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The analysis of glacial retreat of selected mountain regions of New Zealand and natural hazards from GLOFs

Analýza ústupu zalednění vybraných oblastí Nového Zélandu a přírodní ohrožení z GLOFs

Diploma thesis

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Prohlášení:

Prohlašuji, že jsem závěrečnou práci zpracoval samostatně a že jsem uvedl všechny použité informační zdroje a literaturu. Tato práce ani její podstatná část nebyla předložena k získání jiného nebo stejného akademického titulu.

V Praze, 12.04.2017

Ondřej Brambus

Content

Abstract

This thesis analyses the glacier fluctuations in New Zealand since late Pliocen until today (2017) and evaluates GLOFs (Glacier Lake Outburst Floods) hazards from all proglacial lakes of New Zealand. Background research of a wide range of scientific sources was used to describe New Zealand glacier fluctuations during the last ~2.6 Ma, uncover local climatic and tectonic specifics, describe uneven behaviour of different glacier types and summarise current knowledge about climatological forcings to New Zealand glaciers. Compared to the timing of glaciations in the Norther Hemisphere, an earlier onset of LGM (Last Glacial Maximum) and LIA (Little Ice Age) was recorded in New Zealand. A dramatic glacier advance of short- to medium-response time glaciers was recorded between 1983 and 1999. This advance was caused by changes of atmospheric and oceanic circulation patterns around New Zealand.

A detailed study of the past events revealed that really few events were recorded in New Zealand history. While englacial outburst floods are relatively common from Franz Josef Glacier, no moraine dam rupture and only two GLOFs from a proglacial lake were recorded in New Zealand history. Inventory of proglacial lakes of New Zealand completed from remote sensing data was done to further evaluate the hazards of GLOFs. 25 proglacial lakes were located in two highly glaciated regions: Mt Aspiring area and Mt Cook area and their geomorphic properties were described.

New qualitative method assessing the GLOFs hazards was presented to capture local specifics. First, the possible GLOF triggers were evaluated, followed by the assessment of dam stability. The most hazardous lakes are Volta Lake in Mt Aspiring area, and La Perouse Lake and Lyell Lake in Mt Cook area. Risk assessment revealed there are no permanent settlements in the probable flood paths, but several roads, hiking tracks and backcountry huts or shelters are threatened by a potential GLOF. A further, more detailed assessment was recommended for the most hazardous lakes.

<u>Key words:</u>

Mountain glaciation, glacier retreat, natural hazards, GLOFs, climate change

Abstrakt

Tato diplomová práce analyzuje vývoj zalednění Nového Zélandu od období pozdního Pliocénu až do současnosti (2017) a hodnotí přírodní ohrožení z GLOFs (Glacier Lake Outburst Floods) pro všechna proglaciální jezera Nového Zélandu. Široké spektrum vědecké literatury bylo použito v rešeršní části k popisu vývoje zalednění během posledních ~2.6 Ma, k odhalení místních klimatických a tektonických specifik, k popisu rozdílného chování různých ledovcových typů a ke shrnutí dosavadního poznání o působení klimatu na ledovce Nového Zélandu. Nástup LGM (Last Glacial Maximum) a LIA (Little Ice Age) byl, v porovnání s průběhem zalednění na severní polokouli zaznamenán dříve. Mezi lety 1983 a 1999 byl zaznamenán dramatický postup ledovců s krátkým až středním časem odezvy. Tento postup byl způsoben změnami proudění v atmosféře a oceánu v oblasti Nového Zélandu.

Detailní studium historických GLOFs na Novém Zélandu odhalilo pouze několik událostí. Zatímco průvalové povodně englaciálního typu jsou poměrně běžné z ledovce Franz Josef, protržení morénové hráze nebylo v historii Nového Zélandu zaznamenáno. Přelití proglaciálního jezera bylo zaznamenáno pouze ve dvou případech. Za účelem pozouzení míry ohrožení z GLOFs byla dále pomocí dat DPZ provedena inventarizace všech proglaciálních jezer Nového Zélandu. Celkem bylo rozpoznáno 25 proglaciálních jezer ve dvou výrazně zaledněných oblastech. "oblast Mt Aspiring" a "oblast Mt Cook". Geomorfologické vlastnosti těchto jezer byly popsány.

Za účelem podchycení místních specifik byla vytvořena nová kvalitativní metoda posuzující míru ohrožení z GLOFs. Posouzení bylo provedeno ve dvou základních krocích: Posouzení příčin vzniku GLOF a posouzení stability hráze. Nejvíce ohroženými jezery jsou jezero Volta v oblasti Mt Aspiring a jezera La Perouse a Lyell v oblasti Mt Cook. Proces posouzení rizik neodhalil žádná trvalá obydlí v možné povodňové trase, avšak několik silnic, turistických cest, chat a přístřešků je ohroženo potenciální povodní typu GLOFs. Bylo navrženo provést následné, více detailní posouzení pro ta nejvíce ohrožená jezera.

<u>Klíčová slova:</u>

Horské zalednění, ústup zalednění, přírodní ohrožení, GLOFs, změna klimatu

1 Introduction

Human activities and lives have been linked with nature ever since. Recently we have been experiencing a dramatic change in the Earth's climate and environment, influencing our lives even more than before. It is certain that Global Mean Surface Temperature has increased since the late 19th century. Each of the past three decades has been successively warmer at the Earth’s surface than all the previous decades in the instrumental record, and the first decade of the 21st century has been the warmest (IPCC 2013). Rapid glacier retreat has been observed as a common phenomenon of the climate change all over the world. The amount of ice contained in glaciers globally has been declining every year for more than 20 years (IPCC 2013). The consequences are clearly visible in various spheres. Water scarcity, speed-up of global warming through albedo positive feedback, global sea level rise, mass movements, changes in agriculture and hydropower generation, and other consequences of retreating glaciers have been documented by scientists worldwide. One result of glacial retreat has been an increase in the number and size of glacial lakes forming at the glacier terminal ends behind the exposed end moraines. These in turn give rise to an increase in the potential threat of glacial lake outburst floods (GLOFs) occurring (e.g. Samjwal et al. 2007; Carrivick and Tweed 2013).

Glacier response to the climate is not simple. The surface temperature is not the only factor influencing glacier retreat and advance. Amount and form of precipitation, amount of incoming solar radiation (effected e.g. by slope aspect, or cloud cover), wind direction, wind speed, air humidity, topography and the size of a glacier are also important factors influencing the evolution of a glacier. Those parameters fluctuate on a long time (thousands and millions of years) and “short time” (mostly few years) scale. The best example of a long climatic cycle is the cycle of glacials and interglacials. The shorter cycles are rather known as the climate oscillations, like El Niño Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO) or many others. Those climatic cycles interact not only with each other, but also with seasonal climatic oscillations and – to a certain extent – with endogenous forces, especially volcanism and tectonic uplift.

New Zealand glaciers have been evolving in their own unique way. During last two decades of the 20th century, most New Zealand glaciers were expanding or in equilibrium with concurrent climate of that time (Chinn 2001) which does not correspond to the overall glacier recession trend in most mountain regions of the world. The unique geographic position of New Zealand and the complicated combination of climatic cycles create very special conditions for the “life“ of New Zealand glaciers. But, same as in the other parts of the world, New Zealand glaciers are not stagnant. Changes of the ice volume, glacier length, steepness of a glacier or composition of a glacier body are present even in New Zealand. In some cases, the changes are even the fastest in the world. Dramatic change of behaviour of the mighty Tasman glacier and the formation of – more than five kilometres long – Tasman Lake just in the last few decades are the best examples of the dynamic environment of New Zealand.

Changes in the ice volume of New Zealand glaciers and formation of new lakes pose a potential threat to humans and infrastructure. According to the IPCC (2013) and its assessment based on climate models, the global temperature will continue to rise during the 21st century. The increase in the global mean temperatures over the next one hundred years could range from 0.4 to 4.8ºC (depending on the climate model used and on the intervening greenhouse gas emission scenarios). The need for detailed scientific research and assessment of potential hazards related to glacier retreats and advances and especially hazards related to GLOFs (Glacier Lake Outburst Floods) is thus obvious. The goal of such researches is not only to understand nature processes better, but also to mitigate all the threats and protect people and property from potential disasters.

2 Topic of the thesis and objectives

The topic of this thesis is the analysis of glacial retreat of selected mountain regions of New Zealand and natural hazards from GLOFs (Glacial Lake Outburst Floods). This thesis aims to describe and quantify glacial retreat and evaluate hazards and risks related to GLOFs in New Zealand. There are two main parts of this thesis.

1) Background research:

The objectives of this section are

a) to describe the distribution of New Zealand glaciers and lakes (section 4.2 and 4.3), describe their basic geomorphic properties and formation.

b) to summarise knowledge and scientific researches that have been done about glacial retreats and advances during Quaternary with the emphasis on the last few centuries, and especially the last few decades (section 5.1)

c) to describe differences in behaviour between various New Zealand glaciers (section 5.2) and briefly describe the key climate factors influencing glacier fluctuations (section 5.3)

d) to describe selected glaciers into a more detail (section 5.4).

d) to summarise knowledge about recent GLOF events and their impact to the society (section 6).

The background research section aims to answer questions like:

- How has the mass of New Zealand glaciers been evolving?
- When has been the growth and the retreat of New Zealand glaciers the biggest?
- Has there been a spacial variability of glacial mass changes throughout New Zealand or have all New Zealand glaciers been responding in the same way?
- How many GLOFs events have happened in modern history of New Zealand?
- How serious were these events? How many deaths and how big damages were recorded and what was the response to these events?

2) Results:

Evaluation of the hazards related to GLOFs in New Zealand and evaluation of associated risks for the society are the main goals of the results section (section 7, 8 and 9). The key objectives developed to achieve this goal are:

a) to create an inventory of proglacial lakes in New Zealand and assess their relation to the glacier retreat (section 7),

b) to assess the susceptibility to GLOFs of all proglacial lakes according to a chosen method (section 8),

c) to create a list of the most potentially hazardous lakes in NZ (section 8) and

d) to evaluate the overall risks, that potential GLOFs can have for New Zealand society (section 9).

In this section I am trying to find answers to questions like:

- How many proglacial lakes there are in New Zealand and how many of them could be possibly hazardous in the near future?
- How do those lakes differ in origin, geomorphic characteristics, dynamics of evolution and potential hazards?
- Which lakes are the most hazardous?
- Which lakes pose the highest threat to the society?

3 Methods

While this thesis studies both the glacial retreat and the hazards related to GLOFs, wide range of methods needs to be used. Both sections encompass detailed background research, field works and the use of remote sensing data. Furthermore, for the evaluation of GLOFs hazards a detailed method assessing the susceptibility of glacier lakes to outburst floods was used.

3.1 Background research

Background research was done to summarise current knowledge about glacial retreat and GLOFs in New Zealand. The core of the background research was based on the studies of various scientific papers; especially geomorphologic, hydrologic or other scientific journals (see section 12 for full list of references). The research on glacial retreat is pretty much based on life-long work of Trevor Chinn – New Zealand leading glaciologist, but scientific studies of other authors and working groups were used as well. Online databases like the WGMS (World Glacier Monitoring Service) or databases by USGS were used to understand processes more broadly.

Due to the lack of scientific research related to GLOFs in New Zealand (see section 6) maximum effort has been done to find any information about recent and historical GLOF events. In the year 2014 GLOFs events were discussed in person with Professor Sean Fitzsimons and Dr Nicolas Cullen both from University of Otago, New Zealand. Many e-mails have been written to New Zealand scientists and organisations. The key organisations are NIWA (the National Institute of Water and Atmospheric Research), GNS (the Geological and Nuclear Sciences), and DoC (the Department of Conservation). While NIWA is the organisation to refer to regarding atmospheric processes and glacier research, GNS studies natural hazards and risks. DoC has a dense network of visitor centres that deal with every day issues of National Parks and mountain regions. Beside the management of New Zealand Backcountry, DoC cares about protection of the visitors visiting National Parks and other backcountry areas.

3.2 Field work

Selected lakes were personally visited between the years 2014-15 to briefly assess the landscape dynamics, current size of the lakes and glaciers and to observe details that are not clear from topographic maps or Google Earth images, like the slope and erodibility of lateral and terminal moraines, composition of the moraines and other properties. All of the nine glacial lakes bigger than 50 km^2 (see section 4.3) were visited and discussed with other geographers and geologists from University of Otago. From the young glacial lakes, e.g. Tasman Lake, Hooker Lake, Mueller Lake, Lyell Lake and Ramsay Lake were visited and assessed. Lake outlets were assessed for the erodibility and dam freeboard (vertical distance between the dam crest and water level). All the lakes were photographed.

3.3 Use of remote sensing data

Due to the remoteness of some parts of the Southern Alps, field studies are limited and remote sensing data need to be used. Especially west of the Main Divide temperate rainforest flourishes and makes some areas almost inaccessible. Steep topography and high erodibility of recently exposed moraines limit the field works on both sides of the Main Divide. From those reasons remote sensing images were used both for the assessment of glacial retreat and for the evaluation of GLOFs hazard.

Google Earth software was chosen as the source of the remote sensing images and also as a programme for successive analysis. It was concluded that the overall quality and resolution are sufficient for the analysis conducted in this thesis. The data availability, spacial coverage and user-friendly environment are some of the reasons for choosing Google Earth software. On the other hand the time-span of available images is relatively short and insufficient for long-term analysis of retreats and advances of a glacier. From this reason aerial photographs from 1986 supplied by the geography department of University of Otago were used for analysis of selected glaciers. Historical photographs were also studied in 3D using a stereoscope. ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) images were used for more detailed description of Tasman Glacier.

3.4 Methods used for glacial lake inventory

While there are more than 200 glacial lakes in New Zealand (Irwin 1975) – see section 4.3, the inventory and a detailed assessment of hazards related to GLOFs from all of those lakes would be extremely difficult and time-consuming. Due to different ages and types of glacial lakes, various methods would need to be used and probably even new methods introduced. Therefore a primary selection was done to filter-out potentially less hazardous glacial lakes and focus on the most active ones.

Various typologies of glacial lakes has been studied to select group of glacial lakes for the inventory and the detailed assessment. Six categories of modern glacial lakes by Embleton and King (1968) are relatively detailed, but do not well represent the regional characteristics of New Zealand glacial lakes. The typology used by Samjwal et al. (2007) for glacial lakes in Dudh-Koshi, Himalayas was concluded to be more transferable to New Zealand environment, however some categories were still missing in New Zealand and at the same time, by the selection of one category, some possibly dangerous glacial lakes would be excluded.

While the glacial lakes, which are still in contact or very near to the glaciers, are usually dangerous (Campbell et al. 2005), the broad category of proglacial lakes was selected. It was also concluded that the proglacial lake category is probably the most representative in terms of GLOF hazard assessment in the New Zealand context. Encompassing more categories from above mentioned typologies, proglacial lakes appear to show the highest susceptibility to GLOFs due to the high influence of adjacent glacier to lake processes. Due to the little knowledge about subglacial, and englacial lakes, and small extent and ephemerality of supraglacial lakes in New Zealand, only the category of proglacial lakes was chosen for the inventory and hazard assessment. The definition of Carrivick and Tweed (2013) was used to define a proglacial lake even the exact distinction between proglacial and "non-proglacial" lake can be discussed.

"*Proglacial lakes are masses of water impounded at the edge of a glacier or at the margin of an ice sheet... ...including ice contact or ice-marginal lakes, which are physically attached to an ice margin, as well as lakes detached from, or immediately beyond, a contemporary ice margin.*" (Carrivick and Tweed 2013).

According to above mentioned definition, the glacial lake inventory (section 7) and GLOFs hazard assessment (section 8) includes only: **a)** glacial lakes directly connected with a glacier and **b)** glacial lakes with a glacier directly on the slopes above them (lakes threatened by falling ice).

The selection of the criteria above is based on the significance of calving and icefall into a lake to growing hazard of a GLOF. If a proglacial lake forms on the terminus of a glacier, calving occurs and speeds-up the retreat (e.g. Chinn et al. 2012). Additionally calving blocks can cause a displacement wave and lead to a GLOF (e.g. Emmer and Vilímek 2014). Even the causes and mechanisms of GLOFs differ regionally, icefall to a lake is one of the key causes of GLOFs (Emmer and Vilímek 2014). While the inventory (section 7) and hazard assessment (section 8) focus only to one group of potentially hazardous glacial lakes (proglacial lakes) rather than all New Zealand glacial lakes (including the ones in ice-free catchments or old glacial lakes on the margin of the Southern Alps), the GLOFs hazards of the other glacial lakes are being discussed in the discussion section.

The search of New Zealand proglacial lakes was done using online topographic maps by DoC, however, some newly (the last 10-15 years) formed proglacial lakes were not marked in the map and therefore the satellite images from Google Earth were used as the key source for the New Zealand proglacial lake inventory. All the proglacial lakes were located and labelled in Google Earth software and briefly discussed in terms of GLOF hazards. Lake names were used from the topographical maps where available, or named after the main source glacier, valley, or after a stream flowing from the lake. Main source glaciers were detected and so were the main river valleys, to where the lake-water flows before reaching the sea or reaching a local erosion basis. Further, the drainage sea, approximate lake location, and altitude were recorded. While the horizontal location was derived from Google Earth software, the altitude was obtained from the online topographical maps by DoC. Where the spot height of a lake was given, this altitude was recorded, where it was missing, it was measured from 20 m contour lines. In the cases, where the lake was missing in the map at all, combination of Google Earth and topographical map was used to estimate lake level altitude. However, in the case of altitude, the topographical map was preferred due to higher accuracy (Google Earth software showed, in some places, up to 100 m elevation difference within the lake boundary).

To complete the inventory, further geomorphic properties were studied. The length of the lake was measured by ruler tool in Google Earth software. The length was measured as the longest line connecting two points on the lake border, derived from the most recent satellite image available. This value can differ from other length records in the literature or topographic maps. While the proglacial lakes do not have to be directly connected to a glacier (see definition above), the current glacier contact was studied. Results were derived from the latest satellite images provided by Google Earth and the value was described by either “YES” or “NO”. In the case where meltwater channels were visible between the glacier front and the lake, “NO” value was recorded, same as in the case of the presence of hanging glaciers above the lake. Dam type assessment was done by using Google Earth imagery, personal archive of photographs, photographs provided in Google Earth software (with caution) and topographic maps. The dams were classified according the typology of Carrivick and Tweed (2013). While a lake can be dammed by more than one dam type, the dominant one was recorded. In the cases, where two dam types were dominant, both of them were recorded.

3.5 Method used for GLOFs hazard evaluation

An effective GLOF hazard assessment is strongly influenced by the method used. 13 methods (qualitative, semi-quantitative, and quantitative) selected by Emmer and Vilímek (2013) were studied. However, most of those 13 methods deal just with moraine-dammed lakes and do not assess the lakes dammed by bedrock or by an outwash head. Due to the high percentage of proglacial lakes dammed by bedrock or outwash head (see section 7) none of those methods was used in its entirety for the assessment of GLOFs hazards in New Zealand. To reflect the local conditions, a new qualitative, first-order method has been presented.

While, the methods of lake and breach hazard assessment usually include two groups of parameters: the first considers the possibility of a triggering event while the second considers the dam stability (Emmer and Vilímek 2013), the method presented here applies this approach and studies the possible causes and possible flood magnitude (derived from lake and dam properties) separately, before assessing the flood propagation and the risks for society. The method presented is a qualitative, first-order assessment method, that was created as a synthesis of various methods applied to local conditions, in respect of available data. The qualitative probability classification of Huggel et al. (2004) was used and creates the core of this method. Huggel et al. (2004) classify the GLOF probability into three groups: Low (0 points), Medium (1 point), and High (2 points). Those points are assigned for all the lakes in trigger assessment (further described in section 3.5.1), and in the lake/dam stability assessment (further described in section 3.5.2). The points from both categories are further summed together to express the overall GLOF hazard for each lake (see section 8.4).

3.5.1 GLOF trigger evaluation

The possible causes of GLOFs were summarised according to Emmer and Vilímek (2013), and Richardson and Reynolds (2000) before the classification into six categories was done. Each category encompass a natural processes that can possibly cause an outburst flood (a trigger). Whether a process can threaten a lake was assessed in the next step. The severity was defined as a probability of the process to cause an outburst flood for each lake. The process of trigger determination and their hazard assessment is described in figure 1.

<u>Calving</u>

Calving of glaciers into glacial lakes increases the lake area and might generate a displacement wave (Richardson and Reynolds 2000), therefore the possibility of a GLOF from calving was assessed also in this study. This category does not include only calving from the glacier front, but also the release of ice blocks from ice-foot or from the bottom of a lake as described by e.g. Warren and Kirkbride (1998). For the lakes which are not in a direct contact with the lake, the probability is classified as being “Low (0)”, for the lakes with a direct contact with a glacier, but without a distinct (at least 5 meter high) frontal face “Medium (1)” value was recorded, and while the big floating ice tongues are considered to be highly dangerous (Richardson and Reynolds 2000), the lakes, where large glacier frontal faces touch upon the lake, were classified as having “High (2)” probability of a GLOF from calving.

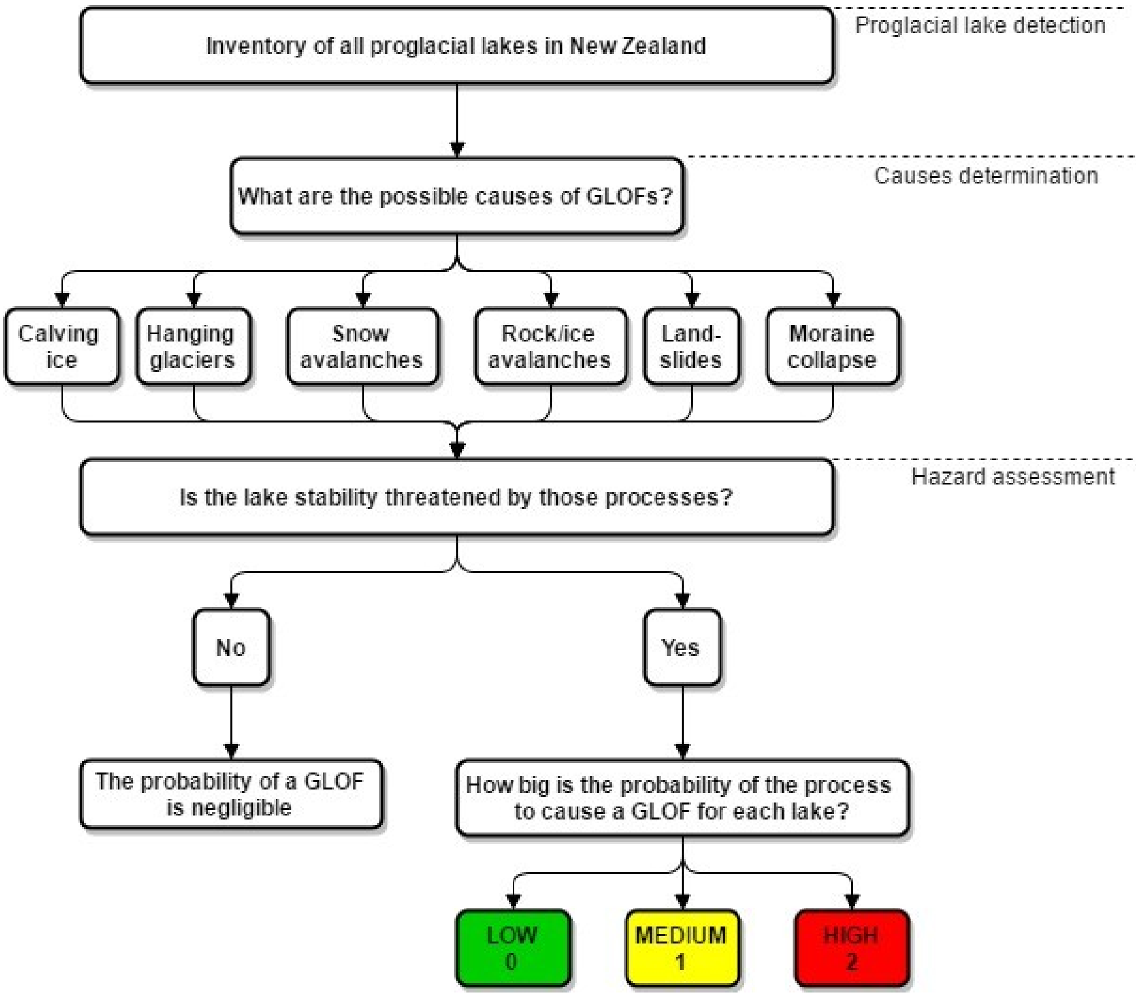

Figure 1: Schematic diagram showing the process of hazard assessment for various GLOF triggers for each lake. This diagram describes the procedure how to find an answer to the question: "Is the lake threatened by a process that can cause a GLOF, and how big is the probability?"

<u>Hanging glaciers</u>

Hanging glaciers above all New Zealand proglacial lakes were monitored, while they pose a significant threat to those lakes trough ice-avalanches. Ice avalanches occur when large masses of ice detach from steep cliff or ramp as frontal block failures, slab failures, or deeper failures at the ice/bedrock interface (Richardson and Reynolds 2000). Compare to calving, the ice avalanches can fall several hundred metres before reaching the lake and thus they can gain significant kinetic energy. The lakes with no hanging glaciers above them were classified as having "Low (0)" probability, where some smaller hanging glaciers were present above smaller cliffs, "Medium (1)" value was selected. "High (2)" probability of a GLOF from ice-avalanches from hanging glaciers was attributed to those lakes, which have large hanging glaciers directly above the lake, or close to them.

Snow avalanches

Most slab avalanches occur on slopes of 30-45°, with the maximum possible slope range of 15-60° (Goddard 2008). The surrounding of all proglacial lakes of New Zealand was thus searched for such slopes. The avalanche runout distance was estimated using the method of Goddard (2008). “Low (0)” probability class belong to the lakes that are not generally threatened by avalanches. “Medium (1)” value was given to all the lakes that are surrounded by smaller, and “not so steep” (generally less than 30°) slopes, or in the cases when the lake is protected from the slope by a moraine. All the lakes directly surrounded by high, steep slopes (estimated to have 15-60°) were classified As “High (2)”.

Rock/ice avalanches

Even the ice avalanches has been included in the “hanging glacier” category, it is necessary to include rock- and rock/ice- avalanches into the assessment process because the rock/ice avalanches have mostly much greater magnitude than ice avalanches from hanging glaciers (Allen et al. 2009) even the frequency is smaller. The prediction of rock/ice avalanches is a challenging scientific problem, however, some regularities have been discovered, leading to a better hazard assessment (e.g. Whitehouse 1983; Allen et al. 2008; 2009; 2011; Huggel et al. 2004; McSaveney 2002). A synthesis of those findings was done to asses the probability of a rock/ice avalanche to produce a GLOF. The most important factors are: Topography, fracturing, permafrost level, rock type, and geometric orientation of the slope towards the lake. Where a lake was surrounded by gently slopping slopes in relatively more stable rocks, in lower elevations and off the main axis with the slope, “Low (0)” value was assigned. Where the lake was surrounded by steep ice-covered, fractured slopes of relatively less stable rocks, and where those slopes were in alignment with the lake, “High (2)” value was assigned. Lakes showing signs of both groups were classified as “Medium (1)”.

Landslides

Landslide-derived outburst floods have been recorded worldwide (e. g. Korup 2005; RCEM 2014), and the magnitude can be comparable with some massive rock/ice avalanches (Allen et al. 2009). The probability of a landslide to cause a GLOF was assessed according to Allen et al. (2009; 2011). Factors like topography, orientation, geological unit, and runout path were considered and lakes classified to one of the three categories used above. Note, that landsliding from freshly uncovered unstable moraines was incorporated as well. However, due to the limited volume of slipping material, lakes threatened by loose moraine material only, were classified as “Medium (1)”.

Moraine collapse

While the moraine collapse is an important cause of a GLOF, the moraine collapse category was included in the trigger assessment even it differs from the other categories listed above. From historical records the failure mechanisms were recorded. The causes of moraine collapse include various mass-movements (as listed above), seepage through the moraine, melting ice-cores, settlement and/or piping within the moraine as a result of earthquake shocks, sudden glacial or meteoric drainage into the lake, and inappropriate engineering works during remediation (Richardson and Reynolds 2000). While all the mass-movements were studied separately, only the processes within the moraine itself were studied in “moraine collapse” category. It is clear, that the assessment of moraine stability can be assessed just for moraine-dammed lakes, thus where the lake is dammed by bedrock or outwash head, the probability of a GLOF from moraine collapse was classified as being “Low (0)”. “Medium (1)” and “High (2)” values are then assigned based on expert assessment of each moraine individually. However, to obtain a detailed hazard assessment of moraine-dam stability, field study, earthquake assessment and many more procedures need to be included. Due to the limited technical, financial, and time possibilities, those were not incorporated in this study.

Points count

After a hazard level assessment, all the “probability points” for each lake were counted. Even the processes have different magnitudes, and frequencies, all the categories have the same weight. However, the “probability points” directly reflect the hazard level. While the points range between 0 – 2, and there are six processes assessed (see figure 1), the total count can range between 0 – 12 points, where 0 signify low probability of a GLOF from all the triggers, and 12 signify that the probability of a GLOF from all the triggers is high.

3.5.2 Lake/dam stability evaluation

Even a significant geomorphic process can have marginal effects in terms of GLOFs. The amount of water retained in the lake, and the dam stability can strongly effect the nature and magnitude of the flood (e.g. Allen et al. 2009). Lake and dam properties were therefore assessed for all the proglacial lakes of New Zealand according the diagram in figure 2.

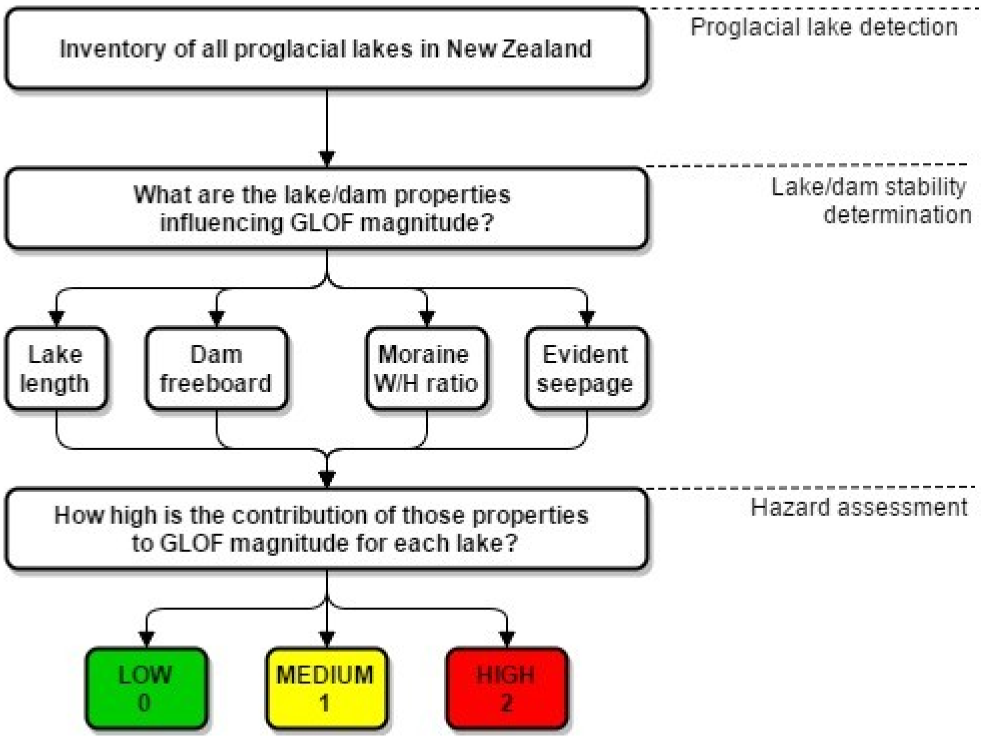

Figure 2: Diagram describing lake and dam stability characteristics effecting the GLOF magnitude.

The criteria assessed in this section reflect the data availability and serve as a first-order assessment, therefore more detailed characteristics need to be encompassed during following studies. While the size of the lake influences the amount of water available for a GLOF (Richardson and Reynolds 2000), lake area, or even better, lake volume needs to be used in the assessment process. While the lake volume is not known for many New Zealand proglacial lakes and area calculations are not available in Google Earth software, lake length (as described above) was used as the approximation of the lake area. The bigger the lake, the greater the amount of water for a GLOF available, thus the potential magnitude increases with the lake volume (lake length in this study). The smallest lakes (shorter than 0.5 km) were therefore classified as having “Low (0)”, medium-size lakes (0.51 – 1.50 km) as having “Medium (1)”, and the biggest lakes (longer than 1.51 km) as having “High (2)” probability to cause a severe

GLOF. However, the thresholds were set subjectively for the local conditions, based on expert assessment, and therefore (if applied elsewhere) needs to be re-adjusted according to local specifics.

While the distinction between different dam types is fundamental for related process discrimination and associated GLOFs probability determination (Huggel et al. 2004), three categories related to dam stability were incorporated in the lake/dam stability assessment. Those are **(1)** moraine dam freeboard, **(2)** Moraine width-to-height ratio, and **(3)** an evidence of seepage through the moraine dam. In the case of the lakes dammed by bedrock or outwash head, the possibility of a catastrophic dam failure is unlikely (Allen et al. 2009; Quincey and Glasser 2009), therefore "Low (0)" probability was assigned to those lakes for **(2)** and **(3)** characteristics.

Dam freeboard (the vertical distance between the lake level and the lowest point of the dam crest) needs to be considered for all the dam types, because the smaller the freeboard is, the bigger probability of a potential displacement wave to overtop the dam exists (Richardson and Reynolds 2000). No threshold was determined, therefore all the probability points in this category were assigned subjectively.

The width-to-height ratio and evident seepage is applicable just for moraine-dammed lakes, because this characteristic describes the dam stability in case of displacement wave propagation. While no moraine-dammed lakes were visited personally, Google Earth Software was used to obtain the results even the detail and accuracy are not sufficient for similar operations. However, the aim of this study is to give a first-order assessment, that enables further – more detailed studies.

3.6 Flood propagation and risk evaluation

The aim of this thesis incorporates also a brief assessment of physical and social vulnerability and thus examines the risks related to a potential flood wave propagation. While a numerical modelling was not included in this study, the assessment was based on visual investigations of the valley floor reaches in person, and interpretation of topographic maps and satellite images. Previous studies from New Zealand (McSaveney 2002; Allen et al. 2009) and other mountain regions worldwide (Westoby et al. 2014; Carrivick 2010) were used to estimate the area effected by a potential flood. Based on the results of McSaveney (2002), a displacement wave of ~10 m was assumed for this assessment process. Various geomorphic characteristics of the valleys bellow the proglacial lakes like overall slope, type of valley floor, width, retention potential and potential blockages were described and summarised.

The risk assessment was performed as an intersection between the estimated area effected and the settlement and infrastructure locations. Topographic map provided by DoC was searched for buildings, roads, railway tracks, bridges, 4x4 tracks, hiking trails, trekking huts, shelters and other man-made structures. The possible interaction between flood wave and settlements or infrastructure was assessed for each structure individually. All the structures effected by a flood-wave were then summarised.

4 Physical settings

4.1 The key drivers: Uplift and climate

New Zealand's diverse geography plays an important role in the evolution of glaciers and lakes of New Zealand. The location of the South Island on an active plate margin is one of the key factors influencing the evolution of New Zealand topography, glaciers and lakes. The collision between Indo-Australian plate and Pacific plate gives rise to ~500 km long mountain range; the Southern Alps. This plate boundary running through the South Island performs a complex movement combined of compression (Thrust faults) and transform (Strike-slip) movements, resulting in the rates of uplift among the fastest in the world (Lowe and Green 1987). At present the rate of rock uplift varies from up to 8–10 mm/year adjacent to the Alpine Fault to 0,8–1,0 mm/year along the southeastern margin of the Southern Alps (Tippett and Kamp 1995). In the geologic time scale, such enormous uplift rates result in a fast orogeny, thus leading to the change of local and regional climate and – between other consequences – to the formation of glaciers.

On top of that, the Southern Alps are placed almost perpendicularly to moisture-laden prevailing westerly winds, acting as a huge barrier causing whole range of geographic processes. Increased precipitation on wind-ward slopes and around the main divide and sharp decrease of precipitation on the lee side, or higher snow accumulation on lee slopes of higher elevations near the main divide are the main climatic consequences. Westerly conditions favour as well to a strong föhn effect. Extreme climate (together with extreme uplift rates) leads also to bigger erosion and sediment transport. New Zealand mountains supply a massive 209 million tonnes a year of sediment to ocean and shelf areas, which is among the highest anywhere in the world (Peat 2010). Strong winds, freeze-thaw processes, glaciers and glacial melt-water combine with the steep topography of quickly uplifted mountain range. It results in numerous landslides, rock avalanches, avalanches and debris flows. Alluvial fans, cliffs, eroded lateral moraines and huge outwash plains are the witnesses of those processes.

The uplift of the Southern Alps, the change of regional climate and the tendency to reach crustal isostatic equilibrium have huge effects on accumulation of snow, distribution and evolution of glaciers and lakes and of course weathering and erosion processes.

While the main factor for the evolution of New Zealand glaciers remains the climate, it is important to highlight some typical climate patterns of New Zealand. New Zealand is placed in the South Pacific Ocean about 1600 km east of Tasmania and 2600 km north of Antarctica in a maritime climate of the temperate zone (Cfb according to Köppen's classification). New Zealand's location within the Southern Oceans mid-latitudes (known as the „roaring forties") and the oceanic Sub-Tropical Convergence Zone (STCZ) makes it sensitive to changes in atmospheric and oceanic circulation (Vandergoes and Fitzsimons 2003). The climate of the Southern Alps is pretty much determined by strong westerly to south-westerly winds blowing in those latitudes due to global circulation patterns. The air has almost no landmass in its way to slow it down, leading to extreme wind speeds. The wind speed record of 250 km/h was recorded at Mt. John, Canterbury on 18 April 1970 (NIWA 2017). The Southern Ocean and the Tasman Sea west of New Zealand also supply the air with moisture which results in high precipitation on the wind-ward side of the Southern Alps. The highest precipitation recorded during one year (any 365 consecutive days) is 18 413 mm. This value was recorded on Cropp Creek weather station on the western side of the Southern Alps (NIWA 2017). It is important to mention, that the Tasman Sea plays an important role in New Zealand weather, because it is a "hotspot" for the generation of cyclonic storms, whose tracks are predominantly eastward. Low-pressure systems often develop south of mainland Australia as a result of the landmass warming rapidly in spring and the sea to the south being at its coldest then (Peat 2010).

However, the climate varies dramatically with the altitude and distance from the west coast. Figure 3 shows the climadiagram for Aoraki / Mt Cook Village located 765 metres above sea level just south of the New Zealand highest mountain Aoraki / Mt Cook, in the vicinity of many long glaciers and glacial lakes. Tasman, Hooker and Mueller glaciers

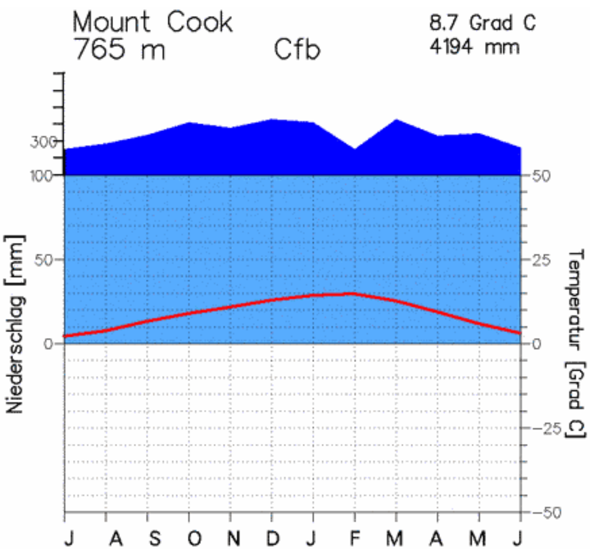

Figure 3: Climadiagram for Aoraki / Mount Cook Village just east of the main divide of the Southern Alps (lee side). Source: www.klimadiagramme.de

are some of the biggest ones. The village itself is placed just below the Holocene moraine of Mueller Glacier called the White Horse Hill. The moraine has been formed during many advances of the glacier in the last 600 years. The highest ridge on White Horse Hill and on the north side of the Mueller was built by a major advance achieving its maximum about A.D. 1740 (Burrows 1973).

The climadiagram shows the average year temperature of 8.7°C and annual precipitation of 4194 mm. However, those values change dramatically both with the altitude and with the distance from the main divide. Figure 4 shows the distribution of annual mean precipitation in New Zealand. We can see that the annual mean precipitation drop dramatically to the southeast of main divide, reaching values bellow 1 000 mm/year just about 50 km away. At the same time, the mean annual precipitation estimate for the summit elevations of the main divide are much higher than the recorded values from the official climatological stations. While the highest mean annual precipitation from the network of the official climatological stations reaches 6 715 mm/year at Milford Sound (NIWA 2017), near the main divide of the Southern Alps the mean precipitation can reach up to 14 000 mm/year (Henderson and Thompson 1999). The drier the climate, the colder the temperatures have to be to maintain a glacier, consequently glacier mean elevation rise from 1 500 m in the west to colder elevations over 2000 m on eastern glaciers (Chinn 2001).

The ratio between summer and winter precipitation is, in most cases, the key factor for assessment of the behaviour of a glacier. Precipitation values in the Southern Alps appear to be relatively balanced throughout the year. However there are no systematic measurements of seasonal snow in the Southern Alps, neither a complex information about its past variability (Fitzharris and Garr 1995). Thus the seasonal snowfall needs to be calculated from precipitation and temperature values.

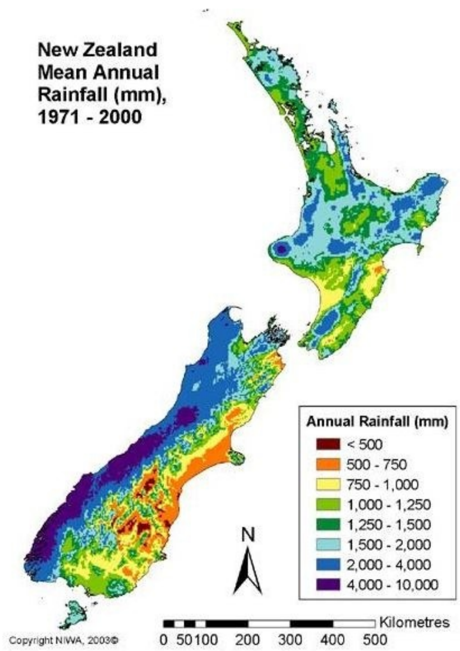

Figure 4: Annual mean precipitation in New Zealand. Note the sharp precipitation drop southeast of the main divide. Source: NIWA

Area-averaged annual snowfall maxima reconstructed from 1931 to 1993 between 1 000 and 2 200 m a.s.l. were calculated to be 366 mm. They show no trend, but large interannual variability from less than 200 to over 650 mm was discovered (Fitzharris and Garr 1995).

4.2 Distribution of New Zealand glaciers

A detailed inventory of the glaciers of New Zealand done by Trevor Chinn (2001) found a total of 3144 glaciers of over 1 ha in area, covering a total area of 1158 km^2. The estimated ice volume of 53.29 km^3 (year 1976 extent) is distributed from Mt. Ruapehu at 39° 15′ S to southern Fiordland at 45° 57′ S (Chinn 2001). While there are just small glaciers on Mt. Ruapehu (the North Island), in Kaikoura Ranges and north of Lewis pass, the Aoraki / Mt Cook and Mt Aspiring areas have many long valley glaciers and plateaus. Between Arthur's and Lewis Passes, none of the summits reach the regional snowline and there are no glaciers (Chinn 2001). The most glaciated area in New Zealand is the core of Aoraki / Mount Cook National Park. With its length more than 20 km, Tasman glacier is the longest glacier in New Zealand, but other valley glaciers (Hooker Gl., Murchison Gl., Darwin Gl. etc.) nest in this area. About 40 % of the total ice mass lies within the Tasman River-Lake Pukaki drainage basin (Gellatly et al. 1988).

