spreads. It indicates nothing more. Meaningful interpretations of hummocky moraine (or, indeed, any type of moraine) must rely on judicious reconstruction of its process origins. In this case, I expect future development of hummocky moraine at Gígjökull (as at Kvíárjökull) to reflect the debris-rich character of its ice margin. The bulk of this debris is of subglacial origin. This sets Gígjökull apart from the majority of glaciers at which debris cover is believed to exert an important influence on mass balance, ice dynamics, style and rate of ice retreat, and style of moraine formation. In these other cases, large volumes of debris are attributed to supraglacial inputs: e.g. Tasman Glacier, Southern Alps, New Zealand (Kirkbride, 1989, 1993, 1995b). Gígjökull is unusual because it ends in an overdeepening which creates a process regime in which flushing is suppressed. Thus both a) the past record of moraine construction, and, b) expected future moraine development, support the central idea of this thesis, namely that flushing efficiency exerts a major - and enduring - control on patterns of ice-marginal sedimentation. This supports my view (Chapter 1.4 and 1.5) that Quaternary studies must take greater account of process geomorphology.

8.4 STEINHOLTSJÖKULL'S NEOGLACIAL MORAINE RECORD

It seems that the Neoglacial record of Gígjökull is explained by extrapolation into the past and intensification of its present-day process regime. Because present-day Steinholtsjökull shares this process regime, it is not unreasonable to expect that it gave rise to a similar Neoglacial moraine record. This is not the case: the Neoglacial moraine record of Steinholtsjökull is very different to that of Gígjökull. Important differences include:

1. There is no single, large moraine rampart, but a nested series of small lateral moraine ridges is found.
2. No moraines which pre-date the Little Ice Age are found.

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[4] See also recent work by Bennett and colleagues which reinterprets Scottish hummocky moraine as stacked slabs of thrust origin (Hambrey *et al.*, 1997). This switches the emphasis to process explanation: i.e. debris delivery by thrust activity. However, if correct, this type of hummocky moraine probably can be read as a climate signal!: fossil evidence of thrusts is thought to indicate that the edges of Loch Lomond Stadial glaciers consisted of cold-based ice.

3. The clast properties of surviving moraines suggests that debris contribution from relict conduit debris bands was not important here.

STEINHOLTSDALUR

The moraine record in front of Steinholtsjökull was wiped out by the 1967 *Steinholtshlaup*. Four terminal moraine ridges, visible on the 1946 aerial photographs were destroyed, or buried beneath the debris which today chokes the valley floor (Dugmore and Kirkbride, draft MS). However, their equivalents survive on higher ground which borders Steinholtsdalur to the south.

SUÐURHLIÐAR

This is the low (~100 m high), flat-topped ridge which separates Steinholtsdalur from the Krossá valley to the north (Fig. 4.1). It is an area which favours preservation of moraines (i.e. an elevated expanse of low relief): an extensive moraine complex is developed here, which indicates that in the past Steinholtsjökull must have been at least 100 m thicker than it is today. Soils developed immediately beyond the outer limits of this moraine complex contain at their base a black Katla ash, probably that of the 1755 eruption, which means that these moraines date from the Little Ice Age (Dugmore, 1989). It is possible that Steinholtsjökull attained a greater extent earlier in the Neoglacial, and built moraines on the slopes of Norðurhliðar/floor of the Krossá valley which were subsequently destroyed by slope and/or river erosion. However, there is no evidence of residual till boulders on these slopes, on which soil profiles exist which are perhaps 7,000 years old. Because of this, Dugmore concludes that Steinholtsjökull's Neoglacial maximum was reached in the Little Ice Age. This means it fits into the category of 'well-behaved' glaciers, which, unlike Gígjökull and Sólheimajökull, exhibit a 'normal' response to regional climate change (Dugmore and Sugden, 1991).

The moraine complex of Suðurhliðar is made up of two distinct components: 1) distinct lateral moraine ridges; and, 2) a semi-fluted till surface.

Lateral ridges

Dugmore and Kirkbride (draft MS) identify eight of these. The larger, outer ridges are ~5 m high, and are visible as distinct wiggles on the aerial photos. Their overall arcuate plan-form indicates a lobe of ice ~1.5 km wide which over-rides the escarpment and pushes out towards the Krossá valley at times of major ice extent. These ridges consist of a relatively loose till, with angular to sub-angular clasts (mean roundness score = 2.46, s.d. ±0.61, n = 96).

Figure 8.4

Steinholtsjökull, Suðurhliðar: stoss-and-lee boulder embedded in a fluted till surface.

Semi-fluted surface

The area inside the outermost lateral ridges consists of an undulating cobble and boulder-rich till surface. Sections exposed in the cliff edge indicate that this till is as much as 3.0-5.0 m thick. If this thickness is sustained, the 1 km wide section of Suðurhliðar between the two meltwater channels cut into the ridge (see below) possibly contains as much as $\sim 1.2 \times 10^6$ m^3 of surface till. If this was deposited in the Little Ice Age over, say, 250 years, then this gives a mean accumulation rate of ~ 5 m^3 m^{-1} yr^{-1}. This is pretty quick: my best-guess estimate for mean growth rate of the Gígjökull moraine over *2,500-5,000* years is ~ 1.0-2.0 m^3 m^{-1} yr^{-1} (Table 8.1). It is important to take note of the order-of-magnitude difference in time-scales here: those periods of the whole Neoglacial in which moraine growth at Gígjökull was active are likely to have involved deposition which occurred with an intensity comparable to, if not greater than, that which I infer at the Suðurhliðar lobe of Little Ice Age Steinholtsjökull.

Elongate ridges superimposed upon the undulations of this till surface qualify as flutes. These flutes have maximum relief of ~1 m, and maximum length of ~80 m. Striated stoss-and-lee boulders (Boulton, 1978; Sharp, 1982) embedded in the flutes are common (Fig. 8.4), but individual flutes are not tied to individual boulders. The flutes consist of a structureless, relatively compact clast-rich till with a sandy matrix. Clasts are largely sub-angular: mean roundness score = 2.85, s.d. ±0.72, n = 250; see Table 8.2. The troughs between ridges contain shallow, poorly-developed soils and sparse vegetation; boulders in the troughs are rare. The distal parts of the flutes merge with the outer lateral ridges; fragments of smaller lateral ridges seem to sit on top of/lie across the flutes in some places.

Table 8.2

Selected clast roundness scores. N.B. 'fresh' = ice-contact, present day; 'LIA' = Little Ice Age.

