Effects of rock avalanches on glacier behaviour and moraine formation

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A B S T R A C T

Although large rock avalanches are infrequent, sediment production in active orogens is dominated by such events, which may strongly influence geomorphic processes. Where rock avalanches fall onto glaciers, they may affect glacier behaviour and moraine formation. We outline the processes of rock avalanche initiation and motion, and show that supraglacial deposits of rock avalanche debris have distinct sedimentological and thermal properties. Laboratory experiments on the effects of such debris on ice ablation are supplemented by field data from two rock avalanches in the Southern Alps, New Zealand. Their effects are compared with those of the thinner supraglacial debris that results from small rockfalls and melt-out of englacial debris. Implications of rock–avalanche debris cover for glacier behaviour are explored using a mass-balance model of the Franz Josef Glacier in New Zealand, demonstrating a likely supraglacial rock avalanche origin for the Waiho Loop moraine, and considering the potential hazard of a large rock avalanche onto the present-day glacier.

1. Introduction

In this paper we examine the effects of the deposits of large rock avalanches on glacier behaviour and moraine formation. Supraglacial rock avalanche deposits on the glacier ablation zone reduce ice-surface ablation, and this can significantly alter the overall glacier mass balance. With a sufficient proportion of the ablation zone covered by rock avalanche debris the glacier mass balance will become significantly less negative, and, as a result, the terminus may advance. This advance may generate a terminal moraine that has no climate-related cause. The much-debated Late-Glacial Waiho Loop moraine, the Franz Josef Glacier, New Zealand (Barrows et al., 2007; Tovar et al., 2008), appears to be the result of such a sequence of events and we use it as an example to demonstrate how a large rock avalanche deposit can reduce ablation causing a glacier to advance and form a prominent frontal moraine.

Rock avalanche deposits on glaciers and their effects have received particular attention only recently. The increasing frequency of recognised historical and prehistorical catastrophic rock avalanche deposits in high mountains areas worldwide (e.g., Post, 1967; Whitehouse and Griffiths, 1983; Hewitt, 1998, 2009; Hermanns et al., 2001; Geertsema et al., 2006; Allen et al., 2011), together with well-defined magnitude–frequency relationships that indicate that very large events contribute more volume over time than do medium-sized ones (Malamud et al., 2004; Korup and Clague, 2009) has established the importance of such events in the sediment delivery and geomorphology of active orogens. Where glaciers are present in mountain valleys, some rock avalanches inevitably fall onto them; however, supraglacial rock avalanche deposits appear to be less frequently reported than those that do not fall onto glaciers. This is partly because of the limited extent of present-day glaciers and also, significantly, because supraglacial rock avalanche deposits are rapidly reworked by glaciers and deposited as moraines whose potential rock–avalanche origins have rarely been considered in the past. By contrast, a rock avalanche depositing into a glacier-free river valley, though progressively removed by river erosion, often leaves long-term recognisable traces of its initial extent in the form of remnants of large angular boulders and vestigial high terraces (Chevalier et al., 2009; Lee et al., 2009). Thus, more likely that the frequency of supraglacial rock avalanche deposits and, hence, their significance to glacier dynamics and behaviour have in the past been underestimated.

Notably, many recently identified rock avalanche deposits, for example those studied by Hewitt (2009) in the Karakoram Himalaya, had earlier been interpreted as moraines and corresponding (incorrect) paleoclimatic inferences drawn (see also McColl and Davies, 2011; Kariya et al., 2011). One of the conclusions of the present work is that prominent moraines can be generated by rock–avalanche-driven glacier advances; these cannot presently be distinguished from moraines reflecting climatically driven advances, and their inclusion in moraine age data may also have led to incorrect paleoclimate interpretations.

In this work we highlight recently-published information about the properties of rock avalanche deposit, their effects on glacier ablation, and glacier response to reduced ablation. We report an
2. Rock avalanches onto glaciers

2.1. Causes

A rock avalanche is a large-scale rock slope failure (of the order of a million cubic metres or more) that leads to a mass of fragmenting rock travelling across a landscape at velocities up to 100 m/s (Voight and Pariseau, 1978; Fort et al., 2009; Hewitt, 2009). Failures of this scale are frequently triggered by major earthquakes, but a seismic trigger is not necessary; for example, the 1991 Aoraki/Mt. Cook, 1991 Mt. Fletcher, 1999 Mt. Adams, and 2007 Young River landslides in the Southern Alps of New Zealand between 1991 and 2007 were all about $10^7$ m$^3$ in volume and neither seismically- nor rainfall-triggered (McSaveney, 2002; Hancox et al., 2005; Allen et al., 2011). Relatively few large-scale rock-slope failures appear to be triggered by rainfall (Melosh, 1987; Soldati et al., 2004). Contrary to convention, recent work suggests that loss of ice buttressing of valley slopes during and following deglaciation may have been overemphasised as a cause of the apparently high frequency of large rock avalanches in the early Holocene (McColl et al., 2010).