On the western side near the main divide there are large glaciated plateaus (Albert Gl., Agassiz Gl.) from which valley glaciers of Fox and Franz Josef spill down to elevations about 400 m a.s.l. Fox and Franz Josef are the most accessible and most studied glaciers on the west from the Main Divide, however other valley glaciers do exist on the western side as well.

North of Aoraki / Mt Cook there are few smaller valley glaciers and two main low-elevated icefields called Garden of Eden and Garden of Allah. Lyell Glacier and Ramsay Glacier are the two main valley glaciers of this area. South of Aoraki / Mt Cook the biggest glaciated area is to be found in Mt Aspiring National Park. Lower and Upper Volta Glacier, Bonar Glacier and Olivine Ice Plateau are high-level valley glaciers. Between Mt Aspiring and Aoraki / Mt Cook, in the Landsborough valley, the McCardell and Dechen glaciers flow radially from the summit of Mt. Dechen in what may be termed an “ice cap”, the only one in New Zealand (Chinn 2001). The biggest glaciated area in Fiordland is to be found north of Milford Sound. Donne Glacier on the northern flanks of Mt. Tutoko is the biggest glacier in Fiordland. A more detailed description of glaciated areas of both islands is given by Gellatly et al. (1988).

New Zealand has comparatively large number of individual glaciers because the majority of the mountain peaks of the Southern Alps barely reach to above the regional snowline. The dryer climate east of the Main Divide results in a low ratio of snowfall to debris supply, therefore many groups of classical rock glaciers, both active and fossil, are to be found (Chinn 2001). However, the amount of debris on the western glaciers (19.3%) is only a little less than the 29.7% cover of the eastern glaciers (Chinn 1991 in Chinn 1996a). On the South Island the snowline rises from 1500 m in Fiordland to 2000 m at Arthur´s Pass. On the North Island the limit is about 2500 m on Mt. Ruapehu. Superimposed on this rise is a steep west-east gradient from 1500 m to 2200 m across the central area of the Southern Alps (Gellatly et al. 1988). However, the fronts of long valley glaciers reach well bellow 1000 m on both sides of the Main Divide. The eastern glaciers of Aoraki / Mt Cook National Park terminate at altitudes about 800 m, whereas the western glaciers at about 400 m well bellow local tree line. The snowline for glaciers with northern aspect is 300 to 320 m higher than the snowline for south-facing glaciers (Chinn 2001). In association with the strong precipitation gradient and asymmetrical uplift at the Alpine Fault (see section 4.1), the western glaciers are generally very steeply sloping, whereas the large valley glaciers east of the Main Divide are generally gentle sloping (Gellatly et al. 1988). Due to the position of the Southern Alps in the zone of interaction between the belts of subtropical highs and subpolar lows (see section 4.1) glaciers of the Southern Alps are highly sensitive to climate variations and act as monitors of atmospheric temperature because of rapid ice throughflow from accumulation to melting zones (Putnam et al. 2012).

Figure 5 shows the distribution of 3153 glaciers in the South Island of New Zealand (shaded areas) and names of some of them after the inventory done in 1991 (Chinn 1996a). With Aoraki / Mt Cook National Park as the most glaciated area standing out in the centre of the South Island.

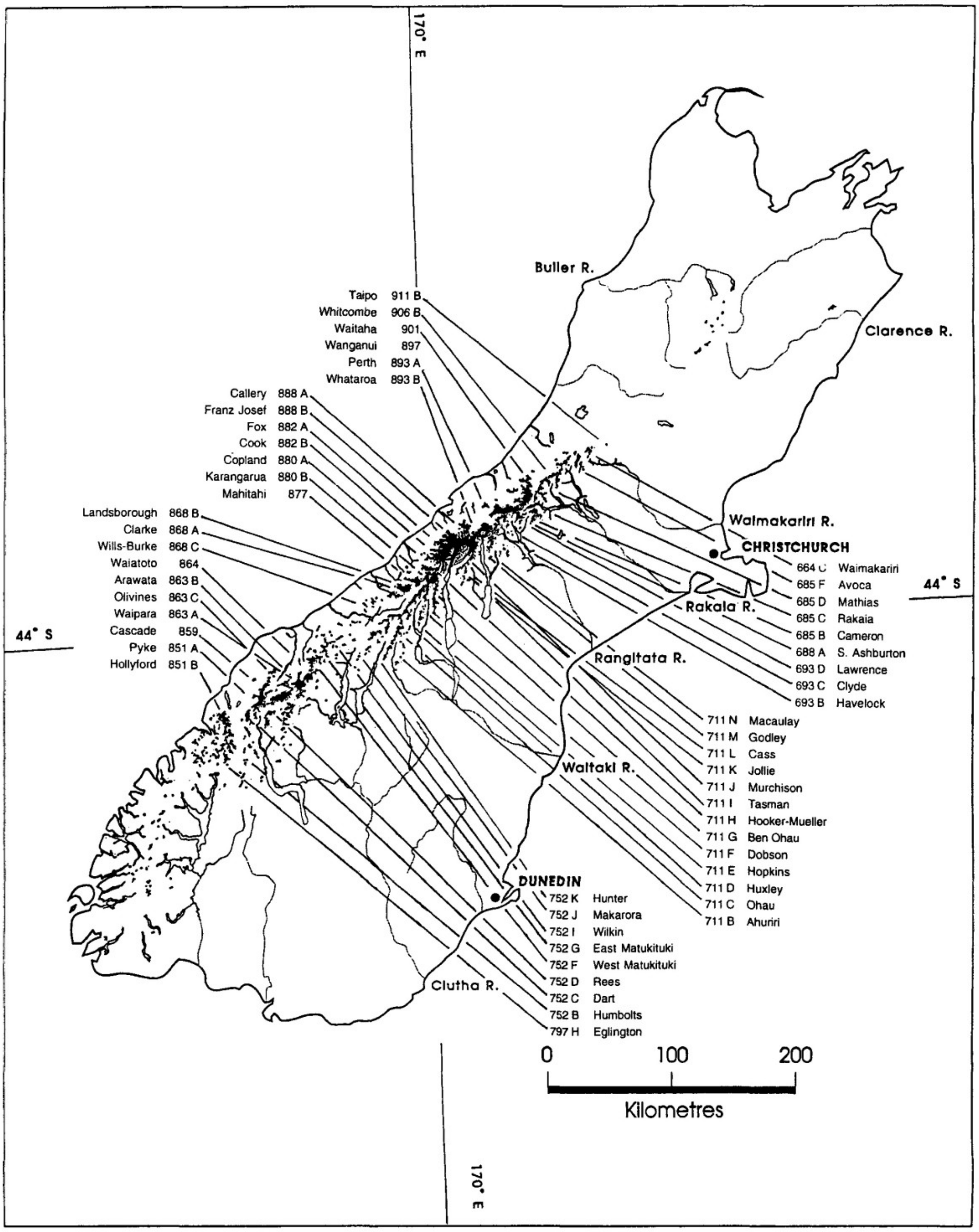

Figure 5: Distribution of selected glaciers of New Zealand as selected by Trevor Chinn (1996a). This map shows the main glaciated areas in New Zealand.

4.3 Distribution of New Zealand lakes

In New Zealand there are 3820 lakes larger than one hectare (McGeorge 2007). Irwin (1975) researched there are 294 lakes in the North Island and 475 in The South Island. Table 1 and table 2 show area frequencies of lake types for the North and South Island. While the predominant lake type in the North Island is the wind-formed, coastal and dune lake (106 out of 294), in the South Island it is the glacial lake (268 out of 475) that dominates (Irwin 1975). However, the bottom threshold for lake area is not given, thus the number of lakes smaller than 0.5 km^2 can vary according the methods used. Lowe and Green (1987) concluded there are 300/476 lakes in the North/South Island respectively (compared to 294/475 mentioned by Irwin (1975)). Even the number of glacial lakes differs according the authors. While Irwin (1975) described 268 glacial lakes (56.4% of all lakes in the South Island), Lowe and Green (1987) concluded the number of glacial lakes is 291 (61.1%).

According to Irwin (1975), of the 268 glacial lakes, most are small in area (172 smaller than 0.5 km^2) and many lie at high altitudes in the southwest of the South Island. But it is important to realise, that the life of glacial lakes is often short. Small glacial lakes are being created between the glacier terminus and its moraines as it retreats. Also the disappearance of glacial lakes is a common phenomenon. Either due to slow seepage through the moraine, sudden release of the water or sedimentation processes.

Figure 6 shows the distribution of New Zealand lakes according to their type. We can see that glacial lakes occur just in the South Island of New Zealand and pretty much correspond with the position of the Southern Alps. It is the result of orography and climate of the South Island. Isolated volcanoes in the North Island do not have sufficient altitude and area to produce large glaciers in current climate, so the glacial lakes are missing in the North Island at all.

During Pleistocene period huge mountain glaciers occupied vast areas around the main divide of the Southern Alps of New Zealand leaving big troughs (glacial valleys) and terminal moraines behind (e.g. Burrows et al. 1976). Many of those valleys were filled by water, leading to formation of glacial lakes, from which nine are bigger than 50 km^2 (Irwin 1975). Those are Lake Te Anau, Lake Manapouri and Lake Hauroko in Southland region, Lake Wakatipu, Lake Wanaka, and Lake Hawea in Otago region, and finally Lake Pukaki, Lake Tekapo and Lake Ohau in Canterbury region.

Lake Type	Area (km^2)					Totals
	<0.5	0.5 - 5	5 - 50	50 - 500	500 - 5000	
Tectonic	2	1		1		4
Volcanic	10	8	9	1	1	29
Landslide	1	1		1		3
Swamp	16	3				19
River	43	13	3			59
Wind	88	18				106
Bar	9	2	1			12
Dam	19	7	3			29
Glacial						
Not determined	29	4				33
Totals	217	57	16	3	1	294

Table 1: Area frequency of lake types, North Island. Source: Irwin (1975)

Lake Type	Area (km^2)					Totals
	<0.5	0.5 - 5	5 - 50	50 - 500	500 - 5000	
Tectonic						-
Volcanic						-
Landslide	5	4				9
Swamp	16	2				18
River	46	15	3			64
Wind	13					13
Bar	13	2	2	1		18
Dam	17	9	3	1		30
Glacial	172	74	13	9		268
Not determined	38	16	1			55
Totals	320	122	22	11	-	475

Table 2: Area frequency of lake types, South Island. Source: Irwin (1975)

Most of the larger glacial lakes in the South Island occupy overdeepened, formerly ice-filled valleys (Lowe and Green 1987). However, their direct connection to a glacier was lost with the onset of Holocene. Above mentioned large glacial lakes dammed by moraines from Pleistocene, or Pleistocene-Holocene boundary appear to have different hydrologic and geomorphologic processes than the younger lakes currently connected with glaciers. Both the moraines and surrounding slopes have been stabilised by wide range of natural processes. Although landslides, rockfalls and avalanches can impact the lakes, icefall from glacier fronts and calving is not present. Due to higher moraine-dam stability and higher level of alteration by humans (dam strengthening, usage of the water for power generation...), GLOFs research worldwide focuses mostly on active glacial lakes directly influenced by glaciers.

The distribution of active glacial lakes is bound with long valley glaciers and cirques in higher altitudes. Most glacial lakes of the South Island of New Zealand lie at altitudes over 610 m a.s.l. (Irwin 1975). While steep granitic mountains of Fiordland and some areas of northwest Nelson favour the presence of small tarns and lakes nested in hollowed floors of cirques, the central Southern Alps are home for many moraine dammed lakes (Lowe and Green 1987).

Figure 6: Map showing the distribution of the main types of lake in New Zealand. The variety of lake types is due to different geological processes. Note that glacial lakes are to be found solely around the axis of the Southern Alps due to presence of glaciers nowadays and in the past. Source: Lowe and Green 1992.

It is important to realize that most of the glacial lakes of New Zealand have originated through more than one mechanism, and many have been modified by non-glacial processes such as alluvial aggradation, faulting, landsliding or coastal processes since their original formation. Usually such lakes may still be classed as glacial because they occupy basins excavated or modified by ice and dammed by drift (moraine) or glacifluvial outwash deposits that originated as a direct consequence of the glacial activity (Lowe and Green 1987).

5 Glacial retreat in New Zealand

5.1 New Zealand glacier history

5.1.1 Glacier changes during late Pliocene and Pleistocene

Fundamental changes to the Earth's climate system and associated biotic response occurred about 2.6 Ma (Gibbard and Head 2010). These fundamental changes have been documented worldwide making this episode the logical start of the Quaternary and the border between Pliocene and Pleistocene epochs (Cohen et al. 2013). Dramatic cooling has been documented on many places, seesaw between colder glacials and warmer interglacials commenced and so did the record of New Zealand glaciation (e.g. Suggate 1990; Chinn 1996b).

The earliest definitive glacial deposits is till of Ross Glaciation found at Ross in Westland. Till layers lie conformably on Pliocene marine beds and are overlain by marine sandstone and conglomerate of late Pliocene. The estimated age of 2.6-2.4 Ma well correspond with the abrupt appearance of subantarctic taxa showing evidence of cooling of the seas in New Zealand (Chinn 1996b). Following the New Zealand late Pliocene glacial events, a hiatus of over one million years occurs (Suggate 1990) where the combined effects of uplift and erosion have almost entirely removed the terrestrial record from ~2.1 Ma until the last 0.35 Ma (Chinn 1996b). Table 3 shows the chronostratigraphic events of New Zealand for the last 2.5 Ma. Both the Ross Glaciation and the gap of glacial deposits during early and mid Pleistocene are obvious.

The mid to Late Pleistocene glaciations took place in a setting substantially similar to that of the present day, where valley glaciers characterise all glaciated catchments. Water was plentiful, enabling enormous fluvioglacial deposits like the Canterbury Plains, rather than periglacial features, to be formed (Chinn 1996b). Deposits of four glaciations commencing at Marine Isotope Stage (MIS) 10 (approx. 410 Ma) have been identified (see Tab. 3). It is believed that the climatic conditions of all glacials during late Pleistocene were basically similar to the last, Otiran Glaciation, when temperature depressions of between 4° and 6° have been estimated (Chinn 1996b).

Figure 7 shows the position of terminal moraines of the last glaciation maximum (LGM) and moraines from the advances near the Pleistocene-Holocene boundary. The “big” lakes mentioned in section 4.3 are clearly visible as well. However the moraines and their outwash surfaces are characteristic mainly for the glaciation east of the Main Divide. In the Westland the glacial record becomes more complex. In the northern sector, uplift has preserved a sequence of interglacial shorelines cut into the glacial outwash plains serving as a base for dating of late Pleistocene events (Chinn 1996b). While the older New Zealand events have been dated by association with fluctuating sea levels and sequence counts, younger New Zealand events, like the Last Glacial Maximum (LGM) are relatively well constrained by radiometric dates (Chinn 1996b). However, due to steep nature of the country and high erosion rates, little evidence has survived from glaciations earlier than the late Otiran (Gellatly et al. 1988).

AGE Ka	$O^{16/18}$ Stage	WORLD EVENTS	EVENTS	STAGES
0 6 12	0 1a 2	*Interglacial*	ARANUIAN	*Waiho
14	2	Glacial	OTIRAN	*Kumara 3 †Poulter ‡Acheron •Hawea
				Minor Interval
16.5 15.3				*Kumara 22 †Blackwater 2 ‡Bayfield 3
18	2b			†Blackwater 1 ‡Bayfield 1,2 •Mt Iron
27	3			*Important Interval*
28 53 60	3a 3c 4			*Kumara 21 †Otarama ‡Tui Ck •Albert Town
70 80 90 100 110 120	5 5a 5b 5c 5d 5e	*Interglacial*	*KAIHINU*	
130 185	6 6f	Glacial	WAIMEAN	*Kumara 1 †Woodstock ‡Woodlands •Luggate
210	7	*Interglacial*	*KARORO*	
280 300	8	Glacial	WAIMAUNGAN	*Hohonu †Avoca
310 341	9	*Interglacial*	Un-named	
340 310	10	Glacial	NEMONAN	*Cockeye Fm ‡Hororata Fm
410 - 445	11	*Interglacial*		
445 - 510	12	Glacial		•Lindis ?
510 - 540	13	*Interglacial*		
540 - 550	14	Glacial		•Lowburn ?
550 - 610	15	*Interglacial*		
610 - 655	16	Glacial		•Lowburn ?
655 - 680	17	*Interglacial*		
680 - 710	18	Glacial		
710 - 730	19	*Interglacial*		
730 - 760	20	Glacial		
760 - 785	21	*Interglacial*		
785 - 900	22	Glacial		
1 - 1.5Ma		Cool		•Northburn
423 Ka > 1Ma				
2.0M			*PORIKA (2.1 - 2.2 Ma)	
2.5M		Worldwide cooling 1.8-2.5 Ma	*ROSS (2.4 - 2.6 Ma)	

Table 3: Chronostratigraphic table of New Zealand during the last 2.5 Ma. Source: Chinn 1996b.

5.1.2 Glacier changes during LGM

In the Southern Alps of New Zealand, glaciers advanced to their LGM limits earlier than the terrestrial (ice-sheet) glaciers in the Northern Hemisphere (William et al. 2015). While Suggate (1990) places New Zealand LGM between 22.3 and 18 ka ^{14}C yr BP (approximately 27-21.7 calendar years BP), William et al. (2015) suggests the LGM advance occurred in late MIS 3, between 31 and 29 ka. The glacial onset was abrupt

Figure 7: Positions of moraines of the last glaciation maximum and other advances near the Pleistocene-Holocene boundary. Main South Island rivers and lakes are shown and labelled. Dashed line marks the Main Divide. Source: Burrows at al. 1976.

with Te Anau Glacier in Fiordland (one of the largest in the Southern Alps of that time) taking only about 2 000 years to achieve its maximum LGM depth (William et al. 2015).

This is probably thanks to wet and cold conditions during the LGM onset. Stable isotope data suggest that the climate then became drier, while remaining cold, and the glaciers progressively ablated. While the surface of Te Anau Glacier had lowered by more than 600 m by ca 18 ka (William et al. 2015), the ice tongue of Lake Ohau Glacier had receded at a mean net rate of about 77 m/yr, lost about 40% of its overall length, and its surface lowered by 200 m between 17 690 and 17 380 ka BP (Putnam et al.

2013). Te Anau and Lake Ohau glaciers are not the only examples. During the LGM a complex system of expanded valley and piedmont glaciers that extended 700 km along the Southern Alps was formed, and averaged 100 km in width (New Zealand Geological Survey 1973). That resulted in the formation of huge terminal moraines damming the biggest glacial lakes of present days (see Fig. 7). Pleistocene glaciers were coalescing in the main mountain valleys to reach the sea in the west and extending through the foothills to the outwash plains in the east (Gellatly et al. 1988) and filling the massive fiords in the southwest (William et al. 2015).

It is important to note that the culmination of LGM in New Zealand may have occurred few thousand years earlier than the terrestrial (ice-sheet) culmination of LGM in the Northern Hemisphere (William et al. 2015). The authors believe it can be possibly due to the faster climate response of alpine glaciers compare to continental ice sheets. The culmination of the LGM in New Zealand occurred in the late MIS 3, about 12 000 years before global ice volume reached its peak. By the time global ice volume was at its maximum (19 ka) glaciers had almost disappeared from major Fiordland valleys in New Zealand (William et al. 2015) and other mountainous areas (Putnam et al. 2013). New Zealand occurred in Last Glacial–Interglacial Transition (LGIT).

5.1.3 Glacier changes during LGIT and early Holocene

The dramatic glacial retreat related to the Last Glacial–Interglacial Transition (LGIT) was interrupted by at least four more advances between 18 and 8 ^{14}C ka BP (Fitzsimons 1997) (approx. 21.7-8.8 cal. ka BP), before the interglacial (Holocene) climatic optimum occurred (Chinn 1996b). A widespread glacier advance, possibly with two maxima, occurred between 16 and 14 ^{14}C ka BP, during which glaciers reached positions close to those attained at the last glacial maximum (Fitzsimons 1997). However, the glacial record differs regionally throughout New Zealand and a different number of advances has been described in different regions. The best preserved evidence for the expansion of Late Quaternary glaciers comes from North Westland (Fitzsimons 1997) and is shown in Figure 8.

One of the best evidence of such an ice advance is provided in south Westland by the Waiho Loop moraine (Vandergoes and Fitzsimons 2003). Waiho Loop moraine (Figure 9) is a tree-covered semicircle of moraine debris about 80 metres high clearly notable on areal photographs. The moraine indicates the extent of the Franz Josef

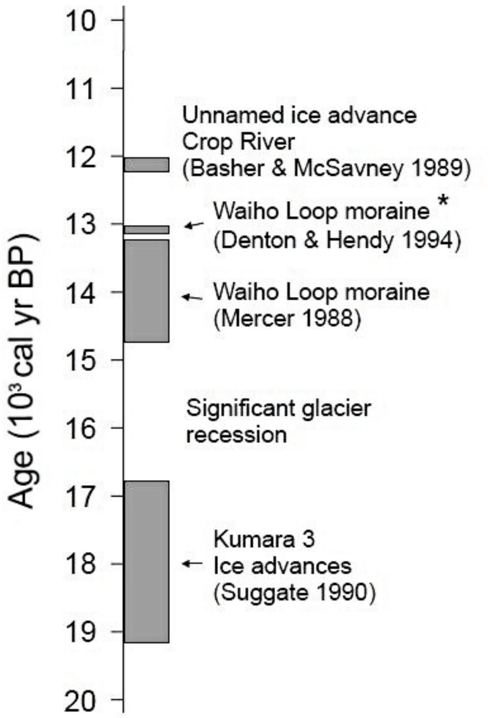

Figure 8: Late-glacial and early Holocene ice advances in north Westland. Edited according to Vandergoes and Fitzsimons 2003.

Glacier, about 12 000 years ago (McSaveney 2007). By studying large moraines in New Zealand, one has to wonder why no equivalent moraines are known elsewhere in Westland. The regional pattern for this important advance has yet to be established (Fitzsimons 1997).

The warming during LGIT is notable also in the biotic response. Pollen record and the radiocarbon AMS (Accelerator Mass Spectrometry) ages show the replacement of grassland by shrubland already prior to ca 18 300 yr BP (15,300 ^{14}C yr BP) followed by the progressive development of broadleaf shrubland and scattered forest prior to ca 14 400 yr BP, indicating that deglaciation commenced in south Westland prior to ca 18 300 yr BP. (Vandergoes and Fitzsimons 2003). The timing of deglaciation and temperature increase prior to ~18,300 yr BP in south Westland is also broadly consistent with pollen and geomorphologic records from similar southern latitudes like Auckland, New Zealand (Sandiford et al. in Vandergoes and Fitzsimons 2003), and central Tasmania (Hopf et al. 2000 in Vandergoes and Fitzsimons 2003) and with other more recent studies (e.g. Williams et al. 2015).

Vegetation changes between ca 14 400 and 11 400 yr BP indicate a period of increased precipitation. Tall podocarp forest, similar to that of the contemporary forest, was well

Figure 9: Waiho Loop moraine in the south Westland. Unlike the surrounding river flats, it was too rugged to be cleared for farming and remains covered in native forest. The braided Waiho River is at right. Source: McSaveney 2007

established by ca 11 400 yr BP. A period of increased westerly circulation over southern New Zealand is considered to be the most likely mechanism for causing the increased precipitation. Increased precipitation would also provide a mechanism for initiating LGIT ice advance in the area (Vandergoes and Fitzsimons 2003).

In accordance to the rapid warming at 14 ^{14}C ka and the demise of the huge Pleistocene glaciers, this age is recognized as the boundary between Otira Glaciation and the Aranui Interglacial (e.g. Chinn 1996b; Fitzsimons 1997).

5.1.4 Holocene glacier changes until the "Little Ice Age"

Throughout the Southern Alps, glaciers contracted rapidly to a small fraction of their Pleistocene extents between 12 and 9 ka but with a number of dated readvances (Chinn 1996b). Those were much smaller than the advances during Pleistocene or the LGIT. A bigger glacier advance and two smaller ice advances may have occurred around 11.0, 10.25 and 8.6 ka BP (Fitzsimons 1997). But the glaciers – during entire Holocene – never regained more than a third of the former Pleistocene maximum ice extent (Gellatly et al. 1988). The period between 8 and 5 ka represents interglacial climatic optimum of the Holocene, however, several moraines (ice advances) have been dated in the period 9 to 7.6 ka BP (Chinn 1996b).

New Zealand neoglacial activity commenced at 5 ka BP and continued to the present, with numerous well-documented glacial fluctuations. Nested moraines indicate that the advances diminished in size only slightly towards the present (Chinn 1996b). Also the vegetational history showed that shortly after 5 ka BP the climate of New Zealand experienced an abrupt change to cooler and drier conditions (Gellatly et al. 1988). Onset of the neoglacial ice advance is time-consistent with many other world mountain regions (Solomina et al. 2015). Although regional patterns of Late-glacial advances are reasonably well established, the Post-glacial advances in New Zealand are known only from single sites and require closer investigation (Fitzsimons 1997). Correlation between separate valleys yet need to be done to understand Holocene glaciation of Southern Alps in its entirety.

Several dating methods has been used by various authors to create the chronology of Holocene glacial history. Ricker et al. (1993) used the method of weathering-rind dating, to investigate the chronology of several moraines of the Craigieburn Range. The first neoglacial moraine was dated to 2.8 to 4.2 ka (Arthur´s

Pass advance) but no moraines dated for the interval 500 B.C. - A.D. 1100. However, there were middle Neoglacial moraines of the latter part of O'Malley time (ca. A.D. 1170 - 1330). The youngest (early Barker) moraines at Craigieburn Range were dated as A.D. 1460 + 100 (Ricker et al. 1993).

Schaefer et al. (2009) identified more than 15 advances of Mueller, Hooker and Tasman glaciers in the last 7 ka, including at least five events during the last millennium: 6.5 ka, 3.6 to 3.2 ka, 2.3 ka, 2.0-1.6 ka (at least three events), 1.4 ka, 1.0 ka, 0.8 ka, 0.6 ka, 0.4 ka (at least two events) and 0.27-0.11 ka. An early study by McGregor (1976) used radiocarbon dating technics and identified four periods of Holocene glacial activity in Ben Ohau Range, Canterbury. In subsequent work, Birkeland (1982) recognised a fifth event and provided ages for each of McGregor (1967) depositional phases. Those phases were dated to 9; 4 and 3 ka BP and 250 and 100 BP. Gellatly et al. (1988) also used radiocarbon method and reported glacial activity around 5; 4.5-4.2; 3.7; 3.5-3; 2.7; 1.8-1.7; 1.5 and 1.1 ka BP. However Solomina et al. (2015) states that Gellatly et al. (1988) summarised the results of radiocarbon dating of fossil wood and soil within the lateral moraines and identified glacier advances at ca 9.0-8.7 ka, 5.7-5.7 ka, 5.3-4.6 ka, 4.1-4.0 ka, 3.8-3.1 ka, 2.8-2.1 ka, 1.8-1.5 ka, 1.4-1.3 ka, 1.0-0.96 ka, 0.91-0.76 ka, 0.67-0.5 ka and 0.4-0.1 ka (Solomina et al. 2015) even though those dates were not mentioned by Gellatly et al. (1988). Inconsistencies in the dating of glacier advances occur also between other studies.

While the Maori people did not leave any written evidence about glacier fluctuations and the European settlement has not started before A. D. 1769 (when James Cook first explored the islands), the Holocene glacial record is mostly based on radiocarbon dating (e.g. Gellatlly et al. 1988; McGregor 1976), weathering-rind dating (e.g. Ricker et al. 1993), or ^{10}Be surface-exposure dating (e.g. Putnam et al. 2012; 2013). But even after A. D. 1769 settlers have not started with glacier studies. Early studies commenced in second half of nineteenth century shortly after the exploration of the Southern Alps (Fitzharris et al. 1992). This period is known as the culmination of the “Little Ice Age” (LIA) – the period of cooler climate with a significant glacier advance of most mountain glaciers in the European Alps (Mann 2002).

5.1.5 Glacier changes since the "Little Ice Age"

While the term "Little Ice Age" describes generally cooler period between 16th and 19th century that lead to a significant glacier advance in Europe, decreased temperatures and glacier advances have been documented also from other mountain regions of the world. However, the timing and nature of those variations are highly variable from region to region (Mann 2002). In New Zealand several LIA ice advances have been documented as well. While Putnam et al. (2012) suggests three small ice advances at 1770, 1864 and 1930 A. D., Schaefer et al. (2009) suggests one advance between A. D. 1680 and 1840. Purdie et al. (2014) described several ice advances of Franz Josef and Fox glaciers during the period ~1600 to 1800 AD and whole range of readvances during the 20th and 21th century, including the biggest one between 1983 and 1999. Chinn (1996a) suggests that the glaciers of New Zealand begun to retreat persistently between about A.D. 1750 and 1890 (at different times on different glaciers, and at different rates) and since then New Zealand glaciers have retreated dramatically. Chinn (1996a) also concluded that since 1978 there has been in most years a positive mass balance of most New Zealand glaciers and states: "Currently, all except a few glacier fronts are thickening and advancing". Glacier advance in a generally warming climate is not unique to the highly responsive New Zealand glaciers. A number of maritime glaciers in Norway have also advanced during the 20th century and advance has been recorded at a small number of Patagonian and Alaskan glaciers (Purdie et al. 2014).

Despite several ice advances, the overall recession trend during the 20th and 21th century is obvious. The present (2014) retreat of Franz Josef and Fox is the fastest retreat in the records of both glaciers (Purdie et al. 2014). Putnam et al. (2012) suggests an average retreat rate of 17.5 m/yr for Cameron Glacier between A. D. 1930 and A. D. 2006. Quincey and Glasser (2009) described retreat rates of over 25 m/yr for the Tasman Glacier. The volume of ice in the Southern Alps has decreased from 54.5 km^3 in 1976 to 46.1 km^3 by 2008. This equates to a rate of −0.3 km^3 a^{-1} over the last three decades, but this is considerably less than the rate of ice volume loss estimated for the previous 100 years (Chinn et al. 2012).

If we look into more detail to the historical ice advances, neither of the advances in 18th, 19th, nor 20th century have been so significant and persistent to be generally recognized as the LIA advances. While Wardle (1973) states that the advances and retreats of the last four centuries constitute a distinct glacial episode culminating between 1600 and 1850 A.D., and thereby equivalent to the "Little Ice Age" of the Northern Hemisphere, Schaefer et al. 2009 suggests that since mid-Holocene the fronts of large eastern valley glaciers of Mount Cook fluctuated just slightly around their mid- to late 19th century position, with a decrease in amplitude during the last millennium. The summer temperature time series based on tree-ring data from the nearby Oroko Swamp shows similarly the coldest period during the past 1100 years around 1000 A. D. (Cook et al. 2002). However the most prominent moraine of the past millennium at Mueller Glacier is about 570 years old (Schaefer et al. 2009) indicating the maximum glacier extent before A. D. 1380. Hence the LIA can not be applied globally in its entirety and the term LIA needs to be used with caution.

5.1.6 Comparison with global glacier fluctuations

While studying the New Zealand glacial history, one can ask: "Were Holocene glacier advances in New Zealand generally coeval with those in the Northern Hemisphere?" As described in section 5.1.5 the largest advances of glaciers in New Zealand occurred in the early Holocene, whereas the LIA advances were more restricted (Solomina et al. 2015). This contrasts with the glacier behaviour of most regions in the Northern Hemisphere (e.g. Mann 2002). Schaefer et al. (2009) compared the dating of New Zealand Holocene moraines with the record from North Hemisphere and found three main conclusions (see figure 10). First, it was concluded there is a notable interhemispheric disparity in the timing of the maximum ice extent. The Mount Cook glaciers were further advanced about 6500 years ago than at any subsequent time. In contrast, most Northern Hemisphere glaciers reached their greatest Holocene extents during the LIA (1300 to 1860 A. D.). Second, several glacier advances beyond the extent of the 19th century termini occurred in New Zealand during northern warm periods characterised by diminished or even smaller-than today northern glaciers, such as between 7500 and 5500 years ago in the Swiss Alps and Scandinavia, during the Bronze Age Optimum, during the Roman Age Optimum and other. Third, the greatest coherency between the Mount Cook and Northern Hemisphere records was during the Dark Ages (300 A .D. to 700 A. D.), and broad similarities were apparent during the past 700 years (the northern LIA), with multiple glacier advances followed by a general termination commencing in the mid- to late 19th century (Schaefer et al. 2009).

Schaefer et al. (2009) concluded that mid- to late Holocene glacier fluctuations were neither in phase nor strictly antiphased between the hemispheres, and therefore it is likely that regional driving or amplifying mechanisms have been an important influence on climate. However, climate controls of glacier behaviour are still being studied (e.g. Fitzharris et al. 1992; Putnam et al. 2012; Chinn et al. 2012) and are described in section 5.3.

Inconsistency between New Zealand and European glacial fluctuations persisted even through 20th and 21th centuries. Contrary to world-wide mass balance trends, the New Zealand data have displayed a cumulative mass balance that is near zero for the last three decades (WGMS 2008). Nevertheless, inter-annual oscillations of both

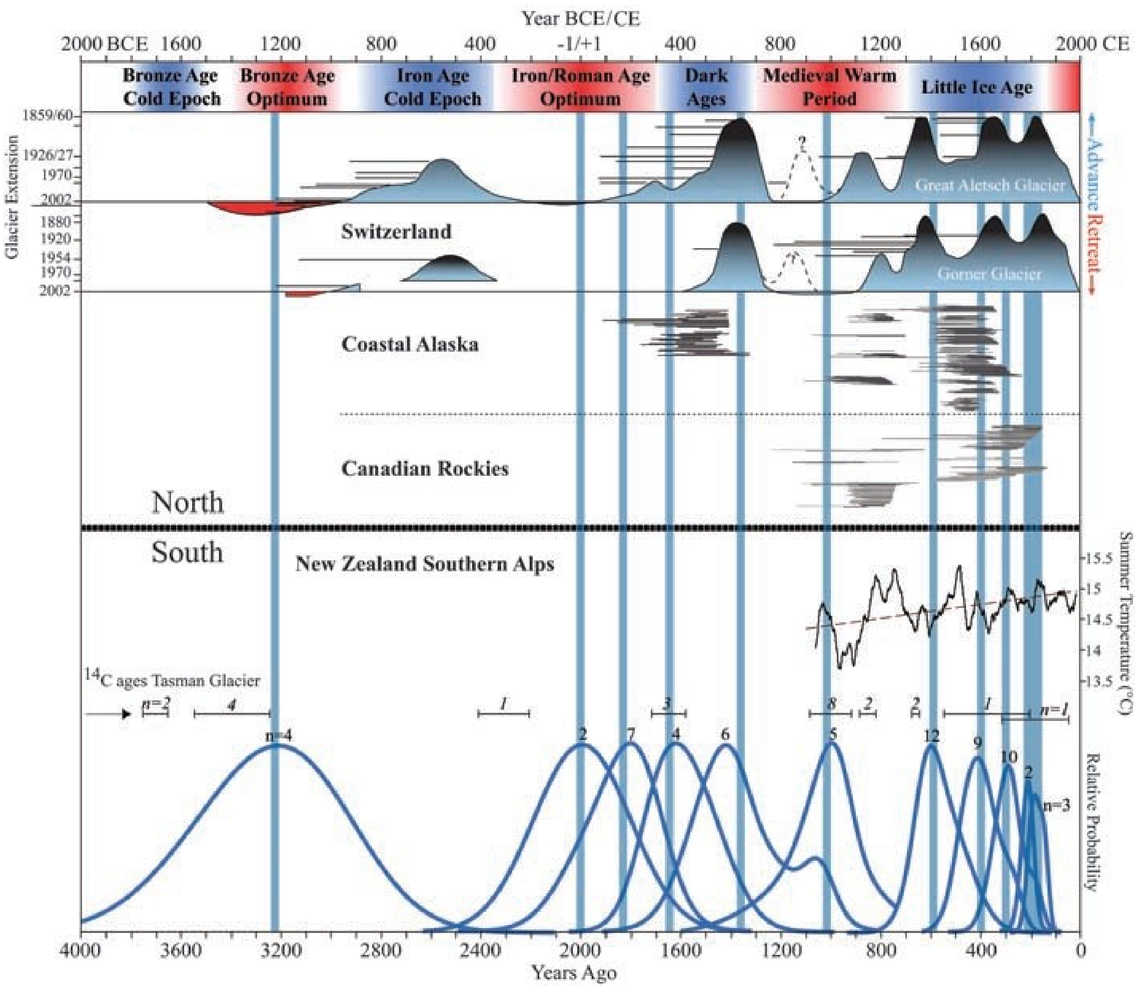

Figure 10: The timing of Holocene glacier fluctuations near Mount Cook in New Zealand's Southern Alps (bottom), compared with glacier fluctuations in the Northern Hemisphere (top). The probability plots at the bottom are summary curves of all individual ^{10}Be boulder ages from each moraine dated in New Zealand. The blue bars show the arithmetic means of the moraine age. The top graph shows fluctuations of two index glaciers in the Swiss Alps, the Great Aletsch Glacier and the Gorner Glacier. The middle graph shows the glacier advances in coastal Alaska and the Canadian Rockies. Source: Schaefer et al. 2009

positive and negative mass balance over this period have provided a wide range of response conditions when compared with the more general world-wide case of constantly receding glaciers (Chinn et al. 2012). This has been explained as the result of regional climate and glacier geometric properties. The majority of New Zealand glaciers are of small or medium size and have mostly short response times (6-20 years). However, glacier responses vary significantly according to glacier type (Chinn et al. 2012) and are described in section 5.2.

Glacier advances in a generally warming climate have been documented on many short response time glaciers throughout New Zealand (further reading). A number of maritime glaciers in Norway and Patagonia and few glaciers in Karakoram and Pamir have also advanced during the 20th century (Mackintosh et al. 2017), however the total advance is reasonably smaller than the global glacial retreat (WGMS 2008). The World Glacier Monitoring Service (WGMS) database shows that 58 glaciers advanced at some point in the 1980s, 1990s and early 2000s and 12 of these glaciers advanced continuously for five or more years. In 2005, when this glacier advance phase neared its end, 15 of the 26 advancing glaciers observed worldwide were in New Zealand (WGMS 2008).

5.2 Glacier retreat in New Zealand according to glacier typology

Even though glaciers are generally sensitive to climate change (e.g. Solomina 2015) the length changes of a single glacier do not represent a climate change. Different glaciers have different response times and different magnitudes, making it inappropriate to compare or to combine the responses of different types of glaciers (Chinn 1996a). Figure 11 describes this variability in glacier behaviour on an example of three characteristic glacier types. While small cirque glaciers respond directly to annual mass balance and snowline variability, medium-size mountain glaciers react dynamically to decadal mass-balance variations in a delayed and strongly smoothed manner. The Franz Josef glacier is of this type. Large valley glaciers damp decadal mass-balance variations but exhibit strong signal of secular developments. The Tasman Glacier is an example of this type (Chinn 1996a).

Therefore the glaciers with relatively short reaction times (<30 years), like the Fox and Franz Josef Glaciers, have gone through multiple advance–retreat cycles, while at the same time those with much longer reaction times (e.g. Tasman and Murchison Glaciers) have absorbed the positive pulses into their growing disequilibrium, with no advance at the terminus, but with the changes in mass volume, especially thining (due to downwasting or thermokarst processes) (Purdie et al. 2014).

In addition, increased debris-cover at retreating glacier termini complicates the retreat process by retarding ice melt once the debris-cover is greater than a few centimetres thick (Östrem 1959). The debris cover on Tasman Glacier is up to 3 m thick leading to significant insulation of the lower tongue and to highest ablation rates to be reported in upper parts of the glacier where there is no, or limited debris cover (Quincey and Glasser

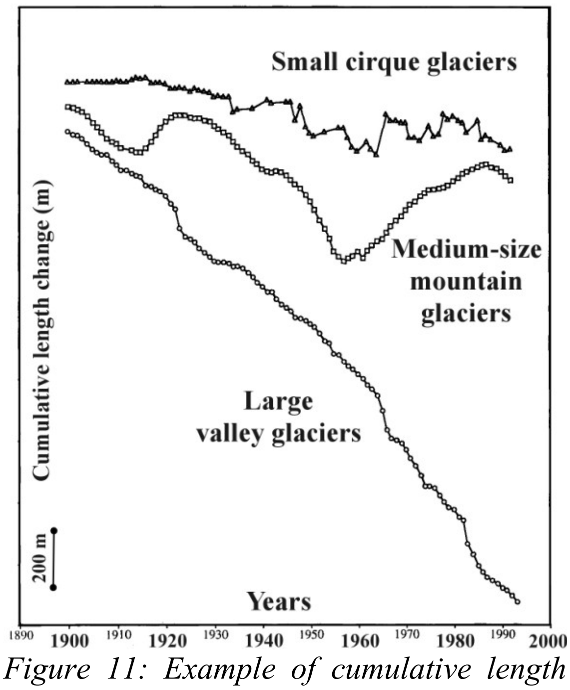

Figure 11: Example of cumulative length changes of three characteristic glacier types. While this example represents three glaciers in the Swiss Alps, it can be as well applied for New Zealand environment even the exact changes can vary. Edited according to Chinn 1996a

2009). The role of glacier geometry is crucial for length changes as well, because the steep glaciers can transport ice mass to lower altitudes where the ablation is much greater. It is clear that the glacier velocities are strongly dependent on the slope and landscape geometry. Therefore, from all those reasons, we can not accept a statement about a certain climate change just from behaviour of a single glacier and more detailed assessment needs to be done.

