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Subglacial sediment accumulation and basal smoothing – a mechanism for initiating glacier surging

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Abstract

We propose a simple and novel hypothesis for the initiation of glacier surging. Based on realistic assumptions about the basal geometry of valley glaciers in actively eroding mountains, it suggests a mechanism for the varying basal friction required to explain the observed cyclicity of glacier surging, in the absence of similarly cyclic behaviour of inputs such as snowfall or temperature.

The hypothesis proposes that a glacier surge begins when the large-scale resistance exerted on the ice by the underlying, uneven surface is sufficiently reduced by sediment buildup in basal depressions that sliding of the ice on the rock and sediment substrates can occur. Ice flow velocity then increases locally and dramatically. Adjacent areas, not yet ready to slide spontaneously, sequentially increase in velocity due to stress transfer, forcing the ice downvalley to form a surge front. Rapid translation of the surge causes intense crevassing and thrust faulting, disrupting the subglacial drainage system and causing storage of water and sediment behind the front. Increased basal water pressures result to the rear of the surge front, assisting the propagation of the surge front by reducing the shear resistance of the basal sediments and bedrock and increasing the water pressure gradient across the front. When the surge front slows down because it either encounters unfilled basal depressions or has outrun the ice supply from upglacier, a concentrated subglacial drainage system re-establishes and water and sediment stored behind the front are released in a turbid flood, evacuating sediment from the basal hollows. Ice then protrudes again into the redeepened bed hollows, causing sliding to cease and flow of the glacier to slow. Buildup of ice then gradually increases the driving stress again, while sediment again accumulates in the subglacial hollows, until sliding again occurs and a further surge begins.

Introduction

The behaviour of surge-type glaciers has been debated intensively for some time, but to date no convincing hypothesis has been advanced to explain why the flow velocity of these glaciers increases by one or two orders of magnitude, with a periodicity of decades to centuries (Benn and Evans, 1998). Current explanations (see e.g. Menzies, 1995) suggest changes in the subglacial drainage system and basal water pressure, or the occurrence of particular rheologies of ice or till. Neither of these provides an explanation for the long-term periodicity of the processes that *initiate* the change from a concentrated to a distributed drainage system, or from one rheological behaviour to another.

Herein we outline a simple and conceptually robust hypothesis that provides a fundamental explanation for the periodic initiation of surges. This hypothesis explains how the basal resistance to glacier motion can vary cyclically at the required time-scales in the absence of correspondingly-varying inputs to the glacial system. It also explains how surges are initiated and propagate; why they are associated with high basal water pressures and distributed drainage systems, and with intense crevassing of the glacier surface; why the cessation of a surge is accompanied by discharge of large quantities of water and sediment; why surges commonly begin in winter; and why surges are sometimes intermittent. Our suggestion proposes that the change from a concentrated to a distributed subglacial drainage system, and the basal water pressure increase, which are commonly thought to initiate glacier surging (Harrison *et al.*, 1994), are not causes but *results* of the development of a surge front.

Hypothesis – outline

We propose that the relatively steady long-term supply of sediment, from subaerial and subglacial erosion processes, to the basal glacier ice and the subglacial drainage system, results in relatively steady long-term sediment accumulation in basal depressions, reducing the protrusion of ice into these depressions. This causes a reduction in the large-scale unevenness of the contact surface between the glacial ice base and the underlying material. *When the basal profile becomes sufficiently smooth at some point, sliding is able to occur and glacier flow velocity increases correspondingly.* Lateral stress transfer then triggers sliding of adjacent areas of ice, that were not yet sufficiently smooth to slide spontaneously, increasing their velocity. Ice at that cross-section then moves rapidly downvalley, compressing the ice ahead of it and forming a surge front. Thus a surge is initiated.

We present evidence from a non-surge-type but rapidly moving glacier (Franz Josef glacier, New Zealand), to show that sediment commonly accumulates in, and is evacuated from, subglacial storages. We show that sliding is likely to occur only when bed depressions become sufficiently infilled with sediment to occlude the downstream lips of the depressions. We rely on published information to support our suggestions with respect to sediment accumulation rates, basal profiles, mechanisms of sediment accumulation and subglacial drainage system modification. First, however, we summarise the behaviour of surging glaciers.

Surging glaciers – behavioural characteristics

Surge-type glaciers are characterised by a number of behaviours not exhibited by non-surge-type glaciers (Kamb *et al*, 1985; Raymond, 1987; Melvold and Hagen, 1998; Menzies, 1995; Nuttall *et al*, 1997; Hewitt, 1969, 1998). These characteristics provide clues about the processes of surging, and also act as criteria that any acceptable explanation must satisfy.
Briefly, surge-type glaciers behave as follows:

1. Long periods (decades to centuries) of "quiescence" occur during which nothing very spectacular happens. In the upper region of the glacier the rate of ice accumulation exceeds the rate of removal of ice by glacier flow and melting; this leads to a gradual increase of glacier surface elevation there. As this ice buildup proceeds, the flow velocity gradually increases, sometimes by way of a series of mini-surges. It is common for the distal part of the glacier to be stagnant during quiescent periods. Just prior to the commencement of a surge the upper part of the glacier has an elevated surface profile.