2.2. Processes

A rock avalanche usually initiates as the detachment of a large mass of rock from the upper part of a mountain edifice by completion of a through-going failure surface. To generate the highly-fragmented debris described below, the volume needs to be in the order of $10^6$ m$^3$ or more (Davies and McSaveney, 2009). The detached mass initially slides under gravity down the mountainside, progressively collapsing as it does so into joint-controlled blocks. However, at an early stage, progressively smaller intact (i.e., uncracked) rocks will start to comminute (or fragment; e.g., Davies and McSaveney, 2009) under the high stresses of the fall, generating fragments of all sizes to submicron diameter; large quantities of fine particles are expelled from the surface to form a growing aerial dust cloud. By the time it reaches the glacier surface, the mass is pervasively fragmenting and travelling at high velocity. The debris then moves rapidly across the ice surface, spreading both laterally and longitudinally as it does so and becoming thinner in the process. A rock avalanche travelling across a glacier continues to comminate until it halts, as demonstrated by the frequent presence of distally located, shattered but undisaggregated clasts; these are the result of fragmentation and, if formed more proximal, would have been disaggregated by the shear strain field in the avalanche (e.g., McSaveney, 1978). Often evidence is found that the ice-surface has been eroded by the rock avalanche passage, which contributes mass to the emplaced deposit (e.g., McSaveney, 1978, 2002).

2.3. Deposits

Supraglacial rock avalanche deposits are usually much thinner relative to their volume than their nonsupraglacial counterparts, and thus larger in area. This is the result of the lower basal friction of rock avalanches that travel over snow and ice (Post, 1967; McSaveney, 1975, 2002; Gordon et al., 1978; Evans and Clague, 1988; Jibson et al., 2004, 2006; Lipovsky et al., 2008; Deline, 2009; Hewitt, 2009), which allows a given volume to spread over a greater area. For example, the average thickness of the $40 \times 10^6$ m$^3$ Sherman Glacier rock avalanche of 1964 (Fig. 1) was about 1.5 m (McSaveney, 1978), whereas the thickness of the majority of reported nonsupraglacial rock avalanches thicknesses varies between 10 m and 60 m or more (Smith et al., 2006). This characteristic allows a rock avalanche to cover a very large area of glacier with a debris carpet some metres thick.

As a result of the intense fragmentation that occurs during fall and runout, the debris of rock avalanches is predominantly extremely fine. For example, McSaveney and Davies (2007) found that 90% by weight (>99% by number) of the surface debris of the 1991 Aoraki/Mt. Cook rock avalanche, New Zealand, was $< 10^{-5}$ m in diameter. The coarser (metre-scale) “carapace” layer is usually at the debris surface (caused by less intense fragmentation at the surface because of lower confining pressure); so the surface appearance is often blocky, but most of the underlying debris is extremely fine, with larger clasts of all sizes forming a fractal grain size distribution (Fig. 2). The bulk density of the subsurface debris is high caused by the wide grain size distribution, and the voids ratio and permeability are correspondingly low, compared to those of other types of supraglacial debris. As a
result, rock avalanche debris has high thermal mass and inertia that is, it requires a large quantity of heat to increase its temperature, so it heats up slowly compared to debris with large volumes of air-filled voids. This gives rock avalanche debris high insulation capacity, compared to the thin melt-out debris common on valley glaciers, as we now describe.

3. Supraglacial debris: Effects on ablation

The major direct effect of a rock avalanche deposit on a glacier is to protect the ice surface beneath the debris from solar radiation, resulting in reduced surface ablation beneath the deposit (Østrem, 1965; McSaveney, 1975; Lundstrom et al., 1993; Mattson et al., 1993; Nakawo and Rana, 1999; Brock et al., 2010; Reid and Brock, 2010; Reznichenko et al., 2010). The heat flux through the debris layer depends on its thickness and physical properties (lithology, bulk density or porosity, albedo, moisture content, and permeability; e.g., Nakawo and Young, 1981, 1982; Kayastha et al., 2000; Nicholson and Benn, 2006), and determines the ice-surface melt rate. Where a large rock avalanche deposit covers a sufficient proportion of the ablation zone with debris of the order of metres thick, surface-ice ablation will almost completely cease and, as a result, the dynamics of the covered glacier will be significantly affected (Shulmeister et al., 2009). Ablation within or beneath the ice will be unaffected by surficial debris, but this is usually a small component of total ablation (of the order of 10%; Alexander et al., in review).