Figure 12 shows the historic length changes for four glaciers in New Zealand showing clearly the behaviour of difererent glacier types to a climate change. Location of those glaciers is to be seen on figure 13 together with mean glacier mass balances,

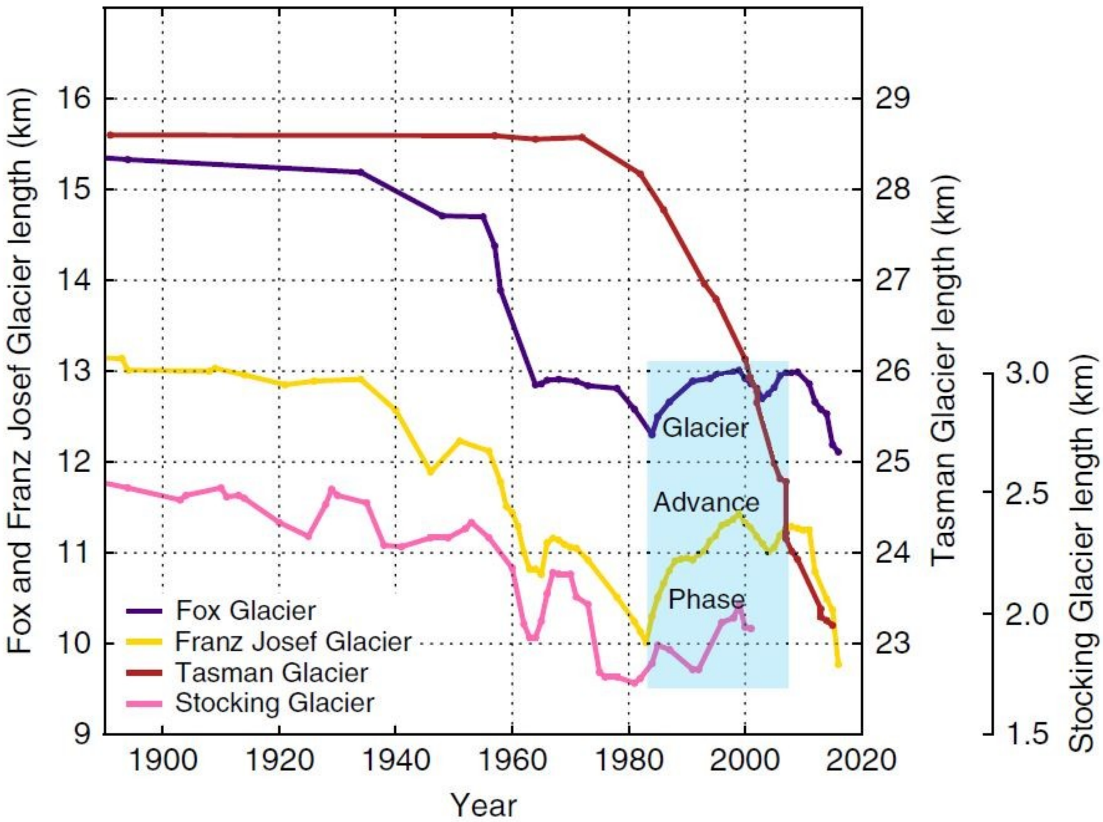

Figure 12: Historic length changes for four glaciers in New Zealand. Franz, Fox, Stocking and Tasman glaciers (see Fig. 13 for glacier locations). All of them retreated during the 20^{th} and early 21^{st} centuries. However, Franz Josef, Fox and Stocking glaciers also experienced periodic re-advances. The climatological drivers of the largest and most recent of these re-advances between 1983 and 2008 (marked by blue shading) is discussed in section 5.3., the behaviour of diferent glacier types in sections 5.2.1 and 5.2.2 and the individual glaciers are in detail discussed in section 5.4. The three glaciers that advanced are all steeply inclined and react swiftly and similarly to climate forcing. Tasman Glacier has a gentle slope and is the largest and thickest glacier in New Zealand. During the twentieth century, Tasman Glacier experienced continuous thinning, followed by retreat via proglacial lake formation since the 1980s. Source: Mackintosh et al. 2017

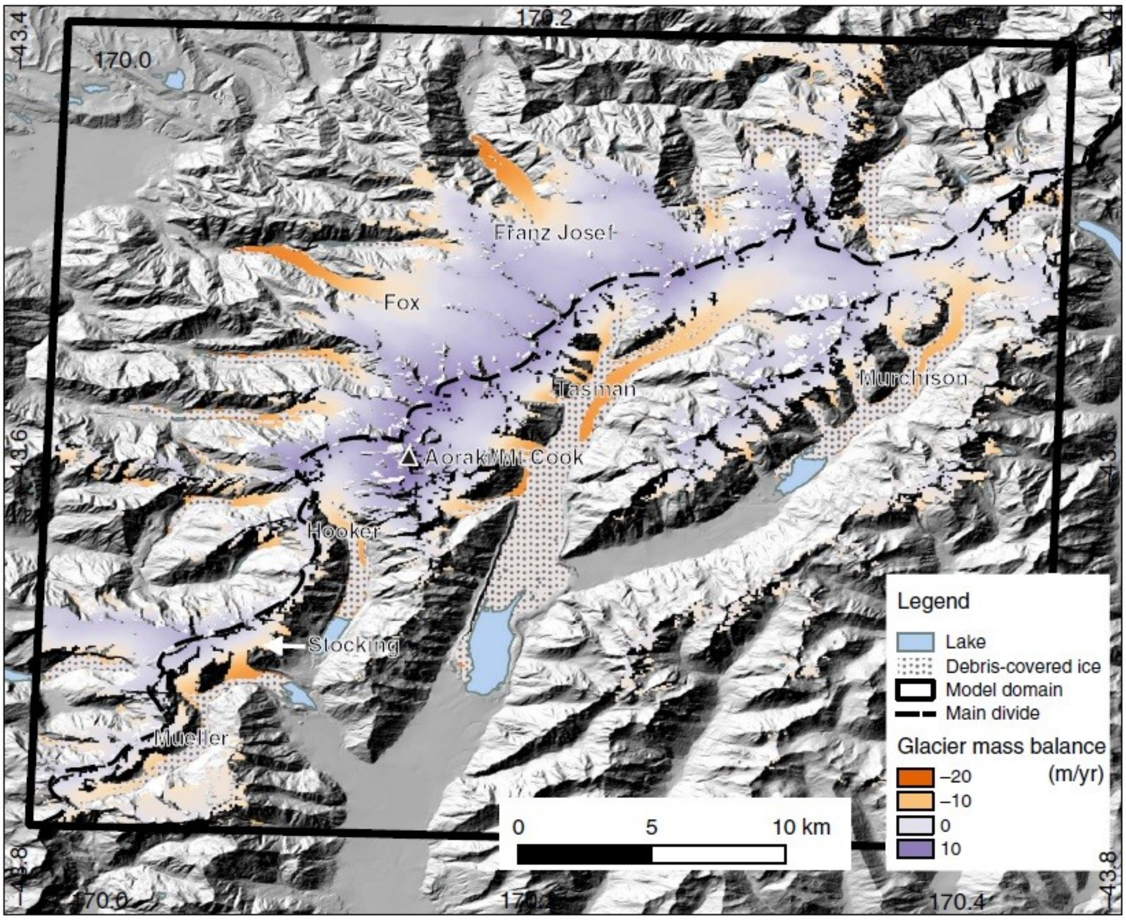

Figure 13: Location of the major glaciers within Mt Cook National Park showing the mean glacier mass balance (1972–2011), surface debris cover and proglacial lakes. Even the Southern Alps contain more than 3000 glaciers, the greatest volume of glacier ice in New Zealand is located within Aoraki/Mt Cook area. Mass balance (metres of water equivalent per year) is shown in red (net melt) and blue colouring (net accumulation). Very large gradients in glacier mass balance exist within the area, depending on glacier elevation and location of the Main Divide. Franz Josef and Fox glaciers to the west and north of the main divide each show snow accumulation rates of ~10 m, and melt rates of ~20m of water equivalent per year. Surface debris covers the lower portion of many glaciers including the Tasman, Hooker, Mueller and Murchison glaciers. Terminal lakes have grown rapidly at these glaciers since the 1980s. Source: Mackintosh et al. 2017

extent of debris cover and location of proglacial lakes.

The three glacier categories mentioned above can be merged into two distinct categories according their response time. The first group encompases all rapid to normal response time glaciers, which tend to be small to medium in size and include most of the glaciers of the Southern Alps (see section 5.2.1). The second category (described in section 5.2.2) refers to the set of large, low gradient valley glaciers with protracted response time (Chinn et al. 2012). While it is highly difficult, expensive and time consuming to monitor all the glaciers, a sample of representative glaciers has been chosen to estimate length changes of the other glaciers and to calculate the total ice volume changes. Throughout New Zealand 50 index glaciers have been selected (Fig. 14) to represent other New Zealand glaciers and to extrapolate data to other New Zealand glaciers. From 50 New Zealand index glaciers selected for mass balance calculations, 38 are smaller "short response time" glaciers and 12 represents the category of large valley glaciers with longer response times (Chinn et al. 2012).

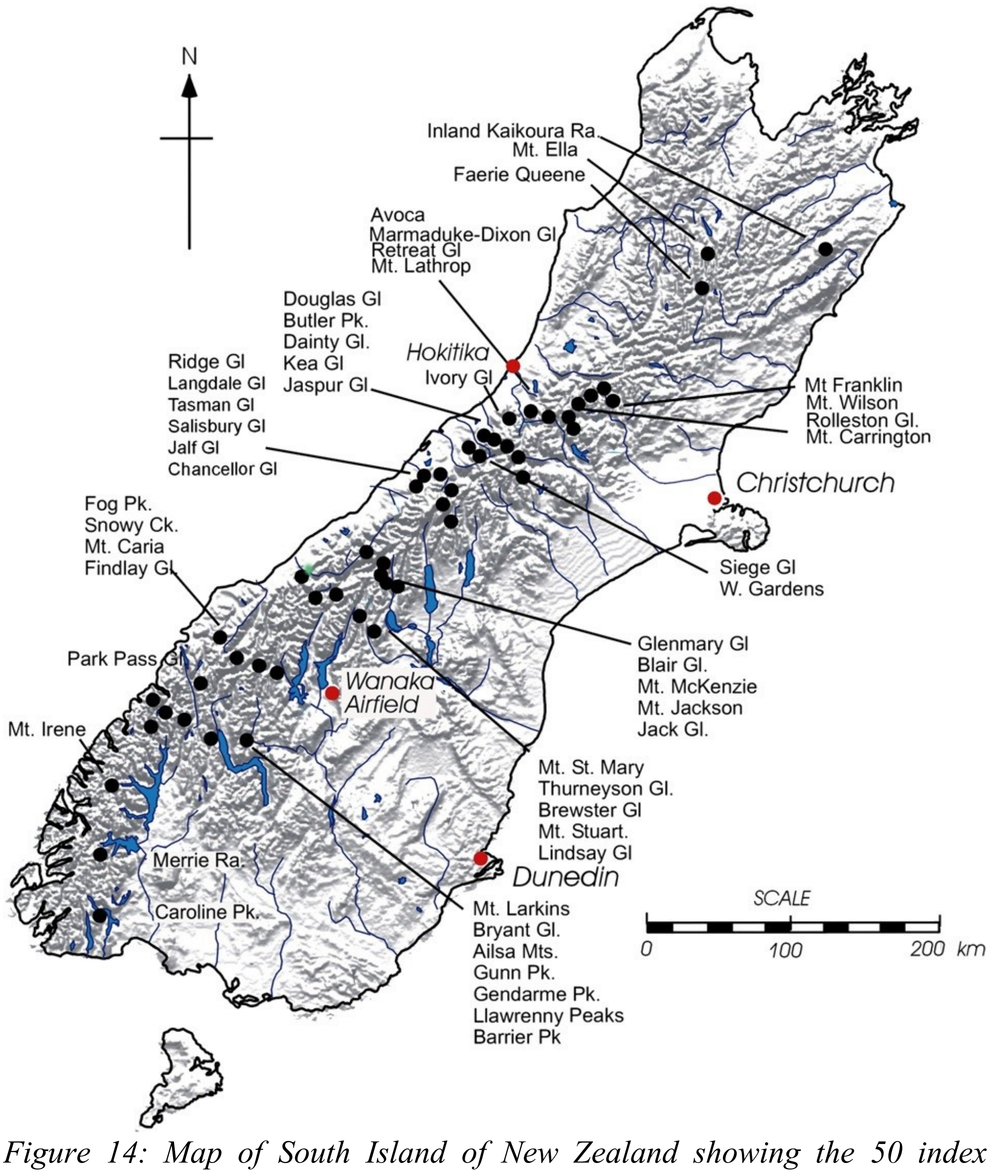

Figure 14: Map of South Island of New Zealand showing the 50 index glaciers (black dots). Red dots represent some important cities. Source: Chinn et al. 2012

5.2.1 Fluctuations of "Short response time" glaciers since 1977

From the total number of 3144 glaciers inventoried by Chinn (2001), most are small cirque glaciers with short response times. Those are to be found on both sides of the Main Divide throughout the Southern Alps. But both the cirque glaciers and the large steep glaciers of the humid zone west of the Main Divide, such as the Franz Josef and Fox glacier were put into this group. This is due to the short response times of those glaciers, short steep profiles, large mass turnovers and high terminus ablation rates. These factors all lead to quick adjustments to climate changes. Thus a large proportion of New Zealand glaciers have reached equilibrium with the present climate.

Those "short response time" glaciers are being monitored by marking the position of "End of Summer Snowline" (EOSS), which roughly corresponds with the "Equilibrium Line Ablation" (ELA), and following calculation of specific mass balance and thus the glacier volume (Chinn et al. 2012). Changes in specific mass balance for the small to medium size glaciers of the Southern Alps between 1977 and 2008 are shown in Fig. 15. Each of the decades of the 1980s, 1990s and 2000s had clusters of consecutive positive mass balance years (ELA is below normal). These clusters are separated by occasional strongly negative mass balance years. These are more pronounced than the positive years and indicate ELA were above normal.

Ice volume changes for "short response time" glaciers show considerable inter-annual variability for the small glaciers of the Southern Alps. There were annual losses of 1 to 2 $km^3\ a^{-1}$ in glacier years 1977/78, 1978/79, 1997/98, 2005/06, and of 2 to 4 $km^3\ a^{-1}$ in 1989/90, 1998/1999, 1999/2000, 2001/02, and 2007/08. There were annual gains of 0.5 to 2 $km^3\ a^{-1}$ from 1982/83–1984/85, 1991/92–1994/95 and 2002/03–2004/05. The cumulative mass balance of "short response time glaciers" shows a total loss of 2.45 km^3 for the Southern Alps over the 1976–2008 period which accounts for just 29% of the overall ice volume loss from the Southern Alps. The remaining 71% is related to the shrinkage of long valley glaciers (Chinn et al. 2012).

However, the selection of observation period is crucial for the results. If the period 1983–2008 (instead of 1976-2008) is selected, the small glaciers experience a total mass and length growth. Franz Josef Glacier, the glacier with the most complete length change record in the Southern Hemisphere, advanced for 19 of the 25 years between 1983 and 2008 (Mackintosh et al. 2017). Figure 12 clearly shows that this period (marked by blue shading in Fig. 12), or more correctly the period between 1983 and 1998 is surely the biggest period of an glacier advance since 1890. Glacial advance

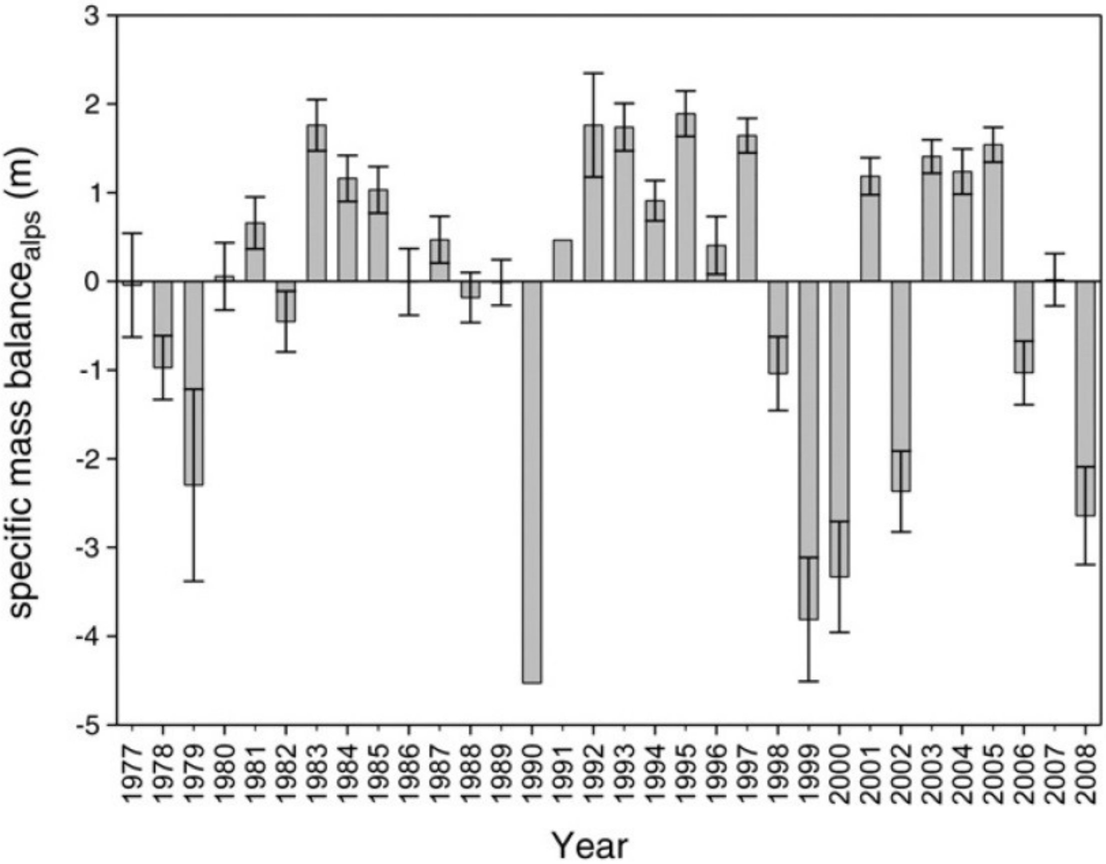

Figure 15: Changes in specific mass balance for the Southern Alps from 1977 to 2008 for the small to medium in size rapid to normal response time glaciers. Also shown are 95% confidence limits based on specific mass balance of 50 individual index glaciers. Estimates for the 1989/90 year are from observations for only two index glaciers, and for one index glacier from 1990/91.

between 1983 and 1998 is obvious also from the specific mass balance values in figure 15. While the negative mass balance in the year 1990 (Fig. 15) corresponds with the retreat of Stocking Glacier (see fig. 12), it can not be extrapolated to the other "short response time" glaciers, because the calculation was made just from two index glaciers. However, the specific mass balance data by Chinn et al. (2012) are consistent with the length change record of Franz Josef Glacier by Purdie et al. (2014).

5.2.2 Fluctuations of "Long response time" glaciers since 1977

For the set of large, low gradient valley glaciers with protracted response time, another method to estimate their volume change needs to be used. The volume estimates according to EOSS can not be used mainly because large, low gradient valley glaciers of New Zealand: (a) have long response times of up to a century or so; (b) carry a thick debris mantle that insulates their tongues; (c) tend to be out of equilibrium with the present climate largely because insulation of ablation areas has delayed their retreat; and (d) are subject to a tipping point once rapid frontal retreat is initiated by lake growth (Chinn et al. 2012).

In contrast to several advances and retreats of the small “short response time” glaciers, the 12 protracted response glaciers maintained their “Little Ice Age” areal extent right up until the 1970s (Chinn et al. 2012). This is despite sustained warming across New Zealand since the late 19th century (e.g. Putnam et al. 2012). As their ELAs rose by warming, they lost ice volume to downwasting, but largely retained their ablation zone area which shifted them to an extreme disequilibrium (Chinn et al. 2012). Retention of ablation areas of these glaciers was a result of: (a) a heavy insulating layer of debris, thickening to as much as 2 m at the termini, which retarded ablation rates; (b) low surface gradients; and (c) very thick ice extending well below the melt water river outlet level, which prevented terminus retreat. The tongues of these 12 glaciers were effectively large, active “ice ponds” lying in over-deepened hollows that could not easily retreat or even reduce size in response to the warmer climate. They lost ice volume by simple downwasting until the terminus ice levels subsided to the outlet river level (see figure 16), which occurred separately between the 1970s and 1990s. This critical development was a tipping point for these large glaciers. Thereafter, they experienced rapid lake expansion, catastrophic ice loss from calving and destruction of their lower trunks (Chinn et al. 2012).

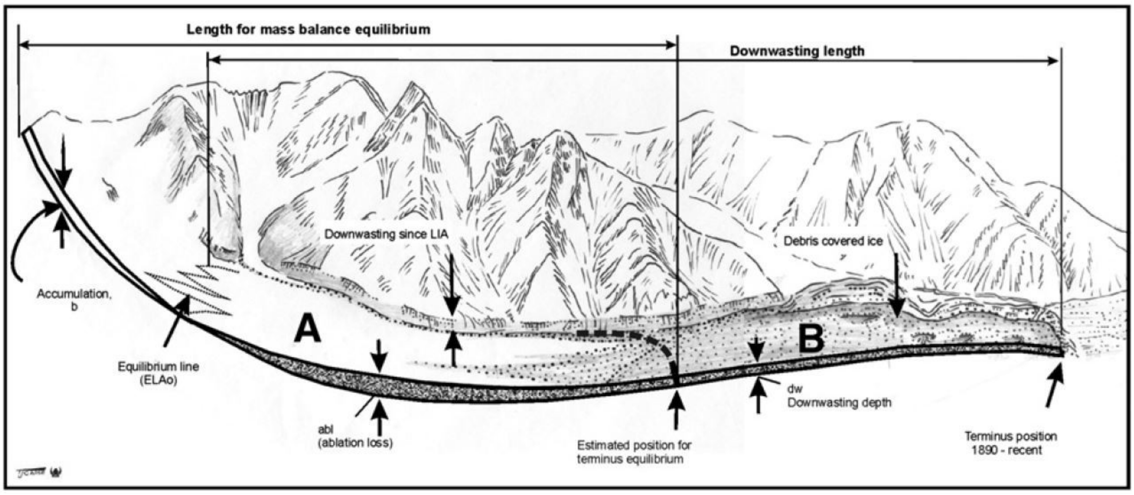

Figure 16: Model of a large debris mantled glaciers typical for the Southern Alps. A = estimated area and terminus position of the glacier for it to be in equilibrium with the present climate. B = the actual area before any lake development. Area B may be considered a measure of the disequilibrium of the glacier and is largely debris covered “relict” but active ice. The loss to downwasting is calculated from the ELA down to the existing terminus. Source: Chinn et al. (2012)

Mass losses in water equivalents for the 12 large protracted response glaciers of the Southern Alps are estimated as 0.75 km^3 for terminus calving and 5.22 km^3 for tongue downwasting over the 1976–2008 period (see figure 17). This represents a total loss of almost 5.96 km^3 over 32 years. This is much larger than the loss of ice from small and medium glaciers of 2.45 km^3 (Fig. 17). These calculations indicate that total ice volume of the Southern Alps has decreased from 54.53 km^3 in 1976 to 46.12 km^3 in 2008 (a loss of 8.41 km^3 or 15%) at a rate of 0.26 km^3 a^{-1}. The melt of the 12 long, low gradient valley glaciers is mainly due to tongue downwasting and terminus calving into expanding pro-glacial lakes (Chinn et al. 2012).

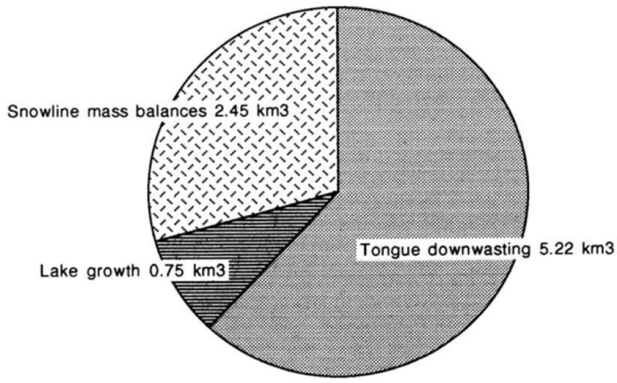

Figure 17: Sources of ice volume lost from the Southern Alps over the 1976 to 2008 period (total 8.4 km^3). According to Chinn et al. (2012)

5.3 Climatological factors of the glacier fluctuations in New Zealand

The fluctuations of New Zealand glaciers during the second half of 20th century have not been synchronous with most Northern Hemisphere mountain glaciers except of few examples (see section 5.1.6). Despite the global warming and general glacier retreat (IPCC 2013) many New Zealand glaciers were advancing between 1982 and 1999 (see section 5.2). The influence of individual climatological drivers has been broadly discussed (e.g. Fitzharris et al. 1992; Mackintosh et al. 2017), but mechanisms of atmospheric and oceanic changes yet need to be discovered.

New Zealand climate is a mosaic of atmospheric and oceanic influences (see section 4.1). Some of them are shown in figure 18. The belt of southern westerlies wrestles with the subtropical ridge shifting the weather patterns with them. Sea Surface Temperatures (SST) of the Tasman Sea are dictated by changes in huge oceanic currents bringing warm or cold water from the tropical or polar regions. The strength of those currents thus significantly influences the regional climate. All atmospheric and oceanic drivers interact together, creating various oscillations, further interacting between each other.

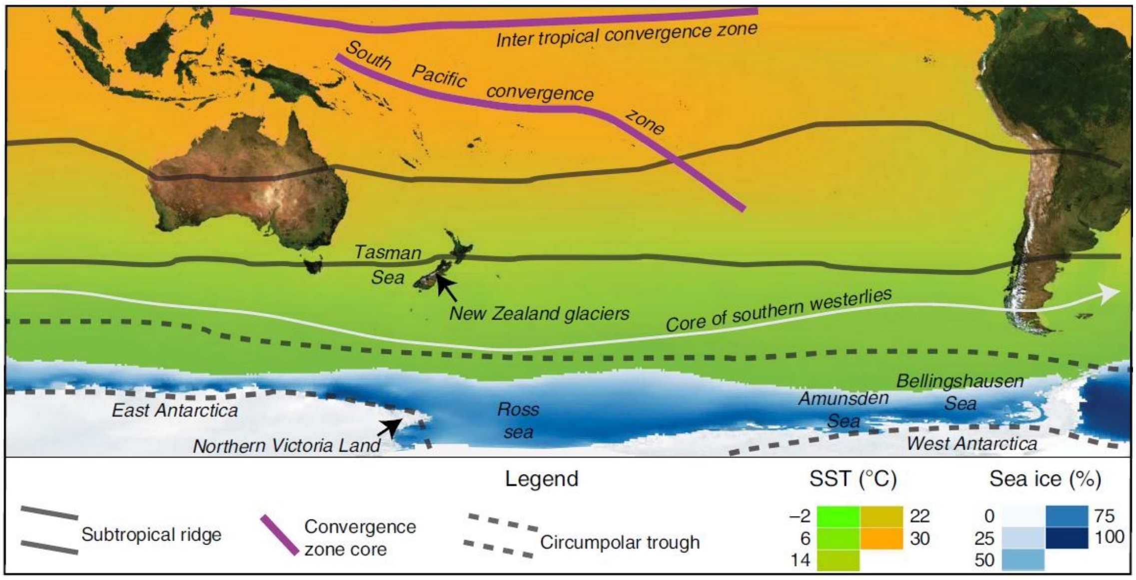

Figure 18: Southern Hemisphere climatological features. To the north lies the subtropical ridge, South Pacific Convergence Zone and Inter Tropical Convergence Zone. To the south lies the core of the westerlies, the circumpolar trough and Antarctic sea ice. All of these features have the potential to influence atmospheric and oceanic conditions in the New Zealand region, and hence Southern Alps glacier mass balance. Sea surface temperature (SST) is from annual mean data, while sea ice data show peak concentration (%) reached in the austral spring. Other climatological features are plotted in their mean annual positions. Source: Mackintosh et al. (2017).

5.3.1 The key climate forcing: Temperature vs precipitation

It has been greatly debated whether recent New Zealand glacier fluctuations relate more to temperature or precipitation forcing. While Salinger et al. (1983) suggest that the retreat of Stocking Glacier has been related to a temperature change, Hessell (1983) in Fitzharris et al. (1992) showed that correlations with temperature are not significant, and that changes in precipitation are probably responsible for the retreat of the Franz Josef Glacier before 1982. Fitzharris et al. (1992) concluded that many New Zealand glaciers advanced in response to a reduction in summer ablation that results from increased cloudiness and precipitation and, therefore, snow accumulation at higher altitudes. Anderson and Mackintosh (2006) used climate modelling for Franz Josef Glacier and concluded that temperature is the dominant control on glacier length. Also Mackintosh et al. (2017) do not support the hypothesis of precipitation forcing and suggest that the advance of glaciers in the Southern Alps between 1983 and 2008 was mostly due to reduced air temperature rather than increased precipitation. Many scientific papers have been documenting this conflict between temperature and precipitation forcing and the arguments are being discussed in more recent studies (e.g. Mackintosh et al. 2017; Purdie et al. 2014).

The most recent study (published during the works on this thesis) by Mackintosh et al. (2017) used diagnostic experiments with energy balance model to evaluate whether precipitation, temperature or another climate variable caused the glacier advance phase between 1980 and 2005. To test the hypothesis that precipitation increase caused the glacier advance phase, the relative contribution of precipitation variability (Fig. 19a) to glacier mass balance was assessed by holding all other climatic variables at their mean values. This experiment illustrates that from 1972 to 1979, precipitation had a negative influence on mass balance (resulting in volume loss), while between 1979 and ~2000, it had a positive influence. Although this is consistent with the hypothesis that increased precipitation caused the glacier advance phase in New Zealand, the diagnostics of Mackintosh et al. (2017) indicate that precipitation variability accounts for only 27% of the total ice volume anomaly during the advance phase (Fig. 19c).

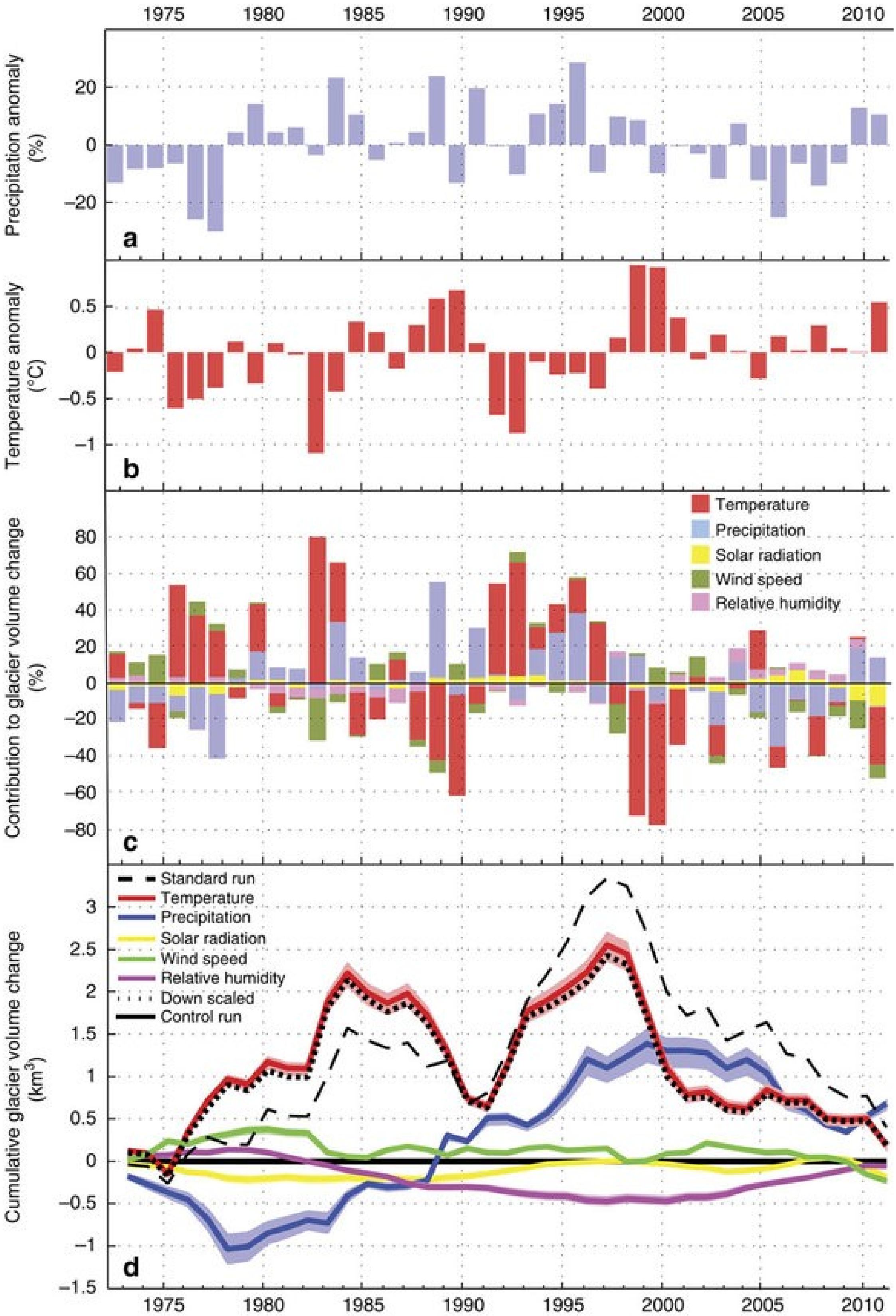

Figure 19: Diagnostic experiments with energy balance model. Annual precipitation (a) and temperature (b) anomalies during the study period. (c) represents the relative contributions of different components of the climate forcing to simulated glacier volume changes. These proportions are derived from the diagnostic experiments (d), which show the relative contribution of each climatic component to the cumulative glacier volume change. Source: Mackintosh et al. (2017).

Mackintosh et al. (2017) repeated the experiment to examine the influence of temperature variability (Fig. 19b) on mass balance changes by holding the other climate variables at their mean annual values. The resulting pattern (Fig. 19d) shows that temperature change is the dominant variable that caused the glacier changes, accounting for 56% of the total volume anomaly during the glacier advance phase (Fig. 19c). The experiment was repeated with wind, cloudiness and relative humidity, but the combined effect of each of these variables resulted in small (17%) changes in glacier volume, compared with temperature (56%) and precipitation (27%).

5.3.2 Climatic processes associated with glacier advance

The Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report also discussed and summarised the conflict between various climatic drivers and their influence to glacier fluctuations, stating: "The exceptional terminus advances of a few individual glaciers in Scandinavia and New Zealand in the 1990s may be related to locally specific climatic conditions such as increased winter precipitation" (IPCC 2013). However, this summation remains speculative because direct glacier mass balance and high-elevation climate data from the Southern Alps are limited (Mackintosh et al. 2017).

The correlation of glacier fluctuations to various synoptic situations gives much more conformable results (see e.g. Fitzharris et al. 1992; Mackintosh et al. 2017). Periods with glacier advances were correlated with an increase in westerly and south-westerly airflow associated with a northward shift of the subtropical high-pressure zone (Fitzharris et al. 1992). The northward displacement of the region of strong westerlies over New Zealand is mostly dominant during the austral summer and is associated with an increase in the frequency of deep lows that pass over and south of the South Island (Mackintosh et al. 2017).

The shifting of circumpolar trough and subtropical ridge is closely tied with the shifts of the convergence zones, South Pacific convergence zone (SPCZ) and Inter tropical convergence zone (ITCZ). Putnam et al. (2012) consider that asynchronous glacier behaviour between the hemispheres is most readily explained by the effects of insolation-driven migrations of the ITCZ. The shift of the belt of southern westerlies northward and an increase south-westerly airflow most probably relate to advection of cool high-latitude surface waters and leads to regional temperature changes in the Tasman Sea (Mackintosh et al. 2017).

The promotion of lower regional sea surface temperatures (SST) during spring and summer contributes to anomalously low temperatures in the Southern Alps (up to 1 °C lower than the 1981–2010 mean), downwind of the Tasman Sea. The lower ambient temperatures favour positive glacier mass balance by increasing the snow component of total precipitation during spring, by lowering the elevation of the temperature-dependent snow/rain threshold. Lower temperatures also reduce melt during summer, thus increasing the length of the accumulation season. Increased snow during spring also increases the glacier albedo, delaying the melt season onset and reducing melt season length (Mackintosh et al. 2017).

Both the mean sea level pressure (MSLP) and sea surface temperature (SST) have their annual, inter-annual, decadal and even much longer variations which are modulated by terrestrial (mostly atmospheric-oceanic) processes but also by extraterrestrial forcing (mainly the amount of incoming solar radiation). Schaefer et al. (2009) suggest that regional ocean-atmosphere oscillations may account for the observed glacier fluctuation patterns and supports the hypothesis that the Interdecadal Pacific Oscillation (IPO) has been an important influence on glacier behaviour in New Zealand over the past few decades. The IPO has been shown to be associated with decadal climate variability over parts of the Pacific Basin, and to modulate interannual El Niño–Southern Oscillation (ENSO)-related climate variability over Australia. Three phases of the IPO have been identified during the 20th century: a positive phase (1922–1944), a negative phase (1946–1977) and another positive phase (1978–1998) (Salinger et al. 2001). During the positive phases colder and wetter conditions were observed in New Zealand's Southern Alps and during negative phases warmer and drier conditions were observed. These changes are well reflected in New Zealand's glacier length fluctuations (Schaefer et al. 2009).

The influence of El Niño–Southern Oscillation (ENSO) has been greatly debated within the scientific community and wide public alike. However, the results differ between authors. Purdie et al. (2014) summarised results of previous studies and concluded that westerly flow anomalies and phases of ENSO relate to glacier mass balance in the Southern Alps. In particular, negative phases of ENSO are associated with positive mass balance, and positive ENSO phases with negative mass balance. However, the degree of phase persistence required to induce a terminus reaction will depend on individual glacier reaction time (Purdie et al. 2014).

On the other hand Mackintosh et al. (2017) do not find such a strong relationship. They agree that ENSO plays a role in setting up the extratropical circulation patterns that ultimately affect Southern Alps glaciers, but suggest that it is only one of several contributors towards generating lower sea surface temperatures in the Tasman Sea. The period of major glacier advance in the 1980s and 1990s saw frequent El Niño activity implying for a strong relationship, however during the 1997/1998 El Niño event, one of the largest of 20th century, Southern Alps glacier mass balance was only weakly positive. This is because the SSTs in the Tasman Sea, and air temperature in the Southern Alps, stayed relatively high during this event. This pattern was again seen during the 2015/16 El Niño event, when the Tasman Sea and air temperature in New Zealand remained unusually high during the austral summer (Mackintosh et al. 2017).

Mackintosh et al. (2017) suggest that ENSO is not the sole (or dominant) influence on local glacier activity or regional climate anomalies, and that anomalous circulation over New Zealand required for glacial advance arises from the interplay of both tropical and extratropical processes, and that these circulation patterns, and associated changes in sea surface temperature have broader impacts across the high southern latitudes.

5.4 Fluctuations of selected New Zealand glaciers

As mentioned in section 5.2, the behaviour of New Zealand glaciers is strongly dependent on their typology. The response of steep, "short response time" glaciers is reasonably different from the large valley glaciers with protracted response times. While Fox and Franz Josef Glaciers are the best examples of the first group, Tasman Glacier is the flagship of the second group. Sections 5.4.1 and 5.4.2 describe the evolution of those glaciers into a more detail. While the behaviour of the other glaciers approaches the evolution of either the duo Fox – Franz Josef Glacier or the Tasman Glacier it was concluded that the description of the other New Zealand Glaciers is redundant.

5.4.1 Franz Josef and Fox Glaciers

Both Franz Josef Glacier and Fox Glacier are situated on the western side of the Southern Alps, in Westland, South Island, New Zealand (Fig. 20). Of the ~3100 glaciers in the Southern Alps, Fox and Franz Josef are two of the best known and are the third and fourth largest glaciers by volume respectively (Chinn 2001). Franz Josef Glacier

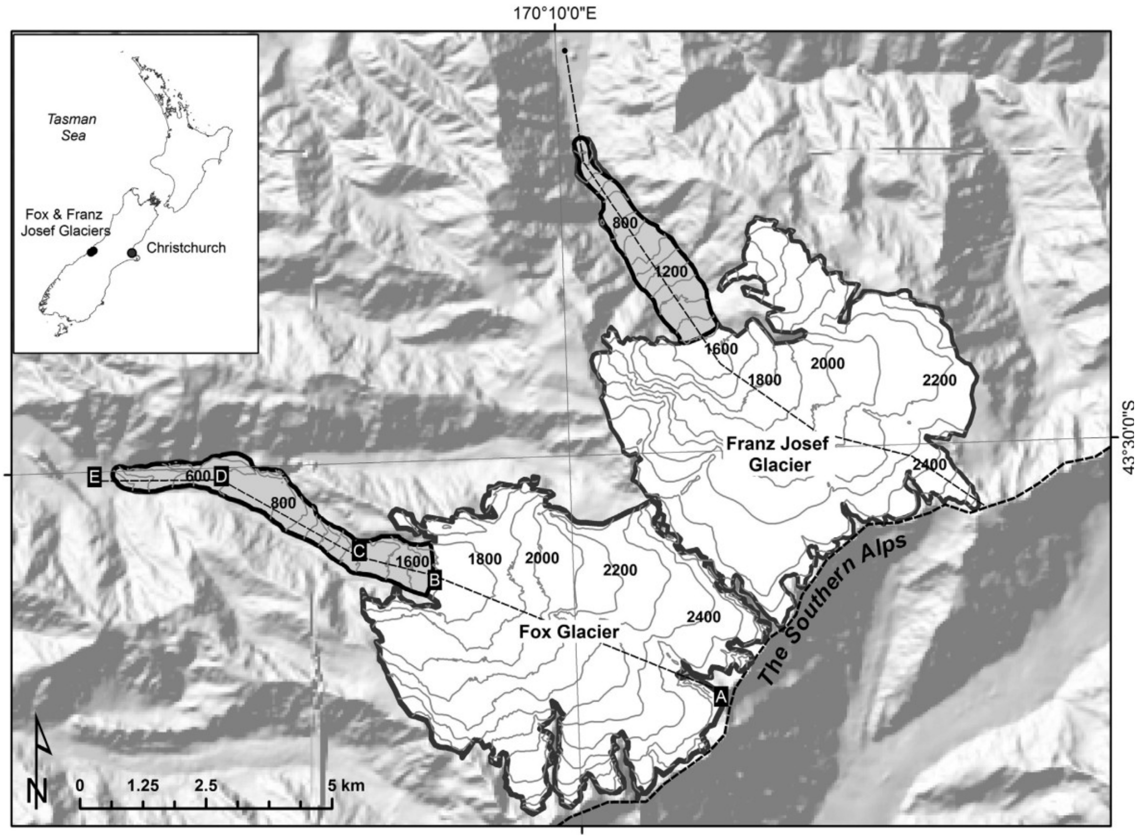

Figure 20: Location of Franz Josef and Fox Glaciers in the Southern Alps, New Zealand. Glacier outlines are derived from ASTER imagery in 2009. Shaded regions on the lower glacier denote the portion used for area calculation. Dotted lines and associated points define the datum used for length calculations by Purdie et al. 2014.

has also the most detailed record of terminus change for any Southern Hemisphere glacier (Purdie et al. 2014). Franz Josef Glacier is presently (2014) just under 10.5 km long and covers ~35 km^2 on the western flanks of the Southern Alps of New Zealand at 43°29′ S, 170°11′ E. The maximum elevation of the glacier is 2900 m a.s.l., although the bulk of the glacier consists of the upper névé area of broad gently-sloping snowfields at elevations ~1900 to 2400 m a.s.l. The glacier tongue descends steeply down a narrow valley to ~300 m a.s.l. The adjoining Fox Glacier is slightly larger, ~12.5 km long with an area of ~36 km^2. It also has a larger elevation range, with ice feeding from the western face of Mt Tasman at 3497 m a.s.l. Like Franz Josef, Fox Glacier has a broad high-elevation névé that funnels ice down a similar steep narrow tongue, and terminates below 300 m a.s.l. (Purdie et al. 2014).

Located in a high precipitation maritime environment, Franz Josef and Fox glaciers are highly sensitive monitors of atmospheric temperature because of rapid ice throughflow from accumulation to melting zones (Putnam et al. 2012). The extreme precipitation up to 14 000 mm/year (Henderson and Thompson 1999) nourishes the upper parts of the glaciers with heaps of snow. However, the glacier tongues terminate within a temperate rainforest in altitudes where temperatures barely reach bellow zero. Monthly mean temperatures at Franz Josef town range between 5.8 °C in July to 16.8 °C in February (Climate-data.org 2017) thus an intense ablation occurs throughout all seasons (Marcus et al. 1985).

In contrast with Fox Glacier the behaviour of the terminus of the Franz Josef Glacier is well documented already since 1893, when it was first mapped, although photographs exist from 1867, and dated moraines extend the record back to 1750 (Fitzharris et al. 1992). Several studies have described advances of Franz Josef glacier during the Pleistocene and Holocene and the Waiho Loop moraine has been studied extensively (see section 5.1.3).

