The Last Glacial Maximum and Deglaciation in Southern New Zealand: New Pollen-climate Reconstructions

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Abstract

The project builds upon existing knowledge of late Quaternary palaeoenvironmental change and tests the recently developed New Zealand INTIMATE (**I**ntegration of **I**ce **M**arine and **T**errestrial archive) climate event stratigraphy (NZ-I CES; 30-8 ka). Four pollen and sediment records from three climatically contrasting regions in the South Island provide a vegetation and climate history for this area between 38-4 ka. In this study, the Last Glacial Cold Period (LGCP; c. 31.4-18.9 ka) is characterised by a two step cooling, with the coldest conditions, reaching possibly $>5.3^{\circ}\text{C}$ cooling, occurring between 21-19 ka, marking the Last Glacial Maximum. A new precipitation proxy using macrophyte pollen concentrations at an eastern South Island site suggests dominantly dry conditions prevailed during the LGCP except for two periods of wetter climate around 26-24 ka and 21 ka. The dry periods correspond with evidence of glacial advance, colder environments and possibly increased intensity of the southern westerlies. Conversely, the wet periods coincide with reduced glacial activity, milder climates and decreased westerly wind intensity. Deglaciation began between 18.9-18.4 ka followed by rapid climate amelioration culminating with *Dacrydium cupressinum*-dominant lowland forest at western sites as early as 11.9 ka, indicative of the start of the Holocene. A disturbance in forest development occurs between 13.4-11.9 ka in one record and may be indicative of a minor cooling within the timeframe of a late glacial climate reversal recognised in the NZI-CES.

Overall the project results (timing and pattern of climate change) broadly align with the NZ-I CES. However, there are some disparities, in particular during the LGCP, which this study suggests began at least 3-4 ka earlier than concluded in the NZ-I CES. The NZ-I CES oversimplifies the complexity of the LGCP which contains evidence of significant climate variability that may be important for an understanding of the possible forcing factors on climate change. The chronology derived from the current study supports recent evidence that points towards a younger, refined age of 25.4 ka for the Kawakawa/Oruanui Tephra, a key chronostratigraphic marker for the LGCP. Pollen-climate models and

Environmental Lapse Rates were used to quantify changes in mean annual temperatures with sometimes conflicting results. This research reveals some limitations of the current New Zealand pollen-climate transfer function when applied to reconstruction of cold climate periods in particular. These include a lack of limitations with modern analogues and a number of wide-ranging pollen taxa that encompass a broad climate envelope. The current research also highlights the potential of regional climate regimes and spatial differences in vegetation and inferred climate reconstructions. These differences pose a major limitation for a New Zealand-wide composite. While the NZ-I CES provides a valuable framework of climate change during a period of large climate variability, results of this study highlight aspects that need further consideration and revision.

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List of Abbreviations

AAA	Acid-Alkali-Acid
ACC	Antarctic Circumpolar Current
ACR	Antarctic Cold Reversal
AMS	Accelerated Mass Spectrometry
BP	Before present
D-O	Dansgaard-Oeschger
ELA	Equilibrium-Line Altitude
ELR	Environmental Lapse Rate
ENSO	El Niño-Southern Oscillation
EPICA	European Project for Ice Coring in Antarctica
ESI	East South Island
GISP2	Greenland Ice Sheet Project 2
GRIP	Greenland Ice Core Project
INTIMATE	Integration of Ice Marine and Terrestrial archives
IPO	Inter-decadal Pacific Oscillation
ITCZ	Inter-Tropical Convergence Zone
KOT	Kawakawa/Oruanui tephra
LGCP	Last Glacial Cold Period
LGIT	Last Glacial/Interglacial Transition
LGM	Last Glacial Maximum
LGR	Late Glacial Reversal
LOI	Loss on Ignition
MAP	Mean Annual Precipitation
MAT	Mean Annual Temperature
MIS	Marine Isotope Stage
NGRIP	North Greenland Ice core Project
NNI	Northern North Island
NSI	Northern South Island
NZ-I	New Zealand INTIMATE group
NZ-I CES	New Zealand INTIMATE Climate Event Stratigraphy
ODP	Ocean Drilling Project
OSL	Optically Stimulated Luminescence
PC	Principal Component Axes
PCA	Principle Component Analysis
PLS	Partial Least Squares Regression
PSA	Particle Size Analysis
RMSE	Root Mean Square Error
RMSEP	Root Mean Square of Predictions Error
SAW	Sub-Antarctic Surface Waters
SNI	Southern North Island
SSS	Sea Surface Salinity
SST	Sea Surface Temperature
STF	Sub-Tropical Front

STW	Sub-Tropical Surface Waters
TDP	Total Dry Land Pollen
TVZ	Taupo Volcanic Zone
W-MAT	Weighted Modern Analogue Technique
WSI	West South Island
YDC	Younger Dryas Chronozone

Chapter 1: Introduction and Rational

1.1 Global pattern of climate change

It is now generally accepted that changes in the Earth's orbital geometry are a major driver of climate change, and the glacial/interglacial cycles observed during the Quaternary (past c. 2.58 Ma, Gibbard *et al*, 2010; Imbrie and Imbrie, 1979). Since c. 700 ka a 100 ka eccentricity/precisely dominated glacial/interglacial cycle is recorded globally (Imbrie and Imbrie, 1979; Huybers and Wunsch, 2005; Liu *et al*, 2008) with a major assumption that the global pattern of climate change occurs synchronously between the hemispheres. Indeed, this has been validated by ice core records (North Greenland Ice Core Project (NGRIP) members, 2004; EPICA community members, 2006) deep-sea records (Kitamura and Kanoto, 2007; Crundwell *et al*, 2008; Herbert *et al*, 2010) and terrestrial archives (Sun *et al*, 2005; Prokopenko *et al*, 2009). However, superimposed upon the astronomical global cycles are numerous millennial scale, abrupt climate oscillations which point towards a phasing of climate change between the poles (North Greenland Ice Core Project members, 2004; EPICA community members, 2006).

GRIP (Greenland Ice Core Project; Dansgaard *et al*, 1993; Johnsen *et al*, 2001) and GISP2 (Greenland Ice Sheet Project 2; Grootes *et al*, 1993) ice core records have shown the existence of numerous millennial-scale climate oscillations. Between 110-14 ka, 25 Dansgaard/Oeschger events (D-O events), periods of abrupt warming to interstadial conditions over a few centuries followed by gradual cooling to glacial temperatures, have been identified (Dansgaard *et al*, 1993; Grootes *et al*, 1993; Johnsen *et al*, 2001; NGRIP members, 2004). Likewise, similar warming events have been identified in Antarctic ice core records (EPICA community members, 2004). However, the pattern and timing of these events between the poles appear to be out of phase with each other. This elucidated that an asynchronous millennial scale pattern of climate change is controlled by variations in the oceanic hydrological circulations through the bipolar seesaw effect (EPICA community members, 2006).

