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Richard I. Waller
Keele University

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R.I. Waller*

School of Earth and Environmental Sciences, University of Greenwich, Medway University Campus, Chatham Maritime, Kent ME4 4TB, UK

Abstract

There exists within glaciology a widely held assumption that basal sliding and bed deformation do not operate beneath cold-based ice and that their basal velocity is therefore zero, irrespective of bed conditions. Consequently, their ability to erode, entrain and transport sediment and thereby alter the landscape is assumed to be limited. Consequently, very little research has been focused towards describing and understanding the motion of cold-based ice-masses and our knowledge of their behaviour remains poor and the assumption of zero basal velocity, unchallenged.

In this review paper, it is argued that this assumption is not universally applicable and that in certain circumstances, basal processes not only remain active at sub-freezing temperatures, but can significantly influence glacier motion. This is particularly the case in glaciers where sub-freezing basal thermal conditions coincide with the presence of fine grained, ice-rich subglacial sediments. Due to the lack of work undertaken on contemporary cold-based glaciers, much of the information used to support this argument is derived from field research in permafrost areas and on Quaternary glacial sediments, and from the laboratory testing of ice/sediment mixtures.

It is concluded that the assumption of zero basal velocity beneath cold-based ice is overly simplistic and that in reality, the situation is likely to be much more complex. Work is therefore urgently required: firstly, to investigate the circumstances in which basal motion remains active at sub-freezing temperatures and to determine their influence on glacier motion and; secondly, to examine the likely extent of subglacial permafrost in both the contemporary and Quaternary glacial environment.

1. Introduction

It is widely accepted that processes operating at the base of glaciers and ice-sheets exert a controlling influence on their dynamic behaviour (Table 1) (Boulton, 1972; Paterson, 1994). Consequently, considerable research effort has been targeted at understanding the role of basal processes in determining patterns and rates of glacier motion (Raymond, 1971; Engelhardt et al., 1978; Vivian, 1980; Iken and Bindschadler, 1986; Humphrey et al., 1993; Harper and Humphrey, 1995; Iverson et al., 1995). Particular attention has been paid to identifying the processes involved in instigating, maintaining and terminating states of fast ice-flow (Clarke, 1987) associated with surge-type glaciers (Clarke et al., 1984; Kamb et al., 1985; Kamb, 1987; Nolan and Echelmeyer, 1999), ice-streams (Alley et al., 1986; Engelhardt et al., 1990; Engelhardt and Kamb, 1998) and tidewater glaciers (Humphrey et al., 1993; Hart and Smith, 1997).

The main basal processes associated with fast ice-flow are sliding and subglacial bed deformation (Table 1). Initially, investigations focused primarily on ice-masses with rigid beds, such that basal sliding was considered the main cause of fast glacier velocities (e.g. Engelhardt et al., 1979). Over the last two decades, research has increasingly focused upon glaciers with unconsolidated beds and the influence of subglacial sediment deformation on glacier motion (Clark, 1993, 1995). Field investigations of bed deformation at Breiðamerkurjökull, Iceland (Boulton and Jones, 1979; Boulton and Hindmarsh, 1987) and Ice Stream B, Antarctica (Alley et al., 1986, 1987) in particular, indicated that the pervasive deformation of a layer of saturated, subglacial sediments could accommodate the bulk of glacier motion. Most recently, research has demonstrated basal sliding beneath soft-bedded glaciers (Hooke et al., 1992; Iverson et al., 1995; Engelhardt and Kamb, 1998), suggesting that fast ice-flow may be associated with the

*Tel.: +44-181-331-9827; fax: +44-181-331-9805.

E-mail address: r.waller@greenwich.ac.uk (R.I. Waller).

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Table 1
Selection of sites where basal processes have been shown to dominate glacier motion

Glacier name	Basal thermal regime	Bed type	Dominant basal process	Percentage of surface velocity accommodated at or below the bed	Source
Østerdalsisen, Norway	Warm-based	Rigid-bed	Basal sliding	> 50	Theakstone (1967)
Athabasca Glacier, Canadian Rockies	Warm-based	Rigid-bed	Basal sliding	84	Raymond (1971)
Breiðamerkurjökull, sourth-west Iceland	Warm-based	Soft-bed	Bed deformation	88	Boulton and Jones (1979)
Variegated Glacier, Alaska, USA	Warm-based	Rigid-bed	Basal sliding	Quiescent phase = 53 Surge phase = 95	Engelhardt and Kamb (1979)
Glacier d'Argentière, French Alps	Warm-based	Rigid-bed	Basal sliding	c. 70	Vivian (1980)
Urumqi Glacier No. 1, China	Cold-based	Soft-bed	Shear within subglacial "ice-laden drift"	c. 60	Echelmeyer and Zhongxiang, (1987)
Storglaciaren, Sweden	Warm-based	Spatially variable	Basal sliding or shear of thin layer of basal till	85	Hooke et al. (1992)
Ice-stream "B", Siple Coast, Antarctica	Warm-based	Soft-bed	Basal sliding or shear of thin layer of basal till	69	Engelhardt et al. (1998)

intense shear of a thin subglacial layer, rather than the pervasive deformation of a thick layer of sediment (Table 1). Whether such a condition is more closely related to basal sliding or basal sediment deformation is unclear however, as it is frequently difficult to distinguish between the operation of the two processes and therefore establish their relative importance (Savage and Paterson, 1963; Robin, 1986).