Franz Josef Glacier has also a series of well-preserved moraines up the valley enabling estimation of glacier maximum lengths during the period ~1600 to 1800 A. D. At this time the glacier was around 14 km long. Figure 21 shows the length changes of Fox and Franz Josef glaciers since the first survey conducted in 1893. Since then glacial fluctuations of Franz Josef Glacier were monitored regularly. First the retreat was recorded, followed by an advance in 1907 (reaching its maximum in 1909–1910). Retreat until 1921 was followed by a minor re-advance in 1926, which ended around 1934. After 1934 there was a period of rapid retreat, and by 1946 the terminus was over

1 km up-valley from its 1934 position. This was followed by a small advance (340 m) between 1946 and 1951. From the early 1950s to early 1965 the glacier retreated, but an advance occurred from late 1965 to early 1968, which left the glacier just over 11 km in length. The 1970s were dominated by retreat, and by 1983 Franz Josef Glacier was the shortest that it had been since measurement began. Another advance phase was recorded in monthly detail between 1984 and 1990. Although this advance appeared to cease temporarily in 1991, the glacier continued to expand through the mid-1990s, and by 1999 the Franz Josef Glacier was over 11.4 km long, a length not seen since 1960. After 1999 the glacier retreated over 400 m until in 2005 the next (and most recent) advance began. This most recent advance ceased in 2008 and the glacier is currently retreating (Purdie et al. 2014).

The current (2008-2016) retreat of Franz Josef glacier is demonstrated in figure 22 and the retreat since 1800 in figure 23. Retreating trend since 2008 is clearly

Figure 21: Summary of length changes at Fox and Franz Josef Glaciers from1894 to present. The overall pattern of advance and retreat is generally synchronous, but Franz Josef Glacier tends to lead phase changes by around a year. Where data are of annual frequency they are connected by a solid black line. Dotted lines represent intermittent measurement — as can be seen, an advance recorded at Franz Josef in the early 1950s was not recorded at Fox Glacier, but this is likely due to a lack of monitoring. Source: Purdie et al. 2014

Figure 22: Recent retreat of Franz Josef Glacier. The massive retreat that commenced in 2008 continues to the present. The two pictures above were photographed from information panel by DoC and the three pictures bellow were used from NZ Herald web page (Morton 2017)

obvious not even from the terminus position, but also from the newly uncovered valley slopes. Even now (2017) Fox and Franz Josef Glaciers are still receding. According to Dr Brian Anderson of Victoria University's Antarctic Research Centre, Fox Glacier and Franz Josef Glaciers had shrunk to record levels (Morton 2017). "Winter 2015 had pretty average snow accumulation on the glaciers, but summer 2016 was really warm, and by the end of the summer the glaciers had lost a lot of snow and ice." (Anderson in Morton 2017). The most recent observations show, there is a continuing retreat even in 2017. "All I can really say is it's likely to keep retreating because of the warm summer, but I can't really say how much until we can check it again." (Anderson in Morton 2017).

Unlike that of Franz Josef Glacier, the Fox Glacier record suffers from a lack of regular surveys, with only thirty-seven data points covering over 250 years. Distinctive trim-lines in present-day vegetation and glacial moraines have helped to estimate the extent of the glacier around the time of the LIA. The modern advances presumably coincide with advances at Franz Josef Glacier, the dates of which are better constrained. Figure 21 shows the approximate fluctuations, however the error until 1965 (when the first photogrammetrically-derived topographic map was created) can be higher due to the lack of measurements. The dramatic advance since 1983 until 1999 is consistent with the one of Franz Josef Glacier and the Fox Glacier was once again 13 km in length. The fluctuations of Fox Glacier are consistent with the fluctuations of Franz Josef Glacier also in the 21st century (Purdie et al. 2014).

Area changes were summarised by Purdie et al. (2014). It was concluded, there is a similarity in the overall area change between the Fox and Franz Josef Glaciers between the late 1890s and 1987. However, the most recent advance has resulted in greater area change at Franz Josef, probably due in part to glacier geometry and the very confined valley in which it terminates. (Purdie et al. 2014).

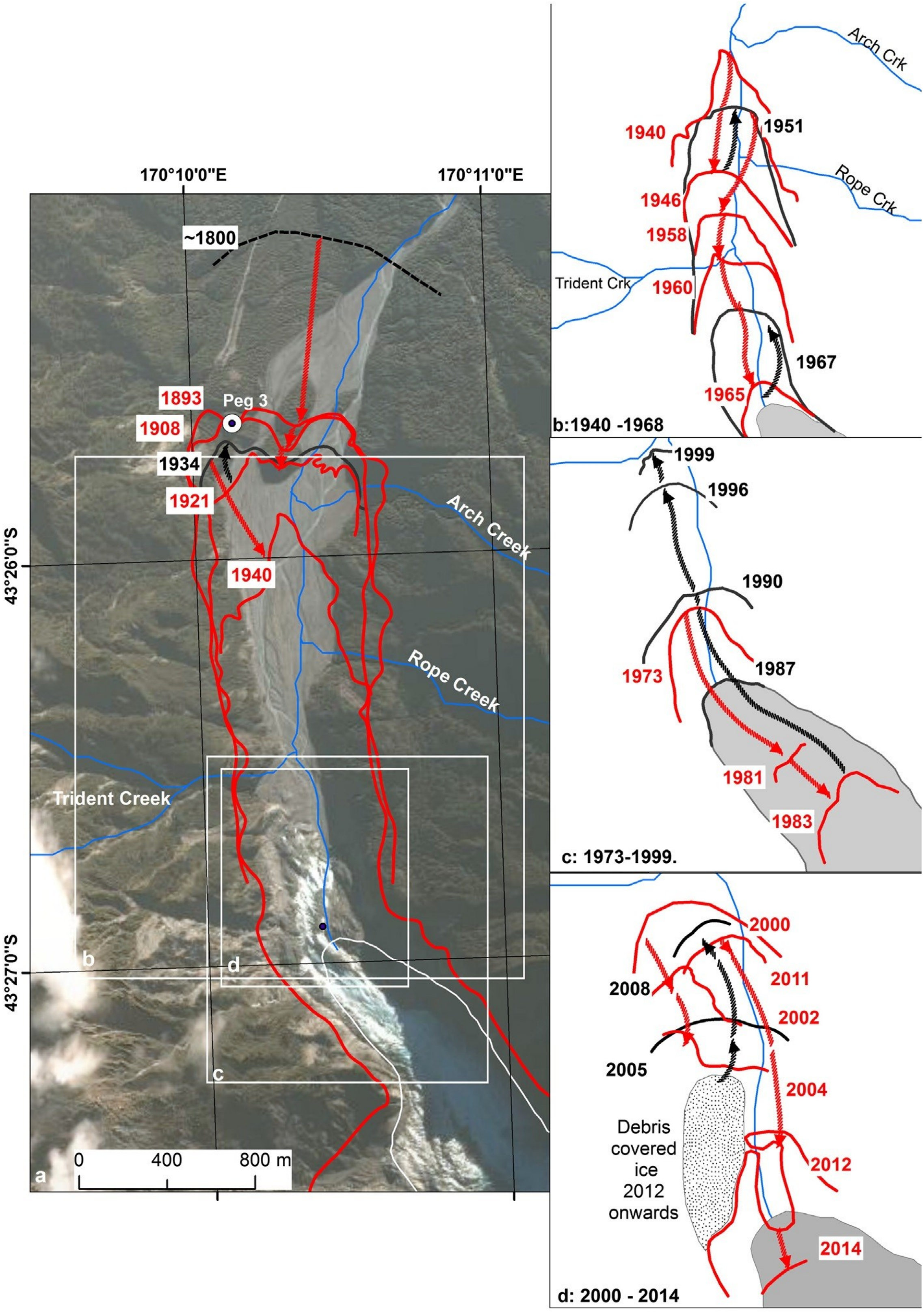

Figure 23: Changes at the Franz Josef Glacier terminus since 1800 to present. Black lines represent periods of advance and red lines periods of retreat. The 1987 glacier outline (derived from aerial photography) is shown for reference. (a) 1800–1940, general retreat with an advance in 1934. (b) 1940–1967, advances in the early 1950s and late 1960s. (c) 1973–1999, minimum (1983) and maximum (1999) extent recorded during recent time. (d) 2000–2014, the latest advance (2008) and present day retreat. Purdie et al. 2014.

5.4.2 Tasman Glacier

While Franz Josef Glacier has the most detailed record of terminus change for any Southern Hemisphere glacier (Purdie et al. 2014), Tasman Glacier has the longest and most detailed historical record of any New Zealand glacier (Kirkbride and Warren 1999). With the current length of about 22 km, width approximately 2 km at its terminus and an area of about 55 km^2 (220 km^2 if tributary glaciers are included) the Tasman Glacier in the Mt Cook National Park is the largest glacier in New Zealand (Purdie and Fitzharris 1999). The glacier descends from an altitude of 2400 m to ~730 m near the terminus (Hochstein et al. 1995). This huge, partly debris-covered, ice mass drains the eastern and southern flanks of the New Zealand Southern Alps (Fig. 24). The glacier is fed by three major tributaries, hereon referred to as Flow Units 1–3 (Fig. 25).

According to Quincey and Glasser (2009) flow Unit 1 is located at the head of the Tasman Glacier catchment and is fed predominantly by snowfields between approximately 2400 m and 3100 m in elevation below the peaks of Minarets (3040 m), Mt Green (2837 m), Mt Elie de Beaumont (3109 m) and Mt Abel (2688 m). Flow Unit 1 descends to the Hochstetter icefall, a distance of 16.5 km, with the lowermost 20% of ice being mantled in debris. Flow Unit 2 joins Flow Unit 1 down-glacier. It comprises the Rudolph Glacier, which is a heavily debris-covered ice mass, sourced from a high-elevation (2950 m) cirque. The Rudolph Glacier supplies a large percentage of the

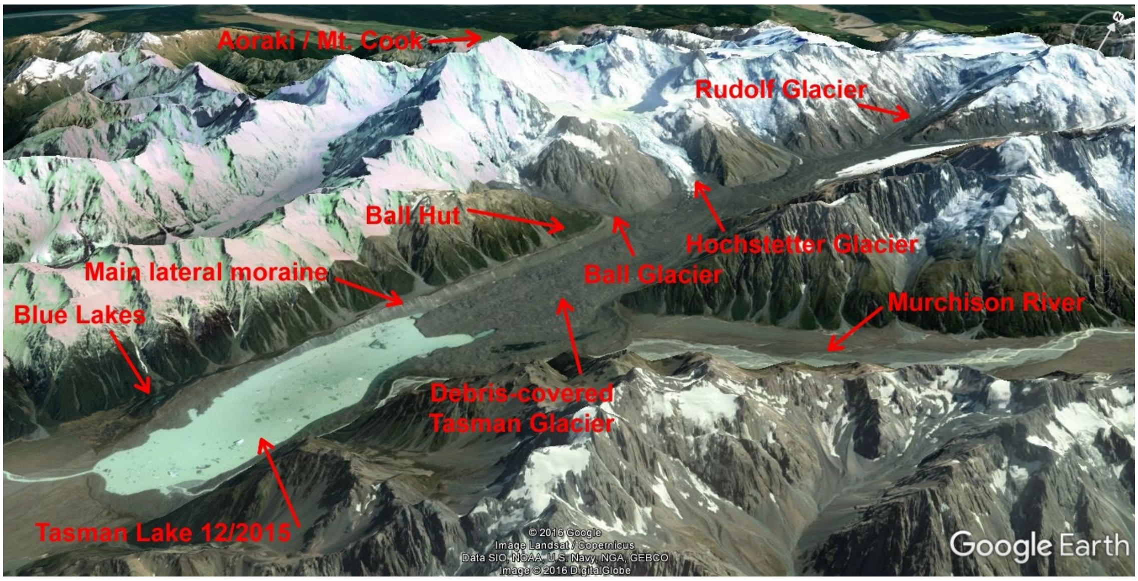

Figure 24: The position of Tasman Glacier within the Southern Alps of New Zealand. Key geomorphic features are labelled. The size of Tasman Lake represents the state in December 2015. Background by Google Earth, labels by author.

debris that mantles the lower parts of the main glacier tongue.

Flow Unit 3 is sourced by the Grand Plateau and the Hochstetter icefall, which together drain the eastern flanks of Mount Cook (3754 m), the highest mountain in New Zealand, and Mount Tasman (3497 m). While the units 1 and 2 terminate adjacent to the icefall in a zone of stagnant ice, the lowermost 12 km of the glacier tongue appears to be fed exclusively by the Hochstetter icefall, which provides sufficient ice to sustain flow from this point almost down to the calving ice front. It is debris-free in its upper parts,

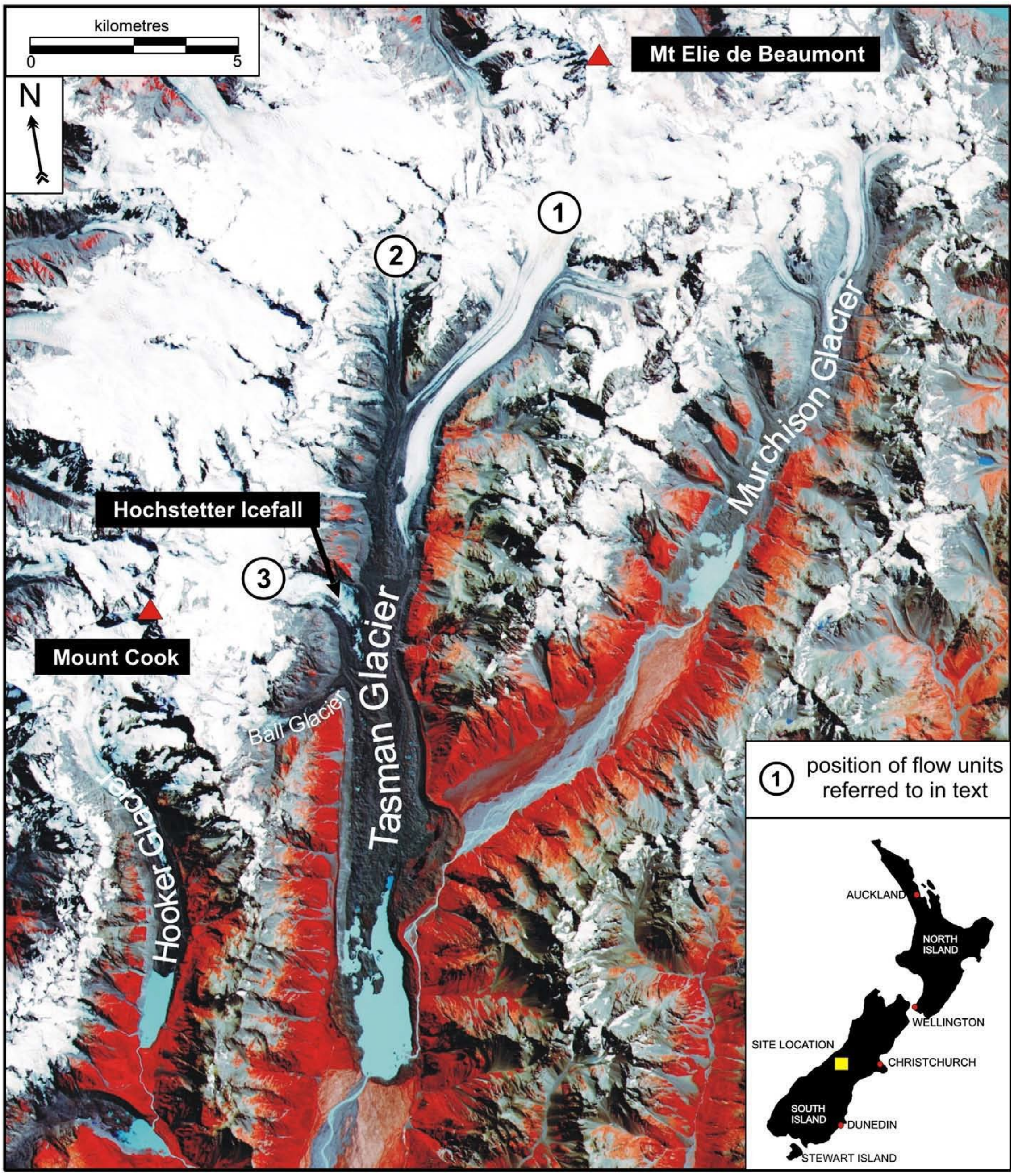

Figure 25: Location of the Tasman Glacier in the Mount Cook area of New Zealand. Base image is an ASTER scene acquired in January 2006. Positions of flow units are marked by numbers 1-3. Source: Quincey and Glasser (2009)

but carries large volumes of rockfall debris through the icefall and onto the lower section of the Tasman Glacier tongue. From this point, the main glacier tongue is pitted with supraglacial ponds, kettle holes and exposed ice cliffs. The glacier terminates in a large, calving ice cliff that abuts the developing proglacial lake (Fig. 24 and 25).

Schaefer et al. (2009) concluded that an inner moraine ridge of Tasman Glacier yield an age of 1040 years BP. When the first Europeans explored the glacier in the latter half of the 19th century, they found that it was either still advancing or was close to its maximum extent (Fitzharris et al. 1992). The terminus position was first surveyed in 1862 and the glacier photographed in 1869 by von Haast, and the lower glacier was mapped in 1883, and again in 1890. Both maps record the extent of the debris cover and surveyed surface altitudes. The glacier surface was reaching the main crest of lateral moraines or even overtopping it (Skinner 1964) meaning that the glacier maintained both its length and mass since the maximum extent around 910 A. D.

Thereafter, limited information is available until the earliest aerial photographs were taken in 1957, apart from short reports detailing thinning of the glacier (Kirkbride and Warren 1999). Since the first explorations in the 19th century, the large Tasman Glacier has rapidly downwasted (Skinner 1964), although the position of its terminus has changed little (Fitzharris et al. 1992) (Fig. 26). At an altitude of 1100 m an average

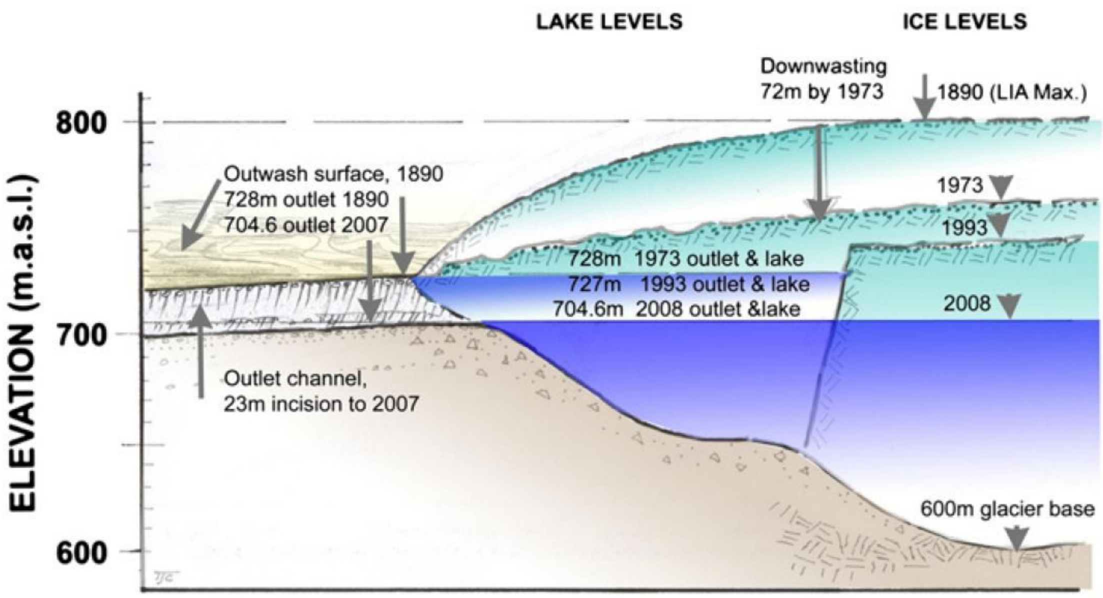

Figure 26: Progressive stages of ice loss from Tasman Glacier associated with regional warming. At the LIA, the meltwater river feeds directly on to an alluvial outwash gravel fan head. Subsequent stages are indicated: downwasting from LIA levels to formation of thermokarst features on the terminus; formation of proglacial lake; calving and retreat of ice cliff. Source: Chinn et al. 2012.

thickness of 82.1 m of ice has been lost between 1890 and 1962 (Skinner 1964). Thinning of the glacier continued even in the second half of 20th century. Between 1890 and 1986 the surface has been lowered of ca. 185 m (from 737 to 552 m glacier thickness) at the 10-km transect, and by ca. 115 m (from 287 to 172 m) at the 2-km transect (measured from the 1890 terminus). Therefore the form of the long profile has changed from one which steepened towards the terminus in 1890 to one which became progressively more gentle downstream by 1986 (see Fig 26). Glacier thinning has produced a marked change from a convex to a concave profile, but with no change in terminus position until the early 1980s. (Kirkbride and Warren 1999).

It was also not until the late 1950s that supraglacial ponding was clearly evident around the terminus (Kirkbride and Warren 1999). By 1971, however, the craters had grown in size and were filled with grey-coloured meltwater (Hochstein et al. 1995). Since the early 1980s dramatic changes of the terminus region have occurred. Melting on a scale not previously reported for any temperate glacier in the Southern Hemisphere has occurred (Hochstein et al. 1995) and progressive expansion of numerous supraglacial and proglacial lakes has been recorded from this time (e.g. Hochstein et al. 1995, Kirkbride and Warren 1999) (see Fig 27).

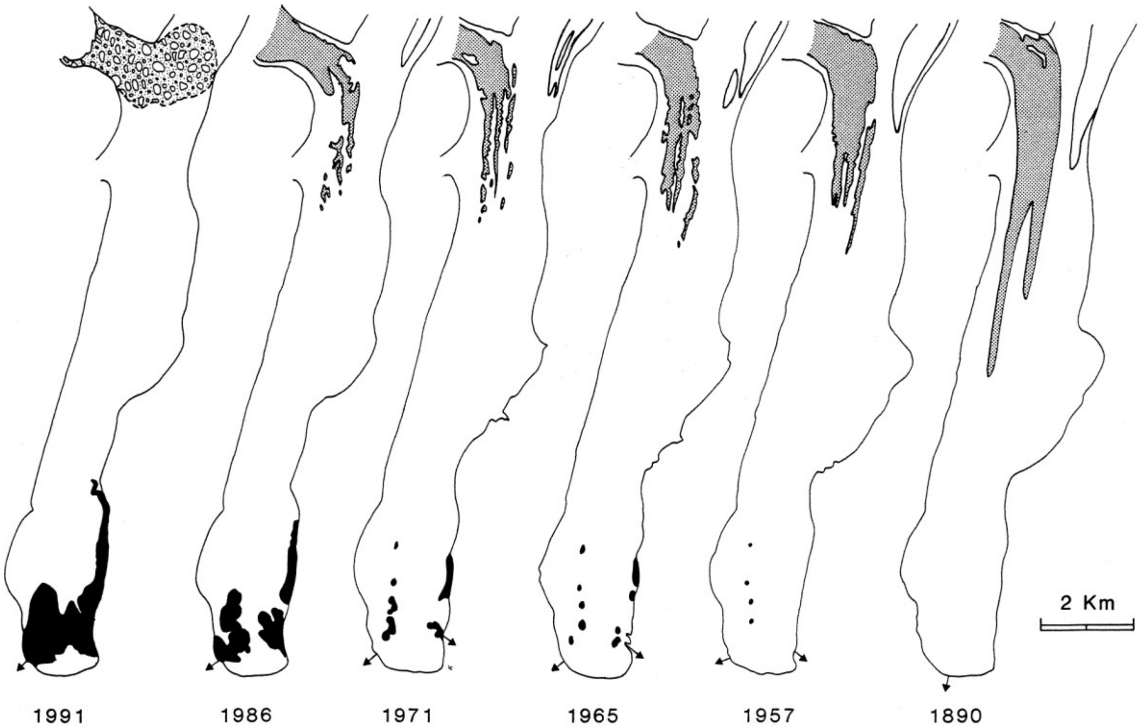

Figure 27: Map showing the upglacier extension of the debris-covered area of Tasman Glacier between 1890 and 1991. The bare-ice area is shaded, lakes are shown in black. The rest of the glacier surface is debris-covered. The 1991 map shows the area of a large rock avalanche deposit which fell in December 1991. Arrows indicate locations of contemporary outwash streams. North to top. Source: Kirkbride and Warren (1999)

Aerial photographs from 1986 (Fig. 28) and Landsat imagery from 1989 showed that melt ponds had become quite extensive, although they had yet to coalesce into a single body of water (Purdie et al. 2016). Supraglacial and proglacial lakes gradually coalesced into one large lake, named the "Tasman Lake" (Purdie and Fitzharris 1999). By December 1990, the newly formed Tasman Lake covered an area of around 1.65 km^2 (Purdie et al. 2016).

When the Tasman Lake was formed, iceberg calving started and became the dominant ablation mechanism of the terminus area. Glacier retreat thus became largely decoupled from climatic influences, initiating independent cycles of advance and retreat (Purdie and Fitzharris 1999). Changes in the terminal position then reflect local topographic and bathymetric controls (Warren 1991). Since the first observations in the end of 19th century the glacier velocity has been decreasing and much of the ice close to the glacier terminus had almost ceased to move by 1986. However, this extensive stagnating zone has been reactivated with the appearance of proglacial lake. The velocity around the terminus increased leading to an extensive crevassing. The new crevasses indicate extending flow towards the ice front related to the growth of the ice-contact lake, and indicate that the velocity increase is not simply a short-term or seasonal effect (Kirkbride and Warren 1999).

With the onset of calving in early 1990s a dramatic retreat commenced and so the growth of the Tasman Lake. In April 1993 the lake covered an area of 1.95 km^2, when its depth was approximately 120 m (Hochstein et al. 1995). By 2008 Tasman Lake increased to 5.96 km^2 and attained a maximum depth of 240 m. In April 2013 Tasman Lake covered an area of 6.88 km^2, was 4.9 km in length, and had a maximum depth of 241 m. By

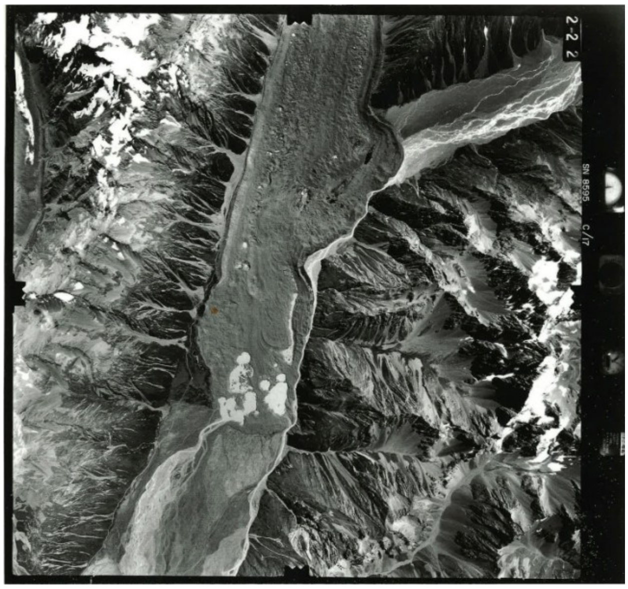

Figure 28: Areal photograph of lower Tasman Glacier from 1986 showing the growing size of melting ponds and the iniciation of Tasman Lake formation. Photograph supplied by the geography department of the Otago University, Dunedin.

May 2014, lake area had increased to 7.12 km^2, and was 4.97 m long. The maximum depth in 2014 was 247 m (Purdie et al. 2016). The lake bathymetry in 2013 (picture A) and in 2014 (picture B) is shown in figure 29.

While the Tasman Glacier proglacial lake expanded at a steady rate between initiation and the mid-1990s (Quincey and Glasser 2009) the second half of 1990s and the 21st century has been characterised by periods of enhanced glacial retreat and thus lake enlargement. The initial growth was predominantly focused around the terminus of the glacier and, subsequently, along the true left glacier margin, up to a point approximately 2 km from the terminus. The fast melting on the eastern side of the glacier (see Fig. 27) is a result of the through flow of the "warm" groundwater from Murchison valley through the alluvial gravel aquifers beneath the lateral moraine (Chinn 2009).

More recently, the lake has expanded rapidly westwards, mainly by calving from the active glacier front, such that the lake has replaced the majority of the lowermost 4 km of the glacier tongue and the lake area has doubled between 2000 and 2007

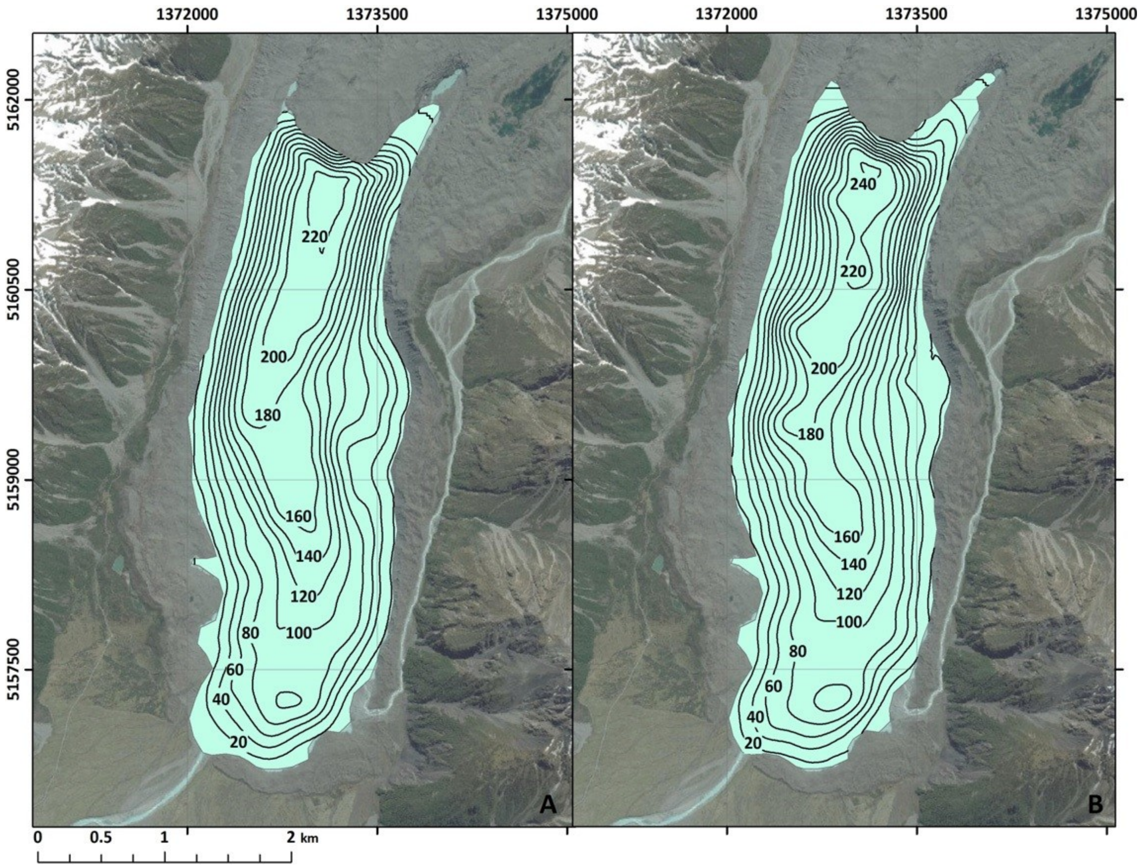

Figure 29: Bathymetric maps for Tasman Lake in A. 2013 and B. 2014. Source: Purdie et al. (2016)

(Quincey and Glasser 2009). The sudden increase in lake area between those years can be attributed solely to a very rapid disintegration of the western half of the calving glacier tongue, which had not been predicted by previous studies and is interesting in style.

The retreat of Tasman Glacier between 2000 and 2008 has occurred in two distinct periods: an initial period of relatively slow retreat prior to 2006, followed by a secondary period of rapid retreat between 2006 and 2008 (Dykes et al. 2010). Terminus full width retreat for the period 2000–2006 occurred at a rate of 54 m a^{-1}, accelerating to 144 m a^{-1} during 2006–2008. During the period 2000–2006, the controlling process of ice loss at the terminus was iceberg calving resulting from thermal undercutting. In contrast, the retreat between 2006 and 2008 was probably controlled by buoyancy-driven iceberg calving caused by decreased overburden pressure as a result of supraglacial pond growth. As a result, the surface area of Tasman Lake has increased by 86% over the period 2000–2008, with lake volume increasing by 284% between 1995 and 2008. In 2008 the volume of Tasman Lake was 510×10^6 m^3 (Dykes et al. 2010). The maximum terminus retreat occurred during a 9 month period between 29/04/2006 to 23/01/2007 (see Fig. 30), when the true right of the glacier retreated of about 1600 m. This equates to a mean retreat of 5 m d^{-1}, though in reality, the terminus has retreated non-linearly over these shorter timescales.

Since 2008 the overall rate of calving retreat has slowed to around 90 m a^{-1} and

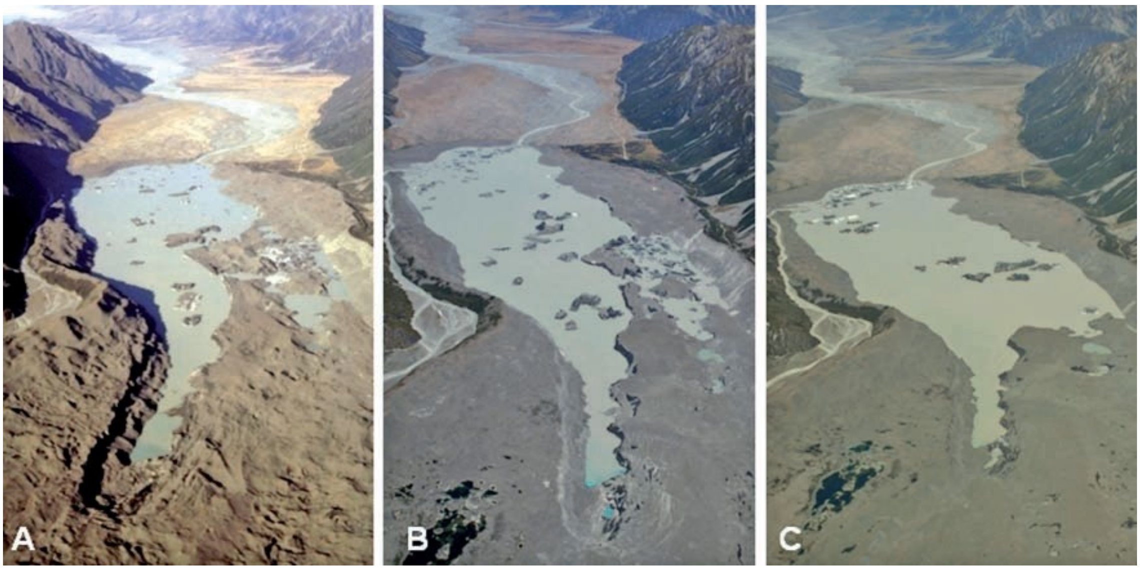

Figure 30: Oblique aerial photographs showing the disintegration of the lower Tasman Glacier terminus between (A) March 2006, (B) March 2007, and (C) March 2008. Photos: S. Winkler, Source: Dykes et al. (2010).

the further deepening towards the terminus became the most notable change in terms of the geometry of Tasman Lake. At the moment, Tasman Lake is by far the largest proglacial/supraglacial lake in the Southern Alps (Dykes et al. 2010) and the Tasman Glacier continues to retreat and the lake expands into the over-deepened glacial trough (Purdie et al. 2016). The size of Tasman Lake, position of Tasman Glacier terminus and other properties were studied during 2014 in person by the author (Fig. 31).

The formation of proglacial lake has also its sedimentary implication. All the debris carried by the glacier was suddenly captured by the lake and the sediment delivery to Tasman River dropped dramatically. The sedimentary river regime immediately changed from aggradation to incision of its channel. The down-cutting was slow to start, only 1 m in the first 10 years, but since then the channel has been incised by another 22 metres (Chinn 2009). Currently (2014), the elevation of the outlet and lake (705 m a.s.l.) are similar to the elevation of the proglacial area where the Tasman River starts to braid 1 km downstream. Therefore, no further incision would be expected unless the base level of the river drops downstream (Purdie et al. 2016).

Chinn (2009) suggests that the Tasman Glacier is some 25-30 % larger than what it needs to be to regain equilibrium with the current climate. To reach the equilibrium, the front of Tasman Glacier has to retreat back to somewhere near the Ball Glacier, where the valley floor is so deep that the ice will be floating. However at this location the Hochstetter Glacier supplies the majority of the ice which must prevent further retreat. One possible scenario is that the upper trunk may collapse and develop a second lake above the Hochstetter. Today the glacier has lost only half the area it needs to loose to regain equilibrium. This implies that the lake has yet to double in size by continuing to grow with another 30 – 40 years of melting (Chinn 2009).

Figure 31: Between 1990 and 2014 the Tasman Glacier retreated significantly and the length of the Tasman Lake increased from 1.6 to almost 5 km. The lower glacier tongue is strongly debris-covered slowing the ice melt from the surface. Inner sides of the lateral moraines are higly unstable leading to regular mass-movements to the lake. The lake surface is currently >100 m bellow the crest of lateral moraines. Source: Archive of the author, 2014.

The other debris-covered glaciers

of the Southern Alps have experienced a similar evolution. Whereas the rate of terminal retreat and lake enlargement slowed down at some of the glaciers (e.g. Godley, Hooker and Ramsay Glaciers) reaching a form of balance and/or responding to positive net balances during the past few decades, the development at Tasman Lake is different. Mueller and Murchison glaciers, on the other hand, have just entered a period of fast lake growth and will most certainly experience comparable fast retreat rates in the near future (Dykes et al. 2010).

6 GLOFs in New Zealand

6.1 New Zealand GLOFs from the global perspective

Glacier thinning and retreat over the past century has led to the formation and growth of lakes at the margins of glaciers and moraines in all high mountain regions of the world (IPCC 2013). One of the results of such a glacial retreat has been an increase in the number and size of proglacial lakes dammed – in many cases – by unstable glacial moraines, further increasing the potential threat of glacial lake outburst floods (GLOFs) to occur (e.g. Samjwal et al. 2007). GLOFs are not only of a geomorphic interest but also a matter of concern for economic and life losses in the river valleys (Campbell et al. 2005).

Mountain lake outburst floods have been studied in most world mountain regions. GLOFs in the Swiss Alps were studied e.g. by Haeberli (1983) and Huggel et al. (2004), in Scandinavia by Breien et al. (2008) and in Iceland e.g. by Björnsson (2003). The scientific research and education regarding GLOFs in Hindu Kush Himalayan (HKH) region is overarched by International Centre for Integrated Mountain Development (ICIMOD). ICIMOD is trying to support regional transboundary programmes between eight member countries of the HKH: Afghanistan, Bangladesh, Bhutan, China, India, Myanmar, Nepal, and Pakistan (see e.g. Campbell et al. 2005, Samjwal et al. 2007). GLOFs in Tien-Shan mountains have been well studied by Janský et al. (2010), the Peruvian Andes by e.g. Emmer et al. (2016), Patagonian Andes by Harrison et al. (2006) and GLOFs in Alaska by Post et al. (1971).

Despite the growing attention related to GLOFs hazards worldwide, scientific studies about New Zealand GLOFs are very sporadic. But not only scientific studies! No break of a glacial moraine was described by neither scientists, nor media. And a dam overtopping was recorded just in few cases, however, mostly just a brief description exists. Within the Southern Alps, outburst floods from Franz Josef Glacier are between the best studied ones (Davies et al. 2003 and Goodsell et al. 2005). However, Franz Josef Glacier has no proglacial lake and the outburst floods are limited just to a release of englacial waters. Furthermore, in some cases the flooding is probably caused just due to re-routing of subglacial channel to the supraglacial position and thus should not be labelled by the term “outburst” (Goodsell et al. 2005).

If we leave the glacial lakes behind and rather focus to other lake types, the number of studies increases and the studies became much more detailed. The first to mention is a lake surrounded by glaciers in the centre of the North Island – Crater Lake on the top of Mt Ruapehu. While the Crater lake on Mt Ruapehu has its volcanic origin, and this thesis focuses just on outburst floods from glacial lakes, Crater Lake of Mt Ruapehu will not be studied here into a detail, however, it will be discussed in section 10.3.3.

Between the other outburst floods, the ones from landslide-dammed lakes (caused by the natural blockage of river channels by hillslope-derived mass movements) are amongst the most obvious and widely recognized of such features in New Zealand (Korup 2004). Due to the threat that landslides pose to glacial lakes, the effect of landslides will be briefly discussed in discussion section.

A maximum effort has been made to research all the information about previous GLOFs events in New Zealand. While online scientific databases and libraries provided studies just about some of above mentioned events, personal communication had to take the place. E-mails were sent to various research organisations like NIWA, DoC, University of Otago Dunedin, Victoria University of Wellington (VUW) and to many scientists personally. Some parts of those personal communications are shown bellow and reflect the poor scientific interest about GLOFs in New Zealand.

> *"...As far as I'm aware there isn't much published information about glacial lake outburst floods in New Zealand. There are quite a few papers relating to flood events caused by landslide dam failure (and there are a large number of landslide dam's which have been caused by the recent Kaikoura earthquake)..."* (Richard Measures, NIWA scientist, personal communication 2017)

> *"...GLOFs have not been frequent in NZ, at least in historical time. I know few works dealing with this topic and probably you already found them..."* (Pablo Iribarren, VUW PhD student, personal communication 2014)

Since 1970s many new glacier lakes were formed and rapid lake expansion has occurred at the termini of many larger glaciers, where LIA moraines form natural but unstable topographic barriers (Irvine-Fynn 2015). Some of the lakes are still rapidly growing (see section 5). A further glacier retreat can lead to the formation of more glacial lakes, both dammed by a moraine and by a glacier tongue (Chinn 2009). Together, these factors produce ideal conditions for the generation of glacier outburst floods (Irvine-Fynn 2015). Scientific studies from all over the world confirm the fact that GLOFs are a common phenomenon and the release of water can have serious consequences, therefore we can assume that the assessment of New Zealand glacial lakes is of high importance.

However, it is important to realise, that the glacial lakes significantly differ throughout the world and so does the degree of the hazard from GLOFs. Various geometric and geomorphic properties influence the susceptibility of a lake to a GLOF. Even not many scientific studies have been dedicated directly to GLOFs hazard assessment in New Zealand, whole range of characteristics have been studied extensively on New Zealand proglacial lakes (e.g. Kirkbride 1993; Hochstein et al. 1995; Warren and Kirkbride 1998; Chinn 2009; Dykes et al. 2010; Purdie et al. 2016). However, all the studies focus just on glacial lakes of Mt Cook region and other glaciated areas remain almost unnoticed.

But even in Mt Cook region just few comments have been done regarding the susceptibility to GLOFs. Those statements are mostly based on the overall stability of a certain dam type rather than a holistic assessment. Even small terminal moraines are evident by many large proglacial lakes of Mt Cook region, the lakes are not directly impounded by the moraine material. Rather, the lake basins are developing behind large Holocene outwash heads (Kirkbride and Warren 1999). Quincey and Glasser (2009) therefore suggest that this fact removes the possibility of catastrophic dam failure. Also Allen et al. (2009) suggest, that natural dam failures are considered unlikely, but point out the fact, that with the growth of the lakes increases also the potential for large magnitude impacts from ice, debris, or rock, producing displacement waves, overtopping and flooding.

6.2 Kea Point outburst flood

One of the oldest GLOF described in New Zealand history is the outburst flood at Kea Point close to Aoraki/Mt Cook Village. The flood occurred in 1913 in response to exceptional rainfall events. Waters from Mueller Glacier overtopped the LIA moraine at Kea Point (see Fig. 32 and 33), sending debris through a narrow chute and creating an alluvial fan between the LIA moraine and Mt Ollivier (Irvine-Fynn 2015).