The precise link between the hemispheres is still uncertain with the idea of asynchronous phasing of millennial scale climate change remaining controversial (Alley *et al*, 2002; Schmittner *et al*, 2003; Huybers, 2004; Alley, 2007). The Last Glacial Maximum (LGM) and Last Glacial/Interglacial Transition (LGIT) provide a valuable testing point for assessing the pattern of global climate change. Modelling efforts for this period have helped to clarify the pattern as the outputs show the bipolar ice-core signals can be accurately reproduced using various modelling techniques (Stocker, 2000; Stocker and Johnsen, 2003; Schmittner *et al*, 2003; Knutti *et al*, 2004). This bipolar seesaw pattern is not restricted to ice core records with evidence suggesting the pattern extends into the mid-latitudinal regions as well. The Global multiproxy data set indicates a globally (near) synchronous ice age cycle with a superimposed north-south millennial bipolar see-saw pattern (Pahnke *et al*, 2003; Lamy *et al*, 2004 Pahnke and Zahn, 2005). However, there is uncertainty and conflicting evidence from mid-latitude sites with some records observing synchronous climatic oscillations to the Northern Hemispheric signal (e.g. Denton and Hendy, 1994; Ivy-Ochs *et al*, 1999) or slightly offset from the Northern Hemisphere signal (e.g. McCulloch *et al*, 2000). The paucity of results and controversial observations makes it difficult to assess the nature and extent of any asynchronous interhemispheric pattern of climate within the Southern Hemisphere mid-latitudes. There is, therefore, a need to increase the reliability and robustness of the proxy climatic reconstructions for the Southern Hemisphere and to produce regional climatic stratigraphies of the last glacial period before answers to the nature and pattern of interhemispheric pattern of climate change can be tackled with any certainty. The New Zealand INTIMATE (INTEgration of Ice core Marine and Terrestrial archives) group was set up in 2003 to accomplish just that.

1.2 New Zealand INTIMATE group (NZ-I)

New Zealand is a temperate, sub-continental, oceanic island located in the south east Pacific Ocean. It lies in the path of the southern westerly zonal winds, the strongest mean oceanic winds in the world driving the deep Antarctic Circumpolar Current (ACC) eastwards around the Antarctic continent, and pushing surface waters away from Antarctica causing upwelling of intermediate water south of the

ACC (Russell *et al*, 2006). The westerlies link climatic processes of New Zealand with those from the Antarctic domain. This provides potential to examine any Antarctic influence on New Zealand and therefore the Southern Hemispheric mid latitudes (Sugden *et al*, 2005). New Zealand's glaciers, vegetation and fauna have been observed to respond quickly to climate change. These responses have been preserved with palaeoenvironmental records capturing and ageing any climate transitions and millennial scale events (Vandergoes and Fitzsimons, 2003). New Zealand is also one of the few land masses located within the oceanic dominant Southern Hemisphere. Therefore terrestrial climate records can be compared and related to marine and ice-core records. New Zealand therefore holds an important strategic location for providing palaeoclimatical information and for assessing the pattern of climate change for the region. Even so there are still relatively few records that span the Last Glacial Maximum (LGM) and the Last Glacial/Interglacial Transition (LGIT). However, the NZ-INTIMATE (NZ-I) group identified key onshore and offshore proxy climate reconstructions, which were used to produce the NZ-INTIMATE climate event scheme (from now on known as NZ-I CES).

The NZ-I group focused on the period spanning 30-8 ka to encompass the onset of the coldest part of the last glacial up to the early Holocene epoch, a period of known large climate variability. (Alloway *et al*, 2007; Lowe *et al*, 2008). This project included seven high resolution continuous proxy records from Auckland Maars, Kaipo, Otamangakau wetlands, Okarito wetland, two speleothem records from northern Westland and the marine core MD97-2121 record (Figure 1.1). Numerous fragmentary records from glacial and fluvial sequences, loess deposits and aeolian quartz accumulations were used to aid the production of the NZ-I CES. Three distinct climate phases were recognised: the last glacial cold period (LGCP) spanning 28-18 ka and includes the Last Glacial Maximum (LGM, 21-18 ka); the Last Glacial/Interglacial Transition (LGIT, 18-11.5 ka) and the Holocene warm epoch (11.5 ka to present). However, superimposed upon these temporally long climate transitions are millennial scale climatic oscillations including the warm and variable climate interval during the LGCP (27-21 ka), and the Late Glacial Reversal (LGR 15-11.5, Figure 1.1). The timing of these climatic changes appears to closely follow an Antarctic pattern and timing (EPICA Community Members,

2006; Alloway *et al*, 2007). The timing of the onset of the LGCP and the LGR, as concluded by the NZ-I group, appear to occur asynchronously to their Northern Hemispheric counterparts. However, this is only a provisional scheme and the NZ-I group recognises the need to improve upon the results by the addition of more high resolution, chronologically robust records that incorporate new methodologies such as quantitative pollen-climate reconstructions and refinement to glacial records (Alloway *et al*, 2007).

1.2.2 Limitations to the NZ-I CES

The provisional NZ-I CES has produced a useful model of climate change in New Zealand for the past 30 ka. The process of synthesising multiple records to produce a single CES is not straightforward, however, the NZ-I CES has provided a useful CES that can be tested. Nevertheless, there are some limitations with the provisional NZ-I CES and areas that can be improved. These are discussed further below.

A limited number of high resolution, continuous records spanning the LGCP are available in New Zealand and therefore in the NZ-I CES. Of the continuous records used, only five extend beyond the LGM. As such the robustness of the hypothesised CED is restricted. The records used for this period also show some uncertainties regarding the timing of climate periods and millennial scale climate oscillations (Table 1.1). For instance the timing of the LGCP onset and the pattern or existence of the LGCP warmer variable period is not clear (refer to Table 1.1; Alloway *et al*, 2007). An increase to both the fragmentary and continuous records that span the LGCP would help produce a more accurate and robust conclusion of climate change during this period.

The Late Glacial Reversal (LGR) is a controversial climate event in New Zealand (refer to section 2.5.3). and as a consequence makes it a difficult event to constrain in the NZ-I CES. Nevertheless the NZ-I CES concluded that the LGR occurred between 13.5-11.6 ka. However, following closer inspection of the records used by the NZ-I group, both the continuous (Table 1.1) and fragmentary records do not so clearly constrain the timing of the LGR and an overall age range

of this event spans 15-11 ka. This is significant as the concluded 13.5-11.6 ka age is suggestive of an asynchronous 'Antarctic' timing of climate change for New Zealand. However the 15-11 ka age would include both an Antarctic Cold Reversal (ACR) and Younger Dryas chronozone (YDC) timing (Alloway *et al*, 2007) making conclusions, as to whether New Zealand follows an Antarctic or Greenland pattern of change, difficult to elucidate convincingly. For this event, higher resolution records that have better age constraint is needed to aid the NZ-I CES and refine the timing of the LGR.

Recent research by Kidson (2000) and Lorrey *et al* (2007) have elucidated that New Zealand can be separated into modern climate regions. The Australasian-INTIMATE meeting in June 2008 (Vandergoes and Barrell, 2008) suggest that this could also be applied to the NZ-I CES. However, as already mentioned there is a paucity of sites used in the project, particularly in the South Island, which can be split into three climate regions, but currently only one climate region with continuous proxy records is used in the CES. Therefore, to achieve a more rounded regional picture of the different climate zones New Zealand an increase in reliable sites, for the South Island in particular, is required.

In summary, although the preliminary NZ-I CES has made the best of the available data, there are still some limitations with the CES. There is a paucity of high resolution terrestrial records used in the NZ-I CES, particularly from the South Island and during the LGCP. There is also uncertainty concerning some important climate transitions and oscillations including:

- Timing of the onset of the LGCP.
- Climate variability patterns during the LGCP.
- Timing and duration of the LGR.

It will also be useful to more robustly confirm the following:

- Timing of Termination 1.
- The timing of the Holocene.
- Pattern of climate change during the deglaciation.

Record	LGCP onset	LGCP 'warming complex'	Termination I	LGR	Early-Holocene warm period	Mid-Holocene warm period
Auckland Maars	27-28	21-27	18-19	11-14	10-11	6.2
Kaipō	-	-	18	12.4-13.8	11	-
Speleothem	26	21-24	17.9	11.6-13.5	10.8-11.6	6.5-6.8
Okarito	28-29	21-27	18	11-15	11-12	-
MD97-2121	ca 27	-	18	11.5-13.7 (surface water) 12.5-14.1 (bottom waters)	7.9-9.0 SST- alkenone	-

Table 1.1: A summary of the timing of the key climate events in New Zealand terrestrial and marine records. All ages are expressed as ka unless otherwise stated and SST = Sea Surface Temperatures (taken from Alloway *et al*, 2007).

