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# An Interhemispheric Comparison of the Recession of Mountain Glaciers in the Last 150 Years

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# AN INTERHEMISPHERIC COMPARISON OF THE RECESSION OF MOUNTAIN GLACIERS IN THE LAST 150 YEARS


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# AN INTRHEMISPHERIC COMPRISON OF THE RECESSION OF MOUNTAIN GLACIERS IN THE LAST 150 YEARS

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An Abstract of the Thesis Presented
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Historical records in the Northern Hemisphere show overall glacier retreat since about AD 1860-1890. A glacial retreat of similar timing and magnitude in the Southern Hemisphere is less well established. Comparison of the timing and magnitude of glacial recession in the two polar hemispheres over the past century can elucidate mechanisms driving global climatic change. In order to determine the recession patterns of Murchison, Hooker, and Tasman Glaciers in the Southern Alps of New Zealand, geomorphic maps were constructed and recent glacial deposits were dated using lichenometry. Since the mid-to-late nineteenth century, each glacier terminus has retreated about 1.5 km. The timing and magnitude of this recession of New Zealand glaciers are similar to that of the recent glacial retreat in the European Alps and elsewhere in the Northern Hemisphere. Such near-synchronous recession is consistent

with global, not merely hemispheric, warming. Mechanisms other than the Broecker ocean circulation model and the bipolar see-saw must be used to explain the synchronous retreat of glaciers in both polar hemispheres. Some of the recent warming could be linked to decreased global albedo, increased concentrations of trace gases in the atmosphere, and variations in solar activity,

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# 1. INTRODUCTION

The advance and retreat of glacial margins are used widely as proxies for climate change. Variations in climate, particularly in summer temperature (Oerlemans, 2001), affect the elevation of equilibrium line altitudes (ELA) and, accordingly, the extent of glacier termini. Since the mid-nineteenth century, most temperate alpine glaciers in the Northern Hemisphere have undergone general recession. Although not all glaciers have receded steadily, due to local climate variability, it is believed that the overall trend of recession is indicative of hemisphere-wide warming. In terms of temperature, this marks the end of what has come to be known as the Little Ice Age (LIA), a period of colder climate beginning in about AD 1280 and ending in about AD 1860 (Holzhauser and Zumbuhl, 1999).

Glacier recession is documented best in the European Alps by historical records and geomorphic evidence. Glaciers underwent overall retreat through out the Alps from the mid-nineteenth century to the present (fig. 1). Haeberli (1994) calculated a 30 to 40% loss of ice mass, a 100 m rise in equilibrium line elevation, and an annual temperature rise of 0.5 °C in the European Alps since AD 1850. For example, between AD 1850 and 1977, the termini of the Aletsch and Hufi Glaciers each have retreated more than 2.5 km (Kasser and Haeberli, 1981). The Glacier de Trient on the Montblanc Massif receded

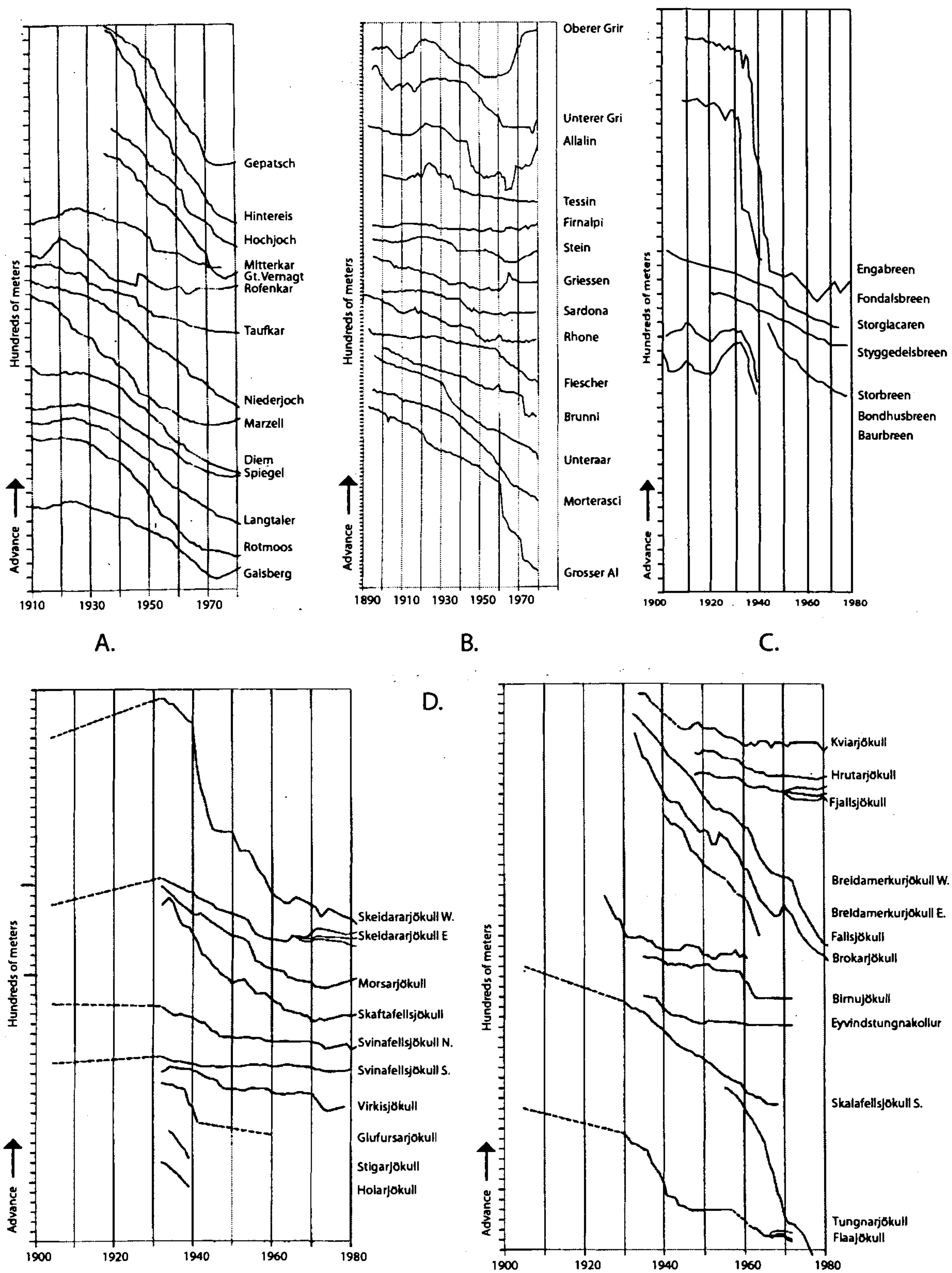


Figure 1: Recent retreat of selected glaciers. A) Austrian Alps. B) Swiss Alps. C) Scandinavia. D) Iceland Figures from Grove, 1988.

~1.4 km between AD 1845 and 1960 (Grove, 1988). The Kesselwandferner and Hintereisferner Glaciers in the Austrian Alps were joined during the LIA. Between AD 1894 and 1978 each glacier retreated ~2.2 km; the two glaciers separated in AD 1939 (Grove, 1988)..

In Scandinavia there was a corresponding glacial recession in the last century and a half (fig. 1). In Swedish Lapland, the latest LIA glacier expansion occurred in about AD 1916. Between AD 1916 and 1976, some termini have retreated as much as two kilometers (Karlén and Denton, 1976). Between AD 1874 and 1972 Tunsbergdalbreen in Southern Norway receded about two kilometers (Mottershead and White, 1972). Nigardsbreen in southern Norway has retreated nearly four kilometers between AD 1845 and 1974 (Östrem *et al.*, 1976). As another example from southern Norway, Austerdalsbreen receded 1.7 km between AD 1865 and 1958 (King, 1959); while Storbeen retreated 1.1 km between AD 1858 and 1974 (Mathews, 1974).

Glacial retreat in Iceland also coincides with warming at the end of temperature component of the LIA. Breidamerkurjökull retreated 2.3 km between AD 1894 and 1968 (Sigbjarnarson, 1970) Outlet glaciers from Vatnajökull also have retreated extensively since the end of the nineteenth century (Grove, 1988) (fig 1).

In Greenland, glacial retreat from the mid-nineteenth century to AD 1968/69 was measured by Gordon (1980). Of the nine glaciers studied, all retreated between 0.5 and 1.5 km.

In the Canadian Rockies, glaciers reached their maximum Holocene extent during two periods of LIA advance. The first period culminated about AD 1725, and the second period, generally marked by greater ice extension, culminated about AD 1850 (Luckman, 2000). Harding (1985) (cited in Luckman, 2000) calculated an areal loss of 25% of ice in the Premier Range between the LIA maximum in AD 1850 and the glacial margins of AD 1970. In the Premier Range, McCarthy and Smith (1994) found an average reduction in glacier length of 15-80% between AD 1916 and AD 1988.

In central Alaska, the Gulkana Glacier reached a LIA maximum around AD 1890 -1900. Between this time and AD 1983, the glacier has retreated eight kilometers (Péwé and Reger, 1983). On a more regional scale, snowlines were depressed between 150 and 200 m from their current elevations in the Brooks Range, Wrangell Mountains, and St. Elias Mountains, as well as in other areas in the western United States, during the LIA (Porter, 1981).

The Southern Hemisphere has fewer locations than the Northern Hemisphere in which to examine temperate alpine glaciers. However, New Zealand is one such place. Glacial valleys in the Southern Alps of New Zealand are ideally situated in the mid-latitudes of the Southern Hemisphere on nearly the opposite side of the earth from the type example of the recent glacial retreat in the European Alps (fig. 2). There is no historical record of glacier fluctuations for much this period in New Zealand. However,

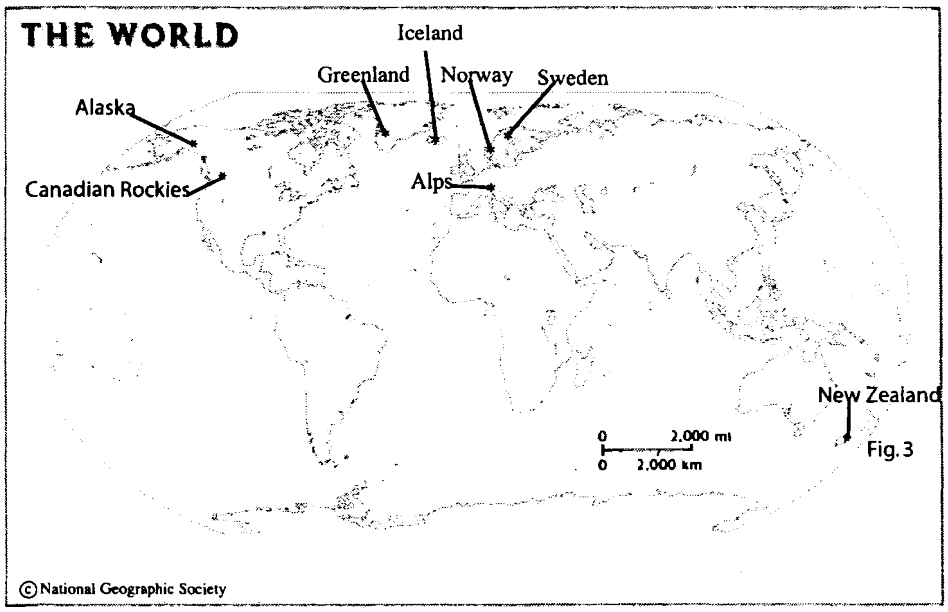


Figure 2: World map showing locations of areas mentioned in the text with well-documented glacier retreat attributed to the warming of the last 150 years in the Northern Hemisphere. New Zealand is on the opposite side of the world from the type location of the recent retreat in the European Alps.

glacier retreat in New Zealand has been noted since the late nineteenth century (Gellatly, 1985; Harper, 1935).

A comparison of the timing and magnitude of glacial retreat between the two polar hemispheres potentially can illuminate the mechanisms driving climate change over the past century and a half. The bipolar seesaw theory (Broecker, 1998) dictates that climate change should be out of phase between the polar hemispheres. In this model, the hemisphere which produces more hyper-saline bottom water in the Atlantic captures more warm tropical surface water. By this mechanism heat is pumped into the hemisphere supplying the most bottom water to the Atlantic . The hemisphere producing a lesser amount of bottom water is cooled due to the net loss of warm tropical surface water. However, a record of similar alpine glacier retreat in both polar hemispheres would imply that climate change is synchronous between the hemispheres. In such a case, heat exchange by ocean circulation would not have been the dominant driving force in the climate change of the last 150 years. Rather, a global mechanism, such as changes in planetary albedo, in atmospheric trace gas concentrations, or in solar activity may have caused the recent warming.

In order to correlate the timing and magnitude of glacial recession in the Northern and Southern Hemispheres, it is necessary to understand the magnitude and timing of the most recent phase of glacial retreat in the Southern Hemisphere. To this end, an intensive study of recent moraines was undertaken using historical records and lichenometry in the


Figure 3: Index map of the Central Southern Alps of New Zealand showing the locations of the field areas that were mapped in detail (adapted from map by Institute for Geological and Nuclear Sciences, New Zealand).