2. The onset of the surge (which usually becomes apparent in winter; Menzies, 1995) is indicated by the appearance of a steep, high surge front partway down the glacier. This reflects the high velocity that has developed in one part of the glacier, causing compression of the slower-moving ice in front of it.

3. The surge front and the heavily crevassed ice to its rear move rapidly downvalley. This causes compression and thrust faulting in the surge front, reflected by intense crevassing and till brought to the surface. Lateral shear at the glacier margins results from the rapid translation of the ice body. Water pressures within the surging part of the glacier are often reported to be unusually high at this stage.

4. The upper reach of glacier now has a much higher flow velocity than previously, so the surface elevation begins to fall again some distance behind the surge front. If the surge front reaches the terminus, the glacier advances; otherwise the surge front slows, becomes lower and disappears, and the surge ends. During this period a series of minor surges may

propagate rapidly along the glacier. Surges usually terminate during summer (Menzies, 1995).

5. Unusually large floods of very turbid water and large quantities of sediment are delivered from the terminal portal(s) of the glacier at the end of a surge. Lesser, but still noticeable, turbid floods occur when the glacier motion slows down temporarily during the course of a surge. The entire surge process occupies one to several years, depending on the length of the surge cycle (Menzies, 1995).
6. Following the end of the surge the glacier becomes quiescent. The reduced ice velocity results in gradual buildup of the surface level as snow accumulation exceeds its rate of removal, and after a time interval of the order of decades to centuries the glacier surges again.

Factors influencing surge interval

The length of the inter-surge interval—of the order of decades or centuries—is both a distinct puzzle and a vital clue to the problem. There are three physical components of a glacier: snow/ice, water and sediment. Since periodic surging is clearly related to periodic volume variation, its cause should reflect the ability of the quantities of these components to vary on the required timescale.

The quantities of snow and ice entering the system can vary only in response to the (areally large-scale) meteorological factors that control precipitation and melting, but the surging of glaciers in the same locality is commonly asynchronous, showing that surging is not initiated by these factors. Water in the glacier can vary in quantity due to variations in storage in the glacier drainage system, even with steady input; but not on a time-scale of decades or greater, as en-glacial and subglacial storage is limited and water moves rapidly through the drainage systems of steep glaciers. By contrast, the subglacial sediment present, which is supplied to the glacier at a slow but fairly constant rate by erosion processes in the glacier drainage basin, is capable of accumulating significantly within the glacier over a period of decades. It seems conceivable that variation of the amount of sediment present between the ice and its rock bed could affect the sliding velocity. In the following we thus focus on the effect of subglacial sediment storage variations on the basal resistance to glacier flow. Hindmarsh (1998) reached a similar conclusion for ice sheets from theoretical consideration of the variability of the ice-till-water system.

The general association between surge-type glaciers and relatively erodible sedimentary bedrock lithology (Raymond, 1987; Hamilton and Dowdeswell, 1996) suggests that the supply of substantial quantities of sediment may be necessary to the surging process. Further supporting this idea is relationship between surge period and erosion rate. Dowdeswell *et al* (1991) report that the surge periods of glaciers in Svalbard tend to be much longer (of the order of centuries) than those of glaciers in S.E. Alaska (of the order of decades), while Hewitt (1998) shows that glaciers in the Karakoram Himalaya surge at intervals of 30 – 110 years, intermediate between those of Svalbard and Alaska. Hallet *et al* (1996) list erosion rates in glacial basins in several regions of the world: the rates in Svalbard range vary from 0.1 – 1 mm a^{-1} with an average of 0.38 mm a^{-1}, while those in S.E. Alaska range from 5 to 50 mm a^{-1} with an average of 25 mm a^{-1}. Erosion rates measured in glacial catchments in the Karakoram are about 2 – 9 mm a^{-1} (Hallet *et al*, 1996; Goudie *et al*, 1984)—again intermediate between the erosion rates of Svalbard and Alaska. There is thus a positive regional correlation between basin erosion rates and surge frequencies, again suggesting that sediment supply rate may play a part in determining surge period, and therefore in initiating surges.

Controls on glacier flow velocity

The flow velocity of a glacier results from the balance between the downvalley component of the weight of the ice and the upvalley resisting force resulting from the interaction between the ice and its solid boundaries. If the depth of the ice, or its surface slope, increases, so will the velocity; if the resisting force decreases, the velocity will increase. If the supply of ice from upstream is constant, an increase in flow velocity (such as occurs during a glacier surge) requires a decrease in boundary resistance.