Figure 1.1: A collation of simplified summary diagrams from sites around New Zealand with b), d), e), f), h) and j) the main continuous records used in the NZ-I CES. a) Vandergoes and Fitzsimons, 2003; b) Newnham *et al.*, 2007a; c) Vandergoes *et al.*, 2008; d) Williams *et al.*, 2005; e) Alloway *et al.*, 2007; f) Sandiford *et al.*, 2003; g) Newnham *et al.*, 2007c; h) Newnham and Lowe, 2000; i) Turney *et al.*, 2003; j) Carter *et al.*, 2002.

1.3 Aim

This thesis therefore aims to test the hypothesised NZ-I CES by producing continuous, high resolution, well-dated palaeoenvironmental reconstructions. Three new sites spread across the South Island will be used to produce continuous records with a fourth providing a fragmentary record formed by three river terrace risers. The four sites collectively span >30-8 ka, but individually spanning part of this period. These sites represent altitudinal, north-south and west-east contrasts that today are characterised by strong contrasts in temperature and precipitation, and lie in the three different climate regions of the South Island as depicted by Kidson (2000) and Lorrey *et al* (2007). It is a secondary aim to investigate whether regional variations occur between the palaeoenvironmental record.

1.4 Rationale for the site location

NZ-I CES has highlighted the key areas in which improvement is required. Now that these limitations are recognised site selection targeting these limitations (increase the spatial resolution, increase the resolution of the palaeoenvironmental reconstructions or go back far enough to date the onset of the LGCP) can be chosen with more discretion. By using tarns, peat bogs or lakes constrained by moraines, geological age constraint can ensure a record can incorporate all, or some, of these events. The Kawakawa/Oruanui tephra (KOT) layer has a best age of ca 25.4 ka (Vandergoes *et al*, 2011). If this layer is found then the deposit must predate the LGM and has the potential to go back beyond the onset of the LGCP. The sites predating the LGM are located near the ice margin during the LGCP and will therefore provide a sensitive record of climate change for this period.

Two of the four sites are located in the Westland region, South Island (Figure. 1.2). Lake Mudgie, is a glacially impounded lake constrained by a pre-LGM moraine complex. This is confirmed by the presence of the KOT layer at 593-595 cm depth in the core. Therefore, Lake Mudgie can provide a continuous record spanning from c. 30-8 ka. Manks Tarn, the second site, lies within a LGCP lateral moraine and provides a climate record spanning from the start of ice retreat to c. 8 ka. This site has an incredibly high sedimentation rate with c. 13 m of sediment

accumulated over the past 18 ka and has the potential to provide the highest resolution record for the LGIT and early Holocene period in the region.

Forks Lagoon is located east of the Southern Alps in the Mackenzie Basin, south-west Canterbury. This is a glacially impounded lake bounded by a moraine predating the LGCP. This is confirmed by the presence of the KOT layer located at 2.5 m in the core. This site's climate contrasts strongly with the Westland region and enables us to analyse the climatic changes for a region that is currently drier than present-day West Coast. The palaeoenvironmental record will concentrate on the LGCP, and will potentially date the onset of the LGCP, assess the climate variability during the LGCP, the timing of LGM and possibly the timing of Termination I. This is an important site as it is the only site which extends beyond the LGM in the region.

The Howard valley site is located in the Nelson lakes region and is composed of three river terrace risers with exposed sections. These risers have been previously dated and range from c. 34-18 ka (Marra and Thackray, 2010). A 10 cm thick KOT layer is located within cover-beds of the third terrace riser. This site will provide a climatic reconstruction beginning pre-LGCP and will hopefully provide a rough estimate for the onset of the LGCP, the pattern of climate during the LGCP including the LGM. Although a fragmentary record will be produced meaningful interpretations can still be made.

Figure 1.2: South Island New Zealand showing the location of the four sites: A. Lake Mudgie, B. Manks Tarn, C. Howard Valley and D. Forks Lagoon. Inset shows the South Island (boxed) in context of the Southern Hemisphere and the EPICA Dome C core site (EPICA Community Members, 2004).

1.5 Objectives

Palynology, the study of pollen, will be the primary proxy used for palaeoenvironmental reconstructions. The pollen assemblage at a particular time and space relates to regional vegetation which, in turn, is controlled largely by climatic parameters. Therefore, pollen can be used to reconstruct past climates (Birks, 1981). This project will apply the recently developed pollen-based transfer function for New Zealand (Wilmshurst *et al*, 2007) to generate quantitative reconstructions of mean annual temperature at these sites. This method was not previously used in the NZ-I CES but was mentioned in the conclusion as a way to improve the results. Sedimentological analysis, including Loss on Ignition (LOI) and Particle Size Analysis (PSA), will aid the pollen interpretation. Dating of the sequences will rely primarily on radiocarbon dating along with one tephra layer of known age thereby aiding the chronology in three sites.

1.5.1 Principal research objectives

1. Produce high resolution palynological and sedimentological records from four sites of contrasting climates in the South Island, New Zealand (see map, Figure 1.2).
2. Apply the pollen transfer function to reconstruct Mean Annual Temperature (MAT) at the four sites between 30-8 ka.
3. Use multivariate analysis to enable objective comparisons of the pollen sequences from the four sites.
4. Develop reliable chronology for all four stratigraphies thereby enabling accurate determination of the key climatic transitions and millennial scale oscillations in New Zealand. This includes the following:
 - Onset of the LGCP.
 - The LGCP milder interval.
 - Beginning and end of the LGM.
 - Beginning and end of the LGR.
 - Onset of the Holocene.

5. Determine the pattern and extent of the climatic changes during the LGCP and LGIT for the South Island of New Zealand and assess whether any regional variation exists in these interpretations.
6. Assess the usefulness of the NZ-I CES at explaining the timing and pattern of South Island climate change between 30-8 ka.

This project will therefore provide further evidence and increase the reliability of the timing, magnitude and pattern of climate change in New Zealand by testing the hypothesised model created by the NZ-I group. By increasing the spatial resolution, the robustness of the scheme can be tested, with the data adding to the model to produce a more refined stratigraphy of New Zealand as a whole. An increase in temporal resolution, and records spanning beyond the LGCP will enable both the LGCP and LGR to be specifically targeted. With the known events targeted, more detailed dating can occur to produce a reliable chronology. The project will also expand the use of the existing New Zealand quantitative pollen-climate reconstruction models by attempting to use this methodology on samples predating 18 ka, the limit of use in the Wilmhurst *et al*, (2007) study. The use of this method was advised by the NZ-I group as a way to progress the provisional scheme to provide a more objective statistical interpretation of the pollen results.

Regarding the wider context, it would be unrealistic to produce conclusions on the causation and timing of global climate changes for the past 30 ka based upon the results of this thesis. However, the results will add validity and argument to this debated topic and highlight any connections which exist between New Zealand and Antarctica by assessing and comparing the timing and pattern exhibited in both localities. It may also provide insights into the possible forcing factors of climate change in the Southern Hemisphere. These points will be highlighted later in the discussion chapter (Chapter 8).

Chapter 2: Background

The previous chapter provides an introduction to the project by introducing late Quaternary climate change concepts and patterns both globally and in New Zealand, identifying current gaps and uncertainties in previous research and provides the aim, objectives and justification for the project. This chapter follows on from the previous chapter by providing geographical, geological and environmental context for the project. This chapter also provides a review of previous late-Quaternary research from New Zealand and the surrounding oceans, which are relevant to the discussion sections in Chapters 4-8.

