Geomorphic Hazards associated with Glacial Change,
Aoraki/Mount Cook region, Southern Alps, New Zealand

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Frontispiece: Looking southeast down the middle section of the Hooker Glacier. This view exemplifies many of the impacts of glacial change in this environment: a downwasting glacier, rockfall deposits across the centre of the glacier, moraine wall instability and ice avalanche debris in the left foreground. Photo: S. Allen (19/01/07).
Abstract

Glacial floods and mass movements of ice, rock or debris are a significant hazard in many populated mountainous regions, often with devastating impacts upon human settlements and infrastructure. In response to atmospheric warming, glacial retreat and permafrost thaw are expected to alter high mountain geomorphic processes, and related instabilities. In the Aoraki/Mount Cook region of New Zealand’s Southern Alps, a first investigation of geomorphic hazards associated with glacial change is undertaken and is based primarily on the use of remote sensing and Geographic Information Systems (GIS) for mapping, modelling, and analysing related processes and terrain.

Following a comprehensive review of available techniques, remote sensing methods involving the use Advanced Spaceborne Thermal Emission and Radiometer (ASTER) imagery were applied to map glacial ice, lakes and debris accumulations in the Aoraki/Mount Cook region. Glacial lakes were mapped from two separate classification techniques using visible near infrared wavelengths, capturing highly turbid and clearer water bodies. Large volume ($10^6 - 10^8$ m$^3$) proglacial lakes have developed rapidly over recent decades, with an overall 20 % increase in lake area recorded between 2002 and 2006, increasing the potential for large mass movement impacts and flooding from displaced water. Where significant long-term glacial recession has occurred, steep moraines have been exposed, and large talus slopes occupy formerly glaciated slopes at higher elevations. At the regional-scale, these potential source areas for debris instabilities were distinguished from surrounding bedrock slopes based on image texture variance. For debris and ice covered slopes, potentially unstable situations were classified using critical slope thresholds established from international studies.

GIS-based flow routing was used to explore possible intersections between zones of human use and mass movement or flood events, assuming worst-case, probable maximum runout distances. Where glacial lakes are dammed by steep moraine or outwash gravel, primarily in cirque basins east of the Main Divide, modelled debris flows initiated by potential flood events did not reach any infrastructure. Other potential peri- and para-glacial debris flows from steep moraines or talus slopes can reach main roads and buildings. The direct hazard from ice avalanches is restricted to backcountry huts and walking tracks, but impacts into large glacial lakes are
possible, and could produce a far reaching hazard, with modelled clear water flood-waves capable of reaching village infrastructure and main roads both east and west of the Main Divide. A numerical modelling approach for simulating large bedrock failures has been introduced, and offers potential with which to examine possible lake impacts and related scenarios.

Over 500 bedrock slope failures were analysed within a GIS inventory, revealing distinct patterns in geological and topographic distribution. Rock avalanches have occurred most frequently from greywacke slopes about and east of the Main Divide, particularly from slopes steeper than 50°, and appear the only large-magnitude failure mechanism above 2500 m. In the schist terrain west of the Main Divide, and at lower elevations, other failure types predominate. The prehistoric distribution of all failure types suggests a preference for slopes facing west to northwest, and is likely to be strongly influenced by earthquake generated failures. Over the past 100 years, seismicity has not been a factor, and the most failures have been as rock avalanches from slopes facing east to southeast, particularly evident from the glaciated, and potentially permafrost affected hangingwall of the Main Divide Fault Zone. An initial estimate of permafrost distribution based on topo-climatic relationships and calibrated locally using mean annual air temperature suggested permafrost may extend down to elevations of 3000 m on sunny slopes, and as low as 2200 m on shaded slopes near the Main Divide. A network of 15 near-surface rock temperature sensors was installed on steep rock walls, revealing marginal permafrost conditions (approaching 0 °C) extending over a much larger elevation range, occurring even where air temperature is likely to remain positive, owing to extreme topographic shading. From 19 rock failures observed over the past 100 years, 13 detachment zones were located on slopes characterized by marginal permafrost conditions, including a sequence of 4 failures that occurred during summer 2007/08, in which modelled bedrock temperatures near the base of the detachments were in the range of -1.4 to +2.5 °C.

Ongoing monitoring of glacial and permafrost conditions in the Aoraki/Mount Cook region is encouraged, with more than 45 km² of extremely steep slopes (>50°) currently ice covered or above modelled permafrost elevation limits. Approaches towards modelling and analysing glacial hazards in this region are considered to be most applicable within other remote mountain regions, where seismicity and steep topography combine with possible destabilizing influences of glacial recession and permafrost degradation.
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PART II Research Publications

Paper I
terrain analyses and hazard assessment in the Aoraki Mount Cook region, New Zealand.
Paper II
distribution in the Mount Cook region of New Zealand’s Southern Alps. In: Kane D.L.
and Hinkel K.M. (eds) Proceedings of the Ninth International Conference on Permafrost,
Institute of Northern Engineering, University of Alaska, Fairbanks, pp. 37 – 42.

Paper III
Allen, S.K., Cox, S.C., and Owens, I.F. (In review) Rock avalanches and other landslides in the
central Southern Alps of New Zealand: A regional-scale study of distribution patterns
and potential climate change impacts. Landslides.

Paper IV
Allen, S.K., Gruber, S., and Owens, I.F. (In press, 2009) Exploring steep bedrock permafrost and
its relationship with recent slope failures in the Southern Alps of New Zealand.

Paper V
Allen, S.K., Schneider, D, and Owens, I.F. (2009) First approaches towards modelling glacial
hazards in the Mount Cook region of New Zealand’s Southern Alps. Natural Hazards
and Earth System Sciences, 9, 481 – 499.

Associated Publications

Paper VI
Alps, New Zealand. Landslides, 6, 161-166.
PART I

Research Synopsis
1. Introduction

1.1 Background to the problem

Since the end of the Little Ice Age (LIA) between the 17th and mid-19th Centuries, glaciers around the world have undergone significant mass wastage, with particularly dramatic ice loss recognized since the mid-1980s, and continuing into the 21st Century (WGMS 2008). In many instances, glacial retreat has already continued beyond any historical precedent (Haeberli et al. 2002a), and expectations are for accelerated and irreversible atmospheric warming over the coming centuries (IPCC 2007). While the rapid disintegration of mountain glaciers and formation of large glacial lakes are amongst the most recognizable impacts of warming in the alpine environment, the degradation of permanently frozen ground (permafrost) represents an additional concern occurring at depth beneath the land surface (Harris and Vonder Mühll 2001; Harris et al. 2003). In combination, glacial recession and permafrost degradation are altering the geomorphic process activity occurring in the highly sensitive cryospheric environment, shifting zones of mass movement initiation and glacial flood potential (Evans and Clague 1994; Kääb et al. 2005a).

Retreating glaciers can uncover large accumulations of poorly consolidated debris, remove support provided to adjacent lateral moraines or rock walls, and expose fresh rock surfaces to mechanical and thermal erosion (Ballantyne 2002). Warming and altered geometry of steep glacial ice may initiate new zones of ice avalanche activity (Haeberli and Burn 2002), while glacial lakes forming on, at the side, or in deglaciated forefields may be poorly constrained and vulnerable to impacts from ice, rock or debris failures (Clague and Evans 2000; Richardson and Reynolds 2000). Meanwhile, the degradation of permafrost within steep rock walls represents a recently recognized, potentially fast mechanism with which warming may influence bedrock stability (Gruber et al. 2004b; Gruber and Haeberli 2007). When human systems intersect with these processes, a serious potential exists for natural hazards, with Kääb et al. (2005a) estimating average annual global damage and mitigation costs in the order of €100 million from related disasters.
Glacial floods, and mass movements of ice, rock or debris have posed a longstanding threat to many populated alpine regions, for example in the European Alps (Haebler 1983; Alean 1985; Rickenmann and Zimmermann 1993), South America (Reynolds 1992; Carey 2005), or central Asia (Hewitt 1982; Richardson and Reynolds 2000), while the most catastrophic events have involved complex transformations of mass movements or floods into rapid debris flows (eg. Lliboutry et al. 1977; Huggel et al. 2005). In other instances, problems have resulted directly from rapid surge-type glacial advances (Hewitt 1998; Haebler et al. 2002b). In addition to the direct threat to human populations, resource management implications relate to the quality and quantity of glacial meltwater, and potential damages to hydropower infrastructure (Huggel et al. 2003b; Singh and Bengtsson 2004; Mark and Seltzer 2005). With tourism development and population expansion into high mountain regions set to continue over coming decades, and human demand on limited hydrological resources intensifying, further conflicts with the natural environment are expected (Huggel et al. 2008a).

Although New Zealand has significant areas of glaciated alpine terrain, the population is almost entirely located in lowland areas, and therefore, scientific attention has traditionally focussed on flood, seismic, volcanic, coastal, and landslide problems potentially impacting upon populated areas (Speden and Crozier 1984). In contrast, glacial geomorphic processes associated with mass movements and floods within the higher elevation terrain of the Southern Alps have received comparatively less scientific study, despite recognition of significant 20th Century glacial recession (Hochstein et al. 1995; Chinn 1996). For example, although the historical distribution and importance of landsliding from steep, fractured slopes of the Southern Alps have been well established (eg, Whitehouse and Griffiths 1983; Korup 2005b), there has been only brief discussion regarding warming, ice recession and possible implications for slope instability and impacts into glacial lakes (McSaveney 2002). Similarly, the occurrence of permafrost in steep terrain of the Southern Alps has never previously been studied, and therefore any relevance of permafrost degradation to past or future mass movements of rock or debris is unknown. Small volume ice avalanches are observed frequently from steep ice cliffs and threaten many popular climbing routes (Iseli 1991), but there are no known precedents to suggest whether or not larger magnitude events may reach infrastructure or areas of high tourism activity within the Aoraki/Mount Cook National Park. Similar uncertainties are associated with debris flows from glacial origins, where reworking of deposits by fluvial activity and snow avalanching limits the recognition of past activity in the region. Therefore, establishing worst-case scenarios for mass
movements and flood events in the Southern Alps must to some extent rely on the implementation of empirical observations from other high mountain regions, where a much longer history of monitoring and documenting glacial hazards exists.

Previous studies have established the importance of applying modern remote sensing and geographic information systems (GIS) to the study of natural hazards in glacial environments, allowing processes and interactions to be studied over large areas often characterised by harsh and difficult access (Kääb et al. 2005b; Quincey et al. 2005). The Southern Alps of New Zealand provide an opportunity to further develop methodologies for detecting and modelling mass movement and flood events, on the basis of remotely sensed data and GIS based analyses. Specifically, the active tectonic setting of the Southern Alps requires a broad, integrative study of glacial-related mass movements, extending beyond previous approaches developed primarily in the European Alps (eg, Huggel et al. 2004a) to incorporate analyses of bedrock instability problems. Hence, the distribution of mountain permafrost within the Southern Alps must be explored for the first time. This is considered an important focus of the current study, because the high frequency of rockfalls and rock avalanches in the Southern Alps (McSaveney 2002) provides an opportunity for New Zealand scientists to contribute significantly to ongoing discussions linking atmospheric warming and bedrock stability in glacial environments.

1.2 Objectives of this research

Consideration of previous work both in New Zealand and internationally, directs the current study towards addressing the following key themes and related research objectives:

I. Remote sensing based mapping of glacial terrain. Studies completed in mountainous regions such as the European Alps and the Himalayas have shown that an initial analytical step towards investigating glacial-related floods and mass movements should distinguish potential initiation zones on the basis of mapped terrain surfaces such as ice, lake water, rock and debris. This study therefore reviews existing remote sensing based methodologies, and explores new possibilities, with the aim of establishing the most appropriate techniques for application at the regional-scale, and enabling ongoing monitoring of changing glacial conditions in the Southern Alps of New Zealand.
II. Validated modelling of mountain permafrost distribution. Quantifying altitudinal limits within which steep slopes are likely to be characterized by permafrost conditions will enable the possible influences of climate induced permafrost degradation to be explored in the Southern Alps. Initial empirically based estimates will aim to provide a basis for subsequent field based measurement of rock wall temperatures and regional-scale modelling of mountain permafrost distribution. Of particular importance is the need to distinguish marginal permafrost terrain, where temperatures approaching 0 °C are likely to indicate warm, degrading conditions.

III. GIS based analyses of landslide distribution and related glacial influences. A landslide inventory represents the spatial distribution of slope failure events, and provides the necessary framework within which instability can be compared with local predispositional factors such as slope topography and geological setting. Such analyses are enhanced within the GIS environment, where landslide affected terrain can be mapped and integrated with digital terrain data and geological mapping. In addition, inclusion of permafrost mapping and reconstructed glacial extents will allow the first informed discussion regarding climate change, related glacial terrain change, and potential influences on past and future slope instabilities in this region.

IV. Regional-scale modelling of glacial-related geomorphic hazards. Remote sensing based terrain mapping, GIS spatial analyses and flow modelling will be combined to explore the spatial distribution of potential ice avalanches, debris flows, and glacial lake floods. It is beyond the scope of this study to provide a detailed assessment of individual events or case-studies. Instead, the primary objective is to apply worst-case scenario modelling at the regional-scale, thereby establishing the extent of the problem posed by glacial-related mass movements and floods, and directing more detailed investigations towards critical situations and affected areas.

1.3 Research and thesis structure

Research methodologies described for glacial hazards commonly fit within a downscaling strategy that includes first-order regional-scale approaches (> 1000 km²) aimed at detecting unstable terrain predisposed to flood or mass movement failure, a second level of investigation
where broad hazard assessments are made and affected areas are recognized, leading on to detailed local-scale analyses to establish failure susceptibility, impact characteristics, vulnerability and remedial measures (e.g., Huggel et al. 2004b; McKillop and Clague 2007a; Quincey et al. 2007). This research addresses the main considerations and methodologies appropriate at the first and second levels of a regional hazard investigation, with results published in a series of peer-reviewed international journal papers. The project therefore encompasses the entire Aoraki/Mt Cook study region (Section 1.4), but local-scale analyses and discussions are included where relevant and where this provides appropriate direction for further research.

The thesis is presented in two parts. Part I introduces the research, describes the study region, and establishes the scientific context surrounding the main themes of this research. This leads to a synthesis of the main results arising from this study. In most instances, the reader is directed towards the associated publication(s) in which a more detailed treatment of the results is provided, although some unpublished material is also presented in the results synthesis. Conclusions from this study are then given, with perspectives pertaining to both local and international development in glacial hazard research.

Part II of the thesis contains the five peer reviewed scientific papers in which the research results are presented and discussed in detail. The order in which these papers appear follows the treatment of the research objectives and is not governed by publication date. Four papers have been published or accepted for publication at the time of thesis completion, while one other remains under peer-review. These five papers have all been primarily written by the current author, with minor written contributions and/or substantial research supervision provided by listed co-authors. An additional publication (Paper VI) presents collaborative work strongly related to the research objectives, in which the current author contributed as the second author.

1.4 Aoraki/Mount Cook study region

The majority of remote sensing based analyses and field research completed in this project focuses upon a large (~3500 km²) area referred to here as the Aoraki/Mount Cook region (Fig. 1), although data from a larger region of the central Southern Alps are included within the analyses of bedrock slope failures (Paper III). The Aoraki/Mount Cook region is broadly defined
here to encompass the entire Aoraki/Mount Cook National Park, extending west of the Main Divide into the Westland National Park and further south towards the Lake Pukaki.

![Map of Southern Alps](image1)

Figure 1. A) Overview of the central Southern Alps, showing land cover as mapped in the national database (Ministry for the Environment 2004). Aoraki/Mount Cook village (MCV), Fox village (FV), and Franz Josef village (FJV) are indicated. B) ASTER false colour composite of the Aoraki/Mount Cook region (January 2006), with some large glaciers indicated. Inset shows the position of the Australian-Pacific plate boundary.

A popular tourist and adventure sport destination, the region also provides the catchment area for a significant proportion of the nation’s hydro-electricity generation. The region includes the highest mountains and the most heavily glaciated terrain of New Zealand’s Southern Alps.
Permanent snow covered peaks are found between 2500 and 3754 m, with local relief in the order of 1000 - 2700 m (McSaveney 2002).

1.4.1 Weather and climate

The weather and climate of the central Southern Alps is dominated by a moist westerly atmospheric circulation which generates very high orographic precipitation ( > 12000 mm yr\(^{-1}\)) as prevailing winds meet the Main Divide of the Alps. However, an extreme precipitation gradient leeward of the Alps produces < 2000 mm yr\(^{-1}\) only 20 km further to the southeast as a consequence of the foehn effect (Henderson and Thompson 1999). Typical synoptic weather patterns involve a strengthening warm west to northwest foehn wind across the Alps ahead of a cold front moving up the South Island, followed by significant cooling as wind directions turn towards the south after the frontal system passes (Sturman and Tapper 1996). Above 1500 m, the majority of winter precipitation falls as snow, although freezing levels fluctuate significantly during storm cycles and rain or snow is possible at all elevations at any time of year. Storm cycles are commonly separated by large slow moving high pressure systems, bringing fine, calm, settled weather, and lower level temperature inversions during the winter months. Temperature records from Aoraki/Mount Cook village (765 m) for the years 1971 – 2000 reveal mean maximum daily temperatures of ~20 °C during January and February, decreasing to only 6.5 °C in July (NIWA 2009). Mean minimum daily temperatures are ~9 °C during January and February, and -2 °C during July. The most reliable long term climate records come from near sea level in Hokitika, where warming of 0.7 °C has been documented during the period 1920 – 1990 (Salinger et al. 1995).

1.4.2 Geology

The geological landscape of the central Southern Alps is young and vigorously evolving, having been strongly influenced by the effects of global climate fluctuations, local weather patterns, and tectonic activity (Whitehouse 1988; Fitzsimons and Veit 2001). The Alpine Fault is a major active fault traversing the northwestern edge of the study area, on which most of the ongoing tectonic movement between the Australian Plate (to the northwest) and the Pacific Plate (to the southeast) is concentrated (Fig. 1). Neogene displacement of up to 470 km along the Alpine Fault has brought together two different pre-Cretaceous geological provinces (Cox and Barrell
Northwest of the Alpine Fault, there are small amounts of exposed Paleozoic metasedimentary and plutonic basement rocks that are fragments of the Gondwanaland supercontinent. Southeast of the Alpine Fault, basement rocks belong to the Torlesse composite terrain. They comprise thick, deformed packages of sandstone and mudstone that were deposited and accreted to Gondwanaland during the Carboniferous to Early Cretaceous, and have locally been metamorphosed into semischist or schist. Convergence across the Australian-Pacific plate boundary in South Island pushes thinned and submerged crust upward into the path of prevailing moist westerly winds (Whitehouse 1988). Differential uplift, erosion and rock exhumation across the Southern Alps has exposed transitions from uncleaved greywackes in the east, through weakly cleaved or fractured greywacke and foliated semischist about the hangingwall of the Main Divide Fault Zone, to strongly-foliated amphibolite facies schist in the west adjacent to the Alpine Fault (Cox and Findlay 1995). Uplift of the Southern Alps by folding and faulting continues to the present day at up to 10 mm yr\(^{-1}\), and the entire area is subject to episodic shaking from high-magnitude M7-8 earthquakes every 200 – 300 years (Wells et al. 1999).

1.4.3 Glacial history

Late Pleistocene glacial cycles have carved and shaped the central Southern Alps landscape, leaving a mantle of cover deposits throughout the region. Glaciations probably began in the Late Pliocene (Suggate 1990) and by the Late Pleistocene an extensive system of glaciers extended almost uninterrupted 700 km along, and 100 km across the Southern Alps (Newham et al. 1999). Glaciers coalesced in the main mountain valleys to form piedmont lobes in the west and extended through the foothills linking to alluvial outwash plains in the east. A series of relatively well-defined moraines and till deposits mark the extent of ice during the Last Glacial Maximum (LGM) around 26-23 ka (cal. yr. BP) and there is widespread evidence in the landscape for at least four major periods of glacial advance since the LGM (Fitzsimons 1997). In addition, at least six further minor advances are thought to have occurred during the Holocene (Gellatly et al. 1988). Timing of the Little Ice Age (LIA) maximum varied across the region, but between 1750 and 1890 widespread retreat had begun, becoming most evident during the mid-20\(^{th}\) Century (McKinzy et al. 2004; Winkler 2004). In total a 49 % loss in ice area and 61 % volume loss has been estimated since 1850 (Hoelzle et al. 2007), and valley glaciers have receded an average of 1.8 km in length (Chinn 1996). Franz Josef and Fox Glaciers are the most notable exceptions to the general pattern of retreat. Both are highly responsive glaciers, and therefore have shown
significant periods of terminus growth during recent years in response to short term variations in synoptic airflow, bringing above average snowfall to the large accumulation areas west of the Main Divide (Chinn et al. 2005).

Figure 2. Aoraki/Mount Cook village (765 m) is located on sloping fans at the foot of the Sealy Range (left), with a camping ground and other recreational infrastructure located in front of the vegetated LIA moraines (right). Mount Sefton (3151 m) and The Footstool (2764 m) are the prominent peaks of the Main Divide appearing behind. Photo: S. Allen (10/04/08).

1.4.4 Infrastructure

The Aoraki/Mount Cook village is the closest service and accommodation centre east of the Main Divide, with a small permanent population of ~100 and a larger seasonal component linked to tourism activities in the region (Figs. 1 and 2). West of the Main Divide, the largest tourist and residential centres are Franz Josef and Fox Glacier villages, with ~300 permanent residents in each village. Because much of the study area is contained within government managed national park land, tourism activities and infrastructural development are tightly regulated. However, a network of backcountry huts and shelters provide accommodation throughout the region. Figures provided by the Department of Conservation indicate that nearly 10,000 nights per year were spent by visitors within Aoraki/Mount Cook National Park during the monitoring period 1982 to
2002. The most popular locations were Mueller Hut (1108 nights) (Fig. 3), Plateau Hut (1143), Tasman Saddle (1043) and Kelman Huts (1677).

West of the Main Divide, within the boundaries of the Westland National Park, 4,500 visitors explored the Copland Valley track during 2007, and extremely popular guided excursions operate on the lower reaches of the Franz Josef and Fox Glaciers. For example, an estimated 250,000 people visited the terminus of the Fox Glacier during 2007. Main roads connect to all tourist and residential areas throughout the region, but the nearest transalpine passes are 80 km to the southwest and 150 km to the northeast. Numerous unsealed vehicle tracks give access to more remote areas and farm buildings. The study region incorporates catchment areas feeding into Lakes Tekapo, Pukaki, and Ohau, all contributing to the nationally significant Waitaki hydroelectric power scheme.

Figure 3. Mueller Hut, looking north towards the Hooker Glacier and summit of Aoraki/Mount Cook (3754 m). Photo: S. Allen (22/12/08).
2. Research Context

2.1 Glacial hazards

The short term ‘glacial hazards’ is employed in this study to encompass a broad range of potentially threatening geomorphic processes originating from peri-glacial, para-glacial or currently glacierised terrain, although a variety of alternative terminology such as ‘glacier hazards’ or ‘glacier and permafrost related hazards’ have been used in other studies (eg, Tufnell 1984; Kääb et al. 2005b). Glacial hazards have elsewhere been defined as any glacial or glacier-feature or process that adversely affect human activities (Reynolds 1992). In the context of this study, the relevant processes include: glacial floods, debris flows, ice avalanches and bedrock failures, while recognizing that complex interactions are possible between all individual processes (eg, Huggel et al. 2004a). The term para-glacial in this thesis is used in reference to ‘present day’ para-glacial processes or landforms, defined by Ballantyne (2002) as nonglacial earth-surface processes, sediment accumulations or landforms that have been directly conditioned by glaciation and deglaciation since the LIA, approximately within the past 100 – 200 years.

2.1.1 Glacial floods

Glacial floods refer to the sudden discharge of a water reservoir that has formed either underneath, at the side, in front, within, or on the surface of a glacier, and related blockages can form within ice, moraine or bedrock (Richardson and Reynolds 2000). The Icelandic term ‘jökulhlaup’ is widely used to cover a spectrum of flood types originating from a glacier, although more correctly should only be applied to events originating subglacially (Thorarinsson 1939). For reservoirs developing as lakes on or at the margins of glaciers, remote sensing at suitable spatial and temporal resolution has proven an appropriate tool for monitoring hazardous developments (Huggel et al. 2002; Wessels et al. 2002). Where floods have initiated from the catastrophic failure of moraine dammed lakes, the term glacial lake outburst flood (GLOF) has been adopted, initially coined in relation to Himalayan events (Richardson and Reynolds 2000). GLOFs frequently transform into hyperconcentrated or debris flows following the rapid
entrainment of para-glacial debris, and are a well documented hazard for Central Asia (Hewitt 1982; Ding and Liu 1992; Quincey et al. 2007), the Andes (Lliboutry et al. 1977; Reynolds 1992), North America (Clague and Evans 2000; O’Connor et al. 2001), and Europe (Haeberli 1983; Haeberli et al. 2001; Huggel et al. 2002). Failure of moraine dammed lakes occurs when the material strength of the dam structure is exceeded by driving forces, including the weight of the impounded water mass, shear stresses from seepage, and overtopping or additional momentum from displacement waves (Korup and Tweed 2007). Such displacement waves, particularly from large impacts of ice or rock are thought to contribute to over 50 % of catastrophic moraine dam failures occurring in the Himalaya’s for example (Richardson and Reynolds 2000), but can also result in overtopping of bedrock dammed lakes that would otherwise be considered stable.

Figure 4. A large lake has formed at the terminus of the Tasman Glacier. The length from the outlet to the highest extension of the lake arm is ~5.5 km at the time of the photo. Photo: S. Allen (11/01/09).

Flood hazard problems have been described pertaining to the Franz Josef and Fox Glaciers, located in the far west of the Aoraki/Mount Cook study region (Davies et al. 2003; Goodsell et al. 2005; Purdie et al. 2008). In these instances, lake formation has not been visible at the ice
surface, and reservoirs have instead formed sub- or englacially, often associated with heavy rainfall and ice collapse (Davies et al. 2003). Extreme hydraulic pressure may reroute subglacial water to the surface, as was the likely situation in a 2003 supraglacial flood (Goodsell et al. 2005). East of the Main Divide, large proglacial lakes have formed in response to terminal recession during recent decades, and are dammed by moraine or outwash debris (Fig. 4) (Hochstein et al. 1998; Warren and Kirkbride 1998; Röhl 2006). There has been limited discussion of any flood potential from these lakes, although a 10 m high flood-wave was described from a rock avalanche depositing into a lake at the terminus of the Maud Glacier, travelling 45 km downstream within the braided Godley River valley and damaging a vehicle track (McSaveney 2002).

At the global scale, it is predicted that both the number and size of glacial lakes will increase in response to global warming and glacial recession, increasing the potential loss of life and infrastructural damage (Kääb et al. 2005a). In the Himalayas for example, significant floods occurred roughly once per decade in the mid-20th Century, increasing to once every 3 years in the 1990s, and are expected to occur annually by 2010 (Richardson and Reynolds 2000). Examples from the European Alps indicate that glacial lakes may even form and significantly enlarge during the course of a single melt-season (Kääb et al. 2003), challenging monitoring and mitigation practices to keep pace with rapidly evolving hazardous developments.

2.1.2 Debris flows

Debris flows consist of fast flowing mixtures of sediment and water, comprising of one or several pulses (Iverson 1997). Diagnostic features include a substantial erosion capacity, transportation of large boulders, poorly sorted deposits and levee formation in response to flow deceleration on flatter terrain (Hungr et al. 2001; Goudie 2004) (Fig. 5). Peri- and para-glacial zones are often characterised by large accumulations of loose, unconsolidated material, and are therefore particularly prone to debris flow formation (Evans and Clague 1994; Haeberli et al. 1997). Glacial retreat is important in this context, uncovering large volumes of unstable sediment. For example, an investigation of damaging debris flow disasters in the Swiss Alps during 1987 concluded that more than 50 % of these events originated in zones that had deglaciated within the previous 150 years (Zimmermann and Haeberli 1992). Re-vegetation of deglaciated terrain is slow and therefore moraine deposits remain unprotected against erosion
over extensive time periods of several decades or more (Kääb et al. 2005a). Additional concerns relate to the thermal sensitivity of marginal permafrost sites in response to warming, potentially leading to destabilisation of steep talus, moraine, or rock glacier fronts (Harris 2005). Permafrost thaw may lead to a loss of cohesion, an increase in poor water pressure, and subsidence in non-consolidated debris following the disappearance of massive ground ice, increasing the potential frequency and magnitude of debris flow events (Haeberli and Burn 2002). Unlike debris flows on lowland hillslopes, events initiating from the glacial environment may activate via several mechanisms, including snow or ice melt and high intensity rainfall (Rickenmann and Zimmermann 1993; Chiarle et al. 2007), as a direct result of permafrost thaw (Harris 2005), or from catastrophic entrainment within glacial floods (Clague and Evans 1994; O’Connor et al. 2001). However, once activated a debris flow within the glacial environment is expected to display flow dynamics comparable to non-glacial debris flows, for which a variety of empirical, physical and numerical understanding has been established (eg, Takahashi 1991; Iverson 1997; Hungr et al. 2005).

Figure 5. Scientists study a large debris flow channel, originating from a peri-glacial zone thought to be characterised by marginal permafrost, and depositing onto the Lower Grindelwald Glacier, Switzerland. Note the levee formations, and large boulders transported within the channel. Photo: S. Allen (10/10/07).
Most studies regarding debris flow activity in the Southern Alps have focussed on non-glacial hillslopes in areas near to and west of the Main Divide, where higher precipitation and seismicity enhances sediment yields and the frequency of shallow slope failures (Whitehouse 1988; Korup 2004; Korup et al. 2005). Within the study region, debris flows initiated during storm events are known to potentially endanger infrastructure within the Aoraki/Mount Cook village, resulting in the implementation of engineered defences to protect the village (Whitehouse 1982; Skermer et al. 2002; McSaveney and Davies 2005) (Fig. 6). Where catastrophic glacial downwasting has occurred, surrounding moraine walls have been left steep and exposed to erosion from debris flows, unconfined debris avalanches, and complex slope failures, destroying or necessitating the removal of infrastructure located on the terraces above (Blair 1994). In all recorded instances of significant moraine wall erosion or debris flow activity affecting the Aoraki/Mount Cook region, events have been linked to spring snow melting or large rainfall events.