In attempting to understand glacier surging, we need to consider the resistance to ice motion offered by the basal ice-rock or ice-sediment contact, since this is the dominant control on the velocity of warm-bed glaciers. The effect of valley-wall resistance is not considered explicitly herein, but will in principle act in the same way as basal resistance. However, the direct stress on the valley walls due to the weight of the ice will be much less than on the base, with a concomitant reduction in the wall resistance to motion.

Glacier ice can move by internal deformation, controlled by its rheology which is described by a flow law (e.g Glen's law; Weertman, 1957); and it can move bodily by sliding between its lower surface and the rock or till boundary, controlled by the friction between the ice and the boundary. Ice can move past bed obstacles (protuberances or depressions) by pressure melting and regelation, but the size of the obstacles limits the flow velocity that can develop. For example, obstacles greater than about 1 m in vertical dimension distributed at about 4 m spacings limit the ice velocity to less than 1 m a^{-1} (Weertman, 1957). The stresses caused by the presence of distributed obstacles can allow fairly rapid flow of the ice between them, but this is not significant if the obstacles extend laterally across the path of the glacier (Lliboutry, 1968). It is generally accepted that rapid glacier motion requires a major sliding component; because of the high internal shear resistance of ice, rapid internal deformation of ice requires higher shear stresses (> about 1 bar; Weertman, 1957) than are normally found in glaciers. It is therefore to be expected that glacier motion during surges is dominated by sliding.

If the bed of the glacier (assumed to be rock; a till bed is unlikely to exhibit long-lasting unevenness) is very uneven at a large scale (of the order of tens to hundreds of meters), then ice will protrude downwards into bed depressions. Basal cavities (Lliboutry, 1968) will, if present, occupy only the upstream part of the depression and ice will be in contact with the downstream face of the depression. This ice will be unable to move longitudinally, so the ice above the level of the lip of the depression can only move by internal deformation relative to the stationary ice below it (Fig. 1). Its motion will therefore be relatively slow. To achieve high velocity, the ice above the level of the depressions must be able to slide. This will only be possible if the height of the lip of the depression becomes small so that the lip does not obstruct the ice. The base level of the ice can alter in the required way if the depression infills with sediment. Note that the infilling need not be parallel to the glacier surface; if the sediment accumulates as a sloping ramp, ice can slide upwards on this ramp well before the depression is completely infilled (Fig. 2). The exact geometry that allows sliding to occur under specified conditions of ice stresses and rheology cannot be defined, but it is clear that sliding is increasingly likely to occur as the height of the obstacles present in the form of downstream depression lips approaches zero.

Sliding of this sort might take place as motion of the lower ice boundary past the upper layer of sediment; or it might involve shearing of the sediment itself, if shear resistance within the sediment is lower than shear resistance between the sediment and the ice. Resistance to sliding motion is very dependent on the presence or otherwise of water at high pressure, reducing the direct stress and thus friction between the surfaces in contact. If water is present at high pressure at the base of the glacier the sliding friction coefficient reduces significantly, so if the basal contact surface is planar this high-pressure water would allow much faster sliding. Alternatively, an increase in water pressure might allow sliding to occur over a slightly more uneven surface.

Current explanations for increased glacier velocity during surges concentrate on factors that can reduce sliding friction, particularly the effect of increased water pressure in reducing the direct contact stress (and thus

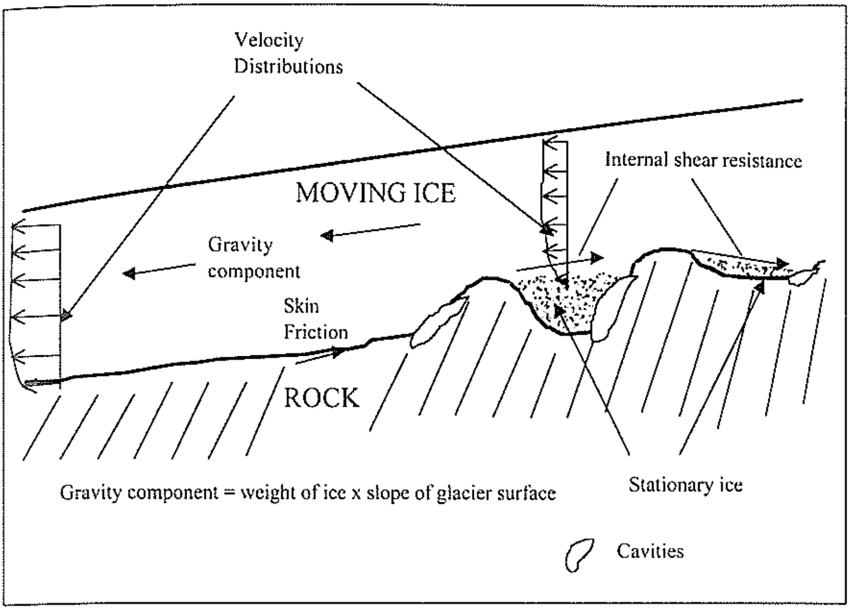

Figure 1 – Forces acting on a glacier over a rock bed.