This chapter begins with a geological background to New Zealand which has primary and secondary influences on the contemporary climate and vegetation of New Zealand. A summary of the late-Cenozoic history of New Zealand will follow, providing an understanding of the terrestrial, marine and climatic changes experienced in New Zealand. This leads onto the contemporary climate. An understanding of New Zealand's vegetative communities and ecological restriction is an important aspect for this project as ultimately the pollen record will be used to produce vegetation and inferred climatic interpretations at the four sites. Therefore, a section on the contemporary vegetation summarising the different floristic communities and factors that impede or promote certain species growth is discussed. An emphasis is placed on South Island communities, particularly those that are relevant for this project. Likewise, an understanding of past vegetative response to climatic change is also necessary and therefore a section on late-Cenozoic vegetation follows the contemporary vegetation section. The final section of the chapter provides a summary of established South Island proxy records. This section is split into the three regions that are studied in this project and provide an overview of previous research from each region. The limitations of the available research at each site are reviewed. A summary section concludes this chapter.

2.1 Geological context of New Zealand

The geological processes, both in the past and ongoing, experienced in New Zealand are significant to palaeoenvironmental reconstructions. The isolation of the New Zealand continent from other large landmasses has had a significant influence to its biogeography. Likewise, tectonic activity has controlled the morphology of New Zealand and significantly influences climate and vegetation patterns. The geological processes in New Zealand also directly influence the stratigraphy of the palaeorecords. Therefore it is important to summarise the geological history and the ongoing processes in New Zealand.

New Zealand is a relatively young archipelago dating from the Mesozoic onwards. Prior to 80 Ma New Zealand was part of the super continent of Gondwana, formed by the continents of Australia, Antarctica, Africa and South America. During the Cretaceous period (60-80 Ma) New Zealand separated and drifted away from Australia and Antarctica (Molnar *et al*, 1975) and subsequently became isolated from both continents by the Tasman Sea, spanning nearly 2000 km (Pole, 1994). During the Oligocene the New Zealand continental plateau thinned reaching a thickness of 17-25 km thick (Adams, 1962). Continued thinning and progressive submergence of the landmass occurred between the Cretaceous and Oligocene (Waters and Craw, 2006). The geological nature of New Zealand has been dominated and controlled for the last 30 Ma by the Pacific and Indian/Australian plate boundary upon which it lies. Movement along what is known today as the Alpine Fault began around 30-25 Ma and continued through the Miocene. Around 10 Ma extension and thinning of the continent reduced and compression began along the alpine fault. During the Pleistocene (past 2.59 Ma) the compression significantly increased causing rapid uplift and mountain building in New Zealand (Carter and Norris, 1976).

The period of rapid mountain building coincides with a period of large climate variability during the late Pliocene and Pleistocene. The climatic variability is characterised by a succession of glacial/interglacial cycles (refer to Section 2.2.1). The large climate variability also induced glacio/eustatic sea-level changes. In turn, this affected the base level, which, combined with extensive glacial activity

and large precipitation values, has resulted in high levels of erosion and weathering of the landscape. This activity has produced a youthful and heavily dissected landscape containing large Quaternary aged deposits and erosional features (see Sections 2.2.1 and 2.6). Tectonic activity continues to uplift the mountains and is responsible for the high relief of the country. Over 75% of New Zealand lies $\geq 250\text{m}$ (McGlone, 1988; Soons and Selby, 1992).

Today only around 10% of the New Zealand continent is emergent (Waters and Crow, 2006) forming a collection of small temperate islands spanning the mid-latitudes ($34\text{--}47^\circ \text{S}$) in the Pacific Ocean. It is a long (approximately 1500 km) thin, north-east trending archipelago. New Zealand still straddles the Pacific and Indian/Australian Plate boundaries (Figure 2.1), forming a convergence zone causing both uplift and subduction throughout the country. The Alpine fault marks the boundary between the two plates and this is trended in a north-east – south-west direction throughout the country (Walcott, 1984; Figure 2.1). Immediately to the east of this fault lies the Southern Alps mountain range in the South Island, formed by the Pacific Plate rising over the Indian/Australian Plate (Adams, 1980). The Southern Alps dominate the South Island and have a profound influence on the climate (see Section 2.3).

Whilst the South Island is dominated by tectonic uplift and mountain building, the North Island is characterised by its recent volcanic activity which, during the Quaternary, has been an important geomorphic influence. Three major types of volcanism is experience in New Zealand and are classified as basaltic, andesitic and rhyolitic. These three types of volcanic activity act in different ways which influences the type, extent and chemical composition of a volcanic deposit (Houghton *et al*, 1995). The Taupo Volcanic Zone (TVZ) in central North Island spans an area 300 km long and 60 km wide (Figure 2.1) and is caused by the subduction of the Pacific Plate under the North Island. Earliest activity in this zone was andesitic beginning around 2 Ma. By 1.6 Ma rhyolitic activity became dominant (Wilson *et al*, 1995, Carter *et al*, 2003; Alloway *et al*, 2005) with the TVZ being the most active and productive rhyolitic magmatic system in the world (Houghton *et al*, 1995). Since volcanic activity began in the area approximately 15-20,000 km³ of volcanoclastic deposits has been produced (Wilson *et al*, 1995). This

material is dominated by gas- and water-supported mass flow deposits as well as airfall tephra which are widely distributed across the landscape and can also form secondary redeposited volcanoclastic materials. These tephra layers have become important stratigraphic and chronological markers which are useful for this project (refer to Section 3.5.2 for more information).

Figure 2.1: A simplified location map of the major tectonic activity in New Zealand. Taupo Volcanic Zone (TVZ) location is based upon Wilson *et al* (1995). Plate boundaries and rates of movement are from DeMets *et al* (1994) and Beavan *et al* (2001).

2.2 Late Cenozoic history of New Zealand

2.2.1 Glacial activity

2.2.1.1 Quaternary glaciations

There are over 3000 glaciers exceeding 0.01 km² in mainland New Zealand today containing a total ice volume of 55.3 km³ (Chinn, 1989). The largest of these is the Tasman Glacier extending 29 km from the main divide down the eastern side of the Southern Alps (Porter, 1975a). However, these existing glaciers are only small remnants of the extensive ice cover experienced during the last glacial. The sediment record from Wanganui Basin shows at least 58 unconformity bounded glacial and interglacial couplets on a 41 ka or 100 ka timescale spanning back 3.6 Ma (Naish and Kamp, 1997; Naish *et al*, 1998). In the terrestrial realm a fragmented and incomplete record of glacial/interglacial activity is preserved (Suggate, 1990). The most complete records are found in north Westland, South Island. Here, a pattern of glacial outwash aggradation surfaces, interbedded with coastal sediments that are interpreted as interglacial high sea-level high events (Suggate, 1965; 1990).

The oldest glacial deposits, found by Gage (1945) in north west South Island, are assigned to the Ross glaciation occurring approximately 2.62.4 Ma. Porika glacial occurred after this between ~2.2-2.1 Ma (Suggate, 1990). A combination of erosion and regional uplift produce a gap in the record until 0.35 Ma, after which evidence of a further four glacial are identified. These are known in New Zealand as the Nemona, Waimaunga, Waimea, and, the most recent glacial period, the Otira (Suggate, 1990; see Table 2.1). The marine and terrestrial proxy records can be correlated and appear to show good correspondence with the geomorphic evidence including vegetation change associated with warm and cold periods, and the $\delta^{18}\text{O}$ marine isotope records (e.g. Mildenhall *et al*, 2004; see Section 2.2.2.3).