The area between Kea Point and alluvial fan has been studied by Irvine-Fynn et al. (2015), however the research has yet not finished, therefore just preliminary results are available. The authors mapped the flood channel, taking two full along-channel profiles and multiple channel cross sections. These data were supplemented with ground-based digital photography and measurements of flood channel sedimentological characteristics. A total of ~1000 clasts across the study area were recorded providing details of roundness, size and orientation. Due to the complex morphology of the GLOF channel and surrounding terrain, imagery from the initial survey were not sufficient to resolve the entire channel course, therefore further works yet need to be undergone. Currently the digital elevation model is being developed and the potential for numerical modelling of the flood is being explored (Irvine-Fynn et al. 2015).

The 1913 GLOF also destroyed the original Hermitage Hotel placed bellow the moraine of Mueller glacier (see fig. 32 and 34 for location). Although reports indicated the flood that destroyed the Hermitage was triggered by a rainfall event alleged to have provided 24” (ca 600 mm) of rain in 24 hours, further investigation revealed there were other, smaller flood events during 1913 (Irvine-Fynn et al. 2015). According to the

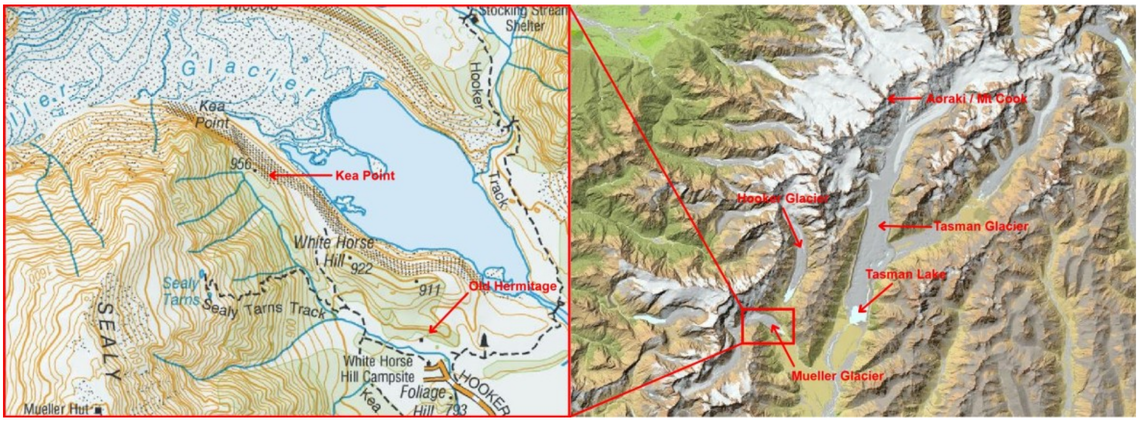

Figure 32: Location of Kea Point within Aoraki / Mt. Cook National Park. The GLOF at Kea Point is the only documented GLOF from proglacial lake in New Zealand. Source of the maps: Doc. Labels: Personal work

Digital Collection of Christchurch City Libraries the hotel was damaged by flood in January 1913, and two months later was destroyed beyond repair by a second flood (Christchurch City Libraries).

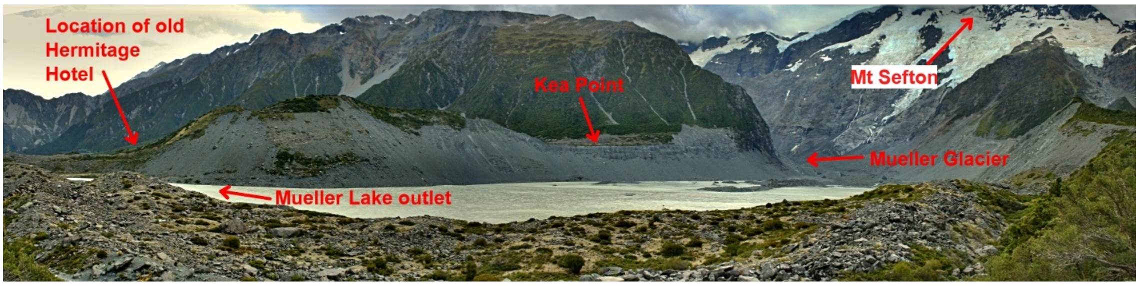

Figure 33: Panoramic photograph of Mueller Lake and Kea Point in 2014. Exactly 101 years ago (this photograph was taken) a GLOF exited Mueller Glacier through Kea Point and destroyed old Hermitage Hotel placed behind the LIA moraine of Mueller Glacier. Photo and labels: Archiv of author.

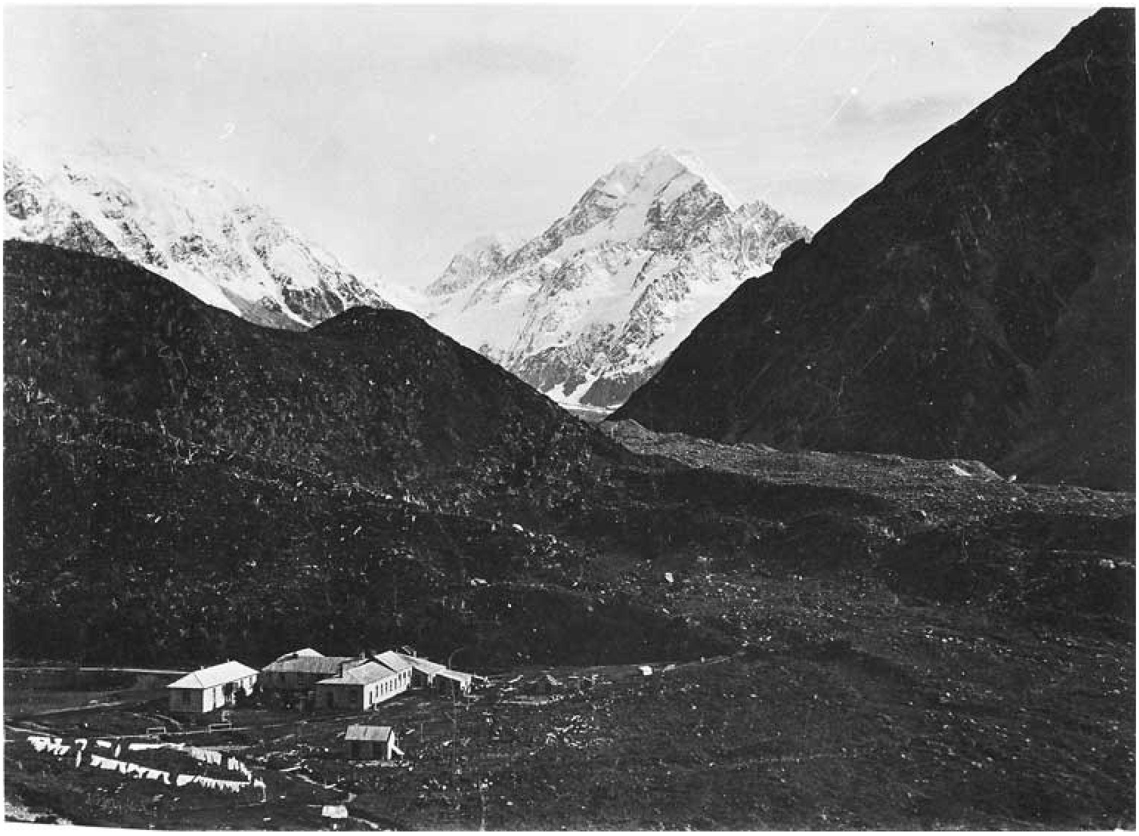
Figure 34: Mt. Cook and the old Hermitage Hotel before it was destroyed by flooding in 1913 (photo ca. from 1910). Source: Christchurch City Libraries

6.3 Franz Josef Glacier outburst floods

Franz Josef Glacier, Westland, New Zealand (see section 5.4.1 for location and description of the glacier), has a history of catastrophic sediment-laden outburst floods. However, the sediment deposits and mechanisms differ from those commonly associated with other glacial outburst floods (Davies et al. 2003). In the absence of lakes, the outburst events at Franz Josef are clearly sub- or englacial in origin, but contrary to other sub- or englacial outburst floods, those at Franz Josef are associated with extreme rainfall events. The flow regime of glacier meltwaters exhibits sustained annual runoff, a high frequency of large floods, and a dominant periodicity of several days rather than diurnal (Davies et al. 2003). Outburst floods from the Franz Josef Glacier are relatively common, but only a limited number have been documented in sufficient detail to reveal something of the processes (Table 4). Many more events are likely to have gone unrecorded (Goodsell et al. 2005).

We can see that all of the floods are associated with precipitation events and except of the supraglacial flood in 1981 all the events occurred during the advancing phase. The supraglacial flood in 1981 was the only well-described outburst event at Franz Josef not associated with a large sediment emplacement (Davies et al. 2003) and the only supraglacial flooding recorded at Franz Josef Glacier (Goodsell et al. 2005). The other outburst floods have frequently delivered, mobilized and deposited sediment in the valley (Davies et al. 2003). During the most significant glacier advance between 1983–1999, the upper Waiho valley aggraded substantially (of the order of 15 m during this period). Aggradation decreased down valley, so that the outwash surface of the upper valley steepened. The river is usually incised into this surface, but major sediment delivery events cause the river to aggrade and rework much of the braid plain surface. Much of the aggradation during advance appears to have occurred during outburst events (Davies et al. 2003).

Davies et al. (2003) recognize three distinct modes of outburst from Franz Josef Glacier. First, the water can be released by surface overflow, second, the water can escape due to expulsion of the blockage and third the flood can occur thanks to uplift of the glacier near the snout. All of those modes are a result of a pressure build up caused by blockage of englacial conduits. Each mode produces a different type of flood and deposits different type of sediments.

Table 4: Summary of recorded outburst floods and ice collapses at Franz Josef Glacier. Source: Goodsell et al. 2005

Date	Source(s)	Description	Type*	Advance or Retreat (Burrows 2001)	Rainfall	Rainfall in township
Summer, 1920s	Alack (1974), Benn (1990)	Collapse of ice at terminal face creating temporary lake and subsequent glacial lake outburst flood.	1	Advance	"One specially wet summer"	No record
Feb 1949	Suggate (1952), Benn (1990)	Large flood infilling proglacial lake with sediment. Large sinkhole appears at base of main icefall.	2/3	Advance	Followed 6 days of heavy rain	No record
19 Dec 1965	Sara (1968, 1974), Benn (1990)	Large-scale flood at snout and western margin following a large collapse of ice at the terminal face.	2	c. 6 months after advance began	280 mm during 16 and 17 Dec	295 mm on 16 and 17 Dec
Late Jan 1967	Sara (1974)	Cave-in of ice at base of icefall.	3	Advance	Following "2 periods of rain" of 920 mm in township	675 mm from 22 to 25 Jan, 162 mm on 30 Jan
Mar 1967	Sara (1974)	Re-opening of hole at base of icefall.	3	Advance	"After another period of heavy rain"	740 mm from 5 to 11 Mar
2 Jun 1981	Marcus et al. (1985)	Sudden debris-laden torrent along western margin of glacier; flowed from 11 a.m. until evening.	4	Retreat/standstill	306 mm in 12 h at Luncheon Rock	36 mm on 2 Jun
Jan 1994	Coates (2002)	Photograph of ice blocks and flooding in proglacial area.	1	Advance	Caused by "heavy rain"	389 mm between 8 and 10 Jan, 601 mm between 19 and 24 Jan
13 Dec 1995	Turnbull (1998)	Outburst flood at terminal face during rainfall-induced flooding.	1	Advance	195 mm on 12, 397 mm on 13 in township	195 mm on 12 Dec, 397 mm on 13 Dec
1990s	M. Melsop (pers. comm. 2003)	Series of four events in which a large hole opened at the base of the icefall, with associated flooding and aggradation of sediment in the Waiho River.	3	Advance	All followed periods of heavy rain	
14 Feb–5 Mar 2003	This paper, Goodsell et al. (2003)	Re-routing of subglacial drainage to supraglacial configuration associated with ice collapse at base of main icefall.	4	Retreat	At Luncheon Rock: 71 mm on 14 Feb, 389 mm between 18 and 21 Feb	80 mm between 12 and 14 Feb, 121 mm between 18 and 21 Feb, 163 mm between 2 and 6 Mar (Fig. 10)

*Flood types: (1) flooding of the proglacial area; (2) ice collapses and flooding at the glacier snout; (3) ice collapses at the base of the main icefall; (4) supraglacial floods.

The overflow through marginal tributaries following closure of the primary conduit rapidly cuts down into the ice and thus relieve the excess pressure in the conduit (Davies et al. 2003). That was the example of outburst flood 1965, 1981 and 1998. During the outburst flood 1981 a stream on the glacier's true left margin was developed between ice and lateral moraine. Within minutes, the stream grew from non-existence into a steep torrent which was discoloured due to its high sediment load. The torrent cut down into the ice and undercut the marginal portions of the glacier. Blocks of ice broke off and were carried away down the torrent. The torrent had eroded away a portion of the glacier margin approximately 50 m wide and about 25 m deep, and deposited a bed of coarse sediment (Marcus et al. 1985). Deposition from such events is limited to development of a fan at the snout, composed of material eroded from the glacier margin (Davies et al. 2003).

In cases where the outburst results from removal of the blockage, materials are ejected exclusively from the portal. Coarse material is immediately deposited on exit. Such conditions are considered responsible for deposits of massive well sorted, rounded, imbricated boulders emplaced in 1997 and 1998 (Fig. 35).

Figure 35: Boulder deposit of May 1997. This deposit extended about 200 m along the true left of the proglacial river. Source: Davies et al. 2003

Where the glacier is lifted, it does not appear to induce wholesale drainage or widespread surging of the glacier. This suggests that uplift is limited – perhaps on the order of a few centimetres, limiting the size of transported material. Uplift thus generally generates hyperconcentrated flows leaving gravel deposits of constrained size. Nevertheless, the overall volume of material removed may be large, given the large pressure gradient, the extensive area of the bed involved, and the possible release of previously confined sediments. Uplift and removal of basal support may also account for massive ice break-up during outbursts (Davies et al. 2003).

However, the outburst floods from Franz Josef can be also categorised differently. Goodsell et al. (2005) classified outburst floods from Franz Josef Glacier to four categories: (1) flooding of the proglacial area; (2) ice collapses and flooding at the glacier snout; (3) ice collapses at the base of the main icefall and (4) supraglacial floods. All of those types have been recorded in Franz Josef outburst history and in some cases have been overlapping each other. Gentle, small magnitude floods have been alternating with dramatic flood and ice collapses. Alack (1974) in Goodsell et al. (2005) described how on a daily trip to view the glacier during “one specially wet summer”, the Waiho River was “boiling from under the whole length of the ice face”, which was estimated at c. 30 m high at the time. A “gigantic slice of the face, the full width of the glacier” collapsed from the terminus of the glacier. The fallen ice blocked the river as it became wedged, and a lake rapidly formed behind. The ice dam burst catastrophically soon afterwards, with large blocks of ice rising into the air: “Great trees were smashed, others uprooted as the frozen bombardment struck them ... for many days after the storm we were able to locate masses of ice well up in the forest” (Alack 1974 in Goodsell et al. 2005). Similar event was recorded also by Sara (1974) in Goodsell et al. (2005).

Probably the best described outburst event of Franz Josef Glacier is the flooding during February and March 2003 when ice collapses at the base of the main icefall (type 3 above) and supraglacial flooding (type 4 above) occurred. Goodsell et al. (2005) studied the event and suggested several scenarios. The schematic interpretation of the most preferred one is shown in figure 36. The authors interpreted the event as the temporary re-routing of a major subglacial channel to a supraglacial position. This re-routing is thought to have followed the collapse of ice above the channel, which blocked the usual subglacial drainage route. Close to a semi-permanent protruding bedrock feature locally known as the Black Hole, the ice-collapse uncovered another part of the bedrock and thus called New Black Hole (best obvious in figure 36, C and D).

The development of such a feature is probably controlled by the topography of the bedrock. The bedrock slope exposed as the New Black Hole was c. 60-90°, which is steeper than the surface slope of the ice flowing over it before collapse. This configuration is likely to lead to ice thinning and formation of cavities between the ice in the icefall and the underlying bedrock. Thinning in the layer of ice moving over the

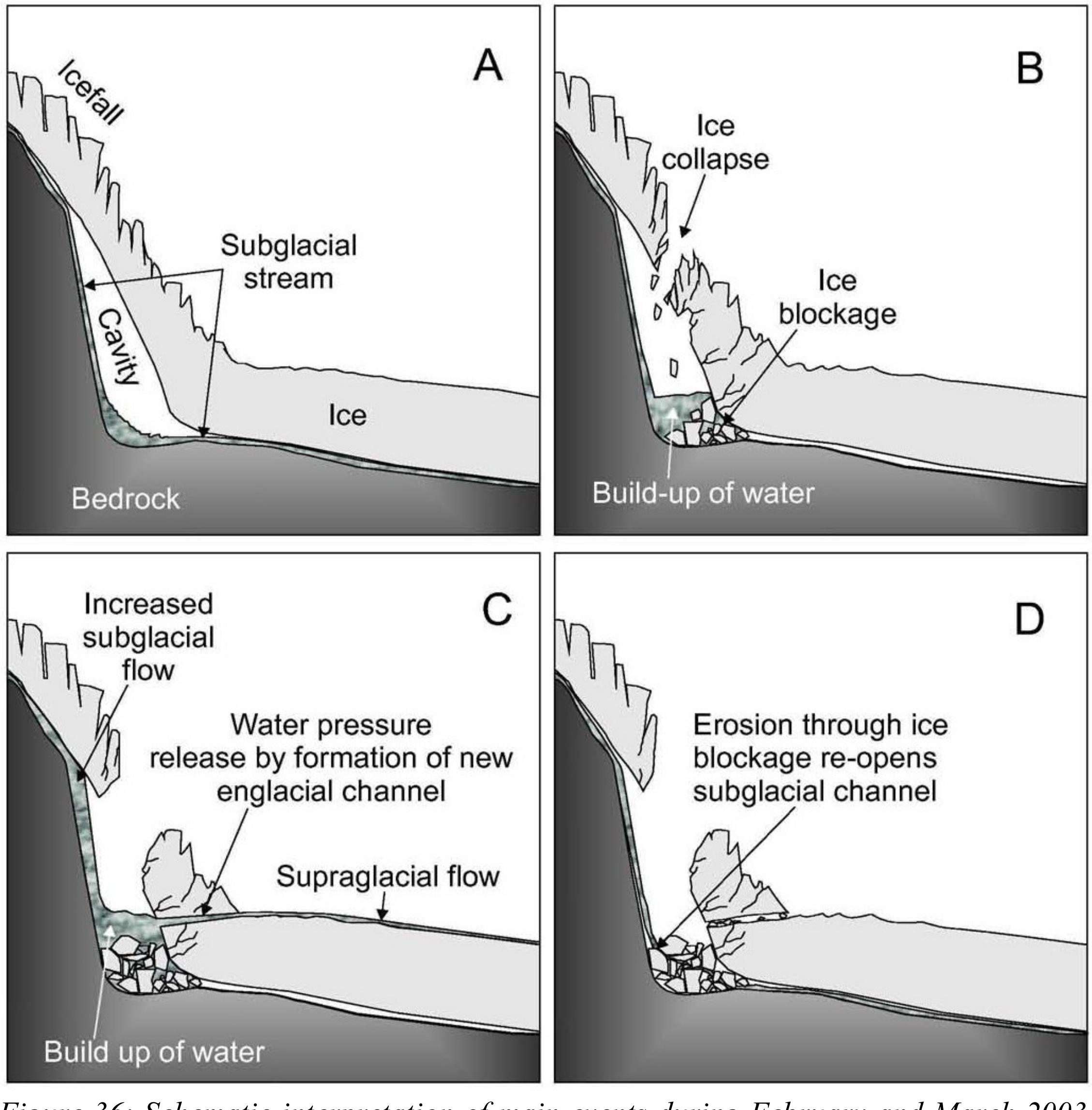

Figure 36: Schematic interpretation of main events during February and March 2003. A, Normal configuration: steep bedrock beneath the base of the main icefall, with ice flowing down detached from the back wall, and subglacial stream beneath the ice. B, Increased subglacial flow in response to rainfall together with collapse of thinned ice into cavity, blocking the subglacial stream, causing pooling. C, Pressure of water build-up forces a route for the subglacial stream over the surface of the glacier. D, Subglacial stream erodes down to a subglacial position. Stages B, C, and D are repeated on a variety of scales until a permanent subglacial channel evolved once more. Source: Goodsell et al. (2005)

bedrock cliff at Franz Josef Glacier could cause the ice to fail and internal collapse to occur (Goodsell et al. 2005).

While Davies et al. (2003) suggest that all outbursts from the Franz Josef Glacier are associated with intense runoff, Goodsell et al. (2005), examining the 2003 outburst flood, found, that the precipitation values were rather normal than extreme, similarly as the rainfalls measured before the 1965 event. The precipitation forcing does not appear to show such a clear correlation, because many large storms (e.g. the three-day total of 1810 mm in March 1982) (Henderson and Thompson 1999) have not induced outbursts from the Franz Josef Glacier (Davies et al. 2003). Hydrologic forcing is therefore a necessary but not sufficient condition.

Davies et al. (2003) believe that the association between outburst events and advanced ice position also indicate a necessary condition, because as the glacier has advanced across its own forefield gravels, the primary drainage conduit has developed a reach of negative slope favouring the blockage. However, Goodsell et al. (2005) argued that there is no clear relationship between glacier advance, retreat, and the formation and extent of “Black Holes” near the base of the main icefall, suggesting that outburst floods related to ice collapse near the base of the main icefall can occur both during the retreat and advance phase. However, both Davies et al. (2003) and Goodsell et al. (2005) agreed that the majority of outburst floods at Fanz Josef has occurred during advanced phases and after a rainfall.

While Franz Josef Glacier is a popular tourist destination, and the glacier termini is well accessible, the Department of Conservation (DoC) is putting a big effort to assure public safety. The 2.7 km long access track is in most places designed to allow quick escape to higher ground, and dangerous places (mostly due to rockfalls) are signposted. The location of the final lookout point has been shifted many times according to the glacier termini position, and is clearly signposted, and enclosed within a rope fence to avoid visitor access closer to the glacier front. Few people have already lost their lives by going to the dangerous places of the glacier front, and those stories are well explained at the lookout point to protect other lives. Further understanding of the process related to outburst floods should lead to discoveries of some regularities enabling more precise decision making. A special attention needs to be dedicated to hazard assessment education to protect lives even on the places where no signposts are present.

6.4 Maud Lake outburst flood

Between 1991 and 1996 several large rock and ice avalanches occurred in the Aoraki/Mount Cook National Park (e.g. McSaveney 2002; Hancox and Thompson 2013). In December 1991, rockfalls began at Mount Fletcher (2467 m) 30 km northwest of Mount Cook (Fig. 37). They continued until the ridge north of Mount Fletcher fell in rock avalanches in May and September 1992. They dropped 1440 m along similar 3.8 km paths in 50 s, generating magnitude 2.8 and 2.7 earthquakes and terminating in a proglacial lake at the terminus of Maud Glacier (McSaveney 2002), in this thesis named "Maud Lake". The speed of the rock avalanche was thus estimated to be ~120 m/s at the foot of the slope and ~65 m/s at the bend on Maud Glacier. While the volume of the two collapses was estimated to be ~11 × 10^6 m^3 and both rock avalanches went into the lake, it remains unknown how big proportion of the debris entered the proglacial lake at the glacier terminus (McSaveney 2002).

McSaveney (2002) described the story of two witnesses staying overnight at Godley Hut near the glacier terminus. They watched several rockfalls from the buttress already during the day on May 2, 1992. The main collapse was at night. Remembering descriptions of showers of sparks at Mount Cook a few months earlier, the witnesses went outside to look but saw nothing: even the stars were obscured by thick dust. The following morning, with dust still hanging in the valley, they found icebergs stranded ~20 m above the lake near the hut. In making their way down the valley, they found the four-wheel-drive access track partly washed away. Those are some of the evidences proofing that The Mount Fletcher avalanches created flood waves from "Maud Lake" (unnamed lake in the literature) at the terminus of Maud Glacier (McSaveney 2002).

But not only the icebergs and 4x4 track, scattered sparse grassy vegetation was widely ripped from around the lake outlet over a width of ~600 m across the valley. The braided gravel bed of Godley River was extensively scoured for several kilometres (see the dotted area in fig. 37), and the flood travelled 45 km to Lake Tekapo, where it raised the lake level for ~90 mm (McSaveney 2002). From the lake level rise the volume of displacement wave was calculated. The first rock avalanche displaced 7.8 × 10^6 m^3 of water from the lake, and the second, ~5 × 10^6 m^3. The fastest lake rise after compensation for seiching was 16 mm in 1.25 h, equivalent to inflow of 300 m^3/s.

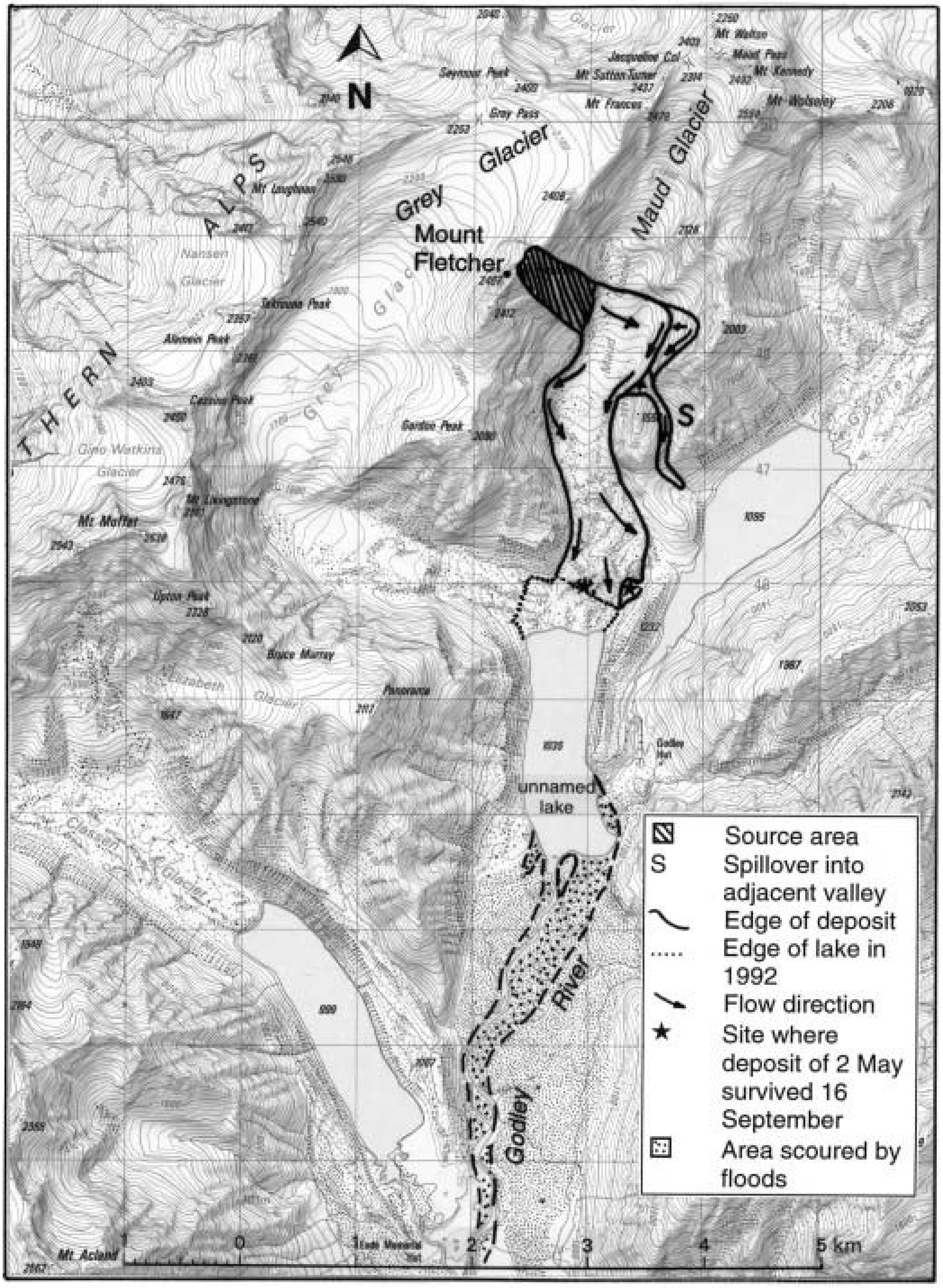

Figure 37: Location of Mt Fletcher, Maud Glacier and an unnamed lake (refered as Maud Lake in this Thesis) within Aoraki / Mount Cook National Park. Source: McSaveney (2002)

However, the peak outflow from the point of impact would have been several orders of magnitude greater. The water arrived over a day and is the water displaced from Maud Lake by an equivalent volume of rock-avalanche debris, including eroded ice. Any surge associated with the avalanche impact had its volume entering Lake Tekapo compensated within the day by correspondingly diminished flow as Maud Lake refilled (McSaveney 2002).

Debris entered the lake across a front of 780 m (Fig. 38), so the volume of displaced water in the second flood could have been contained in a wave ~1.9 km long (assumed triangular section 7 m high). If the first had a similar wavelength, its height could have been ~10 m (the height of the second wave is well shown by silty snow at the shoreline). The witnesses' descriptions of where icebergs were deposited include runup of water and iceberg momentum, and there is uncertainty as to whether "20 m above the lake" refers to a height of 20 m or that ice was stranded 20 m from the lake on the steep shores (McSaveney 2002).

About 13 × 10^6 m^3 was displaced from the lake in the two floods. This is larger than the estimated 11 × 10^6 m^3 that fell from the source, but the latter volume has much uncertainty, and the former includes ice eroded from the substrate (McSaveney 2002).

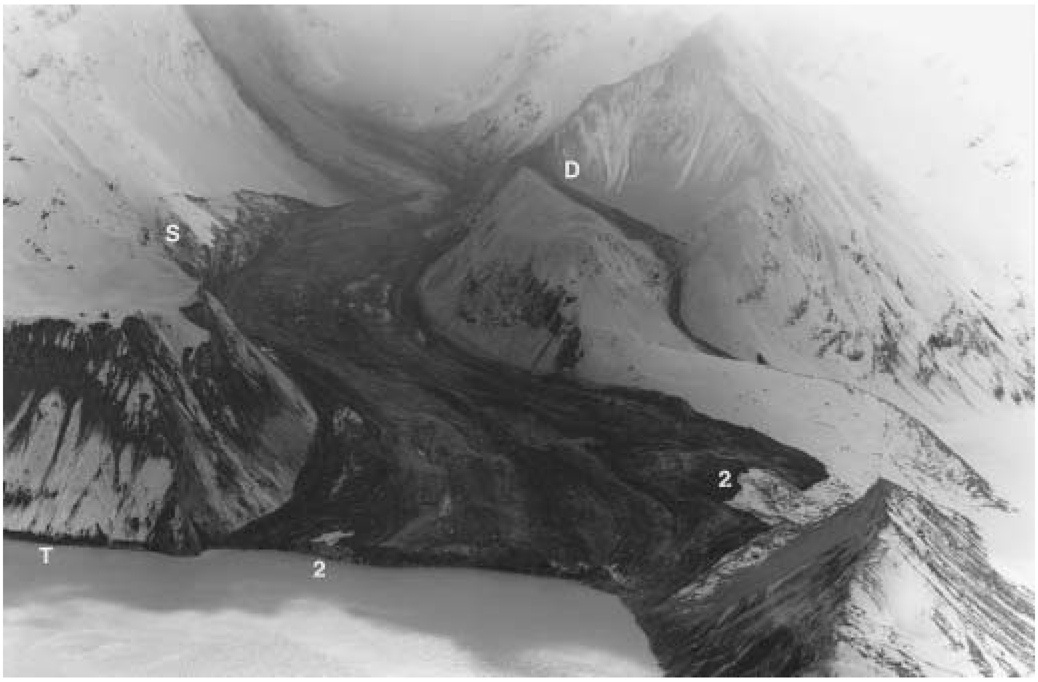

Figure 38. Sinuous path of Mount Fletcher rock avalanche of September 16, 1992. Note thin stream of debris in side valley (D), and superelevation of landslide debris (S) at bend in avalanche flow path on Maud Glacier. Points at 2 are snow-covered remnants of rock-avalanche deposit of May 2. Black trimline (T) to 7 m above lake is from wave generated by avalanche that displaced ~5 × 106 m^3 of water from lake. Source: McSaveney 2002.

6.5 GLOFs modelling in New Zealand

The fact that GLOFs research in New Zealand has been strongly neglected is supported by the fact that the first attempts to model glacial hazards in New Zealand occurred as late as the beginning of 21st century. Allen et al. (2009) presented the first results of glacial hazard modelling in Aoraki/Mt Cook National Park including the modelling of potential GLOFs. However, this study focuses just on one glaciated region within entire Southern Alps and does not describe other highly glaciated regions like the Mt Aspiring National Park.

Allen et al. (2009) studied all the lakes larger than 1500 m^2 within the Mt Cook region regardless their origin or the dam type. They recognized 54 lakes from which most exist at low elevations (750–1000 m). The majority of outlet areas were classified as "debris", with only one non-debris (bedrock) outlet identified. Nearly 70% of terrain contained within outlet channel areas is characterised by slope gradients less than 10°. Because of those characteristics, 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 wave 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. The debris flows were modelled just for the few lakes with steep outlet areas.

Vegetated outlet areas characterise most large proglacial lakes which have formed within low gradient moraine and outwash gravels during the past two decades (Allen et al. 2009). While the vegetation can reduce the erodibility of the channel area, limiting the likelihood of natural dam failure (Allen et al. 2009), the presence of vegetation implies some longer term stability of the outlet area, indicating that no recent disturbances are likely to have occurred. This is exemplified by the Maud Glacier lake, where the current lack of vegetation provides evidence of the flood event that occurred in the year 1992 (see section 6.4).

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 (Allen et al. 2009). Modelling results of Allen et al. (2009) show that the potential floodwaves from selected proglacial lakes of Mt Cook region could influence human activities (Fig. 39 and 40), giving the extent of human infrastructure intersecting with modelled flow paths for the worst case probable maximum runout. Relative to modelled clear water floodwaves, flood triggered debris flows from smaller cirque lakes all appear to terminate well before huts, vehicle tracks, or other infrastructure are reached (Fig. 39 and 40).

Furthermore, 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 (Allen et al. 2009).

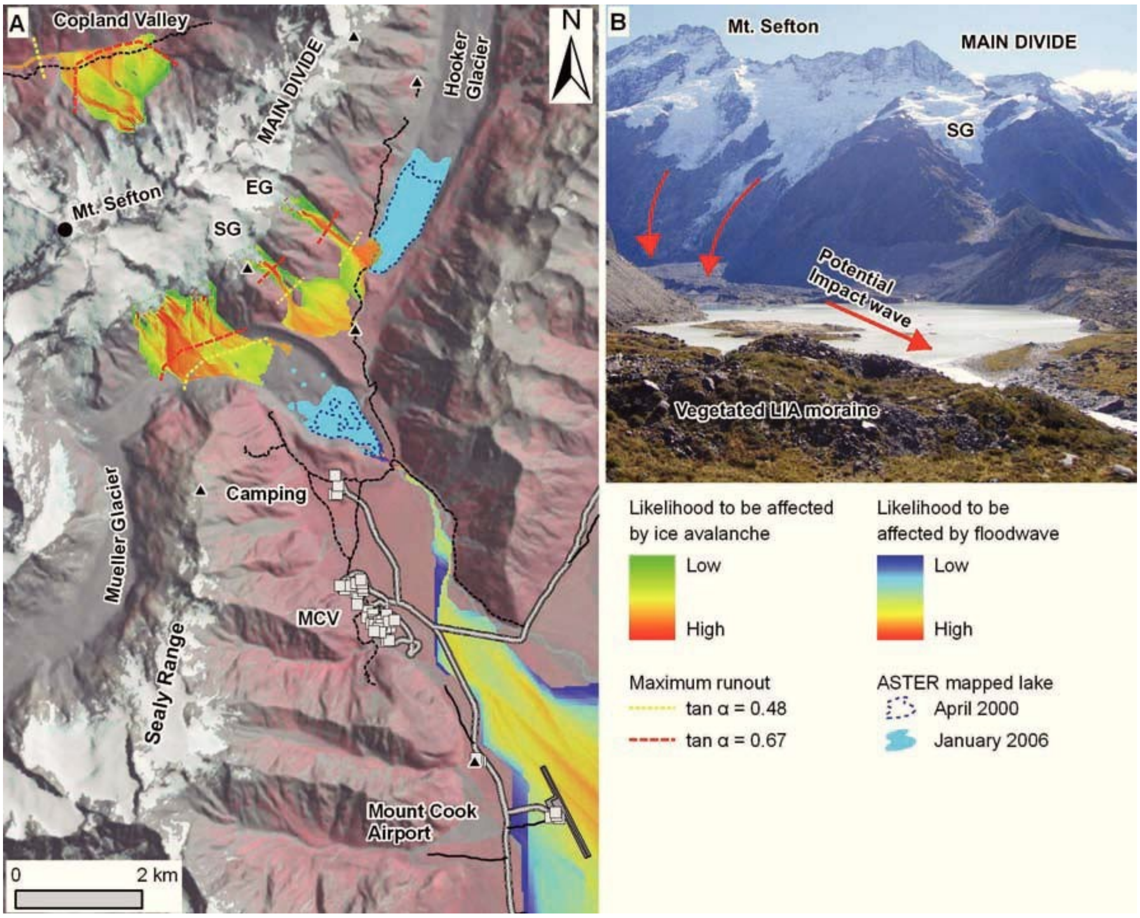

Figure 39: A: Modelling of selected ice avalanches affecting the lower Mueller and Hooker Glaciers, and upper Copland Valley and the potential flood path propagating from the Mueller Lake. B: Outlet area of the Mueller Lake looking towards the main divide, indicating potential ice avalanche trajectories towards the lake from beneath Mt Sefton. Source: Allen et al. (2009).

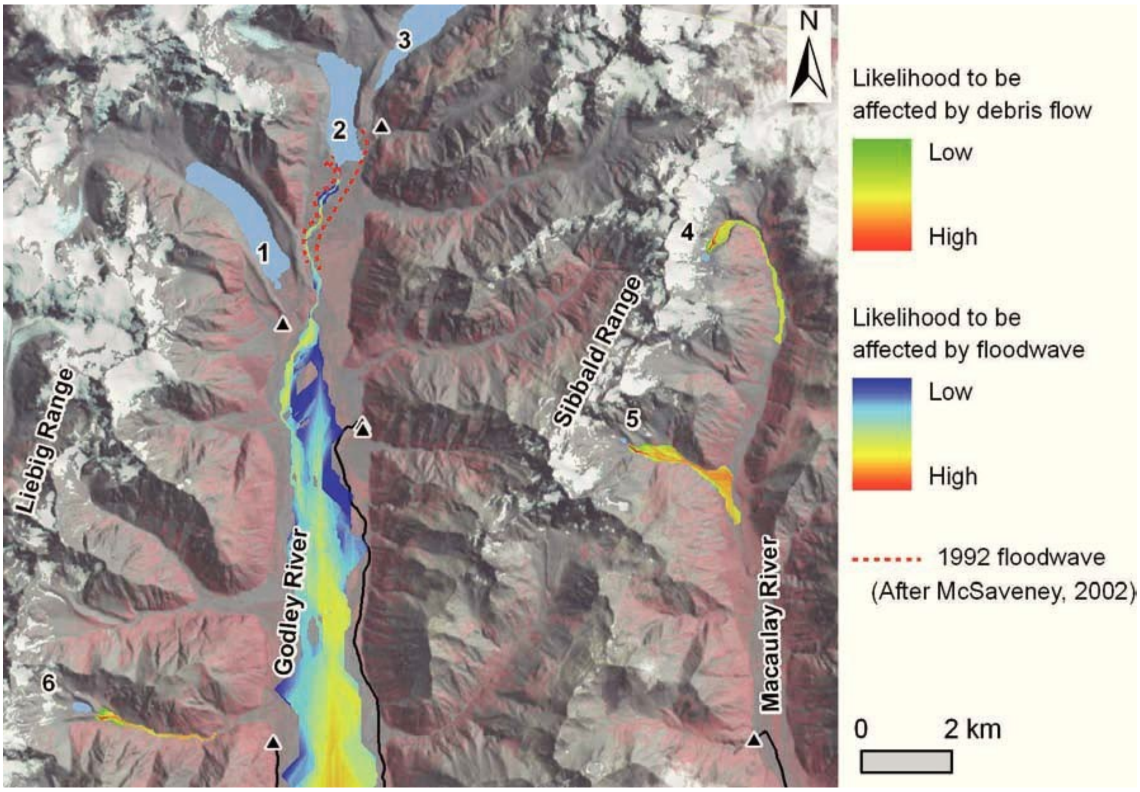

Figure 40: Modelling of potential GLOF triggered debris flow events from 3 unnamed lakes where the dam area and outlet channel is formed within steep morainic debris. Also modelled is the most likely path for a floodwave initiating from the Maud Glacier lake (2). In 1992, a rock avalanche from Mt Fletcher produced a displacement wave from this lake. The background is provided by a 2006 ASTER image. Source: Allen et al. (2009)

7 Inventory of proglacial lakes in New Zealand

7.1 Distribution of proglacial lakes

An inventory of all proglacial lakes in New Zealand was done by using the methods described in section 3.4. The total of 25 proglacial lakes were identified; all of them within the limits of the Southern Alps of New Zealand (Fig. 41). However, the proglacial lakes are to be found just in the central part of the Southern Alps within few kilometres from the Main Divide. The proglacial lake distribution is bound to two main glaciated areas of New Zealand; Mt Aspiring area with 12 proglacial lakes in the southern part (Fig. 42) and Mt Cook area with 13 proglacial lakes in the central part of the Southern Alps (Fig. 43).

Table 5 shows the coordinates, altitude and other geographic details of all 25 proglacial lakes. While 13 (9 from Mt Aspiring area and 4 from Mt Cook area) out of total 25 lakes drain to the Tasman Sea (west of the Main Divide), the other 12 (3 from Mt Aspiring area and 9 from Mt Cook area) drain to Southwest Pacific on the east. In table 5, the proglacial lakes are ordered from south to north and the lakes of Mt Aspiring area are shaded.