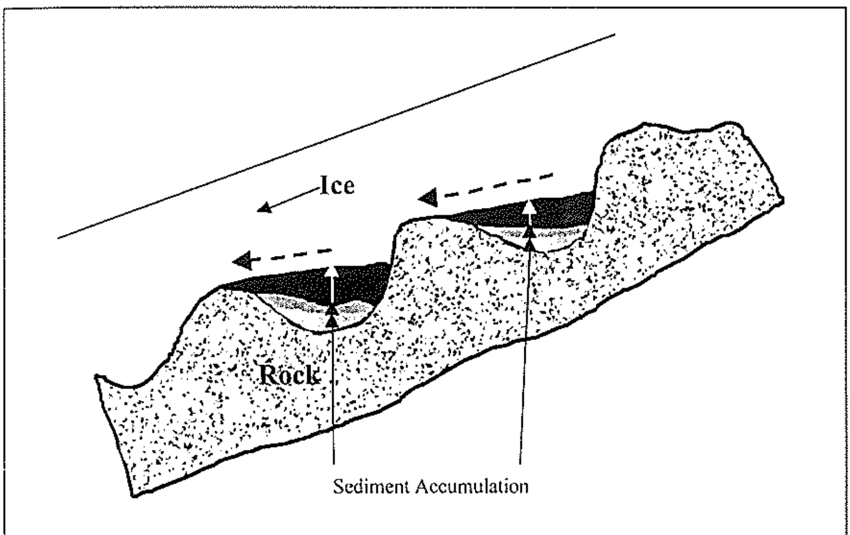

Figure 2 – Sediment accumulation over time develops a lower ice surface profile that allows sliding to occur over hollows (dashed arrows)

the friction force) between the sliding surfaces, and the effect of either increased water content or nonlinear rheology in reducing the internal shearing resistance of ice or till (Hewitt, 1998). By contrast, we focus herein on possible transformations from motion limited by internal deformation, to motion due to sliding, resulting from sediment accumulation altering the geometry of the basal ice profile.

Many surge-type glaciers are found in regions of tectonic activity where faulting is likely to cause very uneven bedrock profiles. Although data on glacier bed profiles are scarce, it is reported that some surge-type glaciers have very uneven basal profiles (Nuttall *et al.*, 1997; Murray *et al.*, 1998), and Flowers and Clarke (1999) describe "small basins" under the Trapridge Glacier in which water—and also, presumably, sediment—can accumulate. Non-surge-type valley glaciers also demonstrate irregular beds with basal cavities capable of accumulating large quantities of sediment. Photographic records of the bed of the Franz Josef Glacier, New Zealand, while at its historic maximum retreat, show that large riegels exist below its present advanced ice surface (Hambrey, 1994). In general it is to be expected that glaciers in mountain valleys will have substantial hollows in their rock beds which, *unless filled by deposited sediment*, will prevent widespread sliding and thus constrain the flow velocity of the glacier to that achievable by internal deformation of the ice.

Subglacial sediment accumulation in basal depressions

Subglacial sediment accumulation and delivery—field evidence

During the 1982–1999 advance of the Franz Josef Glacier, a 10-km long, steep temperate valley glacier on the west coast of South Island, New Zealand, a number of sediment delivery episodes took place that clearly resulted from gradual sediment accumulation in subglacial storages and its subsequent rapid evacuation during rainstorms (Davies *et al.*, 2003). Unusually intense storms discharged large sediment volumes, for example, the $0.25 - 0.5 \times 10^6$ m^3 delivered in December 1995 (Turnbull, 1998). Such volumes are much too large to have been generated by erosion during the individual storms, and much of the debris in the 1995 deposit was sub-rounded and had clearly resided beneath the glacier for some time. Local accumulations of poorly-sorted rounded boulders also suggest that significant sediment sorting had occurred beneath the glacier. It is known from earlier retreat phases that the bedrock profile beneath the Franz Josef Glacier is extremely uneven and contains prominent riegels (Hambrey, 1994, Fig. 4.12).

These events support the concept that a steep valley glacier is able to accumulate large volumes of sediment in subglacial depressions, as required by our hypothesis; and that high glacial drainage flow rates are able to evacuate substantial proportions of this sediment. Franz Josef glacier has a normal flow velocity of the order of 1 m/day; it is not usually thought of as a surge-type glacier, although kinematic waves similar to mini-surges have been observed (McSaveney and Gage, 1968) and its latest (1982 – 1999) advance was rapid. Its relevance to the present discussion is to demonstrate that valley glaciers can accumulate sediment in subglacial depressions, and evacuate it in substantial quantities in high water flows.