Figure 6. Infrastructure of the Aoraki/Mount Cook village is located on composite fans, considered vulnerable to debris flows originating from the Sealy Range. Photo: S.Allen (10/04/08).
2.1.3 Ice avalanches

Ice avalanches occur when large masses of ice break off from steep ice cliffs and hanging glaciers (Alean 1985), detaching as frontal block failures, slab failures, or deeper failures at the ice/rock interface (Richardson and Reynolds 2000). Compared to glacier floods, ice avalanches typically travel much shorter distances, and therefore direct impacts are normally restricted to high alpine villages and infrastructure (Röthlisberger 1977). However, in combination with other mass movements or via flow transformations, ice avalanches have the potential for far-reaching disasters, as eventuated from the Peruvian Huascaran disaster of 1970, in which 18,000 people were killed from an ice/rock avalanche triggered mudflow (Carey 2005). More recently, the enormous Kolk-Karmadon ice/rock avalanche transformed into a highly mobile debris flow, resulting in approximately 140 deaths (Haeberli et al. 2004; Huggel et al. 2005). This event was exceptional in that it involved the entire mobilization of a relatively moderately sloping glacial tongue, indicating that ice avalanche problems are not necessarily restricted to small, steep ice masses. Given expectations of an atmospheric warming, there is particular concern relating to the warming and potential instability of hanging glaciers, because thermal conditions and geometry of the steep ice front is considered to exert a paramount control on glacier stability (Huggel 2004). Responding to this concern, recent GIS and remote sensing applications in the Swiss Alps have enabled the first regional assessment and modelling of ice avalanche hazard to proceed based on the recognition of such steep ice masses (Huggel et al. 2004a; Salzmann et al. 2004), although accurately predicting ice avalanche failure still requires local-scale monitoring (Wegmann et al. 2003).

In New Zealand, at least twenty-three ice avalanche fatalities have been documented, primarily associated with recreational activities, and all but one of these has occurred in the Aoraki/Mount Cook region (Irwin et al. 2002). Past research has suggested that ice avalanche activity in the region predominates as low magnitude (< 1000 m³), high frequency events (1-8 per hour), occurring from steep ice cliffs (Iseli 1991) (Fig. 7). However, the potential hazard from larger magnitude events in the region remains unknown, and in particular, whether or not significant impacts into glacial lakes are possible. Elsewhere, ice avalanches have proven an important trigger of glacial floods, in particular the Cordillera Blanca, Peru (Reynolds 1992), Canadian Cordillera (Reynolds 1992) and the Himalayas (Richardson and Reynolds 2000). As glaciers respond to atmospheric warming, the zone of potential ice avalanche activity will undoubtedly
shift, and in some instances current threatening glaciers could disappear entirely, while other new instabilities might develop at higher elevations (Kääb et al. 2005a).

Figure 7. Frequent ice avalanches falling from steep cliffs on the east face of Mount Sefton are a popular sight viewed from Mueller Hut. Photo: S. Allen (14/01/08).

2.1.4. Bedrock failures

Stress adjustment and destabilisation of bedrock walls associated with glacial ice recession and permafrost thaw can contribute to a range of slope failure mechanisms operating over different time scales, from large deep seated creeping deformation of entire mountain sides, through to small individual rockfalls (Evans and Clague 1994; Harris 2005). Large rock avalanches have received particular focus, owing to potential catastrophic flow transformations and secondary hazards (Huggel et al. 2005; Korup and Tweed 2007), exceptional runout distances observed over snow and ice surfaces (Evans and Clague 1988; Noetzli et al. 2006; Strom and Korup 2006; Lipovský et al. 2008), and potential short or longer term influences on glacier dynamics and landscape modification (Hewitt et al. 2008; Hewitt 2009; Shulmeister et al. 2009). In addition, rock avalanches and large rockfalls detach from a relatively well defined source area, from which related permafrost and glacial conditions can be reconstructed and analysed in many instances (Noetzli et al. 2003; Fischer et al. 2006; Huggel 2008).
Bedrock walls in glacial environments are typically oversteepened, with their lower flanks eroded by glacial plucking (Ballantyne 2002). Subsequent retreat of glacial ice can induce changes in the stress field of the surrounding rock walls and expose previously insulated surfaces to mechanical and thermal erosion (Haeberli et al. 1997; Wegmann et al. 1998). The influence of permafrost degradation within steep rock walls was first considered within theoretical discussions linking atmospheric warming, permafrost degradation, and slope instability (Haeberli et al. 1997; Harris and Vonder Mühll 2001). Meanwhile, laboratory studies quantitatively demonstrated that the shear strength of an ice bonded rock discontinuity significantly reduces with warming, revealing a minimum factor of safety at temperatures between -1.5 and 0 °C, where a rock discontinuity may be less stable than when in a completely thawed state (Davies et al. 2001). Crucially, an absence of snow or debris cover on steep slopes means that sub-surface temperatures and related bedrock stability will respond rapidly to atmospheric warming (Gruber and Haeberli 2007), as was witnessed during the extremely warm European summer of 2003, when numerous rockfalls were associated with extreme heat fluxes in the active layer of the permafrost (Gruber et al. 2004b). Deeper warming of alpine permafrost is delayed (Harris et al. 2003), but has been linked to rock avalanches from corresponding deeper detachment surfaces occurring primarily in estimated zones of warm or marginal permafrost (Dramis et al. 1995; Bottino et al. 2002; Noetzli et al. 2003).

Bedrock slope failures are widespread throughout the Southern Alps of New Zealand, and have been catalogued in relation to large earthquakes (Speight 1933; Adams 1981; Bull 1996; Bull and Brandon 1998; Hancox et al. 2003), and mapped regionally within Westland (Korup 2005b), Fiordland (Hovius et al. 1997; Korup 2005b), and the central Southern Alps (Whitehouse 1983). The geomorphological importance of these mass movements in the Southern Alps is well established (Korup et al. 2004; Korup et al. 2005), and significant secondary hazards resulting from river blockage and breakout floods have been described (Davies and Scott 1997; Hancox et al. 2005; Korup 2005a). Whitehouse (1988) described the zonation that occurs across the Southern Alps, from large landslides and shallow slope failures in the schist terrain west of the Main Divide, to predominant rock avalanching occurring about and east of the Main Divide in a zone referred to as the ‘Axial Alps’, which includes the hangingwall of the Main Divide Fault Zone (Cox and Findlay 1995). It is from within this zone that several large rock avalanches have been recorded over recent years, including the spectacular 12 x 10^6 m^3 summit collapse of Aoraki/Mt Cook (McSaveney 2002) (Fig. 8). Although glacial recession has been recognized as
a likely precursor in some recent bedrock failures, and the increasing potential for rock impacts into newly formed lakes has been briefly discussed (McSaveney 2002), these considerations, including any role of permafrost degradation generally remain unexplored.

Figure 8. A large rock avalanche fell from the east face of Aoraki/Mt Cook (3754 m), December 14, 1991. The failure reduced the elevation of New Zealand’s highest mountain by ~10 m. Photo: Ian Owens (16/12/91).

2.2 Applications of GIS and Remote Sensing

The application of GIS and remote sensing techniques are best suited at the first and second levels of a glacial hazard investigation, for the purposes of glacial terrain mapping, permafrost distribution modelling, analysing past events for the construction of hazard inventories, and ultimately combining this knowledge to model potential mass movements and flood events.

2.2.1 Glacial terrain mapping

A fundamental starting point for glacial hazard investigations begins with the early identification of potentially unstable terrain on the basis of mapped glacial margins and associated terrain
features, identifying for example, expanding glacial lakes and receding or advancing glacial ice (Kääb et al. 2005b). For this purpose, optical remote sensing (0.4 – 14.0 μm) has proven extremely popular and successful, owing to a multitude of available sensors operating at various spatial resolutions, opportunities to capture additional three-dimensional data, and temporal resolutions allowing regular, long term monitoring (Quincey et al. 2005) (Table 1). While traditional aerial photography remains suited for fine delineation of glacial features (Chinn 1995; Casassa et al. 2002), and constructing high resolution digital elevation models (DEM) for detailed geometric investigations (Kääb and Vollmer 2000; Kääb 2002), satellite data from Landsat, Advanced Spaceborne Thermal Emission and Radiometer (ASTER), and Systeme Pour l’Observatoire de la Terre (SPOT) sensors provide cost effective alternatives for regional-scale applications (Gao and Liu 2002). In fact, Landsat and ASTER sensors (15 – 30 m minimum spatial resolution) form the basis of the Global Land Ice Measurements from Space (GLIMS) international project aiming to map and monitor changing glacial environments around the world (Raup et al. 2007). Available techniques for satellite based mapping of glacial ice generally make use of the high reflectance of snow or ice in the visible wavelengths, relative to low reflection in the short-wave infrared, using simple band algebra and ratio images (eg, Hall et al. 1988; Bayr et al. 1994; Paul et al. 2002). However, these techniques have proven sensitive in areas of cast shadow, and are unable to distinguish debris covered glacial ice from surrounding terrain, something that may be partially overcome by considering other factors such as vegetation coverage and glacier topography (Paul et al. 2004a).

Mapping the development of glacial lakes has similarly evolved from historical observations using aerial photography (Kirkbride 1993; Mool 1995; Kirkbride and Warren 1999), through to more recent use of multi-spectral satellite image classification (Wessels et al. 2002; Quincey et al. 2007). In particular, the use of multi-temporal images has provided information regarding not only lake growth, but also the stability of surrounding terrain (Haeberti et al. 2001), and the temporal resolution of ASTER has proven suitable for monitoring even the most rapidly evolving situations (Kääb et al. 2003). As a subsurface phenomenon, permafrost can hardly be mapped directly using remote sensing, although related terrain surface features such as thaw lakes and rock glaciers have been analysed on the basis of aerial or satellite imagery (Brazier et al. 1998; Lewkowicz and Duguay 1999; Kääb and Vollmer 2000). Regional-scale aerial and satellite remote sensing of rock walls is also limited and subject to strong distortion on steep slopes, although the disintegration of glacial ice leading to fresh bedrock exposure can be
monitored (Paul et al. 2004b), past changes reconstructed (Fischer et al. 2006), and even lithological variations detected within exposed rock faces (Huggel et al. 2007). In addition, novel approaches based on image texture analyses have been able to distinguish debris accumulations from surrounding bedrock at higher spatial resolutions (Huggel et al. 2004a).

Table 1. Selected glacial hazard studies where optical remote sensing has been used to map and/or monitor glacial terrain features. See Kääb et al. (2005b) and Quincey et al. (2005) for a more substantial listing, including applications involving three-dimensional analyses of terrain geometry.

<table>
<thead>
<tr>
<th>Glacial Hazard</th>
<th>Location</th>
<th>Data</th>
<th>Application</th>
<th>Author</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Several glaciers of the Tibetan and Nepalese Himalaya</td>
<td>ASTER</td>
<td>Automatic classification of lakes, turbidity levels and temperature</td>
<td>(Wessels et al. 2002)</td>
</tr>
<tr>
<td></td>
<td>Several glaciers of the Tibetan and Nepalese Himalaya</td>
<td>Aerial Photos and SPOT-X combined with Synthetic Aperture Radar (SAR)</td>
<td>Early detection and prediction of lake formation</td>
<td>(Quincey et al. 2007)</td>
</tr>
</tbody>
</table>

| Ice avalanches                  | British Columbia, Canada                      | Aerial Photos                                  | Post-failure analyses of ice avalanche triggered GLOF   | (Clague and Evans 2000)       |
|                                 | Kolk/Karmadon, Russian Caucasus              | ASTER and QuickBird                            | Post-failure analyses of massive ice/rock avalanche    | (Huggel et al. 2005)          |
|                                 | Bernese Alps, Switzerland                     | Landsat TM                                     | Automatic classification of steep ice                  | (Salzmann et al. 2004)        |
|                                 | Grimsel Pass, Switzerland                     | Landsat TM                                     | Automatic classification of steep ice                  | (Huggel et al. 2004a)         |

| Debris flows                    | Grimsel Pass, Switzerland                     | IKONOS                                         | Automatic classification of debris accumulations using texture algorithm | (Huggel et al. 2004a)         |

| Bedrock slope failures          | Monte Rosa east face, Italian Alps            | Aerial Photos                                  | Reconstruction of glacial recession associated with rockfall activity | (Fischer et al. 2006)         |
|                                 | Iliamna Volcano, Alaska                       | Landsat ETM+                                   | Mapping of failure zone lithologies                     | (Huggel et al. 2007)          |
2.2.2 Permafrost distribution modelling

Permafrost is formally defined as ground that remains at or below 0 °C for two or more consecutive years (Permafrost Subcommittee 1988). Because of the sensitivity of upper permafrost layers to atmospheric warming, and potential impacts of mountain permafrost degradation on landscape dynamics and related hazards, accurately modelling and mapping the distribution of mountain permafrost is considered fundamentally important (Hoelzle et al. 2001). As a thermal phenomenon, mountain permafrost is predominantly influenced by air temperature and solar radiation. Hence the DEM, and associated topographic derivatives of slope angle and slope aspect are fundamental requirements for GIS based permafrost distribution modelling at the micro- to meso-scale (<25 – 250 m resolution) (Etzelmüller et al. 2001b; Salzmann et al. 2007). A multitude of GIS based approaches have stemmed from empirical observations in the European Alps and Scandinavia linking diagnostic permafrost features (rock glaciers, thermokarst etc) or proxy indicators (eg, basal temperature of the snowpack) to topo-climatic factors that can be easily modelled in a GIS (eg, Keller 1992; Hoelzle and Haeberli 1995; Imhof 1996; Etzelmüller et al. 1998). Haeberli’s (1975) early investigations in the Swiss Alps were influential in this regard, establishing ‘rules of thumb’ for predicting permafrost occurrence which implicitly considered governing physical factors, including aspect dependent radiation effects, altitudinal changes in air temperature, and topographically related snow cover variation. More recently, modelling methodologies built upon similar concepts have been applied in North America (Lewkowicz and Ednie 2004; Janke 2005), and remotely sensed datasets have been included to improve modelling accuracies (Etzelmüller et al. 2001a; Gruber and Hoelzle 2001).

In recognition of the complex topography that governs discontinuous permafrost distribution and its evolution in steep alpine terrain, focus in recent years has tended towards process-oriented distribution modelling that provides more detailed understanding of energy fluxes between the atmosphere and permafrost layers (Hoelzle et al. 2001). Specifically in relation to steep rock walls, Gruber (2004a) combined the energy balance model PERMEBAL (Stocker-Mittaz et al. 2002) with near-surface rock temperature measurements to investigate the spatial distribution of steep rock wall temperatures in the Swiss Alps. This work provided exciting insights into topographically driven temperature variations and demonstrated links between rockfall activity and permafrost thaw (Gruber et al. 2004b). While meteorological parameters required to drive energy balance models may be lacking for many remote, undeveloped mountain regions such as
New Zealand, the pioneering approach of directly measuring rock wall temperatures (Gruber et al. 2003) offers significant potential for relating measured temperatures to local GIS modelled or extrapolated parameters such as global radiation or air temperature.

2.2.3 Hazard inventories and event reconstruction

Predicting and assessing future slope instabilities or flood problems must to a large extent rely on empirical information learned from past events, while recognizing that unprecedented failures can and do occur, as illustrated by the 2002 Kolka-Karmadon ice/rock avalanche (Haeberli et al. 2004). In any case, identifying and reconstructing past failures is a challenging task given the rapid mass turnover that characterizes dynamic glacial environments and often remote locations which mean even significant mass movements can remain unnoticed for several years (eg, Cox and Allen 2009). Archived aerial photography and more recent satellite imagery provide opportunities to retrospectively study past events across remote mountain regions, aiding future hazard assessment. The resulting data may be used, for example, to statistically evaluate the probability of moraine-dammed lakes catastrophically failing (McKillop and Clague 2007b), to establish empirical relationships defining event magnitudes (Huggel et al. 2004b), or to establish the critical factors predisposing high mountain slopes to failure (Fischer and Huggel 2008). Because large bedrock failures leave a significant imprint on the landscape (Hewitt et al. 2008), landslide inventories are a particularly useful tool for hazard assessment, often mapped entirely on the basis of aerial or satellite image interpretation (eg, Mantovani et al. 1996; Parise 2001; Hancox et al. 2003; Singhroy 2004; Korup 2005b). In combination with a GIS, a landslide inventory provides the basis with which slope stability can be related to predisposing factors such as slope angle, structural orientation, fault related weaknesses, lithology, landuse practices etc (eg, Carrara et al. 1995; Hermanns and Strecker 1999; Pande et al. 2002; Pike et al. 2003). Although more difficult to quantify, additional predisposing factors relating to glacial and permafrost conditions are increasingly being considered in high mountain environments (Evans and Clague 1988; Bottino et al. 2002; Noetzli et al. 2003; Deline 2008).

2.2.4 Event modelling and hazard assessment

Following the recognition of glacial, permafrost, topographic, and geological factors that may indicate potential slope instabilities and/or unstable glacial lakes, attention turns towards the
likely impact the resulting flood or mass movement will have on the surrounding environment, and human use systems. Physical understanding of processes which govern the behaviour of debris flows (eg, Hungr et al. 1984; Iverson 1997), glacial floods (eg, Walder and Costa 1996; Walder and O’Connor 1997; Kershaw et al. 2005), bedrock failures (eg, Dorren 2003; McDougall and Hungr 2004; McSaveney and Davies 2006) and to a lesser extent ice avalanches (Alea 1985; Margreth and Funk 1999) have been described. These processes have been at least partially replicated within physically based mass movement models (O’Brien et al. 1993; Hungr et al. 1995; Christen et al. 2008), but in general the input parameters and computational requirements limit the suitability of these models for large scale application where multiple events are to be simulated.

Because glacial-related mass movements and floods are essentially gravity driven flows, an alternative approach for regional-scale hazard assessment might consider the use of hydrological flow-routing models, many of which are embedded in commercial GIS packages. GIS based flow-routing algorithms have been used to model ice avalanche paths (Salzmann et al. 2004), glacial lake outbursts (Huggel et al. 2003b; Huggel et al. 2003a; Huggel et al. 2004a), debris flows (Huggel et al. 2004a; Stolz and Huggel 2008), and rock avalanches from glacierised slopes (Noetlzi et al. 2006; Huggel et al. 2007). In addition, successful reconstruction and simulation of potential lahars have been achieved (Huggel et al. 2008b; Schneider et al. 2008a). Fundamental empirical parameters regarding debris flows (Rickenmann 1999; Rickenmann 2005), ice avalanches (Alea 1985), and glacial floods (Haebeli 1983; Clague and Evans 1994; Clague and Evans 2000) may assist the selection of source areas (eg, weak dam geometries and compositions, critically steep ice and debris accumulations etc), and determination of maximum travel distances. While not representing the true physical characteristics of a mass movement or flood, modifications to strictly gravity driven flow-routing have enabled for example, flow spreading and deposition to be simulated (Huggel et al. 2003a), significantly improving the reliability of the models in unconfined topography. Sensitivities relating to the underlying DEM remain, and while the highest level of topographic detail can be extracted from photogrammetric analyses of aerial images, a 25 m gridded DEM has been considered best suited to regional-scale approaches (Huggel et al. 2003a). In all instances, these studies have recognized the powerful capabilities of GIS to integrate remotely sensed surface mapping, digital terrain modelling, and combined hydrological and empirical approaches to provide ‘worst-case’ event path modelling, recognizing affected terrain within a maximum defined runout distance.
3. Research Results

3.1 Remote sensing of glacial terrain (Papers I and V)

Methods were aimed at mapping and monitoring glacial ice, glacial lakes, and glacial-related debris accumulations, as an initial step towards the detection of potential source areas where ice avalanches, glacial floods, and debris instabilities might originate. Because of the spatial resolution and coverage (15 m pixel, 60 km swath), temporal resolution (16 day revisiting time), and free data sharing agreements associated with ASTER, imagery from this sensor is well suited for the requirements of regional-scale mapping, and ongoing monitoring in the Southern Alps of New Zealand. Analyses were all based on a scene from January 29, 2002, with repeat imagery used from January 24, 2006, April 08, 2007, and February 17, 2009.

3.1.1 Glacial lakes

Lakes in the Aoraki/Mount Cook region vary from small dark blue mountain lakes and supraglacial ponds, to large, highly turbid, silty-grey proglacial lakes forming at the termini of valley glaciers, particularly east of the Main Divide. These wide-ranging turbidity levels meant that no single wavelength combination provided satisfactory classification of all lakes in the region, with the spectral signature of highly turbid water appearing closer to that of glacial ice rather than other clearer water bodies (Paper I). Attention was therefore given to establishing a sequence or combination of image analyses, to best map all lakes in the region. For lakes with high suspended sediment content, the Normalised Difference Vegetation Index (NDVI) using the Near Infrared (NIR) (3) and red (2) bands of ASTER has previously been suggested as a potential classification method (Huggel 2004), and has also been used extensively in the current study in a more traditional role for vegetation mapping (Papers, I, II, and V). Using the NDVI to map turbid lakes required an additional threshold of the green (1) band, removing misclassifications over large areas of bluish glacial ice and crevassed areas (also showing a blue hue). Mapping the remaining clearer water bodies was best achieved using a ratio between red (2) and green (1) bands. The combined approaches identified over 60 lakes within the glaciated landscape of the Aoraki/Mount Cook region, of which 54 were larger than 1500 m² (Paper V).
Volume measurements have only been made sporadically for some of the larger proglacial lakes (Hochstein et al. 1995; Warren and Kirkbride 1998; Röhl 2005), but estimates based on an empirical relationship between lake area and volume (after Huggel et al. 2002) suggest the majority of lakes in the region currently contain small water volumes (< 0.1 x 10^6 m^3), while 14 lakes have estimated volumes larger than 1 x 10^6 m^3, with the Tasman Glacier lake containing more than 200 x 10^6 m^3 (Paper V). These larger volume lakes prevail in low elevation proglacial areas (below 1000 m) where slope gradients are generally low, although smaller volume lakes are observed in higher elevation steep cirques, primarily east of the Main Divide.

![A) January 2006](image1) ![B) April 2007](image2)

Figure 9. Oblique view of the Tasman Glacier showing the spectacular breakup of glacial ice into the lake captured by repeat ASTER images. The length from the outlet to the highest extension of the lake arm is ~5.5 km in both images. Aoraki/Mount Cook (3754 m) is visible top left, with the terminus of the Murchison Glacier visible top right.

Repeat analyses have shown that large proglacial lakes at the termini of the Mueller, Hooker, Tasman and Murchison glaciers have expanded rapidly over the past decade (Papers I and V), highlighted by an overall 20 % increase in lake area measured within the Aoraki/Mount Cook region between 2002 and 2006. This expansion has continued over recent summers with often spectacular ice calving observed at the termini of large valley glaciers (Fig. 9). Analyses of the Tasman terminus suggest the lake is now squaring off at the calving front, just as the Hooker Glacier lake has done previously, and therefore, might similarly begin to show a deceleration in growth as the length of the calving front reduces (Fig. 10). Because these large proglacial lakes are typically dammed by well vegetated, gently sloping outwash gravels (Paper V), natural dam
failure is considered unlikely. However, as the lakes continue to grow, so does the potential for large magnitude impacts from ice, debris, or rock, producing displacement waves and flooding.

![Glacial lake development at the termini of the Mueller, Hooker, and Tasman Glaciers, 1994 – 2009](image)

Figure 10. Glacial lake development at the termini of the Mueller, Hooker, and Tasman Glaciers, 1994 – 2009 (Based on methodologies and results presented in Paper I).

### 3.1.2 Glacial ice

Two separate methods based on the Visible Near Infrared (VNIR) and Shortwave Infrared (SWIR) wavelengths of ASTER have been used to map perennial ice in the Aoraki/Mount Cook region (Paper I). The Normalised Difference Snow Index (NDSI) (Crane and Anderson 1984) makes use of the green (1) and SWIR (4) bands of ASTER. Alternatively, a simple ratio between Near Infrared (NIR) (3), and SWIR (4) (Kääb et al. 2002; Paul et al. 2002) has been used, with comparisons between the two methods supported using glacier outlines digitized manually from high resolution (1 m) IKONOS imagery of the Hooker Valley (Paper I). Particular emphasis was given to the performance of the two methods on steep faces where shadowing can be
problematic. In general, the ratio method provided better distinction of glacial ice in areas of shadow, and proved more sensitive around rock outcrops and along glacier margins where thin, sporadic debris cover can occur and ‘mixed’ pixels become evident. An overall 90 % confusion matrix accuracy was recognized between the manually digitized glacier ice and the ASTER ratio-derived glacier map. Some steep snow filled gullies and snow covered rock faces proved difficult to distinguish from surrounding ice, either with automated classification or manual digitization of high resolution data. Current optically based methods are also unable to distinguish thick debris covered glacial ice from surrounding terrain. Although this distinction may be possible via a combination of multi-spectral and topographic analyses (Paul et al. 2004a), for applications relating to ice avalanche detection, the exclusion of predominantly flat, debris covered ablation areas is considered appropriate.

Over 400 km$^2$ of debris-free glacial ice was mapped across the entire Aoraki/Mount Cook region, of which ~200 km$^2$ formed slopes steeper than 25°; a conservative slope threshold used to classify potentially unstable glacier ice in previous studies (Huggel et al. 2004a; Salzmann et al. 2004), on the basis of empirical evidence from the European Alps (Alean 1985). Subsequent ice avalanche modelling was restricted to larger potential events (> 2500 m$^2$), thereby excluding falsely classified small seasonal snow patches from the analyses (Paper V). Despite a general pattern of ice retreat since the Little Ice Age (LIA) (Hoelzle et al. 2007), multi-temporal glacial mapping revealed some small steep glaciers have significantly advanced between 2002 and 2006, up to 100 – 140 m for the Stocking and Eugenie Glaciers, east of the Main Divide (Paper I). This glacial advance continues a trend shown by highly responsive glaciers during the late-20th Century (Salinger et al. 1983; Chinn et al. 2005). In general, the potential for ice instabilities is likely to increase as glacier tongues retreat to higher elevations, where topographic slope gradients are steeper (Paper V). In addition to the direct application towards ice avalanche modelling, ASTER derived glacial ice extents were also used to examine historical ice recession occurring near the source areas of several recent bedrock slope failures in the Aoraki/Mount Cook region (see Section 3.3.2, and Paper III).

3.1.3 Debris accumulations

Mapping and distinguishing debris covered surfaces is an initial step towards identifying where potentially steep and unstable accumulations might provide the source area for para- or peri-
glacial debris flows or debris avalanches (Papers I and V). In addition, remote sensing based
debris mapping has been used in the recognition of steep moraine dammed lakes (Paper V), from
where rapid entrainment of debris is likely to accompany any flood event. Following the
mapping of all lake, perennial ice, and vegetated surfaces, an image texture algorithm (based on
pixel variance within a moving window) was used to distinguish the relative uniformity of non-
vegetated debris surfaces, compared to the rough, angular appearance of bedrock (Paper I).
While a similar approach had earlier been introduced with higher resolution IKONOS imagery
(Huggel et al. 2004a), an application using ASTER provides better potential for regional-scale
applications, although recognized surface features are restricted to larger accumulations,
excluding smaller debris cones for example. The procedure was improved with the addition of a
shadow mask, calculated from a 25 m DEM using sun altitude and azimuth values at the time of
satellite overpass. Visual comparisons made with IKONOS imagery revealed remaining
problems on some particularly uniform greywacke bedrock slopes, where debris was falsely
classified by the automated procedure, and debris within narrow channel confines could not be
recognized. However, major stream networks could be extracted using DEM-based hydrologic
stream ordering (Fig. 11) (Paper I), and in a subsequent step, inspected with higher resolution
imagery to detect significant sediment build-up.

A distance buffer extracted all debris accumulations within 750 m of ASTER mapped glacial ice,
thereby restricting subsequent debris flow modelling towards situations where current and recent
glacial processes might directly influence debris accumulation or mobilization into a mass
movement (Paper V). A slope threshold range of 25 – 38° has typified initiation zones observed
internationally for debris flow events (eg. Hungr et al. 1984; Takahashi 1991; Rickenmann and
Zimmermann 1993). This slope range encompasses the 35° angle of repose identified for typical
talus slopes in the central Southern Alps (Whitehouse and McSaveney 1983), and a 29 – 32°
modal slope angle characterizing these slopes (Paper III).
Figure 11. A) ASTER-mapped steep para- and peri-glacial debris accumulations (orange) and topographically-mapped higher order stream channels (blue) within the lower Hooker and Mueller glacial valleys. B), IKONOS sub-scene showing Stewart (S), Hayter (H), and Eugenie (E) streams (April 2002). Debris avalanching, debris flows, and snow avalanching have eroded the western lateral moraine of the Hooker Glacier, destroying the former access to the Hooker Hut. Recent debris flow lobes (yellow arrow and circle) have originated from a steep talus slope at the base of a bedrock gully. C) Debris instabilities are viewed from the opposite moraine wall, November 2007 (S. Allen). (Based on results presented in Papers I and V).
Steep para- or peri-glacial debris accumulations identified in the Aoraki/Mount Cook region included numerous talus slopes in close proximity to glaciers or large perennial snow patches, but also lateral moraines above downwasting valley glaciers, terminal deposits from steeper glaciers, and cirque glacial moraines (Fig. 11). Steep accumulations predominate within an elevation range between 1000 – 2000 m, with gentle slope gradients prevailing at lower elevations, and slopes above 2000 m typically either ice covered or too steep to allow significant debris accumulation (Paper V).

3.2 Permafrost distribution modelling (Papers II and IV)

Establishing the distribution of permanently frozen ground in the Aoraki/Mount Cook region provides the necessary basis from which discussions regarding atmospheric warming, permafrost degradation, and slope instabilities can proceed.