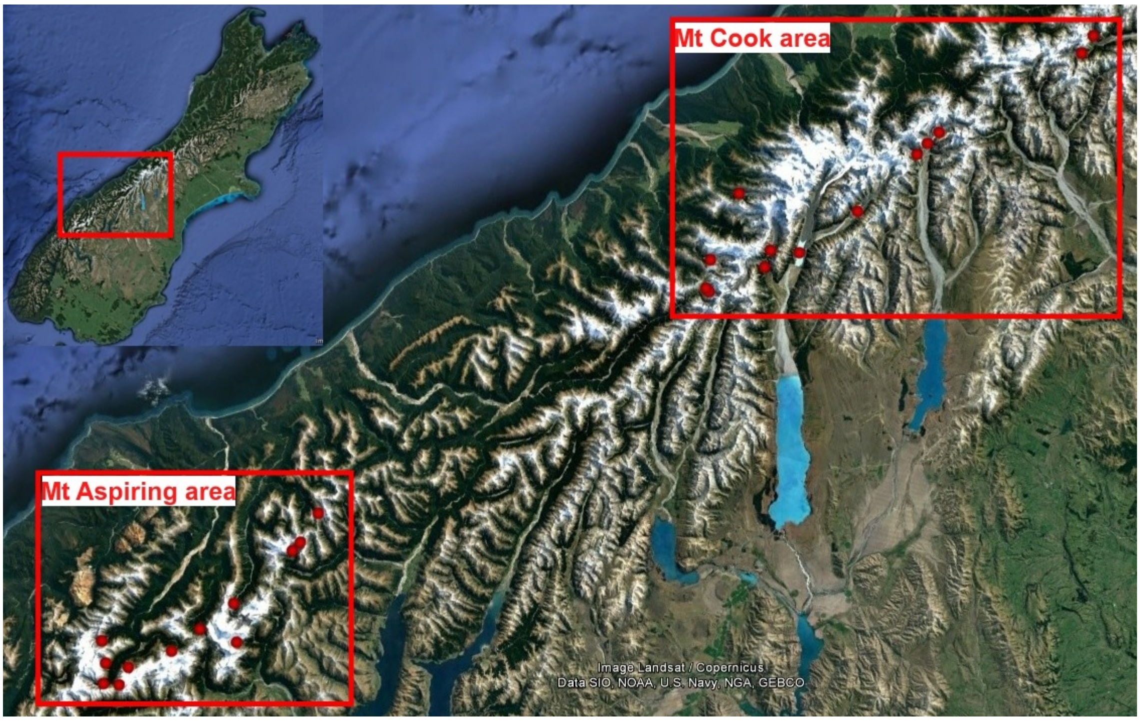

Figure 41: Distribution of proglacial lakes of New Zealand. 25 proglacial lakes have been idetified in two distinct clusters; Mt Aspiring area and Mt Cook area. Source: Google Earth images and personal work.

Figure 42: The detailed satellite image of Mt Aspiring area with all the proglacial lakes labelled and marked by red dots. There are 12 proglacial lakes within this area. Source: satellite image by Google Earth, location of lakes: Personal work

Figure 43: The detailed satellite image of Mt Cook area with all the proglacial lakes labelled and marked by red dots. There are 13 proglacial lakes within this area. Source: satellite image by Google Earth, location of lakes: Personal work

Lake number	Lake name	Main source glacier(s)	Main river catchment	Drainage sea	Approx. Location	Altitude (m)
1	Joe Lake	Joe Glacier	Arawhata River	Tasman Sea	44°29′ S, 168°24′ E	850
2	Victor Lake	Victoria Glacier, Marion Plateau	Arawhata River	Tasman Sea	44°29′ S, 168°27′ E	1180
3	Snowball Lake	Snowball Glacier, Marion Plateau	Arawhata River	Tasman Sea	44°27′ S, 168°28′ E	1320
4	John Inglis Lake	John Inglis Glacier	Arawhata River	Tasman Sea	44°26′ S, 168°25′ E	1040
5	Snow White Lake	Snow White Glacier	Arawhata River	Tasman Sea	44°25′ S, 168°35′ E	1300
6	Kitchener Lake	Kitchener Glacier, Hood Glacier	Matukituki River	Southwest Pacific	44°24′ S, 168°45′ E	1036
7	Williamson Lake	Andy Glacier (Olivine Ice Plateau)	Arawhata River	Tasman Sea	44°24′ S, 168°24′ E	655
8	Waipara Lake	Bonar Glacier	Arawhata River	Tasman Sea	44°23′ S, 168°39′ E	517
9	Volta Lake	Volta Glacier, Therma Glacier	Waiatoto River	Tasman Sea	44°20′ S, 168°45′ E	663
10	Lucidus Lake	Pollux Glacier	Makarora River	Southwest Pacific	44°14′ S, 168°54′ E	828
11	Lake Castalia	Leda Glacier	Makarora River	Southwest Pacific	44°13′ S, 168°55′ E	1120
12	Lake Axius	Axius Glacier	Waiatoto River	Tasman Sea	44°10′ S, 168°58′ E	1538
13	Poet Lake	Poet Glacier	Haast River	Tasman Sea	43°45′ S, 169°56′ E	1200
14	Spence Lake	Spence Glacier	Haast River	Tasman Sea	43°45′ S, 169°57′ E	1320
15	Mueller Lake	Mueller Glacier, Frind Glacier, Huddleston Glacier	Tasman River	Southwest Pacific	43°42′ S, 170°06′ E	750
16	Douglas Lake	Douglas Glacier	Karangarua River	Tasman Sea	43°41′ S, 169°57′ E	950
17	Hooker Lake	Hooker Glacier	Tasman River	Southwest Pacific	43°41′ S, 170°06′ E	870
18	Tasman Lake	Tasman Glacier, Hochstetter Glacier + others	Tasman River	Southwest Pacific	43°41′ S, 170°11′ E	715
19	Murchison Lake	Murchison Glacier, Dixon Glacier, Mannering Glacier	Tasman River	Southwest Pacific	43°36′ S, 170°19′ E	1025
20	La Perouse Lake	La Perouse Glacier	Cook River	Tasman Sea	43°34′ S, 170°01′ E	850
21	Classen Lake	Classen Glacier, Easter Glacier, Sustins Glacier	Godley River	Southwest Pacific	43°30′ S, 170°28′ E	999
22	Maud Lake	Grey Glacier, Maud Glacier	Godley River	Southwest Pacific	43°28′ S, 170°30′ E	1035
23	Godley Lake	Godley Glacier, Ruth Glacier, Amherst Glacier	Godley River	Southwest Pacific	43°27′ S, 170°31′ E	1095
24	Lyell Lake	Lyell Glacier, Cockayne Glacier, Heim Plateau	Rakaia River	Southwest Pacific	43°18′ S, 170°53′ E	1015
25	Ramsay Lake	Ramsay Glacier	Rakaia River	Southwest Pacific	43°16′ S, 170°55′ E	935

Table 5: Basic geographic characteristics of all proglacial lakes in New Zealand. Description in the text

While there is no big Pleistocene or early Holocene glacial lake on the west of the Main Divide, all the proglacial lakes west of the Main Divide drain straight to the Tasman Sea. The eastern side of the Main Divide is characteristic by numerous late-Pleistocene – early-Holocene glacial lakes (see section 4.3), and therefore all the eastern proglacial lakes except of Lyell and Ramsay (Rakaia river catchment) drain to those big late-Pleistocene – early-Holocene glacial lakes. The character of the source glaciers also differs significantly between the areas. While the proglacial lakes in Mt Aspiring area are fed mainly by shorter and steeper glaciers, the eastern lakes of Mt Cook area are fed by long, gently sloped, valley glaciers. The location of proglacial lakes ranges from 43°16´ to 44°29´ South and from 168°24´ to 170° 55´ East.

The altitude of New Zealand proglacial lakes ranges from 517 metres a.s.l. for Waipara Lake (Mt Aspiring area) to 1538 m a.s.l. for Lake Axius (Mt Aspiring area). However, the average altitude of proglacial lakes in Mt Aspiring area is slightly higher (1004 m) than the average altitude of proglacial lakes in Mt Cook area (981 m). The relationship between the altitude and latitude of New Zealand proglacial lakes is described in figure 44. Note, that numbers 1 – 12 represent proglacial lakes of Mt Aspiring area, and numbers 13 – 25 proglacial lakes of Mt Cook area. The average altitude of all New Zealand proglacial lakes is 922 m a.s.l.

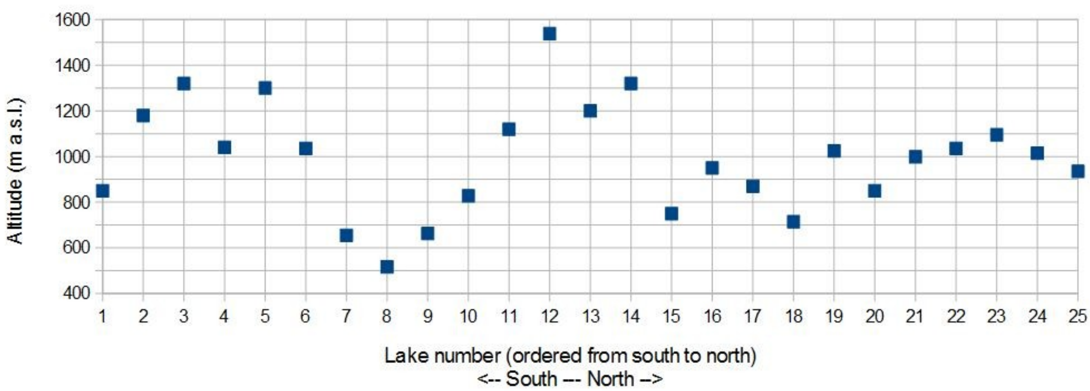

Figure 44: Relationship between latitude of New Zealand proglacial lakes and their altitude. The lakes are ordered from south to north as in table 5.

7.2 Geomorphic properties of New Zealand proglacial lakes

Selected geometric and geomorphic properties of all New Zealand proglacial lakes were studied (Tab. 6). The length of the lakes (see methods section for more detailed info) ranges from 150 metres in the case of Joe Lake (Mt Aspiring area) to 5.7 km in the case of Tasman Lake. While the average length was calculated to be 1.63 km, the lake length (and so the lake area) significantly varies between the areas. While the mean lake length in Mt Aspiring area was calculated to be 812.5 m, in Mt Cook area it is almost three times more (2.38 km). Figure 45 shows the relationship between latitudinal lake position and the lake length. Note, that numbers 1 – 12 represent proglacial lakes of Mt Aspiring area, and numbers 13 – 25 proglacial lakes of Mt Cook area.

Lake number	Lake name	Current lake length (km)	Current glacier contact	Dam type
1	Joe Lake	0.15	NO	Outwash Head
2	Victor Lake	0.45	NO	Bedrock
3	Snowball Lake	0.60	NO	Bedrock
4	John Inglis Lake	0.40	YES	Terminal Moraine
5	Snow White Lake	0.45	NO	Bedrock
6	Kitchener Lake	0.70	NO	Bedrock
7	Williamson Lake	1.15	NO	Bedrock
8	Waipara Lake	1.20	NO	Outwash Head
9	Volta Lake	1.50	YES	Outwash Head
10	Lucidus Lake	1.65	NO	Terminal Moraine
11	Lake Castalia	0.40	NO	Bedrock
12	Lake Axius	1.10	NO	Bedrock
13	Poet Lake	0.20	NO	Outwash Head
14	Spence Lake	0.60	YES	Terminal Moraine
15	Mueller Lake	1.80	YES	Moraines and Outwash head
16	Douglas Lake	2.60	YES	Terminal Moraine
17	Hooker Lake	2.30	YES	Moraines and Outwash head
18	Tasman Lake	5.70	YES	Outwash Head
19	Murchison Lake	2.80	YES	Outwash Head
20	La Perouse Lake	2.00	YES	Terminal Moraine
21	Classen Lake	3.00	YES	Moraines and Outwash head
22	Maud Lake	2.50	YES	Outwash Head
23	Godley Lake	4.10	YES	Moraines and Outwash head
24	Lyell Lake	1.80	YES	Outwash Head
25	Ramsay Lake	1.60	YES	Outwash Head

Table 6: Selected geometric and geomorphic properties of all 25 proglacial lakes of New Zealand. Lakes ordered from south to north. The lakes of Mt Aspiring area are shaded. Detailed description in the text

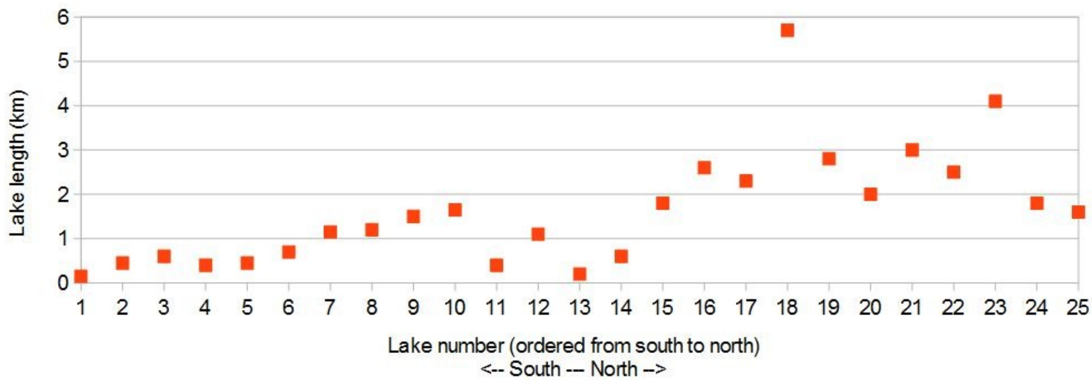

Figure 45: Relationship between latitude of New Zealand proglacial lakes and their length. The lakes are ordered from south to north as in table 6. Note the exceptional size of rapidly growing Tasman Lake (lake nr. 18).

"Current glacier contact" investigation revealed, that 11 out of total 25 proglacial lakes are currently not in a direct contact with its source glacier. However, from the nature of the definition of a proglacial lake (see section 3.4), all of them are directly influenced by a glacier. There is a strong connection between "current glacier contact" and the area where the lake is situated. While the Mt Aspiring area has just two lakes (about 17 % of the total Mt Aspiring proglacial lakes) directly connected with a glacier, the Mt Cook area consists almost ultimately of proglacial lakes with a direct glacier contact. In the Mt Cook area, 12 out of total 13 (about 92 %) proglacial lakes were classified as lakes with a direct glacier contact. The only exception is the small (200 m long) Poet Lake situated in the vicinity of a small (just about 500 m long) cirque glacier. The behaviour of small cirque glaciers significantly differs from the long, gently sloped, valley glaciers (see section 5.2) and therefore the Poet Lake (together with nearby Spence Lake), is highly unrepresentative of the whole Mt Cook area.

While the classification of proglacial lakes according the dam type is one of the most common (Carrivick and Tweed 2013), the type of a dam was used also in the inventory of New Zealand proglacial lakes. Proglacial lakes are most commonly dammed by ice, bedrock, moraine debris or landslide debris. Less commonly, proglacial lakes can be dammed by other sediments; for example as a glacier retreats and thins behind the ice-contact slope of a glacifluvial fan or apron (Carrivick and Tweed 2013). In New Zealand, those glacifluvial sediments are also commonly named by the term "outwash head" (Kirkbride 1993).

Within the whole Southern Alps of New Zealand no ice-dammed, or landslide-dammed proglacial lake were detected. Nine lakes were classified as being dammed by outwash head, 7 by bedrock, 5 by terminal moraine, and 4 by the combination of moraines (terminal or lateral) and outwash head (see table 6). However, it was concluded that the results can vary slightly due to the subjective method of dam type determination (see discussion section). Figure 46 clearly shows, that the majority of New Zealand proglacial lakes are being dammed by an outwash head, which corresponds with the findings of Kirkbride (1993) and Allen et al. (2009). However, Mt Aspiring and Mt Cook area differ significantly. While in the Mt Aspiring area proglacial lakes dammed by bedrock dominate, and the lakes dammed by the combination of a moraine and outwash head are entirely missing, in the Mt Cook area all the proglacial lakes are being emplaced behind glacial or glacifluvial sediments, especially outwash surfaces.

All New Zealand proglacial lakes inherited properties from their source glaciers. The retreat of small, high elevated, steep glaciers of Mt Aspiring area led to the formation of generally small, bedrock-dammed lakes, whereas large valley glaciers of Mt Cook area, located especially east of the Main Divide formed mostly large proglacial lakes dammed by their own outwash sediments as the glacier retreated.

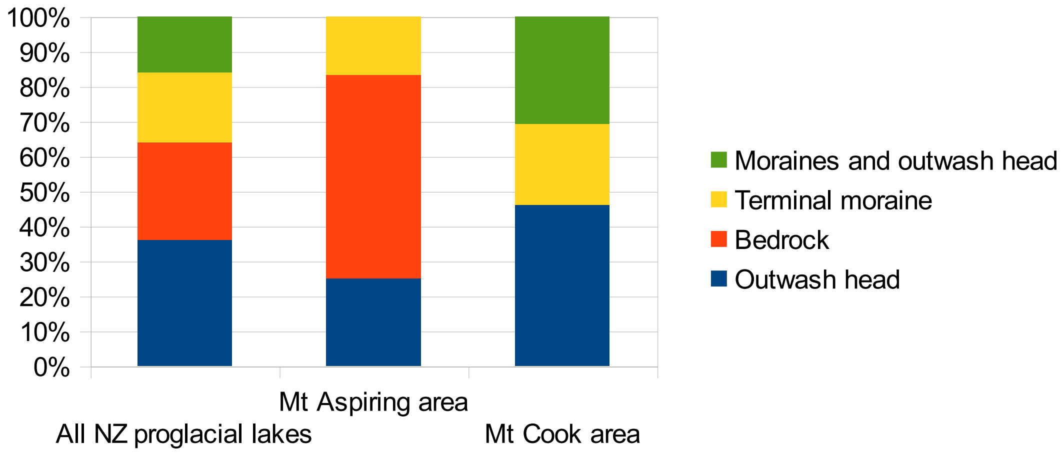

Figure 46: Classification of New Zealand proglacial lakes according to dam type. All New Zealand proglacial lakes (first column) are compared with the proglacial lakes of Mt Aspiring area (second column) and proglacial lakes of Mt Cook area (third column).

8 GLOFs hazard evaluation of selected glacial lakes

GLOFs hazard evaluation is a necessary step in the process of risk assessment and mitigation of all the threats to people, nature and property. While the protection of people and property is superimposed above the basic geomorphic interest, the flood modelling and other studies downstream of a lake are considered to be of high importance. On the other hand, some lakes appear to be more susceptible to dam burst or a displacement wave rise than others and thus should receive immediate attention. The hazard assessment is thus a necessary step in the process of risk mitigation. While various GLOF triggers can have various consequences for each lake, the triggers and the actual dam stability were studied separately in the section 8.2 and 8.3 respectively.

For example, the probability of an ice/rock avalanche impact into a lake and following displacement wave can be really high, but the effect of such a wave can be marginal and vice versa. The assessment of past events appears to be a crucial step in overall hazard assessment often enabling the recognition of the most hazardous areas and regularities. To reach this goal, past events are being studied in section 8.1.

8.1 Hazardous lakes according to past GLOFs events

The study of previous outburst floods in New Zealand (see section 6) revealed very little about the nature and local regularities of this phenomenon. This is mainly due to limited number of events recorded. If we consider just the floods from proglacial lakes, only two events (from a single lake) have been briefly described. Those were the two outburst floods from Maud Lake after the rock avalanches in 1992 (see section 6.4). There is a really limited information about the outburst flood at Kea Point more than 100 years ago caused by heavy rain (see section 6.2), so it is currently not possible to study any regularlities from this event. Relatively well described are the englacial outburst floods from Franz Josef Glacier, however, no other glacier in New Zealand show simillar processes and thus the results can not be applied generally.

From all those events we can preliminarily conclude, that the probability of GLOFs is higher within the Mt Cook National Park, east of the Main Divide, especially after an extreme rainfall, or following a rock avalanche. However, it would be incorrect to accept this statement as a dogma due to the insufficient number of events recorded. On the other hand it is also incorrect to assume that New Zealand glacial lakes do not pose any hazards related to GLOFs. It is important to realise that many of the large proglacial lakes of the Southern Alps have been formed just during the last ~40 years or even less (see section 5.4.2), so the number of mass movements, earthquakes and other potential triggers has been also limited.

The other factor to consider while assessing GLOFs hazards according to past events is the remoteness of the Southern Alps of New Zealand. There are almost no settlements in the vicinity of the glaciers and proglacial lakes and only small sparse villages and towns exist further downstream. This strongly contrasts with relatively more densely populated valleys of the Himalayan region, European Alps, North-American Cordillera and some other regions. Therefore, many outburst floods of New Zealand history have probably remained unnoticed.

While, historical GLOF events can be reconstructed from sedimentological records (e.g. Irvine-Fynn et al. 2015) and other geomorphic features (e.g. McSaveney 2002), many past events can be discovered in the future. However, today no GLOFs predictions can be done based only on the distribution of previous events and thus other factors need to be considered.

8.2 Hazardous lakes according to GLOF causes

Potentially dangerous lakes typically require a trigger mechanism to initiate a flood (Richardson and Reynolds 2000). Between one of the most common causes of a GLOF is a displacement wave from ice or rock avalanche that collapsed into the lake from hanging or calving glaciers (Richardson and Reynolds 2000) or from a rock face above the glacier (McSaveney 2002). Moraine collapse due to seepage or due to melting ice-cores, settlement and/or piping within the moraine as a result of earthquake shocks, sudden glacial or meteoric drainage into the lake, and inappropriate engineering works account for the remainder of recorded events (Richardson and Reynolds 2000). A qualitative method described in section 3.5 was used to assess the probability of various triggers to cause a GLOF. The results are shown in table 7.

8.2.1 Calving

The first trigger mechanism assessed was the process of breaking ice-blocks from the glacier front into the lake – the calving. While 14 out of 25 glaciers in New Zealand currently terminate in a proglacial lake (see section 7), the probability of a

Lake number	Lake name	Lake stability threatened by: Falling ice from calving front	Falling ice from hanging glacier	snow avalanches	rock/ice avalanches	landslides	Moraine collapse	The total of probability points
1	Joe Lake	0	0	2	1	1	0	**4**
2	Victor Lake	0	1	2	1	1	0	**5**
3	Snowball Lake	0	0	1	0	0	0	**1**
4	John Inglis Lake	1	0	2	2	1	1	**7**
5	Snow White Lake	0	0	1	0	0	0	**1**
6	Kitchener Lake	0	2	2	2	2	0	**8**
7	Williamson Lake	0	0	2	2	1	0	**5**
8	Waipara Lake	0	0	2	1	2	0	**5**
9	Volta Lake	1	2	2	2	2	0	**9**
10	Lucidus Lake	0	2	2	2	1	1	**8**
11	Lake Castalia	0	1	1	2	1	0	**5**
12	Lake Axius	0	0	1	0	0	0	**1**
13	Poet Lake	0	0	2	2	1	0	**5**
14	Spence Lake	1	0	2	2	1	1	**7**
15	Mueller Lake	1	2	2	2	2	0	**9**
16	Douglas Lake	1	1	2	2	1	1	**8**
17	Hooker Lake	2	0	2	2	1	0	**7**
18	Tasman Lake	2	0	1	1	1	0	**5**
19	Murchison Lake	1	0	2	1	2	0	**6**
20	La Perouse Lake	2	0	2	2	2	1	**9**
21	Classen Lake	1	1	2	2	1	0	**7**
22	Maud Lake	1	0	2	2	1	0	**6**
23	Godley Lake	2	0	2	2	1	0	**7**
24	Lyell Lake	1	2	2	2	2	0	**9**
25	Ramsay Lake	1	0	2	2	1	0	**6**

Table 7: The probability of various triggers to cause a GLOF for all 25 proglacial lakes of New Zealand.

GLOF triggered by calving was recorded for those 14 lakes. However, some glacier fronts are really thin without a frontal face and the ice-blocks too small to produce a serious GLOF, thus the probability of those lakes to produce a GLOF was classified as “Medium (1)”.

From various satellite imagery and personal visits it was concluded that calving occurs on many glacier fronts, however the displacement waves have been rather small to produce a GLOF. The isostatic release of an ice block from ice-foot or from the bottom of a lake as described by e.g. Warren and Kirkbride (1998) was concluded to be potentially hazardous for the large proglacial lakes of Mt Cook region. However, the prediction of such events is currently at its beginnings. The “High (2)” probability of a GLOF from calving was assigned to four proglacial lakes. Those are: Tasman Lake, Hooker Lake, La Perouse Lake and Godley Lake.

8.2.2 Hanging glaciers

The investigation of GLOF probability triggered by ice avalanches from hanging glaciers revealed that 9 out of total 25 proglacial lakes are surrounded by hanging glaciers. However, the probability of an ice-fall to a lake or its vicinity differs significantly between the lakes. The potentially most dangerous hanging glaciers were discovered in the vicinity of Kitchener Lake, Volta Lake, and Lucidus Lake in Mt Aspiring area and Mueller and Lyell lakes in Mt Cook area. In many cases the ice avalanches are occurring, but have not caused a GLOF yet. The potential for large magnitude ice avalanches combined with the predicted lake growth appear to be a significant threat and might result in the increase of GLOF probability triggered by ice avalanches from hanging glaciers.

8.2.3 Snow avalanches

All the proglacial lakes of New Zealand are located in mountainous landscape with relatively common avalanche activity, and the avalanches are one of the possible GLOF triggers (Emmer and Vilímek 2013), the effects of snow avalanches on lake stability were evaluated. According the method described in section 3.5, the potential of an avalanche to reach a lake was recorded for all 25 proglacial lakes of New Zealand, from which 20 were classified as having “High (2)” probability of GLOF occurrence from an avalanche impact.

Even no GLOFs caused by snow avalanches were recorded in New Zealand (see section 6), some lakes are directly surrounded by slopes of 30-45°; the slopes most susceptible to snow avalanches (Goddard 2008), and thus even large slab avalanches can be expected.

8.2.4 Rock/ice avalanches

While **(1)** ice/rock avalanches are among the most common causes of GLOFs (Emmer and Vilímek 2013), **(2)** far reaching disasters have resulted from transformations of ice/rock avalanches into debris or mudflows (Allen 2009), **(3)** ice/rock avalanches are a common phenomenon around the Main Divide of the Southern

Alps of New Zealand (Whitehouse 1983), **(4)** an ice/rock avalanche caused the only two GLOFs recorded in New Zealand during the last 100 years (McSaveney 2002), the hazards that ice/rock avalanches pose to proglacial lakes need to be properly evaluated.

The assessment of all 25 proglacial lakes revealed that 17 out of 25 proglacial lakes have “High (2)” probability of a GLOF from ice/rock avalanches, 5 lakes were classified as “Medium (1)”, and 3 lakes as “Low (0)”. All the lakes classified as having “Low (0)” probability are located in high altitudes of Mt Aspiring area and dammed by bedrock. With the altitudes >1300 m, those lakes are the highest of all the proglacial lakes of Mt Aspiring area and between the four highest proglacial lakes in New Zealand. Even the lakes classified as “High (2)” are located in both areas, in Mt Cook Area 11 out of 13 proglacial lakes fit into this category, whereas in Mt Aspiring area “just” 6 out of 12 proglacial lakes were classified as having high probability of an ice/rock avalanche to cause a GLOF.

8.2.5 Landslides

An investigation of landslide hazards according the methods described in section 3.5 revealed that three lakes have “Low (0)”, 15 lakes “Medium (1)”, and seven lakes “High (2)” probability of a GLOF from landsliding. While large magnitude landslides require more detailed assessment (Korup 2005), landsliding from freshly uncovered moraine walls can be easily recognized from satellite images. Even the unconsolidated moraine material includes also large boulders (>2 m in diameter) landsliding from moraine walls has not caused a serious GLOF that would be recorded in New Zealand. Due to the slope instability, steep slope angle, and vertical distance more than 100 metres (Fig. 47), the effects of landsliding from moraine walls can not be neglected.

Figure 47: Inner side of Tasman Glacier lateral moraine. Due to rapid downwasting of many large New Zealand glaciers, steep unstable slopes were uncovered and landsliding was initiated. Currently the level of Tasman lake is about 100 m bellow the moraine crest. Source: Archive of author

However, the hazards related to large

magnitude landslides are several orders greater. Such landslides can cause huge outburst floods often leading to disasters (RCEM 2014). To assess the hazard correctly, more detailed evaluation is needed. Therefore all the lakes classified as having “High (2)” probability of a GLOF from landslides should be studied into a more detail.

8.2.6 Moraine-dam collapse

The last GLOF trigger assessed – the moraine dam collapse – is probably the most complex one, because it encompasses many various processes (see section 3.5). It is clear that a moraine dam can collapse due to a displacement wave caused by mass-movements, but in this thesis mass-movements are studied separately (see above), and therefore this section includes only the processes mentioned in section 3.5. While just five proglacial lakes in New Zealand are dammed by a terminal moraine, the assessment of moraine dam collapse probability can be done just for those lakes. Those are: John Inglis Lake and Lucidus Lake in Mt Aspiring area, and Spence Lake, Douglas Lake, and La Perouse Lake in Mt Cook area. All the five moraine-dammed lakes were classified as having “Medium (1)” probability of dam collapse, however, more detailed assessment including field study was suggested. Except of Lake Lucidus all the moraine-dammed lakes are located west of the Main Divide, where the valleys are generally steeper, without large braided-river reaches.

8.2.7 The most hazardous lakes according the causes

The summation of all the points for each proglacial lake revealed a broad variation between the number and/or probability of various triggers to cause a GLOF. The total count of the points ranges from 1 to 9 points (see table 7 and figure 48). The total number of lakes (y axis) that gained a certain number of points in the hazard level assessment according the GLOF triggers is shown in figure 48. While the mean total points count for Mt Aspiring area lakes is c. 4.9, for Mt Cook area it is 7, indicating that the proglacial lakes of Mt Cook area are threatened by either more processes, and/or the probability of those processes to cause a GLOF is higher.

While no lake gained 0 points, the minimum point count is 1 (Snowball Lake, Snow White Lake, and Lake Axius). All those lakes are bedrock-dammed, high altitude (1300-1538 m a.s.l.) proglacial lakes of Mt Aspiring area. Due to their location on a slope rather than in a valley, no rock or ice faces threaten their stability and the only hazard that remains is the avalanche activity.

The total of four points were assigned to Joe Lake; a small lake of Mt Aspiring area embedded in glacifluvial sediment and surrounded by steep slopes. Even the snow avalanches were recognized as the biggest hazard for the lake stability, rock/ice avalanches can occur on the north-facing slopes south of the lake and landslides on the south-facing slopes north of the lake.

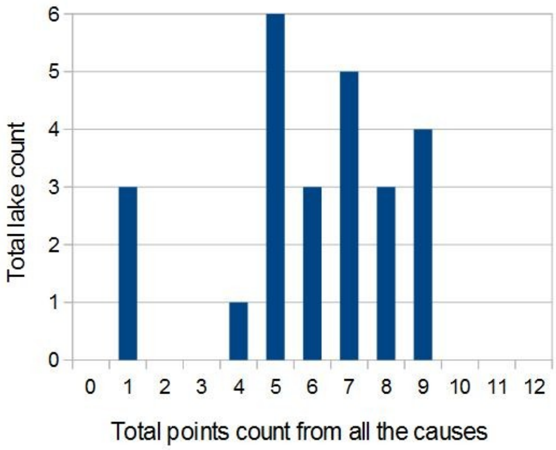

Figure 48: Lake frequencies according the trigger hazard assessment. The total number of lakes (y axis) that gained a certain number of points in the hazard level assessment according the GLOF triggers (x axis).

The total of five points were recorded for six different lakes. Four (Victor Lake, Williamson Lake, Waipara Lake, and Lake Castalia) from Mt Aspiring area and two (Poet Lake, and Tasman Lake) from Mt Cook area. This group includes both the bedrock-dammed lakes and the lakes dammed by outwash head in the altitudes between 517 – 1200 m a.s.l. The lake length also varies significantly (0.4 – 5.7 km) and so does the range of hazards that can potentially cause a GLOF. While snow avalanches, ice/rock avalanches and landslides appear to be the dominant hazards for most of the lakes of this group, calving has been recognized as the dominant hazard for Tasman Lake.

Six points were recorded just in the Mt Cook area, especially in its north-eastern part. Three lakes (Murchison Lake, Maud Lake, and Ramsay Lake) fit into this group. All those lakes are dammed by outwash head, and have really similar geographic and geometric properties. The altitude ranges between 935 – 1035 m a.s.l., and the lake length is currently between 1.6 and 2.8 km. Maud Lake and Ramsay Lake have even the same probability points for all the triggers with the highest probabilities recorded for snow- and ice/rock- avalanches. Whereas Murchison Lake (apart from snow avalanches) appears to have the highest probability of triggering a GLOF from landslide processes.

Seven points were recorded for five New Zealand proglacial lakes. One (John Inglis Lake) within Mt Aspiring area an four (Spence Lake, Hooker Lake, Classen Lake, and Godley Lake) within Mt Cook area. All of those lakes have well developed terminal moraines, but just two lakes are by the moraines really impounded. All the other lakes are impounded by outwash head. Altitudes range between 870 – 1320 m a.s.l., and lake lengths between 0.4 – 4.1 km. While all the lakes have “High (2)” probability of a GLOF from snow avalanches and ice/rock avalanches, Hooker Lake and Godley Lake (the two biggest in this group) have additionally “High (2)” probability of a GLOF also from the process of calving.

Eight points were assigned to three proglacial lakes of New Zealand. Kitchener Lake, and Lucidus Lake in Mt Aspiring area, and Douglas Lake in Mt Cook area. While Kitchener Lake and Lucidus Lake are both cirque lakes (dammed by bedrock and a terminal moraine respectively), Douglas Lake is a moraine-dammed lake located in a narrow valley. Altitudes range between 828 – 1036 m a.s.l., and lake lengths between 0.7 – 2.6 km. All the lakes in this category have “High (2)” probability of a GLOF from snow avalanches and ice/rock avalanches, and the lakes of Mt Aspiring area also from hanging glaciers. Kitchener Lake have two points also in the hazards from landslides. Douglas Lake have “Medium (1)” probability for all the causes studied, except of snow- and ice/rock- avalanches were there is the “High (2)” value.

The highest number of points (9) were counted for four proglacial lakes in New Zealand. Volta Lake in Mt Aspiring Area, Mueller Lake, La Perouse Lake and Lyell Lake in Mt Cook area. Outwash head is the dominant type of the dam in this group, but La Perouse Lake is dammed by a terminal moraine. The lake length is relatively consistent and ranges between 1.5 – 2.0 km. The altitude ranges between 663 – 1015 m a.s.l. All the lakes in this category have “High (2)” probability of a GLOF from snow avalanches, ice/rock avalanches, landslides, and hanging glaciers, except of La Perouse Lake which does not have hanging glaciers in its vicinity, but faces the hazards from calving instead.

8.3 Hazardous lakes according to lake/dam properties

Lake/dam properties influencing the magnitude of a GLOF were assessed according the method described in section 3.5. The results (Tab 8) show that lake size and dam freeboard are the main characteristics influencing the severity of a potential GLOF.

8.3.1 Lake size

The lake length investigation (used as an approximation to lake volume) revealed, that six proglacial lakes are small lakes shorter than 0.50 km (probability value “Low (0)”). Except of Poet Lake, all of them are located within Mt Aspiring area. The “Middle (1)” probability was assigned to another six lakes (0.51 – 1.50 km long), from which five are located in Mt Aspiring area and one (Spence Lake) in Mt Cook area. “High (2)” probability to cause a severe GLOF based on lake size was assigned to 13 proglacial lakes of New Zealand, from which just two are located within Mt Aspiring area. Those are Lake Volta and Lake Lucidus. In Mt Cook area all the lakes except of Poet Lake and Spence Lake were classified as having “High (2)” probability to cause a severe GLOF.

8.3.2 Dam freeboard

Dam freeboard was assessed for all the proglacial lakes regardless the dam type. While all New Zealand proglacial lakes have a surface outflow, freeboard (based on its geomorphic definition – see section 3.5) would be 0 and the lake probability to overflow “High (2)”. From those reasons the characteristic was assessed qualitatively rather than quantitatively and dam freeboard was approached as a complex of topographic characteristics. The “dam freeboard evaluation step” was thus approached as the probability of a displacement wave to overtop the dam, rather than a single value measurement indicating the dam freeboard. It was concluded that this approach enables the correct evaluation of dam overtopping.

The “Low (0)” probability was assigned just to two lakes: Williamson Lake and Lake Axius: bedrock-dammed lakes in Mt Aspiring area. 10 lakes were declared to have “Medium (1)” probability to cause a severe GLOF due to dam overtopping. Those are lakes of both areas, dammed by a moraine, bedrock, or by a combination of moraines and outwash head (see section 7). “High (2)” probability was assigned to 13 proglacial lakes dammed by outwash head. While all the lakes dammed by outwash head do not

Lake number	Lake name	Contribution to cause a severe GLOF: Lake length	Dam freeboard	Moraine width-to-height ratio	Evident seepage	The total of probability points
1	Joe Lake	0	2	0	0	**2**
2	Victor Lake	0	1	0	0	**1**
3	Snowball Lake	1	1	0	0	**2**
4	John Inglis Lake	0	1	1	0	**2**
5	Snow White Lake	0	2	0	0	**2**
6	Kitchener Lake	1	2	0	0	**3**
7	Williamson Lake	1	0	0	0	**1**
8	Waipara Lake	1	2	0	0	**3**
9	Volta Lake	2	2	0	0	**4**
10	Lucidus Lake	2	1	1	0	**4**
11	Lake Castalia	0	2	0	0	**2**
12	Lake Axius	1	0	0	0	**1**
13	Poet Lake	0	2	0	0	**2**
14	Spence Lake	1	1	1	0	**3**
15	Mueller Lake	2	1	0	0	**3**
16	Douglas Lake	2	1	1	0	**4**
17	Hooker Lake	2	1	0	0	**3**
18	Tasman Lake	2	2	0	0	**4**
19	Murchison Lake	2	2	0	0	**4**
20	La Perouse Lake	2	1	1	0	**4**
21	Classen Lake	2	1	0	0	**3**
22	Maud Lake	2	2	0	0	**4**
23	Godley Lake	2	2	0	0	**4**
24	Lyell Lake	2	2	0	0	**4**
25	Ramsay Lake	2	2	0	0	**4**

Table 8: The probability of various lake/dam properties to cause a severe GLOF (increase the flood magnitude) for all 25 proglacial lakes of New Zealand.

have any distinct dam (moraine or bedrock), the displacement wave do not need to overcome any barrier and the wave propagation is therefore more probable.

8.3.3 Moraine width-to-height ratio and evident seepage

The last two categories describe characteristics of moraine-dammed lakes. While there are just five moraine-dammed proglacial lakes in New Zealand (see section 7), the results limit to those lakes only. The width-to-height ratio assessment (done qualitatively) revealed, there are no high moraines damming the proglacial lakes, leading to high width-to-height ratio, and thus only "Medium (1)" probability of a severe GLOF from dam break was assigned. The study of various satellite images in Google Earth software has not revealed any evident seepage through the moraine dams, indicating a higher dam stability. All the proglacial lakes have a surface outflow.

8.3.4 The most hazardous lakes according the lake/dam stability

The summation of all the "lake/dam stability" points for each proglacial lake revealed relatively consistent results. The total count of the points ranges from 1 to 4 points (see table 8 and figure 49). The total number of lakes (y axis) that gained a certain number of points in the hazard level assessment according the lake/dam stability is shown on figure 49. While the mean total points count for Mt Aspiring area lakes is 2.25, for Mt Cook area it is c. 3.5, indicating that the proglacial lakes of Mt Cook area are generally more susceptible to dam overtopping or dam collapse in the case of displacement wave propagation or moraine collapse itself.

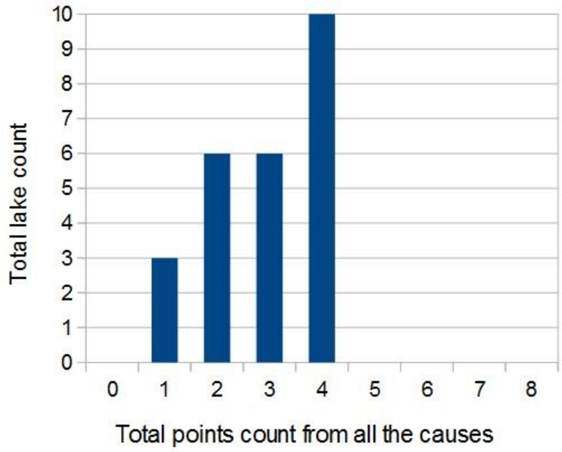

Figure 49: Lake frequencies according the lake/dam stability assessment. The total number of lakes for each category is shown on the y axis, the categories according the result points from lake/dam stability assessment are shown on the x axis.

While no lake gained 0 points, the minimum point count is 1 (Victor Lake, Williamson Lake, and Lake Axius). All those lakes are bedrock-dammed proglacial lakes of Mt Aspiring area. Victor lake is relatively small (0.45 km long) lake with a stable bedrock dam, therefore the only GLOF mechanism probable is the dam overtopping following a mass movement into the lake. Due to the small lake size, the potential GLOF is expected to be small as well. Williamson lake and Lake Axius are bigger in size (the length of 1.15 and 1.10 km respectively), however the bedrock dam surrounding the outlet is relatively high, decreasing the probability of overtopping.

Two points were assigned to six proglacial lakes of New Zealand (see table 8 for the list). Five in Mt Aspiring area, and one (Poet Lake) in Mt Cook area. Five lakes were classified as small, and one (Snowball Lake) as medium. Four lakes of those six lakes have the freeboard close to zero, and therefore they are more prone to dam overtopping (2 points assigned).

Three points were assigned also to six proglacial lakes. Two in Mt Aspiring area (Kitchener Lake and Waipara Lake) and four in Mt Cook area (Spence Lake, Mueller Lake, Hooker Lake and Classen Lake). Three of them classified as medium-size lakes and the other three as large. While those two lakes of Mt Aspiring area have lower freeboard (classified as having “High (2)” probability of overtopping), the remaining four lakes of Mt Cook area were classified as having “Medium (1)” probability of overtopping based on freeboard assessment.

Four points were assigned to ten lakes, from which two are located within Mt Aspiring area and eight in Mt Cook area (see table 8 for the list). All the lakes in this category were classified as large (the length of 1.50 km and more). While three lakes of this category are moraine-dammed lakes, the dam freeboard is higher, compared to the lakes dammed by outwash head. Therefore those moraine-dammed lakes were classified as having “Medium (1)” probability of dam overtopping.

8.4 The most hazardous proglacial lakes of New Zealand

To create an overall GLOFs hazard assessment, the result from trigger assessment and lake/dam stability assessment were put together. Table 9 shows the most hazardous proglacial lakes of New Zealand based on both assessments.

Lake number	Lake name	The total of trigger probability points	The total of lake/dam stability points	The total GLOFs hazard points
1	Joe Lake	4	2	6
2	Victor Lake	5	1	6
3	Snowball Lake	1	2	3
4	John Inglis Lake	7	2	9
5	Snow White Lake	1	2	3
6	Kitchener Lake	8	3	11
7	Williamson Lake	5	1	6
8	Waipara Lake	5	3	8
9	Volta Lake	9	4	13
10	Lucidus Lake	8	4	12
11	Lake Castalia	5	2	7
12	Lake Axius	1	1	2
13	Poet Lake	5	2	7
14	Spence Lake	7	3	10
15	Mueller Lake	9	3	12
16	Douglas Lake	8	4	12
17	Hooker Lake	7	3	10
18	Tasman Lake	5	4	9
19	Murchison Lake	6	4	10
20	La Perouse Lake	9	4	13
21	Classen Lake	7	3	10
22	Maud Lake	6	4	10
23	Godley Lake	7	4	11
24	Lyell Lake	9	4	13
25	Ramsay Lake	6	4	10

Table 9: The overall GLOF hazard assessment for all 25 proglacial lakes of New Zealand ordered from south to north as in section 7. The trigger probability hazard points range between 1-9, lake/dam stability points between 1-4, and the total GLOF hazard points between 2-13.

The results show that the “Total hazard points” range from 2 points for the least hazardous lakes and up to 13 points for the most hazardous ones. The level of the hazard varies geographically, generally implying lower hazard for the proglacial lakes of Mt Aspiring area (the average of 7.17 points), and higher hazard for the proglacial lakes of Mt Cook area (the average of 10.54 points). This geographic variability is expressed in figure 50.

The least hazardous lake is a bedrock-dammed, high-altitude proglacial lake of Mt Aspiring area: Lake Axius (Lake number 12). Lake Axius reached only two points, one from each assessment (more detailed description in section 8.2 and 8.3). Lake Axius is followed by other two bedrock-dammed lakes of Mt Aspiring area: Snowball Lake (Nr. 3) and Snow White Lake (Nr. 5). Both of them reached three points due to lower dam freeboard. While no lake reached four or five points, the next group of lakes was classified as having six points. Those are: Joe Lake (Nr. 1), Victor Lake (Nr. 2), and Williamson Lake (Nr. 7) of Mt Aspiring area. Seven points were calculated for two small lakes. Lake Castalia in Mt Aspiring area (Nr. 11) and Poet Lake in Mt Cook area (Nr. 13). While there is no potential for the growth of those lakes, it is believed that sedimentation will further lower their GLOF hazard.