During the glacial periods the ice volume increased significantly in the South Island. The glacial complex in the Southern Alps widened by 100 km whilst the valley glaciers extended into the lowland areas forming piedmont ice lobes reaching up to 50 km wide in the Westland region (Porter, 1975b). Associated with the glacial advances is an increase in sediment source, transport and deposition

leading to large gravel deposits and extensive outwash aggradation surfaces. These fluvioglacial derived material can be traced up valley to moraines particularly for younger glacial events. When excluding the Quaternary volcanic deposits around 10% of New Zealand is covered by glacial deposits with a concentration on the lowland areas, particularly in the South Island (Suggate, 1965). In contrast, the glacials had limited geological impact on the North Island where only Mt Ruapehu (McArthur and Shepherd, 1990) and the Tararua ranges (Brook and Brock, 2003) reach sufficient altitudes for permanent snow or ice accumulations. Permafrost did not occur in the North Island but periglacial activity, fluvial and wind erosion had notable impact on the geomorphology (Soons and Selby, 1992).

MIS	Glacial (New Zealand)	Interglacial (New Zealand)	Westland formations	Mackenzie Basin formations	Upper Buller formations
1		Aranui		Birch Hill	
2	Otira		Moana	Tekapo	
3			Larrikins	Mt John	Speargrass
4			Loopline	Balmoral 1	Roundall
5		Kaihinu			
6	Waimea		Waimea	Balmoral 2	Tophouse
7		Karoro			
8	Waimaunga		Tansey	Wolds	Harry

Table 2.1: Names and timing of the glacial and interglacial periods from terrestrial evidence in New Zealand related to the Marine Isotope Stages (MIS) (Suggate, 1990; Challis *et al*, 1994; Suggate and Almond 2005).

The last glacial period (the Otira Glacial) spans 70-12 ka and includes the Marine Isotope Stages (MIS) 4, 3, and 2 (Soons, 1979). The period of maximum ice extent occurred between 28-18 ka and is referred to as the LGCP by the NZ-I CES (Alloway *et al*, 2007; see also Section 1.2). This intensive cold period is followed by rapid deglaciation and climatic amelioration in New Zealand. During the LGCP glaciers extended to within 2 km onshore from the current coastline, and in some instances extended beyond the coastline e.g. between Hokitika and Greymouth (Suggate, 1990; Suggate and Almond, 2005). During these cold periods and lower sea level outwash material was deposited far offshore (Suggate, 1990). Moraine, loess and other proxy records indicate the LGCP was climatically complex (Almond *et al*, 2001; Alloway *et al*, 2007). Suggate and Almond (2005) find evidence for 3 periods of ice advance occurring between 34-28 ka, 24-21.5 ka and 20.5-19 ka, with milder condition interrupting these cold periods. The period 21-18 ka, the LGM *sensu-stricto*, marks the period of coldest climate conditions and most extensive period of ice advance (Alloway *et al*, 2007). The moraine records suggest glacial activity remained complex throughout the deglaciation with a number of halts and minor readvancements (Soons and Selby, 1992).

2.2.2 Oceanic circulation and marine records

New Zealand is an oceanic Island surrounded on all sides by water. Therefore, oceanic water masses, fronts and circulations influence New Zealand's contemporary environment. Past changes in latitudinal location of the water masses and fronts and the intensity of the circulations have been shown to impact upon New Zealand climate. Therefore an understanding of these oceanic factors influences on New Zealand is important for climatic reconstructions. Marine sediments also produce valuable proxy reconstructions which can be linked to and complement terrestrial records. The following section provides a brief overview of ocean water masses and fronts in New Zealand followed by a review of the marine records. An emphasis will be placed upon late Quaternary records, particularly the last glacial/interglacial period, MIS 4-1.

Figure 2.2: The current location of the different surface water masses and their bounding oceanic fronts from Carter (2001).

2.2.2.1 Current and past oceanic fronts and water masses

New Zealand straddles two water masses, the warm Sub-Tropical Surface Waters (STW) and the Sub-Antarctic Surface Waters (SAW; see Figure 2.2). These water masses are separated by the 150-400 km wide, globally circulating, Sub-Tropical Front (STF), which bends around the base of the South Island flowing northward and then eastwards along the crest of the Chatham Rise (Carter, 2001). The Tasman Front, marking the boundary between the warm Coral Sea and cold Tasman Sea, flows from Australia to northern North Island bringing warm water from the sub-tropical region to New Zealand. A continuation of this front, the East Auckland Current, flows south along the east coast of New Zealand forming large

anticyclonic eddies by the East Cape (Tomczak and Godfrey, 1994). South of both New Zealand and the STF is the globally circulating Antarctic Circumpolar Current (ACC) at 52° S. This current is disrupted by the Campbell Plateau causing large eddies of cold water to flow northwards along the plateau margin until c. 48° S where the current continues to flow eastwards (Carter, 2001).

The development of New Zealand's oceanic fronts to current conditions occurred over a long time period. Between 65-35 Ma (Paleocene and Eocene) the New Zealand sector of the Southern Ocean (NZSSO) was mostly influence by sub-tropical waters. No detectable oceanic fronts are evident in the marine cores during this period. At 35 Ma the SAW and subtropical waters are separated by a proto-STF at ~65 °S. During the Oligocene the Antarctic polar front develops, pushing the STF northwards and marking a period of global cooling. At ~31 Ma the Tasman gateway fully opens and the proto AACP front develops. The Antarctic waters flowed northwards onto the New Zealand plateau. Since the opening of the Tasman gateway and three part division of the Antarctic, subantarctic and subtropical waters have persisted in the NZSSO. Between 21-15 Ma, the STW shifted southwards into the northern NZSSO area possibly associated with reduced ice volume or the redirection of warm surface flows southwards into the Tasman Sea from higher latitudes. From 15-5 Ma the STF and SAF became established into their relative contemporary positions (Nelson and Cooke, 2001)

During the climatically variable Quaternary period relative locations of the STF and SAF were probably in similar positions during interglacials as they are today. However, little is known about the movements of these fronts during glacial periods (Nelson and Cook, 2001). Evidence from ocean core sites around New Zealand show a considerable, but temporary, northward shift (5-10 °) of the ACC and STF during glacial episodes (Nelson *et al*, 1993; Weaver *et al*, 1998). The compression of SAW by this shift would have likely intensified oceanic flows which contributed to the increased westerly winds or vice versa and added to the harsh conditions experienced during these cold periods (Nelson *et al*, 1993). Carter (2001) suggests there was a northward migration of the ocean fronts and wind belts during the LGM. The Tasman Front migrated 600 km northward reducing the warm water transport to northern North Island and possibly aiding the temperature

depression on land. The STF also moved northwards in the Tasman Sea, but a northward shift east of the South Island is impeded by the Chatham Rise. During the LGM the ACC also strengthened increasing the amount of cold water eddying northward reaching Chatham Rise and flowing into the STF (Carter, 2001). During this period it is believed the colder waters allowed northward drift of icebergs as far north as the Chatham Rise (Carter *et al*, 2002; Crundwell *et al*, 2008). The northward movement of colder water would have helped maintain low temperatures or reduced them further during the LGM (Carter, 2001).

2.2.2.2 Marine records

The longest “marine” record in New Zealand, and one of the most complete late Neogene records in the world, is from the Wanganui Basin and extends over the past 3.6 Ma, spanning MIS 1 to MG6. Wanganui Basin is a 200 km² sedimentary basin location in western North Island. Tectonic activity has caused in upwarping of the basin’s eastern margin resulting in onland exposures of the marine sediments spanning 58 cyclothem. Each 40 ka (MIS 100-18) and 100 ka (MG6-G2 and MIS 17-2) cyclothem consists of a glacial/interglacial couplet. The sediment facies represent deposition during highstands and falling caused by glacio-eustatic sea-level change. This cyclothem record illustrates the New Zealand experience effects caused by major climate and sea-level changes on orbital timescales since 3.6 (Naish and Kamp, 1997; Naish *et al*, 1998). Other temporally shorter marine records from both the eastern and western sides of New Zealand also show a Milankovitch scale climatic change (Pahnke *et al*, 2003; Pahnke and Zahn, 2005; Pelejero *et al*, 2006; Barrows *et al*, 2007; Elderfield *et al*, 2010).