Clast roundness					
Glacier	**Debris Type**	**Age**	**Sample Size**	**Mean Score**	**S.D.**
Steinholtsjökull	Lateral moraine ridge	LIA	96	2.46	0.61
Sreinholtsjökull	Lodgement/fluted surface	LIA	250	2.85	0.72
Steinholtsjökull	Rock-fall debris	Fresh	154	2.34	0.60
Gígjökull	Rock-fall debris	Fresh	248	2.19	0.62
Gígjökull	Basal ice debris	Fresh	396	2.60	0.54
Sólheimajökull	Basal ice debris	Fresh	63	2.80	0.41
Hofðabrekkujökull	Basal ice debris	Fresh	123	2.67	0.67
Tindfjallajökull	Lodgement surface	C20?	131	2.82	0.48

Flute orientations suggest radial ice flow: orientation is ~350° magnetic beside the meltwater channel which confines the till surface to the east, and swings to ~330° magnetic beside the western meltwater channel; the edge of the escarpment runs at ~290° magnetic. This is consistent with the image of a small lobe of ice pushed out over the ridge surface. The mean orientation of crest striations measured on 31 prominent stoss-and-lee boulders was 335° magnetic, s.d. ±28.5°. Crest striation orientation tends to coincide with boulder orientation; few boulders show clear signs of multiple sets of striations. My overall impression is that striations and boulder alignments match that of the flutes; however, there is sufficient mismatch to imply that boulders were put in place by flow episode A before the till surface was part-remodelled by flow episode B, in which ice flow direction was slightly different. If anything, boulder and striation orientation is skewed towards Steinholtsdalur, whereas flute orientations suggest more pronounced lobate flow towards the Krossá valley. Thus it is possible that emplacement of boulders and striations reflects episodes of flow associated with thick (debris-rich?) ice, with flow directed more towards Steinholtsdalur. Subsequent thinning of the ice permits local topography to exert a greater influence on flow, so flow vectors within a thinner ice lobe turn slightly towards the north. Flute formation reflects this later phase of ice flow. However, it is likely that this area was reoccupied by ice - and so remodelled - several times in the Little Ice Age (Dugmore and Kirkbride, draft MS).

Interpretation

This till surface is the product of lodgement and deformation (see Hart, 1995a, for a review). The character of the till is consistent with transport of sediment in the basal traction zone. The lateral ridges are probably dump moraines (cf. Small, 1983). This is supported by the contrasts in till compaction (the till of the ridge is relatively loose) and clast angularity. It seems that the ridges contain a greater proportion of angular clasts, which implies the presence of rock-fall debris, as expected in a dump moraine, but not in a lodgement/deformation till: see Table 8.2). However, I am not certain if this difference is genuine: statistical tests give ambiguous results. The Kolmogorov-Smirnov test fails to differentiate between the two sample distributions, but a difference-of-means test rejects the null hypothesis with vanishing probability of a Type I error (see Box 5.1 for details of these tests). However, the contrast is consistent with sediment transport relationships likely at such a field site. In total, the Suðurhlíðar moraine complex closely resembles the forefield of Tindfjallajökull's thin, gently-sloping western lobe, exposed by twentieth century ice retreat. (Tindfjallajökull is a small mountain ice-cap located the other side of the Markafljót valley: Fig. 1.4) This is made up of an undulating till which contains stoss-and-lee boulders. Both zones resemble lodgement/deformation till surfaces exposed at Vestari-Hagafellsjökull, Iceland (Hart, 1995b), and Columbia Glacier, USA (Hart and Smith, 1997).

8.5 DISCUSSION: A POSSIBLE DOUBLE SWITCH IN MELTWATER BEHAVIOUR?

STEINHOLTSDALUR: THE PROGLACIAL RECORD

Photographic evidence of separate, relatively small terminal moraine ridges prior to the 1967 flood shows that the moraine dam effect was not a factor at Steinholtsjökull. Factors of sediment supply did not control the response of Steinholtsjökull to Neoglacial climate change. The absence of a single large ridge/moraine dam effect indicates significant divergence between the Neoglacial behaviour of Gígjökull and Steinholtsjökull. This possibly reflects the fact that Steinholtsjökull is/was a smaller, less active glacier, so it did not generate sufficient debris to give rise to the moraine dam effect. However, debris supply to the ice edge must reflect flushing activity also. I think that the fact that the debris accumulation threshold at which the moraine dam takes effect in was not crossed at Steinholtsjökull probably indicates that, whereas at Gígjökull flushing efficiency is believed to have diminished in the Neoglacial, at Steinholtsjökull I expect it was increased:

- As Steinholtsjökull expands and thickens, the importance of its overdeepening is reduced. The transition between escarpment/ice-fall and *sandur* is far less pronounced at Steinholtsjökull than it is at Gígjökull, and the 1990 map - Sheet 1812: III, 'Eyjafjallajökull - compiled using data collected at a time when Steinholtsjökull was ~750 m further back than it is at present, shows that the terminal overdeepening must be short and shallow (Fig. 4.1). In the Little Ice Age, Steinholtsjökull was at least 100 m thicker than it is today, and perhaps extended as much as 1.5 km further down Steinholtsdalur (Dugmore and Kirkbride, draft MS). Under these conditions, the perturbations to ice, water and sediment transfer introduced by the presence of the overdeepening are likely to have been less significant than they are today.
- The fact that in the Little Ice Age the overdeepening was also located some distance upglacier is likely to have been important. The process model I describe in the previous chapter works only if the glacier terminates at the overdeepening. Significant lengths of glacier beyond the overdeepening permit any changes to water-sediment transfer relationships to be undone.
- As Steinholtsjökull expands it is increasingly restricted by the impermeable bedrock walls of its trough. Except where ice can spill over onto the ridge of Suðurhliðar (see below) this will confine meltwater and sediment in transit, and so increase the chances of contact between the two. Given that water and debris are likely to return to the bed of the glacier beyond the overdeepening (if they ever leave the bed), this should increase the incidence and efficiency of flushing.

- It is difficult to be sure because of the impact of the 1967 flood. However, the absence of rounded debris in moraines of Neoglacial age at Steinholtsjökull fits the idea of reduced importance of the overdeepening (i.e. englacial drainage is less widespread), with much enhanced debris flushing.