Tasman, Murchison, and Hooker Valleys on the eastern flank of the Southern Alps of New Zealand (fig. 3). The objectives of the project are threefold. First, geomorphic maps were constructed for each of the glacial valleys in order to understand the relative age relationships among glacial landforms. Second, the late Holocene marginal deposits of the glaciers were dated using the fixed area largest lichen (FALL) method of lichenometry (Bull and Brandon, 1998) applied to the growth curve of Lowell *et al.* (2003). Third, the geometry of the glaciers were calculated in order to determine the amount of glacier retreat since over the last century and a half.

# 2. GEOMORPHOLOGY OF THE FIELD AREAS

## 2.1 Glacial Geomorphology of Murchison Valley

Murchison Valley separates the Malte Brun Range on the north from the Liebig Range on the south. The valley is 1-2 km wide. It trends northeast to southwest. Murchison Glacier descends from Tasman Saddle down Murchison Valley to terminate in a position nearly opposite the mouth of the Onslow tributary valley (fig. 4, fig. 7).

Proglacial Lake Murchison fronts Murchison Glacier. Alongside the south end of the lake are raised shorelines and lacustrine deposits that merge into an area of generally hummocky terrain. This overall hummocky landform dominates the southern half of the glacial foreplain. Along the southern edge of the hummocky terrain is a small segment of outwash that grades to an end moraine with a well-preserved ice-contact slope. The outwash banked against this ice-contact slope has two terraces that descend toward the current channel of Murchison River. A relic channel cuts the terraces.

Two sets of moraines protrude from the hummocky terrain. The proximal set is composed of two main ridges and several smaller moraines. Of the two moraines, the northernmost extends into the fan/outwash from Onslow Glacier, whereas the southernmost ridge is cut by a channel that grades to one of the smaller moraines.

The ice-distal moraine set that rises from the hummocky terrain contains larger and more complex moraines than the ice-proximal moraine set. The ice-distal moraine set is bounded by the hummocky terrain on the northeast and by outwash on the southwest. The largest and most complete moraine, located at the southern end of the most ice-distal moraine set, has an erosional scarp along its contact with the outwash. This moraine features a large rotational slump scarp its ice-distal slope. The northern portion of the outermost moraine set is cut in several places by ice-contact fans and associated channels. There is also a small area of outwash that is ice-distal to the northern part of this inner moraine set. On the northern side of the valley, the outer (most ice-distal) end moraine set is continuous with the lateral/terminal moraine upon which the Onslow Hut is located (fig. 4). The Onslow Hut moraine is abutted on the ice-proximal side by the areally extensive fan/outwash deposit from Onslow Valley.

The east side of the mouth of Onslow Valley has deposits that originate from both Murchison and Onslow Glaciers. Here, two fragments of lateral moraine from Murchison Glacier are separated by a relic channel. It is possible that the westernmost moraine fragment is at least partially a result of deposition from Onslow Glacier. The primary deposits of Onslow Glacier in this area are ice-contact terraces, five of which extend various distances up Onslow Valley. In places, these terraces are cut by channels or have small moraines approximately two meters high perched on them.

North of the stream that enters Onslow Valley from the east is a set of seven small moraines, none of which is more than two meters high. Farther upvalley, above and beside the modern ice margin, are three additional small moraines and an ice-contact terrace. All are smeared onto the bedrock of the valley wall.

A lateral erosional slope/trimline, formed by the downwasting of the ice surface, extends up Murchison Valley from the east side of Onslow Valley. This feature truncates the glacial deposits of Cascade Valley (fig. 4). On the west side of Cascade Valley, the deposits consist of a series of fluvial terraces rising from the valley bottom to the base of a relatively continuous ice-proximal/erosional slope from Cascade Glacier. West of the erosional slope are three heavily vegetated lateral moraines from Cascade Glacier.

On the east side of Cascade Valley, a single fluvial terrace is located just above the valley floor. Two ice-contact terraces composed of till occur south the fluvial terrace. The upper ice-contact terrace forms the westernmost portion of a platform separating Cascade, Murchison, and Baker Valleys (fig. 4). Distal to this ice-contact terrace is a well-preserved lateral moraine associated with Cascade Glacier. This moraine grades into the continuous lateral moraine of Cascade Glacier, which has an erosional proximal slope.

The deposits that compose the Cascade/Baker platform east of the Cascade lateral moraine originate either from Murchison Glacier or from Baker Glacier. Along the south and southeast margins of the platform is a continuous lateral moraine deposited by Baker and/or Murchison Glaciers. Four small channels are cut into but not through this moraine. Distal to the continuous moraines are two lateral moraines from Baker Glacier.

North of these two moraines are two heavily vegetated lobate moraines, possibly formed at a time when Baker and Cascade Glaciers joined at this elevation.

On the east side of Baker Valley is another platform composed of deposits from Baker, Murchison, and Wheeler Glaciers (fig. 4). A continuous lateral moraine of Baker Glacier delimits the western margin of this platform. The proximal slope of the moraine is erosional, and, in places, large sections of intact moraine have slumped onto this erosional slope. Distal to this continuous lateral moraine is a discontinuous ice-contact feature of Baker Glacier. In some places, this feature is a moraine or an ice-contact terrace, and in other places it is absent.

The lateral erosional slope of Murchison Glacier flanks the southern margin of the Baker/Wheeler platform. Distal to the erosional slope is a continuous moraine crest that extends into Wheeler Valley before being truncated by the erosional slope of the Wheeler lateral moraine. On the southeast corner of the platform, two moraines extend side by side, separated only by a relic channel that cuts the more distal moraine. This ice-distal moraine continues to the west of where it is cut by the channel, but it is poorly preserved.

The Baker/Wheeler platform also has two large, grassy lateral moraines that span almost the entire length of the platform in a north-south direction. The eastern of these two moraines was deposited by Wheeler Glacier. It is unclear which glacier deposited the westernmost of these two moraines. North of these moraines are two smaller lateral moraines of Wheeler Glacier. Several slumps in this area strike approximately east-west and dip to the south. Slump scarps of up to three meters are exposed in places. These scarps crosscut the unconsolidated glacial sediments of the platform. It is believed that the slump scarps represent mass wasting of large blocks of slope material into Murchison Valley. South of the Baker/Wheeler platform, hummocky ice-contact and ice-cored terraces descend to Murchison Glacier.

Across Wheeler Stream from the Baker/Wheeler platform is yet another platform flanked by Wheeler, Murchison, and Dixon Glaciers. The lateral moraine of Wheeler Glacier delimits the western margin of this platform. The southern margin is marked by a relatively continuous lateral moraine of Murchison Glacier. The lateral moraine of Dixon Glacier flanks the eastern side of the platform. Ice-contact/erosional slopes characterize all three of these lateral moraines. Distal to the lateral moraines is an expanse of outwash, that is separated from the bedrock by two small, well-vegetated moraines. This entire platform is crosscut in several locations by faults that strike approximately east-west and dip south.

## 2.2 Glacial Geomorphology of Hooker Valley

Hooker Glacier is fed by tributaries that drain ice from the western slope of Mount Cook. Hooker Valley is 800-1000 m wide and trends nearly north to south. It separates the Main Divide on the west from the Mount Cook Range on the east (fig. 5, fig. 7).

The Hooker Glacier foreplain is separated from the ice itself by proglacial Hooker Lake. South of the lake are three sets of moraines. The most ice-proximal moraine set is distinctive due to its lack of vegetation. Just south of the lake is a small unvegetated moraine. Ice-distal to this moraine is a well-preserved ice-contact slope that abuts an area of hummocky terrain, which in turn wraps around the eastern shore of the lake. Another unvegetated moraine immediately ice-distal to the hummocky terrain marks the southern extent of the ice-proximal moraine set.

The boundary between the ice-proximal moraine set and the intermediate moraine set is marked by a pronounced increase in vegetation. The ice-proximal moraine set is covered only in wooly moss, whereas the intermediate moraine set is heavily vegetated with brush and small trees. The heavily vegetated moraine set is composed of two continuous ridges that are separated by a channel on the east end. Two smaller moraines occur on the distal slope of the southern moraine in this set. South of the intermediate moraine set there is an area of hummocky terrain cut by a large relic channel of Hooker River.

The third and most ice-distal set of end moraines is located on the opposite side of the modern-day Hooker River from the other two sets of end moraines. The long axis of the northernmost moraine of this third set is oriented northwest to southeast and grades into an ice-contact slope along its northwestern end. The end moraines south of the current river channel are heavily dissected by channels and are bounded to the south by outwash. Fragments of two lateral moraines that may be associated with the end moraines are present along the west side of the valley. These lateral moraines from Hooker Glacier are truncated by the lateral moraine from Eugenie Glacier (fig. 5).

There are two large right lateral moraines from Eugenie Glacier. Both of these moraines have been dissected by melt water at the base of their ice-proximal slopes. East of these moraines in Hooker Valley is a complex set of channels cut by melt-water from Eugenie Glacier. These channels have dissected heavily the lateral and terminal moraines of Hooker Glacier, so that only small fragments remain of the heavily vegetated Hooker moraines.

A series of small moraines and ice-contact slopes lie on the floor of Eugenie Valley and are smeared against the ice-proximal slope of the left lateral moraine of Eugenie Glacier. Distal to these small moraines are two larger, fragmented lateral moraines.

The trimline/erosional slope left by Hooker Glacier after significant recent down-wasting dominates the western side of Hooker Valley. Moraines are preserved in places west of the erosional slope. Many slump scarps exist in the unconsolidated glacial sediments of this area.

A flight of moraines and ice-contact slopes characterize the deposits of eastern Hooker Valley. In places, as many as six relatively continuous lateral moraines are on the east slope of the valley. In some places, channels or scree obscures the moraines. At the southern end of the valley are two lateral moraines.

Stocking Valley joins the southwest corner of Hooker Valley. Upper Stocking Valley features several low terraces as well as lateral moraines. These moraines are located on a bedrock knoll well below the present-day position of the glacier. However, during a time of greater-than-present glacier extent, ice passed on both sides of this knoll, depositing three moraines on the western side of the knoll and two moraines on the eastern side of the knoll (fig.5).

### 2.3 Glacial Geomorphology of Tasman Valley

Tasman Valley trends north to south and separates the Main Divide and Mount Cook Range from the Malte Brun and Liebig Ranges. Tasman Glacier, the largest in New Zealand, fills the valley floor. The glacier is one to two kilometers wide, and about 26 km long (fig. 6, fig. 7).

Tasman Glacier is bounded on its southern margin by proglacial Tasman Lake. The present-day outlet from the lake cuts through relic outwash along the southwest shore. A small rocky lateral moraine occurs just north of the current lake outlet. A channel cuts this moraine. North of and distal to this moraine is a continuous rocky lateral moraine complex that extends upvalley nearly as far as the Department of Conservation (DOC) overlook (fig. 6). The most ice-proximal ice-contact slope in this moraine complex is continuous with ice-contact/erosional slopes that extend several kilometers upvalley from the DOC overlook. Ice-contact fans and small moraines make up the distal portion of this moraine complex.

The rocky lateral moraine complex with the DOC overlook on it is abutted on its ice-distal side by a relic channel. Beyond the relic channel is pair of higher well-vegetated lateral moraines that intersect each other at nearly right angles. These two higher moraines are the most ice-distal ice-contact deposits in the immediate area. Outwash, vegetated with grass and brush, occurs southwest of these moraines.

The pair of nearly perpendicular moraines is cut along the northern edge by a relic channel that trends east-west at this location. However, at its head below an ice-contact fan the channels direction is north-south. Here, the channel extends between the most ice-proximal lateral moraine set and another lateral moraine. Three small moraines are located adjacent to the margins of the ice-contact fan at the head of the channel.

Along the active erosional slope, north of this relatively large ice-contact fan, are two small rocky lateral moraines situated alongside each other. Just north of these moraines is the DOC overlook (fig. 6). Distal to the large erosional slope are three primary low, rocky, discontinuous lateral moraines. Other smaller moraines also exist in this area. A continuous lateral moraine makes up the proximal shorelines of a series of four kettle lakes known as Blue Lakes (fig. 6). A heavily vegetated lobate moraine isolates the southernmost Blue Lake from the second Blue Lake. This same moraine divides the second lake from the third lake. A separate and more discontinuous vegetated moraine isolates the third Blue Lake from the fourth, and northernmost, Blue Lake. The Blue Lakes are bounded on the west by a discontinuous, heavily vegetated moraine.

The moraine that makes up the west shores of the Blue Lakes is banked against a massive grass-and-brush-covered moraine that is the largest ice-marginal feature on the Tasman foreplain. Two ice-contact fans occur on the distal slope of this moraine. The northernmost of the fans also has smaller associated moraines. Some smaller moraines are located southeast of the southernmost ice-contact fan. North of the largest moraine are two heavily vegetated lateral moraine fragments between the valley wall and the most ice-proximal Tasman lateral moraine.