3.1. Melt-out debris

The ablation zones of many mountain valley glaciers are covered with extensive supraglacial debris. This results mostly from the melt-out of englacially or subglacially transported debris originating from subglacial or extraglacial sources (Rogerson et al., 1986; Winkler, 2009); it is commonly several to a few tens of centimeters in thickness and characterized by angular, coarse clasts with a small proportion of fines and, as a result, high permeability (Drewry, 1986), low bulk density and relatively low thermal inertia. Melt-out debris increases in thickness toward the terminus from lateral displacement and continued emergence of englacial debris and if it achieves sufficient thickness can reverse the increasing rate of clear-ice ablation toward the terminus caused by reduction in glacier surface elevation. The high debris porosity favours air and water circulation, reducing the insulation effect of the debris; for example, Sharp (1949) found that coarse debris 15 cm deep has the same insulating effect on ice as sand 3 cm deep. Debris-covered glacier snouts usually are very rough, so a several-centimetres-thick debris layer forms a mosaic with thin layers of debris, clear patches of the ice, and exposed ice cliffs. Differential melting amplifies the topographic unevenness of the glacier and complicates the estimation of total melting rates under this cover (Fig. 3).

3.2. Rock avalanche debris

The effect of rock avalanche debris on underlying glacier ice is distinctly different from that of melt-out debris owing to its greater...
thickness (meters) and thermal inertia (because of the fractal grain size distribution and high proportion of very fine grains). The effect of high thermal mass is to require a large quantity of heat to be input before the mass can achieve steady-state heat transfer, so this takes a long time. If this time exceeds the diurnal radiation cycle time, steady-state heat transfer to the underlying ice-surface is never achieved (Reznichenko et al., 2010). The major effect of a thick rock avalanche deposit on a glacier ablation zone is thus to dramatically reduce the surface ablation beneath the debris by comparison with adjacent bare or thinly-debris-covered ice (see next section). This leads to thickening of ice (in fact, a reduction of ice thinning) under the deposit relative to neighbouring clear ice, which begins immediately and is often clearly evident a year after emplacement of the deposit. If the deposit thickness exceeds 2–3 m, the underlying ice-surface ablation reduces effectively to zero (Bozhinskii et al., 1986). In this situation the rate of relative ice thickening beneath the deposit equals the surface ablation rate of clear ice at that location.

A rock avalanche deposit also contributes a large mass of sediment to the debris transportation system of the glacier almost instantaneously, whereas the gradual exhumation of melt-out debris cover takes place slowly and quasicontinuously without adding mass to the system.

If the glacier surface is crevassed, part of the rock avalanche deposit may enter the sub- and englacial transportation systems of the glacier (Hewitt, 2009). However, most of the debris remains supraglacial and will eventually be carried to the terminus by glacier flow (e.g., the Sherman Glacier, Alaska; Figs. 1 and 4).

3.3. Insulation mechanisms

In order to study the role of the debris cover in insulating the underlying ice from melting, we conducted laboratory experiments in which ice blocks with clear surfaces and identical sizes covered with either medium sand (representing typical highly permeable melt-out debris) or rock avalanche debris were exposed to identical radiation, both steady and diurnally-cyclic (Reznichenko et al., 2010). These experiments showed that the insulating effect of debris on ice depends on the occurrence of cyclic radiation input. For example, under specific laboratory conditions with continuous, steady-state radiation, sand cover greater than 5 cm thick delays the onset of ice-surface melting by more than 12 h as the sand layer warms up, but after that heat transfer through the debris is steady and ablation rates become similar to those for bare ice. By contrast, when realistic diurnal cycles of radiation are imposed the debris cover continues to reduce the rate of ice melt over the long term (Fig. 5A).

Our study showed that sandy supraglacial cover 5 cm thick (representing melt-out debris) reduced melting to 50% of that with bare ice under same conditions, and 9 cm of sand reduced ablation to 25% (Fig. 5A). Using rock avalanche debris instead of sand under diurnally-cyclic radiation (Reznichenko et al., 2010) increased the insulating effect substantially. The presence of rock avalanche debris also significantly reduced rainfall-induced ice-surface ablation, whereas this effect did not occur with sand (Fig. 5B).

In summary, rock avalanche debris cover on a glacier ablation zone dramatically reduces total ablation beneath the debris; the reduction is much less under normal (melt-out) debris cover.

3.4. Field studies

In order to compare the laboratory results of Reznichenko et al. (2010) with effects in the field, we surveyed rock avalanche deposits from 1991 Aoraki/Mt. Cook and 2004 Mt. Beatrice on the Tasman and Hooker Glaciers, respectively (Southern Alps of New Zealand, Fig. 6). Both avalanches were typical of aseismic slope mass failures in the Southern Alps (McSaveney, 2002; Cox et al., 2008).

Ground penetrating radar (GPR) surveys of these deposits were carried out to measure the effect of rock avalanche debris on the surface-ice ablation. GPR images of two perpendicular profiles along the rock avalanche deposit and adjacent clear or debris-covered ice provided information on the thickness of the rock avalanche deposits, the elevation of the underlying ice surfaces, the modification of the deposits since emplacement, and the interaction of the rock avalanche deposits with adjacent clear or debris-covered ice.