SUÐURHLIÐAR: FLUTE FORMATION

The elongate undulations which make up the till surface are thought to reflect some element of differential deformation of till. Till deformation can occur at the same time as emplacement of debris ('lodgement' and deformation coincide), or afterwards as debris is re-mobilised. Hypotheses of flute formation include:

1. Pre-existing till is squeezed into cavities which form in the lee of boulders or bedrock obstacles (Boulton, 1976; Benn, 1994).
2. Flutes form in response to convergent-divergent flow cells set up within basal ice. Convergent ice flow drives deforming sediment together to form an elongate ridge. Sediment in this ridge is supplied both by lateral erosion and transport of till, and by inputs of debris from upglacier. This hypothesis does not require the presence of a boulder to initiate flutes (Rose, 1989).
3. Streaming of basal ice gives rise to elongate zones of above-average debris concentration. Flutes form as this debris-rich ice melts (Gordon *et al.*, 1992).

All of these hypotheses are plausible. However, the possible mismatch between boulder/striation alignment and flute alignment at Suðurhliðar suggests that (3) alone is insufficient, whereas the absence of a one boulder, one flute relationship casts doubt on the relevance of hypothesis (1) (the flutes of Suðurhliðar are too large to fit this idea). Rose's ice flow cell idea possibly fits here: a) it offers a possible explanation of the apparent sorting of boulders, and, b) mismatch of boulder/striation and flute alignments possibly reflects re-alignments of the basal ice flow cells as ice geometry changes. However, it is perhaps better to view these flutes as small, elongate examples of deforming bed drumlins (e.g. Boulton, 1987; Hart, 1995b), created by differential deformation of basal sediment associated with contrasts in till rheology (and possibly conditioned by lodged boulder cores), rather than by differential deformation of basal ice. Indeed, some combination of the two mechanisms, involving also the direct addition of material from debris-rich basal ice, perhaps repeated over several cycles of ice advance, is not impossible.

A full analysis of flute formation is not necessary here. My data are not of sufficient quality nor quantity, and the full problem is beyond the scope of this thesis. However, I am interested in the wider conditions of debris supply and basal hydrology which make construction of subglacial till forms possible. I wish to propose a stable process domain exists, associated with subglacial deformation, and part-conditioned by meltwater activity, much as I identify process domains which favour construction of lateral-terminal moraines (e.g. Gígjökull) or proglacial fluvial features (e.g. Sólheimajökull) (see Chapter 9.1).

The Suðurhliðar moraine complex (and the forefield of Tindfjallajökull) represent a kind of ice marginal (i.e. subglacial) sedimentation I have not yet considered. The relevant facts are:

1. The till is made up of sediments transported as part of the basal traction zone.
2. The quantity of sediment emplaced implies relatively rapid and sustained deposition at the time-scale of the Little Ice Age.
3. It is not known how the till was first emplaced. However, it seems that deposition was never able to restrict ice advance as at Gígjökull: i.e. a) growth of a potentially deformable bed enhanced flow, and/or b) any terminal moraine was too small/too weak to hold ice back.
4. At some time, subglacial till was subject to (low-grade?) deformation.
5. A considerable thickness of till has survived cover by ice. Subglacial supply of sediment must have out-stripped subglacial sediment removal.

The role of meltwater is important here at three stages:

- Large quantities of till emplaced rapidly imply that debris which occupies transport pathways upglacier is relatively immune to removal by contact with meltwaters.
- Deformation of sediment implies that till water content/pore water pressures were quite high, but perhaps not so high as to induce pervasive deformation of sediment and possible piping failure: i.e. the subglacial till was reasonably well-drained. This ties in with my expectations that both the hydraulic conductivity of the till (sandy till matrix) and the hydraulic gradient driving water flow (ice surface slope) were fairly high.
- Meltwater flow through the till once it was emplaced was sufficiently dispersed so as not to remove large quantities of sediment.

Bedrock controls of water flow

It is likely that the impact of bedrock topography in the Little Ice Age was to channel the bulk of meltwaters through/beneath the main part of Steinholtsjökull confined within Steinholtsdalur (see above). The mean angle of the escarpment which overlooks Steinholtsjökull today is ~30-45^{o}; slope degradation following ice retreat possibly means that it was steeper in the Little Ice Age. The bedrock here seems unusually weak. Today, local parts of the cliff approach the vertical. This makes it possible that the surface slope of Little Ice Age Steinholtsjökull driving water flow was insufficient to force water up the escarpment to flow over Suðurhliðar (Shreve, 1972; Chapter 1.1). If not, it is still probable that conduit flow was unstable: i.e. the adverse slope of the escarpment was twice or more that of the ice surface slope (Chapter 5.6). Thus it is likely that water either escaped downvalley, or through one of the two bedrock channels cut back into Suðurhliðar. The possible impact of these merits further discussion.

Bedrock meltwater channels. Ice-marginal gorges incised into bedrock imply the throughput of large quantities of meltwater and sediment. Two gorges are cut back into Suðurhliðar within the Little Ice Age limits of Steinholtsjökull: their heads are at GRs 715 613 and 705 616 (Sheet 1812: III; Fig. 4.1). These may pre-date the Little Ice Age, but it is likely that they did evacuate large quantities of water and sediment in this time. The eastern channel is incised by more than 20 m, and cuts right into Suðurhliðar so that today the (dry) gorge begins just a short distance from the ice edge. It is possible that a large (gradient?) conduit fed into this gorge, so that much of the water and sediment arriving from upglacier was diverted to the Krossá valley, and did not pass beneath the terminus or Suðurhliðar lobe of Little Ice Age Steinholtsjökull. Similarly it is possible that the western channel, located above the overdeepening which today contains the lake, also acted as a 'by-pass' route which captured water and sediment in the vicinity. This possibly helps to explain the absence of rounded clasts in the surviving tills: i.e. debris carried by subglacial/englacial streams was washed away as part of an efficient conduit network controlled by major, low pressure axis escape routes cut into bedrock.

'Dry' inter-channel zones. I infer that the elevated area of Suðurhliðar represents an isolated, 'dry' bed sub-system which favours accumulation of basal till. Thick till and flutes are developed between the two gorges. Available meltwater was probably restricted to that quantity derived locally, because the bulk of discharge ran in bedrock-controlled channels elsewhere (i.e. in the two gorges, or in Steinholtsdalur). Till accumulation reflects local subglacial erosion and restricted local flushing. As the lobe of ice pushes towards the Krossá valley, subglacial debris pathways and subglacial water routes diverge, so the state of water-sediment separation required to build big moraines is achieved. Till accumulates beneath a small, thin lobe of ice, drained at its base by dispersed pore-water flow. This is perfectly consistent with ideas of low intensity subglacial deformation: sufficient water is available to reduce till strength, the ice surface is sufficently steep to generate a shear stress which can mobilise till, but the ice is not so thick so as to give rise to values of normal force which render till immobile (e.g. Boulton *et al.*, 1974). The quantity of meltwater present is insufficient to threaten the integrity of the till layer. In this respect, the till is likely to be self-stabilising: i.e. it generates pore-water flow (rather than some more powerful style of flow) which is incapable of major erosion.