Sediment accumulation by fluvial transport and deposition

The motion of sediment at the base of a temperate glacier is to a large extent the result of the flow of water in closed subglacial conduits. The erosion, transport and deposition of sediment by subglacial streams are likely to be complex compared to the equivalent processes in subaerial streams. Water flow in closed conduits can be pressurised, the conduits are subject to the processes of boundary melting and plastic flow under pressure differential between the water and the ice, and the conduits themselves are likely to have reaches of zero and negative slope. Detailed analytical demonstration that sediment transport processes in such conduits will lead to preferential sediment accumulation in bedrock depressions is presently unattainable. However, the application of simple overriding considerations suggests that this will indeed be the case.

The uneven basal bedrock surfaces inferred beneath valley glaciers imply that the subglacial drainage conduits in such glaciers will vary in slope, being steeper on the downvalley sides of local ridges and flatter on the upvalley sides. There is some (unresolved) doubt as to the extent to which a conduit will follow the basal contact in traversing a deep depression (Lliboutry, 1983; Röthlisberger and Lang, 1987; Fountain and Walder, 1998; Hooke and Pohloja, 1994). However the presence of relict N-channels in bedrock (e.g. Hambrey, 1994, Fig. 5.2) indicates that at least some conduits penetrate deep into the depressions, if not to the very bottom. This shows that the gradient variation in the conduits is substantial.

The sediment that can be deposited in conduits is the coarser fraction of the total load, or "bedload"; much of the finer fraction ("suspended load") will be washed straight through the subglacial drainage system without being deposited. The bedload sediment transport capacity of water flow in closed conduits depends on the slope of the conduit, as well as on the pressure gradient driving the flow. It is more difficult for a given flow in a full conduit to move bedload sediment of a given size at a low, or negative, slope than it is at a high positive slope (Graf, 1977) at the same pressure gradient. The bedload transport capacity of subglacial conduits will therefore vary spatially, being greatest where the conduit slope is steep, less where the slope is low and much less where it is negative. This applies whether the conduit is flowing full, i.e. under pressure with an energy gradient greater than the conduit slope, or partly full, with an energy gradient equal to the water surface slope. However, a conduit with negative slope will always flow full. Sediment transported through steeper reaches will therefore tend to deposit preferentially in conduit reaches of low or negative slope (Fig. 3). This has been shown experimentally by Davies *et al* (2003). Shreve (1972) reached a slightly different conclusion—he assumed that water flowing in a subglacial conduit achieves pressure equilibrium with the surrounding ice, and that transport capability is determined only by velocity. While the former assumption may be valid in steady conditions, the latter one ignores the effect of the gravity component of sediment weight reducing the transport capacity in conduits of negative slope.

Sediment is therefore likely to be deposited preferentially in low- and negative-slope reaches of a subglacial conduit system in which bedload is transported. When a layer of sediment is deposited on the bed of the conduit, the cross-sectional flow area is thereby immediately reduced and the longitudinal pressure gradient and flow velocity both increase. These increases will not be sufficient to erode the deposited sediment; if they were, the sediment would not have become stationary in the first place. The increases in pressure gradient and velocity will tend to cause the conduit melt rate to increase, and this melt will take place at the sides and roof because the sediment deposits prevent flow against the bottom ice (if any) of the conduit. The resulting increase of cross-sectional area will reduce the flow velocity again, so that further sediment can be deposited, resulting in further increase in the elevation of the conduit bed and ceiling (and its width) and therefore of the ice base level. This mechanism assumes that sediment can be deposited from a decelerated flow faster than conduit walls can respond by accretion of ice: since the former process is virtually instantaneous, the assumption seems reasonable. Thus in low or negative gradient reaches, which tend to be in the depressions upvalley of basal protuberances, sediment accumulation will tend to cause the conduit to widen and translate upward. This tendency will result in progressive infilling of the depression with sediment to the level of the top of the downstream rim of the depression (Fig. 3).

Lateral migration and spreading of the conduit will cause this effect to extend across the width of the depression (Ng, 2000). From considerations of conduit equilibrium with ice pressure, Hooke (1984) postulated that conduits with negative slope will have wide, flat cross-sections. In this case, sediment will probably deposit asymmetrically on the conduit bed, giving rise to a tendency for the conduit to migrate laterally. This process is exactly analogous to the spread of sediment across the width of a gradually aggrading