3.4.1. Mt. Beatrice rock avalanche deposit

The 2004 rock avalanche from Mt. Beatrice (2528 m) was emplaced onto the debris-free part of the ablation zone of the Hooker Glacier, close to an earlier deposit that presumably originated from the same source (Fig. 6A). Satellite images show that between 2004 and 2009 the deposit moved about 800 m down the valley, indicating an average ice flow velocity of about 160 m/y at that location, which is similar to previously estimated velocities at the same altitudes (higher than 1200 msl.) on the Tasman Glacier (Quincey and Glasser, 2009).

In December 2009 we obtained the longitudinal and cross-profiles of the Mt. Beatrice deposit shown in Fig. 6A. Profiles revealed that five years after emplacement an elevated ice platform, varying between 20 and 30 m above the adjacent upstream and downstream ice surfaces, respectively, was present under the 3–7 m thick rock avalanche deposit (Figs. 6 and 7), indicating an average clear-ice ablation rate of...
Fig. 5. (A) Example of surface lowering of bare ice and ice under 1, 5, 9, and 13 cm of debris cover under diurnal-cycle conditions. (B) Comparison of the surface lowering of the bare ice, ice under debris cover of sand and rock avalanche material 9 cm deep with diurnal-cycle conditions and diurnal rainfall effect (Reznichenko et al., 2010).

Fig. 6. Location of the two rock avalanche deposits where GPR studies were conducted (red lines indicate GPR profiles, black arrows indicate ice flow directions, satellite image of 2001, LandSat): (A) Mt. Beatrice (2004) rock avalanche deposit on the Hooker Glacier, 2007 (aerial photo by S. Winkler); and (B) Araki/Mt. Cook rock avalanche (1991) deposit on the Tasman Glacier with Hochstetter and Tasman ice indicated (satellite image 2009, ASTER).
The adjacent upglacier, debris-covered ice.

ice-surface ablation.

experimental result that rock avalanche debris dramatically reduces Tasman ice had moved only 500 m (Fig. 6B). On the latter deposit we moved more than 2200 m down the valley by 2009, whereas that on the avalanche deposit into two main parts: that on the Hochstetter ice had after deposition, the Hochstetter ice stream deformed the original avalanche deposit following emplacement was complicated. Shortly after emplacement (McSaveney, 2002). The GPR survey showed the upstream edge of the deposit; this was not present immediately in thickness and a 25-m-high debris-covered ice ridge has formed at the adjacent glacier surface. The GPR data also indicate shearing of the thickened underlying ice owing to the differential ice flow (Fig. 7).

3.4.2. Aoraki/Mt. Cook rock avalanche deposit

The 1991 Aoraki/Mt. Cook rock avalanche, whose volume was originally estimated at 11.8 ± 2.4 × 10⁶ m³ (McSaveney, 2002), travelled across the Grand Plateau, down the Hochstetter icefall, and spread onto and across the Tasman Glacier (Fig. 6B). The motion of the avalanche deposit following emplacement was complicated. Shortly after deposition, the Hochstetter ice stream deformed the original avalanche deposit into two main parts: that on the Hochstetter ice had moved more than 2200 m down the valley by 2009, whereas on Tasman ice had moved only 500 m (Fig. 6B). On the latter deposit we obtained two perpendicular GPR profiles through the avalanche and the adjacent upglacier, debris-covered ice.

The Aoraki/Mt. Cook rock avalanche deposit (Fig. 8) is up to 10 m in thickness and a 25-m-high debris-covered ice ridge has formed at the upstream edge of the deposit; this was not present immediately after emplacement (McSaveney, 2002). The GPR survey showed meltwater channels under the deposit, which can cause local collapses. Therefore, caused by differential ice flow velocities in the deposition area, complications from tributary ice flow and these meltwater channels, the relative ice thickening from reduced ablation is hard to estimate.

Similar ice thickening has been reported on other glaciers, with the formation of platforms up to 30 m above surrounding ice surfaces from reduced ice-surface ablation (Post, 1968; Molnia, 2008). Hewitt (2009) reported that the rock avalanche deposit onto the Bualtar Glacier, Karakoram, increased the surface elevation by 5–15 m in one year by comparison with adjacent ice. Field data, thus, confirm the experimental result that rock avalanche debris dramatically reduces ice-surface ablation.

4. Effects of rock avalanches on glacier dynamics

4.1. Mass balance

The net mass balance of a glacier is determined by ice-mass gain in the accumulation zone and ice-mass loss in the ablation zone. A positive net mass balance over a certain period will cause glacier thickening and frontal advance, while negative net mass balance (i.e. ablation exceeds accumulation) will usually cause the glacier to thin along its profile and the terminus may retreat. Both advance and retreat have response times specific to each glacier. However, in some circumstances (e.g., a long, low-gradient glacier tongue with extensive debris cover), negative net mass balance can cause vertical surface lowering while the terminus remains stationary, as with the downwasting Tasman Glacier (Kirkbride, 1993).