North of the Blue Lakes area, the most ice-proximal lateral moraine is associated with two small ice-contact fans. Fragments of two heavily vegetated lateral moraines trend nearly parallel to the ice-contact/erosional slope. These moraines are cut in places by channels associated with ice-contact fans that grade to the ice-contact/erosional slope. There is a third heavily vegetated lateral moraine west of the other two. It is banked against the valley wall and nearly buried by an alluvial fan.

The most ice-proximal lateral moraine with the ice-contact/erosional slope continues from the Blue Lakes area to the north. Large portions of the crest have slumped into the valley. Slumps are common north of the Blue Lakes area. A similar ice-contact/erosional slope also exists on the eastern side of Tasman Glacier. North of the mouth of Murchison Valley, the ice-contact/erosional slope is the dominant glacial feature (fig. 6). However, the lateral moraine system of Tasman Glacier extends eastward into the mouth of Murchison Valley, forming a complex lobe of lateral moraine deposits.

The northernmost portion of the moraine lobe in Murchison Valley is cut by two large ice-contact fans that are separated from each other at their heads by two small rocky moraines. South of the fans, the large lateral moraine system is characterized by long, low, rocky moraine ridges trending parallel to the crest of the larger moraine that makes up the core of the deposit. The ice-distal slope of the core moraine is vegetated and also has long, low moraines deposited on it. The ice-proximal slope of the core moraine is a till plain with several continuous ice-contact slopes cut into it. Several small lakes lie within the boundaries of the till plain. Distal to the Tasman lateral moraine lobe is outwash from Murchison Glacier.

High and steep chevron moraines characterize the southernmost portion of the Tasman ice lobe that projects into the mouth of Murchison Valley. All of the moraines are cut by a series of west-east trending channels. South of the channels is the last of the series of chevron moraines. This is the southernmost of the smaller moraines that are superimposed over the core lateral moraine on the east side of Tasman Glacier.

South of the southernmost chevron moraine, the ice-contact/erosional slope recurs on the proximal side of the large continuous core moraine. Numerous ice-contact slopes are etched into the proximal slope of the lateral moraine. The channel of Murchison River flows between the eastern slope of the lateral moraine and the valley wall. The southernmost portion of the lateral moraine complex is hummocky without a well-defined moraine crest. However, the ice-contact/erosional slope is present. The Murchison River channel has cut the hummocky deposits after a recent rock fall dammed the previous channel. The river now empties into Tasman Lake.

South of the Murchison River inlet is a series of relic channels that formerly drained water from Tasman Glacier into Murchison River. A single piece of Tasman lateral moraine remains in the midst of the channel system. The frontal deposits of Tasman Glacier begin west of this moraine.

Two small rocky lateral moraine fragments occur beside the channel system. South of Tasman Lake three frontal moraines rise above an area of hummocky deposits. The relatively extensive hummocky deposits are bounded to the north by Tasman Lake and to the south by a continuous moraine ridge. The continuous moraine has smaller ridges just south of it on the east side of the valley. Ice-contact fan deposits spill off of the distal slope of the continuous moraine and into relic channels that cut the outwash. The continuous moraine and the hummocky deposits are truncated on the west side by the present channel system of Tasman River.

# 3. THE HISTORICAL RECORD

Unlike the situation in the European Alps, a relatively short historic record exists for glacial fluctuations in the Southern Alps of New Zealand. Prior to AD 1862 there is essentially no historical record of glacial marginal positions, as only a few herders and explorers had ever seen the glaciers. In 1862, Julius von Haast, a government surveyor, made the first map of Tasman Valley (Gellatly, 1985). From this time until the present, a fairly continuous historical record is preserved in maps, photographgraphs, and written accounts by early travelers. However, the material commonly is unpublished and held in special collections. For this reason, secondhand accounts are used where primary documents are not available.

## 3.1 The Historical Record of Murchison Valley

In AD 1890, G.E. Mannering and A.P. Harper were members of the first European party to venture into Murchison Valley. Mannering documented the journey in a series of reports. Although he makes little reference to the position of Murchison Glacier, he does mention the extended positions of tributary glaciers. Mannering points out that Onslow and Cascade Glaciers both joined Murchison Glacier at that time (Mannering, 1890) (fig. 4).

The geomorphology of the Murchison Glacier terminus is such that if Onslow and Murchison Glaciers were joined, then Murchison ice would have abutted the lateral moraine that now has the Onslow Hut on it. This lateral moraine traces to the outer loop of end moraines that cross the valley. A large-scale sketch map also indicates the viability of this connection (Mannering, 1890).

In AD 1891, T.W. Brodrick, a government surveyor, visited Murchison Valley in search of grazing land. In his report of the survey, he mentions little about the terminal position of the Murchison, but he does mention the tributary glaciers. He writes, "The Onslow, nearest to the terminal face of the Murchison (Glacier) is apparently stationary at present. The Cascade and Wheeler contribute nothing to the general stream and have the appearance of dying out" (Brodrick, 1891, p.3). Brodrick's account, taken in conjunction with that of Mannering, indicates that the Onslow, Cascade, and Wheeler tributary glaciers still joined the Murchison despite the fact that they were in a state of retreat by AD 1891.

The data for the Mt. Cook Alpine Regions map are believed to have been collected about AD 1934 (Crawford, 1953). This map indicates that Onslow and Cascade Glaciers had separated from the Murchison, but that the Wheeler was still joined at that time. However, an aerial photographgraph from AD 1965 (New Zealand Aerial Mapping, survey number 1580, photograph number 3725/33) shows that all three tributary glaciers had separated from the Murchison. The Onslow and Cascade margins had receded up the lower part of their respective valleys to locations above the previous icefalls. The Wheeler terminus was approximately half way up the lower portion of its valley. The terminus of the Murchison appears to have been about even with the western edge of Onslow Valley.

Between AD 1965 and 1986, when the next aerial photographgraph was taken (New Zealand Aerial Mapping survey number 8595, photograph number B/13), there is little difference in the marginal position of Murchison Glacier.

## 3.2 The Historical Record of Hooker Valley

H.G. Wright made a sketch map of Hooker Valley in AD 1884. The map is of relatively low quality, but it shows the terminus of Hooker Glacier approximately 100 m inside the most ice-proximal vegetated moraine (Gellatly, 1985).

A photographgraph from AD 1894, held by the Hocken Library in Dunedin, shows Stocking Glacier in an advanced position. The terminus of Stocking Glacier rests against the south side of the east-west trending bedrock knoll in Stocking Valley. However, the ice does not spill over the knoll into the north side of the valley.

A photographgraph taken by F.G. Radcliffe about AD 1910 shows significant downwasting of Hooker Glacier near the end of the nineteenth century (Black, 2001). Although little retreat had taken place since the deposition of the outermost lightly vegetated moraine, the ice surface had lowered to expose the ice-contact slope that flanks the area of hummocky deposits on the ice-proximal side.

Harrington (1952, p. 141) visited Stocking Glacier and compared its margins at that time to those in a photographgraph taken in AD 1913. He concluded that "no significant differences" can be seen in the location of the terminus of Stocking Glacier between AD 1913 and 1951. Another early twentieth century source, the Mt. Cook Alpine Regions map of AD 1934, shows that the terminus of Hooker Glacier extended down valley to a position beyond where Eugenie Stream enters the Hooker River (Crawford, 1953).

Compared to some Northern Hemisphere glaciers, the terminus of Hooker Glacier retreated very little between the late nineteenth century and AD 1965 when an aerial photographgraph was taken (New Zealand Aerial Mapping, survey number 1580, photograph number, 3724/38) showing the terminus of the Hooker Glacier at approximately the southern end of the current lake. However, there was significant downwasting of the ice surface. By the time that the AD 1986 aerial photographgraph (New Zealand Aerial Mapping, survey number 8595, photograph number C/15) was taken the main ice front had retreated approximately 800 m from the most distal lightly vegetated moraine. Hooker Lake had begun to form in the area of recent glacial retreat.

### 3.3 The Historical Record of Tasman Valley

When von Haast first came to the Tasman Valley in AD 1863, he made a plane-table survey of the glacier terminus (Gellatly, 1985). However, due to the large scale of the map, there are few reference points from which to determine the relative location of the glacier terminus. In his historical account of the survey, von Haast "discovered lying amongst the western lateral moraine deep below us, three very pretty lagoons, their water of an intense blue colour" (von Haast, p. 31, 1879). It is important to note that he saw the Blue Lakes while on Tasman Glacier. This means that in AD 1863 the Tasman had not begun to downwaste from its nineteenth century maximum height.

In AD 1884, Dr. R. von Lendenfeld (1884) visited and surveyed Tasman Glacier. On his map, the Tasman Glacier terminus is banked against the outer end moraine distal to the hummocky deposits.

Between AD 1884 (Lendenfeld, 1884) and an AD 1965 aerial photographgraph (New Zealand Aerial Mapping, survey number 1580, photograph number 3724/38), the terminus of Tasman Glacier retreated approximately 230 m to about the proximal side of the hummocky deposits. In contrast to the slow terminal retreat, there was significant downwasting of the ice surface. In AD 1890, when Harper explored the Murchison Valley with Mannering, he reported that Tasman ice was at the top of the lateral moraine that extends into Murchison Valley (Harper, 1935). Between 1890 and 1936 Harper estimated that the surface of Tasman Glacier dropped 70 ft (21.5 m).

An aerial photographgraph from AD 1986 (New Zealand Aerial Mapping, survey number 8585, photograph number C/17) shows essentially no retreat of the Tasman terminus since the AD 1965 aerial photographgraph (New Zealand Aerial Mapping, survey number 1580, photograph number 3724/38). However, pro/englacial lakes were growing near the ice margin, and continued downwasting of the ice surface is evident.

# 4. PREVIOUS WORK

Considerable work has been done to date glacial deposits in the Mount Cook region. Investigators have used various dating methods, both relative and absolute, ranging from geomorphic relationships to radiocarbon measurements.

McGregor (1967) subdivided glacial deposits in the Ben Ohau Formation, named for the Ben Ohau Range of the Southern Alps (fig. 3). In the high, steep valleys of the Ben Ohau Range, small changes in equilibrium-line elevation create large fluctuations in glacial margins relative to the size of the valley. From oldest to youngest, McGregor (1967) subdivided the Ben Ohau Formation into the Ferintosh, the Jacks Stream, and the Dun Fiunary Members based on geomorphic evidence. The proposed members were not assigned absolute ages.

Birkeland (1982) used relative dating techniques on surface clasts on moraines and rock glacier deposits to differentiate among the Ferintosh, the Jacks Stream, the Whale Stream (new), and the Dun Fiunary Members of the Ben Ohau Formation. These techniques included lichenometry, rind thickness, quartz vein height, surface oxidation (color), and percent of clasts with surface pitting. The rind thickness data were used in conjunction with the rock-weathering rind development curve of Chinn (1981) to assign approximate ages to each member. By this procedure, the Ferintosh Member was given an age of ~4000 BP, the Jacks Stream Member ~3000 BP, the Whale Stream Member ~250 BP, the Dun Fiunary Member ~100 BP.

Burrows (1973) conducted a lichenometric study in Mueller, Hooker, and Tasman Valleys east of Mt. Cook. No glacial deposits were assigned ages older than the mid-twelfth century (Burrows, 1973). However, Burrows (1980) later rescinded his previous conclusions and stated that he had underestimated the ages of some moraines

Gellatly (1984) used the rock-weathering rind development curve of Chinn (1981) to date Holocene moraines in Mueller, Hooker, Tasman, Classen, and Godley Valleys. Near the Main Divide of the Southern Alps, 15 periods of glacial advance were identified over the past 7200 years. In a separate study, Gellatly *et al.* (1985) obtained radiocarbon dates from organic material extracted from lateral moraines of Tasman Glacier. These data indicate 11 periods of glacial advance over the past 5000 years.

Recently, geomorphic interpretations made by Gellatly (1984) have come into question. After careful geomorphic mapping, Schoenenberger (2001) reinterpreted sections of the west lateral moraine of Classen Glacier to have been overrun by ice during the past 150 years. Gellatly (1984) indicated that the outermost west lateral moraine was hundreds of years older than the inner moraines that Schoenenberger (2001) interpreted to be overrun. On the west side of Hooker Valley, Gellatly (1984) identified seven lateral moraines. The moraine most proximal to Hooker Glacier yielded a rind date of <100 YBP; the most distal moraine yielded a rind date of 2160 YBP. Black (2001) mapped this area as a fan deposit from Eugenie Glacier and made no indication of multiple ages. In addition to disputing the geomorphic interpretations of Gellatly (1984), these findings also brings into question the accuracy of at least parts of the rock-weathering rind curve (Chinn, 1981). Black (2001) undertook an intensive lichenometric study in Muller and Hooker Valleys, and, by this technique, found no glacial deposits older than AD 1637 on the Hooker glacial foreplain.