Basal sliding and subglacial sediment deformation are frequently assumed to be restricted to areas of the glacier sole where the basal ice is at the pressure melting point and net basal melting occurs. Conversely, ice-masses with basal ice at temperatures below the pressure melting point are usually assumed to display "dry" bed conditions and zero basal velocity (i.e. both basal sliding and bed deformation are inactive). This assumption is evident within the basal boundary conditions prescribed for many ice-sheet models (Paterson, 1994; Payne, 1995). In the absence of these basal processes, cold-based glaciers and ice-sheets are typically considered slow moving (Echelmeyer and Zhongxiang, 1987). This glaciological division between "fast" and "slow" glaciers according to their basal thermal regime (Clark, 1995), has provided the basis for theories concerning surge initiation (Clarke, 1976) and ice-marginal thrusting (Moran et al., 1980; Hambrey and Huddart, 1995).

The perceived inactivity of cold-based glaciers is also evident within the geomorphological and sedimentological literature. Rates of erosion and sediment transport beneath cold-based ice are thought to be extremely low (e.g. Goldthwait, 1960; Dowdeswell and Siegert, 1999), due to the assumed absence of basal sliding, bed deformation and thermally-regulated, debris entrainment mechanisms (e.g. Boulton, 1972; Hubbard and Sharp, 1989). As a result, their influence upon the landscape is thought to be limited, such that pre-glacial landforms are readily preserved (e.g. Kleman, 1994).

A growing body of field evidence suggests that basal processes may remain active beneath cold-based, ice-masses. This evidence has been derived from both contemporary (e.g. Echelmeyer and Zhongxiang, 1987; Cuffey et al., 1999) and Quaternary, glacial environments (e.g. Broster and Clague, 1987; Astakhov et al., 1996; Richards, 2000). The aim of this paper is threefold:

- First, to describe traditional views concerning the dynamic behaviour of cold-based ice-masses and to examine the growing body of evidence that suggests that basal processes may remain active beneath them;
- Second, to investigate the potential causes of basal motion at temperatures below the pressure melting point; and
- Third, to examine the glaciological implications of basal sliding and sediment deformation at sub-freezing temperatures.

2. The dynamic behaviour of cold-based ice-masses

For the purpose of this paper, a cold-based, ice-mass is defined as an ice-mass whose basal ice is at a temperature below the pressure melting point. The occurrence and persistence of cold-based ice is dependent primarily upon the thermal boundary conditions present at the glacier bed. If the amount of heat added to the basal ice by the geothermal heat flux, frictional sliding and ice deformation, is exceeded by the amount of heat removed by conduction up through the overlying glacier ice, the basal temperature of the ice-mass will fall. Once the temperature of the basal ice drops below the pressure melting point, then the glacier becomes cold based. It is important to note that the basal conditions beneath glaciers and ice-sheets often display considerable variations and that they are therefore frequently polythermal, with both areas of warm-based ice and cold-based ice.

It is traditionally assumed that basal processes are inoperative beneath cold-based ice (Paterson, 1994; Payne, 1995). For the purposes of this paper, the term "basal processes" is used to describe the combination of basal sliding and subglacial sediment deformation (see Savage and Paterson, 1963). Basal sliding is thought to be prevented by the creation of strong adhesion between the ice and the bed (e.g. Weertman, 1961; Boulton, 1972). Subglacial sediment deformation is similarly believed to cease for two reasons. First, the lack of basal melting prevents saturation of the subglacial sediments and the attainment of high porewater pressures. Second, the presence of interstitial ice greatly increases sediment cohesion, prevents dilation, and enhances the material's shear strength (Williams and Smith, 1989). Consequently, research investigating the glaciological influence of subglacial sediment deformation has tended to consider its operation solely beneath warm ice (e.g. Boulton and Jones, 1979; Beget, 1986, 1987; Clark et al., 1996). As a result, cold-based glaciers are conceptualised as being frozen to their beds, and to move solely through the creep deformation of the ice-mass itself (e.g. Paterson, 1994). Research investigating the dynamic behaviour of such ice-masses in the High Arctic and Antarctica has tended to corroborate this view (e.g. Holdsworth, 1974; Fisher and Koerner, 1986).

Although limited in extent, there is growing field evidence that directly contradicts this assumption that basal processes cease to operate beneath cold-based ice. One of the most important and informative contributions is provided by Echelmeyer and Zhongxiang (1987) (Table 1). They describe observations made in a tunnel at the base of the predominantly cold-based Urumqi Glacier No. 1 in China. The basal ice in the vicinity of the tunnel displayed a temperature below the pressure melting point ($<-1.75°C$), whilst the bed comprised "ice-laden drift", containing between 21% and 39% ice by weight. Experimental data suggest that in this situation, the subglacial sediment should be significantly stronger than the clean glacier ice (e.g. Nickling and Bennett, 1984), and consequently, its deformation should be very limited. However, Echelmeyer and Zhongxiang (1987) found that over 60% of the glacier's surface motion occurred through deformation of the subglacial sediment, which displayed a hundred-fold reduction in creep strength relative to clean ice at the same temperature. A small but significant amount of basal sliding between the glacier sole and the subglacial drift was also measured (c. $0.5\,\mathrm{mm\,d^{-1}}$ or 5% of total surface motion). Very slow, but detectable rates of sliding have also been measured beneath the Meserve Glacier, Antarctica, at –17°C (Cuffey et al., 1999). These measurements endorse the findings of Shreve (1984), who mathematically predicted non-zero sliding speeds at sub-freezing temperatures.