A 280 ka long alkenone sea surface temperature (SST) record from marine core SO136-GC3, from Challenger Rise, western New Zealand reveal a 7 °C temperature amplitude between glacial and interglacial periods. A maximum temperature of 19 °C and 19.5 °C lasting 5 ka long was recorded for MIS 7e and 5e respectively. The temperature subsequently reduced to 14.5 °C and 15.4 °C respectively. Holocene temperatures only reached a maximum 16.9 °C, significantly lower than the previous interglacials and averaged 3 °C lower than MIS 5e (Barrows *et al*, 2007). A similar pattern is recorded in marine records from

eastern New Zealand. A benthic Mg/Ca palaeotemperature reconstruction and $\delta^{18}\text{O}$ record from Ocean Drilling Project (ODP) site 1123, Chatham Rise, reconstruct higher temperatures during the interglacials MIS 11, 9, 7 and 5e compared to the Holocene (Crundwell *et al*, 2008; Elderfield *et al*, 2010). Pahnke *et al* (2003) and Pahnke and Zahn (2005) MD97-2120 marine core record, also from the Chatham Rise, again, records higher temperatures for MIS 9c, 7e, 5e than the Holocene, although a smaller range in temperature from glacial to interglacial periods compared to the SO136-GC3 (Barrows *et al*, 2007). The pattern of SST in SO136-GC3 during the glacial show a progressively colder minimum temperature with age whereby MIS 8, 6, 4, and 2 experienced a minimum temperature of 11.9 °C, 11.6 °C, 11.3 °C and 9.8 °C respectively. Also prior to maximum cooling each glacial experiences a small increase in SST (Pelejero *et al*, 2006). This occurs also in eastern New Zealand with The $\delta^{18}\text{O}$ curve from ODP 1123 core becoming more positive, suggesting a slight warming, prior to maximum cooling (Elderfield *et al*, 2010).

Barrows *et al* (2007) SST record from SO163-GC3 using planktonic foraminifera found millennial scale climate scale fluctuations superimposed on the larger Milankovitch scale climate transitions. For instance during MIS 3 warming periods were recorded every 6-7 ka between 57.8-38.9 ka. The timing of these warmer intervals appeared to lead the equivalent Dansgaard-Oeschger (D-O) events recorded in the GISP2 ice core record suggesting an asynchronous pattern of climate change between the hemispheres on millennial timescales. Another noticeable discrepancy between the Northern Hemisphere ice core records and SST record of SO136-GC3 is the pattern and timing of deglaciation at the end of the LGM. SST reconstructions show rapid increases in temperature marking the end of the LGM with the majority of warming occurring by 15 ka. This precedes the Bølling Allerød warming chronozone, identified in the GISP2 ice core record, by 3-4 ka. The pattern of warming is not continuous but contains a series of minor reversals. After 15 ka there does not appear to be a strong cooling trend and it is difficult to distinguish an ACR cooling event. However, the period marked by the Younger Dryas Chronozone in the Northern Hemisphere is a period of warming in the eastern Tasman Sea, and again precedes a Northern Hemisphere warming.

The core MD97-2120 lies south of the Inter-Tropical Convergence Zone (ITCZ) which separates the warm, saline rich subtropical surface waters and the fresher, cold sub-Antarctic water, a perfect location to assess movement in ocean circulations as well as temperature change. Sea Surface Salinity (SSS) measurements have also been calculated with the results providing insight to oceanic changes coinciding with climate change. SSS decreases during glacial and stadials and increases during interglacial. Pahnke *et al* (2003) noted that terrestrial records, such as vegetation reconstructions, indicate increased aridity during glacials. This would indicate less fresh water and higher evaporation rates and would therefore be assumed to cause an increase in ocean salinity. However, the opposite occurs and suggests a northward migration of fresher, colder sub-Antarctic surface waters during glacials causing the lower SSS. MD97-2120 also uniquely records a 1.1 °C cooling coinciding with an ACR timing of cooling (Pahnke and Zahn, 2005), which is not recorded in the other marine records.

2.2.2.3 Marine records containing dominantly terrestrially derived material

Sediment deposited in the marine realm in part can be derived from and contain terrestrial material providing an opportunity to link marine records and terrestrial signals of climate change. The oldest palynomorph record is from ODP1123 and extends back 5 Ma (Mildenhall, 2003; Mildenhall *et al*, 2004). Chronology for this core is constrained using benthic foraminifera $\delta^{18}\text{O}$ records produced by Hall *et al* (2001). Like the records already mentioned in the previous section a clear Milankovitch scale climate change is recorded in the core. The pattern of glacial/interglacial cycles changes from low amplitude, high frequency 40 ka obliquity cycles to saw toothed high amplitude, low frequency 100 ka eccentricity cycles between 0.92-0.62 Ma (Raymo, 1997).

Marine core DSDP 594 from Chatham Rise produces a 750 ka long silt, carbonate and $\delta^{18}\text{O}$ record (Nelson *et al*, 1985, 1986) and a 350 ka long pollen record (Heusser and Van de Geer, 1994). The following patterns from these records are observed. Pollen assemblages during the interglacials (MIS 1, 3, 5, 7 and 9) are dominated by arboreal pollen, like the palynomorph records from ODP 1123 (Mildenhall, 2003; Mildenhall *et al*, 2004). This vegetation pattern corresponds with an increase in pollen concentrations and silt minima suggesting a large vegetation

cover and low amounts of erosion. The pollen assemblage suggests MIS 7 was the coldest interglacial period caused by the dominance of sub-montane taxa, whilst MIS 1 and 5 were the warmest caused by the large abundance of temperate rainforest taxa. Conversely the glacials (MIS 2, 4, 6, 8, and 10) were dominated by herb and grassland taxa. Evidence of forest refugia, contributing up to 20% of the pollen assemblage, was recorded during all glacial periods. Unlike the interglacial periods the pollen assemblage remained the same suggesting similar conditions between glacial periods. During the glacials pollen concentration decreased significantly whilst silt abundance increased indicative of reduced vegetation cover and increased erosion (Heusser and Van de Geer, 1994).

On shorter timescales Wright *et al* (1995) and McGlone (2001a) produced a marine palynomorph record, offshore from the North Island, extending 59 ka and 30.7 ka respectively. From 59-46 ka conifer-hardwood forest dominated the North Island indicating warm interglacial conditions prior to the glacial maximum. Abrupt and significant cooling occurred between 46-28.8 ka, characterised by increased $\delta^{18}\text{O}$ and pronounced vegetation change to a grass, herb and shrub dominated vegetation community (Wright, *et al*, 1995). Coinciding with the vegetation change was an increase in aeolian derived quartz deposits indicative of windy glacial conditions (Stewart and Neall, 1984). Temperatures reach a minimum between 22.8-18.3 ka, representing the last glacial maximum. This period had the lowest representation of arboreal pollen in both records (Wright *et al*, 1995; McGlone, 2001a), however scattered patches of forest refugia are evident. The end of the glaciation is marked by an abrupt transition from full glacial open vegetation landscape to a dominance of podocarp/hardwood forest (McGlone, 2001a) and a decline in quartz, suggesting a reduced strength or a shift in westerlies (Stewart and Neall, 1984).