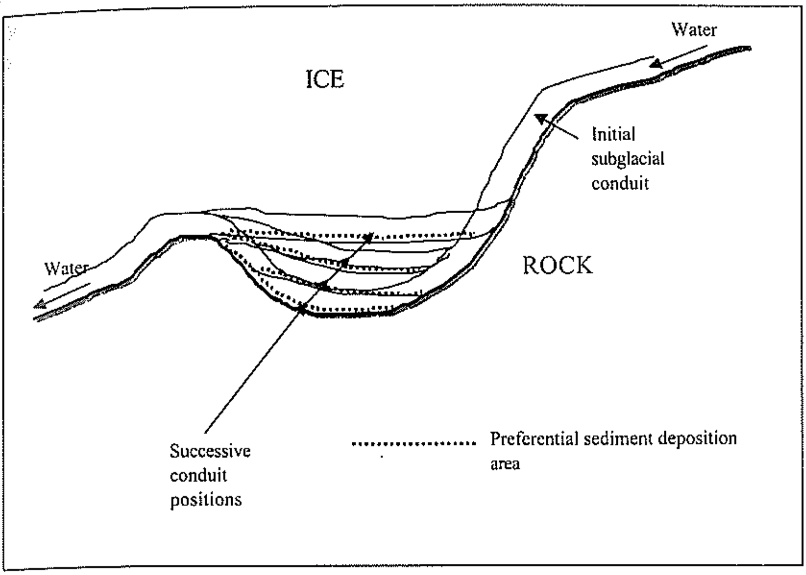

Figure 3 – Upward migration of a subglacial conduit by sediment deposition on its bed in reaches of small or negative slope.

subaerial river or alluvial fan. Thus the increase in elevation of the glacier base will extend across the width of a basal depression. Meltout sediment from the basal ice will enter the conduit as it melts upwards, adding sediment to the flow and increasing the deposition rate.

The upward movement of the conduit will cease when it becomes so elevated that it has a uniform positive slope over the top of the bedrock elevation downvalley (Fig. 3). At this stage preferential sediment deposition will cease and the basal ice surface will be flat instead of undulating. We anticipate that sliding of the glacier will occur somewhat before this state is achieved.

Sediment accumulation rate

The realism of the concept that sediment can infill basal hollows to initiate surging can be tested by an order-of-magnitude sediment volume calculation, using the Variegated Glacier (Raymond and Harrison, 1987; Lawson *et al.*, 1994) as an example. This glacier is known to surge fairly regularly, with a period of about 18 years. The erosion rate of its glacial basin has been found to be about 5 mm a^{-1} (Hallet *et al.*, 1996) from measurements of suspended sediment leaving the glacier in meltwater. The area over which erosion occurs to deliver sediment to the part of the glacier in which surges initiate is about

10 km^2, giving a supply rate of the order of 50 000 m^{3}a^{-1}. This reported erosion rate for the Variegated Glacier is derived from measurements of suspended load leaving the catchment in the glacial drainage water, and so does *not* include sediment deposited beneath the glacier. As an order of magnitude estimate, however, it is reasonable to assume that an equal volume of sediment (50 000 m^3) is deposited beneath the glacier annually (Humphrey and Raymond, 1994). Thus, during an 18-year inter-surge interval, of the order of 9×10^5 m^3 of sediment are deposited beneath the glacier. This is enough to create a roughly 90 cm deep layer over an area 1 km by 1 km, which appears to be the order of size of the area in which surging first becomes apparent.

The bed topography of the Variegated Glacier is not known in detail, and the surveys reported by Bindschadler *et al.* (1977) are too sparse (0.5 km spacing) to be useful at the scale we require. Thus we have to make some reasonable assumptions as to the scale and extent of rock depressions or protuberances. If depressions are present over, say 50% of a 1 km^2 bed area, and are roughly tetrahedral in shape, then the volume of sediment delivered to the bed over 18 years is sufficient to fill them completely from empty if they are about 6 m deep. Thus there is sufficient sediment available for the hypothesis to be reasonable if the average bed depression depth is of the order of 10 m or so, given that the depressions will be by no means completely emptied of sediment following the post-surge sediment-laden flood. This calculation, though necessarily extremely approximate, indicates that the concept of surge initiation by depression infilling is not unrealistic, in that the rate of sediment supply is sufficient to initiate the postulated surging process.

If the hypothesis is correct, then after a surge has halted there should be a large quantity of sediment deposited on the glacier forefield. It has certainly been reported that post-surge flood discharges carry large quantities of sediment; the above order-of-magnitude budget suggests that surges of the Variegated Glacier could yield of the order of 1 million m^3 of sediment. Such deposits would probably be noticeable, but if the degree of depression infilling and evacuation were in fact considerably smaller, which is possible without invalidating the hypothesis, the deposits would not be very conspicuous. Better information on sediment yield, bed topography and surge geometry would allow a more quantitative test of the hypothesis.

Status of hypothesis

There remain many unknowns that prevent more quantitative examination of the present hypothesis. Subglacial bedrock topography is almost completely unknown at the scale required (1 m – 10 m); so are the degree and rate of infilling of bedrock depressions by sediment, and the ability of high drainage flows to excavate them. The rheological and frictional behaviours of debris-rich ice and till are poorly known, so the response of a glacier to depression infilling, though broadly clear, is unknown in detail. Substantial progress is required in these areas before the present suggestion can be either supported or invalidated. The concept is advanced herein as an incentive for further investigation of these topics.

Discussion

The complete sequence of the proposed surging process is illustrated diagrammatically in Figure 4 (over page).