Waller and Hart (1999) measured the patterns of motion occurring in the basal part of the Russell Glacier, a polythermal outlet of the Greenland Ice Sheet, thought to be cold-based near its margin (van Tatenhove, 1993). They found that very little motion was accommodated by creep, either within the basal ice layer present, or within the overlying glacier ice. Instead, the majority of the total surface motion was accommodated within the basal zone, either as basal sliding or subglacial sediment deformation (Fig. 1). A dramatic increase in the velocities measured during and immediately after a prolonged rainfall event (Fig. 2) was interpreted as indicating that basal sliding in particular was occurring. The increase in velocity at the ice margin may have reflected increased sliding velocities in the

Fig. 1. The basal ice layer at the Russell Glacier; note the movement markers emplaced in a vertical array. The majority of the surface motion was found to occur below the lowest marker.

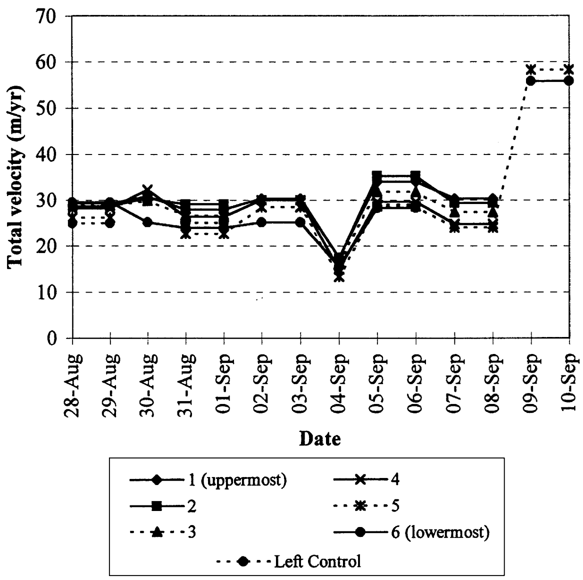

Fig. 2. Total motion of the markers in the basal ice layer during the course of the experiment. Notice the rapid acceleration between 8 and 9 September, caused by a heavy and prolonged rainfall event.

Fig. 3. Example of glacially deformed permafrost from Liverpool Bay, Canadian N.W.T. Note the recumbent fold in the upper part of the massive ice body located just above the ice-axe (65 cm long).

warm-based interior of the glacier, in turn, causing increased deformation of the sediment underlying the cold-based margin.

Evidence for the subglacial deformation of frozen sediments has also been described from Quaternary ice-sheet reconstructions. Russian researchers in particular have investigated the interaction between ice lobes and pre-existing permafrost in Siberia (Kaplyanskaya and Tarnogradskiy, 1986; Astakhov et al., 1996). Astakhov et al. (1996) for example, have described the ductile deformation of extensive areas of permafrozen sediment in Western Siberia, suggesting that the frozen sediment acted as a subglacial deforming bed. Despite occurring in the proglacial environment, the formation of ice-thrust moraine ridges is also suggested to indicate the potential for interaction between cold-based ice and permafrost. Numerous examples have been described around the margins of the Pleistocene, Laurentide Ice Sheet (Kupsch, 1961; Moran et al., 1980; Mooers, 1990). Mackay (1956, 1959) for example, has described the large-scale glaciotectonic deformation of thick permafrost sequences close to the maximum extent of the ice-sheet's north-western margin (Fig. 3). He suggested that at Nicholson Peninsula and Herschel Island, both on the Beaufort Sea coast, thick slabs of frozen sediment were mobilised by the advancing ice, compressed and tilted into an imbricate sequence of thrust sheets. This produced large thrust moraines and, in the case of Herschel Island, an equally-sized depression immediately up-glacier of the moraine from which the thrust sheets were derived. Finally, Kleman and Hattestrand (1999) have recently argued that rogen moraines reflect the brittle fracture of frozen subglacial sediments in response to the generation of high stress concentrations at boundaries between frozen and thawed beds. They can therefore be considered another manifestation of the interactions between cold-based ice and subglacial permafrost.

In summary, field research in both contemporary and Quaternary glacial environments suggests that the assumption that basal processes cease to operate when the temperature of basal ice drops below the pressure melting point, is questionable in certain circumstances. In the following section, the possible mechanisms and conditions associated with active basal motion beneath cold-based ice will be examined.

3. The potential causes of basal motion beneath cold-based ice

It is generally accepted that basal sliding is likely to be limited where cold-based ice rests on a rigid-bed (e.g. Boulton, 1972). Although Cuffey et al. (1999) have measured basal sliding at a temperature of $-17°C$, the rates they recorded are extremely low. Consequently, this section will focus on cold-based glaciers resting on beds of sediment at sub-freezing temperatures and examine the likelihood of this type of substrate deforming in response to glacier-induced stresses. Due to the lack of field research, the viability and potential significance of frozen sediment deformation beneath cold-based ice will be appraised by examining published literature on the rheology ice/sediment mixtures, and conceptual interactions between glaciers and permafrost.