2.2.2.4 Summary

The proxy terrestrial evidence in the marine record contains a number of glacial/interglacial cycles occurring throughout the past 5 Ma shifting from a dominant 40 ka cycle to 100 ka cycle. These cycles corresponds to Milankovitch timescale and synchronously with the Northern Hemisphere. However, on a millennial timescale there are disparities evident between the hemispheres. The

marine records from both the eastern and western sides of New Zealand record a similar pattern of climate change and all recognise the Holocene to be colder than the previous interglacials. A northward migration of the Antarctic cold water during the glacials is also evident from the SSS record.

2.3 Climate of New Zealand

New Zealand is located in the south-eastern Pacific Ocean and lying within the Southern Hemisphere temperate zone. New Zealand also lies in the path of the Southern Hemisphere prevailing westerlies (Salinger and Mullan, 1999). The weather is dominantly controlled by the progression of eastward migrating anticyclones, depressions and fronts (Fowler *et al*, 2008). However, blocking of the eastward migration does occur generally during the winter months. New Zealand is an oceanic island and regional weather and climate variations are largely influenced by the humid, moisture laden maritime air masses flowing from the sub-tropical and sub-Antarctic latitudes (Salinger and Mullan, 1999).

Climate is largely influenced by New Zealand's topography. The north-east-south-west trending mountain ranges run the length of New Zealand and have been found, in both islands, to have had a profound effect on regional climate by forming a barrier obstructing the westerly circulations (Maunder, 1971). This produces a marked contrast in precipitation between the western and eastern side of the ranges (Close *et al*, 1978; Griffiths and McSaveney, 1983; Figure 2.3). Therefore, regional-scale climate, particularly precipitation, is extremely sensitive to changes in the direction of the prevailing wind and variations in circulation strength (Lorrey *et al*, 2007). During periods of strong westerly flow, rainfall increases in the west and south coast of the South Island and central North Island whilst decreasing on the east coast of the North Island (Salinger 1980a; Salinger 1980b). If these westerlies are sufficiently strong, föhn winds are experienced in the Canterbury region with associated high temperatures and low, if any rainfall (Maunder, 1971). Conversely, during periods of low westerly or high easterly flows, these same areas experience lower and high rainfalls respectively (Salinger 1980a; Salinger 1980b). The North Island axial mountain ranges receive the majority of rainfall during northerly or north-easterly cyclonic storms (Soons and

Selby, 1992). The west to east precipitation gradient is accompanied by a north to south latitudinal gradient in temperature (Close *et al*, 1978; Figure 2.3). New Zealand's climate is also sensitive to global climate phenomena that originate outside the mid southern latitudes. This includes teleconnections to the El Niño Southern Oscillation (ENSO) and Inter-decadal Pacific Oscillation (IPO) (Lorrey *et al*, 2007).

Kidson (2000) split New Zealand into eight rainfall regions and three temperature regions, resultants of the orographic effect. He also recognised 12 synoptic circulation types which have been divided into three basic regimes: trough, blocking and zonal types (see Figure 2.4). Trough regimes are dominated by frequent troughs moving over New Zealand. During these regimes precipitation and cloudiness increases in all regions and climates tend to be cooler. Blocking regimes are dominated by highs over the South Island which block the westerly wind flow. During these regimes New Zealand receives an increased easterly flow resulting in increased precipitation in east and north North Island and drier conditions in western regions and east South Island (ESI). Finally, zonal regimes are characterised by high pressure located over northern New Zealand and increase westerly wind flow. This is accompanied by increased precipitation in western South Island and drier conditions over the majority of the North Island and in the north and eastern South Island (Kidson, 2000; Lorrey *et al*, 2007; see Figure 2.4). Lorrey *et al* (2007) took this idea further and tested the hypothesis that regional climate regimes identified in the contemporary environment would have been similar pattern in the palaeoenvironment. Their research indicates that long-term spatial climate signals could be identified in the palaeorecords. This idea implies that palaeoclimate records may also contain regional differences due to changing wind directions and circulation strength. This is an important point as it can help explain 1. the processes involved in New Zealand's palaeoclimates and 2. why there are many disparities between records from different regions. Regional climate regimes in the South Island were an important factor for site selection in this study, but this will also be important when interpreting the palaeoenvironmental records and its relevance.

Figure 2.3: Mean Annual Temperature (MAT) and precipitation (MAP) maps of the South Island (adapted from NIWA, 2001).

Regional weather
regime classification
of Kidson (2000)

Regional
precipitation
anomaly

Figure 2.4: The different rainfall anomalies during different climate regimes for each rainfall region taken from Lorrey *et al* (2007, pp 413).

2.4 New Zealand vegetation

Before accurate vegetation, environmental and climatic reconstructions can be inferred from palaeopalynological records an understanding of vegetation in its contemporary environment is necessary. This includes; the different types of vegetation communities, plant assemblage in each community type, the way species interact with each other, specific niches and environmental limitations acting upon different species, altitudinal and latitudinal limits of species and the climatic factors which individual species and communities can tolerate. This section provides a summary of the points above for New Zealand's plant communities which will be used to reconstruct changing vegetation and infer climate. The section concentrates on the regions and vegetation communities that are representative of the plant assemblages in the pollen records and relevant to this project and excludes the North Island. The following is based primarily on Wardle (1991) and McGlone (1988) and the reader is directed to these sources for more information. Each site chapter contains a detailed vegetation description for the region (see Chapters 4-7 for more detail).

2.4.1 Forest communities; lowland-treeline

Since human occupation, and in particular since European colonisation, New Zealand's forest cover has been severely reduced. This is especially the case in the lowland and easily accessible locations where only small, vulnerable remnants remain. Extensive native forest cover today is restricted to the remote, rugged areas. Only on the west coast of the South Island can unbroken forest exist from sea level to alpine tree line limit (Wardle, 1991). The remaining native forests can be split into two sub-categories: conifer-broadleaved forest and *Nothofagus* (beech) forest.

2.4.1.1 Conifer/hardwood forest

These forests are highly diverse and species rich and occur throughout the whole of New Zealand and in all types of climate, although the complexity of a community can vary greatly with altitude, temperature and soil type/quality. This forest type includes the podocarp conifer species mostly belonging to the family *Podocarpaceae* and a selection of hardwood angiosperm (flowering plants; Godley, 1975). These forests are composed of a number of different vegetative tiers caused by the maximum growth limitation of each species. The tallest tier is formed from conifer species where the canopy ranges from being continuous to completely absent. The second tier is characterised by the shorter broadleaved trees with the density of cover varying greatly. A further tree tier is formed by the smaller trees and associated tree ferns. This can form the dominant canopy cover in localised regions but is suppressed if the canopy above is continuous. Beneath the tree cover, tall herb species are located with the ground level comprising mostly ferns, mosses and liverworts, although small angiosperms can also be present (Wardle, 1991).

Elevation has a significant influence on the conifer/hardwood forest community composition with clear elevation zones evident in the Westland region. The lowland forest (0-400 m.a.s.l) is dominated by emergent podocarp species, including *Dacrydium cupressinum* and *Prumnopitys*. Hardwood taxa from the sub-canopy and shrubs (*Coprosma*, *Pseudopanax* and *Myrsine*) produce a diverse sub-canopy. Tree ferns are also an important component of the sub-canopy whilst ground ferns cover the forest floor. In waterlogged/wetland areas *Leptospermum*

and *Halocarpus* shrub tier grows. Between 400-800 m.a.s.l a montane hardwood forest dominates the vegetation community and many lowland podocarp taxa have limited growth. The sub-canopy is made up of low lying broadleaved trees, shrubs and tree ferns (Wardle, 1991). At 800-900 m.a.s.l the vegetation becomes dominated by sub-alpine shrubs and low forest up until 1200 m.a.s.l, marking the end of the treeline (Wardle, 1979). The distinct elevation zones are useful when estimating elevational vegetation depression to infer temperature depression using Environmental Lapse Rates (ELR) (see Section 3.2.6 for more detail).