The subglacial conditions I infer at Suðurhliðar in the Little Ice Age are the same I believe to have held at Tindfjallajökull. The till surface exposed in front of the ice here today is split by a deep bedrock gorge which carries the proglacial stream. At the time the till was emplaced/deformed this must have been a subglacial Nye-channel which carried the majority of water derived from upglacier. This must have isolated the area either side, so it was safe from major flushing events. As a result, it developed a thick basal till under the relatively thin ice of the gentle terminal lobe.

CHAPTER 8: SUMMARY

- The large Neoglacial moraine rampart at Gígjökull can be explained by extrapolation into the past, and intensification of the sediment transfer regime which exists today.
- Reduced flushing efficiencies associated with the ice-fall/overdeepening system explain the quantity of debris present. Clast properties indicate that the moraine is built up of a mixture of basal ice debris and relict conduit debris.
- The size and form of the composite moraine rampart reflects repeated reoccupation by ice and localised episodes of deposition fixed by the presence of the moraine itself. It acts as a barrier beyond which Gígjökull cannot advance. Instead it thickens, which serves only to instigate a process of positive feedback which reinforces what I call the moraine dam effect.
- The moraine dam effect refines Dugmore's (1989) original topographic pinning point hypothesis. If correct, this explains the anomalous 'under-advances' of Gígjökull in response to Neoglacial climate change.
- If it is high, debris supply factors are likely to control the location and style of moraine accumulation at times of ice advance and ice retreat. This makes it dangerous to read the moraine record as a record of climate change. Before the climate signal can be identified, it is first necessary to interpret moraines at the process level of debris delivery.
- The Neoglacial moraine record at Steinholtsjökull probably reflects bedrock controls on flushing efficiency. As Steinholtsjökull expands it is likely to be increasingly confined by the impermeable bedrock walls of Steinholtsdalur; simultaneously the impact its overdeepening has on drainage diminishes. Smaller moraines, the absence of a moraine dam effect as at Gígjökull, and the absence of relict conduit debris of Neoglacial age are thought to reflect enhanced meltwater activity within the confines of Steinholtsjökull's bedrock trough.
- A thick cover of fluted basal till was developed beneath the Little Ice Age Suðurhlíðar lobe. This is thought to reflect the past existence of an isolated sub-system conditioned by preferential routing of meltwaters through bedrock channels elsewhere. Flushing is limited to that which can be achieved by the dispersed, low magnitude discharge of local meltwaters. This permits widespread survival of debris in transit at the glacier bed, which contributes to rapid build up of a thick subglacial till.
- The conditions of debris supply and debris (non-) removal which permit a deforming till layer to accumulate must in part reflect meltwater activity. High intensity flushing is incompatible with a thick subglacial till. Landforms of subglacial deposition/deformation

thus reflect a process domain part-controlled by meltwater. This represents an alternative to meltwater process domains conducive to formation of lateral moraines or proglacial fluvial landforms as discussed in previous chapters.

IV. CONCLUSIONS

CHAPTER 9
Conclusions

9.1 MELTWATER IS IMPORTANT!

The case studies which make up this thesis provide firm support for the view that the style of drainage exerts a major control on the pattern of ice-marginal sedimentation. This is because subglacial water flow disrupts the processes of the basal transport zone, and tends to carry away large quantities of sediment. The crucial factor is the prevailing balance between different modes of subglacial sediment transport - by water, within basal ice, or as part of a deforming till layer - which results from the interaction of a wide range of processes which operate at the full-catchment scale. Studies which isolate a specific part of the sediment transfer system cannot pick up on this. The popularity of such studies probably explains why the role of meltwater as an influence on moraine formation has not been studied in detail before.

METAPHOR OF THE ATTRACTOR

Fig. 9.1 summarises the substantive thrust of my thesis. It combines the basic ideas of realism - specifically the distinction between the level of basic process and the level of real-world outcome (Chapter 1.4) - with the idea of the attractor as used in complexity theory (e.g. Cohen and Stewart, 1994; 9.2, below). Ice, rock and water combine to create a number of different processes which act to generate and redistribute debris at the bed of the glacier. These processes operate unevenly in time and space, at a wide range of intensities (e.g. rock fracture can be rapid, or it can be non-existent), at a wide range of levels (e.g. debris transport of any kind presumes prior rock fracture). The chain of events which determines real-world outcomes - in this case, the different styles of ice-marginal sedimentation - is enormously complex. This makes it impossible to isolate any single, basic causes. Nevertheless, despite the fact that any given chain of events is extremely unlikely to occur more than once (e.g. the transport histories of individual clasts will always be unique) certain patterns tend to emerge - or as Cohen and Stewart (1994) put it, chaos tends to collapse. These patterns arise because of the presence of attractors upon which the causal chains tend to converge. I think it is reasonable to identify three attractors attached to different subglacial process regimes which favour 1) *ice-marginal* deposition (lateral moraines), 2) *proglacial* deposition (*sandar*/valley trains/braid plains), or 3) *subglacial* deposition (flutes, etc.). Individual clasts which form part of the catchment sediment