3.1. The rheology of frozen sediment

In order for subglacial sediment deformation to significantly influence a glacier's motion, the substrate must display a lower yield strength than that of the overlying glacier ice. Consequently, under an applied basal shear stress, failure will occur preferentially within the subglacial material (e.g. Beget, 1986; Murray, 1997). Given sufficient time, the subglacial sediment layer will be characterised by a higher strain rate than that of the overlying glacier ice, such that the bulk of the glacier's surface motion is accommodated by subglacial deformation.

Testing of unfrozen subglacial tills with high, saturated porosities, has revealed exceptionally low yield stresses (down to 3 kPa). This is primarily due to the attainment of high porewater pressures balancing the normal stress induced by loading, resulting in low effective pressures and an associated reduction in interparticle contact forces (Boulton and Dent, 1974; Beget, 1986). Consequently, these water-saturated, subglacial sediments display yield stresses significantly lower than that of ice, thereby promoting high velocities at low basal shear stresses and accommodating the majority of ice surface motion (Alley et al., 1986; Boulton and Hindmarsh, 1987) (Table 1).

In general, the shear strength of frozen sediment is significantly greater than that of unfrozen sediment. This is due to the growth of interstitial ice enhancing sediment cohesion and preventing the generation of high porewater pressures to counteract the normal stress caused by loading (Mathews and Mackay, 1960). It is assumed that when frozen, the shear strength of the resultant ice/sediment mixtures is substantially higher than that of glacier ice. In addition, it is also assumed that the transition from a state of comparative weakness to one of comparative rigidity occurs at the pressure melting point. Such an assumption is most appropriate if the subglacial sediments are coarse-grained in nature. Liquid water present within such sediments effectively freezes instantaneously once the temperature drops below the pressure melting point, resulting in a rapid change in their geotechnical properties (Williams and Smith, 1989).

Goughnour and Andersland (1968) tested the geotechnical properties of a sand–ice system and related their characteristics to those of sediment-free, polycrystalline ice. They found that with the exception of sand concentrations between 2% and 3% by volume, the addition of sand increased the shear strength of the ice/sediment mixture relative to that of sediment-free ice, especially when the ice content was similar to the sample's porosity. They suggested that this reflected an increase in sediment cohesion, caused by the formation of an ice matrix. In contrast to Goughnour and Andersland (1968), who performed their tests at –12°C, Nickling and Bennett (1984) investigated the comparative shear strengths of a wide variety of ice/sediment mixtures at temperatures much closer to the pressure melting point (−1°C). Despite the different conditions, both in terms of temperature and calibre of the sediment used (coarse-grained diamicton), their tests produced similar results. As with Goughnour and Andersland (1968), they found that peak shear strength increases rapidly as the ice content rises from 0% to 25% by volume, reaching a maximum value at 25% by volume (equivalent to the sample porosity) and decreasing thereafter. They concluded that ice/sediment mixtures display higher peak shear strengths, due both to the influence of the ice matrix on sample cohesion, and to the increase in internal friction caused by the presence of sediment particles.

Therefore, laboratory-based experiments appear to support the assumption that frozen, coarse-grained sediments are characterised by shear strengths exceeding those of sediment-free, polycrystalline ice, particularly when their ice contents are similar to the sediment's porosity. The geotechnical properties of frozen sediments comprising fine-grained debris are however, markedly different from those containing coarse-grained material.

Fine-grained materials, especially clays, retain significant volumes of liquid water down to temperatures substantially below the pressure melting point. This is due to the ability of water to remain adsorbed onto the surface of the constituent particles as an unfrozen water film, thereby reducing their strength (Mathews and Mackay, 1960; Williams and Smith, 1989). Andersland and Alnouri (1970) investigated the creep characteristics of frozen sand and clay-sized sediment under constant shear stresses and strain rates at –12°C. They demonstrated that at this low temperature, frozen sand displays a much higher resistance to loading than frozen clay. They attributed this difference in behaviour to the clay retaining 12% unfrozen water by volume, whilst the sand was completely frozen at the same temperature. Due to this experiment being performed at very low temperatures (−12°C), such results contrasting the rheological behaviour of frozen sand and clay, may not apply at temperatures closer to the pressure melting point. However, the underlying explanation for this difference, that fine-grained sediments contain significant amounts of liquid water at sub-freezing temperatures, whereas coarse-grained materials freeze very rapidly below the pressure melting point, should be applicable at a wide range of temperatures. Such an assertion is corroborated by the limited field data collected. Echelmeyer and Zhongxiang (1987) suggested that the retention of liquid water within subglacial sediment beneath the cold-based Urumqi No. 1 Glacier layer reduced its creep strength by over one hundred times, relative to ice at a similar temperature and stress level, thereby enabling its rapid deformation. Lawson (1996) has also suggested that at temperatures close to the pressure melting point, the presence of debris enhances the strain rate of ice/sediment mixtures relative to those of debris-free ice. This is thought to reflect the formation of high stress levels around the constituent particles and consequent regelation around them.