Although periodic glacial advances are thought to have taken place throughout the Holocene in the small, high, valleys of the Ben Ohau Range, much debate remains about the ages of glacial deposits in the low, large, valleys east of Mt. Cook. This study seeks to differentiate between the glacial deposits of the last 200 years and earlier geomorphic features. Once the maximum extent of the glaciers in the last 200 years has been determined, the timing of retreat can be reconstructed.

# 5. METHODS

In order to calculate the amount of glacial retreat over the last couple centuries, it is obviously necessary to determine the position of glacier margins during this time period. Geomorphic mapping of Murchison, Hooker, and Tasman Valleys identified past ice-marginal glacial deposits. Lichenometry was used to determine their absolute ages.

Lichenometry is the measuring of lichen diameters to determine the time at which lichen colonized the substrate on which they are growing. This gives a close minimum age of the landform. Relative ages can be obtained by comparing lichen diameters from different geomorphic surfaces, or absolute minimum ages can be obtained after a lichen growth-rate curve is established.

Lichen data were collected using a modified fixed-area largest-lichen (FALL) method (Bull and Brandon, 1998). The longest axis of the largest Rhizocarpon lichen was measured on the surface of a boulder (our fixed area) using Mitutoyo digital calipers. This process was continued along a transect across a geomorphic surface until 100 measurements were made. The large sample size averages out irregularities in the data related to microclimate, colonization time, inherited lichens, and merged lichens (Bull and Brandon, 1998). Some data sets were collected using an alternate method in which the five or ten largest lichens on a landform were sought. These data sets were few and they were not used to reconstruct past glacial margins.

In addition to measurements of lichen diameters, accompanying data were collected to mimic better the technique of Bull and Brandon (1998). Before being measured, each lichen was assigned a qualitative rating, between one and five, of the overall quality of the lichen. High quality lichens (5's) were nearly round, were isolated from other lichens on the rock, and had a continuous black rim. Lichen diameters included the black rim. Lichens with quality ratings of 4, 3, or 2 exhibited the above characteristics to sequentially lesser degrees. Lichens with a quality rating of 1 were lacking in the above characteristics to such an extent that they were not measured. The aspect (north, south, east, west, or top) which each lichen faced was recorded. Finally, the name of the person who made each measurement also was noted.

Lichenometry does not necessarily date the time at which a geomorphic surface forms. Rather, it dates the time when the surface was stable enough to support lichen colonization. In choosing a sample site, care was taken to collect data on a single geomorphic surface that stabilized over a short period of time following the removal of ice or water from the substrate. Such geomorphic surfaces include moraine ridges, relic channels, and ice-contact fans.

Bull and Brandon (1998) used lichenometry to date rock fall deposits believed to be the result of earthquakes. They sought multiple ages within a single data set. However, each of our data sets was collected on single-aged landforms. Therefore, different methods of statistical analysis were used here as compared to those of Bull and Brandon (1998).

Lowell *et al.* (2003) investigated various means of analyzing lichenometric data. Five growth curves were established using a geometric regression line fitted to measurements of 1) the largest lichen in a data set, 2) the average of the five largest lichens in a data set, 3) the average of the ten largest lichens in a data set, 4) the average of all lichens in a data set, and 5) the 98th percentile of lichens in a data set. Lowell *et al.* (2003) found that the 98th percentile curve is a reliable and statistically robust method of data analysis. The 98th percentile growth curve of Lowell *et al.* (2003) is employed here (fig. 8). Lowell *et al.* (2003) did not account for aspect and quality of lichens in the creation of their growth-rate curve. Therefore, in order to simulate the method of Lowell *et al.* (2003) aspect and quality data were not taken into account in this study. All lichen measurements, regardless of quality, were used to determine the age of landforms.

The 98th percentile of each lichen data set is plotted on the Lowell *et al.* (2003) regression line $y = -1.0012515\ x^2 + 0.7838\ x - 31.0853$, where y = the 98th percentile from each data set, and x = years before present (YBP). The regression line does not work for data sets with a 98th percentile diameter higher than 91.64 mm. This correlates to more than 313 YBP. Surfaces with lichenometric ages older than 313 YBP were considered too old to date with lichenometry.

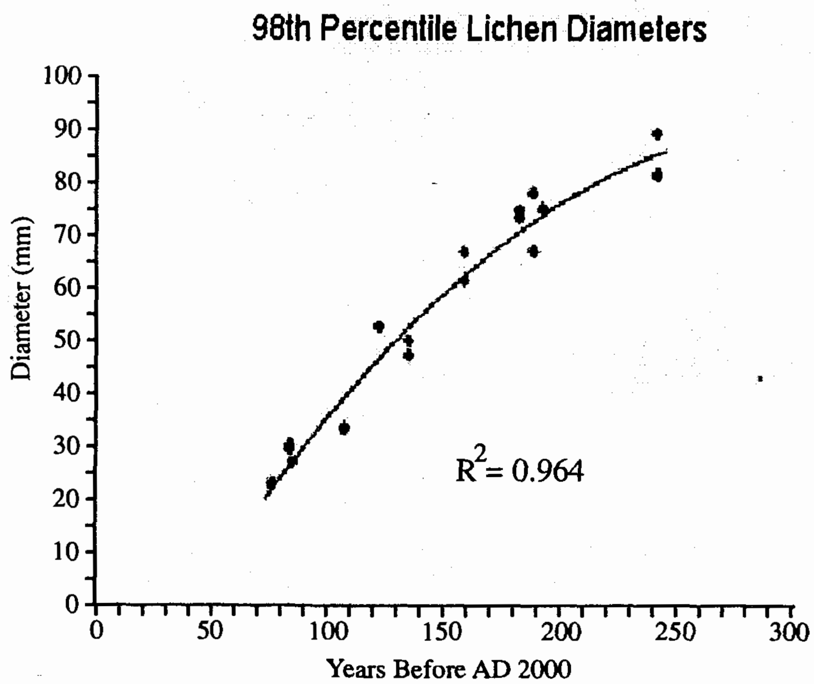


Figure 8: The lichen growth curve constructed by Lowell *et al.*, (2003) using the fixed area largest lichen method in the Southern Alps.

Additional statistical analysis included z-tests to determine if there was a significant difference between the means of measurements taken by each operator on a single landform. Z-tests were conducted on data from five randomly selected landforms.

The geomorphic mapping in the valleys created a relative chronology of the glacial deposits. This relative chronology was used to check the accuracy of the absolute lichenometric ages. Also, the geomorphic relationships determined during mapping allowed us to determine partially a relative chronology of landforms in a valley. This enabled us to reconstruct better the past glacier margins.

# 6. RESULTS

Lichenometric results presented in Tables 1 through 3 are interpreted best with the aid of the geomorphic maps (figs. 4-6). Most of the data is internally consistent. However, there are instances where multiple lichen counts on a single geomorphic surface predict different ages. Also, lichenometric ages do not always correlate with the geomorphic sequence of landforms. Years before present is years before AD 2000. Ages listed as "unknown" are the result of the 98th percentile lichen diameter being too large for the growth curve of Lowell *et al.* (2003) (fig. 8).

It was hypothesized that measurements taken by different operators are statistically the same. A z-test was used to test this hypothesis at the 95% confidence level. The largest value that allows for acceptance of this hypothesis is z = 1.96. Of the five data sets, sampled at random, the highest z value is 1.09 (table 4, fig. 4-6). Therefore, different people measuring lichen diameters did not introduce error into the determination of lichenometric ages.

Table 1: Murchison Lichenometric Data. See fig. 4 for site locations.

| Lichen Count | N | N(98) | 98th Percentile | Years BP | Calendar years | Useful for reconstructing past ice margins? |
|---|---|---|---|---|---|---|
| MR 1 | 50 | 49.00 | 20.935 | 75 | 1925 | no; unstable surface |
| MR 2 | 100 | 98.00 | 25.35 | 83 | 1917 | no; unstable surface |
| MR 3 | 50 | 49.00 | 31.52 | 94 | 1906 | yes |
| MR 4 | 100 | 98.00 | 44.51 | 119 | 1881 | yes |
| MR 5 | 50 | 49.00 | 38.955 | 108 | 1892 | yes |
| MR 6 | 100 | 98.00 | 36.49 | 103 | 1897 | yes |
| MR 7 | 100 | 98.00 | 27.115 | 86 | 1914 | yes |
| MR 8 | 100 | 98.00 | 48.105 | 127 | 1873 | yes |
| MR 9 | 100 | 98.00 | 50.76 | 132 | 1868 | yes |
| MR 10 | 100 | 98.00 | 50.06 | 131 | 1869 | yes |
| MR 11 | 100 | 98.00 | 42.33 | 115 | 1885 | yes |
| MR 12 | 100 | 98.00 | 39.44 | 109 | 1891 | yes |
| MR 13 | 100 | 98.00 | 28.62 | 89 | 1911 | yes |
| MR 14 | 101 | 98.98 | 27.41 | 87 | 1913 | yes |
| MR 15a | 100 | 98.00 | 12.705 | 62 | 1938 | no; located on present ice |
| MR 15b | 100 | 98.00 | 16.555 | 68 | 1932 | no; located on present ice |
| MR 16 | 100 | 98.00 | 28.37 | 88 | 1912 | yes |
| MR 17 | 50 | 49.00 | 19.765 | 74 | 1926 | yes |
| MR 18 | 100 | 98.00 | 33.895 | 98 | 1902 | yes |
| MR 201 | 100 | 98.00 | 36.085 | 102 | 1898 | yes |
| MR 202 | 100 | 98.00 | 30.44 | 92 | 1908 | yes |
| MR 203 | 100 | 98.00 | 22.67 | 78 | 1922 | yes |
| MR 204 | 100 | 98.00 | 28.1 | 88 | 1912 | yes |
| MR 205 | 100 | 98.00 | 32.355 | 96 | 1904 | yes |
| MR 206 | 101 | 98.98 | 26.71 | 85 | 1915 | yes |
| MR 207 | 100 | 98.00 | 29.42 | 90 | 1910 | yes |
| MR 208 | 101 | 98.98 | 37.1 | 104 | 1896 | yes |
| MR 209 | 105 | 102.90 | 25.26 | 83 | 1917 | no; significantly older date exists |
| MR 210 | 100 | 98.00 | 36.07 | 102 | 1898 | yes |

Table 1: Continued

| | | | | | | |
|---|---|---|---|---|---|---|
| MR 211 | 100 | 98.00 | 37.73 | 106 | 1894 | yes |
| MR 212 | 100 | 98.00 | 32.695 | 96 | 1904 | yes |
| MR 213 | 99 | 97.02 | 70.81 | 184 | 1816 | yes |
| MR 214 | 100 | 98.00 | 42.05 | 114 | 1886 | yes |
| MR 215 | 100 | 98.00 | 26.61 | 85 | 1915 | yes |
| MR 216 | 100 | 98.00 | 32.305 | 95 | 1905 | yes |
| MR 217 | 101 | 98.98 | 39.12 | 108 | 1892 | yes |
| MR 218 | 75 | 73.50 | 49.32 | 129 | 1871 | yes |
| MR221 | 100 | 98.00 | 44.475 | 119 | 1881 | yes |
| MR 222 | 100 | 98.00 | 39.375 | 109 | 1891 | yes |
| MR 224 | 100 | 98.00 | 58.68 | 151 | 1849 | no; unstable surface |
| MR226 | 100 | 98.00 | 28.735 | 89 | 1911 | no; significantly older date exists |
| MR228 | 100 | 98.00 | 44.82 | 120 | 1880 | yes |
| MR 229 | 100 | 98.00 | 42.07 | 114 | 1886 | yes |
| MR 230 | 104 | 101.92 | 39.56 | 109 | 1891 | yes |
| MR 231 | 101 | 98.98 | 34.56 | 100 | 1900 | yes |
| MR 232 | 100 | 98.00 | 48.955 | 128 | 1872 | yes |
| MR 233 | 100 | 98.00 | 41.56 | 113 | 1887 | yes |
| MR 235 | 101 | 98.98 | 50.33 | 131 | 1869 | yes |
| MR 236 | 100 | 98.00 | 47.845 | 126 | 1874 | yes |
| MR 237 | 100 | 98.00 | 27.17 | 86 | 1914 | yes |
| MR 238 | 100 | 98.00 | 26.1 | 84 | 1916 | yes |
| MR 240 | 100 | 98.00 | 37.94 | 106 | 1894 | yes |
| MR 241 | 101 | 98.98 | 45.17 | 120 | 1880 | yes |
| MR 244 | 100 | 98.00 | 30.345 | 92 | 1908 | yes |
| MR 245 | 100 | 98.00 | 25.97 | 84 | 1916 | yes |

Table 2: Hooker Lichenometric Data. See fig. 5 for site locations.