Two main changes in the glacier system result from the emplacement of a rock avalanche deposit on the ablation zone surface:

(i) the net mass balance becomes less negative or more positive by reduction of ice-surface ablation in the ablation zone, assuming there is no change in accumulation rate; and

(ii) additional mass is provided (a) immediately, by the rock avalanche deposit, and (b) gradually, by the increasing relative ice thickness caused by reduced ablation under the rock avalanche deposit.

Rock avalanche debris deposited in the glacier accumulation zone rapidly becomes buried by snow and ice (and entrained in the englacial transport pathway), so it only affects the ablation rate for a short time. It will most likely emerge in the ablation zone as melt-out englacial or subglacial debris after being transported and dispersed in the en/subglacial transport pathways.

The reduction of total ablation under a rock avalanche deposit can be large (e.g., Reid, 1969; McSaveney, 1975; Reznichenko et al., 2010; Alexander et al., 2011). Ice-surface ablation usually reduces to close to zero beneath a rock avalanche deposit. While sub- and englacial ablation by meltwater heat flow, heat conductivity and other processes will be unaffected by the surface debris; they contribute only in a minor way (~10–20%; Alexander et al., in review) to total ablation. Thus, if a sufficient area of the ablation zone is covered, the glacier net mass balance can be significantly altered. Because the
clear-ice ablation rate increases toward the terminus (Schytt, 1967) and the rock avalanche deposit will be progressively carried down-glacier, the reduction in surface ablation per unit area will increase as the debris approaches the terminus. At the same time, the upglacier limit of the debris cover is moving downvalley and clear-ice ablation thereby increasing, so the net variation of mass balance with time is not simple. For the present we note only that when the rear of the debris cover has advected downvalley of the original terminus position, the glacier mass balance is re-established and the debris-covered ice downvalley thereafter behaves independently of the glacier.

4.2. Glacier response to positive mass balance

A rock avalanche covered 50% of the ablation zone of the Sherman Glacier to an average depth of 1.5 m during the Alaska earthquake of 1964, and ablation beneath the deposit was reduced by about 80% (McSaveney, 1975). The resulting alteration in the net mass balance changed previous glacier retreat to steady-state conditions and, after some years, a slow advance became evident (Post, 1968; McSaveney, 1975), which still continued in 2009 (M.J. McSaveney, GNS Science, PO Box 30368 Lower Hutt, NZ, pers. comm. 2009). A rock avalanche onto the Bualtar Glacier, Karakoram, in 1986, caused substantial relative ice thickening under rock avalanche debris covering only 15% of the ablation zone, but the positive impact on mass balance was equivalent to a 20% increase in annual accumulation (Hewitt, 2009). As a result, the Bualtar Glacier (which had been retreating for several decades) surged during the first 2 years after the rock avalanche and continued to advance during the following 12 years.

The mass added to the glacier both immediately (rock avalanche debris) and gradually (relative ice thickening) will tend to increase the glacier flow velocity. Firstly, the rock avalanche material itself is equivalent to about double the weight of the same volume of ice (Shulmeister et al., 2009). Secondly, the relative thickening of ice under the deposit is added to the normal (ice) mass of the glacier. For example, the area of glacier covered by the Mt. Beatrice rock avalanche 5 years after emplacement has the equivalent of about $2 \times (1.4 - 2 \times 10^5)$ m$^3$ additional ice weight from the rock avalanche deposit plus about $13 \times 10^5$ m$^3$ of ice (if the average additional ice thickness is 27 m), which is increasing annually. Totally it is equivalent of an extra $17 \times 10^5$ m$^3$ of ice on an area of about 5 ha, or 34 m of ice depth. This additional mass must alter the glacier dynamics by increasing basal sliding, and this will occur in proportion to the percentage of the ablation zone affected and in inverse proportion to the original ice depth.

The sliding velocity of a warm-based valley glacier is proportional to the square of the bed shear stress, which depends on the ice-surface slope and the mass of the ice per unit area of bed (Paterson, 1994). Thus 10% additional ice depth should increase flow velocity by 20% at a given location. In addition, extra sediment finding its way to the glacier base via crevasses and moulins may affect the subglacial topography and drainage and hence the basal water pressure, which in turn affects sliding (Turnbull and Davies, 2002; Davies and Smart, 2007); however, the magnitude of this effect is difficult to determine. This effect was first observed after the St. Elias earthquake by Tarr and Martin (1914), who proposed the “earthquake advance theory” when seven glaciers were surging and advancing 10 years after the event. The advances were attributed to added mass from rock, snow, and ice avalanches, and also to the uplift of the St. Elias Mountains by 14 m during the earthquake and the corresponding increase in glacier gradients (McSaveney, 1975).