In permafrost regions, fine-grained soils frequently contain excess ice, in which the ice content exceeds the available pore space. Both Mathews and Mackay (1960) and Astakhov et al. (1996) have suggested that such ice-rich permafrost is more easily deformable, with areas of excess ice providing potential zones of weakness. The ability of ice-rich permafrost sequences to deform under low applied stresses is illustrated by Dallimore et al. (1996). They measured the in situ creep deformation of a stratigraphic sequence in the Western Canadian Arctic, comprising an 2 m thick upper unit of frozen clay-rich till, overlying a layer of massive ice 22 m thick. They recorded downslope creep of up to $4\,\mathrm{mm\,yr^{-1}}$, with the majority of movement occurring close to the base of the massive ice layer. If creep deformation can occur in response to gravitational-induced stresses, it is highly likely that such a sequence, which is common within the Western Canadian Arctic (Mackay, 1971), will deform beneath an overriding ice-mass. This conclusion is consistent with geological evidence of glaciotectonic deformation in the region (e.g. Mackay, 1956, 1959; Mackay et al., 1972, Pollard, 1990) and in Siberia (Kaplyanskaya and Tarnogradskiy, 1986; Astakhov and Isayeva, 1988; Astakhov et al., 1996).

In summary, the geotechnical properties of frozen ice/sediment mixtures are strongly dependent upon the particle-size characteristics of the material involved. Although the assumption that frozen sediment is characterised by a shear strength in excess of that for sediment-free ice, appears reasonable for coarse-grained materials, its validity with reference to frozen, fine-grained materials is less clear. The concept of particular types of permafrost being more easily deformed than others is described by Tsytovich (1975), who differentiates between *hard-frozen* and *plastic-frozen* ground. He describes hard-frozen materials as those containing little unfrozen water, whilst plastic-frozen materials contain high unfrozen water contents and display relatively rapid creep rates. He also indicates that particle-size determines the thermal boundary between these two rheological states, with it occurring at lower temperatures in finer-grained sediments.

It is suggested that ice/sediment mixtures containing fine-grained materials, significant amounts of excess ice and occurring at relatively warm temperatures, are likely to be the most susceptible to creep deformation at low applied stresses, and the most easily deformable subglacially. This combination of conditions maximises the material's unfrozen water content, and may result in them being weaker than debris-free ice at equivalent temperatures and able to accommodate a significant percentage of a glacier's total surface motion.

3.2. Migration of the effective bed

Conceptually, cold-based glaciers are thought to display zero basal velocities as they are frozen to their beds. The preclusion of basal sliding at the ice-bed interface suggests that cold-based ice will bond effectively to subglacial, permafrozen sediments, allowing the effective transmission of basal shear stress into the substrate. A number of authors have suggested that as a result, cold-based glaciers can mobilise and entrain large bodies of frozen sediment and rock *en masse* (e.g. Mathews and Mackay, 1960; Hughes, 1973; Boulton, 1972; Moran et al., 1980; Eyles et al., 1989; Mooers, 1990; Hambrey et al., 1996; Astakhov et al., 1996; Hambrey et al., 1997). It has been argued that the more efficient transmission of basal shear stress across the ice-bed boundary, in combination with mechanically rigid permafrost, results in a migration of the effective bed (the boundary between glacially mobilised and static material), to the base of the underlying permafrost (Mathews and Mackay, 1960; Astakhov et al., 1996). Consequently, a layer of subglacial permafrost is entrained and mobilised as an intact layer, with little or no internal deformation occurring. Hambrey et al. (1996) have suggested that once entrained, such layers can be elevated into the overlying ice by thrusting, in response to deceleration and flow compression.

It has been suggested that this migration of the effective bed to the boundary between subglacial permafrost and the underlying unfrozen sediments, is likely to be facilitated by the development of high porewater pressures at the permafrost table. The generation of high porewater pressures at the boundary between subglacial permafrost and underlying unfrozen sediment, provides an obvious plane of décollement (Mathews and Mackay, 1960; Menzies, 1981; Mooers, 1990; Boulton et al., 1993). Mathews and Mackay (1960) and Menzies (1981) describe how such high porewater pressures might result from the aggradation of the permafrost table down into the substrate and associated expulsion of porewater ahead of the freezing front. Alternatively, Mooers (1990) and Boulton et al. (1993) have indicated how elevated porewater pressures are promoted at the base of pre-existing permafrost as subglacial drainage is focused along this interface. Subglacial drainage along the base of permafrost, rather than the ice–bed interface, has been demonstrated beneath cold-based zones of the Trapridge Glacier in Canada (Clarke et al., 1984). Consequently, the boundary between frozen and unfrozen material, rather than the ice–bed interface, becomes the dominant zone in terms of both subglacial drainage and basal motion. In this respect, Astakhov et al. (1996) suggest that interaction between the Pleistocene, Barents Sea ice-sheet and pre-existing permafrost, resulted in the entrainment and deformation of up to 300 m of the stratigraphy, as the effective bed was driven down to the base of the pre-existing permafrost. Moran et al. (1980), Eyles et al. (1989) and Mooers (1990) have also described the proglacial detachment of intact sheets of permafrozen sediment and bedrock around the margins of the Laurentide and British ice-sheets, in response to coupling between cold-based ice and pre-existing permafrost.

In summary, it is suggested that cold-based glaciers can couple with permafrozen materials in two different ways. If the permafrozen materials are readily deformable (*plastic-frozen*; Tsytovich, 1975), then this will encourage their pervasive subglacial deformation. In contrast, when interacting with permafrost characterised by comparative rigidity (*hard-frozen permafrost*; Tsytovich, 1975), it is suggested that cold-based glaciers are able to mobilise and entrain the permafrozen, subglacial materials *en masse*, due to the effective transmission of basal shear stress across the ice–bed interface.