Waipara Lake of Mt Aspiring area (Nr. 8) reached eight points, John Inglis Lake

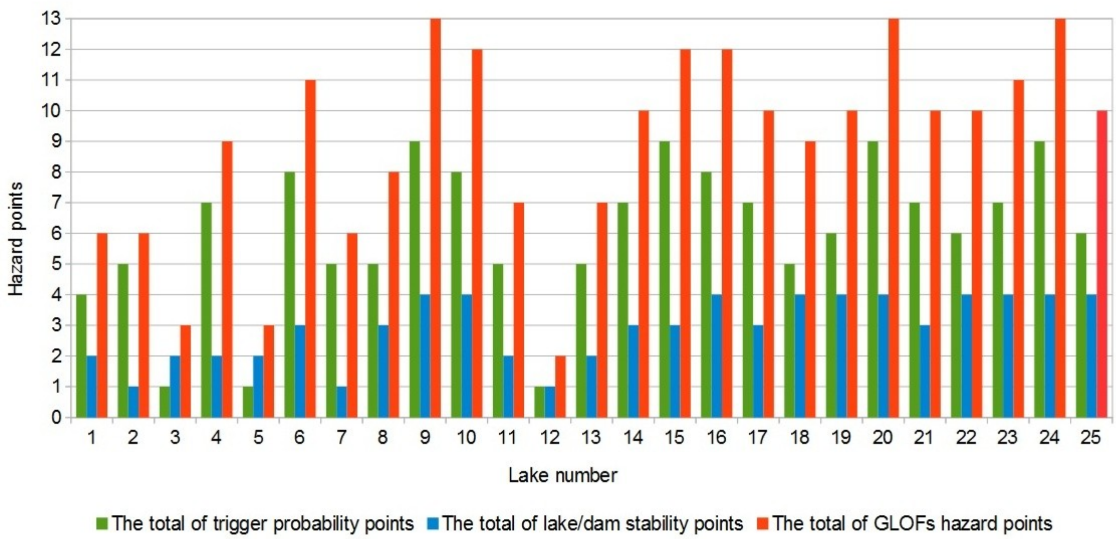

Figure 50: The total hazard point count for all 25 proglacial lakes of New Zealand. The results of trigger assessment are displayed in green, the results of lake/dam stability assessment in blue, and the total of GLOF hazard points (summation of the two previous) in red.

(Nr. 4) of Mt Aspiring area and Tasman Lake (Nr. 18) of Mt Cook area nine points. While John Inglis Lake is one of the smallest lakes studied, Tasman Lake is the biggest one with a a dramatic growth during the last few decades (see section 5.4.2). While the lake size and a rapid lake growth may indicate increased GLOFs hazard, Tasman Lake reached "only" nine points. This is mainly due to the topographic setting resulting in "Low (0)", or "Medium (1)" probability of mass-movements to reach the lake. However, the low dam freeboard indicates "High (2)" probability of dam overtopping in the case of displacement wave propagation. Dam collapse is highly improbable due to the fact that the lake is dammed by thick layer of the glacifluvial sediments of Tasman Glacier.

More than a half of the total number of lakes studied (14 out of 25) reached ten or more points, which was classified as a "High (2)" GLOF probability. From those 14 lakes, just three are within Mt Aspiring area, and the rest is located within Mt Cook area. Ten points were reached by six lakes (see table 9), from which all are proglacial lakes of Mt Cook area. 11 points were reached by relatively small Kitchener Lake (Nr. 6) located in the centre of Mt Aspiring area and by relatively large Godley lake (Nr. 23) of Mt Cook area. While the size of Kitchener Lake is relatively small, the lake is dammed by bedrock and the probability of most of the mass-movements to reach the lake is "High (2)", the probability of a small displacement wave overtopping the dam was thus concluded to be high.

Three lakes reached the sum of 12 points. Lucidus Lake (Nr. 10) of Mt Aspiring area, and Mueller Lake (Nr. 15) and Douglas Lake (Nr. 16) of Mt Cook area. Except of generally high probability of mass-movements to reach the lake, Lake Lucidus is also threatened by the collapse of the moraine dam. Mueller and Douglas Lakes are further threatened by the process of calving. All of those lakes are of reasonable size (1.65 – 2.80 km in length).

The most hazardous lakes according to the method used are Volta Lake (Nr. 9) of Mt Aspiring area, and La Perouse Lake (Nr. 20) and Lyell Lake (Nr. 24) of Mt Cook area. All of those lakes gained 13 points. All of those lakes are threatened by snow avalanches, ice/rock avalanches and landslides from steep slopes surrounding the lakes, mostly in the longitudinal axis of the lake. While La Perouse Lake has no hanging glaciers above, Volta Lake and Lyell Lake have high probability of falling ice from hanging glaciers. Calving occurs in all three cases, however La Perouse Lake has higher probability of a serious displacement wave due to calving. Also the probability of moraine collapse was recorded just for La Perouse lake. On the other hand, La Perouse Lake has higher dam freeboard resulting in the equal points count as Volta Lake and Lyell Lake. All those three hazardous lakes reached 9 points in the trigger assessment, and 4 points in the lake/dam stability assessment, which is – in both cases – the maximum reached. The lake location and its geomorphic features are shown in figure 51 (Volta Lake), figure 52 (La Perouse Lake) and figure 53 (Lyell Lake). All topographic maps are oriented to the north, and all the satellite images show the view from downstream.

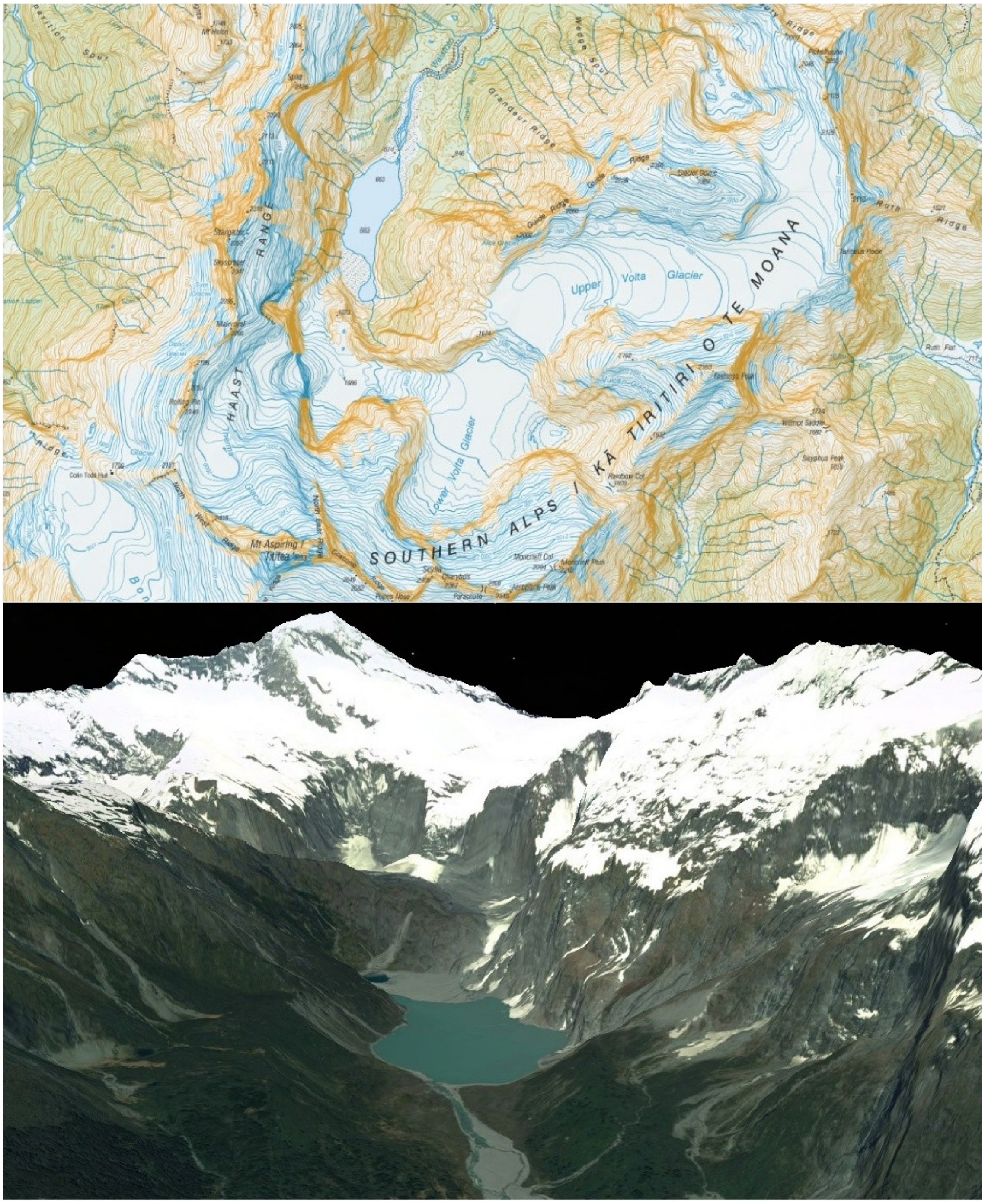

Figure 51: Location and geomorphic features threatening the stability of Volta Lake in terms of GLOFs. Source of the map: DoC, Source of the satellite image: Google Earth

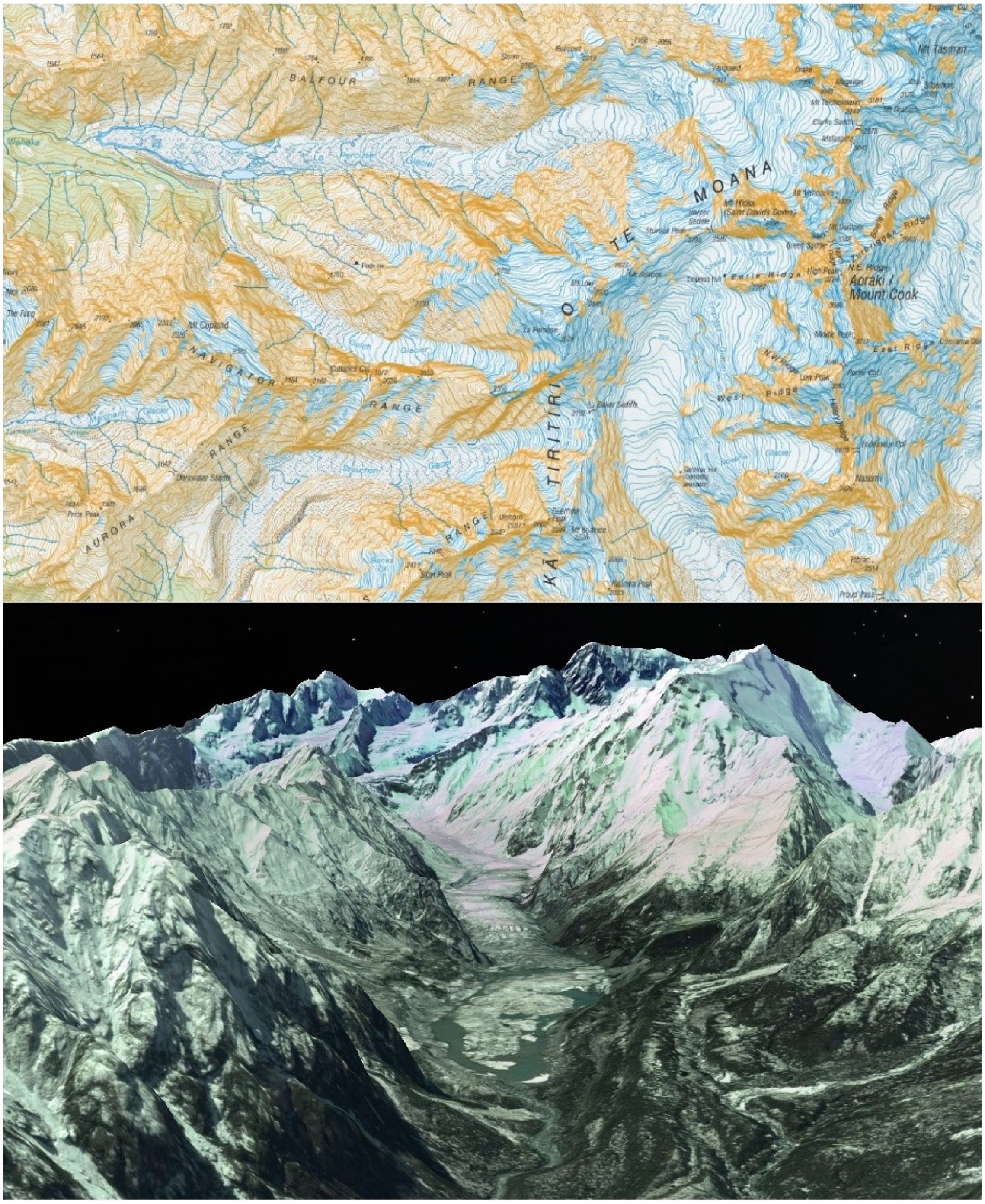

Figure 52: Location and geomorphic features threatening the stability of La Perouse Lake in terms of GLOFs. Source of the map: DoC, Source of the satellite image: Google Earth

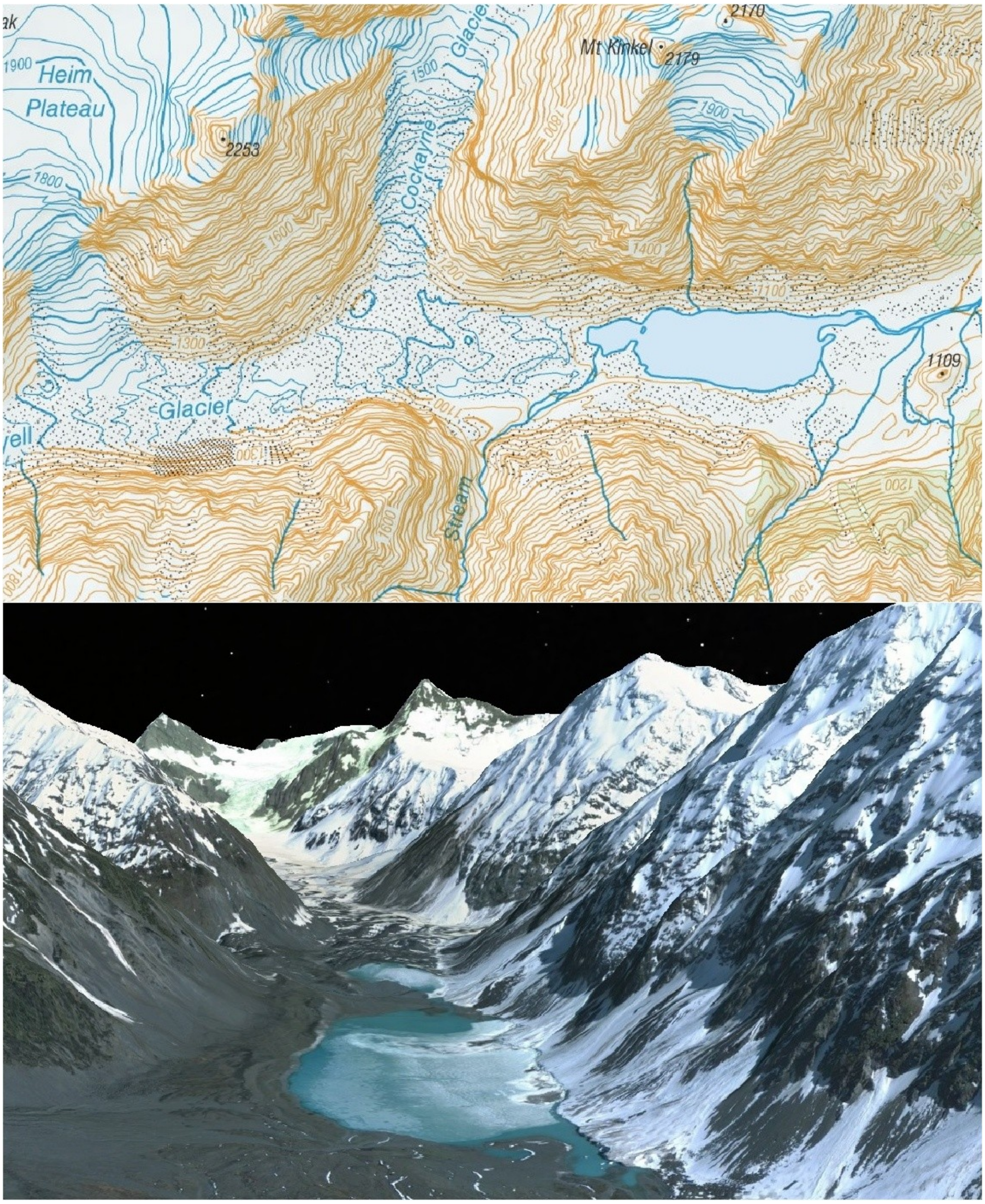

Figure 53: Location and geomorphic features threatening the stability of Lyell Lake in terms of GLOFs. Source of the map: DoC, Source of the satellite image: Google Earth

9 The potential floodwave propagation and related risks

The key goal of a GLOF hazard assessment is to obtain GLOFs probability and magnitude results. Those should further serve as input data for floodwave propagation modelling and risk assessment. Outburst floods pose a serious threat to life, property and infrastructure and can exert intense and widespread landscape change through erosion of both unconsolidated sediments and bedrock (Carrivick 2010). However, GLOFs are extremely complex phenomena varying significantly event from event. The nature of a GLOF is characterised by many factors like the triggering mechanism(s), reservoir hypsometry, the geometry, type, composition and structural integrity of the dam, as well as the topography and geology of the flood path (Westoby et al. 2014).

The social vulnerability then influences the severity of an event from the point of costs on property and losses of life (Adger 2006). Social vulnerability varies significantly by region and local conditions. Generally speaking, a high magnitude, violent GLOF, for instance, passing through uninhabited land presents little danger, while on the other hand, a relatively small GLOF can pose significant risks to human life and can result in great economic losses in densely populated areas (Donner and Havidán 2011).

Glacial lake outburst floods have had serious impact to human activities, life and property (e.g. Richardson and Reynolds 2000). “Some 32,000 people have been killed by glacial lake outbursts in Peru during the 20th century; hundreds of people and livestock have died in the Himalayas in the last 50 years by being swept away in the catastrophic discharges from lakes high in the mountains. Commercial projects in Asian countries are inadvertently extending into areas prone to glacial hazards in response to increased land use pressures and natural resource exploitation” (Richardson and Reynolds 2000).

Continuous economic and tourism growth in New Zealand (MBIE 2017) also creates a pressure to expand to more hazardous areas of the Southern Alps. In New Zealand, the expansion is most notable on the example of tourist infrastructure development. As human activities extend further into the high mountainous regions, conflicts with glacial hazards are becoming more apparent (Richardson and Reynolds 2000). However, overall hazard and risk assessments are being made around many inhabited places of New Zealand (e.g. DoC 2009) and the future developments are being effectively regulated.

9.1 Flood channel assessment

Flood channel assessment revealed some significant differences in downstream valley properties between proglacial lakes of New Zealand. The first characteristic assessed was the longitudinal river profile expressing the overall slope. While all the rivers have large braided rivers in the lower reaches, some glacial rivers flowing from generally smaller proglacial lakes have steep river reaches close to the lake outlets. This investigation is important mainly for the determination whether a debris flow can occur. Allen et al. (2009) suggest that the critical slope for a debris flow is 10°, therefore all the river reaches steeper than ~10° were located. However, for a debris flow to occur the valley needs to be covered by a loose material (Allen et al. 2009), therefore steep rock walls bellow some proglacial lakes do not pose a hazard of debris flow. A debris flow from a proglacial lake can potentially occur bellow: Joe Lake, Victor Lake, John Inglis Lake, and Kitchener Lake. All of those lakes are relatively small, remote lakes of Mt Aspiring area.

Together with the slope, the valley floor type varies as well. All the rivers draining Mt Aspiring area proglacial lakes flow in relatively narrow valleys, without braided river reaches for several kilometres. The shortest section of a narrow valley (before a braided river reach starts) was measured for Lake Williamson, where the braided river starts ~15 km downstream from the lake. The width measurement of braided river reaches showed that the Waiatoto river has the narrowest valley floor of all braided rivers within Mt Aspiring area. Just few braided river reaches ranging from 100 to 600 metres are present on the Waiatoto river, thus inhibiting the flood wave dissipation and water infiltration. The Waiatoto river drains Volta Lake and Lake Axius and terminates in the Tasman see on the west coast. Before reaching the sea, the flow velocity decreases and the river starts to meander.

In Mt Cook area, the valley floors west of the Main Divide differ significantly from those east of the Main Divide. All the rivers draining west have reaches of narrow valley and reaches of braided rivers further downstream. However, the braided river reaches are relatively short and narrow (< 1 km) compared to the braided rivers of eastern lakes. Therefore a potential flood from the western lakes of Mt Cook area (Poet, Spence, Douglas, and La Perouse Lakes) has less space to spread resulting in higher peak flood level. Three rivers drain those four western lakes. Haast river drains Poet and Spence Lakes, Karangarua river drains Douglas Lake and Cook river drains La Perouse lake. While the Haast river has several wider flats in its course, the physical vulnerability of the Karangarua and Cook rivers were concluded to be higher.

East of the Main Divide, three main braided rivers drain nine proglacial lakes. Those are the Tasman river (draining Mueller, Hooker, Tasman, and Murchison lakes), the Godley river (draining Classen, Maud, and Godley lakes), and the Rakaia river (draining Ramsay and Lyell Lakes). While the overall width of those braided rivers ranges from one to two kilometres (see figure 54), in some reaches the maximum width was measured to be up to four kilometres (e.g. the Rakaia River or the Tasman river). The retention potential of such a big outwash plains is relatively high, however this question is quite complex and requires its own assessment procedure with high demands on input data (Emmer and Vilímek 2014).

McSaveney (2002) described the outburst flood from Maud Lake in 1992 and the effects on landscape (see section 6.4). The braided gravel bed of Godley River was extensively scoured for several kilometres, and the flood travelled 45 km to Lake Tekapo, where it raised the lake level for ~90 mm. Due to the simillar lake properties within Mt Cook area (see section 7) and simillar hazards of GLOFs (see section 8), simillar effects can be expected to occur in case of a future events. While Lake Pukaki is used for hydropower generation, same as Lake Tekapo, the water level oscillates significantly (McSaveney 2002) and the lake level rise of few centimetres appears to be marginal from the safety point of view.

The only exception is the Rakaia river – the biggest braided river in New Zealand. Compared to the Tasman and Godley rivers, the Rakaia river does not have a large Pleistocene/Holocene glacial lake in its way enabling the flood wave to travel tens of kilometres before reaching the sea. If the potential flood magnitude is extremely high, the flood wave could reach Rakaia Gorge ~80 km downstream (see figure 55 a). The gorge could cause flood water rise, and significant erosion due to undercutting of unstable sedimentary deposits on its southern end of the gorge (fig 55 b).

Figure 54: Outwash plain of Tasman River as seen from the Holocene terminal moraine of Tasman glacier – the longest glacier in New Zealand. With the width up to 3 km the water retention within the flood plain is relatively high. Source: Archive of the author.

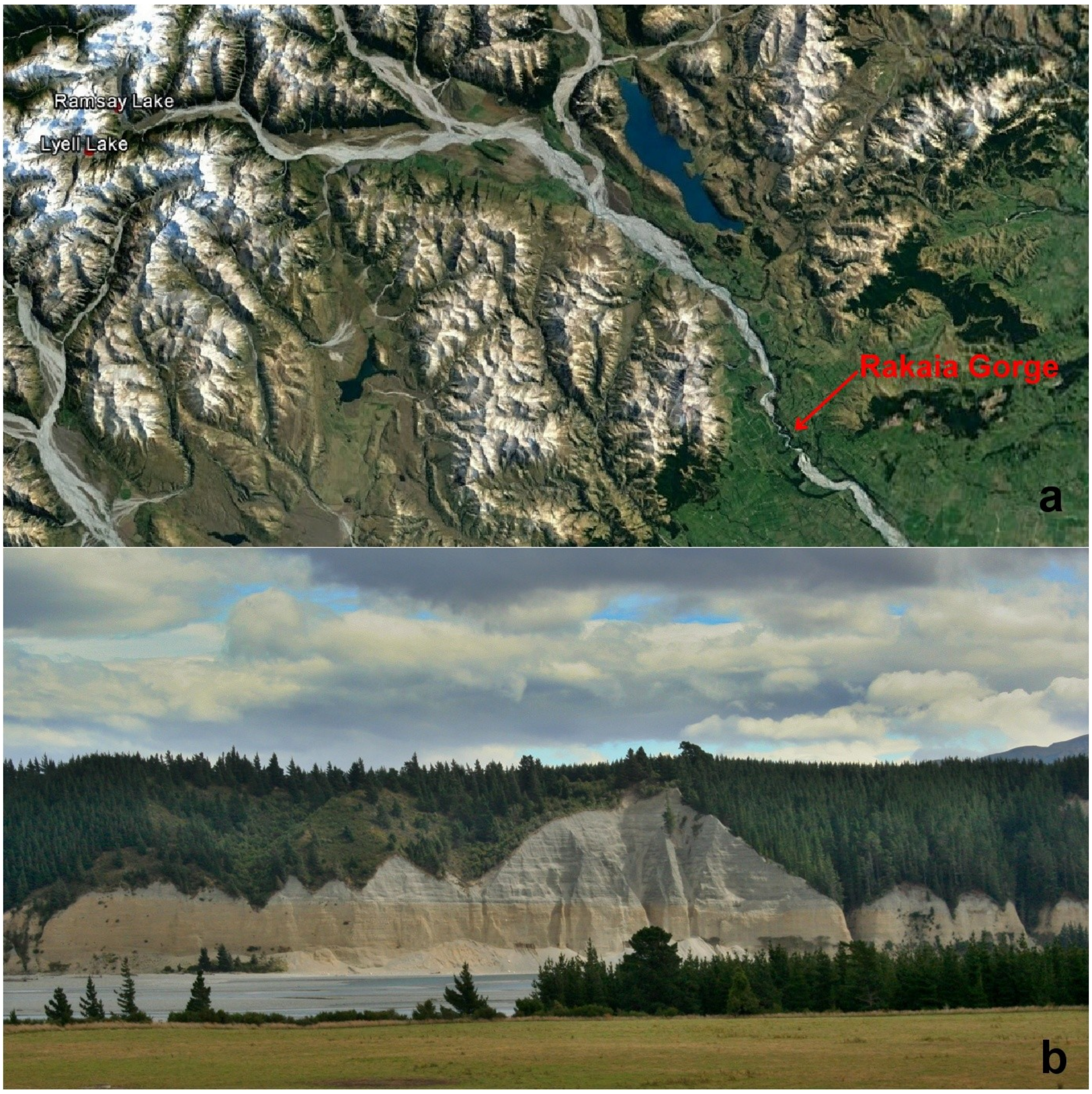

Figure 55: (a) Location of Rakaia Gorge. This narrow river reach is surrounded by soft sedimentary formations (b). Source of satellite image: Google Earth, Source of the photo: Archive of the author.

9.2 GLOFs Risk for society in New Zealand

While the risk level is influenced by various socio-economic factors like density of population and the level of hazard awareness (Adger 2006), all those factors should be studied individually. However, most of the socio-economic indicators are reflected in the nature. The intersection of infrastructure and settlement areas with potentially hazardous areas may for instance indicate poor policies, poor decision making, high population density, corruption, high tourism increase, or combination of many of those factors.

With the total population of about 4.7 million (2016) and area about 271,000 km^2 (STATS-NZ 2017), the overall New Zealand's population density is about 17.3 people/km^2. However, most of the population is concentrated in few large cities, and vast areas of New Zealand remain almost uninhabited (Fig 56). Furthermore, almost all mountainous areas of the South Island has less than one person per square km (Fig 56). Low population density together with high socio-economic development of New Zealand contributes to generally low social vulnerability to GLOFs hazards.

Possible flood paths (see section 9.1) were searched for any sttlements and infrastructure. The results show that really few objects are located in the possible flood paths, even the worst-case scenario was considered. While no towns or villages are located in the runout paths of potencial GLOFs, the destruction of roads and bridges was concluded to be the main threat to the society.

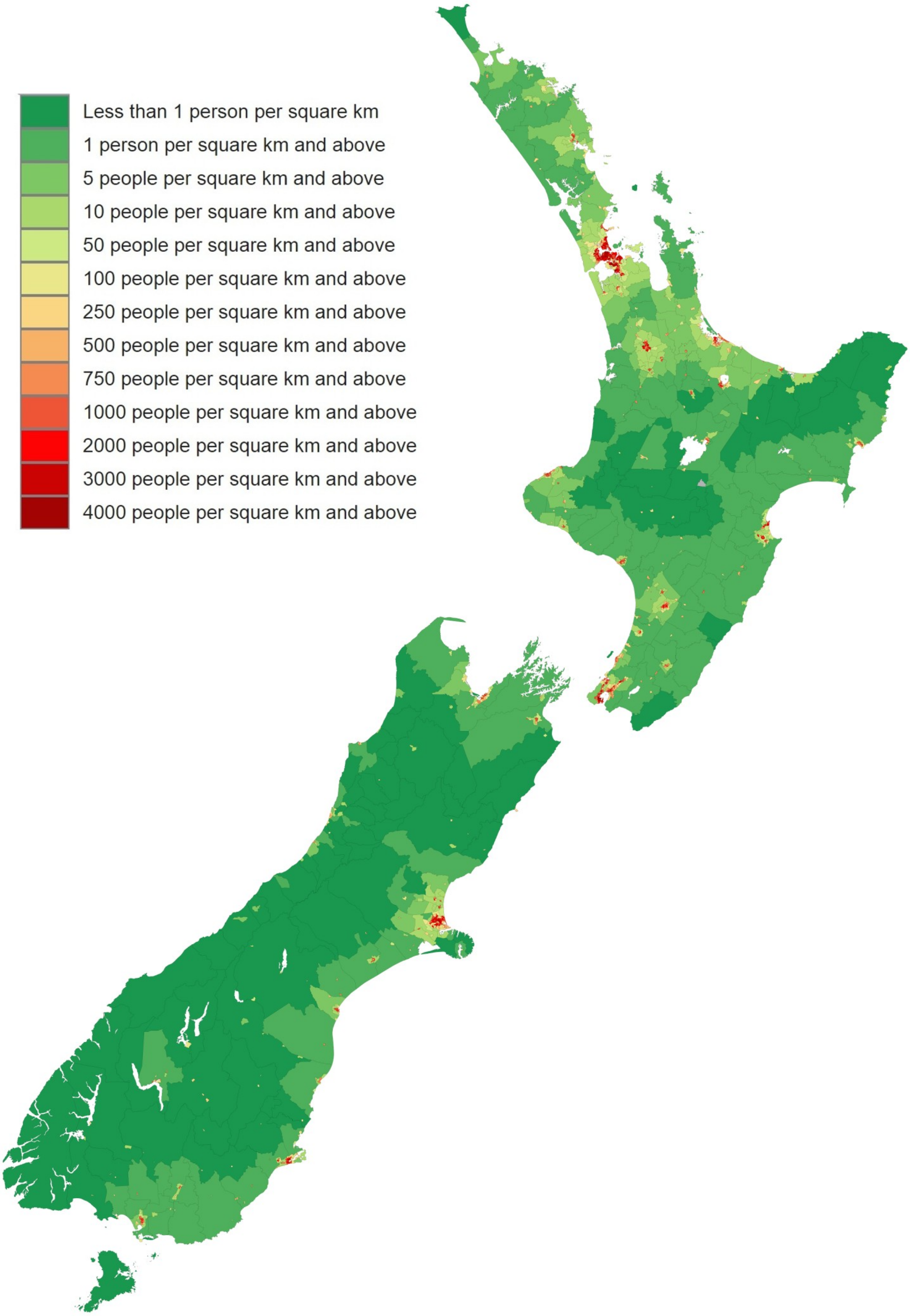

Figure 56: Map showing population density of New Zealand (by Statistics NZ Area Unit) as of the 2006 census. Source: Vardion (2008)

9.2.1 Risk for major roads

A potential GLOF from any Mt Aspiring area proglacial lake has no major road (State Highway – SH) in its way, thus only local roads cross or follow the potential flood paths. In Mt Cook area, State Highways follow the coast, both on the west and on the east. On the west, State Highway 6 (SH-6) is the only road connecting north with the south, west of the Main Divide. It is a popular tourist route, followed by thousands of visitors a year. On the east, State Highway 1 (SH-1) is the spine of the South Island road network, connecting the biggest cities and towns of the South Island. A key railway line follows SH1 for most of its course.

Figure 57 shows the intersections between potential flood paths and main roads of the South Island. While there are no main roads crossing or following the potential flood paths of Mt Aspiring area, all the threatened places are within Mt Cook area. Figure 57 **(a)** shows the location of SH-6 crossing the Karangarua river. In case of a GLOF from Douglas lake, the flood water would eventually flow to the Karangarua river and if it is a severe GLOF the bridge on SH-6 might be damaged or even destroyed. The same case might happen few kilometres northeast, where SH-6 crosses the Cook river – see Figure 57 **(b)**. The Cook river drains La Perouse Lake – one of the most hazardous proglacial lakes of New Zealand (see section 8.4).

East of the main divide, no major roads are crossing the potential flood paths

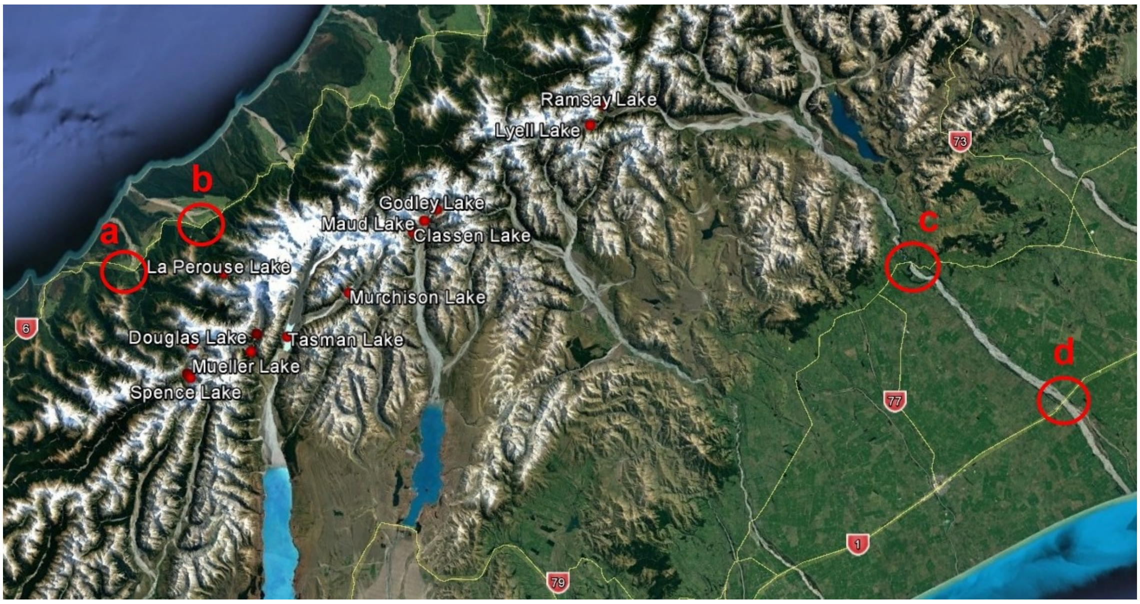

Figure 57: Map showing the intersection of potential flood paths with main roads of the South Island. (a) shows the location of SH-6 crossing the Karangarua river, (b) the crossing of SH-6 with the Cook river, (c) the crossing of the Rakaia river with SH-71, and (d) the crossing of the Rakaia river with SH-1. Source of the satellite image: Google Earth.

from Mueller, Hooker, Tasman, Murchison, Classen, Maud, and Godley lakes. Only Lyell and Ramsay lakes, drained by the Rakaia river pose a threat for to some major roads, in case of a GLOF. Two road bridges over the Rakaia river were located. Figure 57 **(c)** shows the location of SH-71 crossing the river in a section called Rakaia Gorge (see section 9.1) and **(d)** shows the location of SH-1 crossing the river southwest of Christchurch. Due to the extreme width of the braided Rakaia river (see section 9.1) a potential flood wave is expected to cause minimum harm to above mentioned bridges. However, the stability of both bridges should be assessed during GLOF propagation modelling.

9.2.2 Risk for minor roads, walking tracks, and other structures

A minor road, 4x4 track, a walking track, or a backcountry hut was recorded in flood paths of all proglacial lakes. While some roads or track are commonly visited, the other are visited barely few times a year. In Mt Aspiring area, only one minor road and two official walking tracks with small backcountry huts and shelters were located in the probable flood paths west of the Main Divide. East of the Main Divide, two access roads to Matukituki Valley and several walking tracks and huts were located.

The results from Mt Cook area show slightly higher number of minor roads, 4x4 tracks, walking tracks, backcountry huts and shelters and other structures which are located in the potential flood path. The most vulnerable area was detected around Mt Cook Village (see figure 58). During a serious GLOF from Hooker and/or Mueller Lake, highly visited walking tracks can be threatened, especially Hooker Track leading from White Horse Hill

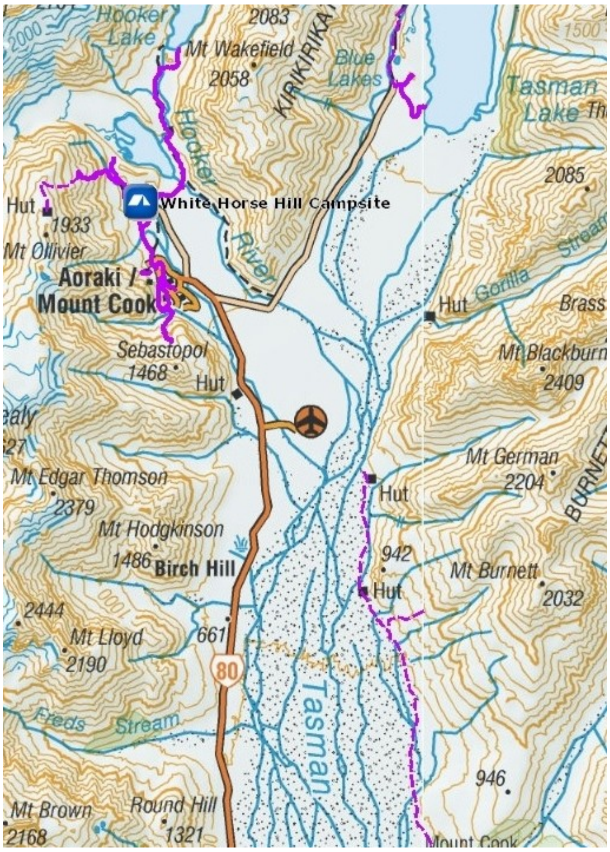

Figure 58: Topographic map showing the locations of some vulnerable features around Mt Cook Village. Source of the map: DoC

campsite to Hooker Lake. Stability of several swing bridges might be threatened and losses of life might occur. However, the results show that the probability of a GLOF to reach the campsite is relatively low, because the campsite is located ~5-20 m higher than the outlet of Mueller Lake. The campsite is also well hidden behind huge, well vegetated lateral moraine of Mueller lake with dam freeboard more than 100 m. The results suggested here are consistent with the results of Allen et al. (2009).

The results show that Mt Cook village is out of the potential flood wave reach, however, the access road (nr. 80) intersect on several places with the worst-case scenario potential flood path. Also the small Mt Cook airport (in the centre of figure 58) might be effected by a severe GLOF from Mueller, Hooker, and Tasman lakes. Small backcountry huts and shelters are often placed on elevated places or on the margins of braided rivers (see figure 59). Even the locations of those hiking and trekking facilities are being carefully selected to minimise a damage from floods and other natural hazards, the damage or destruction by a severe GLOF can not be neglected.

Figure 59: Location of Louper Bivouac on the margin of the braided Rakaia river. Well developed vegetation around the bivi indicates recent low flooding activity. During a severe GLOF flood wave might reach the hut. Lyell and Ramsay lakes are located few kilometres upstream (to the left on the background photograph) from the bivi. Source of the photographs: Archive of the author.

10 Discussion

Any scientific discipline is strongly influenced by the level of current knowledge, historical and cultural background and by the goals that want to be achieved. For instance the glacier fluctuations are currently being explained by changes of temperature and/or precipitation, however new forcings can be discovered and thus change the overall preception of current (western) science. Even the scientific knowledge is growing, there will never be an aswer to all the questions. This is exactly the case with GLOFs. We know that a GLOF can be triggered by an earthquake, but we do not know when it strikes, why, where exactly and what magnitude it will have. Is there a possibility of an earthquake with a magnitude e.g. 12? The fact that we have not experienced it yet does not mean it is not possible. If such an earthquake strikes, all the GLOF research that has been done might need to be re-evaluated.

The other common phenomenon of western science is the act of classification. Classification enables us to divide some objects (e.g. glaciers) according some characteristics they have (more or less) in common. Thereafter we can easily describe some properties, trends, or behaviour. On the other hand, classification disregards the uniqueness of the object and average various parameters. Lake origin is a good example (see section 10.2.3). All those limitations need to be considered while doing any assessments and detailed scientific studies. Some key limitations are discussed bellow.

10.1 Glacier fluctuations

Even the climate is the key factor influencing glacier mass balance, other factors needs to be used to explain glacier length changes. While many smaller and/or steeper New Zealand glaciers retreated and advanced several times during the 20th century, the fronts of large, low gradient valley glaciers remained unchanged until the 1970s, when they started to retreat rapidly (see section 5.2). Many “Glacier warming deniers” commonly use the 1983-2008 advance of some New Zealand glaciers to prove that no global warming is occurring. However, during this period the mighty Tasman glacier retreated of almost 5 kilometres, and similarly did many other glaciers of the same type. It is important to realise, that length of smaller, short response time glaciers respond swiftly to the annual mass balance and thus giving a good information about interannual climate variability. While the response might by enhanced by local, or regional conditions, and by a whole range of climate oscilations, we can not use those fluctuations as indicators of a global climate change.

To assess wether a significant climate change is occurring or not, long (at least 30 years) trends need to be used. While long response time glaciers generally show the trend directly by the distance the glacier retreated or advanced, short response time glaciers fluctuate back and forth annually and therefore the trends need to be constructed from regular glacier length observations.

Special caution needs to be given to factors and processes that decouple glacier fluctuations from climate forcings. One of the most significant process for the long, low gradient glaciers is the formation of proglacial lakes and initiation of calving. Calving speeds up the glacier retreat and modulates the overall recesion trend. Steeper glaciers are strongly influenced by underlying topography and by valley width changes. Apart of flow velocity adjustments, various terraine features can for instance divide a glacier to more sections during a retreat phase and reconnect them during an advance phase. Methodology of the length record then becomes complex and compromises often need to be done.

The interpretation of glacier fluctuations rise many challenges. Length fluctuations of a single glacier can not be used as an evidence of a climate change. A holistic approach needs to be used to better understand interactions between all Earth-spheres and to determine the rate of human contribution to glacier fluctuations.

10.2 Limitations of the lake selection for inventory

Lake classifications and inventories face several scientific challenges. Between the most important are: (1) minimum size threshold determination, (2) lake ephemerality assessment, (3) rapidity of the lake area/volume changes, (4) accurate lake type classification (e.g. according to lake origin).