| Lichen Count | N | N(98) | 98th Percentile | Years BP | Calendar years | Useful for reconstructing past ice margins? |
|---|---|---|---|---|---|---|
| H 1 | 42 | 41.16 | 65.94 | 170 | 1830 | no; lichen mats on substrait |
| H 4 | 100 | 98.00 | 32.35 | 95 | 1905 | yes |
| H 5 | 100 | 98.00 | 46.75 | 124 | 1876 | yes |
| H 6 | 100 | 98.00 | 115.95 | Unknown | Unknown | no; lichen mats on substrait |
| H 7 | 61 | 59.78 | 108.6 | Unknown | Unknown | no; lichen mats on substrait |
| H 8 | 100 | 98.00 | 115.585 | Unknown | Unknown | no; lichen mats on substrait |
| H 9 | 99 | 97.02 | 127.88 | Unknown | Unknown | no; lichen mats on substrait |
| H 10 | 100 | 98.00 | 93.1 | Unknown | Unknown | no; lichen mats on substrait |
| H 11 | 100 | 98.00 | 75.135 | 198 | 1802 | no; lichen mats on substrait |
| H 12 | 100 | 98.00 | 94.69 | Unknown | Unknown | no; lichen mats on substrait |
| H 13 | 100 | 98.00 | 95.275 | Unknown | Unknown | no; lichen mats on substrait |
| H 14 | 100 | 98.00 | 29.115 | 90 | 1910 | no; lichen mats on substrait |
| H 15 | 100 | 98.00 | 43.265 | 117 | 1883 | yes |
| H 16 | 100 | 98.00 | 51.485 | 134 | 1866 | yes |
| H 17 | 100 | 98.00 | 44.1 | 118 | 1882 | yes |
| H 18 | 100 | 98.00 | 42.735 | 115 | 1885 | yes |
| H 19 | 101 | 98.98 | 51.98 | 135 | 1865 | no; lichen mats on substrait |
| H 20 | 100 | 98.00 | 78.555 | 211 | 1789 | no; lichen mats on substrait |
| H 21 | 100 | 98.00 | 29.94 | 91 | 1909 | no; lichen mats on substrait |
| H 22 | 100 | 98.00 | 65.51 | 169 | 1831 | no; lichen mats on substrait |
| H 23 | 100 | 98.00 | 64.41 | 166 | 1834 | no; lichen mats on substrait |
| H 24 | 65 | 63.70 | 92.24 | Unknown | Unknown | no; lichen mats on substrait |
| H 25 | 100 | 98.00 | 38.355 | 107 | 1893 | yes |
| H 26 | 75 | 73.50 | 40.77 | 112 | 1888 | no; signigicantly older date exists |
| H 27 | 103 | 100.94 | 74.14 | 195 | 1805 | no; lichen mats on substrait |
| H 28 | 87 | 85.26 | 68.42 | 177 | 1823 | no; lichen mats on substrait |
| H 29 | 100 | 98.00 | 59.83 | 154 | 1846 | no; lichen mats on substrait |
| H 30 | 100 | 98.00 | 105.325 | Unknown | Unknown | no; lichen mats on substrait |
| H 31 | 100 | 98.00 | 147.11 | Unknown | Unknown | no; lichen mats on substrait |

Table 2: Continued

| | | | | | | |
|---|---|---|---|---|---|---|
| H 301 | 100 | 98.00 | 33.985 | 99 | 1901 | yes |
| H 302 | 51 | 49.98 | 17.44 | 70 | 1930 | yes |
| H 303 | 100 | 98.00 | 38.79 | 108 | 1892 | yes |
| H 304 | 100 | 98.00 | 49.465 | 130 | 1870 | yes |
| H 307 | 100 | 98.00 | 79.165 | 213 | 1787 | yes |
| H 310 | 100 | 98.00 | 78.005 | 209 | 1791 | no; unstable surface |
| H 313 | 100 | 98.00 | 55.795 | 144 | 1856 | no; unstable surface |
| H 314 | 101 | 98.98 | 36 | 102 | 1898 | yes |
| H 316 | 100 | 98.00 | 79.51 | 215 | 1785 | no; lichen mats on substrait |
| H 320 | 100 | 98.00 | 60.115 | 154 | 1846 | yes |
| H 321 | 99 | 97.02 | 103.59 | Unknown | Unknown | no; lichen mats on substrait |
| H 323 | 100 | 98.00 | 71.06 | 185 | 1815 | yes |
| H 324 | 100 | 98.00 | 44.69 | 119 | 1881 | no; significantly older date exists |
| H 325 | 100 | 98.00 | 38.96 | 108 | 1892 | yes |
| H 327 | 100 | 98.00 | 73.885 | 194 | 1806 | yes |
| H 329 | 14 | 13.72 | 68.98 | 179 | 1821 | no; different sampling method was used |
| H 330 | 10 | 9.80 | 122.45 | Unknown | Unknown | no; different sampling method was used |
| H 331 | 13 | 12.74 | 42.74 | 115 | 1885 | no; different sampling method was used |
| H 332 | 12 | 11.76 | 24.85 | 82 | 1918 | no; different sampling method was used |
| H 333 | 5 | 4.90 | 129.46 | Unknown | Unknown | no; different sampling method was used |
| H 338 | 101 | 98.98 | 68.65 | 178 | 1822 | no; lichen mats on substrait |
| H 339 | 100 | 98.00 | 79.065 | 213 | 1787 | no; lichen mats on substrait |
| H 341 | 100 | 98.00 | 52.605 | 137 | 1863 | yes |
| H 342 | 99 | 97.02 | 34.93 | 100 | 1900 | yes |
| H 343 | 15 | 14.70 | 177.65 | Unknown | Unknown | no; different sampling method was used |

Table 3: Tasman Lichenometric Data. See fig. 6 for site locations.

| Lichen Count | N | N(98) | 98th Percentile | Years BP | Calendar years | Useful for reconstructing past ice margins? |
|---|---|---|---|---|---|---|
| T 1 | 50 | 49.00 | 15.85 | 67 | 1933 | |
| T 2 | 100 | 98.00 | 24.65 | 82 | 1918 | |
| T 10 | 100 | 98.00 | 36.07 | 102 | 1898 | yes |
| T 11 | 100 | 98.00 | 41.92 | 114 | 1886 | yes |
| T 12 | 104 | 101.92 | 26.47 | 85 | 1915 | |
| T 13 | 104 | 101.92 | 22.79 | 79 | 1921 | yes |
| T 14 | 100 | 98.00 | 46.66 | 124 | 1876 | |
| T 15 | 100 | 98.00 | 33.39 | 97 | 1903 | yes |
| T 16 | 100 | 98.00 | 32.06 | 95 | 1905 | yes |
| T 17a | 100 | 98.00 | 97.69 | Unknown | Unknown | no; lichen mats on substrait |
| T 17b | 100 | 98.00 | 91.54 | 304 | 1696 | no; lichen mats on substrait |
| T 18 | 100 | 98.00 | 33.36 | 97 | 1903 | yes |
| T 19 | 100 | 98.00 | 46.29 | 123 | 1877 | yes |
| T 20 | 100 | 98.00 | 63.52 | 163 | 1837 | no; significantly older date exists |
| T 21a | 100 | 98.00 | 65.51 | 169 | 1831 | no; lichen mats on substrait |
| T 22 | 100 | 98.00 | 29.93 | 91 | 1909 | no; lichen mats on substrait |
| T 301 | 100 | 98.00 | 37.03 | 104 | 1896 | no; unstable surface |
| T 302 | 101 | 98.98 | 27.72 | 87 | 1913 | yes |
| T 303 | 107 | 104.86 | 31.08 | 93 | 1907 | yes |
| T 304 | 104 | 101.92 | 36.60 | 103 | 1897 | yes |
| T 305 | 107 | 104.86 | 47.35 | 125 | 1875 | yes |
| T 306 | 103 | 100.94 | 61.26 | 157 | 1843 | no; significantly older date exists |
| T 307 | 100 | 98.00 | 112.24 | Unknown | Unknown | no; lichen mats on substrait |
| T 307A | 35 | 33.25 | 43.46 | 117 | 1883 | no; unstable surface |
| T 308 | 100 | 98.00 | 74.99 | 198 | 1802 | yes |
| T 309 | 111 | 108.78 | 31.06 | 93 | 1907 | no; unstable surface |
| T 310 | 97 | 95.06 | 202.29 | Unknown | Unknown | no; lichen mats on substrait |
| T 311 | 100 | 98.00 | 3.69 | 48 | 1952 | yes |
| T 312 | 100 | 98.00 | 4.38 | 49 | 1951 | yes |

Table 3: Continued

| | | | | | | |
|---|---|---|---|---|---|---|
| T 313 | 100 | 98.00 | 3.99 | 49 | 1951 | yes |
| T 314 | 99 | 97.02 | 3.46 | 48 | 1952 | yes |
| T 315 | 100 | 98.00 | 67.85 | 175 | 1825 | yes |
| T 316 | 100 | 98.00 | 55.75 | 144 | 1856 | no; significantly older date exists |
| T 317 | 100 | 98.00 | 31.22 | 93 | 1907 | yes |
| T 319 | 100 | 98.00 | 31.52 | 94 | 1906 | yes |
| T 320 | 100 | 98.00 | 81.12 | 221 | 1779 | yes |
| T 321 | 100 | 98.00 | 70.50 | 183 | 1817 | yes |
| T 322 | 99 | 97.02 | 55.71 | 144 | 1856 | yes |
| T 323 | 100 | 98.00 | 57.76 | 149 | 1851 | yes |
| T 324 | 100 | 98.00 | 35.01 | 100 | 1900 | yes |
| T 325 | 100 | 98.00 | 36.17 | 103 | 1897 | yes |
| T 326 | 101 | 98.98 | 38.67 | 107 | 1893 | yes |
| T 327 | 100 | 98.00 | 48.00 | 126 | 1874 | no; significantly older date exists |
| T 328 | 98 | 96.04 | 54.60 | 141 | 1859 | no; significantly older date exists |
| T 329 | 100 | 98.00 | 30.62 | 92 | 1908 | yes |
| T 330 | 100 | 98.00 | 59.44 | 153 | 1847 | yes |
| T 331 | 100 | 98.00 | 33.60 | 98 | 1902 | yes |
| T 332 | 99 | 97.02 | 31.85 | 95 | 1905 | yes |
| T 333 | 100 | 98.00 | 27.29 | 86 | 1914 | yes |
| T 334 | 50 | 49.00 | 22.87 | 79 | 1921 | no; unstable surface |
| T 335 | 100 | 98.00 | 37.12 | 104 | 1896 | no; significantly older date exists |
| T 336 | 100 | 98.00 | 141.41 | Unknown | Unknown | no; lichen mats on substrait |
| T 338 | 100 | 98.00 | 44.41 | 119 | 1881 | no; significantly older date exists |
| T 339 | 100 | 98.00 | 46.76 | 124 | 1876 | yes |
| T 340 | 50 | 49.00 | 23.47 | 80 | 1920 | yes |
| T 341 | 100 | 98.00 | 41.27 | 113 | 1887 | yes |
| T 342 | 75 | 73.50 | 67.98 | 176 | 1824 | yes |
| T 343 | 100 | 98.00 | 34.41 | 99 | 1901 | yes |
| T 344 | 101 | 98.98 | 46.64 | 124 | 1876 | no; significantly older date exists |

Table 3: Continued

| | | | | | | |
|---|---|---|---|---|---|---|
| T 345 | 100 | 98.00 | 39.29 | 109 | 1891 | yes |
| T 346 | 100 | 98.00 | 38.85 | 108 | 1892 | yes |
| T 347 | 100 | 98.00 | 37.07 | 104 | 1896 | yes |
| T 348 | 100 | 98.00 | 23.27 | 79 | 1921 | yes |
| T 349 | 100 | 98.00 | 25.56 | 83 | 1917 | yes |
| T 356 | 9 | 8.82 | 113.10 | Unknown | Unknown | no; different dat colletion method |
| T 357 | 12 | 11.76 | 151.78 | Unknown | Unknown | no; different dat colletion method |
| T 360 | 100 | 98.00 | 133.00 | Unknown | Unknown | yes |

Table 4: z-test Statistical Data of Selected Data Sets.
See fig. 4-6 for site locations.

| | MR 202 | MR 202 C | MR 202 J | |
|---|---|---|---|---|
| mean | 17.2454 | 17.81265306 | 16.70039216 | |
| stdev | 5.082422953 | 5.261718974 | 4.893756414 | |
| variance | 25.83102307 | 27.68568656 | 23.94885184 | |
| z(C-J) | | | | 1.093504497 |
| | | | | |
| | MR 206 | MR 206 C | MR 206 J | |
| mean | 14.79831683 | 15.24196429 | 14.24622222 | |
| stdev | 5.17599982 | 5.64971131 | 4.519656206 | |
| variance | 26.79097414 | 31.91923789 | 20.42729222 | |
| z(C-J) | | | | 0.984039478 |
| | | | | |
| | H 341 | H 341 C | H 341 J | |
| mean | 24.1462 | 24.43982759 | 23.74071429 | |
| stdev | 12.88513252 | 13.34038079 | 12.37688924 | |
| variance | 166.02664 | 177.9657596 | 153.1873873 | |
| z(C-J) | | | | 0.269775259 |
| | | | | |
| | T 319 | T319 C | T 319 J | |
| mean | 17.5713 | 17.84756098 | 17.37932203 | |
| stdev | 5.477597655 | 4.320436194 | 6.184593167 | |
| variance | 30.00407607 | 18.6661689 | 38.24919264 | |
| z(C-J) | | | | 0.445726858 |
| | | | | |
| | T 327 | T 327 C | T 327 J | |
| mean | 26.2109 | 26.11622222 | 26.28836364 | |
| stdev | 10.06763683 | 10.32020862 | 9.951126265 | |
| variance | 101.3573113 | 106.5067059 | 99.02491394 | |
| z(C-J) | | | | -0.084325634 |

# 7. DISCUSSION

There are two primary reasons that a lichenometric date may underestimate the actual age of a geomorphic surface. One is the existence of second-generation lichens. The other is instability of the substrate. There is little reason for a lichenometric age to be older than the actual age of the landform. Therefore, the oldest age is used whenever significantly different (> 20 years) lichenometric ages were obtained for a single landform.