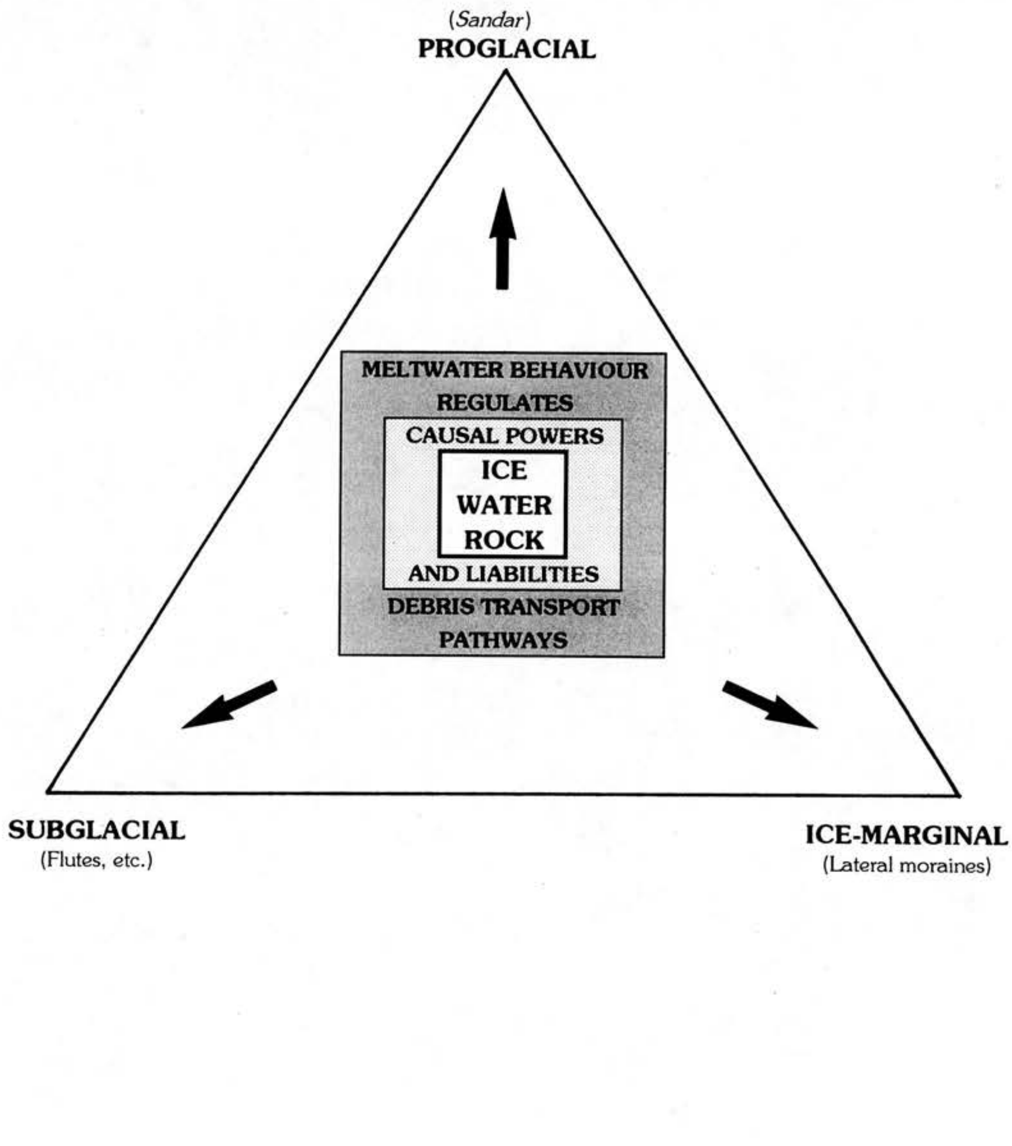


Figure 9.1

Causal structure of the subglacial sediment transfer system, and associated attractors. See text for explanation.

cascade must move towards one - and only one - of these attractors (clasts cannot be in two places at the same time). However, the process by which clasts 'choose' different attractors is not entirely random: some outcomes will tend to be more probable than others. This is because systems tend to exhibit some measure of self-organisation, which tips the balance in favour of a certain outcome. The crucial *configurational* factor which determines the relative strength of the attractors for styles of ice-marginal sedimentation is meltwater flow, as I have argued in detail in preceding chapters.

The exact importance of the attractor metaphor, and related wider frameworks such as complexity theory and non-linear thinking, has yet to be established, but it promises much. Much of its potential rests with the new light it can shed on established ideas, such as that of the equilibrium state (Lane and Richards, 1997). It is clear that the idea of the attractor provides useful guidelines on how to 'do' geomorphology (cf. Brunsden and Thornes, 1979; Brunsden, 1990), as I have tried to show in this thesis. In particular:

- If we can identify the existence of attractors, and what the critical factor(s) is which gives rise to these attractors, we can explain why specific landforms arise using an appropriate level of explanation. For instance, at present ice-marginal sedimentation at Sólheimajökull is dominated by what I call the 'proglacial' attractor, whereas the 'ice-marginal' attractor is of much greater (relative?) importance at Gígjökull. Ice sliding speeds and bedrock geology are thought to be similar, so it is unlikely that contrasts in the intensity of subglacial erosion processes account for the contrasts in moraine formation. However, some difference in the style of drainage, related to the presence at Gígjökull, but not at Sólheimajökull, of a pronounced terminal overdeepening, emerges as a likely candidate which explains the relative strength of the attractors.
- If we want to understand how and why landscapes change over time we need to know a) what causes the balance to switch between different attractors? ('bifurcations'); and, b) (of special relevance to the study of glaciers) what causes the attractor to change the location at which it operates? It is possible that a) and b) are linked. For instance, Gígjökull seems to have been 'locked-on' to the 'ice-marginal' attractor for the duration of the Neoglacial, although the exact location of that attractor (i.e. the position of the ice margin), and possibly also the intensity with which it operates, has varied as mass balance has changed. In contrast, the strength of the 'ice-marginal' attractor at Steinholtsjökull seems to have diminished in the Little Ice Age, with a commensurate increase in the relative strength of the 'subglacial' (i.e. Suðurhliðar) and 'proglacial' (i.e.

Steinholtsdalur) attractors. This switch seems to have been caused by some change in the influence bedrock geometry exerts on water flow, conditioned by mass balance-induced thickening and advance of the ice beyond its terminal overdeepening.

PROVISIONAL KNOWLEDGE?

I believe the ideas presented in this thesis to be robust. However, it is important to stress that this thesis does not fit the orthodox positivist tradition of a successful hypothesis test. Several factors, including the difficulty of direct access to the bed of glaciers, the multiple degrees of freedom characteristic of the subglacial process regime, and the historical aspects of the problem of moraine formation make it unlikely that these ideas can be subject to rigorous testing (cf. Alley *et al.*, 1994; Richards, 1996). Confidence in my ideas stems from:

1. The strength of previous work by others on individual processes.
2. My hope that the way in which I have put previous ideas together is both physically plausible and internally consistent.
3. The fact that general ideas developed on the basis of a single case study seem to work well elsewhere: notably the similarity between Gígjökull and Kvíárjökull.
4. The absence of an alternative body of ideas which obviously contradict mine.