3.2.1 Topo-climatic permafrost distribution estimate

A first approach towards estimating mountain permafrost distribution in the Southern Alps (Paper II) has been based upon a topo-climatic key, incorporating ‘rules of thumb’ describing permafrost occurrence in the European Alps (Haeberli 1975). These fundamental relationships consider the decrease in mean annual air temperature (MAAT) occurring with elevation, variation in solar radiation across different slope aspects, and the influence of snow build-up at the base of steep slopes, all of which will influence mean annual ground temperatures. Values given in the original topo-climatic key describe the lower limit of probable permafrost occurrence (ie, permafrost observed in > 50 % of instances), and also the lower limit of a transitional zone where permafrost remains ‘possible’, and likely to exist only within a marginal or warming state (approaching 0 °C). For each of eight primary slope aspects, the key values were adjusted for the central Southern Alps using local MAAT calculations (1982 – 2007) and associated 0°C isotherm elevations for air temperature (Paper II). To account for the climate variability across the region, MAAT values were obtained at Franz Josef, Aoraki/Mount Cook village, and Lake Tekapo climate stations, with selected free air lapse rates ranging from 0.005 to 0.0075 °C m⁻¹ reflecting the moisture gradient occurring from northwest to southeast. A GIS-based spatial model was created, with 0°C isotherm elevations interpolated between the three climate stations to establish local permafrost limits across the region, and all slope pixels located
above these ‘possible’ and ‘probable’ limits were identified (Fig. 12A). Possible lower permafrost limits are positioned above 3000 m for steep sunny slopes about and west of the Main Divide, decreasing to nearer 2200 m for shaded slopes, and down to 200 lower in mountain ranges further to the southeast (Paper II).

The extreme maritime climate near the Main Divide has prevented rock glacier formation, but in the drier mountain ranges towards the southeast, talus development is widespread and rock glaciers are found, with numerous fossil and active examples documented in the Ben Ohau Range (Brazier et al. 1998). The initiation elevations for all active rock glaciers in the Ben Ohau Range were found to be located 20 – 180 m lower than the estimated possible permafrost limit for these slopes (Paper II). The preservation of permafrost at the shaded foot of steeper slopes could help explain this discrepancy, with late lying snow-cover shielding the underlying talus from spring and early summer warmth, but it also appears likely that at least in the Ben Ohau Range, the approximation of permafrost limits may be too high. Fossil rock glaciers were all identified below the current estimated permafrost limits, and in some instances are known to represent peri-glacial relics from considerably cooler past climates (Birkeland 1982). Remotely sensed vegetation patterns are strongly influenced by snow cover variation and geomorphic disturbances across the region, and therefore offered little potential for validating or improving estimated permafrost limits. On shaded slopes, permafrost limits were found to be at least 200 m higher than the maximum altitude of vegetation growth (scrub, grassland, or forest) on the same slopes, while on sunny slopes, permafrost limits were up to 1000 m higher than corresponding vegetation limits (Paper II).

3.2.2 Rock temperature measurement and permafrost modelling

While first-order estimates of permafrost distribution are a useful step towards better understanding permafrost and slope stability interactions in remote regions (see Section 3.3.2 and Paper III), they have mostly been developed for and validated using data from debris covered slopes, and therefore provide only an approximation of thermal conditions occurring about complex steep bedrock topography. Following procedures developed in the Swiss Alps (Gruber et al. 2003), a network of 15 miniature temperature dataloggers were installed onto steep rock faces in the Aoraki/Mount Cook region and left for a twelve month period in order to measure Mean Annual near-surface Rock Temperature (MART) (Paper IV). Installation sites
were positioned at elevations above and below estimated permafrost limits (after Paper II), and
aimed to sample a variety of slope aspects both near the humid Main Divide, but also from the
drier Liebig Range, located further leeward. Because deeper sub-surface temperatures reflect a
longer-term average of near-surface temperature (Gruber and Haeberli 2007), MART
measurements for the single year were adjusted using longer-term air temperature records from
Aoraki/Mount Cook village (Paper IV).

Figure 12. Modelled permafrost distribution about the Aoraki/Mount Cook massif and Main Divide of the
central Southern Alps, based on A) a topo-climatic estimate using MAAT (Paper II), and B) a distribution
model based on near-surface rock temperature measurements (Paper IV).

MART varied by more then 7 °C between logger sites, with corresponding 0°C isotherm
elevations for rock temperature (E0) ranging from 2465 to 3514 m, although measurements were
lacking from the most super-shaded bedrock slopes. A simple linear regression ($R^2 = 0.78$)
relating measured E0 to GIS-modelled potential annual clear-sky radiation ($R_{pol}$) at each logger
site provided a basis for permafrost distribution modelling across the study region. This
statistical modelling approach identified three permafrost zones; a zone > 300 m above modelled
E0, where terrain is considered mostly permafrost; a zone > 300 m below modelled E0 where
terrain is unlikely to be permafrost; an intermediate zone where marginal or ~0 °C permafrost is
considered possible, and where warm, degrading conditions are expected (Fig. 12B). In theory, greater cloud cover nearer the humid Main Divide was expected to dampen the influence of potential solar radiation compared to drier leeward regions (see Paper II) (Gruber et al. 2004a), but there were no obvious corresponding variations in permafrost limits recognized across the study region (Paper IV). In fact, two outlying datasets from near the Main Divide indicated measured E0 to be more than 500 m higher than modelled E0 at the corresponding logger sites. In these instances, both from sun exposed slopes, maximum daily rock temperatures frequently exceeded 35 °C. This was thought to possibly result from terrain reflected radiation exacerbating the warmth of rock walls which are surrounded by extensive, highly reflective glacial neves near the Main Divide.

Table 2. A comparison of permafrost terrain classifications for the Aoraki/Mount Cook region, central Southern Alps, New Zealand. Classifications are derived from a topo-climatically based estimate using MAAT (Paper II), and from a distribution model based on near-surface rock temperature measurements (Paper IV).

<table>
<thead>
<tr>
<th>Permafrost classification</th>
<th>Total land area</th>
<th>Lower elevation limit (north facing slopes)</th>
<th>Lower elevation limit (south facing slopes)</th>
<th>Proportion &gt; 45° slope</th>
<th>Proportion glacierised</th>
</tr>
</thead>
<tbody>
<tr>
<td>Paper II</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Possible permafrost</td>
<td>12 km²</td>
<td>2550 m</td>
<td>2130 m</td>
<td>37 %</td>
<td>45 %</td>
</tr>
<tr>
<td>Probable permafrost</td>
<td>18 km²</td>
<td>2750 m</td>
<td>2190 m</td>
<td>47 %</td>
<td>41 %</td>
</tr>
<tr>
<td>Paper IV</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Marginal permafrost</td>
<td>42 km²</td>
<td>2240 m</td>
<td>2000 m</td>
<td>41 %</td>
<td>44 %</td>
</tr>
<tr>
<td>Mostly permafrost</td>
<td>9 km²</td>
<td>2970 m</td>
<td>2380 m</td>
<td>75 %</td>
<td>41 %</td>
</tr>
</tbody>
</table>

Compared to the initial MAAT based estimate of permafrost distribution (Paper II), modelling based on actual rock temperature measurements (Paper IV) suggests that the zone in which warm, marginal permafrost conditions might be expected, potentially encompasses a much larger elevation range, restricting colder probable/mostly permafrost conditions to the highest bedrock slopes (Fig. 12) (Table 2). With no significant variation in permafrost limits noted across the study region, marginal conditions might occur down as low as 2000 m even near the humid Main Divide, in situations where extreme topographic shading limits potential radiation inputs. Hence, permafrost is realistically expected at elevations even where MAAT is positive (Paper IV), and even north facing slopes might feature permafrost to low elevations where they are sufficiently
shaded by neighbouring topography (Table 2). Nearly half of all slopes currently located within potential permafrost terrain are covered by glacial ice, and therefore subject to additional warming from latent heat associated with melt. The reduction of perennial ice cover over coming decades may allow permafrost to penetrate into previously insulated slopes (Wegmann et al. 1998), and where glaciers are polythermal (eg. hanging glaciers), complex thermal regimes might exist at the underlying ice-rock interface (Haeberli et al. 2004; Huggel 2008).

3.3 Bedrock slope failure analyses (Papers III, IV, and VI)

510 bedrock slope failures in the central Southern Alps were mapped, compiled, and integrated into a GIS based landslide inventory (Paper III). The majority of failures were extracted from a recently completed digital geological map of the region (Cox and Barrell 2007), but numerous recent failures have been identified and reconstructed on the basis of remotely sensed imagery and field studies (eg, Paper VI). Rock avalanches were distinguished from other landslide types, with the combined dataset providing the basis for a regional-scale study of slope failure distribution (Paper III), and discussion of potential climate change impacts relating to glacier recession and permafrost degradation (Papers III and IV).

3.3.1 Topographic and geological distribution

Nearly 2 % of the land area in the central Southern Alps has been identified as landslide affected terrain, consisting of hillslopes disturbed by either landslide source or deposit areas (Paper III). Evidence for rock avalanche failure has been observed most frequently from greywacke and cleaved greywacke bedrock which outcrops steeply about and east of the Main Divide. For other landslide types, semischist and schist bedrock slopes show the highest proportional evidence of failure, predominating west of the Main Divide where tectonic uplift, precipitation, and erosion are all several orders of magnitude higher than in the east. The distribution pattern and characteristics of landsliding across the Southern Alps were further considered within the western, axial, and eastern geomorphic regions described by Whitehouse (1988). In all three regions, rock avalanches and other landslides have prevailed most frequently from slopes facing west to northwest (Paper III), suggesting a structural control on landslide distribution, with bedding and schistosity predominantly dipping towards this direction. However, rock avalanches occurring over the past 100 years have prevailed from east to southeast facing slopes about the
Main Divide, during what has been a period of relative seismic inactivity. Hence, comparisons between the historic and prehistoric landslide distribution were not considered appropriate, given that the prehistoric record is considered to be significantly influenced by earthquake generated events (Whitehouse and Griffiths 1983; Bull and Brandon 1998; Orwin 1998; Smith et al. 2006), and is also likely to be missing many events for which deposits have been removed by glacial transport, or eroded from river valleys. The altitudinal distribution of slope failures showed that at elevations above 2500 m, rock avalanching appears the only large magnitude mechanism of bedrock failure, supporting earlier suggestions by McSaveney (2002). In comparison, other landslide types have been observed most frequently within a subalpine zone (1000 – 2000 m), where hillslope angles are less steep, glacial ice less prevalent, and permafrost is unlikely to be a relevant factor (Papers III). Slope angle is considered a fundamental driver of bedrock instability, and a notable increase in the proportion of rock avalanche failures occurring on slopes steeper than 50° was observed in the central Southern Alps.

3.3.2 Glacial change and bedrock stability

The impact of lowered glacial surfaces causing deep-seated slumping and complex failure of surrounding valley walls in the Aoraki/Mount Cook region has been recognized (Blair 1994; Cox and Barrell 2007). Within the current study, attention turned towards the potential impacts of glacial ice recession and permafrost degradation relating to catastrophic rock avalanches, given that these events have clearly prevailed from glaciated, steep greywacke and fractured greywacke slopes about the Main Divide over the past century (Paper III). Uncertainties in the prehistoric record of events occurring from these slopes have been acknowledged, and therefore, no statistical basis exists to support any notion of increased frequency or magnitude of recent bedrock failures. However, a comparison of estimated permafrost limits across the region (after Paper II) with the source areas mapped from 19 rock avalanches occurring over the past 100 years, revealed 12 instances in which part of the source area has been located within 300 m of the estimated zone of possible permafrost, where warm, degrading conditions are expected (Paper III). Considering only the lower portion of the source area (where pre-failure stresses are normally greatest), 7 events have appeared to initiate from within 300 m of the lower limit of possible permafrost, including a sequence of four rock avalanches occurring during January, February, and April of 2008. MART values ranging from -1.3 to +2.5 °C were modelled at the base of these four source areas, suggesting warm, degrading conditions were likely (Paper IV).
The Vampire Peak rock avalanche of January 2008 followed a prolonged period of melting conditions near the rock surface (Fig. 13), with nearby air temperature records confirming above average temperatures were experienced throughout January and May of 2008 (Cox et al. 2008). The most recent rock avalanche recognized from ASTER imagery occurred from the southern slopes of Malte Brun during February of 2009, where modelled MART ranging from -2.0 to -0.6 °C characterized the source area (Paper IV).

Figure 13. A) The Vampire Peak rock avalanche of January 2008 occurred in two pulses, and deposited on top of an older event from 2003 (Paper VI). B) Modelled MART (Paper IV) indicated at the top and bottom of the 2008 detachment zone. Running water can be seen seeping from discontinuities in the rock face. The light coloured scar to the right is the source area from the earlier, 2003 failure. C) Temperature records from two nearby rock temperature dataloggers (Paper IV) indicate near-surface temperatures were sustained above 0 °C in the months leading up to the 2008 failures. The source area is facing southeast. Photos: D. Bogie (16/01/08).
Two recent rock avalanches from below Mount Beatrice (Hooker Glacier) and from above the Murchison Glacier, along with numerous smaller rockfalls, have occurred from steep valley walls uncovered and destabilized by ice retreat since the LIA (Paper III). The majority of recent events have however originated from higher elevations above the glacial Equilibrium Line Altitude (ELA), where recognition of surface lowering and/or recession of steep hanging or cliff glaciers is difficult to quantify in the absence of high resolution, current terrain data. Perennial ice extents within the Mueller Glacier valley were reconstructed from aerial photography and recent ASTER-based mapping (after Paper I), revealing notable ice recession near and within the source area of several recent rock avalanches occurring from the pervasively fractured hangingwall of the Main Divide Fault Zone (Paper III).

Considering the various analytical uncertainties associated with the prehistoric and historic landslide distribution in the central Southern Alps, discussions relating to future atmospheric warming and slope stability have been directed towards the most fundamental and robust expectations (Paper III). In particular, a future increase in the frequency of rock failure occurring from the highest slopes about the Main Divide has been suggested, as ice levels and permafrost limits respond to accelerated atmospheric warming (IPCC 2007). With ~45 km² of extremely steep (> 50°) hillslope area currently ice covered or positioned above estimated permafrost limits, the potential influence of future warming is not insignificant (Paper III). However, this influence is likely to be restricted primarily to greywacke slopes at elevations where rock avalanches appear the dominant mode of slope failure, with other lithologies and landslide mechanisms (including debris instabilities) predominating at much lower elevations. On southeast – eastern aspects about the Main Divide Fault Zone, the most pervasively fractured and steepest bedrock prevails, and these slopes are therefore likely to continue to produce a disproportionate frequency of rock mass failure occurring as rock avalanches. Seismicity has been acknowledged as the probable dominant predisposing and triggering mechanism of large bedrock failures across the central Southern Alps, but the role of permafrost degradation and ice recession is likely to add to the likelihood of catastrophic failures occurring from steep, high elevation slopes. With the continued expansion of proglacial lakes, particularly east of the Main Divide (eg. Fig. 10), the future potential for flooding from large magnitude rock impacts into glacial lakes has been illustrated and suitable modelling approaches have been introduced (see Sections 3.4 and 3.5).
3.4 Hazard modelling (Paper V)

Potential flood and mass movements of ice and debris in the Aoraki/Mount Cook region have been modelled using the modified single flow (MSF) routing algorithm introduced by Huggel et al. (2003a; 2004a), which enables regional-scale detection of hazardous intersections between event paths and human use systems. At this level of the hazard investigation, worst-case scenarios were simulated, whereby modelled flow paths were continued until a probable maximum runout was reached, defined by an angle ($\alpha$) measured between the starting and end points of a mass movement. In addition, a physically based numerical mass movement model has also been introduced for simulating potential rock avalanches, providing an accurate representation of flow path dynamics, and with capabilities for estimating potential magnitudes, including impact volumes into glacial lakes.

3.4.1 Glacial floods, ice avalanches, and debris flows

Because the majority of glacial lakes in the Aoraki/Mount Cook region were identified within lower elevation, low gradient proglacial areas, it was considered inappropriate to assume all potential floods would transform into debris flow events. Therefore, remotely sensed debris and vegetation mapping (Paper I) were combined with topographic analyses to distinguish glacial lakes formed behind steep morainic debris, from which catastrophic debris flows were considered possible (Paper V). A distance of 500 m was used to encompass the lake dam and surrounding outlet channel area (after O’Connor et al. 2001), and a slope threshold of 10° distinguished outlet areas in which sediment entrainment was considered likely (Hungr et al. 1984; Clague and Evans 1994). Several examples of lakes formed within steep vegetated or non-vegetated debris were identified, mostly in remote cirque basins east of the Main Divide. Modelled debris flows initiating from lake flooding in these instances (based on a runout distance defined by $\tan \alpha = 0.19$) did not reach any areas of human use (tourist or residential dwellings, remote buildings, vehicle roads or tracks, foot tracks, utilities) (Paper V). In contrast, potential clear water flood-waves from other lakes dammed by bedrock or more commonly formed within low-gradient moraine or outwash gravels, will likely attenuate slowly over exceptional distances (eg, Hewitt 1982), and therefore may impact upon infrastructure located on or near downstream braided river flood-plains both east and west of the Main Divide (Paper V). Although most large, expanding proglacial lakes in the region exhibit well vegetated, historically
stable outlet areas, future dam overtopping and flooding from displacement waves remains possible. The inability of the MSF model to simulate wave-height in these instances is a concern, meaning that any flood-wave is initially confined to the existing lake outlet channel, rather than catastrophically inundating the entire dam area, as has been observed previously from rock avalanche generated displacement waves occurring in the Aoraki/Mount Cook region (McSaveney 2002).

The probable maximum runout for ice avalanche modelling was defined by $\tan \alpha = 0.31$ (Aleam 1985), preventing any possible events from reaching village infrastructure or main roads (Paper V). Numerous backcountry huts and walking tracks located in closer proximity to steep glaciers were identified within potential ice avalanche paths, as was illustrated for the lower regions of the Hooker and Mueller Glaciers, which are easily accessible to day-visitors from the Aoraki/Mount Cook village. West of the Main Divide, in the upper Copland Valley, ice avalanches modelled using considerably smaller travel distances ($\tan \alpha = 0.67$) still appeared capable of reaching the valley floor. For all steep glaciers, direct indicators of instability such as crevasse orientation and rapid displacements are best monitored for individual cases (Wegmann et al. 2003), and direct measurements of hydrological and thermal conditions at the glacier bed are probably unwarranted, but may be inferred from other proxy variables such as air temperature or meltwater outflows. Fundamentally, use of the hydrologically based MSF approach for ice avalanche modelling remains problematic, given that the often significant airborne powder component of an ice avalanche detaching from steep cliff or hanging glaciers is not constrained by topography. As a coarse approximation, buffer distances could be applied to modelled flow paths presented here, in order to visualize and identify potential zones affected by an air pressure blast, including upslope areas.

A probable maximum runout defined by $\tan \alpha = 0.19$ was used for para- and peri-glacial debris flow modelling, based on debris flow observations from alpine regions primarily in the European Alps (Rickenmann and Zimmermann 1993; Rickenmann 2005) (Paper V). However, this runout distance appeared to far exceed the historical extent of debris flow activity that has affected Aoraki/Mount Cook village, and also exceeded probable runout distances modelled on the basis of available catchment size (Rickenmann 2005). Reduced runout distances were therefore also simulated, with minimum $\alpha$ increased by 50 and 100 %. The majority of potential debris flow problems initiating from para- and peri-glacial origins for all runout scenarios were identified
east of the Main Divide, where a drier climate favours extensive talus development and moraine preservation (Whitehouse 1988; Brazier et al. 1998), and vehicle tracks extend further into the headwaters of glaciated valleys where farm or recreational huts are located. West of the Main Divide, debris flows originating from glaciated catchments may not directly reach road or village infrastructure (Paper V), but steep gorges may become blocked, and debris remobilized to create a far reaching secondary hazard (Davies and Scott 1997). With over 80% of all steep debris sources located below 2000 m, and the current lower altitudinal limit of permafrost considered to be at or above 2000 m (Papers II and IV), future permafrost warming is unlikely to be a significant concern relating to debris instabilities in the Aoraki/Mount Cook region (Paper V).

However, ongoing monitoring of glacial change will be required to detect rock masses freshly exposed to altered weathering regimes, and newly uncovered moraines. This may be particularly important where faulted bedrock slopes are uncovered, for example, along the Sealy Range and hangingwall of the Main Divide Fault Zone, from which enhanced sediment delivery can be expected from zones of crushed fault gouge and breccia (Korup 2004). Accumulation depths at higher elevations will however be limited by steep slope gradients which frequently exceed the angle of repose for talus slopes.

3.4.2 Rock avalanches

Given a steep tectonically active geological setting, coupled with potential influences of recent and future glacial changes on bedrock stability (Paper III), approaches towards glacial hazard modelling in the Aoraki/Mount Cook region have included the simulation of large bedrock failures (Paper V). Noetzli et al. (2006) have previously demonstrated the ability of the MSF model to crudely reconstruct the flow paths of several large rock-ice avalanches in the European Alps, providing a useful tool for early hazard recognition. In view of potential impacts into rapidly expanding proglacial lakes, river damming, and major secondary hazards, the focus in this study was given towards methods that provide a higher level of flow detail, including flow dynamics and magnitudes. A numerical rapid mass movement model (RAMMS), developed by the WSL Institute for Snow and Avalanche Research, Davos Dorf, Switzerland, was used for this purpose (Paper V), having been developed primarily on the basis of experiments relating to snow avalanches, but with capabilities to simulate other mass movements of ice, rock, and debris (Christen et al. 2008).
RAMMS was used to reconstruct the 1992 rock avalanches from Mount Fletcher which followed a common path, assuming a failure volume of $8 \times 10^6$ m$^3$ comprised mostly of bedrock with minor amounts of ice and firm (Paper V). Flow velocities and areas of barrier run-up exceeding 300 m were well represented in the simulation, and maximum flow heights enabled an estimation of the mass volume that deposited into the proglacial lake. Although no field measurement of this deposition volume has been possible, nearly $1 \times 10^6$ m$^3$ was calculated from RAMMS to enter the lake at a maximum velocity of 20 ms$^{-1}$ (Paper V), proving responsible for the displacement of a significantly larger volume of water owing to the seiching effect of the impact waves (McSaveney 2002). The capabilities of RAMMS for simulating considerably smaller volume bedrock failures ($1 - 2 \times 10^5$ m$^3$) were evaluated by modelling two recent rock avalanches occurring from Vampire Peak in 2003 and 2008 (Paper V). Both events have been reported in detail by Cox et al. (2008), and Paper VI. For modelling the smaller volume Vampire Peak failures, the limitations of using a 25-m resolution DEM were partially overcome by resampling the data to a 10-m grid, thereby improving model precision (Paper V). However, the accuracy of the DEM remained a limiting factor, because it was unable to represent small scale topographic features that influence surface friction and could not depict the rapidly changing glacial topography.

Calibration of the RAMMS simulations required that the rock avalanche paths aligned geometrically in three dimensions and matched the time signature of the actual events. The later constraint was extracted from seismic data (Paper VI and McSaveney 2002), although this can give an unclear signal of avalanche timing in the case of a progressive failure. In the Vampire Peak simulations, Coulomb $\mu$ and turbulent $\xi$ friction parameters were adjusted to reduce the runout distance of the 2008 failure, relative to the earlier event. The 2008 rock avalanche was thought to be less mobile because of the altered surface friction caused by the earlier 2003 avalanche deposit onto the previously smooth glacier surface (Paper V). Further model runs for both large and smaller magnitude events from New Zealand (Schneider et al. 2008b) and other mountainous regions are expected to establish a range of best-fit friction parameters that can be applied to simulate potential events where hazardous scenarios are recognized (see Section 3.5).
3.4.3 Process interactions

Internationally, the most devastating natural disasters originating from glacial environments have involved chain reaction type events, where one or more processes transform into or initiate a secondary event (e.g. Carey 2005; Huggel et al. 2005). In the Aoraki/Mount Cook region, the most likely scenario originating from a glacial environment would involve a mass movement of ice, rock or debris depositing into a lake and initiating a catastrophic flood-wave (Paper V). While lakes were most frequently observed within the path of potential debris flows, these events are unlikely to enter the lake with the same catastrophic velocity and impact as a mass movement of ice or rock. Thirty-six lakes were identified within the path of ice avalanches travelling a probable maximum runout length of tanα = 0.31, reducing to 11 lakes if a shorter runout distance is considered (tanα = 0.67). The Mueller Glacier lake was given particular attention, because the continued growth of this lake towards the steep ice cliffs beneath Mount Sefton is increasing the potential for ice avalanches to reach the lake (Paper V). Mass movements coming from the Main Divide can in this instance direct full wave energy towards the lake outlet, whereas many other lakes in the region are constrained by U-shaped valleys, with their longitudinal axis forming perpendicular to adjacent mountain slopes, and therefore, are less vulnerable to catastrophic impacts from ice or rock. Low gradient braided rivers flow from all expanding proglacial lakes east of the Main Divide, suggesting flood-waves generated from these lakes are not likely to transform into debris flows. West of the Main Divide, river valleys are steeper and narrow, but lake volumes are lower, and have shown no significant growth over the past decade. Although not analysed within the scope of the current project, narrow river gorges may become obstructed by large mass movements that may or may not originate from para- or peri-glacial slopes (Hancox et al. 2005; Korup 2005a). Hence, glacial hazards, and their interactions, have been appropriately recognized as only one group within a myriad of hazardous mass movement and flood problems affecting the central Southern Alps (Paper V).

3.5 Method and data integration

In combination, the various methods explored in this study lead towards the recognition of potentially hazardous situations, directing local authorities and further scientific investigation towards localities where glacial-related mass movements or floods might affect infrastructure and human activities (Fig. 14). The integration of these methods is illustrated here using the
lower Mueller Glacier as an example. The recent and continued expansion of the glacier lake has been recognized, along with the potential for steep ice cliffs to avalanche into the lake as it expands closer towards the Main Divide slopes (Paper V). Further consideration is given here to the possibility and consequences of a large bedrock failure occurring from these slopes. Across the Aoraki/Mount Cook region, permafrost degradation is considered unlikely to be directly significant in relation to debris instabilities or destabilisation of lake dams (Fig. 14, dashed lines), owing to the predominance of related landforms occurring below current permafrost limits, and steep topography that is unfavourable for lake development or large debris accumulations at higher elevations (Papers III and V). Indirectly however, potential exists for permafrost related bedrock failures to cause catastrophic flooding, river blockage, or transform into rapid debris flows following the entrainment of ice or snow (Fig. 14).

Figure 14. Conceptual framework showing the integration of methodologies used to study glacial hazards in the Aoraki/Mount Cook region. Results presented in this study have primarily related to levels 1 (green) and 2 (yellow) within a downscaling approach towards detailed local-scale hazard assessment.
Figure 15. Lower Mueller Glacier showing historic landslide distribution in relation to A) bedrock geology (Paper III) and B) modelled bedrock temperatures within permafrost terrain (Paper IV). C) RAMMS simulation showing a hypothetical rock avalanche detaching from The Footstool and MSF simulation of a related flood-wave from the proglacial lake (Paper V). D) View of Mueller Glacier from near the summit of The Footstool, showing the path of an earlier flood event through the right lateral moraine of the glacier.
A hypothetical rock avalanche is modelled, detaching from a steep buttress supporting the east–southeast face of The Footstool (Fig. 15). This detachment zone is selected as an example only, and may not necessarily be any more predisposed to failure than many high elevation glaciated bedrock slopes of the Main Divide, particularly where warm, degrading permafrost may be an additional destabilizing factor (Papers III and IV). Therefore, although beyond the scope of the current project, it is recognized that detailed geotechnical studies are required to reasonably identify where rock avalanche detachments are most likely. Nonetheless, at the regional-scale, some broad predisposing factors can be recognized, and current studies in the European Alps are further bridging the gap between regional and local-scale bedrock stability analyses in glacial environments (Fischer and Huggel 2008). In the central Southern Alps, a recent tendency for larger magnitude rockfall and rock avalanches to detach from steep (> 50°) east–southeastern slopes of the Main Divide Fault Zone has been identified, and these instabilities have been particularly evident from above the Mueller Glacier during recent decades (Paper III). Bedrock forming the east–southeast face of The Footstool grades from cleaved greywacke through to schist, and is cross-cut by discontinuities relating to the Main Divide Fault Zone (Fig. 15A). Maximum slope angles exceed 60° within the modelled detachment zone. Steep glacial ice below the face is likely to have thinned over historical times, but no significant change in ice extent has been detectable from oblique photography. Modelled mean annual bedrock temperatures towards the lower portion of the buttress and on more east sloping facets approach marginal temperatures of ~0 °C (Fig. 15B.).

Numerical modelling of a rock avalanche from The Footstool was based on a comparable volume (12 x 10⁶ m³) and friction parameters used to calibrate simulated events from Mount Fletcher (Paper V), and Aoraki/Mount Cook (Schneider et al. 2008b). Steep ice is evident immediately below the detachment zone (Fig. 15C), and hence, as with many potential bedrock failures occurring form the Main Divide, there is a strong likelihood that significant ice entrainment would occur. The resulting ice-rock avalanche path is split by a ridge below the detachment zone. The flow direction towards the southeast travels a greater distance, and deposition could disrupt the river leading from the Hooker Glacier lake, and/or dam the outlet area of the Mueller Glacier lake. The other flow direction towards the south terminates abruptly after impacting onto the Mueller Glacier and dissipating against the opposing valley wall. Under current conditions this flow direction would appear relatively harmless, but a mass movement along this path could impact directly into an expanded glacial lake in the future. Hence, there is a
need for ongoing monitoring of glacial conditions and reassessment of potential process interactions (Fig. 14), for which satellite based techniques have proven well suited.

Importantly, glacial changes need not necessarily imply that hazard potential will increase, as illustrated for the Mueller Glacier. In 1913, a flood originating from the glacier destroyed a hotel located near the current camping area (Wilson 1968) (Fig. 15D), but subsequent 20th Century lowering of the glacier surface and the exposure of large lateral moraine walls now limits the likelihood of any flood or mass movement problems originating from, or overtopping these lateral margins. A sequence of smaller LIA terminal moraines would help initially constrain any flood-wave originating from the current outlet of the Mueller Glacier lake, such that any flood from this lake is unlikely to affect village infrastructure (Fig. 15C). However, further downstream the flood-wave quickly disperses on low gradient alluvial fans, with road and infrastructure at the Mount Cook airport located towards the centre of the flowpath (Paper V).
4. Conclusions

An integrated study of mass movements and floods associated with glacial change in the Aoraki/Mount Cook region has been completed, with a strong emphasis given to the application of remote sensing and GIS based methods for mapping, modelling, and analysing related processes and terrain. A summary of main findings addressing the four primary research themes and related objectives of this thesis is outlined below, followed by implications and perspectives relating to further research directions in the Southern Alps of New Zealand, and more generally within the international field of glacial hazard research.