On many landforms, lichens overlap each other to form extensive lichen mats (fig. 9). Data sets collected from such landforms are not indicative of the age of stabilization. The lichens that originally colonized the landform cannot be measured because they are beneath younger lichens. Therefore, such surfaces are too old to be dated by lichenometry. However, some of our data sets were collected on such landforms in order to compare the results obtained from more stable surfaces. The data sets from older moraines with lichen mats were not used in the reconstruction of past glacial margins (tables 1-3).

As previously mentioned, lichenometry does not necessarily date the age of a landform unless the surface stabilized immediately after formation. Rather, it dates the time when the surface was colonized by lichens. It may take years for moraines or portions of moraines to stabilize enough to allow for lichen colonization. However,

Figure 9: An example of a lichen mat indicative of second generation growth in Hooker Valley during the 2002 field season. Geomorphic landforms with lichen mats cannot be dated with lichenometry.

lichen colonies have been observed on exceptionally stable areas of rock cover on Tasman and Murchison Glaciers. Although unstable locations were avoided when a location to collect a data set was selected, this variable was not completely avoided. In many cases, ice-contact fans that cross cut a moraine have an older lichenometric age than the moraine. This lichenometric age reversal is caused by the fact that moraine slopes continue to shed boulders for years after the ice has retreated. The lower angle fans are stable as soon as they are free of water.

In Murchison Valley, the lateral moraine/ice-contact terrace of Baker Glacier has two different lichenometric ages. The data for MR 213 (fig. 4) were collected on the ice-contact terrace portion of the landform, whereas the data for MR 224 (fig. 4) were collected on the moraine ridge. Instability of the moraine slope led to the lichenometric age of the ice-contact terrace being 33 years older than that of the moraine ridge.

The lichenometeric ages of the outer end moraine set of Murchison Glacier are also out of sequence. Lichenometeric ages at sites MR 1 (fig. 4) and MR 2 (fig. 4) are too young due to instability of the surfaces, or operator errors. Significantly, older lichenometric dates were used for these deposits.

On the east side of Hooker Valley at site H 223 (fig. 5), the ice-contact fan has a lichenometric age that is 66 years older than the H 224 (fig. 5) moraine that it cross cuts. This discrepancy is attributed to the time required for the moraine slope to stabilize.

In Tasman Valley, the moraine forming the eastern shore of the Blue Lakes has four lichenometric ages ranging from AD 1802 to 1876 (fig. 6). Currently, the moraine has consistently low-angle slopes. However, this may not have been not the case when ice receded from the moraine. Localized slope instability may have caused the wide range of ages.

The data from sites T 315 and T 316 (fig. 6) should date the same outwash surface southwest of the terminal moraine. However, T 315 is 31 years older than T 316. No explanation can be found for this age discrepancy. The older date, T 315 was used to reconstruct the past glacial margins.

The outermost small moraine in the Murchison lobe of the Tasman system has three different lichenometric ages that also can be accounted for by slope instability (Fig. 6). The oldest age of AD 1847 was used to reconstruct past ice margins. The ice-contact fan at the north end of the Murchison lobe also has three different lichenometric ages ranging from AD 1851 to 1881 (fig. 6). No explanation can be found to rectify this inconsistency.

Aside from these few discrepancies, the large majority of data were internally consistent and therefore useful for determining minimum ages of glacial landforms. Lichen diameters decrease on landforms closer to the glaciers. This demonstrates the correct sequence of retreat where younger deposits are closer to the present-day ice margin than older deposits. Multiple lichen ages from a single landform are similar. For example, on the foreplain of Tasman Glacier four lichen counts were conducted in relic channels that grade to the outermost frontal moraine. The four lichen ages determined from these counts differed by only one year. Also, lichenometric ages correspond closely with historical dates that were not used in construction of the Lowell *et al.* (2003) growth curve. In Murchison Valley, the outermost moraine was dated with lichenometry to have formed between AD 1881 and 1908 (fig. 10). This agrees closely with the AD 1890 observation of Mannering that Onslow and Murchison Glaciers were joined.

### 7.1 Reconstructions of Former Glacial Margins

Terminal positions of Murchison Glacier are reconstructed for three different times (fig. 10). The outer most moraine belt, with the Onslow Hut, has four lichenometric dates from ice-contact fans and five lichenometric dates from the moraine itself. In most cases, the licheometric ages of the fans are slightly older than the lichenometric ages of the moraines which they cross cut. The dates from the ice-contact fans, AD 1881, 1885, 1892, 1906, give distinct ages for when water no longer flowed down over the fan deposit. The mean of the dates from the moraine, AD 1908, is used to give a minimum age of the moraine. Lichenometric ages from the fans indicate that the ice was situated against this moraine from AD 1881 through 1906, which would be consistent with the historical record. The mean of the three lichenometeric ages on the

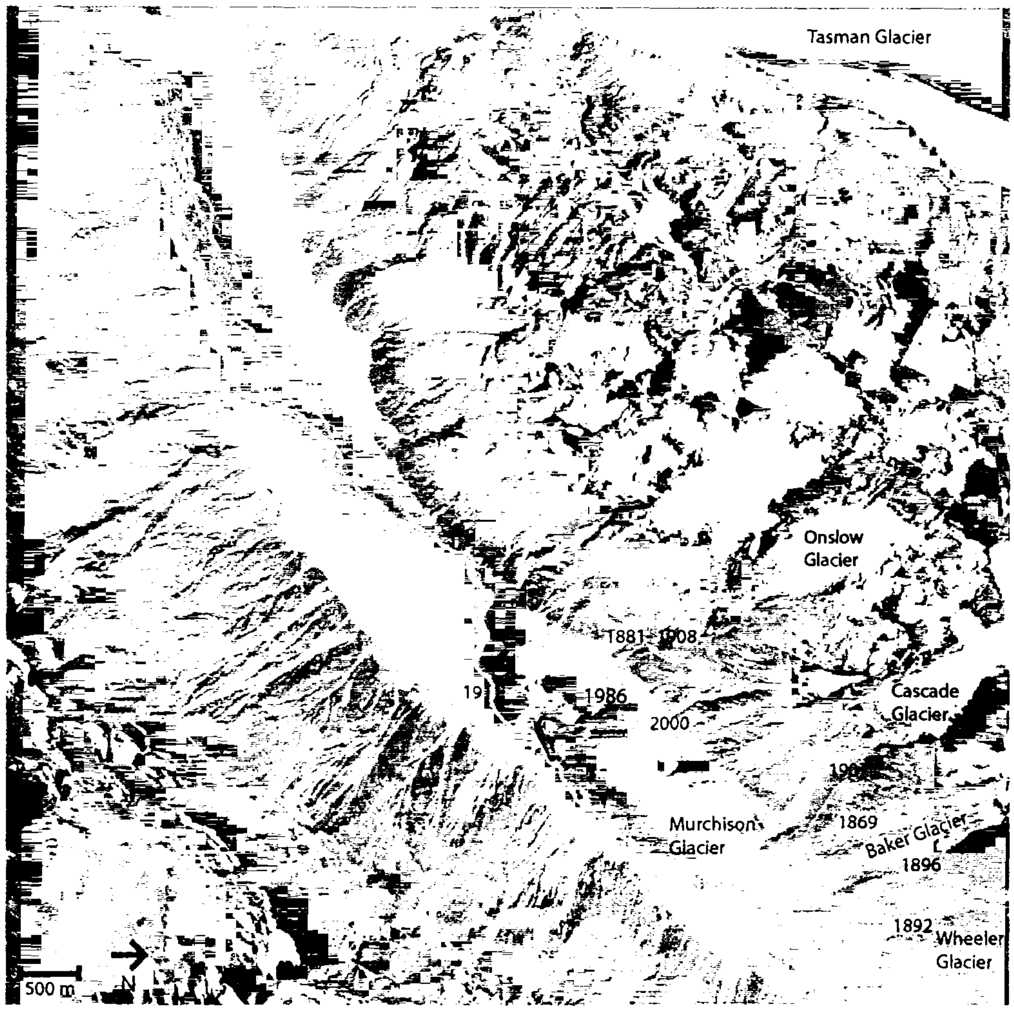


Figure 10: Reconstruction of past ice-margins in Murchison Valley. Aerial photograph from New Zealand Aerial Mapping survey number 8595, photograph number B/12.

moraine is AD 1908. Therefore, the terminus of Murchison Glacier rested against this outer moraine belt from about AD 1881 to 1908.

The more ice-proximal moraine belt is dated to about AD 1911 from one lichenometric age. The AD 1986 ice margin was determined from an aerial photographgraph (New Zealand Aerial Mapping survey number 8595, photograph number B/12) (fig. 10). The ages of the lateral moraines of Cascade, Baker, and Wheeler Glaciers were acquired by averaging the multiple lichenometeric dates.

Ice-marginal positions of Hooker Glacier were reconstructed for three different times (fig. 11). The oldest deposit for which a date was obtained is the outermost lightly vegetated moraine. An ice-contact fan cutting this moraine dates to AD 1815. The moraine itself has a lichenometric date of AD 1846. Therefore, it is concluded that the ice rested against this moraine between these two dates. The innermost lightly vegetated moraine has a lichenometric date of AD 1866. The AD 1986 ice margin was determined from the aerial photographgraph (New Zealand Aerial Mapping, survey number 8595, photograph number,C/16).

The terminus and sides of Tasman Glacier were reconstructed for a total of five different times (fig. 11). The moraine that makes up the eastern shore of the Blue Lakes has a lichenometric date of AD 1802, based on the oldest lichen count on the landform. Old outwash just distal to the outermost terminal moraine has a lichometric date of AD 1825. The mean of four lichenometric dates on channels grading to the outermost


Figure 11: Reconstruction of past ice-margins in Tasman and Hooker Valleys. Aerial Photograph from New Zealand Aerial Mapping survey number 8595, photograph number C/16.

moraine is AD 1952. These data, coupled with historical information, indicate that the terminus of Tasman Glacier rested against the outermost terminal moraine from about AD 1825 through 1952. The outermost continuous lateral moraine of Tasman Glacier in the mouth of Murchison Valley has a minimum age of AD 1851 determined by the average of three lichenometric dates from that surface. On the west side of Tasman Valley, the lateral moraine with the ice-contact/erosional slope can be dated in a manner similar to the outer most terminal moraine of Murchison Glacier. Based on lichonmetric dates derived from four ice-contact fans, the ice was banked against the moraine from AD 1875 to 1913. The lichenometric date derived from the moraine itself indicates ice retreated at about AD 1920. The moraines at the top of the ice-contact/erosional slope on the east side of the valley date to about AD 1904, based on the average of five lichenmetric ages. Finally, the AD 1986 ice margin was reconstructed from the aerial photographgraph (New Zealand Aerial Mapping, survey number 8595, photograph number,C/16) (fig. 12).

### 7.2 Interpretation

The geomorphology of Murchison, Hooker, and Tasman Valleys suggests three episodes of glacial activity (fig. 4-7). The earliest phase is one of glacial advance. During this time, the old, heavily vegetated moraines were deposited in Hooker and Tasman Valleys. Dates of deposition may be as early as 7200 YBP (Gellatly, 1984) or as recent as ~850 YBP (Burrows, 1973). The lightly vegetated moraines at the south ends of the lakes in all three valleys delimit a second period of glacial advance at ~150 to 100 YBP. The glacier surfaces were at the top of the lateral moraine walls during this time, 100 to 150 m higher than present-day ice.

The geomorphology of all three valleys indicates glacial collapse subsequent to the deposition of the unvegetated moraines. Areas of hummocky terrain exist on the ice-proximal sides of these unvegetated terminal moraines in each valley. Hummocky terrain is suggestive of rapid glacier disintegration and burial of stranded ice blocks. Other features indicative of collapse include proglacial lakes in Murchison, Hooker, and Tasman Valleys. In areas of shallow relief, proglacial lakes occur at the termini of rapidly retreating glaciers. The lakes form when glacier ice loss exceeds sediment volume delivered to the terminus. Under these circumstances, a hole separates the glacier terminus from its former moraines and outwash head and subsequently fills to form a proglacial lake. Finally, trimlines and steep erosional slopes that previously were supported by ice flank the lateral margins of Murchison, Hooker, and Tasman Glaciers. These features also suggest recent surface lowering and glacial collapse.