The addition of sufficient rock avalanche debris to a glacier evidently can have significant effects on both mass balance and glacier motion (Shulmeister et al., 2009). An equilibrium glacier may advance; a retreating glacier may retreat more slowly, or come into equilibrium, or advance; and an already advancing glacier may

Fig. 8. GPR profile along the 1991 Aoraki/Mt. Cook rock avalanche and adjacent debris-covered glacier surface in February 2009, the Tasman Glacier, Southern Alps, NZ. The thickness of the rock avalanche deposit is up to 10 m, and the relative ice thickening at the upglacier edge of the deposit is about 25 m. (A) The ridge formed at the upvalley edge of the deposit with the Hochstetter icefall in the left background; and (B) GPR measurements on the upvalley slope of the Aoraki/Mt. Cook rock avalanche deposit.
advance more rapidly. A surge-type glacier may surge prematurely (Hewitt, 1988).

In case of the Mt. Beatrice rock avalanche onto the Hooker Glacier, no changes in glacier behaviour have been observed, because the area covered by the deposit was small proportionally to the ablation area. Shulmeister et al. (2009) estimated that glacier behaviour would be altered by a rock avalanche deposit that covered more than about 10% of the ablation zone area. With smaller deposits, the glacier may experience local effects from the additional supraglacial load, but the overall mass balance will not be significantly affected. The Mt. Beatrice rock avalanche covers about 0.05 km² of the roughly 7 km² of the ablation zone for the Hooker Glacier, which is proportionally <1% of the area (Fig. 6A). The Aoraki/Mt. Cook avalanche deposit covered about 1.7 km², <4% of the ablation zone of the Tasman Glacier (about 45 km²; Kirkbride, 1989). The deposit has since been redistributed and compressed and now covers an even smaller area (Fig. 6B). Therefore, these two deposits are not expected to noticeably affect the overall behaviour of their glaciers.

4.3. Modelling

The effects of rock avalanche debris on glacier behaviour can be predicted quantitatively. Alexander et al. (2011) developed a simple steady-state mass balance model, calibrated on the modern (1999–2010) quasisteady-state conditions of the Franz Josef Glacier and successfully proven against its Little Ice Age terminal moraine positions and trim lines. This was used to test the hypothesis of Tovar et al. (2008) that a catastrophic rock avalanche could have caused the advance of the Franz Josef Glacier (western Southern Alps, New Zealand; Fig. 9) to the location of the Late Pleistocene/Early Holocene Waiho Loop moraine; and thus have formed the moraine. The modelling shows that a steady-state terminus position at the

![Diagram showing modelled rock avalanche debris cover on the Franz Josef Glacier, NZ, showing percentage of ablation reduction, which is staggered below the ELA to a terminus position at the Waiho Loop, assuming temperatures 2 °C less than the present-day, with an ELA of 1480 masl (satellite image of 2001, LandSat). The staggered levels of ablation reflect the accumulation of large quantities of debris toward the terminus, while in the upper-ablation zone, ablation rates will return to those of clean ice once the debris has moved downvalley. Location of the Franz Josef névé is also shown in Fig. 6.](image-url)
Waiho Loop requires mean temperatures about 4–5 °C cooler than present-day. Anderson and Mackintosh (2006), using an ice flow/mass balance model, found that an advance to the Waiho Loop required 3–4 °C cooling from present-day. The results from both studies contradict proxy data, which indicate temperatures no more than ~2.5 °C cooler than present-day during the last glacial-interglacial transition (LIGT) about 12,000–13,000 YBP (Carter et al., 2003; Tovar et al., 2003; Barrows et al., 2007). Direct dating of the Waiho Loop moraine by Barrows et al. (2007) yielded an age of about 10,500 YBP, corresponding to significantly warmer temperatures than those that occurred during the LIGT and making a climatically driven advance even less likely.

To explain the advance of the Franz Josef Glacier from an equilibrium position close to Canavan's Knob to the Waiho Loop position during a time without any cooling (but ~2 °C cooler than present), Tovar et al. (2008) suggested, based on the sedimentology of the Loop moraine, that the ablation zone of the glacier was covered with debris by a rock avalanche. Alexander et al. (2011) showed that 65% reduction in total ablation (or a gradual reduction from 20 to 80% along the glacier) is needed to cause the required 3 km of advance from near Canavan's Knob to an equilibrium position at the Waiho Loop moraine (Fig. 9), assuming constant climate; based on the above data this appears reasonable. A coseismic rock avalanche of the required volume in the Waiho valley is certainly possible; Tovar et al. (2008) provided evidence of sufficiently large-seated, source-area scars in the correct location to match the dominant Loop lithology. Similar events have been reported elsewhere (e.g., Post, 1967; McSaveney, 1975, 2002; Gordon et al., 1978; Evans and Clague, 1988; Orrombeli and Porter, 1988; Gardiner and Hewitt, 1990; Jibson et al., 2004, 2006; Lipovsky et al., 2008; Deline, 2009; Hewitt, 2009).

Vacco et al. (2010) recently presented a one-dimensional numerical model of the Franz Josef Glacier. Though approximate, it confirmed that a debris driven advance could reach the Waiho Loop under certain circumstances and, significantly, that the resulting deposition would have a prominent, high end moraine with an extensive blanket of ablation moraine to its rear, as appears to be the case at the Waiho Loop based on available evidence (Shulmeister et al., 2010).