4.1 Summary of main findings

1. Remote sensing based mapping of glacial terrain

- Previously established methods using SWIR and NIR wavelengths are suitable for mapping steep non-debris-covered glacial ice in the Aoraki/Mount Cook region, from ASTER imagery.

- Large variations in the suspended sediment content of lake water across the region necessitated a new approach to glacial lake detection. This approach, requiring separate image classification for turbid and clearer water bodies ensured that all glacial lakes in the region could be identified and monitored on the basis of repeat imagery.

- There has been substantial enlargement of many large proglacial lakes over recent years, most notably in low gradient terminal regions of valley glaciers east of the Main Divide, where the largest lakes currently have estimated volumes in the order of $10^6$ – $10^8$ m$^3$. An overall 20% increase in lake area has been measured in the Aoraki/Mount Cook region between 2002 and 2006.

- Steep talus slopes, moraine walls, and other debris accumulations can be distinguished from surrounding bedrock slopes using textural variations observed from regional-scale
imagery. GIS proximity buffer analyses can focus attention towards situations where recent glacial retreat may have exposed new slopes, and/or perennial ice melt may initiate debris instabilities.

- The recognition of debris within the surrounding outlet area of glacial lakes can distinguish lakes dammed by moraine or outwash gravels from other bedrock or potentially ice dammed lakes, while the presence of vegetation within the floodplain of most large proglacial lakes in the Aoraki/Mount Cook region, suggests these lakes have been historically stable and undisturbed by mass movement impacts.

- In all instances, the minimum size of mapped ice, debris, and lake surfaces is limited by the 15 m resolution of ASTER VNIR imagery, and in particular, debris classification which is based upon a 3 X 3 pixel window is unable to distinguish smaller talus slopes, debris cones, and sediment build-ups within narrow stream channels.

- Clear altitudinal patterns in glacial terrain were recognized, with lakes predominating in proglacial terrain below 1000 m, steep debris accumulations between 1000 and 2000 m, and steep ice prevailing above 2000 m.

**II. Validated modelling of mountain permafrost distribution**

- An initial estimate of permafrost distribution across the Aoraki/Mount Cook region indicated that a possible zone of warming or marginal permafrost is positioned above 3000 m on sunny slopes, extending down to 2200 m on shaded slopes about and west of the Main Divide.

- In mountain ranges further leeward of the Main Divide, where conditions are drier, lower permafrost limits were estimated, corresponding well with the distribution of active rock glaciers previously mapped in the Ben Ohau Range.

- A network of near-surface temperature measurements from steep rock walls provided a basis for bedrock temperature and permafrost modelling across the Aoraki/Mount Cook region, considering the influences of solar radiation within steep, complex topography.
Data were measured both near the Main Divide, and from a drier, leeward mountain range, but no obvious variations in near-surface temperatures or permafrost limits were identified between these two sub-regions for steep bedrock slopes.

- A zone of marginal permafrost has been distinguished where modelled mean annual rock temperature (MART) ranges from -1.8 to +1.8 °C, while colder temperatures are considered to indicate mostly permafrost conditions. On extremely shaded slopes, marginal permafrost could extend down as low as 2000 m near the Main Divide, even at elevations where mean annual air temperature remains positive.

- Two outlying data points suggest reflected radiation from extensive glaciers occurring about the Main Divide may heat surrounding rock walls to warmer temperatures than would be expected for similar bedrock slopes located in drier leeward areas, where surrounding slopes are predominantly talus with lower albedo.

### III. GIS based analyses of landslide distribution and related glacial influences

- Rock avalanches have occurred most frequently from greywacke and cleaved greywacke slopes about and east of the Main Divide of the central Southern Alps, and appear the dominant slope failure mechanism above 2500 m. Shallow slope failures, complex deep seated failures, and other failure types prevail at lower elevations and in the schist terrain west of the Main Divide.

- The observed proportion of rock avalanche failure significantly increases for slopes steeper than 50°.

- The prehistoric distribution of rock avalanches and other landslides have predominantly been from slopes facing west – northwest, across western, axial, and eastern domains of the central Southern Alps. This may indicate a structural control associated with the dominant dipping direction of bedding and schistosity, and many events are likely to have been generated by large prehistoric earthquakes.
• Rock avalanches observed over the past 100 years have prevailed from east – southeast facing slopes, particularly about the glaciated hangingwall of the Main Divide Fault Zone, and have all been unrelated to any seismic trigger.

• 12 out of 19 recent rock avalanches have occurred from within 300 m of the lower limit of possible permafrost, with 13 detachment zones characterized by marginal MART values within the range of -1.8 to +1.8 °C. The majority of these failures have occurred from near ridges, where the most rapid degradation of permafrost is expected.

• A sequence of 4 bedrock failures occurred during the warm/dry summer of 2007/08, with modelled MART near the base of these detachment zones ranging from -1.3 to +2.5 °C. The largest of these failures occurred from Vampire Peak (~150,000 m³), and followed a prolonged period of probable melting conditions within the detachment zone.

• The majority of recent rock avalanches have detached from within a close proximity to steep mountain glaciers. Historical recession of these glaciers is difficult to quantify without sufficient high resolution terrain data, but in some notable instances illustrated above the Mueller Glacier, ice recession near and within recent detachment zones has been identified.

• Approximately 45 km² of steep (> 50°) bedrock slopes in the central Southern Alps are currently ice covered or above modelled permafrost limits, and therefore vulnerable to future weakening associated with ice recession and permafrost degradation.

IV. Regional-scale modelling of glacial-related geomorphic hazards

• Lakes formed behind steep moraines or outwash gravels are rare, and potential debris flows originating from catastrophic dam failure in these instances appear unlikely to reach any infrastructure. Modelled clear-water flood events from large volume proglacial lakes can potentially impact upon infrastructure located near braided river flood-plains both east and west of the Main Divide.
- Ice avalanches are unlikely to reach areas of permanent human settlement, although backcountry huts and walking tracks may be affected. Thirty-six glacial lakes are vulnerable to impacts from ice avalanches, assuming maximum probable runout distances are reached.

- Debris flows initiating from sediment within para- and peri-glacial zones are most likely to be problematic east of the Main Divide, where vehicle tracks and building infrastructure are located higher within glaciated catchments. Temporary river blockage within narrow gorges and debris remobilization may produce further reaching, secondary hazards west of the Main Divide.

- Numerical modelling with RAMMS provides capabilities for accurately representing rock avalanche dynamics, and can be used to estimate flow velocities and magnitudes, including mass volumes depositing into glacial lakes. Calibration parameters used to reconstruct past events can be applied to simulate future scenarios, where potential bedrock instabilities have been recognized.

- Most large volume proglacial lakes in the Aoraki/Mount Cook region are forming with their longitudinal axis perpendicular to adjacent mountain slopes, such that mass movement impacts will predominantly direct wave energy across these lakes. The Mueller Glacier has been identified as a unique example where a rapidly enlarging proglacial lake is forming directly beneath steep slopes of the Main Divide, with potential mass movements of ice or rock able to direct full wave energy towards the lake outlet.

4.2 Implications of the main findings

ASTER based methods for mapping glacial ice, lakes, and glacial-related debris accumulations have been explored, leading to the recognition of potentially steep and unstable terrain. The distinct altitudinal zonation of these landforms is significant, particularly in relation to permafrost limits which are unlikely to extend below 2000 m. Hence, the stability of lake dams and debris flow potential will mostly be unaffected by future permafrost degradation, a situation that is in contrast with expectations and observations from with other high mountain regions such as the European Alps. The potential role of permafrost degradation in the Aoraki/Mt Cook...
region therefore appears most important in relation to bedrock stability. An initial first-order estimate of permafrost distribution, coupled with subsequent modelling of bedrock temperatures has provided a basis for this relationship to be explored for the first time. Source area elevations on their own provide only a crude measure of possible permafrost conditions, with rock temperature measurements indicating thermal conditions can vary significantly up and down, and across a multifaceted detachment zone.

An incomplete prehistoric record of landslides and potential biases in the prehistoric landslide distribution resulting from high intensity triggers (most notably earthquakes), prevents conclusive, quantitative analyses regarding the recent influences of atmospheric warming and permafrost degradation. In any case, the recognition of marginal permafrost conditions within the detachment zone of many recent bedrock failures does not on its own imply that warming temperatures and permafrost degradation were a factor in these events, particularly given that extremely steep, pervasively fractured bedrock prevails throughout the central Southern Alps. However, cataloguing these events and associated topographic, geological, glacial and permafrost conditions, contributes towards the wider documentation of bedrock failures from glaciated environments around the world. In fact, this study has contributed the first large scale documentation of such failures and related permafrost conditions outside of the European Alps, providing an opportunity now to explore common predisposing factors within vastly different geological settings. The QMAP geological dataset provided an indispensable basis from which landslide analyses and detachment zone studies could proceed in the central Southern Alps. Given suitable aerial and/or satellite imagery, large scale slope failure mapping and monitoring could be implemented in many other remote mountain regions, with robust methods presented here providing direction for regional-scale modelling of permafrost conditions in the absence of high elevation climate data.

Compared to other high mountain regions within countries such as Peru or Switzerland, levels of infrastructural development and human settlement in the Aoraki/Mount Cook region are limited, and will continue to be strongly regulated by national park policies into the foreseeable future. As a consequence, the direct risk posed by glacial change related mass movements and floods in the central Southern Alps is much lower than these regions. Nonetheless, now that potential instabilities have been recognized in the Aoraki/Mount Cook region, monitoring future processes and process interactions can contribute towards improved scientific understanding in the broader
field of glacial hazard research. While the region has been seismically inactive over historical times, a large earthquake is overdue, and highly likely to occur over the coming century. In the event of such a high intensity mass movement trigger, widespread ice, debris, and bedrock failures can be expected, with potential impacts occurring into expanding glacial lakes. Methods for detecting and modelling such process interactions in the Aoraki/Mount Cook region could be most applicable within other tectonically active remote mountain regions, such as the Himalaya, where additive effects of seismicity, glacial recession and permafrost degradation may contribute to enhanced slope instabilities.

Owing to high mass turnover, and limited historical documentation of glacial hazards in the Aoraki/Mount Cook region, the local empirical basis for modelling mass movements of ice and debris is lacking. Snow avalanches and fluvial processes, for example, rapidly rework geomorphic evidence from para- or peri-glacial debris flows. Therefore, consideration of worst-case scenarios has been based upon empirical understanding established from other glacial regions, and in particular, the European Alps, where there has been a long history of awareness and scientific study regarding glacial hazards. When modelling potential debris flows in the vicinity of Aoraki/Mount Cook village, worst-case parameters clearly exceeded any threat to the village recognized from historical fan aggradation and expected runout distances based on available catchment size. Alternative, smaller runout distances modelled for debris flows and ice avalanches have therefore been included to qualitatively guide local authorities towards situations most likely to be affected by instabilities, while the results indicating worst-case scenarios represent a conservative approach to initial hazard assessment.

The tectonically active geological setting of the Southern Alps, together with extreme rainfall and high erosion rates near the Main Divide, contribute to enhanced slope instability and sediment delivery relative to most other mountainous regions. Therefore, mass movements and floods originating from glacierised (or recently glacierised) terrain form only one component of potentially hazardous processes occurring in the Aoraki/Mount Cook region, where failure of hillslopes, river damming, debris and clear water floods not directly linked to any glacial process or landforms remain the most recognized threat to infrastructure and human settlements. The most far-reaching events may result from the interplay of instabilities occurring in glacial and non-glacial environments. For example, a large bedrock failure occurring from glaciated steep slopes of the Main Divide may dam adjacent river valleys forming a temporary lake. Subsequent
heavy rainfall or earthquake triggered destabilisation of the lake dam could then initiate a catastrophic flood, impacting upon lowland populations. Although large bedrock failures have recently appeared most prevalent from steep east – southeast aspects of the Main Divide Fault Zone, the likelihood of catastrophic river damming is probably higher from events occurring west of the divide, where gorges are steep and narrow. Clearly, glacial hazards should not be considered in isolation, and modelling approaches that have been presented here, based on remote sensing and GIS techniques, are well suited for extending analyses over large distances and integrating potential secondary events.

4.3 Perspectives

Where infrastructure has been identified within potential flood or mass movement paths, local authorities are encouraged to proceed with a more detailed local-scale investigation of the hazard in these instances. Equally, any new or modified infrastructural developments in the region can now consider potential threats relating to glacial hazards. It is quite probable that small scale topographic features that were not captured or accounted for within the regional-scale approach to hazard modelling, may in fact divert or dissipate many potential flow paths. This is likely to be true for a majority of backcountry huts, which are often located atop or behind rock outcrops. Subsequent desktop studies using higher resolution imagery will quickly identify such situations, but consideration must also be given to potential barrier run-up or overtopping for larger magnitude events. At the local-scale, estimates of flow magnitudes, erosion capabilities, and deposition volumes are needed to establish the extent of any hazard, in combination with vulnerability and risk calculations. Higher level flood and mass movement models such as FLO-2D (O’Brien et al. 1993) or DAN/DAN3D (Hung et al. 1995) are available for local investigations, while RAMMS has been successfully introduced for rock avalanche modelling in the Aoraki/Mount Cook region.

Further study of glacial hazards in the Southern Alps of New Zealand will contribute significantly to international development in the field of glacial hazard research. Catastrophic glacial floods and glacial-related mass movements are recognized typically as low frequency, high magnitude events, and as such, empirical data is relatively scarce and mostly compiled from experiences in populated mountainous regions. A significant potential for seismic triggered failures, coupled with an expected sensitivity of perennial ice and permafrost to warming over
the coming century, suggests widespread instabilities are likely about the central Southern Alps, providing unparalleled opportunities to study failure mechanisms and process interactions within a remote mountain region. However, fundamental data are currently lacking and are needed to support future glacial hazard analyses in the Southern Alps, and more generally, to advance cryospheric studies in this region. Most importantly, a current, high resolution topographic dataset is required, because the current 25-m national DEM is almost 25 years out of date, and therefore cannot account for lowered glacial surfaces, altered terminal geometries, and significant moraine formations along lateral margins. Elevation data computed from ASTER stereo imagery, or the Shuttle Radar Topography Mission (SRTM) offer cost effective options for ongoing monitoring, but higher resolution aerial or satellite imagery is required to adequately monitor changing surface geometry on the steepest ice covered slopes. Airborne laser scanning has been applied in the European Alps to study mass movements from steep glaciated bedrock walls (Fischer and Huggel 2008), and similar airborne or possibly ground-based methods could be used for local-scale analyses of highly active slopes in the Aoraki/Mount Cook region, for example, above the Mueller Glacier. Improved topographic data should be coupled with a complete, satellite based inventory of New Zealand glaciers, so that future changes in ice extents can be documented with precision, and related influences on slope stability analysed.

Mountain permafrost and its role in bedrock slope failure is a relatively new and evolving field of research. Preliminary modelling of permafrost distribution in the central Southern Alps can now be significantly strengthened by establishing a larger dataset of near-surface rock wall temperatures. Specifically, measurements are currently lacking from the most super-shaded slopes, such that a current lower permafrost limit of 2000 m has been surmised. In addition, the broader significance of permafrost in landscape evolution in the Southern Alps is largely unexplored, and would need to consider past and present permafrost limits along the full length of the Alps. Modelling completed for the central region now provides local validation for any coarser, larger-scale permafrost distribution modelling in the Southern Alps.

In the current study, the difficulties in separating the potential influence of climate change and related permafrost degradation from other geological and topographic factors have been acknowledged. Both internationally, and locally, a wide gulf exists between expectations resulting from theoretical understanding regarding permafrost degradation and slope stability, and quantitative field based evidence. In the central Southern Alps, 13 recent bedrock failures
occurring from slopes likely to be affected by marginal permafrost have been identified, for which thermal conditions within the detachment zones should be further investigated. Geophysical techniques, including electrical resistivity tomography (Krautblatter and Hauck 2007; Krautblatter 2008) could allow direct monitoring of the spatial and temporal variability of permafrost characterizing these source areas and surrounding slopes. Such direct measurements would support 3-dimensional modelling of bedrock temperature, considering permafrost evolution within complex topography. Improved understanding of thermal conditions within the source area of recent slope failures will enable better certainty regarding future expectations of permafrost degradation.

Steep glacial ice and unstable debris accumulations are widespread about the Aoraki/Mount Cook region of New Zealand’s Southern Alps, glacial lakes are expanding, and steep fractured bedrock slopes await the next seismic rupture. Further glacial recession and weakening associated with permafrost degradation can only add to the likelihood of catastrophic mass movements, and potential lake impacts occurring over coming decades. Uncertainties concerning the scale and magnitude of future events must be met with modern investigative approaches for early hazard recognition.
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PART II

Research Publications
PAPER I
Satellite remote sensing procedures for glacial terrain analyses and hazard assessment in the Aoraki Mount Cook region, New Zealand


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Abstract

Hazards originating within glacial environments have received limited attention in New Zealand. In view of rapidly changing environmental conditions and the need for ongoing monitoring, overseas studies have recognised the importance of remote sensing procedures to support related hazard investigations. The initial goal for regional hazard studies is the identification of potential source areas on the basis of terrain analyses. Following a review of remotely sensed imagery and its application in glacial studies, several methods are applied here to map glacial ice, lakes, and debris accumulations in the Mount Cook region of New Zealand using freely available ASTER imagery. A new method for mapping turbid and clear water glacial lakes is presented, and the first application of a method for mapping debris accumulations at the regional scale appears promising for debris flow hazard investigations. Analyses of terrain changes reveal that a 20% expansion in lake area has continued over recent years, increasing the hazard potential from mass movement induced outburst flooding. A framework for the assessment of hazard potential in the region is introduced, requiring the integration of terrain analyses presented here with permafrost distribution modelling, geological mapping, and topographic analyses.

Keywords glacial hazard; glacial change; remote sensing; terrain analyses; ASTER; New Zealand

INTRODUCTION

Satellite remote sensing techniques have previously been used to analyse a range of geophysical hazards in New Zealand, such as landslides (Trotter et al. 1989; Dymond et al. 2006), flood hazards (Fuller 2005), coastal hazards (Bell & Gorman 2003), severe storm events (Leslie et al. 2005), and volcanic hazards (Oppenheimer 1996; White & Hockey 1996). Similar studies have not been undertaken for glacial hazards in New Zealand despite the significant glacial changes that have been observed (Chinn 1996, 1999), and the field-based identification of related hazards, including ice avalanches (Iseli 1991), debris instabilities (Blair 1994), rock avalanches, and their interaction with glacial lake outburst flooding (McSaveney 2002). Overseas remote sensing based studies of glacial hazards have been successfully undertaken for many populated mountain regions of the world (e.g., Europe, Himalayas, Andes, North America) as shown by Kääb et al. (2005b) and Quincey et al. (2005). These studies have shown that for first-order regional hazard assessments, an initial analytical step should investigate possibilities to distinguish potential hazard sources from mapped glacial terrain surfaces such
This paper aims to investigate existing and new satellite remote sensing based techniques for glacial terrain analyses in the Mount Cook region of New Zealand. The intention here is to review and introduce a range of optically based terrain classification procedures that may be used to improve glacial-related investigations in the region. Classification accuracy and limitations are presented, additional information needed to support glacial hazard investigations is discussed, and an integrated procedure for glacial hazard assessment in the region is proposed.
A First Estimate of Mountain Permafrost Distribution in the Mount Cook Region of New Zealand’s Southern Alps

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Abstract

The heavily glaciated Mount Cook Region of New Zealand has experienced several recent large rock instabilities, but permafrost conditions related to these events remain unknown. This work presents the first systematic approach for investigating the distribution of mountain permafrost in New Zealand. At this level of the investigation, a first-order estimate is based upon the adaptation of established topo-climatic relationships from the European Alps. In the south east of the study region, the permafrost estimate gives a reasonable correspondence with mapped rock glacier distribution but the maximum elevation of vegetation growth is situated 200 m beneath the lower limit of estimated permafrost. Extreme climate gradients exist and towards the humid northwest where rock glaciers are absent and vegetation patterns give an unclear climate signal, large uncertainties remain. Data currently being recorded from a network of rock wall temperature measurements will help remove this uncertainty, and will allow distribution modeling that better accounts for the topographic complexities of this steep alpine region.

Keywords: permafrost distribution; rock instabilities; spatial modeling; Southern Alps, New Zealand

Introduction

Internationally, mountain permafrost is a well recognized phenomenon in relatively low-angled debris-covered terrain where active rock glaciers produce distinctive landforms that indicate perennally frozen ground beneath. Previous research based upon the identification of active rock glaciers in the Southern Alps of New Zealand has suggested that permafrost probably occurs only sporadically within a very narrow altitudinal zone in the more arid areas of the Alps (Brazier et al. 1998). However, there has been no scientific consideration in New Zealand given to the likely wider distribution of permafrost within bedrock slopes, and in particular, on steep slopes which dominate at higher elevations throughout the Mount Cook Region (MCR). Recent large rock avalanches, including the spectacular summit failure of Mount Cook (McSaveney 2002) have awakened interest in the understanding of permafrost and slope stability interactions in the region. This paper aims to present the first results from permafrost distribution modeling for the MCR, based upon local application and calibration of topo-climatic relationships established in the European Alps. Initial validation of the estimated permafrost distribution is discussed on the basis of a rock glacier inventory and remote sensing-based vegetation mapping, and future research towards improved modeling in this region is introduced.

Background

Studies of permafrost processes in New Zealand are limited (Soons & Price 1990), with only a few detailed studies emerging from the identification and dating of rock glaciers within the Southern Alps. These studies have served to improve the chronology of Holocene glacial activity in the region (McGregor 1967, Birkeland 1982) and offer some insights regarding possible climate sensitivities of rock glaciers (Jeanneret 1975, Kirkbride & Brazier 1995). From the mapping and classification of periglacial landforms in the Ben Ohau Range, Brazier et al. (1998) suggested that permafrost distribution was more limited than would be expected based upon climatic boundaries identified in the European Alps (Haeberli 1985). In the driest, generally low-elevation mountains east of the Southern Alps, rock glaciers are absent and ice-free talus slopes dominate. In the more humid, maritime regions to the west where precipitation exceeds 10000 mm yr⁻¹, glacier equilibrium line altitudes (ELAs) are low and temperate mountain glaciers dominate, with no active permafrost landforms evident (Augustinus 2002). Between these two extremes, within a narrow zone where precipitation does not exceed 1500 mm yr⁻¹ ELAs are higher, glacial ice is limited to debris-covered cirques, and more numerous permafrost landforms are found (Brazier et al. 1998).

Understanding permafrost distribution within complex steep mountain terrain is a relatively new field of scientific research largely stemming from European based studies in relation to climate warming, permafrost degradation and related slope instability hazards (Gruber & Haeberli 2007). During the past 50 years, several large rock avalanche events have occurred within the MCR (McSaveney 2002), but in the complete absence of data regarding permafrost distribution in the steep terrain of the Southern Alps, the possible role of permafrost weakening within past or future detachment zones is uncertain.

In complex mountain terrain, variations in topography and related permafrost factors such as snow cover result in the necessary use of spatial modeling to achieve any realistic estimate of permafrost distribution (Etselmüller et al. 2001). A hierarchy of modeling procedures has therefore been developed over recent years (Hoelzle et al. 2001). Readily
applied empirical-statistical approaches relate documented permafrost occurrences to easily measured topo-climatic factors such as altitude, slope and aspect, air temperature, and solar radiation (e.g. Imhof 1996) and are particularly well suited for preliminary assessment and planning in relation to geotechnical hazards (Harris et al. 2001). A useful, and essential first step in these studies has been the adaptation of Haebeli’s (1975) original topo-climatic key for predicting permafrost occurrence, which provides an immediate impression of likely permafrost distribution, and a basis from which local validation and more advanced modeling can proceed.

**Study Region**

The MCR is broadly defined here to encompass the 700 km² Aoraki Mount Cook National Park, extending west of the Main Divide into the Westland National Park and south to include the Ben Ohau Range (Fig. 1A). The region includes the highest mountains and the most heavily glacierised terrain of New Zealand’s Southern Alps. Permanent snow covered peaks are found between 2500 and 3754 m.a.s.l, with local relief in the order of 1000-2700 m. Moist westerly airflow perpendicular to the Main Divide, generates very high orographic rainfall amounts and creates an extreme precipitation gradient leeward of the Alps (Griffiths & McSaveney 1983). Glacial retreat since the Little Ice Age maximum has been most rapid during the mid 20th Century, leading to a 25% loss of total ice area in the Southern Alps during this past century, although some highly responsive glaciers have had notable periods of advancement over recent years (Chinn 1996).

**Estimating Permafrost Distribution**

Based on extensive geophysical and morphological investigations of rock glacier phenomena in the eastern Swiss Alps during the 1970s, Haebeli (1975) developed ‘rules of thumb’ for predicting permafrost occurrence. Stemming from these rules was the empirical topo-climatic key, incorporating the primary physical factors which influence permafrost distribution, determining both a zone of probable permafrost and the permafrost limit within a transitional zone termed ‘possible permafrost’. These physical factors include significant aspect dependent radiation effects, altitudinal changes in air temperature, and topographically related snow cover variation (Etzelmüller et al. 2001). In relation to snow cover, gentle terrain situated at the foot of steeper slopes where long-lasting avalanche snow may accumulate, can maintain cooler ground surface temperatures than steeper slopes of the same aspect. On flat terrain, the local permafrost distribution is determined more by air temperature and snow cover than by radiation. Calibration of the key to local conditions in the MCR was based upon calculation of mean annual air temperature (MAAT) for the years 1982 – 2007 and the local 0°C isotherm elevation using daily temperature data from three climate stations located across the region (Fig 1A). For the eight primary slope aspects, the original elevation zones given in the topo-climatic key were either raised or lowered based upon the difference between the local elevation of the 0°C isotherm and the equivalent value from the Swiss Alps where the key was established. Topographic values for all analyses were extracted from the Landcare Research 25 m resolution South Island digital terrain model (DTM).

![GIS-based modeling of possible steep permafrost distribution](image)

**Figure 1. A)** GIS-based modeling of possible steep permafrost distribution (black) in the MCR based upon adjustment of the original topo-climatic key (Haebeli 1975) using MAAT calculated from the Franz Josef (FJ), Mount Cook Village (MCV) and Lake Tekapo (LT) climate stations. Also shown are the locations of Mt Cook (MC, 3754 m) and Mt Sefton (MS, 3151 m). **B** Closer view of the estimated permafrost terrain around the summit area of Mount Sefton, C) repeated for a 0°C isotherm rise of 200 m, and D) a 0°C isotherm lowering of 200 m. The base image used is an ASTER satellite mosaic from 2007.

In an attempt to account for the significant climate gradients associated with the föhn effect which results from initially moist airflow across the Main Divide, the 0°C isotherm was calculated independently using MAAT measured at each of the three climate stations. Hydrological studies in the region have shown that rainfall gradients are approximately parallel to the Main Divide, with maximum rainfall measurements associated with the steepened terrain along the alpine fault (e.g. Griffiths & McSaveney 1983). Because of the effects of decreasing moisture towards the southeast on environmental lapse rates, the following values were used: Franz Josef, 0.005 °C m⁻¹ (Anderson 2003), Mount Cook Village, 0.0065 °C m⁻¹, and Lake Tekapo, 0.0075 °C m⁻¹. The latter two rates have not been directly measured, but were inferred from the nearest available measured locations where precipitation and humidity values are comparable (Brazier et al. 1998). Using a GIS procedure, the 0°C isotherm elevations were then interpolated between the three climate stations in a southeast direction based upon distance from the alpine fault (Fig 1A). This modeling of 0°C isotherm elevations is a crude simplification, but for the purposes of an initial permafrost estimate it is able to provide
an approximation for the influence of moisture gradients prevailing across this region of the Alps.

Table 1 gives the adjusted permafrost elevation limits based upon the 0°C isotherm elevation of 2114 m calculated at the Mount Cook Village. This establishes a lower limit of permafrost in steep terrain (on slopes > 20°) of 2880 m on sunny northern aspects, decreasing to 2230 m on shaded southern aspects. At Franz Josef, where a strong maritime climate prevails, the 0°C isotherm could be positioned up to 270 m higher, with a corresponding rise of the permafrost limits by the same amount, whereas these limits may be 190 m lower towards the drier climate at Lake Tekapo. At the foot of slopes (≤ 20°), permafrost may be possible up to 750 m lower than on corresponding steeper slopes, but is probable at elevations only 400 m lower.

<table>
<thead>
<tr>
<th>Aspect</th>
<th>Permafrost possible</th>
<th>Permafrost probable</th>
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<tbody>
<tr>
<td>Steep</td>
<td>Foot of slopes (S)</td>
<td>Foot of slopes (S)</td>
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<tr>
<td>SE</td>
<td>2280</td>
<td>2180</td>
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<td>NE</td>
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<td>NW</td>
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<td>SW</td>
<td>2230</td>
<td>1930</td>
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<th>Flat Areas</th>
<th>Wind-exposed</th>
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<tr>
<td></td>
<td>2480</td>
<td>2530</td>
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<tr>
<td>Variability</td>
<td>Franz Josef (0.005 °C m⁻¹)</td>
<td>270 m higher</td>
</tr>
<tr>
<td></td>
<td>Lake Tekapo (0.007 °C m⁻¹)</td>
<td>190 m lower</td>
</tr>
</tbody>
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All elevations are given in metres above sea level (m). Values are based upon MAAT at Mount Cook Village (765 m) and an air temperature lapse rate of 0.0065°C m⁻¹. Variability is calculated using MAAT at Franz Josef (155 m) and Lake Tekapo climate stations (762 m) with given lapse rates.

At elevations where possible steep permafrost is predicted along the Main Divide glacial ice dominates much of the terrain with only limited exposed bedrock around ridge tops, on steep faces and rock outcrops (Fig 1A). In this region of the Alps, the average ELA is around 2000 m, well below the lower boundary of permafrost such that temperate glacial ice dominates, but areas of polythermal ice and associated permafrost interactions are likely within the higher elevation cliff and hanging glaciers (Etlzmüller & Hagen 2005). In the drier southeast, glacial growth is restricted with both talus and bedrock surfaces featuring prominently at elevations within the estimated permafrost terrain.