The hypothesis outlined above combines both hard-base and soft-base glacier models; in effect, it suggests that surging results from long-period alternation of the dominance of hard-base and soft-base conditions over an uneven rock bed. With basal depressions only partly infilled with sediment, hard-base conditions dominate and general ice motion is limited by the rate of internal deformation of ice relative to the stationary ice in bed depressions. When the depressions are full of sediment, quasi-planar soft-bed conditions dominate and shearing can occur at the till surface or within the till; thus general sliding is possible and velocity can increase dramatically.

The reported tendency for surging to initiate in winter (Raymond, 1987) seems unlikely to be caused by the presence of high water pressures at the glacier base; one would expect runoff and melt water to be present at higher pressure in summer than in winter. It is in winter, however, that snow accumulates on the glacier surface, increasing the driving stress, and it is in summer that ice velocity and subglacial water flow are greatest, allowing infilling of depressions to occur. The present hypothesis suggests that a surge initiates after a summer in which basal depressions become significantly more filled with sediment, and during a winter in which sufficient snow accumulates to initiate a surge. Slowing and cessation of a surge is likely to occur in summer as the ice mass is reduced due to melting, reducing the driving stress.

In the scenario outlined herein, the observations that basal water pressures increase (Kamb, 1987) and that the subglacial drainage morphology changes from concentrated to distributed during a surge (Kamb, 1987; Björnsson, 1998), are seen as *consequences rather than causes* of surging. The rapid ice motion of the surge will certainly disrupt the subglacial drainage system and reduce its ability to discharge water, as will the compression and faulting of the ice and folding of the basal sediment at the surge front. Water discharge from the subglacial system will therefore decrease, leading to water and sediment storage and increase in water pressure. This will itself tend to

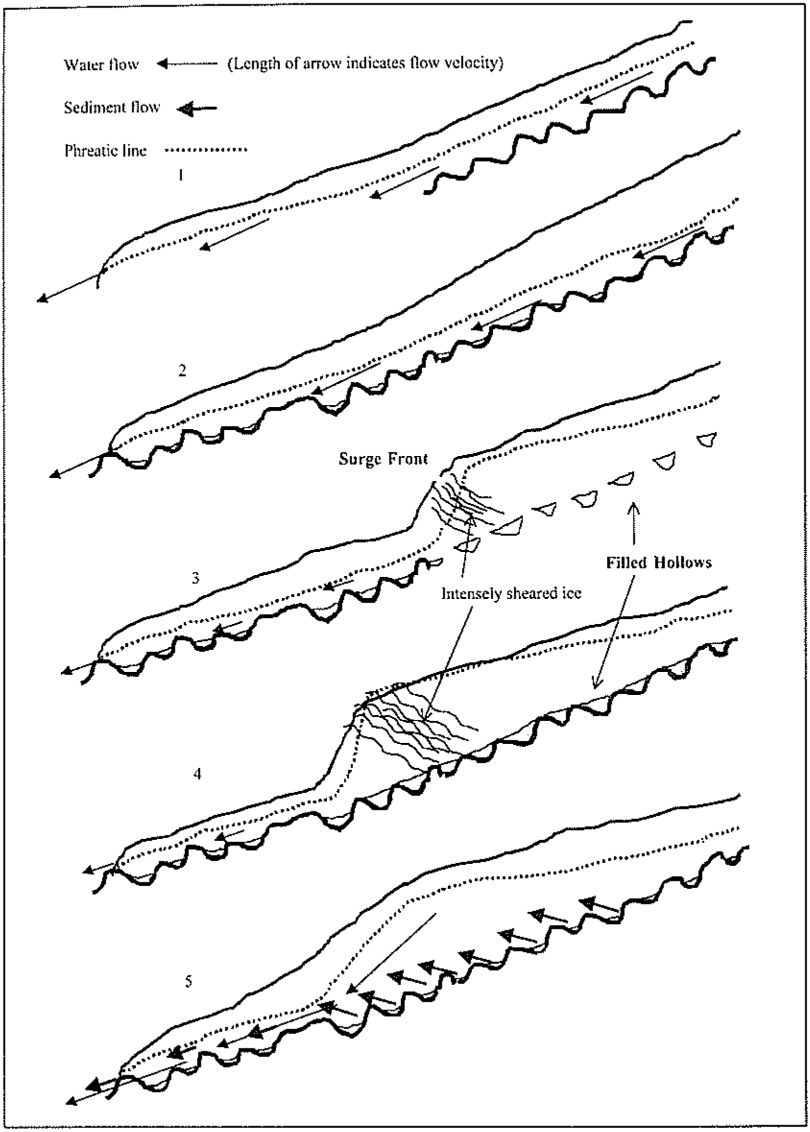

Figure 4 – Diagram illustrating the surge sequence:
1. Beginning of quiescent period: flow is slow, basal hollows are empty, and ice level is gradually rising. **2.** Some years later: basal hollows are filling slowly, ice flow is slow, ice buildup. **3.** Hollows are filled upvalley and fast ice flow results. A surge front develops, and intensely sheared ice blocks internal drainage: there is a steep pressure gradient and high water pressures at and to the rear of the surge respectively. **4.** Surge front moves downvalley. **5.** Surge front slows and flattens, and shearing decreases; the internal drainage system re-establishes, and there is rapid drainage and flushing of sediment under the steep pressure gradient.

increase the surge front velocity, due to the increased hydrostatic pressure gradient across the front. The gradual increase of ice velocity during the quiescent periods is not reported to be accompanied by an increase in basal water pressure; suggesting that water pressure is not the cause of that velocity increase.