4. The significance and implications of basal processes beneath cold-based ice

4.1. Basal boundary conditions

The basal boundary conditions of any ice-mass can be subdivided according to the type of bed present (rigid or soft) and the basal thermal regime (warm-based or cold-based). This provides a four-fold classification of warm/rigid, warm/soft, cold/rigid and cold/soft, which is useful in determining the extent of our knowledge and degree of understanding of these different types of glacier. Table 2 indicates the amount of research performed on each type, along with widely held assumptions of their dynamic behaviour, whilst Fig. 4 schematically illustrates the basal processes associated with each class.

Both types of warm-based glacier have been widely investigated within the literature and our knowledge of their behaviour is comprehensive, although the relative importance of basal sliding and bed deformation in soft-bedded glaciers remains a matter of debate (Boulton, 1986; Robin, 1986; Clark, 1995; Iverson et al., 1995). There are a small number of studies of cold-based, rigid-bedded glaciers, written mostly on small valley glaciers and ice-caps situated in the High Arctic (e.g. Koerner and Fisher, 1979) and Antarctica (e.g. Holdsworth, 1974). Glaciers and ice-sheets characterised by a cold-base and resting on frozen sediments are the least well investigated and the most poorly understood. Despite investigations such as Echelmeyer and Zhongxiang (1987), the revolution in glaciology that has occurred through an appreciation that glaciers can interact with soft beds (Boulton, 1986) has been restricted to warm-based glaciers and has not been extended to cold-based examples. It is still widely assumed that subglacial sediments display two distinct mechanical states, one of rigidity and one of relative weakness, divided by a clearly defined boundary (the pressure melting point). Consequently, despite acknowledgements that frozen subglacial sediment may deform (e.g. Hart et al., 1990), cold-based glaciers with soft-beds tend to be considered in the same light as those with rigid-beds; namely frozen to their beds.

Geocryologists have long appreciated that the assumption that unconsolidated materials are characterised by two highly distinct mechanical states, one of rigidity and one of relative weakness, separated by a clearly defined boundary (the pressure melting point), is a gross oversimplication (e.g. Williams and Smith, 1989). However, many glaciologists and ice-sheet modellers assume that such a bipartite division exists. As outlined earlier, unconsolidated materials in reality display a wide range of mechanical characteristics at sub-freezing temperatures, dependent primarily upon

Table 2
General characteristics of the four main types of glacier

Type of glacier	Research	Movement	Comments
Class 1: Rigid-bedded, warm-based	By far the most studied type of glacier	Relatively fast moving; mostly via basal sliding	The most widely studied type of glacier
Class 2: Rigid-bedded, cold-based	A few reports from Antarctica and Northern Greenland	Very slow moving; only motion via internal creep	Considered very slow moving and geomorphologically ineffective
Class 3: Soft-bedded, warm-based	An increasing number of reports	Possibly the fastest moving; due to rapid bed deformation	Has become the subject of much research since the introduction of deforming bed concept
Class 4: Soft-bedded, cold-based	Very few reports	Slow-medium flow? Possible deformation of the frozen substrate or shear at the subglacial permafrost table	This combination of a soft-bed and a cold-based thermal regime is seldom considered within the literature