The accuracy of the selected lapse rates, MAAT, and resulting 0 °C isotherm calculations will have a significant influence on the permafrost distribution estimate. This sensitivity is well illustrated for the area of the Main Divide around Mount Sefton (Figs. 1B-D). The current model estimates widespread permafrost surrounding the summit pyramid and along the ridges to the west and northeast. However, a 200 m lowering of the isotherm due to a colder MAAT or greater lapse rate selection would significantly increase the estimated possible permafrost distribution along all surrounding ridgelines and on all slope aspects. Under the scenario of a warmer MAAT or lower selected lapse rate resulting in a 200 m rise of the isotherm, the estimated permafrost terrain becomes mostly limited to the shaded aspects high on the summit area of Mount Sefton. While illustrating the sensitivity of the topo-climatic key to local calibration, this also gives some indication of the effect future climate warming could have on permafrost distribution in the region.

**Local Validation Using Rock Glacier Inventory**

Active and fossil periglacial landforms have been previously mapped in the Ben Ohau Range using a threefold classification of debris-covered glaciers, cirque-floor lobe forms, and talus rock glaciers (Brazier et al. 1998, Appendix 1). Although the Ben Ohau Range comprises only a small area of the much larger study region, it does contain the only known active permafrost forms in the region, and therefore is able to provide some initial local-scale validation of the permafrost distribution estimate. The original mapped landform data were transferred into a GIS inventory of active and fossil permafrost forms, with some positions and measurements reassessed using high resolution (0.61 m) QuickBird satellite imagery in combination with a DTM. A total of 70 permafrost forms were mapped according to their Rock Glacier Initiation Line Altitude (RGILA) position as measured from the foot of the talus. Active debris-covered glaciers are not necessarily indicative of permafrost conditions, but are included here to further illustrate the distinct landform zonation that occurs along this range (Brazier et al. 1998).

The active permafrost forms predominate on shaded aspects within a narrow 8.2 km north-south zone towards the centre of the Ben Ohau Range where the highest peaks are just above 2400 m (Fig. 2A). The RGILA of all active landforms would therefore be expected to lie mid-way between the permafrost limits estimated at the southern and northern ends of the range, but instead are positioned at altitudes 20 to 180 m lower (Fig. 2B). Permafrost originating at the foot of slopes might account for the preservation of some of these forms at lower elevations, but the possibility that the model estimate is too high within this area of the study region must also be considered. Although data are unavailable for northerly aspects, the aspect-related variability of the model from east through to west appears to match the distribution of active rock glaciers. No fossil permafrost forms are located within the estimated permafrost terrain and most are positioned more than 200 m below the estimated boundary, which is greater than can be expected from 20th Century temperature warming in this region (Salinger 1979). The numerous fossil forms mapped at the southern end of the range are located at much lower altitudes, with surface dating suggesting that many of these forms are periglacial relics from the late Pleistocene (Birkeland 1982). Towards the northern end of the range, increased snowfall combined with topographic effects enables the growth of
heavily debris covered cirque glaciers at higher elevations and neither active nor fossil permafrost forms are observed here.

Figure 2. A) Modeled permafrost distribution on steep slopes (black) along the Ben Ohau Range compared with an inventory of active (triangle) and fossil (circle) rock glacier forms, and debris covered glaciers (square). Permafrost modeled at the foot of slopes is also shown (grey), based on slope curvature analyses. B) The spatial distribution of rock glacier forms is compared to steep permafrost lower elevation limits at the northern (grey) and southern (black) ends of the Ben Ohau Range.

Comparison with Vegetation Mapping

Modern remote sensing-based mapping techniques are able to provide a crude, indirect indication of likely permafrost distribution (Etzelmüller et al. 2001). The presence or absence of alpine vegetation is a particularly well known indicator of permafrost distribution in mid-latitude mountains (Haebler 1975), and the inclusion of vegetation abundance mapping from satellite imagery has proven a useful parameter for improved distribution modeling (Gruber & Hoelzle 2001). In the current study, orthorectified 15 m resolution Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) imagery from 2007 was used to create a map of vegetation distribution for the entire study region. To achieve this, the Normalised Difference Vegetation Index (NDVI) was used, which is based upon the contrasting spectral response of healthy vegetation between the red and near infrared (NIR) wavelengths. Combined with a DTM, altitudinal patterns in vegetation distribution were analysed and compared with the estimated permafrost distribution. A comparison is made between the maximum altitude of alpine vegetation growth (MAV) and the estimated lower limit of steep permafrost distribution within every 1 km zone in a northwest to southeast direction across the Southern Alps, parallel to the climatic gradient (Fig. 3). To examine the influence of slope aspect on vegetation patterns and any relationship there might be to permafrost distribution patterns, analyses are included for both sunny northern aspects and shaded southwest aspects. A slope curvature threshold was incorporated to ensure that only pixels with a small change in aspect were included, minimizing the risk of erroneous measurements around sharp terrain features.

In the drier mountains east of the Main Divide such as the Liebig and Ben Ohau Ranges, the MAV on steep southwest aspects is positioned at around 2000 m, and is consistently 200 m lower than the estimated lower boundary of possible permafrost for these slopes. Closer towards the Main Divide, this difference increases to over 400 m as the rise in permafrost limits is not matched by any rise in MAV. In fact, the fluctuating MAV decreases closer to the Main Divide possibly as a function of large maritime snowfalls combined with the effects of geomorphic and glacial disturbances. On northern slope aspects the MAV averages 150 m higher than on southwest aspects, with some greater differences measured nearer to the Main Divide, but never coming close to reflecting the much greater differences that are expected between permafrost limits on these contrasting slope aspects. The lack of any strong MAV pattern across the study region suggests that the usefulness of vegetation as an indicator of likely permafrost distribution is very limited in the Southern Alps of New Zealand. While the presence of vegetation confirms the absence of permafrost, the absence of vegetation offers very little conclusive information as to the distribution of permafrost, particularly nearer towards the Main Divide.
Discussion

Application of topo-climatic relationships established within the Swiss Alps to a more maritime alpine region characterized by extreme climate gradients produces many challenges. MAAT and lapse-rate calibration parameters used here are based on low-elevation climate stations located some distance from the Main Divide, and with the absence of additional data, climate gradients across the highest terrain cannot be modeled with certainty. A current study is suggesting that maximum precipitation might exist very close to, and even leeward to the Main Divide (Kerr et al. 2007), which would likely raise the estimated permafrost limits near to this region. In addition to the effect that climate gradients will have on local MAAT, associated differences in factors such as cloudiness and precipitation will also influence the amount of variation within the topo-climatic key from northern to southern aspects. Heavy snowfall and maritime cloud cover near the Main Divide will influence solar radiation patterns at the ground surface, which largely determine the aspect variation within the key. Therefore, the assumption of uniform variation between slope aspects across the study region must be reconsidered in a more advanced approach to modeling permafrost distribution.

Comparison of modeled permafrost distribution with rock glacier inventories is a well established approach (e.g. Imhof 1996), but was restricted in the current study by the limited spatial distribution of these landforms. Modern earth imagery such as QuickBird has significantly improved the ability to recognize and map permafrost features, but some subjectivity and potential for error remains in determining active from inactive landforms and defining parameters such as the RGILA. Vegetation patterns in relation to climate gradients and permafrost distribution were explored here on the fundamental basis of being absent or present, but given the ecological diversity that exists across the Southern Alps, more useful patterns might be observed within individual species or by using the NDVI to explore topographic patterns in plant biomass (Gruber & Hoelzle 2001).

Recent large rock failures in the European Alps have suggested that permafrost degradation is a serious concern in relation to climate warming and slope instability (Gruber et al. 2004). In the Southern Alps of New Zealand, the summit collapse of Mount Cook in 1991 detached from a maximum elevation of 3720 m on an eastern exposure (McSaveney 2002). This is well above the estimated lower boundaries of possible permafrost where 20th Century thawing may be expected. However, the large rock buttress involved in the failure extended down to a much lower elevation, and in complex, steep topography, the effects of three-dimensional thermal gradients must also be considered because of contrasting temperatures between colder and much warmer sunny slope expositions of the terrain (Noetzli et al. 2007). Furthermore, bedrock temperatures are complicated by the presence of steep ice bodies because heat exchanges associated with surface melting can induce significant thermal anomalies within the underlying bedrock (Huggel et al. 2008). Of the other large rock avalanche events occurring in the region over the past 50 years, three originated from within the estimated permafrost terrain, while another was initiated from an elevation approximately 140 m below the estimated lower permafrost boundary. In addition, two recent fatal rockfall events near the summit of Mount Cook and Mount Sealy further south, have also originated from within the estimated permafrost terrain.

On the basis of this initial estimate of permafrost distribution and the uncertainties it has raised, a field campaign was initiated during November 2007 measuring the spatial variation of rock surface temperatures on steep bedrock slopes over a 12 month period. Following the methodologies developed by Gruber et al. (2003), 15 miniature temperature data logs were installed at elevations ranging from 2400 m to 3150 m on various slope aspects both immediately on the Main Divide, and on the drier Liebig Range towards the southeast. In addition, a series of high elevation air temperature loggers have been installed across the region to better establish relationships with the low elevation, long term records from the Mount Cook Village climate station. Information regarding the spatial distribution of rock surface temperatures will be used to more accurately model and validate permafrost distribution across the region.

Conclusions

A first-order approach for permafrost distribution mapping in the Mount Cook Region of New Zealand’s Southern Alps has been presented. Towards the southeast of the region a drier, more continental climate prevails and the permafrost estimate gives reasonable correspondence with the limited distribution of active and fossil rock glaciers. Closer towards the Main Divide where conditions are more humid, the lower boundary of permafrost distribution is expected to be significantly higher, but rock glaciers are absent here preventing any potential for local validation. Similarly, modeled vegetation altitudinal limits across the region showed no relationship with the climatic gradients which are expected to influence permafrost distribution. Future modeling will incorporate rock wall temperature data to facilitate permafrost modeling which better accounts for the effects of solar radiation and topographic shading in this complex, steep mountain environment.

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References


PAPER III
Rock avalanches and other landslides in the central Southern Alps of New Zealand: A regional-scale study of distribution patterns and potential climate change impacts

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Abstract

Slope instabilities in the central Southern Alps, New Zealand, are assessed with reference to possible influences of glacier recession and permafrost degradation. 510 landslides have been identified and their geological and topographic characteristics analysed within a geographic information system (GIS) inventory. Landslides affect 2% of the study area, but their distribution is non-uniform across the alps: rock avalanches are most common from steep (>50°) greywacke slopes about the Main Divide of the alps (1/72 km²), and appear the principal large failure mechanism above 2500 m; other landslides occur most frequently in the schist terrain closer to the Alpine Fault (1/25 km²). For all landslides, the greatest proportion of failures has occurred from slopes facing west – northwest. 19 rock avalanches and large rockfalls were observed during the past 100 years: 18 occurred within 300 vertical metres above or below glacial ice; 12 have source areas within 300 vertical metres of the estimated lower permafrost boundary, where warm degrading conditions are expected. Recent events above Mueller Glacier include retreating ice near several source areas, and degrading permafrost possibly characterises the Vampire Peak failure zones from 2003 and 2008. Effects of climate-change may be ephemeral and difficult to ascertain, however, in comparison to the longer term influences of uplift, erosion and seismicity along an active plate margin.

Keywords: Landslide inventory, rock avalanche, glacial change, permafrost, Southern Alps, New Zealand.
1. Introduction

Landslides are a major process in hillslope evolution (Densmore and Hovius 2000; Korup et al. 2005), impact upon rivers (Korup 2002; Korup 2005a) and glacial systems (Hewitt et al. 2008; Shulmeister et al. 2009), and are a serious hazard in many regions of the world (Nadim et al. 2006). In glaciated alpine terrain, landslides often involve large volumes and travel large distances, owing to high local relief, enhanced travel over ice or snow surfaces, flow transformations and chain-reaction events (eg, Evans and Clague 1988; Haeberli et al. 2004). In addition, the hazard associated with slope instabilities in glacial environments is highly dynamic, because recent atmospheric warming is rapidly altering glacial landscapes, potentially shifting zones of instability and initiation (Kääb et al. 2005). Landslides and other gravitational mass wasting play a significant role in the erosion of New Zealand’s Southern Alps (Hovius et al. 1997; Korup et al. 2004). This mountain chain is being actively formed by the convergence of two tectonic plates. Stress-release fracturing and dilation of mountain rock masses occurs during uplift and exhumation. Over time, effects such as rock-mass dilation, ice wedging, glacier and snow level recession, and the reduction in ice-binding of rock masses combine to lower their resistance to failure (Wegmann et al. 1998; Davies et al. 2001; Gruber and Haeberli 2007). Major earthquakes and storms have triggered failures (Whitehouse and Griffiths 1983; Hancox et al. 2003), but numerous spontaneous events have also occurred (McSaveney 2002; Cox and Allen 2009). Recent completion of a digital geological map and geographic information system (GIS) dataset (Cox and Barrell 2007) creates an opportunity to better study landslide distribution in the Southern Alps, and to discuss potential impacts of glacier recession and permafrost warming in the region. The dataset provides an inventory of over 500 landslides in the central South Island, for which related geology and topography can be explored in a GIS environment. This study begins with a description of the physical setting, before providing an overview of the regional topography characterizing the central Southern Alps. The main objective of the paper is to examine the distribution of extant landslides in relation to topography and local geology. This provides background for a discussion regarding possible influences of recent and future glacial and permafrost terrain changes on slope stability in this seismically active plate collision zone.
2. Background

Slope stability is primarily governed by the geological and related geomechanical setting of the landscape, but in high-mountain environments, geological hazards are increasingly being studied in relation to possible interactions with changing glacial and permafrost conditions (eg, Evans and Clague 1988; Harris 2005; Deline 2008; Huggel 2008). Bedrock walls in glacial environments are typically steep, with erosion of their lower flanks exacerbated by glacial plucking. Subsequent retreat of glacial ice can induce changes in the stress field of the surrounding rock walls and expose previously insulated surfaces to altered mechanical and thermal erosion (Haeberli et al. 1997; Wegmann et al. 1998). The influence of permafrost degradation within steep rock walls is a relatively new field of research stemming from initial theoretical discussions linking atmospheric warming, glacier recession, permafrost degradation, and slope instability (Haeberli et al. 1997; Harris and Vonder Mühll 2001). Laboratory studies have since demonstrated that the shear strength of an ice bonded rock discontinuity significantly reduces with warming, revealing a minimum factor of safety at temperatures between -1.5 and 0 °C, where a discontinuity may be less stable than when in a completely thawed state (Davies et al. 2001). Other factors including ice segregation and volume expansion are likely to predispose a rock discontinuity to subsequent failure upon warming (Gruber and Haeberli 2007), and meltwater or ground-water flow into a previously frozen fracture may cause elevated water pressure and reduced frictional strength (Harris 2005). An absence of snow or debris cover on steep slopes means that sub-surface temperatures and related bedrock stability can respond rapidly (within one season to several years) to atmospheric warming. This was demonstrated during the extremely warm European summer of 2003, when numerous rockfalls were likely associated with extreme heat fluxes in the active layer of the permafrost, with massive ice observed in some detachment zones (Gruber et al. 2004). Deeper warming of alpine permafrost is delayed in the order of decades to centuries, but warming of 0.5 to 0.8 °C in the upper decametres has been observed over the past century in Europe (Harris et al. 2003), and has been linked to rock avalanches from deeper detachment surfaces occurring primarily in estimated areas of warm or degrading permafrost (>1.5 °C) (Dramis et al. 1995; Bottino et al. 2002; Noetzli et al. 2003). Within landslide-inventory studies, slope-failure susceptibility has been explored on the basis of individual or combined causative factors such as slope, lithology, geological structure, slope morphology, and anthropogenic activities (eg, Hermanns and Strecker 1999; Pande et al. 2002; Pike et al. 2003). Relating landslide susceptibility to permafrost and glacial factors is faced with greater
uncertainty because these factors are often difficult to observe or measure and change rapidly over time, but some important linkages have been established from a small inventory of recent landslides in the European Alps (Noetzli et al. 2003), and further insights have been gained from detailed case studies (eg, Gruber et al. 2004; Fischer et al. 2006; Huggel 2008).

Rock avalanches and other landslides in the Southern Alps have been investigated in a number of studies, but few are regional scale. Studies have examined slope failures associated with earthquakes (eg, Speight 1933; Adams 1981; Bull and Brandon 1998; Orwin 1998), or affecting road construction (Paterson 1996). Other detailed works describe individual rock avalanches initiated from high elevation bedrock slopes about the Main Divide during the past twenty years, including the spectacular summit collapse of New Zealand’s highest mountain, Aoraki/Mount Cook (McSaveney 2002; Hancox et al. 2005; Cox and Allen 2009). The first inventory-based study defining long-term occurrence probability was carried out by Whitehouse (1983) using mostly prehistoric rock avalanches. The average occurrence rate of >10⁶ m³ rock avalanches is estimated at 1 per 100 years across the entire Southern Alps (Whitehouse and Griffiths 1983), but a rate of 1 per 20 to 30 years is probably more characteristic from shattered rocks and precipitous slopes near the Main Divide of the alps (McSaveney 2002). More recently, inventories have been developed for landslides in South Westland and Fiordland (Hovius et al. 1997; Hancox et al. 2003; Korup 2005b), and used to examine landslide-related sediment flux. This paper serves to complement existing inventory-based research, within a GIS approach that encompasses the entire central Southern Alps, but more specifically offers the first discussion regarding glacial ice recession, permafrost warming, and potential impacts on landsliding in this mountainous region of New Zealand.

3. Methodology

The QMAP (Quarter-million-scale Map) digital geological map of New Zealand is currently being produced by the Institute of Geological and Nuclear Sciences (GNS Science) to supersede a previous 1:250,000 national geological map series published during the 1960s (Rattenbury et al. 1994). The QMAP project started in 1994 and now includes the recently completed Aoraki sheet describing the central South Island (Cox and Barrell 2007). QMAP is based on geological information plotted on 1:50,000 topographic base maps, compiled from previous studies and new field work. Data are simplified for digitising at the compilation stage, with line work smoothed and
geological units amalgamated to a standard national system, enabling production of 1:250,000 hard-copy maps using ArcInfo®. Published sheets are available as a series of ArcGIS coverages and shapefiles, including data on landslides, geological units and structure. QMAP’s research applications are best suited for the regional level, because of the scales at which data were captured and simplified. Landslides are defined within QMAP as both landslide deposits (within the geological unit layer), and as landslide-affected areas (polygons and arcs in a specific landslide layer). Many landslides have also been identified in earlier inventories (eg, Whitehouse 1983; Korup 2005b), but were in all instances remapped in the field, and/or re-examined from aerial photography to achieve consistency in the QMAP data. Landslides occur on all scales, but due to the map simplification process only those larger than 0.02 km² are digitised, with boundaries smoothed and digitised to an accuracy of ±100 m. In practice, this results in variable uncertainty in landslide area: approximately ±25% for small landslides (<0.1 km²), ±10% for large landslides (1 – 10 km²), and ±1% for very large landslides (>10 km²). Types of landslide include: shallow translational slides and rotational slumps on lowland hillslopes; deeper-seated earthflows, slides or slumps within Late Cretaceous-Cenozoic sedimentary rocks; large rock block-slides and landslide complexes in mountain bedrock; large rock/debris falls from bluffs; and rock avalanches from steep mountain slopes. Rock avalanches are differentiated in the dataset and their deposits comprise unsorted bouldery, silty, sandy debris. Identification of large, deep-seated landslide complexes is easiest where obvious scarp development and slumping has occurred, which is particularly evident where moraine terraces or other cover sequences have been displaced. However, where this evidence is not obvious, potentially very large landslide complexes which have not developed into a rock avalanche may be missing from the inventory. Smaller, shallow slope failures such as debris flows are clearly evident in forested regions, but become more difficult to recognize in alpine regions, and therefore, prehistoric events of this type may also be underrepresented in the inventory.

The QMAP landslide dataset forms the basis for this study, with the addition of several new published (Cox et al. 2008) and unpublished events identified since compilation of Aoraki QMAP. In total, 510 landslides were mapped, 405 of which have no specific failure mechanism defined, and 105 of which were sub-classified as rock avalanches. This represents a minimum catalogue of events in the region, particularly for slope failures occurring within glaciated terrain or depositing into highly erosive watersheds where geomorphic evidence can rapidly become unrecognizable (Hewitt et al. 2008). For the nonspecific landslides, failure scarp digitized as arcs within the QMAP dataset
were enclosed to form polygons approximating the landslide source areas. Rock avalanche source areas were treated to a higher level of discrimination, given that detachment zones were often small and poorly defined by QMAP landslide scarps. Therefore, air photo and satellite image interpretation, topographic maps and shaded relief images were used to directly map the source areas for all 105 rock avalanches, verified wherever possible by field observations and with reference to published studies and reports. Source areas were digitised directly on the shaded relief imagery generated from the NZ 25-m grid cell digital elevation model (DEM). Created from 1986 aerial photography, and assessed to have a root mean square error of 5 – 8 m for hilly, steep terrain (Barringer et al. 2002), this DEM provided the basis for all topographic analyses. All complete landslide affected areas and related source areas for the 405 nonspecific landslides and 105 rock avalanches were linked by a landslide identification code, and a combination of zonal statistics and spatial queries used to derive related topographic and geological information for the inventory. Alluvial deposits and flat (0°) slopes which predominate within valley floors comprise over 40% of the study region, and were excluded from all analyses to provide more meaningful results in the context of a slope instability study. The term hillslope is used herein for these non-alluvium areas with slopes >0°.
Figure 1. Simplified geological classification of the central Southern Alps developed from the Aoraki QMAP geological units, showing nonspecific landslide and rock avalanche distribution (after Cox and Barrell 2007), and three main geomorphic domains described by Whitehouse (1988). Northern (NC) and southern (SC) transects across the alps relate to Figure 2, and also indicated are the locations shown in Figure 9 (Mueller Glacier) and Figure 12 (Hooker Glacier). Landslide magnitude describes the total affected area (source and deposit areas combined). Inset shows the position of the Australian-Pacific plate boundary.
4. Physical setting

This study covers 21,300 km² of the central Southern Alps, centred upon the Aoraki/Mt Cook National Park, but also extending west of the Main Divide of the Southern Alps towards the Alpine Fault, north to the headwaters of Rakaia River, and east to include the lower elevation foothills bordering the Canterbury Plains and MacKenzie District (Fig. 1). It encompasses the highest and most heavily glaciated terrain of the Southern Alps while traversing different climatic regimes from superhumid, maritime conditions in the west (precipitation >12 m y⁻¹), to a drier, more continental climate towards the east (precipitation <2 m y⁻¹) (Griffiths and McSaveney 1983).

The Alpine Fault is a major active fault traversing the northwestern edge of the study area, on which most of the ongoing surface tectonic movement between the Australian Plate (to the northwest) and the Pacific Plate (to the southeast) is concentrated (Fig. 1). Neogene displacement of up to 470 km along the Alpine Fault has brought together two different pre-Cretaceous geological provinces. Northwest of the Alpine Fault, there are small amounts of exposed Paleozoic metasedimentary and plutonic basement rocks that are fragments of the Gondwanaland supercontinent. Southeast of the Alpine Fault, basement rocks belong to the Torlesse composite terrane. They comprise thick, deformed packages of sandstone and mudstone that were deposited and accreted to Gondwanaland during the Carboniferous to Early Cretaceous, and have locally been metamorphosed into greywacke semischist or schist. Convergence across the Australian-Pacific plate boundary pushes thinned and submerged crust upward into the path of a westerly atmospheric circulation pattern. Differential uplift, erosion and rock exhumation across the Southern Alps has exposed transitions from uncleaved greywackes in the east, through weakly cleaved or fractured greywacke and foliated semischist about the Main Divide, to strongly-foliated amphibolite facies schist (almost gneiss) in the west adjacent to the Alpine Fault (Fig. 1). Late Cretaceous – Pliocene sedimentary and volcanic Tertiary rocks occur locally in the Canterbury foothills, beneath the Canterbury Plains, and locally west of the Alpine Fault. Uplift of the Southern Alps by folding and faulting continues to the present day at up to 10 mm y⁻¹. The entire area is subject to episodic shaking from high-magnitude M7 – 8 earthquakes every 200 – 300 years, occurring most recently in 1717 AD (Wells et al. 1999; Sutherland et al. 2007), and appearing seismically quiescent since observations began in the early 1900’s.
Pleistocene – Holocene glacial cycles carved and shaped the central Southern Alps landscape, leaving a mantle of cover deposits throughout the region. Glaciations began at least by the Late Pliocene (Suggate and Wilson 1958) and by the Late Pleistocene, an extensive system of glaciers extended almost uninterrupted 700 km along, and 100 km across the Southern Alps (Newham et al. 1999). Glaciers coalesced in the main mountain valleys to form piedmont lobes in the west and extended through the foothills to alluvial outwash plains in the east. Till deposits mark the extent of ice during the Last Glacial Maximum (LGM) around 26-23 ka (cal. yr. BP); there is widespread evidence for at least four major periods of glacial re-advance since the LGM (Fitzsimons 1997). Since the end of the Little Ice Age (LIA) in the mid-19th Century there has been significant ice-loss (49% decrease in glacier area and 61% decrease in glacier volume) (Hoelzle et al. 2007), and the disappearance of many snowfields, despite episodic local advances in response to changes in atmospheric flow patterns (Chinn et al. 2005). Accompanying and contributing to the disintegration of many glacial tongues has been the formation of large lakes, which continue to grow rapidly in proglacial areas east of the Main Divide (Allen et al. 2009), increasing the potential for flood waves from bedrock failures impacting these lakes (eg, McSaveney 2002).

A current estimate of permafrost distribution in the central Southern Alps (Allen et al. 2008b) has been calculated on the basis of a topo-climatic key used to describe the lower limits of permafrost occurrence in the European Alps (Haeberli 1975). Physical relationships incorporated in this empirical approach consider aspect dependent radiation effects, altitudinal changes in air temperature, and topographically related snow cover variation. The altitudinal limits given in the original key were adjusted using local air temperature records from the past two decades and estimated free air lapse rates for three climate stations located across the study region, and validated on the basis of fossil and active rock glacier distribution (Allen et al. 2008b). Although found only in the drier mountain ranges southeast of the Main Divide, rock glaciers can only be maintained where annual ground temperatures remain <0°C and therefore provide a reliable indication of past or present permafrost. Higher snowfall near the Main Divide favours perennial ice development on low angled slopes and rock glaciers are not observed (Brazier et al. 1998). However, on steeper, largely snow-free shaded aspects near the Main Divide, the lower limit of permafrost occurrence is estimated at 2300 m, rising to nearly 3000 m on sunnier aspects, while a drier climate and higher free air lapse rates are estimated to lower these limits by ~300 m further towards the southeast (Allen et al. 2008b). Long term climate records from Hokitika are often used as a proxy for the
general climate of the central Southern Alps; recent warming of 0.7 °C is documented for the period 1920 – 1990 (Salinger et al. 1995), suggesting slowly degrading permafrost may exist 100 – 150 m below the current estimated permafrost limits (Gruber and Haeberli 2007).

5. Regional topography

Elevations of central South Island extend from sea level up to the highest peaks of the Mount Cook massif above 3700 m. Local relief near the Main Divide rises 1000 – 2700 m within a horizontal distance of less than 5 km. A large proportion of the land area is within lowland valleys where fluvial processes dominate, and slope angles are low. Above 1000 m, hillslope area decreases with elevation, such that only 6% of all slopes are located at elevations above 2000 m (Fig. 2a). However, the higher-elevation slopes are progressively steeper and more extensively covered by perennial ice (Fig. 2b), with modal (φ) and mean (ψ) slope gradients in excess of 45° characterizing the highest slopes in the vicinity of the Main Divide. Between 1000 and 2000 m in the Southern Alps, φ is relatively constant at 33 – 34°, with only a small increase in ψ from 29 – 31°. Although located at a higher elevation, this zone has slope characteristics similar to the subalpine domain where Korup et al. (2005) considered mass movement and fluvial erosion/sedimentation to be the dominant geomorphic processes occurring west of the Main Divide. The effects of relief damping by glacial ice cover that were described for the alpine zone (Korup et al. 2005) are not distinguishable within the larger scale study area. In addition, there is no clear differentiation in slope between the alpine and high alpine zones. Hence, a single alpine altitudinal zone is used here to refer to steep terrain above 2000 m. Ablation areas of the larger valley glaciers extend below 1000 m east of the divide, and nearer to sea level for glaciers in the west, such as Franz Josef and Fox Glaciers. Whitehouse (1988) described three main geomorphic domains occurring across the central Southern Alps, resulting from differential tectonic uplift, precipitation, erosion, and glaciation. These domains provide a basis for analyses here considering topographic and landslide related variances across the alps, but also correspond well with estimated permafrost distribution across the region (Allen et al. 2008b). In the western domain permafrost is unlikely to occur, becoming more frequent about alpine slopes in the axial domain, and restricted to the highest shaded slopes in the eastern domain (Fig. 2c) where the only active rock glaciers are found (Brazier et al. 1998).
Figure 2. a) Hillslope frequency distribution and slope angles calculated in 250-m elevation increments, with three altitudinal zones identified (coloured). b) Proportion of perennial ice cover out of total hillslope area with respect to elevation. c) Maximum elevation transects across the central Southern Alps (refer to Fig. 1), showing altitudinal zones (coloured) and lower limits of permafrost distribution (after Allen et al. 2008b) in relation to the three main geomorphic domains described by Whitehouse (1988).