It is possible that my ideas seem to work well because of the choice of field sites. Conditions specific to southern Iceland - fast-flowing glaciers, weak bedrock, plentiful ice melt, and a paucity of supraglacial debris sources - are likely to provide exactly the right conditions for the kind of process relationships I discuss to operate with maximum effect. The influence of meltwater on ice mass sediment budgets is likely to be less important in different environments, for reasons which will not always be obvious: e.g. it will be difficult to say anything meaningful about the partition of debris between different types of transport, and so different types of ice-marginal landforms, if rates of debris production are low to start with. Such uncertainty is an inevitable part of the 'messy' real world, and science must learn to live with it.

9.2 THE WAY WE THINK IS IMPORTANT TOO!

I wish to end with the text of a paper published in *Earth Surface Processes and Landforms* (Spedding, 1997), written as a response to the 1996 British Geomorphological Research Group symposium on the theme of 'Linking short-term geomorphic processes to landscape evolution'.

This was held as part of the Annual Conference of the RGS-IBG, held at Strathclyde University, Glasgow, in January 1996. The paper was written as a stand-alone piece of work, but it does draw heavily on the ideas I developed whilst writing this thesis. It extends my arguments to cover the wider field of geomorphology, adds extra historical detail, and picks up on recent ideas (in the tradition of 'popular science' writing) in biology which closely match my own. The skirmish here with complexity theory provides the most important addition to the ideas set out in Chapter 1.4, although I think there are strong parallels between this and realism (see above). This thesis was conceived as my attempt to address the problem of satisfactory geomorphic explanation at the level of process-form interaction, so this paper, which sets out the current state of my ideas on this question, as a result of my research, provides what I hope is a fitting and worthwhile conclusion. The text is presented here as it appears in *ESPL*, with the exception of the passages enclosed in square brackets, which were excluded from the published version because of pressure of space. The critical comments of referees Stuart Lane and Barbara Kennedy forced me to clarify many of my arguments, leading to substantial improvements in the manuscript. Barbara Kennedy suggested I borrow the paper's title from the biologist/morphologist D'Arcy Thompson's major work *On Growth and Form* (published 1917).

ON GROWTH AND FORM IN GEOMORPHOLOGY

Standard histories of geomorphology tend to identify a basic conflict between two extremes: (1) (timebound) descriptive regional studies of landscape evolution, and (2) the (timeless) analysis of process mechanics which emerged in the post-war years to challenge the hegemony of its predecessor. This dichotomy represents a pervasive myth within geomorphology which perpetuates the impression that any study must choose between the historical or ahistorical category, and its associated package of research methods, scales of analysis, etc.. To give just one example: the special issue of the *Journal of Glaciology* published to mark the fiftieth anniversary of the International Glaciological Society includes individual papers which separately review the histories of the scientific investigation of glaciers (Clarke, 1987) and glacial geology (Boulton, 1987a). I take this as tacit endorsement of the view that the study of the behaviour of ice stands apart from the study of its concrete time- and place-specific impacts. The legacy of such divisions is seen today in the growing insistence that geomorphology must put itself back together by means of the explicit study of process-form relationships which integrate the mechanistic and the historical: e.g. within glacial geomorphology see Harbor (1993) and Menzies (1995).

If we are to accept that the relative failure of geomorphology to deliver meaningful process-based accounts of landform/landscape development represents something of a crisis, what is surprising is the complacency with which the discipline has let the crisis develop. I think it is fair to say that a large number - perhaps a majority - of professional geomorphologists subscribe to the rhetoric which advocates a redefinition of geomorphology as the explicit study of process-form linkages. If this represents a genuine sense of collective unease, standard accounts of the history and sociology of science might lead us to expect a period of intense debate among geomorphologists as we search for new ways to meet this new challenge. In my view, however, signs of such debate are scarce.

This lack of self-reflection can perhaps be explained by the suggestion that although geomorphology admits to a substantive research problem - how best to bridge the gap between process and developing form? - it refuses to see this also as a methodological problem. The fact that the majority of the participants at the BGRG Strathclyde symposium - myself included - chose to interpret its theme as an invitation to present/discuss substantive issues seems to support this argument. Within the discipline, the publication of a conceptual paper is a rare event: writing in 1978, Chorley noted that thirteen years had passed since a paper of major theoretical significance - Schumm and Lichty (1965) - was last published! 'Classics' such as this, or Wolman and Miller's (1960) work on the magnitude and frequency of geomorphic events are rehearsed diligently to introductory undergraduate classes, but make little impact on contemporary research.

In particular, geomorphology as a discipline tends to distance itself from debates which touch on the philosophy of science. It is common to regard philosophy as an irrelevant diversion from the pragmatic business of research (Chorley, 1978, p. 1; Bassett, 1994; Frodeman, 1995). As a result geomorphology proceeds happily within the research culture to which it was initiated in the 1950s: that of 'scientific method'. Few pause to consider the implications of this: I suspect that most geomorphologists absorb implicitly the conventions of scientific method rather than are instructed explicitly in its use. Fewer still see fit to challenge the hegemony of scientific method as a means of solving geomorphic problems (Richards, 1994).[1]

[1] The publication late in 1996, after this was written, of the proceedings of the 27th Binghampton Symposium, entitled *The Scientific Nature of Geomorphology* (Rhoads and Thorn, 1996), must be seen as an important exception to the arguments made in these two paragraphs!

Geomorphology's association with scientific method ties it to a research programme which is broadly positivist. It is difficult to be more precise, because the move away from descriptive historical studies itself created a new dichotomy between 'empirical/functional' studies and 'rational/realist' studies (Mackin, 1963; Chorley, 1978). Functional (instumentalist) studies seek to express empirical regularities by means of 'law-like' mathematical statements of co-variation, whereas 'realist' studies try to uncover 'law-like' descriptions of fundamental physical processes. [Chorley's use of the term 'realist' disguises the "preoccupation with process" (Chorley, 1978, p. 10) which characterises much of the work he puts in this class. I prefer to use 'reductionist' because realism recognises that basic physics and chemistry alone cannot explain real world events: the diverse factors which intervene to regulate the expression of causal powers must also be taken into account (Chapter 1.4).] Descriptions of trough morphometry using parabolic or quadratic equations (e.g. Graf, 1970) provide one example of the functional approach to glacial geomorphology, whereas the 'realist'-reductionist approach to trough formation begins with the detailed analysis of bedrock erosion processes (e.g. Hallet, 1981; Iverson, 1991b). Both schools share the view - which is challenged by ('true') realist philosophy (e.g. Sayer, 1992) - that the relationship between cause and effect is self-evident. This simple view, with its implicit - if fallacious - assumption of system closure (by which it is meant that all system variables are fixed), drives the search for regularites within a framework which is (supposedly) objective, with priority - indeed, respectability - attached to the use of technology-driven experimentation, simulation and prediction, directed towards the production of quantified data and results.