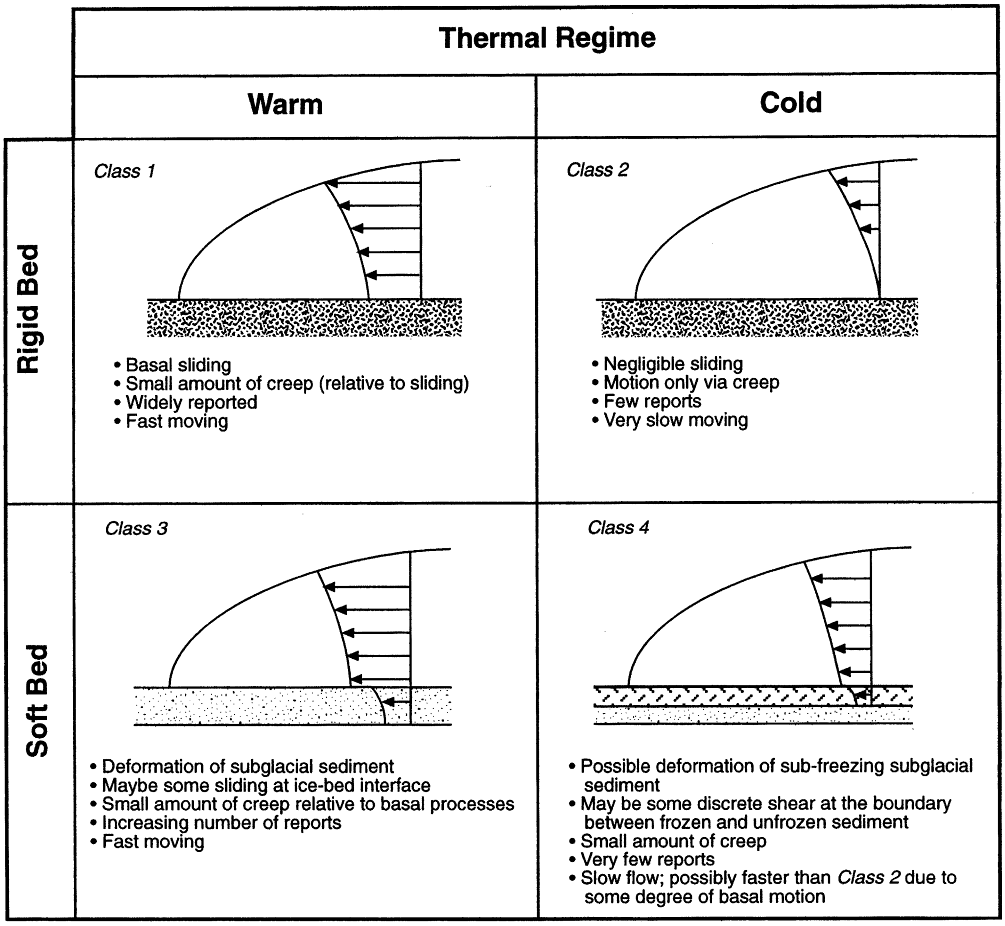

Fig. 4. Diagrammatic representation of the flow profiles and processes thought to characterise different types of glacier.

their unfrozen water content, itself governed by their temperature and their grain-size characteristics (Tsytovich, 1975).

In order to understand and model the motion of cold-based ice-masses resting on frozen sediments more fully, the rheological variability of such sediments needs to be appreciated and incorporated into future models. Increased collaboration with permafrost researchers in general and the utilisation of field data concerning the creep behaviour of permafrost sequences (e.g. Dallimore et al., 1996) in particular, may provide useful data towards this end. In addition, the continued collection of field data, concerning the basal motion occurring beneath contemporary cold-based ice-masses should be considered a research priority.

It is also argued that too much importance has been attributed to the ice/bed interface as the boundary at which movement and water flow is focused. Theoretical studies and field evidence have demonstrated that when a glacier is cold-based and the permafrozen sediment comparatively rigid, the subglacial permafrost table can provide an ideal décollement at which the majority of motion and water flow can be accommodated (Mathews and Mackay, 1960; Clarke et al., 1984; Mooers, 1990; Astakhov et al., 1996). Consequently, rather than preclude basal sliding or deformation, the presence of subglacial permafrost can cause a migration of the effective bed down into the substrate and result in the mobilisation of a layer of permafrost sediment *en masse*.

4.2. *The extent of glacier/permafrost interactions*

It is frequently assumed that the extent of subglacial permafrost is extremely small and therefore that the potential for interaction between glaciers and permafrost is extremely limited. This conclusion stems primarily from numerical modelling results, which suggest that the heat energy generated beneath warm-based ice would rapidly degrade any permafrost overridden (e.g. Sugden, 1977; Beget, 1986; Hindmarsh et al., 1989). Therefore, although some workers agree that subglacial deformation is viable at sub-freezing temperatures, they argue that the spatial and temporal extent of such conditions is limited and by inference, comparatively unimportant (e.g. Hart et al., 1990). As outlined earlier however, there is mounting field evidence which suggests that cold-based conditions were prevalent beneath the continental ice sheets of the Quaternary (Mackay et al., 1972; Moran et al., 1980; Attig et al., 1989; Eyles et al., 1989; Kleman and Hattestrand, 1999; Richards, 2000). Russian researchers in particular have actively argued that subglacial permafrost was extensive beneath the ice-masses covering Siberia during this period (Kaplyanskaya and Tarnogradskiy, 1986; Astakhov and Isayeva, 1988; Astakhov et al., 1996).

For glacier/permafrost interactions to occur, the permafrost must either form subglacially in response to being overridden by cold ice, or exist prior to glaciation due to the influence of subaerial, periglacial conditions. Both Hughes (1973) and Menzies (1981) have described how permafrost may aggrade subglacially. Hughes (1973) also suggests that this *glacial permafrost* was common beneath the Pleistocene Ice Sheets, especially in areas characterised by a combination of thick sedimentary sequences and contemporary permafrost regimes, such as the Western Canadian Arctic.

However, in view of the fact that the initial stages of glacial periods are ordinarily associated with persistent periods of periglacial conditions and permafrost aggradation, the formation of permafrost prior to glaciation seems the most likely scenario. The initial stages of the Anglian Cold Stage are for instance, thought to have been characterised by an intensely cold and dry climate of many thousands of years duration (Jones and Keen, 1993). In combination with the findings of Allen et al. (1988) for example, who argue that permafrost 600 m in thickness can form subaerially under glacial climatic conditions in less than 5000 yr, large thicknesses of permafrost could easily have formed prior to the onset of glaciation. This is especially the case for the more marginal areas of glaciation, where the period of time subjected to subaerial, periglacial conditions is at a maximum.

The basal thermal regime of any ice-mass overriding a low temperature substrate must be affected by its presence, at least temporarily. Heat energy generated at the glacier bed in these circumstances can be conducted both downwards into the permafrost and upwards into the overlying ice, especially if the ice surface temperature and ice thickness is low. Consequently, even if originally warm-based, the overriding of permafrost is likely to promote a shift towards a cold-based thermal regime. Combined with the considerable thermal inertia of thick permafrost sequences, it is argued that subglacial permafrost and glacier/permafrost interactions can be significant both spatially and temporally. In this respect, Mathews and Mackay (1960) calculated that if 400 m thickness of saturated, frozen, unconsolidated sediments, with ground temperatures comparable to the current Mackenzie Delta (c. -10°C) were overridden by a glacier with a basal temperature of 0°C, permafrost c. 100–200 m thick should still persist after a period of 10,000 yr. Recent ice-sheet modelling runs by Cutler et al. (2000) also suggest that both the spatial and temporal influence of subglacial permafrost can be highly significant. Their modelling results indicate that the bed of the Green Bay Lobe of the southern Laurentide Ice Sheet was frozen for 60–200 km upglacier of the margin during the Last Glacial Maximum. They also suggest that when overridden by advancing ice, subglacial permafrost was likely to remain for hundreds to a few thousand years.