For the purpose of analysing landslide distribution in relation to geological setting, a simplified geological classification scheme was developed (Table 1), reflecting both the metamorphic bedrock gradation that occurs southeast of the Alpine Fault, and the existence of various cover sequences resulting form dynamic glacial, periglacial, and fluvial processes. Over half of the region is mapped by QMAP as cover formations (exceeding 50 m in depth, or 1 km² areal extent), which are predominantly alluvium and till deposits, The remaining landmass consists of exposed bedrock slopes having thin or no surface cover, with greywacke predominating in vast areas east of the Main Divide. The distribution of geological units is non-uniform with slope aspect (Fig. 3). Schist and
semischist bedrock is exposed more frequently on slopes facing north through to southwest, while greywacke is mapped relatively evenly across all slope aspects. The higher proportion of cleaved greywacke on east- and southeast-facing slopes reflects uplift and exposure of these rocks in the hangingwall of west-dipping faults, such as the Main Divide Fault Zone (Cox and Findlay 1995) and Great Groove Fault in the Sealy Range (Lillie and Gunn 1964). Schist and semischist rocks are almost entirely exposed west of the Main Divide and exhibit slopes that are noticeably steeper than greywacke and cleaved greywacke for many aspects. The steepest schist and semischist slopes occur on northeast to east facing aspects. The mapped covered surfaces show considerably lower modal slope angles than exposed bedrock, with alluvium gravels, till, and tertiary cover predominating on gentle depositional slopes (< 4°), with alluvium particularly evident in valleys and on outwash plains facing south to southeast. Scree shows a relatively uniform φ = 29 – 32°, which is less than the angle of repose (35°) described for typical talus slopes in the Southern Alps (Whitehouse and McSaveney 1983). The distribution of perennial ice generally reflects the predominance of large glacial neves and low gradient valley glaciers flowing from the Main Divide, but a bimodal slope distribution results from the occurrence of numerous steeper mountain glaciers and ice cliffs. On leeward, and shaded southeast facing slopes, steeper hanging and cliff glaciers actually predominate over flatter, ice covered slopes (Fig. 3c).

Table 1. Simplified geological classification scheme derived for the central Southern Alps.

<table>
<thead>
<tr>
<th>Class</th>
<th>Description</th>
<th>Regional Coverage %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water COVER</td>
<td>Surface water on land – lakes and lagoons</td>
<td>2</td>
</tr>
<tr>
<td>Ice</td>
<td>Glaciers (mapped from 1986 aerial imagery)</td>
<td>3</td>
</tr>
<tr>
<td>Scree</td>
<td>Talus and colluvium on hillslopes, including some hanging-valley till deposits either &gt;50 m thick or, &gt;1 km² areal extent</td>
<td>3</td>
</tr>
<tr>
<td>Alluvium</td>
<td>Fan gravels and outwash alluvium, infilling major valley floors</td>
<td>34</td>
</tr>
<tr>
<td>Till</td>
<td>Thick glacial sequences and moraine deposits &gt;50 m thick</td>
<td>12</td>
</tr>
<tr>
<td>Tertiary cover</td>
<td>Late Cretaceous-Pliocene sedimentary and volcanic rocks</td>
<td>3</td>
</tr>
<tr>
<td>BEDROCK</td>
<td>Interbedded sandstone, siltstone and argillite of the Torlesse composite terrain</td>
<td>27</td>
</tr>
<tr>
<td>Greywacke</td>
<td>Weakly developed cleavage evident</td>
<td>6</td>
</tr>
<tr>
<td>Cleaved greywacke</td>
<td>Cleaved and weakly schistose greywacke and argillite of the Torlesse composite terrain</td>
<td>2</td>
</tr>
<tr>
<td>Semischist</td>
<td>Strongly foliated and metamorphosed schist (almost gneiss)</td>
<td>7</td>
</tr>
<tr>
<td>Paleozoic basement</td>
<td>Meta-sedimentary and plutonic basement rocks northwest of the Alpine Fault</td>
<td>1</td>
</tr>
</tbody>
</table>
6. Landslide distribution

414 km$^2$, or nearly 2% of the total land area in the central Southern Alps is affected by landsliding. The distribution of landslides is centred upon the mountainous terrain and foothills both east and west of the Main Divide, with flat lowland areas such as the Canterbury and Mackenzie District plains largely unaffected by landslide activity (Fig. 1). The size of mapped events (source and deposit areas combined) ranges from 0.02 km$^2$ to more than 14 km$^2$. Over 20% of all events included in the inventory (rock avalanche or other landslide) can be arbitrarily labelled large (>1 km$^2$), and these events constitute 57% of the total landslide-affected area. No landslide volumes are included in the inventory because pre-failure topography cannot be reconstructed in most early historic or prehistoric cases. However, events from recent decades provide an indication of the large magnitudes that have been involved in rock avalanches such as Aoraki/Mt Cook (12 x 10$^6$ m$^3$) (McSaveney 2002), or Mt Adams (10-15 x 10$^6$ m$^3$) (Hancox et al. 2005), or the extreme volumes involved in complex landslides such as the ongoing failure beneath the Mueller Hut on the Sealy
Range (>100 \times 10^6 \text{ m}^3) (Cox and Barrell 2007). 27 of the mapped rock avalanches included in the inventory are listed as prehistoric failures by Whitehouse (1983), for which deposited volumes in the order of \(10^6 – 10^8 \text{ m}^3\) are given. Within the subsequent analyses, two datasets are referred to; ‘rock avalanche’ analyses are based on the 105 source areas identified from mapped rock avalanches, while ‘nonspecific landslide’ analyses are based upon the source areas of the remaining 405 events for which no other specific failure type has been assigned.

### 6.1 Geological distribution

The proportions of different geological units affected by evidence of rock avalanche and nonspecific landslide activity show clear differentiation (Fig. 4). These proportions are calculated as the source area cells (25-m pixels) divided by the total number of cells for each of the geological classes, expressed as a percentage. Many scree slopes in the region are formed within or beneath nonspecific landslide source areas (Fig. 4a), while others have been formed and modified by frequent rockfall, snow avalanche, or smaller debris flows not included within the inventory. Tertiary rocks which form lowland hillslopes bordering the Canterbury Plains also appear unstable and prone to landsliding. Till has been displaced within large landslide complexes, where rapid downwasting of valley glaciers has destabilized adjacent lateral moraines (Blair 1994). In some notable instances above the Tasman, Murchison and Mueller Glaciers, active landslide scarps propagate large distances uphill where bedrock has also destabilized in response to longer term glacial recession. Greywacke shows the least evidence for landsliding, but landslide occurrence increases significantly for the higher-metamorphic-grade bedrock slopes (Fig. 4a). Semischist, which lies in a relatively thin zone near the Main Divide shows the greatest occurrence of landsliding, more than five times higher than greywacke bedrock. In the schist terrain west of the Main Divide, shallow debris failures and very large, slow failure of entire hillslopes are thought to be the common landslide mechanisms, accounting for an average 9 ± 4 mm y\(^{-1}\) mass wasting (Hovius et al. 1997) and contributing exceptional sediment delivery to rivers draining the windward side of the Southern Alps (Korup et al. 2004; Korup 2005b). This geological association is reversed for rock avalanche activity, with source areas most evident from greywacke and cleaved greywacke slopes which predominate in areas along and east of the Main Divide (Fig. 4b), supporting a regional distribution pattern earlier described by Whitehouse (1983). Although geomorphic evidence is less commonly observed, large rock avalanches can occur from schist slopes in the west, as illustrated by the catastrophic 1999
failure of Mt Adams (Hancox et al. 2005), but few events have been mapped from zones of semischist bedrock.

Figure 4. a) Proportion of different geological units affected by evidence of nonspecific landslide source areas, and b) rock avalanche source areas.

6.2 Topographic distribution

The distribution of both rock avalanche and nonspecific landslide source areas predominates from slopes orientated west to northwest across all geomorphic domains of the central Southern Alps (Fig. 5a-c). This trend is most pronounced for slopes within the axial domain of the alps, where 66% of all nonspecific landslide source area cells are observed on west to northwest facing slopes, and nearly 30% of all rock avalanche source areas are orientated to the northwest (Fig. 5b). In this region, a large proportion of rock avalanches are also observed from slopes facing east to southeast (27%), although few other landslides are observed from these orientations. In the much larger eastern domain of the alps, where the greatest proportion of rock avalanche and other landslides have been mapped (Fig. 5d), a more uniform distribution of nonspecific landsliding is observed
across all slope aspects, while the predominance of rock avalanching from west to northwest facing slopes remains evident (Fig. 5c). The spatial density of nonspecific landsliding is highest for hillslopes in the western domain, with 1 event observed within every 25 km², increasing to 27 km² in the axial domain and 37 km² in the eastern domain. Rock avalanches are observed less frequently, with the highest density in the axial domain of 1 event per 72 km², increasing to 1 event per 123 km² further east.

Figure 5. a-c) Distribution of all hillslopes, nonspecific landslide source areas and rock avalanche source areas as a function of slope aspect across three geomorphic domains of the central Southern Alps (after Whitehouse 1988). d) Proportion of all nonspecific landslides, rock avalanches, and hillslopes observed within these geomorphic domains. Only 3 rock avalanches have been observed in the western domain, and therefore, the apparent distribution of these source areas should not be considered a regional trend (a).

Within the western and axial domains, source area slope angles for nonspecific landslides (φ = 32°), appear subdued with respect to general hillslope angles (φ = 36°) (Fig. 6a). Rather than resulting from any geomorphic imprint caused by landsliding, this probably reflects the predominance of nonspecific landslides occurring from within lowland to subalpine elevation zones (Fig. 7.), where hillslope angles are less than at higher elevations. In the eastern domain, a closer correspondence between modal hillslope and nonspecific landslide slope angles is observed (φ = 32°). Rock
Avalanche source areas in the eastern domain are steeper and normally distributed ($\varphi = 35^\circ$), often creating convex cirque type depressions where deep-seated failures have occurred (Turnbull and Davies 2006) (Fig. 6b). In the axial domain a larger proportion of steep slope angles are evident within rock avalanche source areas, reflecting a greater number of failures occurring from the steeper alpine zone above 2000 m (Fig. 7), where many recent events have exposed steep scarps near or at the ridgeline.

Figure 6. a) Slope frequency distributions for all hillslopes and nonspecific landslide source areas across three geomorphic domains of the central Southern Alps (after Whitehouse 1988). b) Slope frequency distributions for all hillslopes and rock avalanche source areas in the axial and eastern domains only. c) Proportions of all hillslopes affected by evidence of nonspecific landslide (red) and rock avalanche (blue) source areas, calculated within 5° slope increments.
As a fundamental driver of instability, slope is frequently incorporated into landslide susceptibility analyses (eg, Donati and Turrini 2002; Pike et al. 2003). By comparing the slope angle distribution for rock avalanche and nonspecific landslide source areas against the distribution for all hillslopes in the region, a notable increase in the proportion of avalanche failures occurring on slopes >50° is revealed. This relationship is not evident for other landslide source areas, because the various failure mechanisms involved in these cases operate across a wider range of lithologies (Fig. 4a) with variable strength properties, and originate from within a lower elevation range (Fig. 7) where slope angles are less. For the steepest slopes, which occur more frequently at higher elevations, rock avalanching and higher frequency/lower magnitude rockfalls appears the dominant modes of slope failure.

![Graph showing frequency distributions for maximum and minimum elevations of landslide and rock avalanche source areas](image)

**Figure 7.** Frequency distributions for the maximum and minimum elevations of nonspecific landslide and rock avalanche source areas calculated within 250-m increments. Also included is the altitudinal distribution of all hillslopes in the central Southern Alps.

7. Glacial change and bedrock stability

7.1 Altitudinal distribution of rock avalanching

Nearly 25% of all rock avalanches have source areas where the maximum elevation is positioned within the alpine zone, and for approximately half of these events, the minimum elevation of the source area has also been positioned above 2000 m (Fig. 7). In contrast, evidence from other landslide mechanisms indicates that source areas in most instances have originated from within
subalpine to lowland elevation zones. In considering the possible influences of permafrost recession and glacial retreat on slope stability, particular focus is therefore given here to the distribution of prehistoric and historic rock avalanches, allowing comparisons to be made with current estimated zones of perennial-ice and permafrost in the central Southern Alps (Fig. 8). The glacial zone in this context is indicated by the steady-state glacier equilibrium line altitude (ELA), defining the longer-term elevation above which ice accumulation is favoured over ablation, and above which the proportion of perennially ice covered slopes increases significantly (Fig. 2b). Because of strong precipitation gradients, this elevation rises steeply from ~1600 m west of the Main Divide to over 2200 m in the east, and includes possible variations of up to 200 m from sunny northern to shaded southern slope aspects (Chinn 1995). While it appears that prehistoric events have not occurred from higher elevation slopes (Fig. 8), this must be considered an observational bias, with prehistoric events depositing onto glaciers being poorly preserved and therefore missing from the inventory. In addition, many prehistoric events identified originally by Whitehouse (1983) were located east of the Main Divide where fewer high elevation slopes exist, and the spatial distribution of prehistoric events is surely influenced by earthquake generated events (see Section 8). However, these reservations should not detract from the clear predominance of recent events (< 100 yr. BP) originating from glaciated slopes and/or permafrost terrain (Fig. 8).

Of the 19 events to have been recorded over the past 100 years in the central Southern Alps, over half have initiated from source areas where the base elevation of the failure has been above 2000 m, and only 2 have involved sources areas located entirely below 2000 m (Table 2). One of these lower elevation exceptions was a rainfall triggered event from the slopes above the downwasting Murchison Glacier, and the other from beneath Mt Beatrice where the retreat of the Hooker Glacier since the LIA has exposed the flanks of steep, U-shaped valley walls. These 19 recent events have initiated from a variety of collapse structures, span a range of magnitudes, feature extremely steep failure slopes (42 – 73°), and have all had source areas located either directly above or below, or within a vertical distance of 300 m from large valley or smaller mountain glaciers (Table 2). In some instances, steep cliff or hanging glaciers have been present within the source areas, and have contributed to the failure mass (McSaveney 2002). The mean elevation difference between rock avalanche source areas and the estimated lower permafrost boundary is 240 m, with a total of 12 events occurring from within 300 m of this lower boundary, where degrading permafrost is expected
(Fig. 8). If only the base level of the source area is considered, 7 events are identified within 300 m of the lower permafrost boundary, including all 4 rock avalanches occurring during summer 2007/08. The 1991 summit failure of Aoraki/Mt Cook, and an earlier event from Mt Vancouver at 3300 m initiated from considerably higher elevations where warming permafrost was unlikely to be a factor, although the presence of hanging glaciers can produce sub-surface warm thermal anomalies (Haeberli et al. 2004; Huggel 2008). With a significant proportion of the event ages remaining unknown, no temporal changes in rock avalanche initiation altitude can or should be inferred.

Figure 8. Mean elevation of 105 rock avalanche source areas. The maximum and minimum elevations of the source area for events occurring about the Main Divide in the past 100 years are also indicated (bars), with the yellow tick denoting events that occurred during the warm dry summer of 2007/08 (Table 2). The estimated permafrost zone is indicated for the Main Divide (after Allen et al. 2008b) and overlaps with the glacial zone. Dates older than 100 yr. BP are after Whitehouse (1983).
Table 2. Significant rock avalanches and rockfalls recorded in the central Southern Alps over the past 100 years.

<table>
<thead>
<tr>
<th>Location</th>
<th>Date</th>
<th>Max elevation source (m)</th>
<th>Min elevation source (m)</th>
<th>Slope aspect source</th>
<th>Median slope source (°)</th>
<th>Source lithology and collapse structure</th>
<th>Estimated permafrost lower limit (m)</th>
<th>Glacial conditions¹</th>
<th>Total area (km²)</th>
<th>α (°)²</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mt Isobel I</td>
<td>1950 – 1955</td>
<td>2180</td>
<td>1780</td>
<td>SE</td>
<td>47</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2440</td>
<td>Gl. directly above Gl. ~280 m below</td>
<td>0.66</td>
<td>23</td>
<td>McSaveney 2002</td>
</tr>
<tr>
<td>Mt Isobel II</td>
<td>c. 1965</td>
<td>2360</td>
<td>2200</td>
<td>E</td>
<td>59</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2590</td>
<td>Steep glacial ice Gl. directly below</td>
<td>0.25</td>
<td>30</td>
<td>McSaveney 2002</td>
</tr>
<tr>
<td>Mt Vancouver</td>
<td>1974 – 1975</td>
<td>3300</td>
<td>3100 (uncertain)</td>
<td>E</td>
<td>61</td>
<td>Greywacke, scarp slope</td>
<td>2590</td>
<td>Steep glacial ice - -</td>
<td>-</td>
<td>-</td>
<td>McSaveney 2002</td>
</tr>
<tr>
<td>Murchison Glacier³</td>
<td>25/12/75</td>
<td>1820</td>
<td>1590</td>
<td>NW</td>
<td>33</td>
<td>Greywacke, partly dip slope</td>
<td>2690</td>
<td>Gl. directly below</td>
<td>0.43</td>
<td>11</td>
<td>Whitehouse &amp; Griffiths, 1983</td>
</tr>
<tr>
<td>Beelzebub Glacier</td>
<td>1980 – 1984</td>
<td>2030</td>
<td>1900</td>
<td>S</td>
<td>42</td>
<td>Cleaved greywacke, dip slope</td>
<td>2390</td>
<td>Gl. directly below</td>
<td>0.24</td>
<td>18</td>
<td>Korup 2005a</td>
</tr>
<tr>
<td>Aoraki/Mt Cook</td>
<td>14/12/91</td>
<td>3760</td>
<td>2870</td>
<td>E</td>
<td>51</td>
<td>Greywacke, scarp slope</td>
<td>2590</td>
<td>Steep glacial ice &amp; hanging glaciers</td>
<td>14.6</td>
<td>22</td>
<td>McSaveney 2002</td>
</tr>
<tr>
<td>Mt Fletcher I &amp; II⁴</td>
<td>02/05/92</td>
<td>2420</td>
<td>1810</td>
<td>E/SE</td>
<td>50</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2590 - 2440</td>
<td>Steep glacial ice, Gl. ~100 m below</td>
<td>2.22</td>
<td>22</td>
<td>McSaveney 2002</td>
</tr>
<tr>
<td>La Perouse</td>
<td>1994 – 1995</td>
<td>2410</td>
<td>2220</td>
<td>NW</td>
<td>47</td>
<td>Greywacke, dip slope</td>
<td>2690</td>
<td>Steep glacial ice, Gl. ~200 m below</td>
<td>0.37</td>
<td>29</td>
<td>This study</td>
</tr>
<tr>
<td>Mt Thomson I</td>
<td>22/02/96</td>
<td>2230</td>
<td>1640</td>
<td>E</td>
<td>59</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2590</td>
<td>Gl. directly above Gl. directly below</td>
<td>0.30</td>
<td>42</td>
<td>McSaveney 2002</td>
</tr>
<tr>
<td>Mt Thomson II</td>
<td>2002 – 2004</td>
<td>2470</td>
<td>2200</td>
<td>E</td>
<td>51</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2590</td>
<td>Steep glacial ice Gl. directly below</td>
<td>0.17</td>
<td>37</td>
<td>This study</td>
</tr>
<tr>
<td>Mt Adams</td>
<td>06/10/99</td>
<td>2130</td>
<td>1190</td>
<td>N</td>
<td>47</td>
<td>Schist, scarp slope</td>
<td>&gt; 3000</td>
<td>Gl. on opposing slope facet</td>
<td>14.7</td>
<td>37</td>
<td>Hancox et al. 2005</td>
</tr>
<tr>
<td>Vampire Peak I</td>
<td>2003</td>
<td>2560</td>
<td>2440</td>
<td>SE/S</td>
<td>65</td>
<td>Semischist, joint controlled scarp slope</td>
<td>2440 - 2390</td>
<td>Gl. ~200 m below</td>
<td>0.42</td>
<td>24</td>
<td>Cox et al. 2008</td>
</tr>
<tr>
<td>Mt Beatrice</td>
<td>23/1/04</td>
<td>1720</td>
<td>1580</td>
<td>NE</td>
<td>53</td>
<td>Cleaved greywacke, toppled scarp slope</td>
<td>2840</td>
<td>Gl. ~100 m below</td>
<td>0.18</td>
<td>20</td>
<td>Cox et al. 2008</td>
</tr>
<tr>
<td>Anzak Peaks (rockfall)</td>
<td>2007 – 2008</td>
<td>2008</td>
<td>1810</td>
<td>SE</td>
<td>43</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2440</td>
<td>Gl. ~50 m above Gl. ~700 m below</td>
<td>-</td>
<td>41</td>
<td>This study</td>
</tr>
<tr>
<td>Vampire II</td>
<td>7-13/1/08</td>
<td>2520</td>
<td>2380</td>
<td>SE</td>
<td>73</td>
<td>Semischist, joint controlled scarp slope</td>
<td>2440</td>
<td>Gl. ~100 m below</td>
<td>0.49</td>
<td>28</td>
<td>Cox et al. 2008</td>
</tr>
<tr>
<td>Douglas Peak</td>
<td>18/02/08</td>
<td>3010</td>
<td>2910</td>
<td>NW</td>
<td>56</td>
<td>Greywacke, dip slope</td>
<td>2690</td>
<td>Gl. ~230 m below</td>
<td>0.03</td>
<td>46</td>
<td>Cox et al. 2008</td>
</tr>
<tr>
<td>Mt Spencer</td>
<td>6-7/04/08</td>
<td>2720</td>
<td>2650</td>
<td>NW</td>
<td>53</td>
<td>Greywacke, dip slope</td>
<td>2690</td>
<td>Gl. ~100 m below</td>
<td>0.05</td>
<td>34</td>
<td>Cox et al. 2008</td>
</tr>
<tr>
<td>Mt Hilscombe I &amp; II⁴</td>
<td>24/04/08</td>
<td>2650</td>
<td>2570</td>
<td>E</td>
<td>45</td>
<td>Cleaved greywacke, scarp slope</td>
<td>2590</td>
<td>Gl. directly below</td>
<td>0.02</td>
<td>35</td>
<td>Cox et al. 2008</td>
</tr>
<tr>
<td>Malte Brun</td>
<td>01/01/09 – 18/02/09</td>
<td>2590</td>
<td>2400</td>
<td>S</td>
<td>58</td>
<td>Greywacke, Scarp slope</td>
<td>2390</td>
<td>Gl. directly above Gl. directly below</td>
<td>0.38</td>
<td>21</td>
<td>This study</td>
</tr>
</tbody>
</table>

¹ Presence of ice in the source area and ice proximity measured above and below the mapped source area, based on 2006 perennial ice mapping (after Allen et al. 2008a).
² α or Farthoobung – Angle of reach measured from avalanche source to toe of the deposit.
³ Rainfall triggered event. ⁴ Events originated from a common source area.
7.2 Recent failures above the Mueller Glacier

The Mueller glacial valley may typify geological and glacial conditions occurring about the Main Divide. The steep, pervasively fractured hangingwall rocks of the Main Divide Fault Zone are cross-cut by numerous secondary faults, partially covered by steep cliff and poly-thermal hanging glaciers, and >1000 m of lateral support has been removed from the valley flanks since the LGM (McSaveney 2002). Historical glacial extents were reconstructed using aerial photography from the years 1965 and 1986, and compared to satellite-based glacial mapping from 2006 (Allen et al. 2008a). Together with inferred permafrost distribution, this information can be compared with the source areas identified from a sequence of spontaneous rock avalanches occurring since the mid-1950s (Fig. 9). Both Mt Thompson events appear to have originated from within zones where some loss of perennial-ice cover is evident between 1965 and 1986, while the earliest Mt Isobel event originated from steep slopes approximately 400 m above the downwasting Mueller Glacier. In contrast, there is no notable loss of perennial ice visible near the source areas of the 1965 Mt Isobel or more recent Vampire Peak events. However, these failures are positioned within, and adjacent to estimated zones of marginal permafrost. The 2008 Vampire Peak event may be significant given that it occurred following a prolonged period of >0 °C air temperature at the elevation of the failure, and was the first of a sequence of high elevation rock avalanches to occur during a particularly warm and dry summer (Cox et al. 2008). Qualitative accounts referred to the abnormally rapid melt of snow/ice during the summer, and there were several reports of abnormally high rockfall activity from local guides and helicopter pilots. Smaller rockfalls and debris avalanches from the steep former glacier supported walls above the Mueller Glacier are widespread (eg, Mt Bannie, Fig. 9), and help maintain a thick supra-glacial debris cover. By comparison, a large and complex deep seated landslide on the true right of the glacier has been creeping slowly over many centuries, with the toe of the slope gradually unloaded since the LGM (Cox and Barrell 2007).
Figure 9. Recent rock avalanche events occurring from Mt Thomson (TM), Mt Isobel (IS), and Vampire Peak (VP), Mueller Glacier, Aoraki/Mt Cook National Park, with year of failure indicated. a) Rock avalanches shown in relation to changing perennial ice extent from 1965 to 2006. b) Rock avalanches shown in relation to zones of probable and possible permafrost (after Allen et al. 2008b). The zone of possible permafrost is expected to be characterized by degrading permafrost. Also indicated is an active rockfall onto the Mueller Glacier from beneath Mt Bannie (yellow dashed), and a large active, deep seated landslide propagating from the crest of the Sealy Range (orange dashed).

7.3 Future expectations relating to atmospheric warming

Rock avalanches recorded over the past century within the central Southern Alps, including those described for the Mueller Glacier, clearly indicate that recent rock-slope failures have predominated from high elevation slopes of the Main Divide (Fig. 8, Table 2). In particular, rock avalanches have been most frequent over this time from east – southeast aspects, which probably reflects the significantly steeper slope of these aspects at elevations above 2000 m (Fig. 10), where rocks dipping towards the west in the hangingwall of the Main Divide Fault Zone are exposed (Cox and
Findlay 1995). The hangingwall rocks are more pervasively fractured, faulted, and deformed than the greywacke of the lower elevation footwall rocks. Also rock-mass bulging is likely to be widespread throughout the highest slopes of the Main Divide, occurring in response to stress release and glacial debutressing (Prebble 1995). Unfortunately, given uncertainties associated with the prehistoric record of rock avalanching, and in particular the absence of evidence for high elevation failures, there can be no statistical basis to determine whether or not the recent frequency and/or magnitude of events occurring from steep high elevation slopes of the Main Divide is significantly greater than at any time in the past. However, in view of fundamental relationships describing atmospheric warming, perennial ice loss, permafrost degradation, and slope stability (eg, Haeberli et al. 1997; Harris 2005; Gruber and Haeberli 2007), a future increase in the frequency of rock failure occurring from the highest slopes of the axial domain could be expected as ice levels and permafrost limits rise in response to predicted atmospheric warming (IPCC 2007) (Fig. 10).

![Modal slope as a function of aspect calculated within 250-m elevation increments. Atmospheric warming, glacial recession, and permafrost degradation are expected to raise the lower elevation limit of the perennial-ice and permafrost zones, increasing the proportion of rock avalanche source areas (dashed) observed from higher elevation slopes. The exposure of hangingwall rocks between 2250 and 3250 m creates steeper slopes on east – southeast facing aspects, where numerous events from the past 100 years have originated.](image-url)
While total hillslope area is less at higher elevations, the bedrock that exists within permafrost terrain will in the short term be vulnerable to rapid active layer thickening, and over the longer term, gradual warming at greater depths (Gruber and Haeberli 2007). For example, above the Mueller Glacier ∼10 km² of hillslope area at elevations over 2000 m is currently ice covered and/or within estimated permafrost terrain, and therefore vulnerable to future adverse affects of atmospheric warming. Across the entire central Southern Alps, 12% (45 km²) of all hillslopes steeper than 50° are currently ice covered and/or within estimated permafrost terrain. Steep slopes above the estimated permafrost limit are almost entirely composed of jointed greywacke and cleaved greywacke bedrock, with only localized zones of schist and semischist outcropping about the Main Divide, south of Aoraki/Mt Cook. Hence, the importance of future permafrost degradation in this region appears relevant only in relation to the failure of steep greywacke slopes, primarily as rock avalanches. Any significance in relation to other forms of landslide failure within schist bedrock, or debris instabilities seems unlikely given these processes and lithologies predominate at subalpine elevations, well below the current permafrost limit.

8. Discussion

Limitations associated with this regional-scale approach to landslide analyses in the central Southern Alps relate primarily to mapping generalizations, uncertainties in mapping glacier and permafrost distribution, DEM resolution, and simplifications associated with the use of zonal statistics. The generalization of higher resolution geological mapping in the creation of 1:250,000 Aoraki QMAP led to a loss of precision in the mapped area of small landslides, loss of lithological detail, and amalgamation of some small or thin cover sequences into mapped bedrock units and vice-versa. Vegetation cover, which is particularly thick in subalpine terrain west of the Main Divide, restricted the direct observation of cover sequences, meaning their mapping was based primarily on the morphology of landforms developed on these surfaces, supplemented with available surface exposures (Cox and Barrell 2007). In a similar manner, identification of landslide affected terrain using air photo interpretation is compromised where thick revegetation of disturbed surfaces has occurred (Brardinoni et al. 2003). These deficiencies, combined with significantly higher erosion and mass turnover west of the Main Divide (leading to shorter survival time of rock avalanche and other landslide evidence), suggest that estimates of mass movement frequency may be negatively biased in the schist terrain of the western domain relative to eastern areas. Despite these concerns,
the spatial density of nonspecific landslides is still highest in the west (Fig. 5d), supporting the assumption of Whitehouse (1983), that a lack of rock avalanche evidence in the west is likely to be true reflection of a significantly lower occurrence rate than in the east, and not a result of mapping biases. Using zonal statistics with landslide source areas enabled a larger range of cells to be sampled for each slope failure than would have been possible using only QMAP scarps, and the influences of spurious values (resulting from positional errors) were minimized using modal and median zone values. Although important slope angle, slope aspect, and lithological variations within a source area would be better distinguished using a more sophisticated approach, the use of zonal statistics better suits the resolution of the underlying topographic and geological data. For example, slope angle is calculated over a 75-m length, and therefore even the most accurately digitised head scarp is unlikely to be represented at this scale.