Glacier margin deposits in the Southern Alps of New Zealand are similar to those created during the ice retreat of the last 150 years in the European Alps. For example, the foreplain of Rhone Glacier contains an area of hummocky terrain. Roseg and Gries Glaciers both terminate in proglacial lakes, and Tsidjiore Nouve, Rhone, Aletsch, Gorner and Tschierva Glaciers all have exposed lateral moraine walls or trimlines associated with them (Kasser and Haeberli, 1981). The close resemblance between the glacial geomorpholgy of the Southern Alps and that of the European Alps suggests similar glacial histories.

Not only is the glacial geomorphology similar but the timing and amount of recession of New Zealand and European glaciers are comparable as well. Since the mid-nineteenth century, the termini of Hooker, Murchison, and Tasman Glaciers each have retreated ~1.5 km. This is comparable to the timing and amount of recession shown by Upper Grindelwald, Argentiere, and Rhone Glaciers in the European Alps (fig. 12), as well as by other glaciers in Europe, Canada, Iceland, Chile, and Pakistan (Oerlemans, 2001). However, recession of glacier margins alone is not always an adequate proxy for climate change. Kirkbride (1993) suggested that retreat of heavily debris-covered glacial termini is a poor indicator of climatic change. Furthermore, glacial termini calving into proglacial lakes may have a reduced sensitivity to climatic change (Kirkbride, 1993). Glaciers studied in this project suffer from both of these problems. The New Zealand glaciers have a thick mantle of surface debris due to the geology of the area. The local bedrock is composed of the Torlesse Supergroup (Suggate, 1978). Despite the well-indurated nature of this sandstone, it is shed readily from the valley walls onto the

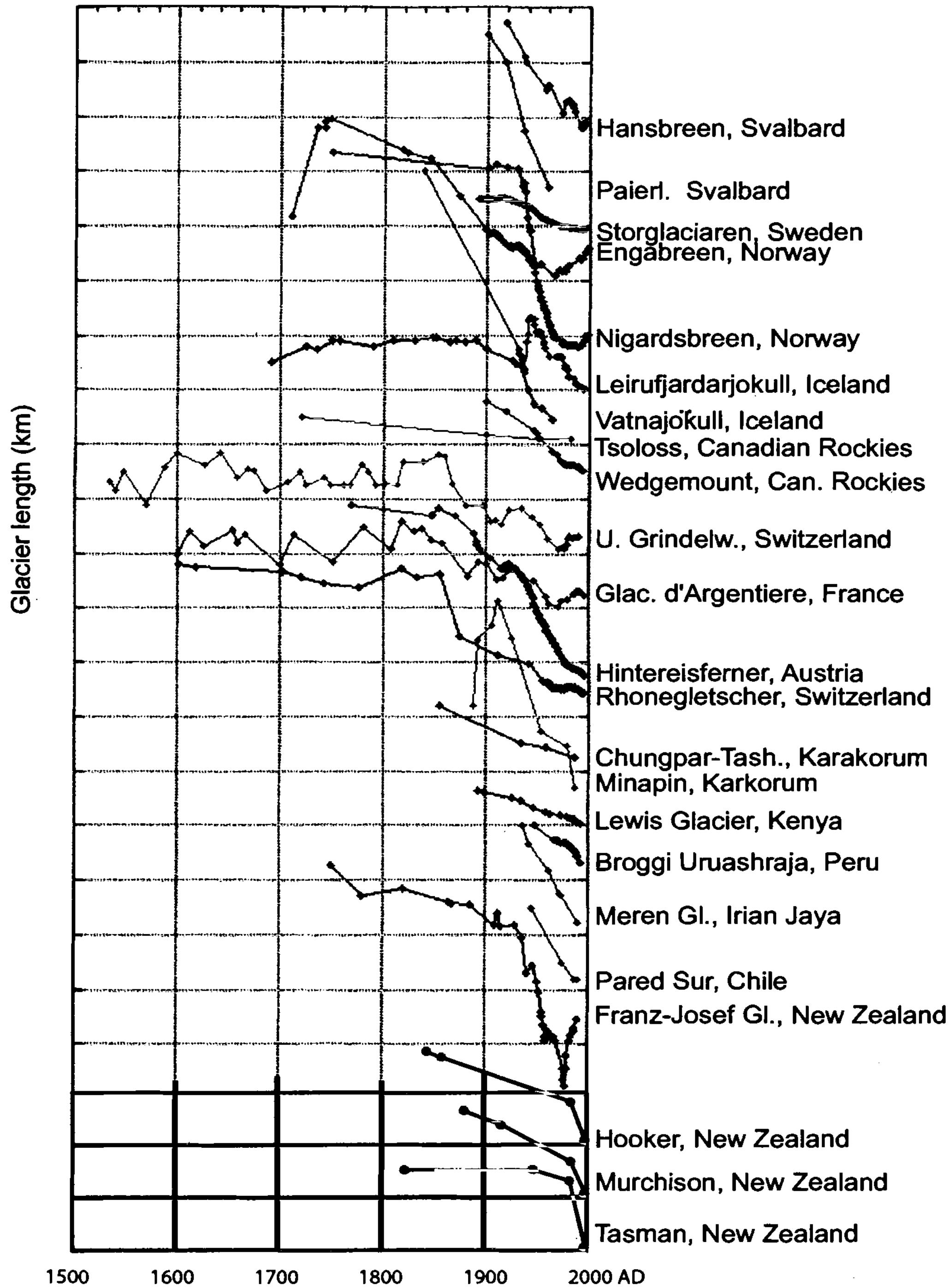


Figure 12: The worldwide retreat of temperate alpine glaciers since AD 1500 from Oerlemans (2002) compared with the findings of this study.

glaciers as a result of the active faulting and rapid uplift of the Southern Alps (more than one centimeter per year post-glacial uplift; Suggate, 1978) Thick debris cover insulates underlying ice and protects it from melting. Chinn (1996) claimed the substantial debris mantle found on glaciers such as Murchison, Hooker, and Tasman can slow melting rates by as much as 90%. Therefore, in order to document the response of these glaciers to recent climate variations, it is necessary to examine changes in addition to terminus retreat.

Significant downwasting of the ice surface has taken place in New Zealand. Laser range finder and Brunton compass measurements taken from lateral moraines near the termini indicate a 95, 130, and 156 m vertical lowering of the ice surface for Hooker, Tasman, and Murchison Glaciers, respectively, since the late nineteenth century. These numbers compare favorably with glacier-surface lowerings in the European Alps. For example, the Rhone Glacier in Switzerland downwasted ~150 m near its terminus between AD 1856 and 1969 (Kasser and Haeberli, 1981). Rapid downwasting is consistent with glacial collapse in both the Northern and Southern Hemispheres.

In summary, the glacial geomorphology and the timing of recent recession are similar in the Southern Alps of New Zealand and in the European Alps. Both areas show evidence of glaciers in advanced positions with surface elevations as much as 150 m higher than at present about a hundred and fifty years ago. At this time, glacier margins in New Zealand rested against the unvegetated end moraines and the crests of the lateral moraines. The presence of hummocky terrain, proglacial lakes, and exposed lateral moraine walls and trimlines indicate rapid ice collapse and downwasting in both New Zealand and the European Alps subsequent to the glacial advance.

Despite the fact that the termini of New Zealand glaciers may not be as responsive as European glaciers due to unusually thick debris cover, they show a similar volume loss. Over the last 150 years the Alps have lost 30 to 40% of their glacierized area, whereas the Southern Alps have lost 23 to 32% over the same period (Chinn, 1996). The retreat of glaciers in the European Alps is attributed to an increase in temperature over the last century and a half (Haeberli, 1994). Comparison of mid-nineteenth century ELAs to present-day snowlines indicates that a warming of 0.5 °C has occurred in the European Alps (Haeberli, 1994) . Based on similarities in the timing and volume of ice collapse, the retreat of New Zealand glaciers is consistent with an increase in temperature in New Zealand similar to that found in Europe. A snowline study conducted by Chinn (1996) found a warming of 0.6 °C has occurred in the Southern Alps since the mid-nineteenth century. Therefore, the current retreat of glacial margins in both the Northern and Southern Hemispheres probably is due to a global temperature increase over the past century and a half.

### 7.3 Climatic Origins of the Glacial Retreat of the Last 150 Years

The Broecker bipolar see-saw theory (Broecker, 1998) is at odds with the findings of this study. The bipolar seesaw theory predicts that when the Northern Hemisphere warmed over the last 150 years then the Southern Hemisphere should have cooled. However, this was not the case, as both hemispheres have warmed over the past 150 years. Therefore, the recent near synchronous global warming must be explained by a different mechanism. Possible forcing mechanisms for the recent retreat of glaciers include decreased global albedo due to reduced volcanic activity, increased trace gas concentrations in the atmosphere, or changes in solar activity.

Small dust particles and sulfuric acid aerosols increase global albedo following large volcanic eruptions. According to modeling experiments, periods of high volcanism may increase global albedo enough to lower Earth's temperature by ~1°C (Pollack *et al.*, 1976). If true, this could account for the 0.5 to 1.0°C cooling experienced during the LIA (Rind and Overpeck, 1993). Porter (1986) used the acidity level in a Greenland ice-core as a proxy for Northern Hemispere volcanic activity. Periods of low pH (high volcanism) in Greenland ice corresponded temporally with periods of glacial advance in the Northern Hemisphere for the past 1000 years (fig. 13). A recent reduction in volcanism would thus explain the bipolar synchronous warming of the last 150 years. The weakness

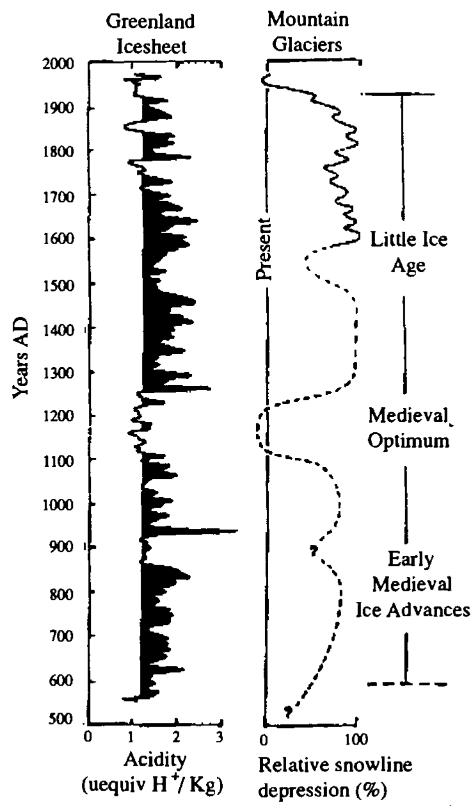


Figure 13: Temporal correlation between volcanic activity and climatic change.
From Porter, 1986.

of the volcanic forcing theory as a whole is that it essentially ignores Southern Hemisphere volcanic eruptions.

Trace gases, such as $CO_2$, in the atmosphere absorb outgoing radiation emitted by the earth and re-emit it back to the earth. This tends to warm the planet. The concentration of $CO_2$ in the atmosphere has increased dramatically since the mid-eighteenth century, due partially to anthropogenic burning of fossil fuels (Neftel *et al.*, 1985) (fig. 14). Based on a modeling experiment conducted by Rind and Overpeck (1993), 0.2 to 0.5 °C warming since the mid-nineteenth century may be attributable to increased concentrations of $CO_2$ and other trace gases in the atmosphere.

Variations in solar activity may also explain the climatic changes of the LIA. Until recently the "solar constant" was thought to be a constant. However, new satellite data acquired over the past few decades indicate variability of ~0.1% attributable to the relatively weak 11 year sun spot cycle (Beer, 2000). These findings introduce the possibility of longer time scale solar oscillations that may be responsible for Holocene climatic change as proposed by Denton and Karlén (1973).

Historical periods during which few sunspots were observed (low solar activity) correlate temporally with periods of climatic cooling. The Maunder (1645-1715 AD), Dalton (1800-1820 AD), and 1900 (1880-1900 AD) minima were times during which few sun spots were observed (fig. 15). These minima correspond temporally with climatic cooling in the Northern Hemisphere (Beer *et al.*, 2000). The recent increase in solar activity since the 1900 minimum could explain the global retreat of alpine glaciers over the last century.