Since the 1970s the majority of studies have tended to fall within the 'realist'-reductionist method, from which preference springs the present debate. At present, large parts of the discipline seem confident in the expectation that building within this existing framework - scaling-up increasingly refined process models - will enable us to solve the problems of landscape development. What is required are better instruments, better dating techniques, better computer models, etc... This is the 'disciplinary immaturity' response frequently used to explain the relative failure of the social sciences: i.e. predictive success is simply a matter of applying the scientific method with greater rigour. Harbor (1993, p. 129), for example, announces that "[a]s we head towards the twenty-first century, glacial geomorphology will advance through the use of three-dimensional numerical models that include ice flow, basal sliding (with explicit consideration of deformable beds), erosion and deposition processes, and underlying material characteristics".

As yet, the computer model which brings together all these factors in a holistic account of glacial landscape development is the subject of science fiction. However, it is not so much technical barriers to model performance which limits their utility, but conceptual constraints. Computer models - or, indeed, specific laboratory experiments or fieldwork case studies - seek to answer certain types of questions in certain types of ways. Models which isolate certain processes, and explore the impact of these, act as a conceptual prop to our understanding, but by their nature cannot provide a complete explanation of landscape development. It is important to accept the partial character of models, and not to place too much faith in their performance or potential, nor to read too much into their conclusions. For example, a computer model of glacial trough development which suggests glaciers erode rapidly (Harbor, 1992; Harbor and Warburton, 1993) does not, in the absence of a comparable model of subaerial valley development, tell us that glaciers erode more rapidly than rivers. Similarly, improved dating techniques are not the Rosetta Stone which magically reveal the processes of landscape development. However powerful, dating techniques are *just* dating techniques: a partial approach which cannot answer questions it does not ask.

As I see it, the chief problem is that of the *context* within which geomorphology chooses to operate. Perhaps the best demonstration of this is Schumm and Lichty's (1965) paper on time, space and causality, the essence of which is that the same behaviour (geomorphic activity within the drainage basin) looks very different depending on the temporal and spatial scales at which we look at it. If we look at long intervals of time applied to large areas we tend to see a sequence of historical changes as the landscape evolves; historical states, however, which tend to conceal the action of geomorphic processes. Conversely, if we study small areas for a short time, we tend to prioritise process behaviour, but lose our sense of history. Subjective choice by the scientist defines both the context - the collection of concepts and techniques used to structure the research problem - and the nature of results. Quantum physics, for example, suggests that at the sub-atomic level matter can behave simultaneously as both waves and particles. What the scientist sees depends on what s/he sets out to see: if s/he asks 'wave' questions, s/he gets 'wave' answers; 'particle' questions, 'particle' answers (Gribbin, 1984). Significantly, to study property A in detail implies s/he cannot study property B of the same thing in detail, and vice-versa: this is the Heisenberg uncertainty principle. A similar effect is often seen in geomorphology: historical studies which prioritise dating and the calculation of bulk process rates (e.g. of erosion or sedimentation) rarely permit us to identify the complex space-time dynamics of individual processes. To extend the logic of this argument, the 'bigger, better, faster, more!' approach, for all the insights into individual aspects of geomorphology it is likely to give, will not necessarily produce the desired synthesis of process and landscape development because its asks a certain type of question within a certain tradition of scientific enquiry. If we are to claim the middle ground within which this synthesis lies, we must ask a new type of question which in turn requires a new context of investigation (see, for example, Kennedy, 1977; Chorley, 1978, p. 10).

Recent work by Goodwin (1994) - writing on the current status of biology - presents an instructive parallel here. He relates a narrative whereby biology developed its strength and identity as an individual scientific discipline by means of the widespread adoption of Darwinism. Thus the central theme of biology became that of tracing the historical sequence of changing life-forms. However, this descriptive, evolutionary approach was replaced by the rise of molecular biology in the 1950s, and the focus of research switched to reductionist investigations into the basic mechanisms of genes. With genes installed as "the basic elements of biological reality", the status of the life-form within biology diminished to the extent that any given historical state of the organism could be explained as the product of "the evolutionary adventures of genes". Within this new paradigm, the exact events by which genes are linked to life-forms and their development/evolution are simply taken for granted.

The parallels here with the standard history of geomorphology are striking: even to the publication just one year apart of the iconic papers taken to symbolise the new eras in geomorphology and biology (Strahler, 1952; see also Strahler, 1992; Watson and Crick, 1953). However, yet more pertinent is Goodwin's argument that if biology is to reinvent itself as a science which provides a meaningful link between the fundamental level of genetics and the level of the organism - the study of the interactive mechanics of morphogenesis (development of form) - it must adopt a new approach. Goodwin redefines the central problem of biology as one of *organisation*, drawing on the science of complexity to create a research framework which emphasises the morphogenetic dynamics inherent to interactive systems. Form develops within a complex generative field, which reflects the influence both of immanent features and wider environmental influences. Feedback between process and form drives the system as the morphogenetic dynamic is continuously redefined in response to the developing form it itself creates. The priority attached to properties of things such as genes, or basic geomorphic processes - the traditional focus of post-war research - disappears: *it is not the building blocks themselves which matter so much as the way in which they are put together.* This recalls Simpson's (1963) distinction between the reductionist, ahistorical (laboratory) sciences and the *compositional* historical sciences, chiefly geology and biology.

Given the current debate within geomorphology, the appeal of Goodwin's emphasis on the dynamics of morphogenesis is clear. Given the traditional niche of geomorphology (within the United Kingdom) as a sub-discipline of geography, Goodwin's redefinition of the appropriate context of investigation is equally attractive. It is the context of space and time which mediates the physical relationships which give rise to morphogenesis: "[w]hat counts in the production of spatial patterns is not the nature of the molecules and other components involved... but the way these interact with one another in time (their kinetics) and in space (their relational order: how the state of one region depends on the state of neighbouring regions). These two properties together define a field, the behaviour of a dynamic system that is extended in space, which describes most real systems" (Goodwin, 1994, p. 49).