The predominance of all landslides from slopes orientated west – northwest suggests a structural control on landslide distribution across the central Southern Alps, given the prevalent bedding and schistosity dip-direction is towards the west/northwest or east/southeast (Cox and Barrell 2007). While landslides occurring in the western domain have been characterized by failure along weaknesses within the schistosity (Whitehouse 1988), relationships linking rock avalanching, other landslides, and structural orientations in the greywacke terrain of the axial and eastern domains are not well established. From the small catalogue of historical events, there is no clear tendency towards dip slope failures but important unanswered questions relate to the role of co-seismic landsliding. Figures 8 and 10 demonstrate differences in the apparent spatial distribution of rock avalanches observed over the past 100 years, compared to the larger record of prehistoric events. The fact that the historical record contains no earthquake generated failures is significant, and suggests that recent failures from steep glaciated scarps on east and southeast aspects of the Main Divide reflect the relatively quiescent seismic activity of the past 100 years. Prehistoric failures from similar slopes would have occurred, but with the evidence not surviving, the prehistoric record appears biased by large, potentially earthquake generated events (eg, Whitehouse and Griffiths 1983; Bull and Brandon 1998; Orwin 1998; Smith et al. 2006), that may in many instances exhibit bedding plane failure (Whitehouse 1983). Catastrophic recent failure of schist rocks from Mount Adams (Table 2) (Hancox et al. 2005) and a landslide into the Young River further south along the alps, are not likely to be related to any recent influence of glacial recession or permafrost degradation. Likewise the shape and slope of many peaks in the central Southern Alps attests to the
importance of large catastrophic rock avalanches in forming the topography of this region during considerably cooler geological time periods. Therefore, although being potentially an important transient influence on slope stability during recent and into future decades, the broader geomorphic significance of glacial/permafrost related slope failure may represent a relatively minor process superposition within a seismically active, vigorously evolving landscape.

Figure 11. Transect across the three main geomorphic domains of the central Southern Alps (after Whitehouse 1988), showing frequency distribution of rock avalanche and other landslides in relation to generalized bedrock lithology, glaciation (inferred from ELA) (Chinn 1995), estimated lower limits of permafrost distribution (Allen et al. 2008b), and patterns of erosion, uplift and precipitation.

Quantifying any recent influence of atmospheric warming, permafrost degradation, and perennial ice melt on slope instabilities, and importantly distinguishing this influence from other tectonic and climatic forcing is difficult. Slope failure distribution across the alps is summarized in Figure 11, highlighting the predominance of nonspecific landslide mechanisms observed within the schist and semischist of the western domain, where uplift and erosion are at their highest levels (Whitehouse 1988; Fitzsimons and Veit 2001), slope stability is locally influenced by weakening along the Alpine Fault (Korup 2004), and precipitation is near its maximum (Griffiths and McSaveney 1983).
Evidence for rock avalanching is rare in this same region, prevailing instead about the Main Divide Fault Zone in the axial domain of the Alps and further east. Considering earthquake – landslide interactions established elsewhere in New Zealand (eg, Pearce and O'Loughlin 1985; Crozier et al. 1995; Hancox et al. 2003), seismicity must also be considered a primary predisposing factor and trigger of mass slope failures throughout the central Southern Alps. Additionally, McSaveney (2002) suggested glacial thinning has led to stress adjustment within some exceptionally steep slopes near the Main Divide. As was surmised for slopes above the Mueller Glacier, other high-elevation events could be more directly associated with altered physical weathering regimes following ice loss within steep detachment zones. However, such historical recession is generally difficult to quantify, with only a limited record of aerial imagery extending back to 1965, and these images often severely distort the steepest slopes. In addition, distinguishing perennial ice from late lying seasonal snow patches can be problematic when interpreting remotely sensed imagery (Allen et al. 2008a). Therefore, while the widespread recession of larger valley glaciers is well documented (eg, Hochstein et al. 1995), and related influences on valley wall stability described (Blair 1994), quantifying surface geometry changes of smaller, steep ice masses and any related influence on slope stability is hampered by a lack of current high resolution terrain data.

An increased frequency of catastrophic landslides over recent decades has been associated with climate change in other mountainous regions such as British Columbia (Geertsema et al. 2006), and at least four large rock avalanches have initiated from within steep permafrost zones of the European Alps during the past decade (Gruber and Haeberli 2007). Although the words ‘extreme’ or ‘exceptional’ are often attributed to high-mountain mass movements, particularly where climate change impacts are implicated, the historical information required to support such inferences is mostly lacking. In the central Southern Alps of New Zealand, large-magnitude, high-elevation events from pre-1900 are known to be missing from the inventory, illustrated by the uncertainty surrounding a large event thought to have originated somewhere near the summit of Aoraki/Mt Cook (Barff 1873). Distinct avalanche lobes visible on the lower Hooker Glacier in photographs from 1893 are located too far down the glacier to realistically correspond to this event, and therefore constitute another sequence of slope failures without any definitive origin (Fig. 12). This loss of evidence from the historical record creates major uncertainty with any attempt to quantify the influence of climate change on slope instabilities within glacial environments.
9. Conclusions

510 landslides have been analysed within a GIS inventory, with combined source and deposit areas affecting ~2% of the central Southern Alps. Rock avalanching appears to be most abundant from greywacke and cleaved greywacke slopes which outcrop steeply along and east of the Main Divide. Within the schist and semischist bedrock which is exposed closer towards the Alpine Fault, other landslide mechanisms predominate, and evidence for rock avalanching is apparently less abundant west of the Main Divide. For all landslide failures, the greatest proportion of events have occurred from slopes facing west – northwest, and the likelihood of rock avalanche failure dramatically increases on bedrock slopes exceeding 50°. Above 2500 m in the alpine zone, rock avalanching appears the primary large magnitude mechanism for bedrock slope failure, and over the past 100 years numerous apparently spontaneous events have initiated from glaciated slopes near the Main Divide, with many detachment zones likely to be characterized by degrading permafrost. These recent rock avalanches have prevailed from east – southeast aspects on the hangingwall of the Main
Divide Fault Zone, but the prehistoric record of landsliding in the region is potentially too incomplete and influenced by large seismic related failures to quantitatively infer that changes in slope failure distribution have occurred in response to recent climate change. Warming predicted for the coming century causing continued downwasting of valley glaciers, destabilization of their adjacent walls, and loss of perennial ice and permafrost degradation at higher elevations will negatively influence bedrock stability in the central Southern Alps, where approximately 45 km$^2$ of steep (>50°) slopes are currently ice covered or within permafrost terrain. Irrespective of any direct role future climate variability might have, the fundamental relationship between rock avalanche detachment and slope implies that the next major earthquake in the region will trigger many catastrophic failures from steep glaciated slopes near the Main Divide. Additional predisposing weaknesses resulting from glacial recession and permafrost warming can only add to the likelihood of failure from these affected slopes.

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Exploring Steep Bedrock Permafrost and its Relationship with Recent Slope Failures in the Southern Alps of New Zealand

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ABSTRACT
The central region of New Zealand’s Southern Alps is characterised by steep, glaciated slopes prone to rock mass failure, but permafrost conditions and any relevance to past or future slope instabilities have received little previous attention. A network of 15 dataloggers was used to record near-surface temperatures on steep rock walls located about the Main Divide of the Alps and further leeward where the climate is much drier. Mean annual rock temperature (MART) ranged from -1.9 to 5.4°C, corresponding to local 0°C elevations (E0) of 2465–3514 m, with no significant difference observed between the humid and drier mountain ranges. On extremely shaded slopes, the permafrost limit may extend down towards 2000 m, but further measurements are needed to confirm this. E0 levels were modelled as a function of potential solar radiation, allowing steep permafrost distribution to be mapped across the region. From an inventory of 19 bedrock failures occurring about the Main Divide since the mid-20th century, 13 have initiated from source areas where MART in the range of +/-1.8°C is considered to indicate marginal permafrost conditions. None of these events was triggered by seismic activity, and mostly exhibit scarp areas that include or originate in close-to-ridge topography, where the most rapid permafrost degradation might be expected. Copyright © 2009 John Wiley & Sons, Ltd.
First approaches towards modelling glacial hazards in the Mount Cook region of New Zealand’s Southern Alps

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Abstract. Flood and mass movements originating from glacial environments are particularly devastating in populated mountain regions of the world, but in the remote Mount Cook region of New Zealand’s Southern Alps minimal attention has been given to these processes. Glacial environments are characterized by high mass turnover and combined with changing climatic conditions, potential problems and process interactions can evolve rapidly. Remote sensing based terrain mapping, geographic information systems and flow path modelling are integrated here to explore the extent of ice avalanche, debris flow and lake flood hazard potential in the Mount Cook region. Numerous proglacial lakes have formed during recent decades, but well vegetated, low gradient outlet areas suggest catastrophic dam failure and flooding is unlikely. However, potential impacts from incoming mass movements of ice, debris or rock could lead to dam overtopping, particularly where lakes are forming directly beneath steep slopes. Physically based numerical modeling with RAMMS was introduced for local scale analyses of rock avalanche events, and was shown to be a useful tool for establishing accurate flow path dynamics and estimating potential event magnitudes. Potential debris flows originating from steep moraine and talus slopes can reach road and built infrastructure when worst-case runout distances are considered, while potential effects from ice avalanches are limited to walking tracks and alpine huts located in close proximity to initiation zones of steep ice. Further local scale studies of these processes are required, leading towards a full hazard assessment, and changing glacial conditions over coming decades will necessitate ongoing monitoring and reassessment of initiation zones and potential impacts.

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1 Introduction

Glacial hazards have been defined as any glacial or glacier-related feature or process that adversely affect human activities (Reynolds, 1992). These hazards are a concern for many high mountain regions of the world and can have severe resource management implications relating to water availability and hydro-power generation (Reynolds, 1992; Richardson and Reynolds, 2000). Glaciers are highly sensitive to climate forcing, and in many regions glacial retreat during the past century has now continued beyond any known historical positions, drastically altering the geomorphic process activity of the environment and shifting zones of hazard initiation (Kääb et al., 2005). Glacial hazards include ice avalanches, glacial floods, and debris flows, although the most catastrophic events have involved complex chain reactions or transformation of mass movements or floods into rapid debris flows (e.g., Huggel et al., 2005). Other hazards can develop directly from surge type glacial movements (Haebelerli et al., 2002), while large bedrock failures are increasingly being studied in relation to changing glacial and permafrost conditions (Harris, 2005). Fundamental research into phenomena such as ice avalanches (Alean, 1985), glacier-related debris flows (Rickenmann, 1999) and flooding (Clague and Evans, 2000; Haebelerli, 1983; Maizels and Russel, 1992) has come from Canada, Central Europe and Iceland, where populated villages and transport infrastructure extend into the glacial environment. Understanding of such processes and potential impacts in the Mount Cook region of New Zealand is comparatively limited, despite recognition of 20th century glacial recession (Chinn, 1996), associated destabilisation of surrounding terrain and lake formation (Blair, 1994; Hochstein et al., 1995), and the potential for large magnitude chain reaction events involving mass movements into glacial lakes (McSaveney, 2002).
This paper aims to use remote sensing, geographic information system (GIS) capabilities and flow path modelling to identify potential initiation zones and explore the spatial distribution of potential ice avalanches, debris flows and catastrophic outburst floods in the Mount Cook region of New Zealand. In addition, the first results from numerical modelling of recent rock avalanches are introduced, providing a tool for assessing future impacts from large bedrock failures in the region. The emphasis of this study is on events originating from the current glacierised and recent paraglacial landscape, which is inferred here to be terrain uncovered since the Little Ice Age (LIA) maximum was reached (see Sect. 3). It is beyond the scope of this contribution to provide a hazard assessment. Instead, the primary objective here is to apply conservative, worst-case scenario modelling of potential events, thereby directing future studies towards locations where infrastructure and human activities may be exposed to mass movements and/or flood inundation. Limitations of the approach will be discussed, and the impact of changing glacial conditions will be considered in relation to future event processes.

2 Background

Glacial floods refer to the sudden discharge of a water reservoir that has formed either underneath, at the side, in front, within, or at the surface of a glacier (Richardson and Reynolds, 2000). For reservoirs developing as lakes on or at the margin of glaciers, remote sensing at suitable spatial and temporal resolution is an appropriate tool for monitoring hazardous developments (e.g., Huggel et al., 2002). Floods are the most far reaching of all glacial hazards, often initiating from the catastrophic failure of moraine dammed lakes, for which the term glacial lake outburst flood (GLOF) has been adopted (Richardson and Reynolds, 2000). GLOFs frequently transform into hyperconcentrated or debris flow events following the entrainment of paraglacial debris, and are a well documented hazard for Central Asia (Ding and Liu, 1992; Quincey et al., 2007), the Andes (Reynolds, 1992), Canada (Clague and Evans, 2000), and Europe (e.g., Haeberli, 1983). Ice avalanches occur when large masses of ice detach from steep cliff or ramp type glaciers (Alecan, 1985) as frontal block failures, slab failures, or deeper failures at the ice/bedrock interface (Richardson and Reynolds, 2000). Relative to glacial floods, ice avalanches typically travel shorter distances and directly affected areas are normally restricted to densely populated alpine regions (Salzmann et al., 2004). However, far reaching disasters have resulted from process interactions involving combined ice/rock avalanches and transformations into debris or mudflows, such as the 1970 Huascaran disaster in Peru (Carey, 2005), or the more recent 2002 Kolkà-Karmadon avalanche in the Caucasus (Huggel et al., 2005). Periglacial and recently uncovered paraglacial environments are prone to debris flow initiation because they are characterised by large accumulations of unconsolidated sediment, in the form of moraine deposits and talus slopes (Evans and Clague, 1994). For example, over 50% of debris flows observed during one particularly severe year in the Swiss Alps initiated from zones that had deglaciated within the previous 150 years (Zimmermann and Haeberli, 1992). Unlike debris flows on lowland hillslopes, glacier-related events can become activated via several mechanisms, including snow or ice melt and high intensity rainfall (Chiarle et al., 2007; Rickenmann and Zimmermann, 1993), permafrost degradation (Harris, 2005), or from catastrophic entrainment within glacial floodwaves (Clague and Evans, 1994; O’Connor et al., 2001).

Although New Zealand has significant areas of mountainous terrain, the population is almost entirely located in lowland areas, and therefore floods and mass movements occurring in high mountain regions have received comparatively less scientific attention. Studies have described flood hazard problems from Franz Josef and Fox Glaciers where the sudden release of sub- or glacial reservoirs can mobilise large amounts of sediment (Davies et al., 2003; Goodsell et al., 2005). Debris flow activity initiated during storm events is known to potentially endanger infrastructure within the Mount Cook Village (Skermmer et al., 2002; Whitehouse, 1982), and the catastrophic downsloping of large valley glaciers is resulting in widespread moraine wall failure and the destruction of backcountry huts and tracks (Blair, 1994). Ice avalanche activity in the Mount Cook region predominates as low magnitude (<1000 m³), high frequency events (1–8 events/h) from steep cliff-type glaciers (Iseli, 1991), producing a significant hazard on many of the well known climbing routes during the summer ablation season (Irwin et al., 2002). Large bedrock failures from glaciated areas of the Southern Alps have received greater scientific attention (e.g., Korup, 2005a; Whitehouse and Griffiths, 1983), partly because road infrastructure can be at risk (Paterson, 1996), downstream chain reactions can endanger lowland areas (e.g., Davies and Scott, 1997; Hancox et al., 2005), and recent glacial changes appear to be increasing the frequency of high magnitude events and likelihood of impacts into glacial lakes (McSaveney, 2002).

An important development in glacial hazard research has been the implementation of multi-hazard approaches within a GIS environment, integrating remote sensing detection of individual ice, debris and flood hazard sources with combined empirical and hydrological approaches to event path modelling (Huggel et al., 2004). Despite the implied suitability of these approaches to large scale, regional applications, to date, results have mainly been illustrated for case-study type scenarios within the European Alps. The Mount Cook region of New Zealand’s Southern Alps provides an opportunity to implement GIS based procedures across a large, dynamic and diverse region where the primary objective is to gain first order knowledge of glacial hazard potential and recognition of affected areas. Particular emphasis
is given to glacial lake flooding, and exploring the use of combined remote sensing/GIS methods to distinguish lakes formed within steep morainic debris. In recognition of the importance of large bedrock failures in the region (e.g., McSaveney, 2002; Whitehouse and Griffiths, 1983) this study is expanded beyond GIS based modelling of glacial lake floods, debris flows and ice avalanches, to include a numerical modelling approach that may be used to simulate potential rock avalanche impacts at a more detailed level of investigation.

3 The Mount Cook region

The Mount Cook region is broadly defined here to encompass the Aoraki Mount Cook National Park, extending west of the main divide into the Westland National Park and further south towards the Ben Ohau Range, with a total land area of over 3500 km² (Fig. 1). The region includes the highest mountains and the most heavily glacierised terrain of New Zealand’s Southern Alps. Permanent snow covered peaks are highest towards the Ben Ohau Range, with a total land area of over 3500 km². The region includes the highest mountains and the most heavily glacierised terrain of New Zealand’s Southern Alps. Permanent snow covered peaks are found between 2500 and 3754 m, with local relief in the order of 1000–2700 m (McSaveney, 2002). Moist westerly airflow generates high orographic precipitation about the main divide (> 12 m y⁻¹), but an extreme leeward gradient produces <4 m y⁻¹ only 20 km further to the southeast (Henderson and Thompson, 1999). The timing of the LIA maximum varied across the region, but between 1750 and 1890 widespread retreat had begun, becoming most evident during the mid 20th century, and in total, a 49% loss in ice area has been estimated, although some highly responsive glaciers have advanced recently owing to short term climate variations (Chinn, 1996). The Alpine Fault is a major active fault traversing the northwestern edge of the study area, on which most of the ongoing tectonic movement between the Australian Plate (to the northwest) and the Pacific Plate (to the southeast) is concentrated. Plate convergence across the Australian-Pacific plate causes continued uplift of the Southern Alps in the order of 10 mm y⁻¹, which is approximately in balance with regional erosion (Koons, 1990). The last major (M 8) earthquake on the Alpine Fault has been dated at 1717, although the expectation is for episodic shaking from high-magnitude (M7-8) events every 200–300 years, (Wells et al., 1999).

The Mount Cook Village is the main service and accommodation centre for visitors that access the region from the east, with a small permanent population of ~100 and a larger seasonal component linked to tourism activities in the region. West of the main divide, the largest tourist and residential centres are Franz Josef and Fox Glacier Villages, with ~300 permanent residents housed in each village. Figures provided by the Department of Conservation indicate that backcountry hut and camp ground usage in the Mount Cook National Park averaged nearly 10 000 bed-nights per year during the monitoring period 1982 to 2002. In the Westland National Park, 4500 visitors walked the Copland Valley track in 2007 (Fig. 1), and 250 000 made the short walk to the terminus of the Fox Glacier (FX), with these numbers slowly increasing over recent years. Main roads connect to all tourist and residential areas, but the nearest transalpine passes are 80 km to the southwest and 150 km to the northeast. Numerous unsealed vehicle tracks give access to more remote areas and farm buildings.

4 Regional hazard modelling

4.1 Modelling strategy

Allen et al. (2008b) applied remote sensing based methods to map the distribution of glacial lakes, ice, and debris accumulations across the Mount Cook region. These procedures used visible, near infrared (NIR) and shortwave infrared (SWIR) bands of the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) satellite sensor, and were based upon an orthorectified scene of 24 January 2006. The inclusion of topographic information from the Landcare Research L2, 25 m gridded South Island digital elevation model (DEM) enables detection of steep, potentially unstable surfaces on the basis of empirically derived values describing critical slope gradients (Table 1). The DEM is derived from stereographic analyses of 1:50 000 aerial photography captured in 1986, and although errors are strongly related to landform, a root mean square error of 5–8 m has been calculated for hilly and steep terrain (Barringer et al., 2002).

Potential debris flow, ice avalanche, and flood events were modelled at the regional scale using the modified single flow (MSF) routing algorithm developed by Huggel et al. (2003). Simple GIS spatial queries could then identify where event paths potentially intersect with infrastructure or where interactions such as ice impacts into glacial lakes might occur (Table 2). The MSF model is based on the D8 flow direction algorithm, modified to provide a quasi-qualitative likelihood that a given cell will be affected by the flow path based on distance from the source area, and allowing for deviation of up to the 45° either side from the path of steepest descent. However, it must be noted that this should not be interpreted as a hazard probability because event frequency and timing is not considered. In addition, the MSF model has no physical basis, and therefore cannot represent more complex behavior of mass movements such as barrier run-up and overtopping. Despite these obvious limitations, the MSF model has proven a useful tool for early recognition of hazard potential, and is particularly well suited for use at larger spatial scales where multiple source areas are identified, and direct field observations may be difficult (e.g., Huggel et al., 2004; Schneider et al., 2008). As with previous studies, a worst-case scenario approach is used, whereby modelled flow paths are continued.
until a probable maximum runout is reached. This maximum runout is described by an angle of reach $\alpha$, which defines the average slope of a line between the starting and end points of a mass movement (Table 1).

4.2 Glacial lake floods

Glacial lakes were mapped with ASTER using the Normalised Difference Vegetation Index (NDVI) \((\text{NIR3-RED2})/(\text{NIR3+RED2})\), and an additional threshold of...
Table 1. Parameters used to identify potential source areas and defining the maximum probable runout of glacial lake floods, debris flows, and ice avalanches.

<table>
<thead>
<tr>
<th></th>
<th>Glacial lake floods</th>
<th>Debris flows</th>
<th>Ice avalanches</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Surface characteristics</strong></td>
<td>Lakes on or at the margins of a glacier</td>
<td>– Debris accumulations occurring within glacial, or recent glacial zones</td>
<td>Steep glacial ice</td>
</tr>
<tr>
<td></td>
<td>Expanding lake area</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Steep moraine dammed lakes</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Critical slope gradient</strong></td>
<td>Sediment entrainment and hyperconcentration:</td>
<td>Flow initiation:</td>
<td>Temperate ice:</td>
</tr>
<tr>
<td></td>
<td>– 10° (Clague and Evans, 1994; Hungr et al., 2005)</td>
<td>– 25–38° (Hung et al., 1984; Rickenmann and Zimmermann, 1993; Takahashi, 1991)</td>
<td>– 25°</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Cold ice:</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>– 45° (Alean, 1985)</td>
</tr>
<tr>
<td><strong>Maximum probable runout</strong></td>
<td>Clear water flood:</td>
<td>– 11° angle of reach (Rickenmann, 2005; Rickenmann and Zimmermann, 1993)</td>
<td>– 17° angle of reach (Alean, 1985; Huggel et al., 2004)</td>
</tr>
<tr>
<td></td>
<td>– May exceed 200 km and attain angle of reach &lt;3°</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>GLOF triggered debris flow:</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>– 11° angle of reach</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Huggel et al., 2002; McKillop and Clague, 2007)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

RED2/GREEN1, providing a useful distinction of all water bodies in the region regardless of turbidity levels (Allen et al., 2008b). In total, 65 glacial lakes were identified but some were small interlinked supraglacial lakes and ponds. To exclude the smallest lakes from the flood modelling, a lake area threshold of 1500 m² was used, leaving 54 remaining lakes. The majority of lake water exists at low elevations (750–1000 m), where large proglacial lakes have formed on the Murchison (MN), Tasman (TN), Hooker (HR) and Mueller (MR) glaciers east of the main divide (Figs. 1 and 2). Because bathymetric measurements have been limited to piecemeal observations, estimation of lake volumes across the wider region must rely upon relationships describing lake volume as a function of surface area, derived from measurements in Canada, North America, Nepal, Central Europe, and South America (O’Connor et al., 2001; Huggel et al., 2002). The available measurements from the Mount Cook region enable some local validation of these relationships (Table 3). Because it results in significantly smaller errors, the relationship of Huggel et al. (2002) is preferred, albeit with the understanding that actual volumes of large proglacial lakes may be underestimated by ~20–50% because of the exceptional depths (>55 m) of these lakes compared with the sample from which the relationship was established. The Mueller Lake appears an anomaly, with very shallow depths possibly relating to greater sedimentation into the lake and/or the younger development phase of the lake. The majority of lakes are estimated to have relatively small volumes (modal volume of 0.05×10⁶ m³) but 14 lakes have estimated volumes larger than 1×10⁶ m³, with the Tasman Lake containing more than 200×10⁶ m³ (Fig. 3a).

Because the majority of glacial lake water in the Mount Cook region is located at lower elevations characterized by gentle slope gradients (Fig. 2), it is considered inappropriate to assume lake failures will transform into debris flow events. Therefore, potential flood events were initially modelled as clear water floodwaves, for which flood attenuation was allowed to continue until the great lakes of Tekapo or Pukaki were reached in the east, or until the ocean was reached in the west. In fact, travel distances in excess of 200 km have been recorded in the Karakorum Himalaya from glacial floodwaves (Hewitt, 1982). In a second step, the modelled flood paths were analysed to identify outlet channel areas
Table 2. Summary of MSF modelling results, giving the extent of human infrastructure intersecting with modelled flow paths for the worst-case probable maximum runout. Also provided is the number of glacial lakes positioned within incoming flow paths, giving an indication of where potential flow transformations or impact waves might be initiated. Modelling was repeated for reduced runout scenarios, with minimum $\alpha$ increased by 50 and 100%.

<table>
<thead>
<tr>
<th>Glacial Floods</th>
<th>Tourist or residential dwellings</th>
<th>Remote buildings</th>
<th>Main roads</th>
<th>Vehicle tracks</th>
<th>Foot tracks</th>
<th>Power/phone Lake impacts</th>
</tr>
</thead>
<tbody>
<tr>
<td>i) Clear water</td>
<td>Mount Cook Village (MCV) and airport. Floodplains southwest of Fox Glacier village.</td>
<td>NIL</td>
<td>17 km</td>
<td>29 km</td>
<td>3 km</td>
<td>12 km</td>
</tr>
<tr>
<td>ii) Debris flows</td>
<td>NIL</td>
<td>NIL</td>
<td>NIL</td>
<td>NIL</td>
<td>NIL</td>
<td>NIL</td>
</tr>
<tr>
<td>11° ($\tan\alpha=0.19$)</td>
<td>MCV and locations along main access road. Glentanner Airport.</td>
<td>17 huts</td>
<td>17 km</td>
<td>33 km</td>
<td>37 km</td>
<td>3 km</td>
</tr>
<tr>
<td>+50% 16.5° ($\tan\alpha=0.30$)</td>
<td>MCV</td>
<td>13 huts</td>
<td>8 km</td>
<td>22 km</td>
<td>31 km</td>
<td>2 km</td>
</tr>
<tr>
<td>+100% 22° ($\tan\alpha=0.40$)</td>
<td>MCV</td>
<td>10 huts</td>
<td>2 km</td>
<td>16 km</td>
<td>21 km</td>
<td>NIL</td>
</tr>
<tr>
<td>Ice Avalanches</td>
<td>NIL</td>
<td>10 huts</td>
<td>2 km</td>
<td>16 km</td>
<td>30 km</td>
<td>NIL</td>
</tr>
<tr>
<td>17° ($\tan\alpha=0.31$)</td>
<td>NIL</td>
<td>10 huts</td>
<td>2 km</td>
<td>16 km</td>
<td>30 km</td>
<td>NIL</td>
</tr>
<tr>
<td>+50% 25.5° ($\tan\alpha=0.48$)</td>
<td>NIL</td>
<td>7 huts</td>
<td>1 km</td>
<td>NIL</td>
<td>23 km</td>
<td>NIL</td>
</tr>
<tr>
<td>+100% 34° ($\tan\alpha=0.67$)</td>
<td>NIL</td>
<td>NIL</td>
<td>NIL</td>
<td>4 km</td>
<td>NIL</td>
<td>NIL</td>
</tr>
</tbody>
</table>

characterized by steep debris, and if appropriate, runout distances could then be re-evaluated for possible debris flow scenarios. A secondary output from the MSF model provides the horizontal distance along the flood path from the lake source. For floods originating from moraine breaches in the Three Sisters Range, Oregon, O’Connor et al. (2001) established 500 m as the horizontal distance in which floodwaves transformed into debris flow via sediment entrainment. This distance will be highly variable depending on local topography and channel morphology, but 500 m is taken here to arbitrarily encompass the dam and surrounding outlet channel areas for all glacial lakes (Fig. 4a).

ASTER-based methods were used to map debris and vegetation coverage within the lake outlet areas, supported by higher resolution satellite imagery and field photography. Large debris accumulations across the region have previously been identified and distinguished from bedrock using an image texture algorithm (Allen et al., 2008b). This idea stems from novel approaches developed with higher resolution imagery in the Swiss Alps (Huggel et al., 2004) and recognizes the uniformity of a debris surface compared to the rough, angular appearance of bedrock. The success of the procedure becomes more limited when applied with lower resolution ASTER imagery, but larger talus slopes, moraine deposits and outwash gravels were all identified. Vegetated surfaces were mapped using the NDVI. In a first step, outlet areas containing less than a combined 20% coverage of vegetation or debris were considered to be non-debris. For all other outlets, a larger presence of vegetation or non-vegetated debris was considered to indicate alluvium or moraine composition. If the vegetation abundance relative to non-vegetated debris exceeded 75%, the outlet area was categorized as ‘vegetated’ otherwise the outlet was categorized as “debris”. A mean slope threshold of 10° was used to further distinguish steep outlet areas in which sediment entrainment is considered likely, based on debris flow studies of Hungr et al. (1984) and observations of GLOF events in Canada (Clague and Evans, 1994).
Table 3. Comparison between measured and estimated proglacial lake volumes in the Mount Cook region using two different empirical relationships.

<table>
<thead>
<tr>
<th>Lake</th>
<th>Observation year</th>
<th>Reference</th>
<th>Area (m² 10⁶)</th>
<th>Mean depth (m)</th>
<th>Volume (m³ 10⁶)</th>
<th>Estimated volume¹ (m³ 10⁶)</th>
<th>Estimated volume² (m³ 10⁶)</th>
<th>Error¹ (%)</th>
<th>Error² (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tasman</td>
<td>1993</td>
<td>Hochstein et al. (1995)</td>
<td>2.0</td>
<td>71</td>
<td>139</td>
<td>89</td>
<td>647</td>
<td>−36</td>
<td>365</td>
</tr>
<tr>
<td></td>
<td>2002</td>
<td>Röhler (2005)</td>
<td>3.5</td>
<td>75</td>
<td>263</td>
<td>205</td>
<td>2075</td>
<td>−22</td>
<td>689</td>
</tr>
<tr>
<td>Hooker</td>
<td>1995</td>
<td>Warren and Kirkbride (1998)</td>
<td>0.7</td>
<td>55</td>
<td>41</td>
<td>23</td>
<td>95</td>
<td>−44</td>
<td>132</td>
</tr>
<tr>
<td></td>
<td>2002</td>
<td>Röhler (2005)</td>
<td>0.9</td>
<td>65</td>
<td>59</td>
<td>30</td>
<td>140</td>
<td>−49</td>
<td>137</td>
</tr>
<tr>
<td>Mueller</td>
<td>2002</td>
<td>Röhler (2005)</td>
<td>0.5</td>
<td>9</td>
<td>4.3</td>
<td>12</td>
<td>40</td>
<td>340</td>
<td>830</td>
</tr>
</tbody>
</table>

¹ Estimate based on the equation of Huggel et al. (2002), where lake volume (V) in m³ is expressed by the relationship:

\[ V = 0.1044A^{1.42}. \]

² Estimate based on the equation of O’Connor et al. (2001) where lake volume (V) in m³ is expressed by the relationship:

\[ V = 3.114A + 0.0001685A^2 \]

where (A) is lake area in m².