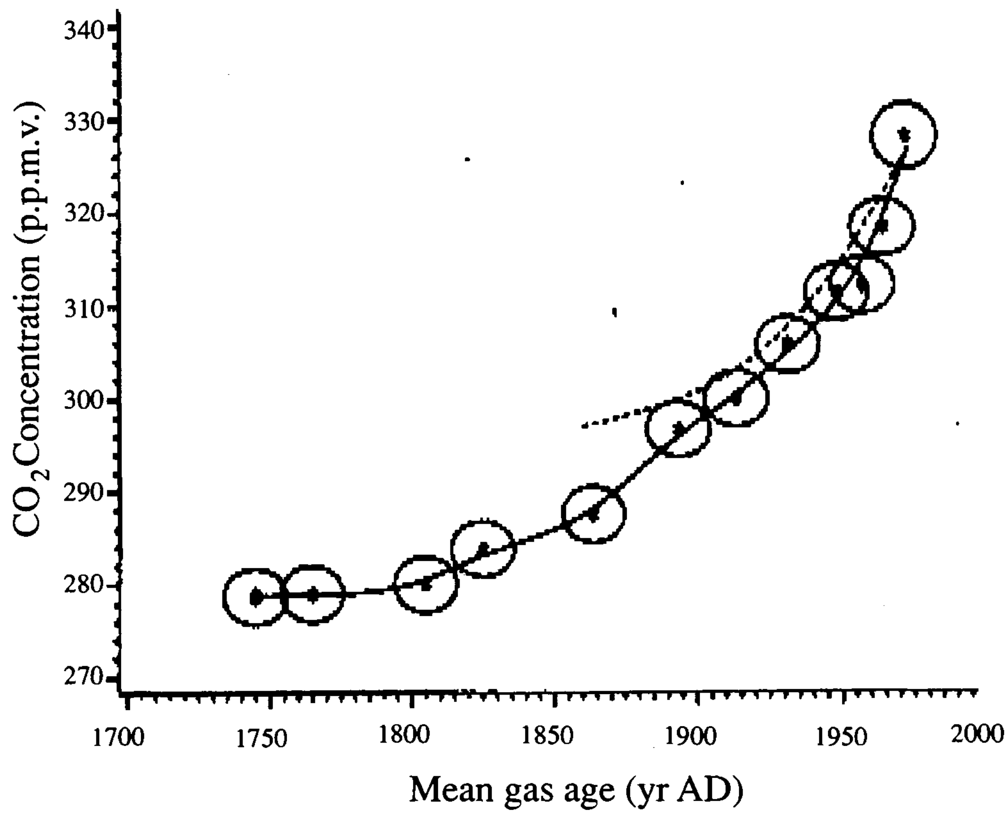


Figure 14: Increasing $CO_2$ concentration in the atmosphere with time. From Neftel et al., 1985.

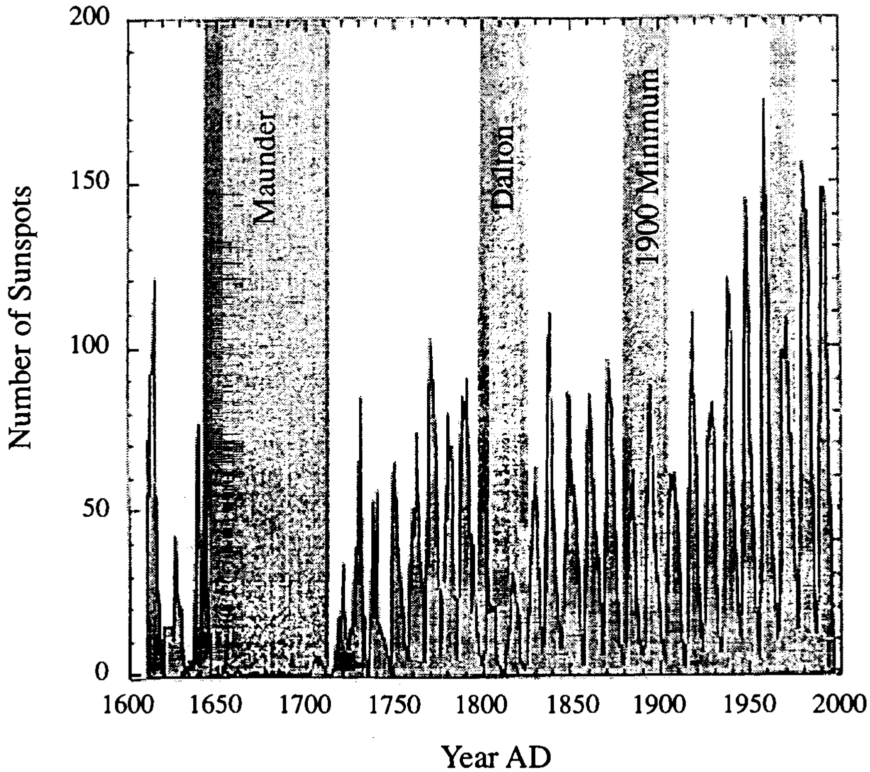


Figure 15:. Temporal correlation between solar activity and climatic change. From Beer et al., 2000.

# 8. CONCLUSIONS

- Marked retreat of Murchison, Hooker and Tasman Glaciers in New Zealand has taken place since the mid-to-late nineteenth century.

- The retreat of the New Zealand glaciers is similar in timing and magnitude to the glacial retreat following the LIA in the European Alps, Scandinavia, and elsewhere in the Northern Hemisphere. This is consistent with synchronous global warming, with the possible exception being large parts of Antarctica.

- Synchronous warming of both polar hemispheres following the LIA cannot be explained by the Broecker bipolar see-saw model. Rather, it may be due to a mechanism that would cause synchronous global warming. The possibilities include reduced albedo due to fewer volcanic eruptions, increased greenhouse gas concentrations in the atmosphere, and changes in solar activity.

## REFERENCES


Beer, J., Mende, W., Stellmacher, R., 2000. The role of sun in climate forcing. Quaternary Science Reviews, 19, 403-415.

Birkeland, P.W., 1982. Subdivision of Holocene glacial deposits, Ben Ohau Range, New Zealand, using relative-dating methods. Geological Society of America Bulletin, 93, 433-449.

Black, J.L., 2001. Can a Little Ice Age Climate Signal Be Detected in the Southern Alps of New Zealand? B.S. Thesis, Institute for Quaternary and Climate Studies, University of Maine, Orono.

Brodrick, T.W., 1891. The Murchison Glacier: The Original Report on the Murchison by Surveyor Brodrick in 1891. Unpublished.

Broecker, W.S., 1998. Paleocean circulation during the last deglaciation: A bipolar seesaw?. Paleoceanography, 13, 119-121.

Bull, W.B., Brandon, M.T., 1998. Lichen Dating of Earthquake-Generated Regional Rockfall Events, Southern Alps, New Zealand. GSA Bulletin, 110, 60-84.

Burrows, C.J., 1973. Studies of some glacial moraines in New Zealand: Two ages of moraines of the Mueller, Hooker, and Tasman Glaciers. New Zealand Journal of Geology and Geophysics, 16, 831-855.

Burrows, C.J., 1980. Radiocarbon dates for post-Otiran glacial activity in the Mount Cook region, New Zealand. New Zealand Journal of Geology and Geophysics, 23, 239-248.

Chinn, T.J., 1981. Use of rock weathering rind thickness for Holocene absolute age-dating in New Zealand. Arctic and Alpine Research, 13, 33-45.

Chinn, T.J., 1996. New Zealand glacier response to climate change of the past century. New Zealand Journal of Geology and Geophysics, 39, 415-428.

Crawford, R.J., 1953. Mt. Cook Alpine Regions. Crown copyright.

Denton, G.H., Karlen, W., 1973. Holocene climatic variations- Their pattern and possible causes, Quaternary Research, 3, 155-205.

Eddy, J. A., 1977. Climate and the changing sun. Climatic Change, 1, 173-190.

Gellatly, A.F., 1984. The use of rock weathering-rind thickness to redate moraines in Mount Cook National Park, New Zealand. Arctic and Alpine Research, 16, 225-232.

Gellatly, A.F., 1985. Historical Records of Glacier Fluctuations in Mt. Cook National Park, New Zealand: A Century of Change. The Geographical Journal, 151, 86-99.

Gellatly, A.F., Röthlisberger, F., Geyh, M.A., 1985. Holocene glacier variations in New Zealand (South Island). Zeitschrift Für Gletscherkunde und Glazialgeologie, 21, 265-273.

Gordon, J.E., 1980. Recent climatic trends and local glacier margin fluctuations in West Greenland: Nature, 284, 157-159.

Grove, J.M., 1988. The Little Ice Age. New York, Methuen.

Haast, J.V., 1879. Geology of the Provinces of Canterbury and Westland, New Zealand. Christchurch, The Times Office: 31.

Haeberli, W., 1994. Glacier and Permafrost Changes in the Alps. Mountain Environments in Changing Climates, ed. Beniston. New York, Routledge: 96.

Harding, K.A., 1985. Glacier Inventory of the Premier Range, British Columbia. B.A., University of Western Ontario, London.

Harper, A.P., 1935. Glacier Retreat. New Zealand Alpine Journal, 5, 322-326.

Harrington, H.J., 1952. Glacier Wasting and Retreat in the Southern Alps of New Zealand. Journal of Glaciology, 2, 140-145.

Holzhauser, H., Zumbuhl, H.J., 1999. Glacier Fluctuations in the Western Swiss and French Alps in the 16th Century. Climatic Change, 43, 223-237.

Institute for Geological and Nuclear Sciences, 2001, Lower Hutt.

Karlén, W., Denton, G.H., 1976. Holocene glacial variations in Sarek National Park, northern Sweden. Boreas, 5, 25-56.

Kasser, P., Haeberli, W., eds., 1981. Switzerland and Her Glaciers, Kümmerly and Frey, Berne.

King, C.A.M, 1959. Geomorphology in Austerdalen, Norway. Geographical Journal, CXXV, 357-369.

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, 232-240.

Lendenfeld, R.V., 1884. Karte des Tasman-Gletscher. Erganzungsheft No. 75 'Petermanns Mitteilungen', 1-80.

Lowell, T.V., Schoenenberger, K., Deddens, J., Denton, G., Black, J., Smith, C., in press. Lichen Calibration Curve for the Aoraki/Mount Cook Area of New Zealand.

Luckman, B.H., 2000. The Little Ice Age in the Canadian Rockies. Geomorphology, 32, 357-384.

Mannering, G.E., 1890. On the Murchison Glacier. Transactions and Proceedings of the New Zealand Institute, 23, 355-366.

Mannering, G.E. Climbing and Explorations in the Southern Alps.

Mathews, J.A., 1974. Families of lichenometric dating curves for the Storbreen gletschervorfeld, Jotunheimen, Norway. Norsk Geografisk Tidsskrift, 28, 215-235.

McCarthy, D.P., Smith, D.J., 1994. Historical glacier activity in the vicinity of Peter Lougheed Provincial Park, Canadian Rocky Mountains. Western Geography, 4, 94-109.

McGregor, V.R., 1967. Holocene moraines and rock glaciers in the central Ben Ohau Range, South Canterbury, New Zealand. Journal of Glaciology, 47, 737-748.

Mottershead, D.N., White, I.D., 1972. The lichenometric dating of glacier recession, Tunsbergdal Southern Norway. Geografiska Annler, 54A, 47-52.

National Geographic Society website, 2003, www.nationalgeographic.com/maps/.

Neftel, A., Moor, E., Oeschger, H., Stauffer, B., 1985. Evidence from polar ice cores for the increase in atmospheric $CO_2$ in the past two centuries. Nature, 315, 45-47.

New Zealand Aerial Mapping, 1986. Hastings.

New Zealand Aerial Mapping, 1965. Hastings.

Oerlemans, J., 2001. Glaciers and Climate Change. Lisse, A.A. Balkema.

Ostrem, G., Liestöl, D., Wold, B., 1976. Glaciological investigations at Nigardsbreen, Norway. Norsk Geografisk Tidsskrift, 30, 187-209.

Péwé, T.L., Reger R.D., 1983. Delta River area, Alaska Range. Richardson and Glenn Highways, Alaska. Guidebook to permafrost and Quaternary geology, eds. Péwé, T.L. and Reger, R.D. Fairbanks, State of Alaska: 47-136.

Pollack, J.B., Toon, O.B., Sagan, C., Summers, A., Baldwin, B., Van Camp, W., 1976. Volcanic explosions and climatic change. A theoretical assessment: Journal of Geophysical Research, 81, 1071-1083.

Porter, S.C., 1981. Glaciological evidence of Holocene climatic change. Climate and History, Wigley ed. Cambridge, Cambridge University Press, 82-110.

Porter, S.C., 1986. Pattern and forcing of Northern Hemisphere glacier variations during the last millennium. Quaternary Research, 26, 27-48.

Rind, D., Overpeck, J., 1993. Hypothesized causes of decade-to-century-scale climate variability. Climate model results: Quaternary Science Reviews, 12, 357-374.

Schoenenberger, K.R., 2001. Little Ice Age Chronology for Classen and Godley Glaciers, Mount Cook National Park, New Zealand. B.S. Thesis, Department of Geology, University of Cincinati, Cincinnati.

Sigbjarnarson, G., 1970. On the recession of Vatnajökull. Jökull, 20, 50-61.

Suggate, R.P., ed., 1978. The Geology of New Zealand, E.C. Keating, Wellington.

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