10.2.1 Minimum size threshold determination and ephemerality

Based on a lake definition, a lake (and also a glacial lake) has no threshold of minimum area or volume, resulting in the fact that after a rain, tens, hundreds, thousands or even millions of lakes are being formed in the depresions of the land surface. Most of them last just for few hours or days, but some others last several months or years. While no spatial or temporal thresholds have been set, classification and final inventory is thus strongly subjective. The situation with supraglacial lakes is even more complicated. Supraglacial lakes change their size and location during only a single year and merge together. In this study no minimum lake size was determined, however, the lakes were selected according to a topographic map and Google Earth images -both with limited resolution. The total number of (proglacial) lakes can thus strongly vary according the input data used.

10.2.2 Rapidity of the lake area/volume changes

While tectonic lakes, or old Pleistocene glacial lakes generally maintain their area and volume for milenias, young glacial lakes often change their properties much faster. New lakes are being formed while some others are being drained. The initiation of calving speeds up glacier retreat and the newly formed proglacial lake starts to grow rapidly. The lake size measurements are then strongly dependent on the time of surveying. Lake volume can double just within few years (see section 5.4.2), which can strongly influence the hazard level estimated. Also a flood modelling can be strongly effected by a rapid lake growth, therefore some older studies needs to be approached with caution.

10.2.3 Accurate lake type classification

While we classiffy lakes according their origin to glacial lakes, landslide-dammed lakes...etc, the distinction between different types is sometimes really weak. The process of moraine formation can be coeval with numerous landslides to the moraine area and an expert then needs to decide wether it is more a glacial lake or a landslide-dammed lake. The classiffication to landslide-dammed category, however, does not mean that an outburst flood can not occur.

The other good example of a scientific challenge regarding lake type classification rises from the definition of a proglacial lake (see section 3.4). While proglacial lakes include lakes, which are physically attached to an ice margin, as well as lakes detached from, or immediately beyond, a contemporary ice margin (Carrivick and Tweed 2013), a question “How far from a glacier can a lake be located to be still classified as proglacial?” needs to be asked. Even more complex the problem starts to be when we consider the hanging glaciers and their “ofsprings” at the cliff bases bellow the hanging glaciers. How big needs a glacier by the cliff base be, to shift an adjacent lake to proglacial lake category?

The dam type classification in New Zealand environment also revealed some challenges. Many proglacial lakes of Mt Cook area are impounded behind outwash heads rather than large terminal moraines (see section 7). However, several small moraines can be detected around the outlets of proglacial lakes. The dam classification then becomes complicated and while no thresholds for minimum moraine height have been developed, the classification is ultimatelly based on an expert assessment.

Those and many similar questions need to be asked if we want to classiffy glacial lakes. While most definitions have their limitations, the classification process remains partly subjective.

10.3 Lakes excluded from GLOFs assessment

While this study focuses just on proglacial lakes of New Zealand, the other glacial lakes and other hazardous non-glacial lakes of New Zealand were excluded from the GLOFs hazard assessment. It was found that two categories of glacial lakes show sufficient stability and do not respond to glacial retreat. Those are all the old Pleistocene (or Pleistocene/Holocene boundary) glacial lakes (section 10.3.1) and tarns and other small glacial lakes in ice-free catchments (section 10.3.2). It was also discovered there are lakes in New Zealand to be prone to outburst floods even they are not of glacial origin. Those are mainly landslide-dammed lakes and then the Crater lake of Mt. Ruapehu!

10.3.1 Big lakes dammed by Pleistocene moraines

Moraine dams and lake banks of old glacial lakes formed during Pleistocene or Pleistocene/Holocene transition (further called “old glacial lakes”) have been stabilised by wide range of natural processes and human activities alike. Most of the big lakes, mentioned in section 4.3 have wide stabilised moraines, or man-made dams. Water from the lakes is mostly used for power generation, and therefore the water level fluctuates significantly. The moraines were often strengthened during the construction of canals and pipelines. Also the spillways and other outlets are often made of concrete. From those reasons the hazard of a dam burst or dam overtopping is assumed to be marginal.

It is also important to realise that moraines around those big glacial lakes are just minor features contributing relatively little to the depth of the lake (Lowe and Green 1987) and the deepest parts of some of them, including Manapouri, Te Anau, and Wakatipu, are below sea level, forming cryptodepressions (Irwin 1972a in Lowe and Green), therefore a complete drainage of those lakes is not possible.

In spite of this stability, there are few processes that can lead to an outburst flood from above mentioned lakes. The first one to be named is a strong earthquake with epicentre located in the vicinity of the lake dam. However, no earthquake in modern history of New Zealand has led to a serious damage on the dam, to a dam rupture or to a dam overtopping of an old glacial lake. In the case of an earthquake with M > 7, we could expect intensities of about VIII (Modified Mercalli Intensity) and greater (USGS 2017). Such intensities occurred during the earthquakes near Christchurch in 2010 (MMI = VIII) and 2011 (MMI = IX) and during the earthquake near Amberley in 2016 (MMI = IX). While the intensity dissipated quickly from the epicentre, no damages have been observed on the lake dams tens of kilometres away. However, the occurrence of an earthquake with an intensity greater than VIII near the lake dam can not be excluded and neither the possibility of a GLOF.

Regarding a massive slope movement into the lake, the situation appears to be even more complex. Whole range of factors controls the susceptibility of a slope to a mass movement. In case of their combination a catastrophic mass movement can occur and create a reservoir displacement wave (Korup 2005). In case of a mass movement of millions or billions of cubic metres even big water bodies can be threatened by dam overtopping or dam failure (e.g. RCEM 2014). Prediction of catastrophic mass movements is a challenging scientific problem and at the same time the probability is relatively low compare to smaller mass movements, therefore the catastrophic mass movements are not usually studied while assessing hazards from GLOFs.

The last factor to consider while assessing the hazards of old glacial lakes is the hazard of a flood wave from a lake situated upstream. Each lake larger than 0.02 km^2 contains at least $6x10^5$ m^3 of water; if it breaches, downstream valleys could suffer hazardous consequences (Samjwal et al. 2007). According to Emmer and Vilímek (2014) there are two components to assess the potential for dam overtopping following a flood wave originating in a lake situated upstream. Those are: (a) retention potential of a lake situated downstream and (b) potential for a flood wave from a lake situated upstream. The retention capacity of the old glacial lakes was estimated to be relatively high due to dam freeboard of several meters and enormous lake areas (figure 60). The retention capacity of the valley between the lakes should be quantified as well. While all the valleys between selected young glacial lakes upstream and big old glacial lakes downstream consist mostly of wide outwash plains with braided rivers (see section 9.1) the retention potential is considered to be high.

Those are the main reasons why old glacial lakes were not studied independently in this thesis. Old glacial lakes are not directly connected to a glacier and there is no glacier on the slopes above them. The volume of water in the lake is still influenced by glacial meltwater but no extreme events are expected due to large volume of water in old glacial lakes compare to the volume of water in the glaciers.

Figure 60: „Old glacial lake“ Pukaki as seen from the Pleistocene terminal moraine. With the area of 178.8 km^2 and water volume of 4.66 km^3 potential flood wave can be spread easily. Aoraki / Mt. Cook in the background. Source: Archive of the author.

10.3.2 Glacial lakes in ice-free catchments

The next type of lakes that were not incorporated are small glacial lakes in the cirques or upper parts of mountains and tarns which are not surrounded by glaciers and have relatively small catchment area. Those lakes are typical in granitic Darran Mountains of Fiordland National Park (e.g. Lake Marian or Lake Adelaide) or schistose Humboldt Mountains (e.g. Lake Harris, Lake Mackenzie or Lake Unknown). While their formation is related to the erosive activity of cirque glaciers, the criteria of glacial origin was met, but absence of a glacier in its vicinity or even in the whole catchment results in the reduction of hazards from GLOFs. Furthermore the lakes are being fed mostly by rainwater and the lakes are mostly too small to create a catastrophic flood wave. Therefore those lakes were excluded from the detailed GLOFs analysis.

10.3.3 Hazardous lakes of non-glacial origin

While the definition of a proglacial lake states that “Proglacial lakes are masses of water impounded at the edge of a glacier...” (see section 3.4) even the lakes of non-glacial origin surrounded by glaciers should be incorporated. In New Zealand there is just one such a lake. It is the Crater lake of Mt Ruapehu – the highest mountain in the the North Island. With the altitude of 2540 m a.s.l. Crater lake is the highest lake in New Zealand. Even surrounded by a glacier, the lake is of volcanic origin and volcanism is the key factor for triggering an outburst flood. Due to high volcanic activity of Mt Ruapehu, Crater Lake poses a significant threat to the society. During an eruption, the Crater Lake generally overflows and create lahars flowing down the slopes of Mt. Ruapehu (e.g. Scott 2013), threatening two popular ski areas which are located on the southern and northwestern flanks.

Also landslide- and rockslide-dammed lakes are of a special interest. Landslide-dammed lakes are amongst the most obvious and widely recognized of such features in New Zealand (Korup 2004) and pose a significant threat to people and infrastructure even of more densely inhabited areas. While this thesis focuses on high mountain environment and glacial lakes, the interaction between landslide-dammed lakes and glacial waters will be briefly outlined.

There are several landslide-induced mechanisms that can lead to an outburst flood. The most common one is the process of a river channel blockage, followed by formation of a landslide-dammed lake and an outburst. Second mechanism of an outburst flood is based on the direct impact of a mass-movement into a lake. Thirdly, a landslide can terminate in a (moraine) dam area of a glacial lake. In that case, the lake outlet can be blocked leading to lake enlargement and possible dam breach. The last mechanism to be aware of is the landsliding to proglacial area. If a landslide blocks meltwater streams flowing from a glacier, new lake can be formed. This lake can grow in size, reach the glacier front upstream and initiate the process of calving, which further speeds up the glacier retreat and potentially the lake enlargement. A landslide can thus initiate formation of proglacial lake – the other example of a proglacial lake of non-glacial origin.

10.4 Limitations of the GLOFs assessment method

The goal of a hazard assessment is to identify natural processes that can potentially cause death, injury, or property loss and further estimate their potential severity (magnitude and frequency). The process of hazard assessment is strongly dependent on the current level of knowledge about the natural process studied, quality of input data, and method used. While the development of all those "foundation stones" is infinite, a hazard assessment can never reach the probability level of 100 %.

Various methods have been developed to assess the GLOFs hazards throughout the world (see section 3.5). The method presented here is a first-order, qualitative method based on remote sensing data. Those characteristics result in whole range of limitations. While first-order methods are used for overall assessment and location of the most hazardous areas, exact discharge volumes, probable outburst times, and other detailed characteristics are not known from the results of those methods.

The method presented here (same as other qualitative methods) has no (or just few) thresholds determined, resulting in high subjectivity level of the method (based on expert assessment). The results are thus strongly influenced by the knowledge of the expert and input data used. Repeated use is thus considerably limited. However, the determination of some thresholds does not necessarily increase the accuracy and transferability of the method. If, for instance, slope threshold for mass movements is determined, other important factors influencing the gravitational movements like lithology, slope aspect, climate, fracturing etc. are neglected, resulting in different results if transferred elsewhere.

Although determining the probability of occurrence is difficult for glacial hazards, for the purpose of practicality it is better to assign a probability, even if approximate and subjective, than not at all (Fell 1994). Probability of various hazards thus varies between the authors. In this study the hazard assessment was based on the summation of "probability points" for each lake. It is important to note, that a higher number in the sum does not necessarily mean a higher GLOF hazard level, because some extreme hazards can be concealed due to the nature of the summation. Where e.g.

“Lake A” has a “Medium (1)” probability for all the triggers (1;1;1;1;1;1) the total count is 6, and “Lake B” has some “Low (0)” and some “High (2)” probabilities e.g. (0;0;2;2;0;0) the total count is 4, however, some processes with “High (2)” probabilities can lead to serious GLOFs, even the probability of the other triggers is considered to be marginal. The interpretation of the total count thus needs to be approached with caution.

Also the limitations of the methods based on remote sensing data are obvious. Limited resolution, time span, readability (due to cloud cover, or snow cover), and other limitations strongly influence the results. The remote sensing data are, however, the key tool for the first-order assessments.

The last important limitation of most GLOFs hazard assessment methods is the fact that the methods focus just on existing lakes, or current lake size. The probability of a mass-movement to a lake, for example, can be classified as low, however, the lake can grow rapidly and within few years the lake surface can be exposed to impacts from slopes above the lake. Those scenarios are often not incorporated in the current methods. Longer term glacial recession and lake expansion will increase the potential for ice, debris or rock impacts into most active glacial lakes in New Zealand (Allen et al. 2009), hence the expected lake growth should be incorporated in detailed hazard assessments.

10.5 Past GLOFs events in New Zealand

Examination of past GLOFs events is a significant tool in the process of hazard assessment (Campbell et al. 2005). New Zealand historical events, however, do not give sufficient information that could be used for further research. The other limitation is the fact that just brief descriptions exist. Especially the outburst flood at Kea Point in 1913 conceals many uncertainties. Irvine-Fynn et al. (2015) described this event as a GLOF, however, the presence of a lake can be highly questionable. While the authors describe a rapid lake expansion at the termini of many larger glaciers since 1970s, and an increase of the GLOFs hazards, at the same time they describe a **glacier lake** outburst flood that appeared in 1913! Nowadays the lake level of Mueller lake is more than 100 metres bellow Kea Point (the lowermost place of the lateral moraine) where the outburst flood occurred in 1913, limiting the possibility of a further GLOF at the same location.

Based on the rates of glacier surface downwasting and the history of proglacial lakes of the Mt Cook region, the presence of a proglacial lake in 1913 was excluded. It is suggested that the surface of Mueller Glacier was as high as Kea Point, or even higher. Two possible explanations of the 1913 outburst flood were then suggested. First, a supraglacial lake might have occupied the surface of Mueller Glacier. An extreme rain event then caused the lake to overflow and emerge trough Kea Point. The second scenario assumes an outburst flood from within the glacier body. An englacial outburst flood similar to outburst floods from Franz Josef Glacier (see section 6.3) might have occurred. If the second scenario is validated, Kea Point outburst flood should not be termed a GLOF.

My findings about Kea Point outburst flood has been sent to Mr. Tristram Irvine-Fynn PhD (the author of the study), but no reply (except of automatic respond shown bellow) had came by the submission date of this thesis.

"I am away from the office until the 24th April, with limited time available to reply to non essential emails, and I will respond on my return. Apologies for any delay.

Best wishes, Tris."

10.6 Limitations of the GLOFs assessment results

Both, the current scientific knowledge, and the quality of input data effect the complexity of method that can be used for GLOFs hazard evaluation. Even based on previous results and expert knowledge, each method is strongly subjective. The number and type of characteristics assessed and threshold determination are the main variables effecting the final result. The method presented here suffers the same limitations. The characteristics used were selected based on knowledge gained from various scientific literature and after an examination of the local specifics of the Southern Alps of New Zealand. The reasons for (not)incorporating some characteristics are given bellow.

For instance, no "hazard points" were assigned for the lakes that previously experienced a GLOF. This decision was made after a detailed assessment of past GLOF events. While the outburst flood at Kea Point in 1913 (see section 6.2) was probably not an outburst flood from a glacial lake (see section 10.5), and currently the lake level is more than 100 metres bellow the moraine crest at Kea Point, the probability of a GLOF on this location is really low. Similarly no "hazard points" were assigned for the GLOF from Maud Lake in 1992. The reason for this decision is simply that Maud Lake itself does not show any signs of instability. The GLOF was triggered by a rock avalanche from Mt Fletcher – a peak located at the Main Divide. The evaluation of possible rock/ice avalanches revealed that Mt Fletcher has no special position between the other peaks of the Main Divide and thus the probability of rock/ice avalanches from the other peaks of the Main Divided was concluded to be equal.

The other characteristic not included in the hazard assessment was the characteristic describing the rate of lake growth. The reason for the incorporation of this characteristic by some authors is simply based on the fact that as the lake volume increases, so does the possibility of a moraine-dam break (e.g. Janský et al. 2010). However, the rapidly growing lakes of the Southern Alps of New Zealand are mostly dammed by outwash heads without any significant moraines present, and thus limiting the possibility of a dam break. For example, the lake level elevation of the rapidly growing Tasman Lake (705 m a.s.l.) is similar to the elevation of the proglacial area where the Tasman River starts to braid 1 km downstream (Purdie et al. 2016) and thus the “dam” overtopping appears to be the only GLOF process possible.

Many other local characteristics were examined to achieve as accurate results as possible for the first-order assessment. However, the results presented, can differ from the results of other authors evaluating the GLOFs hazards in the future. Interdisciplinary approach is encouraged to understand nature processes into a detail and to create a robust background for a detailed hazard assessment of the most hazardous lakes derived from this study.

11 Conclusion

11.1 The analysis of glacier retreat in New Zealand

The analysis of glacier retreat in New Zealand since late Pliocene (~2.6 Ma) was done by using numerous scientific sources. Even the proxy records are not continuous (a hiatus of more than one million years appears), Pleistocene glacial history can be partly reconstructed. Four glaciations commencing at MIS 10 (Marine Isotope Stage) (approx. 410 Ma) have been identified. The LGM (Last Glacial Maximum) occurred during late MIS 3 stage (~30 ka), approximately 12 000 years before the global ice sheet LGM. During the Otiran Glaciation (the last glacial) and LGIT (Last Glacial-Interglacial Transition) large glacial lakes were formed on the margins of the Southern Alps and glaciers retreated rapidly.

There were several large glacier re-advances during the LGIT, and many smaller re-advances during the Holocene indicating continuous climate oscillations. Interglacial climatic optimum of the Holocene was dated to occur between 8 – 5 ka, before the neoglacial activity commenced at about 5 ka BP. Since that time numerous retreats and readvances have been dated, however, the results vary according the study location and methods used. The Holocene record shows notable interhemispheric disparity in the timing of the maximum ice extent. Largest advances of glaciers in New Zealand occurred in the early Holocene, whereas the LIA (Little Ice Age) advances were more restricted. The Mount Cook glaciers were further advanced about 6500 years ago than at any subsequent time. In contrast, most Northern Hemisphere glaciers reached their greatest Holocene extents during the LIA (1300 to 1860 A. D.). While some New Zealand glaciers advanced during northern warm periods characterised by diminished northern glaciers, broad similarities were apparent during the past 700 years (the northern LIA) indicating that mid- to late Holocene glacier fluctuations were neither in phase nor strictly antiphased between the hemispheres.

Since the late 19th century most New Zealand glaciers downwasted and retreated significantly. However, not all the glaciers responded in the same way. Long response time, low gradient valley glaciers downwasted significantly during the first half of 20th century, but maintained their 19th century terminus positions until 1970s when proglacial lakes started to be formed, calving was initiated and the retreat commenced. Even the net mass balance has varied during the last 50 years, long response time glaciers have continued to retreat and lose their total mass. The length and mass evolution of Tasman Glacier was described in detail in this thesis to illustrate the behaviour of long response time glaciers.

On the other hand, steeper, short response time glaciers were retreating continuously since the end of 19th century, with a dramatic drop around the 1940s and 1950s. The most significant advance occurred during the last two decades of the 20th century. 58 New Zealand glaciers advanced at some point in the 1980s, 1990s and early 2000s and 12 of these advanced continuously for five or more years. In 2005, when this glacier advance phase neared its end, 15 of the 26 advancing glaciers observed worldwide were in New Zealand.

Franz Josef Glacier – a glacier with the most detailed length record in New Zealand well represents short response time glaciers. Overall retreat since the 1930s was interrupted by two small re-advances. In 1968 a significant and long-lasting retreat commenced. By 1983 Franz Josef Glacier was the shortest since measurement began. The most significant advance phase since 19th century (interrupted by a small retreat in 1991) occurred between 1983 – 1999, when the Franz Josef Glacier reached a length not seen since 1960. After 1999 the glacier retreated until in 2005 the next (and most recent) advance began. This most recent advance ceased in 2008 and Franz Josef Glacier is currently (2017) retreating. In total, ice volume of the Southern Alps has decreased from 54.53 km^3 in 1976 to 46.12 km^3 in 2008 (a loss of 8.41 km^3 or 15%).

It was concluded that the behaviour of New Zealand glaciers is driven by their response times (influenced by glacier size/thickness), longitudinal profiles (steepness), mass turnovers, and terminus ablation rates (influenced by the ablation zone climate, by the presence of debris cover on the lower glacier tongues, and by the presence of proglacial lake).

While some authors suggest that glacier advances in New Zealand are a result of alternated precipitation patterns, the most recent studies show that from 1972 to 1979, precipitation had a negative influence on mass balance (resulting in volume loss), while between 1979 and ~2000, it had a positive influence. On the other hand temperature correlation appears to be much stronger and it is believed that temperature drop is responsible for the glacier advances between 1983 and 2008. While both factors might be responsible for glacier variations, it was concluded that advances of New Zealand glaciers are a result of decreased summer ablation. Northward shift of the subtropical high-pressure zone and zone of southern westerlies leads to the decrease of SSTs (Sea

Surface Temperatures) in the Tasman Sea during austral spring and summer, leading to higher snow accumulations in spring, and smaller ablation in summer and thus to glacier advances. Interdecadal Pacific Oscillation (IPO) has been an important influence on glacier behaviour in New Zealand over the past few decades. During the positive phases colder and wetter conditions were observed in New Zealand's Southern Alps and during negative phases warmer and drier conditions were observed. These changes are well reflected in New Zealand glacier length fluctuations. The effects of El Niño–Southern Oscillation (ENSO) to New Zealand glaciers are much more complex.

11.2 Evaluation of GLOFs hazards in New Zealand

To process an evaluation of GLOFs hazards in New Zealand and possible risks for society, several steps had to be done. First, a detailed background research related to previous GLOFs in New Zealand has been undertaken. Secondly, a group of potentially hazardous lakes had to be selected. Thirdly, an inventory of those lakes including geomorphic and geometric properties had to be created, and lastly an effective method for GLOFs hazard assessment had to be introduced. After those steps actual hazard and risk evaluation was completed.

Background research about previous GLOFs events revealed very little about the nature, location, and frequency of New Zealand GLOFs. While there were only three GLOFs events, recorded in New Zealand history, outburst floods from Franz Josef Glacier were studied as well. Just a brief description exists about the outburst flood from Mueller Glacier in 1913. Rain induced flood appeared at the place called Kea Point and destroyed old Hermitage Hotel close to Mt Cook Village.

Outburst floods from Franz Josef have occurred several times in the history in three different modes. Most of the floods are associated with high rainfall and advanced phase of Franz Josef Glacier. Even the mechanisms of outburst floods from Franz Josef vary between the cases, most of them are associated with a blockage of englacial conduits. The water is then re-routed to supraglacial position through moulins or escapes the glacier through the front of the glacier tongue.

The only two New Zealand GLOFs recorded in recent history are the floods from Maud Lake in May and September 1992. Rock avalanches from Mt Fletcher reached the lake and caused a displacement wave. In May 1992 the braided gravel bed of Godley River was extensively scoured for several kilometres and the flood travelled

45 km to Lake Tekapo, where it raised the lake level for ~90 mm.

An inventory of all proglacial lakes in New Zealand revealed there are 25 proglacial lakes in New Zealand; all of them located within the limits of the Southern Alps, in two distinct areas: Mt Aspiring area and Mt Cook area. While Mt Aspiring area favours more to small bedrock-dammed proglacial lakes, the dominant dam type of proglacial lakes within the Mt Cook area is outwash head. Even small terminal moraines are evident by many large proglacial lakes of Mt Cook area, the lakes are not directly impounded by the moraine material. Rather, the lake basins are developing behind large Holocene outwash heads, which further results in their high overall stability. Various lake properties were recorded and further used in the hazard assessment process.

The GLOFs hazard assessment of all New Zealand proglacial lakes was divided into two methodological steps. The first step include hazard assessment of potential GLOF triggers and the second include the lake/dam stability assessment of every single proglacial lake. The first step evaluates the probability of a GLOF caused by a) falling ice from calving front, b) falling ice from hanging glacier, c) snow avalanches, d) rock/ice avalanches, e) landslides, f) moraine collapse. The results show that in Mt Aspiring area Volta Lake is the most hazardous from the point of a GLOF trigger probability. In Mt Cook area the most hazardous lakes from a GLOF trigger probability are Mueller Lake, La Perouse Lake and Lyell Lake.

The second step assessing the lake/dam stability was designed to determine the probability of a dam break and displacement wave release alike. The assessment process includes four parameters: a) Lake length, b) dam freeboard, c) moraine width-to-height ratio, and d) evident seepage. Volta Lake and Lucidus Lake were concluded to be the most hazardous lakes in the terms of lake/dam stability in Mt Aspiring area. In Mt Cook area eight lakes reached equal number of total points and were concluded to be the most hazardous in the terms of lake/dam stability. Those are: Douglas Lake, Tasman Lake, Murchison Lake, La Perouse Lake, Maud Lake, Godley Lake, Lyell Lake and Ramsay Lake.

By the summation of the total "hazard points" the overall hazard level was reached. The most hazardous lakes (13 "hazard points" reached) are Volta Lake in Mt Aspiring area, and La Perouse Lake and Lyell Lake in Mt Cook area. It was recommended to conduct further, more detailed studies for those three lakes and also for the group of proglacial lakes that reached 10 or more "hazard points".

Risk assessment was also divided into two steps. First, the flood channel assessment was done to estimate the probable flood path during the worst case scenario. Secondly an intersection between the potential flood path and human infrastructure was searched. It was concluded that no cities, towns, or villages are in the risk of a GLOF. On four locations the possible flood path intersect with major roads (state highways), resulting in possible damage or destruction of four road bridges and one railway bridge. Several local roads, 4x4 tracks, hiking tracks and backcountry huts are in a risk, especially if the worst-case scenario is considered.

Due to high number of visitors to Mt Cook national park from Mt Cook Village, walking track towards Hooker Lake, Mt Cook airport and parts of the access road to Mt Cook Village were concluded to be the most vulnerable if the combined effects of physical and social vulnerability are considered. Further development in this region should be properly assessed; not only from the point of GLOFs hazards, but also from the point of other natural hazards.

12 References

- Adger, W. N., 2006: Vulnerability. Global Environmental Change, 16 (3), 268-281
- Allen, S. K., Owens, I. F., Huggel, C., 2008: A first estimate of mountain permafrost distribution in the Mount Cook region of New Zealand's Southern Alps. In: 9th International Conference on Permafrost , Fairbanks, Alaska, 29 June 2008 - 03 July 2008, 37-42.
- Allen, S. K., Schneider, D., 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
- Allen, S. K., Cox, S. C., Owens, I. F., 2011: Rock avalanches and other landslides in the central Southern Alps of New Zealand: a regional study considering possible climate change impacts. Landslides, 8, 33-48
- Anderson, B., Mackintosh, A., 2006: Temperature change is the major driver of late-glacial and Holocene glacier fluctuations in New Zealand. Geology, 34 (2), 121–124
- Birkeland, P. V., 1982: Subdivision of Holocene glacial deposits, Ben Ohau Range, New Zealand, using relative dating methods. Geological Society of America Bulletin, 93, 433-449
- Björnsson, H., 2003: Subglacial lakes and jökulhlaups in Iceland. Global and Planetary Change, 35/3-4, 255-271
- Breien, H., De Blasio, F.V., Elverhøi, A., Høeg, K., 2008: Erosion and morphology of a debris flow caused by a glacial lake outburst flood, Western Norway. Landslides, 5/3, 271-280
- Burrows, C. J., 1973: Studies on some glacial moraines in New Zealand-2. New Zealand Journal of Geology and Geophysics, 16/4, 831-856.
- Burrows, C. J., Chinn, T., Kelly, M., 1976: Glacial activity in New Zealand near the Pleistocene-Holocene boundary in the light of new radiocarbon dates. Boreas, 5, 57-60.
- Campbell, J. G., Li, X., Tongliang, G., Partap, T., Arora, B. R., Soomro, B., 2005: Inventory of Glaciers and Glacial Lakes and the Identification of Potential Glacial Lake Outburst Floods (GLOFs) Affected by Global Warming in the Mountains of India, Pakistan and China/Tibet Autonomous Region. Final report for APN project 2004-03-CMY-Campbell, ICIMOD, 48 p.
- Carrivick, J. L., Tweed, F. S., 2003: Proglacial lakes: character, behaviour and geological importance. Quaternary Science Reviews, 78, 34-52
- Carrivick, J. L., 2010: Dam break – Outburst flood propagation and transient hydraulics: A geosciences perspective. Journal of Hydrology, 380, 338-355.
- Chinn, T., 1996a: New Zealand glacier responses to climate change of the past century. New Zealand Journal of Geology and Geophysics, 39, 415-428.
- Chinn, T., 1996b: The southern hemisphere glacial record – Antarctica and New Zealand. Papers and Proceedings of the Royal Society of Tasmania, 130(2), 17-24.
- Chinn, T., 2001: Distribution of the glacial water resources of New Zealand. Journal of Hydrology (NZ), 40(2), 139-187.
- Chinn, T., 2009: The (once) Great Tasman Glacier, Department of Conservation Information Brochure.
- Chinn, T., Fitzharris, B. B., Willsman, A., Salinger, M. J., 2012: Annual ice volume changes 1976–2008 for the New Zealand Southern Alps. Global and

Planetary Change, 92-93, 105–118.
- Climate-data.org, Climatic data server: https://en.climate-data.org/location/143093/ accessed: 18.2.2017
- Cohen, K. M., Finney, S. C., Gibbard, P. L., Fan, J. X., 2013: The ICS International Chronostratigraphic Chart. Episodes, 36, 199-204.
- Cook, E. R., Palmer, J. G., D'Arrigo, R. D., 2002: Evidence for a 'Medieval Warm Period' in a 1,100 year tree-ring reconstruction of past austral summer temperatures in New Zealand. Geophysical Research Letters, 29/14, 1667
- Christchurch City Libraries: Mt. Cook and the old Hermitage before it was destroyed by flooding in 1913, File Reference CCL Photo Collection 22, Img00344. http://christchurchcitylibraries.com/Heritage/Photos/Collection22/00344.asp accessed: 7.3.2017
- Davies, T., Smart, C., Turnbull, J., 2003: Water an Sediment Outbursts from Advanced Franz Josef Glacier, New Zealand. Earth Surface Processes and Landforms, 28, 1081-1096.
- Doc – Department of Conservation. Aoraki/Mount Cook Long Term Community Plan 2009-2019. Section 8.3. http://www.doc.govt.nz/about-us/our-policies-and-plans/statutory-plans/statutory-plan-publications/national-park-management/aoraki-mount-cook-long-term-community-plan-2009-2019/
- Donner, W., Havidán R., 2011: Disaster risk and vulnerability: The role and impact of population and society. Population Reference Bureau.
- Dykes, R. C., Brook, M. S., Winkler, S., 2010: The contemporary retreat of Tasman Glacier, Southern Alps, New Zealand, and the evolution of Tasman proglacial lake since AD 2000. Erdkunde, 64 (2), 141-154
- Embleton, C., King, C., 1968: Glacial and Periglacial Geomorphology. Edward Arnold (Publishers) Ltd., 608 p.
- Emmer, A., 2013: Potenciální nebezpečnost vybraných morénami hrazených jezer v pohoří Cordillera Blanca, Peru. Master Thesis. Chrles University in Prague.
- Emmer, A., Vilímek, V., 2013: Review Article: Lake and breach hazard assessment for moraine-dammed lakes: an example from the Cordillera Blanca (Peru). Natural Hazards and Earth System Sciences, 13, 1551–1565
- Emmer, A., Vilímek, V., 2014: New method for assessing the susceptibility of glacial lakes to outburst floods in the Cordillera Blanca, Peru. Hydrology and Earth System Sciences, 18, 3461-3479
- Emmer, A., Klimeš, J., Mergili, M., Vilímek, V., Cochachin, A., 2016: 882 lakes of the Cordillera Blanca: An inventory, classification, evolution and assessment of susceptibility to outburst floods. Catena, 147, 269-279
- Fell, R., 1994: Landslide risk assessment and acceptable risk. Canadian Geotechnical Journal, 31, 261–272.
- Fitzharris, B. B., Hay, J. E., Jones, P. D., 1992: Behaviour of New Zealand glaciers and atmospheric circulation changes over the past 130 years. The Holocene, 2, 97-106
- Fitzharris, B. B., Garr, C. E., 1995: Simulation of Past Variability in Seasonal Snow in the Southern Alps, New Zealand. Annals of Glaciology, 21(1), 377-382
- Fitzsimons, S. J., 1997: Late-glacial and early Holocene glacier activity in the Southern Alps, New Zealand. Quaternary International, 38/39, 69-76
- Gellatly, A. F., Röthlisberger, F., Geyh, M., 1985: Holocene glacier variation in New Zealand (South Island). Zeitschrift für Gletscherkunde und

Glazialgeologie, 21, 265-273

- Gellatly, A. F., Chinn, T. J., Röthlisberger, F., 1988: Holocene glacier variations in New Zealand: A Review. Quaternary Science Reviews, 7, 227-242
- Gibbard, P., Head, M.J., 2010: The newly-ratified definition of the Quaternary System/Period and redefinition of the Pleistocene Series/Epoch, and comparison of proposals advanced prior to formal ratification. Episodes, 33/3, 152-158
- Goddard, P., 2008: Avalanche Awareness in the New Zealand Backcountry. New Zealand Alpine Club, 96 p.
- Haeberli, W., 1983: Frequency and Characteristics of Glacier Floods in the Swiss Alps. Annals of Glaciology, 4 (1), 85-90
- Hancox, G. T., Thompson, R., 2013: The January 2013 Mt Haast Rock Avalanche and Ball Ridge Rock Fall in Aoraki/Mt Cook National Park, New Zealand. GNS Science Report, 2013/33, 26 p.
- Harrison, S., Glasser, N., Winchester, V., Haresign, E., Warren, C., Jansson, K., 2006: A glacial lake outburst flood associated with recent mountain glacier retreat, Patagonian Andes. The Holocene, 16/4, 611-620
- Henderson, R. D., Thompson, S. M., 1999: Extreme rainfalls in the Southern Alps of New Zealand. Journal of Hydrology (New Zealand), 38(2), 309-330.
- Hochstein, M. P., Claridge, D., Henrys, S. A., Pyne, A., Nobes, D. C., Leary, S. F., 1995: Downwasting of the Tasman Glacier, South Island, New Zealand: Changes in the terminus region between 1971 and 1993. New Zealand Journal of Geology and Geophysics, 38/1, 1-16
- Huggel, C., Kääb, A., Haeberli, W., Teysseire, P., and Paul, F., 2004: An assessment procedure for glacial hazards in the Swiss Alps. Can. Geotech. J., 41, 1068–1083
- IPCC, 2013: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp.
- Irvine-Fynn, T., Williams, R., Carr, R., Williams, J., 2015: Reconstructing the Kea Point glacier lake outburst flood, Southern Alps, New Zealand. BSG Grant Reports 2014-2015, 1 p.
- Irwin, J., 1975: Checklist of New Zealand lakes. New Zealand Oceanographic Institute, Wellington, 160 p.
- Janský, B., Šobr, M., Engel, Z., 2010: Outburst flood hazard: Case studies from the Tien-Shan Mountains, Kyrgyzstan. Limnologica, 40, 358-364
- Kirkbride, M. P., 1993: The temporal significance of transitions from melting to calving termini at glaciers in the central Southern Alps of New Zealand. The Holocene, 3 (3), 232-240
- Kirkbride, M. P., Warren, C. R., 1999: Tasman Glacier, New Zealand: 20th-century thinning and predicted calving retreat. Global and Planetary Change, 22, 11–28
- Korup, O., 2004: Geomorphic characteristics of New Zealand landslide dams. Engineering Geology, 73, 13-35
- Korup, O., 2005: Geomorphic hazard assessment of landslide dams in South Westland, New Zealand: Fundamental problems and approaches. Geomorphology, 66, 167-188
- Lowe, D. J., Green, J. D., 1987: Origins and development of the lakes. In A. B.

Viner (Ed.), Inland Waters of New Zealand (p. 1–64). Wellington: New Zealand Department of Scientific and Industrial Research.

- Lowe, D. J., Green, J. D., 1992: 'Lakes.' In Landforms of New Zealand, 2nd ed., 108. Auckland: Longman Paul. Used from Te Ara - The Encyclopedia of New Zealand
- Mackintosh, A. N., Anderson, B. M., Lorrey, A. M., Renwick, J. A., Frei, P., Dean, S. M., 2017: Regional cooling caused recent New Zealand glacier advances in a period of global warming. Nature Communications, 8, 14202
- Mann, M. E., 2002: Little Ice Age. Volume 1, The Earth system: physical and chemical dimensions of global environmental change, pp 504–509. in Encyclopedia of Global Environmental Change. John Wiley & Sons, Ltd, Chichester
- Marcus, M. G., Moore, R. D., Owens, I. F., 1985: Short-term estimates of surface energy transfers and ablation on the lower Franz Josef Glacier, South Westland, New Zealand. New Zealand Journal of Geology and Geophysics, 28, 559-567
- MBIE - Ministry of Business, Innovation and Employment. Key tourism statistics, 2017. http://www.mbie.govt.nz/info-services/sectors-industries/tourism/key-tourism-statistics
- McGeorge, P., 2007: A land of water: rivers & lakes of New Zealand. David Bateman, Auckland, 144 p.
- McGregor, V. R., 1967: Holocene moraines and rock glaciers of the central Ben Ohau Range, South Canterbury, New Zealand. Journal of Glaciology, 6, 737-748
- McSaveney, M. J., 2002: Recent rockfalls and rock avalanches in Mount Cook National Park, New Zealand. Geological Society of America Reviews in Engineering Geology, Volume XV, 35-69
- McSaveney, E., 2007: Glaciers and glaciation - Retreating ice and the glacier legacy, Te Ara - the Encyclopedia of New Zealand, http://www.TeAra.govt.nz/en/photograph/10744/waiho-moraine-loop-franz-josef accessed 8 February 2017
- Morton, J., 2017: PHOTOS: Iconic South Island glaciers retreating, NZ Herald, http://www.newstalkzb.co.nz/photos/national/photos-iconic-south-island-glaciers-retreating/ accessed 15 March 2017
- New Zealand Geological Survey, 1973: Quaternary Geology -- South Island. 1: 1,000,000 map. New Zealand Geological Survey Miscellaneous map Series, Map 6.
- NIWA, 2017: https://www.niwa.co.nz/education-and-training/schools/resources/climate/extreme accessed 22 January 2017
- Östrem, G., 1959: Ice melting under a thin layer of moraine, and the existence of ice cores in moraine ridges. Geografiska Annaler, 41 (4), 228–230
- Peat, N., 2010: The Tasman – Biography of an ocean. Penguin Group (NZ), 239 p.
- Post, A., Mayo, L. R., 1971: Glacier dammed lakes and outburst floods in Alaska. Washington, DC: US Geological Survey, 10 p.
- Purdie, J., Fitzharris, B., 1999: Processes and rates of ice loss at the terminus of Tasman Glacier, New Zealand. Global and Planetary Change, 22, 79–91
- Purdie, H., Anderson, B., Chinn, T., Owens, I., Mackintosh, A., Lawson, W., 2014: Franz Josef and Fox Glaciers, New Zealand: Historic length records. Global and Planetary Change, 121, 41–52

- Putnam, A. E., Schaefer, J. M., Denton, G. H., Barrell, D. J. A., Finkel, R. C., Andersen, B. G., Schwarz, R., Chinn, T. J., Doughty, A. M., 2012: Regional climate control of glaciers in New Zealand and Europe during the pre-industrial Holocene. Nature Geoscience, 5, 627-630
- Putnam, A. E., Schaefer, J. M., Denton, G. H., Barrell, D. J. A., Birkel, S. D., Andersen, B. G., Kaplan, M. R., Finkel, R. C., Schwarz, R., Doughty, A. M., 2013: The Last Glacial Maximum at 44°S documented by a ^{10}Be moraine chronology at Lake Ohau, Southern Alps of New Zealand. Quaternary Science Reviews, 62, 114-141
- Quincey, D. J., Glasser, N. F., 2009: Morphological and ice-dynamical changes on the Tasman Glacier, New Zealand, 1990–2007. Global and Planetary Change, 68, 185-197
- RCEM – Reclamation Consequence Estimating Methodology: U.S. Department of the Interior Bureau of Reclamation, 2014
- Richardson, S. D., Reynolds, J. M., 2000: An overview of glacial hazards in the Himalayas. Quaternary International, 65/66, 31-47
- Ricker, K. E., Chinn, T. J., McSaveney, M. J., 1993: A late Quaternary moraine sequence dated by rock weathering rinds, Craigieburn Range, New Zealand. Canadian Journal of Earth Sciences, 30, 1861-1869
- Salinger, M. J., Heine, M. J., Burrows, C. J., 1983: Variations of the Stocking (Te Wae Wae) Glacier, Mount Cook, and climatic relationships. New Zealand Journal of Science, 26, 321-38
- Salinger, M. J., Renwick, J. A., Mullan, A. B., 2001: Interdecadal Pacific Oscillation and South Pacific Climate. International Journal of Climatology, 21 (14), 1705–1721
- Samjwal, R. B., Pradeep, K. M., Basanta, R. S., 2007: Impact of climate change on Himalayan glaciers and glacial lakes: Case studies on GLOF and associated hazards in Nepal and Bhutan, ICIMOD, 133 p.
- Scott, B. J., 2013: A revised catalogue of Ruapehu volcano eruptive activity: 1830-2012, GNS Science Report 2013/45. 113 p.
- Skinner, B. E., 1964: Measurement of Twentieth Century Ice Loss on the Tasman Glacier, New Zealand. New Zealand Journal of Geology and Geophysics, 7 (4), 796-803
- Solomina, O. N., Bradley, R. S., Hodgson, D. A., Ivy-Ochs, S., Jomelli, V., Mackintosh, A. N., Nesje, A., Owen, L. A., Wanner, H., Wiles, G. C., Young, N. E., 2015: Holocene glacier fluctuations. Quaternary Science Reviews, 111, 9-34
- Stats-NZ, 4.4.2017, http://www.stats.govt.nz/browse_for_stats/population/estimates_and_projections/NationalPopulationEstimates_HOTPAt30Jun16.aspx
- Suggate, R. P., 1990: Late Pliocene and Quaternary glaciations of New Zealand. Quaternary Science Reviews, 9, issues 2-3, 175-197.
- Tippett, M., Kamp, P., 1995: Geomorphic evolution of the Southern Alps, New Zealand. Earth Surface Processes and Landforms, 20, issue 2, 177-192.
- USGS, 6.1.2017, https://earthquake.usgs.gov/learn/topics/mag_vs_int.php
- Vandergoes, M. J., Fitzsimons, S. J., 2003: The Last Glacial–Interglacial Transition (LGIT) in south Westland, New Zealand: paleoecological insight into mid-latitude Southern Hemisphere climate change. Quaternary Science Reviews, 22, 1461-1476
- Vardion, 2008: New Zealand Population Density. Available online:

https://commons.wikimedia.org/wiki/File:NewZealandPopulationDensity.png

- Wardle, P., 1973: Variations of the glaciers of Westland National Park and the Hooker Range, New Zealand. New Zealand Journal of Botany, 11 (2), 349–388
- Warren, C. R., 1991: Terminal environment, topographic control and fluctuations of West Greenland glaciers. Boreas 20, 1–17
- Westoby, M. J., Glasser, N. F., Brasington, J., Hambrey, M. J., Quincey, D. J., Reynolds, J. M., 2014: Modelling outburst floods from moraine-dammed glacial lakes. Earth-Science Reviews, 134, 137-159
- WGMS, 2008: Global Glacier Changes: facts and Figures (United Nations Environmental Programme and WGMS)
- Williams, P. W., McGlone, M., Neil, H., Zhao, J., 2015: A review of New Zealand palaeoclimate from the Last Interglacial to the global Last Glacial Maximum. Quaternary Science Reviews, 110, 92-106