* Error is calculated as the difference between measured and estimated volumes, divided by the measured volume.

The majority of outlet areas were classified as “debris”, with only one non-debris (bedrock) outlet identified. Lakes formed within debris mantled glacial ice could not be automatically excluded from the classification. These lakes have small volumes, and occur on low gradient supraglacial areas of the larger valley glaciers. Nearly 70% of terrain contained within outlet channel areas is characterized by slope gradients less than 10° (Fig. 3b), but several examples of lakes formed within steep vegetated and non-vegetated morainic debris occur in cirque basins primarily east of the main divide. In these instances, potential debris flow initiation is considered possible, and paths were therefore remodelled using a maximum runout defined by an 11° angle of reach recognized from GLOF triggered debris flow observations in the European Alps and Canada (Huggel et al., 2002; McKillop and Clague, 2007). Relative to modelled clear water floodwaves, flood triggered debris flows all appear to terminate well before huts, vehicle tracks, or other infrastructure are reached (Fig. 5, Table 2). Because the MSF model does not consider wave height and barrier overtopping cannot be simulated, the modelled floodwave down the Godley River is initially confined within a shallow channel leading from the lake outlet. However, evidence from a flood event which occurred from a rock avalanche impact in 1992 indicated that a 7–10 m high wave completely inundated the flat area surrounding the lake outlet before rapidly dispersing downstream on the wide braided river plains (McSaveney, 2002).

Vegetated outlet areas characterize most large proglacial lakes which have formed within low gradient moraine and outwash gravels during the past two decades (e.g., Figs. 4c and 7b). The presence of vegetation implies some longer

![Fig. 3. A Frequency distribution of glacial lake volumes in the Mount Cook region, January 2006, estimated from ASTER mapped lake areas using an empirical relationship between lake area and volume (after Huggel et al., 2002). B Frequency distribution of slope gradients for all pixels contained within and surrounding lake outlet channel areas.](www.nat-hazards-earth-syst-sci.net/9/481/2009/)
erodibility of the channel area, limiting the likelihood of natural dam failure. However, when considering worst-case scenarios, future dam overtopping from displacement waves generated by mass movement impacts cannot be excluded. This would be most concerning where permanent building and road infrastructure are positioned within the flood plains of larger volume lakes, which occurs on the West Coast southwest of Fox Glacier Village, and along the road leading into Mount Cook Village (Table 2).

4.3 Debris flows

Steep ASTER classified debris accumulations that might give rise to a flow event were distinguished within a 25°–38° slope range. Although specific to the source lithology involved and the associated angle of repose, this slope range typifies starting conditions observed internationally for debris flow events (e.g., Hungr et al., 1984; Rickenmann and Zimmermann, 1993), and corresponds to slope gradients measured for talus slopes in the Southern Alps of New Zealand (Dunning, 1996; Whitehouse and McSaveney, 1983). Steep debris accumulations are widespread across the region, but the emphasis of this study is towards situations where current and recent glacial processes directly influence debris accumulation and where ice or perennial snow melt can lead to flow initiation. To achieve this focus, a GIS buffer was used, extracting all debris accumulations within a maximum distance of 750 m from ASTER mapped glacial ice, including manually digitized debris covered ice. This distance is based upon average glacial length changes observed throughout the Southern Alps since the LIA, and also encompasses the likely extent of vertical recession over this time (Chinn, 1996). To remove isolated pixels, debris accumulations used as input into the MSF model were restricted to a minimum area of 1 × 2 pixels, or ∼1250 m², representing areas of lateral moraine above the downwasting valley glaciers, terminal moraine deposits from
steeper glaciers, cirque glacial moraines, and talus slopes beneath glaciers or large perennial snow patches (Fig. 1).

The most comprehensive observations of debris flow travel distances in alpine terrain comes from the Swiss Alps, where a minimum $\alpha=11^\circ$ (tan $\alpha=0.19$) has been recognized for coarser grained flows (Rickenmann and Zimmermann, 1993), and subsequently used to define the maximum runout for modelled periglacial events (Huggel et al., 2004). Application of this worst-case scenario to potential events in the Mount Cook region identifies large sections of roading, tracks, huts, shelters and farm buildings positioned within modelled debris flow paths (Table 2). This is particularly evident east of the main divide, where a drier climate favours extensive talus development (Whitehouse, 1988) and vehicle tracks extend higher into the headwaters of braided river valleys (Fig. 1). Built infrastructure located within the Mount Cook Village, and nearby roading appears to intersect with potential paths originating from high on the Sealy Range where remnant cirque glaciers and perennial snow remain (Fig. 6). However, the maximum probable runout distance for these events exceeds any recognized threat to the village (McSaveney and Davies, 2005) and appears to overestimate flow propagation far beyond the composite fans upon which infrastructure is located. Minimum and maximum slope gradients observed on these fans range from <2 to 7°.

West of the main divide, narrow steep gorges transport potential flows originating in the glacierised catchments down towards, but stopping short of directly reaching the state highway. In these instances, temporary blockage, dambreak, and remobilization of debris are possible within the confines of a steep gorge, and hence, a far reaching hazard potential may exist to not only the road, but villages such as Franz Josef located nearby (e.g., Davies and Scott, 1997).

Application of the MSF for multiple events at the regional level allows only a single worst-case maximum runout parameter, but an alternative estimate for individual events might consider available catchment area ($A_c$) above the debris source. For this purpose, $\tan \alpha=0.20 A_c^{-0.26}$ has been used to describe the minimum $\alpha$ observed for debris flows in alpine areas of Switzerland and Canada (Rickenmann, 2005). Based on catchment areas calculated using ArcGIS watershed function, a significantly reduced maximum runout is estimated for modelled events affecting the Mount Cook Village and nearby areas (Fig. 6). For example, with a catchment area of $\sim0.09 \text{ km}^2$, a potential event originating from talus debris in the upper reaches of Kitchener Creek establishes a minimum $\tan \alpha=0.37$, although this reach is still in excess of 300–500 m beyond any recent aggradation visi-
Fig. 6. A MSF modelling of selected debris flows initiating from glacial zones. The flow paths are terminated using the maximum probable runout defined by a minimum $\alpha=11^\circ$ ($\tan \alpha = 0.19$). Additional runout distances are indicated, corresponding to a minimum $\alpha$ increase of 50% ($\tan \alpha = 0.30$) and 100% ($\tan \alpha = 0.40$), and also using a minimum $\alpha$ related to catchment area (after Rickenmann, 2005). B Closer inspection of a potential source area from a snow covered talus slope at the head of Kitchener Creek (KN), and C deposition area with village infrastructure located towards lower right. High resolution QuickBird image is from 4 May 2006.
Fig. 7. A MSF modelling of selected ice avalanches affecting the lower Mueller and Hooker Glaciers, and upper Copland Valley. The flow paths are terminated using the maximum probable runout defined by a minimum \( \alpha = 17^\circ \) (\( \tan \alpha = 0.31 \)). Additional runout distances are indicated, corresponding to a minimum \( \alpha \) increase of 50\% (\( \tan \alpha = 0.48 \)) and 100\% (\( \tan \alpha = 0.67 \)). A potential flood path propagating from the Mueller Lake is also shown. B Outlet area of the Mueller Lake looking towards the main divide, indicating potential ice avalanche trajectories towards the lake from beneath Mt Sefton (photo: S. Allen, April 2008).

The future loss of remaining glacial ice and perennial snow in these ranges will expose new rock masses to weathering, and uncover further accumulations of morainic debris.

4.4 Ice avalanche and lake interactions

ASTER-based mapping of glacial ice has been achieved using a single band ratio (NIR3/SWIR4), providing the best distinction of glacial surfaces in steep, rocky terrain (Allen et al., 2008b). Debris covered tongues are not included within the automated classification, but these areas occur predominantly on low elevation gentle slopes and are therefore unlikely to be significant in relation to avalanche hazard. Modelling potential ice instabilities originating from >400 km\(^2\) of glacier covered terrain is extremely difficult because important indicators such as crevasse patterns, ice displacements, hydrological and thermal conditions are best established through local scale analyses and field studies (e.g., Wegmann et al., 2003). At the regional scale, approaches have applied fundamental empirical observations from the Swiss Alps relating ice instability to topographic slope gradient (Alean, 1985), with a slope threshold of 25° used to classify potentially unstable steep ice (Huggel et al., 2004;
Salzmann et al., 2004). In reality, cold based ice, as occurs on steep cliffs and beneath the frontal section of many hanging glaciers at elevations where ground temperature remains <0°C (above ∼2500–3000 m) (Allen et al., 2008a), should remain stable at higher slope gradients (Table 1). However, the application of a single slope threshold is considered an appropriate worst-case scenario where polythermal glaciers are likely to occur. Steep glacial ice in the region predominates above 2000 m where modal slope gradients significantly increase (Fig. 2). To restrict the analyses to the largest potential events and exclude isolated seasonal snow, only glacial areas >2500 m² were used as input to the MSF model. The modelled avalanche runouts were terminated when a minimum α = 17° (tan α = 0.31) was achieved, corresponding to observed values in the Swiss Alps (Alean, 1985).

The smaller reach of ice avalanches combined with initiation zones which are generally located higher within the alpine valleys, excludes potential events from reaching any village infrastructure or major roads, and only a small section of the Ball Shelter vehicle track is located within the path of a far reaching event (Table 2). However, backcountry huts/shelters, and foot tracks located in closer proximity to starting zones are intersected by ice avalanche paths, as evident in the lower areas of the Hooker and Mueller Glaciers (Fig. 7). In these instances, events originating from the tongues of the Stocking (SG) and Eugenie (EG) Glaciers must travel the maximum probable distance to reach areas of human activity, while in the upper Copland Valley, smaller travel distances are sufficient to reach the walking track.

Ice avalanches from the steep cliff glaciers beneath Mount Sefton frequently deposit onto the lower Mueller Glacier (Iseli, 1991), but these low magnitude events do not exceed a runout beyond a minimum tan α = 0.67. However, a proglacial lake is currently expanding towards the ice cliffs, increasing the future potential for displacement waves, particularly from a larger magnitude event, with direct implications for adventure tourism activities operating on or near to the lake.

Ice avalanches on the lower Mueller Glacier are one example of the potential (current or future) for mass movements of ice, rock or debris to produce displacement waves, and lake flooding. A large proportion of lakes in the region show potential for interaction with incoming mass movements from ice and debris (Table 2), and bedrock instabilities are common from steep slopes of the main divide (e.g., McSaveney, 2002). The potential hazard from mass movements of ice (and/or rock) depositing into the Mueller Lake is exemplified because of the rapid enlargement of the lake over recent years, close proximity to the Mount Cook Village, and visitor infrastructure surrounding the lake (Fig. 7). In addition, incoming mass movements from the main divide can enter the Mueller Lake in line with the longitudinal axis of the lake, enabling wave energy to propagate directly towards the outlet channel. At other large proglacial lakes, on the Tasman, Hooker, and Murchison Glaciers, lakes have formed parallel to the mountain slopes, so that incoming mass movements are possible only from the lake sides, in which case, significant wave energy will be dissipated on the opposing bank.

The Mueller Lake has formed within LIA moraines, which extend up to 150 m higher than the lake level on the true right hand side, providing a natural defense structure for the camping area and village located in behind. Nearer the outlet channel, the moraine is only 10–30 m higher than the lake level, and following some initial confinement, the modelled floodwave quickly disperses on the low gradient alluvial fan, intersecting with main roads and infrastructure of the Mount Cook airport, from where tourist flights operate. Low slope gradients in all the braided river valleys suggest that transformations into debris flows within the flood path from any larger proglacial lake are unlikely.

## 5 Rock avalanche modelling

Most steep rock walls in the Mount Cook region are heavily fractured and dilated, while many have become over-steepened as a result of glacial plucking and subsequent late Holocene ice retreat from their lower flanks (McSaveney, 2002). Given the possibility for rock avalanches to deposit into expanding proglacial lakes, and the potential for rivers to become blocked, leading to catastrophic dam failure and secondary mass movement hazards (Davies and Scott, 1997; Korup, 2005b), appropriate methods at a reduced spatial scale are needed to assess rock avalanche impacts. For simulating individual avalanche events, advanced two-dimensional mass movement modelling approaches are well suited for recognition of detailed flow patterns and dynamics. A numerical rapid mass movement model (RAMMS) developed by the WSL Institute for Snow Avalanche Research, Davos Dorf, Switzerland, meets these requirements, and the simulated output is easily integrated into a GIS environment (Christen et al., 2008). This physically based dynamic model uses a finite volume scheme to solve the 2-D shallow water equations for granular flows. The frictional resistance $S_{fx}$ in x-direction and $S_{fy}$ in y-direction which is acting against gravitational acceleration, is described by using a Voellmy approach which incorporates a dry Coulomb friction $\mu$ and a turbulent friction $\xi$ (Bartelt et al., 1999):

$$S_{fx} = gH\mu \cos \alpha \tan \alpha \left( \frac{g \cos \alpha (U_x^2 + U_y^2)}{\xi} \right) \frac{U_x}{\sqrt{U_x^2 + U_y^2}}$$

$$S_{fy} = gH\mu \cos \alpha \tan \alpha \left( \frac{g \cos \alpha (U_x^2 + U_y^2)}{\xi} \right) \frac{U_y}{\sqrt{U_x^2 + U_y^2}}$$

where $g$ is the gravitational acceleration, $H$ the flow height, $\alpha$ the slope angle, and $U_x$ and $U_y$ the velocity components in x- and y-direction respectively. Details concerning momentum balance and mass conservation equations, including the matter of erosion are described by Christen et al. (2008).
The RAMMS code is based on extensive experiments within snow avalanche chutes (Kern et al., 2004) and field-based measurements (Sovilla et al., 2006), but the model is also intended to simulate mass movements other than pure snow avalanches, including large rock/ice avalanche events. Here, these model capabilities are illustrated for the first time in relation to a large rock-ice avalanche event which led to lake outburst flooding, and two smaller recent rock avalanches in the Mount Cook region.

In May and September of 1992, two separate rock avalanches with a combined volume of \(\sim 11 \times 10^6\) m\(^3\) fell from the summit area of Mount Fletcher, depositing into a nearby proglacial lake, producing flowwaves that traveled 35 km down the Godley Valley, damaging a vehicle track. Both events followed a similar path, and for numerical modeling an initial starting volume of \(8 \times 10^6\) m\(^3\) was selected, corresponding to estimates completed for the May event (McSaveney, 2002). Average density was set to 2200 kg m\(^{-3}\) representing typical greywacke bedrock \((2650\) kg m\(^{-3}\)) mixed with minor amounts of ice and firm \((500–900\) kg m\(^{-3}\) which were present in the source area. The density remains constant in the model, so that any segregation of the material during the flow process and within the deposition volume should be considered when interpreting the results. Furthermore, entrainment was not included as there was no data available to support the likely magnitude involved. However, erosion of glacial ice might have played a significant role, especially during the first impact to a large rock-ice avalanche event which led to lake outburst flooding, and two smaller recent rock avalanches in the Mount Cook region.

Wider lateral spread of the 2008 flow immediately below the moraine from the right hand edge of the flow as it traveled 350 m up the true left flank of the valley, before the majority of the mass deflects back towards the opposite wall, and then down the glacier into the lake (Fig. 8). A component of the mass reaches and overtops the ridge crest, spilling towards but stopping just short of the adjacent Godley Lake. This spillover and other lateral spreading characteristics at the lower section of the flow were used to calibrate the model geometrically. Velocity calibration was achieved on the basis of available seismic data, although the exact duration of the avalanche \((\sim 180\) s) was not clearly derived from the seismogram as the collapse was progressive (McSaveney, 2002). The best fit frictional parameters for the Mt. Fletcher avalanche were achieved using \(\mu = 0.19\) and \(\xi = 2100\) m s\(^{-2}\).

Modelled maximum flow heights correspond well with ar
derection to the larger scale modelling. This is most evident midway along the rock avalanche path where the flow travels 350 m up the true left flank of the valley, before the majority of the mass deflects back towards the opposite wall, and then down the glacier into the lake (Fig. 8). A component of the mass reaches and overtops the ridge crest, spilling towards but stopping just short of the adjacent Godley Lake. This spillover and other lateral spreading characteristics at the lower section of the flow were used to calibrate the model geometrically. Velocity calibration was achieved on the basis of available seismic data, although the exact duration of the avalanche \((\sim 180\) s) was not clearly derived from the seismogram as the collapse was progressive (McSaveney, 2002). The best fit frictional parameters for the Mt. Fletcher avalanche were achieved using \(\mu = 0.19\) and \(\xi = 2100\) m s\(^{-2}\).

Modelled maximum flow heights correspond well with areas of run-up and the general flow direction as indicated by McSaveney (2002), although the initial run-up height and spillover extent is slightly overestimated by the model. Exaggerated spread of the flow from the initiation area results from the modelled release of the detachment mass on top of the topography, rather than from within the rock face as occurs in reality. Modelled maximum flow velocities of \(102\) m s\(^{-1}\) below the area of initial run-up corresponds to estimates of \(\sim 120\) m s\(^{-1}\) made by McSaveney, while modelled mass entry into the lake occurred at a maximum velocity of \(20\) m s\(^{-1}\). By isolating the area of the flow that intersects with the lake and observing the maximum flow heights from within this zone, an estimation of the volume deposited into the lake was made. For the \(8 \times 10^6\) m\(^3\) event, only \(0.76 \times 10^6\) m\(^3\) was calculated to enter the lake. Although no measurement of the actual deposit into the lake has been possible, given the absence of any dam freeboard height (Fig. 4b) and the seiching effect of the impact wave, it is not unexpected that a displaced water volume several orders of magnitude greater than the incoming mass was recorded.

The capability of RAMMS to simulate avalanches at a much reduced volumetric magnitude was tested by modelling the two recent Vampire Peak rock avalanches described by Cox and Allen (2009). These events occurred in early summer 2003 and January 2008, with both having similar failure volumes of \(1–2 \times 10^5\) m\(^3\), and deposition volumes approaching \(3 \times 10^5\) m\(^3\). The initiation areas were located between 2500 and 2600 m, and were separated by a horizontal distance of less than 150 m. Average density was set to \(2650\) kg m\(^{-3}\) for both events, as there was no surface ice present in the source areas. The modelled flow paths accurately reflect the stronger run-up of the 2008 event, as it rose up to 80 m across a spur above the Bannie icefall (Fig. 9). Wider lateral spread of the 2008 flow immediately below the detachment resulted in reduced maximum flow heights, compared to the straighter, more direct path of the 2003 event. The runout distance of the 2008 event was also smaller, most probably relating to altered surface friction resulting from the presence of the earlier avalanche deposit. In the RAMMS model these differences were addressed using altered frictional input parameters. While the Coulomb friction \(\mu\) was set to 0.14 for both avalanches, the turbulent friction \(\xi\) was set to \(3000\) m s\(^{-2}\) for the 2003 event and to \(2000\) m s\(^{-2}\) for the 2008 event. With the lower \(\xi\)-value for the second event, the avalanche mobility is sufficiently reduced, terminating \(\sim 400\) m earlier then the 2003 event.

The limitations of using a 25 m resolution DEM for modelling low volume avalanche events were partially solved by resampling the elevation data to a 10 m grid. Although this method does not improve the underlying DEM accuracy, it does enable a higher level of precision in dynamic avalanche modelling. However, the results must be evaluated carefully considering the geometry of actual flow paths can be affected by small differences in topography which are not represented in such a resampled DEM. For example, both modelled paths show a third lobe deviating over the medial moraine from the right hand edge of the flow as it traveled through the lower Bannie icefall, but this phenomenon was not evident in either of the mapped deposits. This model artifact may result from a failure of the DEM to represent the
Fig. 8. A RAMMS simulation of the 1992 Mount Fletcher rock avalanche showing maximum flow heights and B maximum velocity. Mapped avalanche extent is after McSaveney (2002). C View of the avalanche path observed from the glacial lake where partial deposition occurred, resulting in a displacement wave and flooding (photo: M. McSaveney, September 1992).

Fig. 9. RAMMS simulations of the A 2003, and B 2008 rock avalanches from Vampire Peak, Mueller Glacier, showing maximum flow heights. The 2003 avalanche and remobilised debris deposits are reconstructed from satellite imagery from January 2006 (Cox and Allen, 2009). As indicated by the red arrows, significant displacement of the avalanche deposit down the glacier is likely to have occurred between 2003 and 2006.
micro-topography of the steep, chaotic icefall, the effects of different surface friction within the icefall that were not accounted for in the model run, or potential changes in the icefall topography that have occurred since the elevation data was captured in 1986.

Calibrating model simulations to reconstruct past events requires that the modelled flow paths fit geometrically in three dimensions as well as in time. The examples presented here form a small component of a calibration initiative for various rock/ice avalanche events from different glacial environments around the world, establishing an initial range of best fit frictional input parameters (Fig. 10) that may be used to simulate potential events where hazardous scenarios are recognized. This range of values encompass volumetric differences of two orders of magnitudes, unequal fractions of ice and water content and extremely variable topography. The development and expansion of this dataset over time, is expected to provide an indispensable basis for future model evaluation and scenario simulations with RAMMS.

6 Discussion

A first order GIS approach to glacial hazard modelling in the Mount Cook region allows non-specialist local authorities involved in regional management to quickly identify situations or locations where more detailed and comprehensive studies should be directed. With all GIS-based procedures, the reliability of the output is determined by the accuracy and characteristics of the input data used, model assumptions and limitations. Hazard sources used as input to the MSF model were derived on the basis of satellite terrain mapping and slope gradient thresholds, where the main limitations relate to the 25 m resolution of the DEM, and filtering procedures that operate across several pixel values further increasing the minimum size of recognized surface features (Allen et al., 2008b). In addition, the New Zealand 25 m DTM is developed from 1986 aerial photography, and significant topographic changes have occurred within the glacial landscape over recent decades, most evidently surrounding the larger valley glaciers, where ice surfaces have lowered up to 4 m per year (Blair, 1994). Elevation data computed from ASTER stereo imagery (Kääb et al., 2002) or the 90 m resolution shuttle radar topography mission (SRTM) (Schneider et al., 2008), can keep pace with changing environmental conditions, but provide reduced topographic detail and can have severe errors in steep terrain. For more advanced numerical models, such as RAMMS, topographic sensitivity becomes more pronounced, and therefore a lack of recent high resolution elevation data from the Southern Alps remains a limiting factor in representing complex flow dynamics.

The limitations of the MSF model are well documented by the developer (Huggel et al., 2003), and in the context of the Mount Cook region where empirical information is lacking, specifically relate to the inability of the model to give any direct information regarding potential flow magnitudes, erosion capabilities, and deposition volumes. At the next level of hazard investigation, this information may be calculated using combined empirical relationships and field observations (e.g., McKillop and Clague, 2007; Rickmann, 1999). Preferably, calculations will be verified using physically based flood and mass movement models such as FLO-2D (O’Brien et al., 1993) or DAN/DAN3D (Hungr, 1995), while the capabilities of RAMMS have been introduced for rock avalanche modeling in glacial environments. Greater input and computational requirements make these models best suited for local scale modelling of individual events or scenarios.

Empirical relationships represent a simplification of complex natural processes but remain useful for initial investigations and are most suitable for integration into GIS and remote sensing based approaches (Huggel et al., 2004). However, numerous climatic, topographic, geological, and glaciological factors differentiate the physical characteristics of processes occurring in one alpine region from another. Therefore, while often necessary, the extrapolation of well grounded empirical relationships must be treated cautiously and may not always be appropriate. Although not unique to New Zealand, the active tectonic setting and extreme precipitation gradient occurring across the Southern Alps has a

![Fig. 10. Turbulent and coulomb friction best fit parameters established from simulating rock avalanche events in New Zealand (NZ) and other glacial regions using RAMMS. These values were iteratively found with the RAMMS program and should not be confused with the physical properties of the material as the internal and basal coefficients of friction. The Coulomb friction also can not directly be transformed to the widely used ratio of fall height to runout length (H/L). In a Voellmy model, the final runout distance also depends on the turbulent friction term (see Eqs. 1 and 2).]
paramount influence on geomorphic processes in the region, and related natural hazards. In particular, sediment yields may be several orders of magnitude higher than other regions (Fitzsimons and Veit, 2001) and locally enhanced by landslide activity and fault proximity (Hovius et al., 1997; Korup, 2004), while high intensity triggers (rain or earthquake) of mass movements may occur more frequently. As a consequence, there will be uncertainties associated with any attempt to quantify the longer term influence of climate and glacial change on event frequency and/or magnitude in this region. In addition, events initiating from glacial origins are only one component of mass movement processes in this mountainous region, where failure of hillslopes, river damming, debris and clear water floods not directly linked to glacial processes remain the most recognized threats to lowland populations (e.g., Davies and Scott, 1997; Korup, 2005b). Predicted climatic change, potentially bringing even greater rainfall in the west and dryer conditions in the east (Mullan et al., 2001), adds to the complexities that scientists must consider in relation to future mass movement and flood hazards in the Southern Alps.

A comparison of twentieth century glacial changes in the New Zealand and European Alps reveals comparable reductions in ice extent although New Zealand glaciers may be more sensitive to future climate warming (Hoelzle et al., 2007). However, lower equilibrium line altitudes and therefore restricted ablation zones in the Southern Alps means that lake development and moraine destabilization associated with glacial mass wastage has so far been limited to lower elevation zones east and west of the main divide. The steeper topographic expression of the Southern Alps inhibits lake development or significant moraine deposition at higher elevations. In contrast, glacial retreat in the European Alps has resulted in lake formation and unstable moraines in high elevation hanging valleys or cirque basins, often with populated valleys located below (e.g., Huggel et al., 2003). In addition, permafrost and its climate induced degradation is expected to cause further destabilization of unconsolidated moraine deposits and talus slopes, leading to increased debris flow potential in the European Alps (Harris, 2005). Results presented here suggest this is not a concern in the Mount Cook region, given that no potential debris flow source areas were identified at elevations above 2500 m where permafrost is expected on steeper slopes (Allen et al., 2008a). Permafrost degradation in the Southern Alps is therefore most relevant concerning a possible role in destabilization of high elevation bedrock slopes (e.g., Cox and Allen, 2009). This is alarming given the potential for impact induced floodwaves as many proglacial lakes in the region continue to expand. Furthermore, the expected probability of a large earthquake (>M7) striking the region during the next 50 years may be as high as 35% (Cox and Barrell, 2007), providing a trigger for future large magnitude movements of ice and rock, and possible floodwave initiation.

7 Conclusions

GIS based procedures were applied to gain first order knowledge of glacial flood, debris flow, and ice avalanche potential across the Mount Cook region of New Zealand’s Southern Alps. Large volume lakes dammed by moraine and Quaternary gravels occur in proglacial areas below 1000 m where gentle slope gradients and well vegetated outlet channels indicate relatively stable conditions. Where smaller cirque lakes are dammed by steep morainic debris, debris flow scenarios were considered, but maximum probable runouts in these instances were well short of affecting any infrastructure. Potential debris flows were also modelled from glacial and recent paraglacial sediment accumulations, and in many instances runout paths intersected with human activity. However, examples illustrated near the Mount Cook village indicated a large discrepancy is possible between the static worst-case approach to runout modelling and maximum runout distances expected on the basis of catchment area available above the debris source. Direct impacts from potential high magnitude ice avalanches do not extend beyond walking tracks and some huts/shelters located in close proximity to steep ice, although numerous lakes are positioned within the runout paths from even smaller magnitude events, providing potential for displacement waves and flooding. The rapidly expanding Mueller Lake was illustrated as one example in close proximity to tourist and residential infrastructure, but longer term glacial recession and lake expansion will increase the potential for ice, debris or rock impacts into most lakes across the region. Physically based rock avalanche modelling with RAMMS provides a useful tool for recognizing detailed flow patterns, calculating potential flow magnitudes and velocities, and the application of this model for predictive purposes will benefit from further calibration studies. Further investigations are required at the local scale, assessing the likelihood of high magnitude flood, ice and debris hazards, and providing a comprehensive assessment of possible climate change impacts. Given an unstable geological setting and high potential for both seismic and rainfall triggered slope failures, a broader approach considering topographic, geological and glacial factors relating to bedrock failure susceptibility is encouraged. Uncertainty related to climate change and its influence on remote glacial regions should be met with robust methods for early recognition and monitoring of potential impacts.

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Abstract Rock avalanches fell from Vampire (2,645 m) Peak in the Southern Alps of New Zealand during January 2008. There were no direct witnesses, casualties or damage to infrastructure. Field observations indicate about 150,000 m\(^3\) (±50,000) of indurated greywacke collapsed retrogressively from a 73° slope between 2,380 and 2,520 m. Debris fell 800 m down Vampire’s south face and out 1.7 km across Mueller Glacier, with a 27.5° angle of reach. The resulting 300,000 m\(^2\) avalanche deposit contains three distinct lobes. The national seismograph network recorded two pulses of avalanche-type shaking, equivalent in amplitude to a M\(_{L}\) 2.4 tectonic earthquake, for 60 s on Monday 7 January at 2349 hours (NZDT); then 45 s of shaking at M\(_{L}\) 2.5 on Sunday 13 January at 0923 hours (NZDT). Deposit lobes are inferred to relate directly with shaking episodes. The avalanche fell across the debris from an older avalanche, which was also unwitnessed and fell from a different source on Vampire’s south face between February and November 2003. The 2003 avalanche involved 120,000 m\(^3\) (±40,000) of interlayered sandstone and mudstone which collapsed from a 65° slope between 2,440 and 2,560 m, then fell 890 m down across Mueller Glacier at a 24° angle of reach. Prolonged above-freezing temperatures were recorded during January 2008, but no direct trigger has been identified. The event appears to be a spontaneous, gravitationally induced, stress failure.