It is therefore suggested that glaciologists have tended to underestimate the extent, thickness and thermal inertia of pre-glacial permafrost. It is traditionally assumed that except in the most marginal areas of an ice-mass, pre-existing permafrost is either non-existent or almost instantaneously destroyed as it is overridden by warm-based ice. However, a number of theoretical studies and recent modelling attempts have suggested that permafrost may remain intact beneath substantial areas of an ice-margin for periods up to a few thousand years. Additional field-based investigations are now required to ascertain both the temporal and spatial extent of subglacial permafrost and to check the validity of these assertions.

4.3. The influence of active basal processes beneath cold-based ice

The active deformation of permafrozen sediments beneath cold-based ice has a wide range of potential implications for our understanding of such ice-masses. If the frozen materials are fine-grained, ice-rich and relatively warm, then they may form a comparatively weak subglacial layer, causing either an increase in ice velocity, a reduction in ice surface slope, or a combination of both. In other words, despite existing at sub-freezing temperatures, they may act as a subglacial deforming layer and thereby modify the glacier's dynamic behaviour (Hart et al., 1990; Beget, 1986; Astakhov et al., 1996). In absolute terms, the influence of basal processes beneath cold-based ice are likely to be relatively minor, only accommodating flow rates of the order of up to a few metres per year. Consequently, their operation will not promote the flow velocities demonstrated by fast flowing, warm-based glaciers (e.g. Clarke, 1987). However, as demonstrated by Echelmeyer and Wang Zhongxiang (1987), their contribution to the total surface motion of a cold-based ice-mass, may be extremely important. The presence of subglacial permafrost may also influence the routing of subglacial drainage. Indeed, it has been suggested that the presence of permafrost precludes drainage at the ice–bed interface, which is as a consequence is forced to flow below the permafrost, through the groundwater system (e.g. Clarke et al., 1984; Boulton et al., 1993).

Even if the rate of deformation of frozen subglacial sediments is slow and their impact on glacier motion relatively small, they may still influence the rate of erosion and subglacial sediment fluxes associated with these ice-masses. If, as envisaged by Mathews and Mackay (1960) and Astakhov et al. (1996), the thickness of the glacially deformed layer is very large ($\geqslant 100$ m), then comparatively small rates of movement will result in the advection of large quantities of sediment. In addition, as indicated by Cuffey et al. (2000), given sufficient time to operate, a very slow process can still have a significant influence.

A premise implicit within many ice-sheet reconstructions, is that the presence of striated bedrock or streamlined bedforms indicates the presence of warm-based ice, whilst conversely, the absence of such features and the preservation of pre-glacial forms indicates the occurrence of cold-based ice (e.g. Attig et al., 1989; Kleman, 1994; Kleman and Borgström, 1996). However, as pointed out by Cuffey et al. (2000), the presence of striated bedrock need not indicate the presence of warm-based ice. Additionally, as argued within this paper, cold-based ice is not necessarily associated with an absence of basal processes. Therefore, although some cold-based glaciers are likely to be associated with negligible erosion and sediment transport rates and therefore a limited impact on the landscape, it is oversimplistic to assume that this is universally the case. Consequently, caution should be exercised when attempting to reconstruct ice-sheet basal thermal regimes from the geomorphological record and additional lines of evidence should be employed where possible.

5. Concluding remarks

Despite our poor understanding of the basal conditions and dynamic behaviour associated with cold-based glaciers, there is a growing body of evidence which suggests that basal processes may remain active beneath such ice-masses and exert a significant influence on their dynamic behaviour and sediment fluxes. Firstly, there are a number of reports of the glaciotectonic deformation of permafrozen sediments in the marginal areas of the Pleistocene Ice Sheets. Russian research in particular has indicated how permafrost can be deformed subglacially, especially when ice-rich or fine-grained in nature. Secondly, laboratory data describing the geotechnical properties of ice/sediment mixtures suggests that in certain circumstances, these mixtures may be more easily deformed than pure ice at similar temperatures. Research indicates that the presence of a fine-grained matrix in particular, may facilitate deformation by preserving substantial amounts of liquid water at sub-freezing temperatures, adsorbed onto the surface of the constituent particles. Thirdly, data describing the gravitational creep of frozen massive ice and frozen soils in permafrost environments suggests that the deformation of such material in a subglacial environment where the shear stresses are much greater, is highly likely.

Research is required to investigate the extent to which the deformation of subglacial materials at sub-freezing temperatures can influence the motion of cold-based glaciers. In order to determine the glaciological significance of deformation beneath cold-based ice, work is also required to examine the extent of subglacial permafrost, both in the contemporary and Quaternary glacial environments. The acquisition of such information is vital if we are to correctly interpret and predict the behaviour and stability of cold-based ice and to reconstruct its occurrence in the past.

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