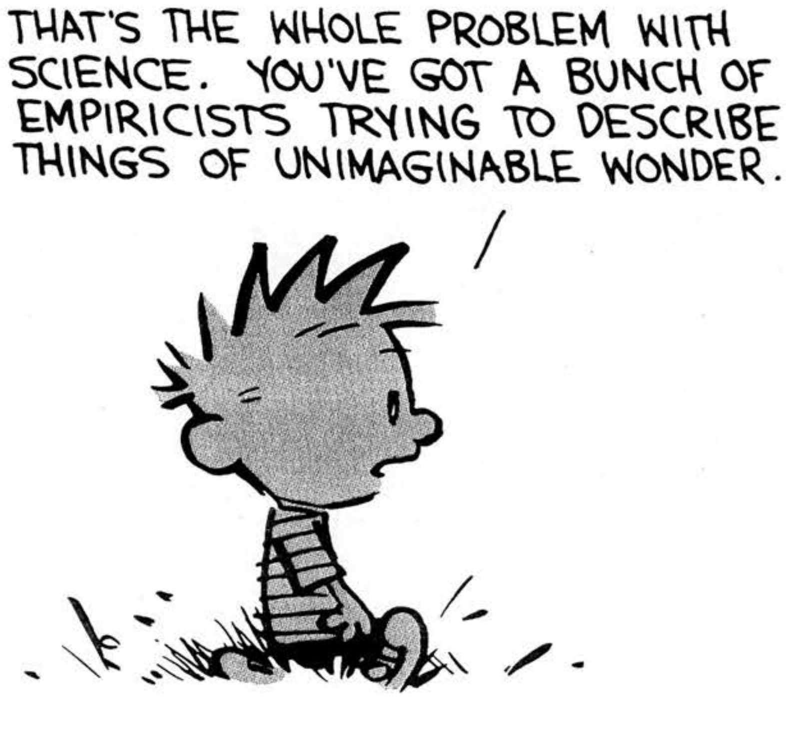
To redefine biology - or geomorphology - in terms of the study of organisational dynamics is to restore its natural context, which carries within it a critique of the established reductionist tendency in science. The weakness of classic reductionist method is that it is both aspatial and atemporal: thereby it denies the complexity of the real world it purports to explain. Thus Goodwin (p. 168) asserts that "[t]he reductionist insistence on some basic material level of cause and explanation... can be recognised for what it is - an unfortunate fashion or prejudice that is actually bad science". This is perhaps too strong: Cohen and Stewart (1994, p. 3), in a similar analysis to Goodwin's, take a less extreme view: "We think [the reductionist story] is right as far as it goes, but it doesn't go far enough". This gives us criteria by which we can judge the utility of reductionist science: it is useful to the extent it is able to identify the key interactions which control the trajectory of landscape development. Progress need not necessarily await further advances in our knowledge of process mechanics; instead, it depends upon the skill with which we are able to tie together critical causal factors within coherent explanatory structures. Contextual relationships replace non-contextual mechanistic relationships as the basis of explanation, at which level the fine details of process can be taken for granted. Harbor's (1992) computer model simulates effectively the development of glacial trough cross-sections using a simple relationship between the square of basal ice velocity and the rate of erosion which is some way removed from the abstract and in-depth mathematical analysis of Hallet (1981), but it is nevertheless consistent with Hallet's reductionist formulations. Hooke (1991) draws on Iverson's (1991b) quantitative analysis of quarrying to propose a *qualitative* model of glacial overdeepening which proceeds by means of a positive feedback involving ice dynamics, hydrology and topography. Similarly, Kirkbride and Spedding (1996) suggest that the presence of overdeepenings can interact with drainage and sediment transfer to establish a particular pattern of moraine formation [Chapters 4, 5, 6 and 7 of this thesis].

To rediscover the context of space-time relationships is simultaneously to retreat from the rigorous, quantitative analysis celebrated by traditional scientific method. The switch towards conceptual, qualitative analysis - at least to begin with - is a central theme of the books by Goodwin and Cohen and Stewart. It fits also with the distinction between the positivist emphasis on explanation as prediction and control, relative to the realist's quest for explanation as understanding (Richards, 1990a; Sayer, 1992). Both realism and complexity theory share an interest in 'emergent' phenomena: the pervasive reproduction of certain robust structures (landforms) within apparently diverse systems (the creation of 'order out of chaos'), the genesis of which is not immediately obvious from the study of the system's individual components. Many moraines are examples of emergent landforms. Delivery of substantial quantities of sediment to the ice margin is necessary to build a large moraine. This implies both that 1) bedrock erosion creates large quantities of debris, and, 2) thereafter, the debris is entrained, and retained within the ice, and not flushed beyond the glacier by the action of meltwater. Variance in moraine size cannot be partitioned between the two factors, with, say, 50% of the variance attributed to debris production, and 50% to debris retention. The two factors are not independent and additive in terms of their causal properties, but are intimately linked within the emergent structure which constitutes the debris transfer system of a glaciated catchment. The debris production and retention factors themselves represent the emergent product of several interdependent variables, including glacier dynamics, bedrock topography and configuration of the drainage network (see previous paragraph).

The extent to which the study of emergence remains consistent with the analytical rigour implied by the realist emphasis on the recovery of 'ontological depth' is likely to depend on the sub-discipline of geomorphology involved. The 'new' fluvial geomorphology succeeds in retaining a strong quantitative, quasi-reductionist element *within* the perspective of a spatially- and temporally-distributed approach by means of its choice of a morphologically dynamic, small-scale, readily accessible study environment (frequently a proglacial meltwater stream), allied with new techniques and instruments such as photogrammetry, digital elevation modelling and the use of electromagnetic current meters (Lane, 1995). Progress in glacial geomorphology, in which process-form interactions evolve much more slowly, and are often concealed beneath the ice, is likely to be far less spectacular in this respect.

In geomorphology, as with geology, it is important to restore the respectability of conceptual, qualitative methods, and so free us from the absurd position of feeling we have to apologise for the interpretative, historical character of our science (Frodeman, 1995). These characteristics follow directly from the fact that geomorphology - *if* we are to define it as the process-led study of landform development - *is* a spatially- and temporally-located science: i.e. it deals with an open system in the realist sense of system closure. Herein lies both its attraction and its strength. Richards (1990a) and Sugden (1996) suggest that geomorphology has perhaps as much in common with the social sciences as it has with the physical sciences. Arguably, human geography became the richer because of its willingness to reinvent itself, to explore beyond its traditional boundaries, and ask new questions within a contextual framework of dynamic socio-spatial organisation (e.g. Dear, 1989; Gregory, 1994). Is now not the time for geomorphology to try the same?

9.3 A FINAL THOUGHT... (Watterson, 1993, p. 101)



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THE END

Endpiece
Sunset from Heimaland, Vestur-Eyjafjallahreppur, Iceland.