In summary, there is substantial evidence that a rock avalanche deposit covering a sufficient proportion of an ablation zone can cause glaciers mass balance fluctuations and its nonclimatically driven advance.

5. Consequences

5.1. Terminal moraine formation

Here we describe theoretically how a rock avalanche may cause glacial advance and frontal moraine formation without any hindrances. A rock−avalanche-driven glacier advance will be short-lived; perhaps in the order of decades, compared to centuries to a millennium in the case of an advance driven by climate such as the Younger Dryas or the Little Ice Age. As the rear of the debris deposit is carried downvalley by ice flow, ice-surface ablation will increase and mass balance will gradually become less positive. It will return to normal conditions, or the mass balance as before the event, when all the debris has been carried past the original terminus. Then the advance of the detached debris-covered ice will eventually slow and stop, during which time substantial quantities of rock avalanche debris will reach the new terminus and form a moraine. The sequence of events is illustrated conceptually in Fig. 10: ice covered with rock avalanche deposit reduces in elevation less slowly because of suppressed melting than the previous clean or thinly-covered ice, so it flows more rapidly. In addition, ice at the terminus has compressive flow which is less rapid than that farther upvalley, so the length of debris-covered ice reduces and its thickness increases accordingly.

Eventually the original glacier ablation zone is debris-free because all the debris has been carried beyond the original terminus. The debris-covered ice downvalley is now independent of the glacier and continues to flow slowly under its own surface gradient while slowly melting, forming a terminal moraine; it eventually becomes stagnant and slowly melts to form a sheet of ablation moraine to the rear of the terminal moraine. This form is similar to that generated by the model of Vacco et al. (2010), and is strikingly similar to that displayed by the small Classen terminal moraine in New Zealand shown in Fig. 11.

Supraglacial rock avalanche debris is normally between 1 and 10 m thick; this allows estimation of the size of a rock avalanche required to cover a specified proportion of the ablation zone of a given glacier and suppress ablation accordingly. For example, the Franz Josef Glacier ablation zone area is about 6 km², so a rock avalanche of the order of 10⁶ m³ in volume would be needed to cover 10% of the ablation zone to a depth of >1 m and cause a significant change in glacier behaviour. Such a volume of rock avalanche deposit arriving at the approximately 0.4 km wide terminus could form a moraine of cross-sectional area ~10³−10⁴ m²; this would certainly be a prominent feature in the landscape.

How moraines are formed by climatic glacier advances is also incompletely understood. Bulldozing is able to generate small “push” moraines of the order of several metres high (Molnia, 2008; Winkler and Matthews, 2010), while sluicing of debris by subglacial water flow generates characteristic stratified “outwash” moraines (Hyatt, 2010); however, “dumping” of supraglacial debris, which can build very high and steep lateral moraines (Benn and Evans, 2010) seems to be the most likely potential cause of very large, steep-fronted terminal moraines such as the Waiho Loop (Vacco et al., 2010), and the requirement for large volumes of supraglacial debris to do this suggests that a rock avalanche may be the most likely source of such moraines.

![Fig. 10. Response of a glacier to rock avalanche debris covering the ablation zone:](image)
Thus, a rock sediment able to be delivered to the moraine, the larger it will be. This will remain after retreat has begun. The greater the volume of debris that undergoes an advance-retreat sequence and is supplied with the glacier terminus to form a large moraine in a short time, whereas to form the same size of moraine without rock avalanche debris would require the terminus to remain in one position for a relatively long time.

In the same way, moraines left by surge-type and tidewater glaciers may record nonclimatic ice-front fluctuations (Yde and Paasche, 2010). Therefore the terminal moraines that are used as paleoclimatic indicators must be known to be generated by climatic variation rather than by nonclimatic processes. To date, however, no such investigations of this topic have been reported, and all terminal moraines are assumed de facto to result from climatic variations. Given the crucial contemporary importance of paleoclimatology in understanding climate change, this is potentially a serious deficiency. The various difficulties outlined herein can only be resolved when the processes of formation of both climatically driven and rock–avalanche-driven moraines are much better understood, so that criteria can be developed for distinguishing them in the field.

Given the potential for rock avalanches to generate terminal moraines, improved understanding of moraine formation processes by rock–avalanche-driven advances is an urgent requirement.

### 5.2. Paleoclimatology

Terminal moraines are commonly used as paleoclimatic indicators. This method is based on the assumption that all moraines owe their existence to climatically driven advance-retreat sequences, but our work above clearly shows this assumption to be potentially in error. Any purely rock–avalanche-generated moraine whose date is assigned paleoclimatic significance generates an error in the inferred paleoclimate record. Of course, a rock avalanche can fall onto a glacier that is in the course of a climatically driven advance-retreat sequence, in which case the moraine age will have paleoclimatic relevance.