Because of compression and active faulting in the surge front, the front is relatively impermeable to water, as conduits are continuously distorted and occluded by ice motion. The front effectively acts as a dam holding back a considerable volume of water and sediment; the easiest path past the surge front for water from upglacier is at the glacier surface. Discharge from the glacier terminus during a surge will therefore be relatively free of sediment, as reported by Raymond (1987), because the water has to move up to the ice surface past the surge front and down again, and will tend to leave sediment behind as it traverses distributed rising conduits in the rear of the surge front (Graf, 1977). As the surge front slows at the cessation of a surge, the disruption to the subglacial drainage system is able to be reversed under the slower ice shear: a concentrated drainage system re-establishes itself under the high water pressure gradients still present, and the stored water and sediment are released in the form of a large and turbid flood, as reported by Harrison *et al* (1994) and Kamb *et al* (1985). This high flow in the subglacial conduits re-evacuates the infilled storages.

The postulated onset of sliding on upwards-sloping surfaces (Fig. 3) causes local increases in elevation at the glacier surface as ice locally moves upwards over rock high points. This will cause the glacier surface to immediately become very uneven and crevassed, as commonly noted in surges, and, if sliding is caused by shearing within the underlying till, will bring till to the surface. Intense general crevassing of the glacier surface indicates rapid motion over remaining unoccluded bed depressions or protuberances.

The progress of a surge, once initiated, will depend on whether or not the bed over which the surge front propagates is itself smooth because its depressions have already been sufficiently filled with sediment. If it is, then there will be little resistance to the increased ice velocity and the surge front will propagate rapidly. If, on the other hand, the surge front develops at the downvalley end of the infilled reach, then it may have to propagate over a series of as-yet unfilled depressions, the basal resistance of which will be high. In this case the surge will propagate relatively slowly, and may even halt, as the basal sediment beneath the surge front becomes stored in the depressions. The propensity for rapid ice movement still remains upglacier, however, and it is inevitable that a further surge will occur which will propagate further. A series of small, short-lived surges (Raymond, 1987) might occur as a precursor to a large surge. A mini-surge front thus generated will translate downglacier as a compression wave but will not accelerate as it does so. Minisurges are interspersed with small releases of turbid water, indicating that the small surge front is capable of retarding the flow of subglacial water and sediment while the front exists.

The initial increase in velocity of the ice at a particular location will not only cause a compression wave to move downvalley as a surge front; it may also generate a rarefaction wave that moves upglacier (Meier and Post, 1969). This will generate a local steepening of the ice surface, which may itself be sufficient to cause rapid ice acceleration if the ice traverses almost-full basal depressions. Thus further mini-surges can be initiated upstream of the primary surge, and the surge can propagate upglacier as well as downglacier (as in the Bering Glacier, Alaska; Herzfeld and Mayer, 1997).

We propose that the mechanism of basal geometry change by sediment accumulation may act as a trigger for the surging process in glaciers whose bed topography is uneven. We acknowledge that surging may be due to other mechanisms in other circumstances; for example if a glacier has a very smooth rock bed then our mechanism cannot apply (e.g. the rapid sliding events in the Balmhorngletscher described by Röthlisberger (1987)). Likewise, some glaciers (e.g. Franz Josef) have bedrock depressions and ample sediment supply but do not obviously surge in the accepted sense of the term, perhaps indicating that sliding is in that case the dominant form of motion under normal circumstances. The usually rapid motion of this glacier (~ 1 m/day: McSaveney and Gage, 1968) encourages this suggestion.

Conclusions

1. Bedload sediment transport capacity is low in low-slope or negative-slope subglacial drainage conduits on the downvalley side of basal depressions. Sediment deposition occurs preferentially in such locations. Basal lodgement and melt processes also deposit sediment in these locations. Sediment can be evacuated from such basal storages by high water flows.
2. When infilling of basal depressions has sufficiently reduced the unevenness of the basal ice contact, sliding occurs and allows higher ice velocities to occur locally and a surge front to develop.
3. Increased glacier velocity and ice tectonics in the surge front, as well as disturbance of the basal till, render the surge front relatively impermeable, causing water and sediment to accumulate upstream of the front. Basal water pressure therefore increases also, causing the surge front velocity to increase further.
4. Water and sediment accumulated upstream of the surge front are released as a turbid front when the glacial drainage system reorganises itself as the surge front slows and halts.

5. Field data support the suggestion that the sediment accumulates in basal depressions at approximately the rate required to initiate surges at time intervals of the order of decades to centuries.

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