Terminal moraines form when a glacier reaches an advanced position; when it retreats again the moraine records the previous maximum extent (Benn and Evans, 2010). If a glacier terminus undergoes an advance-retreat sequence and is supplied with sediment by any process, a terminal moraine is likely to form that will remain after retreat has begun. The greater the volume of sediment able to be delivered to the moraine, the larger it will be. Thus, a rock–avalanche-driven advance will be capable of generating a substantial terminal moraine because it involves a very large volume of debris. Such a moraine will have no necessary association with any climatic change (Larsen et al., 2005; Tovar et al., 2008; Shulmeister et al., 2009). Even if a rock avalanche deposit does not cause a glacier to advance, the eventual arrival of a large volume of supraglacial debris at the terminus can form a large moraine in a short time, whereas to form the same size of moraine without rock avalanche debris would require the terminus to remain in one position for a relatively long time.

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### 5.3. Effects on society

The potential for large rock avalanches to affect some present-day glaciers may result in significant impacts on society and business, because of the increasing level of tourism-related human activity in the vicinity of accessible glaciers (which are often the larger ones that extend to low altitudes). A good example is the Franz Josef Glacier because of its location in a tectonically active setting that frequently experiences intense seismicity (Shulmeister et al., 2009) and because of its popularity as a tourist destination owing to its easily accessible terminus (at ~300 masl).

Catastrophic landslides have previously been caused by earthquakes on the Alpine fault, a 700-km-long plate boundary fault running along the western rangefront of the Southern Alps of New Zealand (e.g., Chevalier et al., 2009; Dufresne et al., 2010). This fault generates earthquakes of M–8 several times per millennium on average, and the last one was in 1717 according to paleoseismic information (Rhoades and Van Dissen, 2003). The large number of deep-seated landslide source-area depressions are observed in the Southern Alps, whose morphology indicates that their origin was most likely coseismic (Tovar et al., 2008). Shulmeister et al. (2009) suggested that a large rock avalanche could be expected in the order of once per millennium on average in the Franz Josef Glacier catchment. However, given that since 1999 six rock avalanches have occurred in the Southern Alps, all of the order of 107 m3, and none were triggered by either earthquake or rain (Aoraki/Mt. Cook and Mt. Fletcher I and II; McSaveney, 2002; Mt. Adams; Hancox et al., 2005; Young River; un referenced; Joe River; un referenced), the probability of any kind of rock avalanche in any specific mountain valley is clearly nontrivial.

The ever-present and immediate hazard from a rock avalanche onto the Franz Josef Glacier (and perhaps running out along the proglacial upper Waiho valley) is to the lives of people on the glacier and in the valley (several hundred visitors at any one time on a fine day), since warning of such an event may only be a minute or so for people at the terminus and less for people on the glacier. This therefore represents a not insignificant threat to peoples’ safety, given the ~250,000 tourists that visit the valley, terminus, and ice surface annually (Corbett, 2001).

To estimate the potential effect of a rock avalanche on the present-day glacier, Alexander et al. (2011) assumed that debris from a rock avalanche in the mid-upper catchment is voluminous enough to cover the entire present–day ablation zone (area about 6 km2) to metre-scale depth. This would require a volume of about 107 m3. If total ablation was reduced by 80% as a result, the modelling indicates that the Franz Josef Glacier would advance almost as far as the township of Franz Josef in perhaps a decade; human life would not be threatened by such an advance, but the main highway along the west coast of South Island would be. The consequent effects on the behaviour of the already troublesome Waiho and Callery rivers (McSaveney and Davies, 1998; Davies and McSaveney, 2001) would also need to be investigated—when the Franz Josef Glacier advanced about 1.5 km between 1982 and 1999, the proglacial area aggraded of the order of tens of metres. Clearly, there is the potential for major disruption of tourism and other business in this iconic tourist destination.

This scenario has not to date been part of any assessment carried out for the area, and, therefore, no management strategies are developed. A similar threat could exist for the locality of the Fox Glacier, 20 km south of the Franz Josef Glacier, whose glacial and geomorphological characteristics are similar. Evidently, if a rock-
avalanche-generated glacier advance were to approach Franz Josef township, the carefully planned, ordered, and structured relocation of the settlement and of the highway are required. No active mitigation methods appear to be practicable. The requirement for further investigation of this scenario is clear.

6. Conclusions

(i) Large rock avalanches falling onto glaciers can affect glacier behaviour significantly by reducing the total ablation beneath the deposit by up to 90%, causing substantial changes in glacier behaviour such as advances and surging.

(ii) Rock–avalanche-driven glacier advances can form prominent terminal moraines.

(iii) Terminal moraines generated by rock–avalanche-driven advances have no necessary climatic association; therefore, the validity of paleoclimatic inferences from moraines depends on identifying the origin of the moraine.

(iv) Further investigation is required into the processes of terminal moraine formation caused by rock–avalanche-driven and climatically driven advances and into criteria for distinguishing moraines formed by these processes.

(v) Rock–avalanche-driven glacier advances have the potential to cause serious disruption to society and business over decadal timescales.

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