2.4.1.2 Beech forest

New Zealand contains four species of beech: *Nothofagus fusca*, *N. menziesii*, *N. truncata*, and *N. solandri* (includes the var. *solandri* found in the lowland-montane elevations, and var. *cliffortioides* inhabiting the upland to sub-alpine zone) (McGlone, 1988; Wardle 1991). Ecologically, *Nothofagus* spp tend to flourish in regions climatically less favourable for other plant growth when compared to areas dominated by conifer-broadleaved forests (McGlone, 1988). Therefore, they are often located in the upland areas and form the majority of timberlines. Beech forests can extend from the sub-alpine zone to low altitudinal areas along river banks. They tend to inhabit recently disturbed areas where other species have not had the chance to dominate. With reduced altitude they become increasingly confined to shallow soils found on crests, ridges and spurs. Beech forest can mix with the conifer-broadleaved species although this tends to occur in the cool-temperate belt in areas with high-moderate rainfall. There are unexplained gaps in the beech distribution in some regions where the conditions are still favourable for beech growth. These major gaps include areas from Mt Taranaki, some areas of the Volcanic Plateau and Westland between the latitudes 42° 40' S and 43° 30' S, and on Stewart Island (Wardle, 1991).

In comparison to conifer-broadleaved forest, beech forests structurally tend to be more simplistic. In the most extreme environments the canopy is solely made from *Nothofagus* with few shrubs and a very sparse understory. Each species has its own ecological preference. *N. menziesii* is generally found in wetter areas of south and central New Zealand and forms unbroken cover from sea level to treeline. In contrast, *N. fusca* and *N. truncata* are both less tolerant to colder environments.

As a result *N. fusca* tends to be limited to the warm midslope locations and *N. truncata* is most common north of 38° S. In areas of poorly drained, low fertility sites *N. solandri* var. *cliffortioides* can be common (Wardle, 1991). The Howard Valley is located in the Nelson Lakes region which is dominated by mixed beech forest. A more detailed vegetation description is provided in Section 6.1.

2.4.2 Forest transition, sub-alpine and alpine zone

The character of forest transition into the alpine vegetation zone varies between beech and conifer-broadleaved forests. Beech forests tend to end abruptly with vegetation changing immediately into the tussock grasses. In contrast, the conifer-broadleaved tree line tends to be lower in altitude and has a more gradual transition into alpine vegetation with a highly diverse sub-alpine shrub zone. This zone includes: podocarp shrubs, composite shrubs, small leaved *Coprosma* spp, *Dracophyllum*, *Hebe*, *Myrsine*, *Olearia* *Pseudopanax* and others. This zone rarely exceeds 100 m higher than the tree line.

The alpine zone is dominated by tussock grasses such as *Chionochloa* spp. These grasslands have an abundance of herbs and shrubs including species from Apiaceae (*Aciphylla*, *Anistotome*, *Hydrocotyle*) Epacrideae (*Dracophyllum*, *Cyathodes*, *Leucopogon*), Compositae (*Celmisia*), Thymelaeaceae, *Drapetes*, *Pimelia*, *Coprosma*, *Ephilobium*, *Geranium* and *Hebe* (McGlone 1988). The alpine zone begins at 1200 m.a.s.l in the Westland region (Wardle, 1977), c. 1400 m.a.s.l in Nelson Lake National Park (Wardle, 1991), and at 1500 m.a.s.l at Mt Cook (Wardle, 1991).

2.5 Cenozoic vegetation history of New Zealand: overview

Since separating from the super continent Gondwana over c. 60 Ma the extent and length of New Zealand's isolation has led to a unique floristic community (Cooper and Millener, 1993). The traditional view is that origins of New Zealand's endemic flora are remnants from Gondwana times (e.g. Wardle, 1991). However, trans-oceanic dispersal from Australia, South America and Hawaii is the favoured contemporary view (McGlone *et al*, 2006; McDowell, 2008). The geological history

of New Zealand (see Section 2.1) and past climatic changes have had a large influence on the vegetation composition in the past and therefore on the contemporary vegetation. The following section is a brief overview of the history of New Zealand's vegetation change during the Cenozoic. An emphasis is upon Quaternary vegetation change and the response of vegetation to climate changes during this period. These patterns of change are relevant to the discussion of the vegetation and inferred climate change for each site section in this project (Chapter 4-7) and the main discussion chapter (Chapter 8).

2.5.1 Palaeogene/Neogene

During the early to mid Cenozoic period New Zealand was a low lying archipelago located in the high latitudes 76-68° S. The flora was dominated by temperate and sub-tropical species suggestive of an unseasonal and warm climatic regime (Mildenhall, 1980). The climate was very similar to that experienced by Australia and many floristic species were related (McGlone *et al*, 2006). This warm climate continued through the Miocene (Mildenhall, 1980).

Approximately 5 Ma (late Miocene/Early Pliocene) the westerly circulation increased in intensity, coinciding with major polar ice sheet development in Antarctica (Kennett and Brunner, 1973; McGlone, 1985). During the Miocene/Pliocene, New Zealand moved into its current latitudinal position whilst contracting in length and entering a mountain building phase, which is continuing today (see Section 2.1). It was during this period that the beginnings of the rain shadow effect were felt in areas in the lee of the mountains. These alterations in latitude, altitude, climate and land area resulted in large floristic changes (Mildenhall and Pocknall, 1983), with conditions favouring the herbaceous and easily dispersible species. Consequently, these species increased in abundance (McGlone *et al*, 2006), with shrub and grassland dominating the vegetation (Mildenhall, 1980). Substantial plant extinction, due to climatic deterioration was experienced throughout New Zealand (McGlone, 1985; Lee *et al*, 2001). Warm-temperate flora, often associated with Australia, was most affected. These extinctions continued throughout the Quaternary glaciations (McDowell, 2008).

2.5.2 Quaternary (past 2.58 Ma)

The Quaternary is characterised, around the world, by increased climatic variability than the Cenozoic. The climate was predominately cool which had a large impact on the vegetation. The last tropical/sub-tropical species disappeared from the fossil record by c. 1.7 Ma (Mildenhall, 1980). Full interglacial conditions such as MIS stages 1, 5e, 7a and 7b are characterised by widespread development of conifer-angiosperm forests. By contrast, the glacial episodes experienced an expansion of grassland and shrub taxa. The most detailed and widely studied period of the Quaternary is the most recent glacial / interglacial cycle (Newnham *et al*, 1999) and is described below.

2.5.3 Last glacial period - present (Figure 2.5)

Recent evidence suggests an earlier onset of LGM cooling in New Zealand. Alloway *et al* (2007) renamed this period the LGCP to include the earlier cooling phase beginning at 28 ka with the coldest period, 21-18 ka designated as the LGM. However, evidence from other studies suggest an even earlier onset of the LGCP e.g. 30.5 ka (Vandergoes *et al*, 2005; Newnham *et al*, 2007c). This section will produce an overview of New Zealand vegetation change from 30.5 ka to the present. More detailed vegetation records concentrating on the regions studied in this project will follow in Section 2.6.

The onset of the LGCP is marked by a change from sub-alpine shrub dominated vegetation, and forest floristic element evident at some sites (Wright *et al*, 1995; Moar and Suggate, 1996; Vandergoes *et al*, 2005; Burge and Shulmeister, 2007a; Marra and Thackray, 2010) to an open grass dominated environment with alpine herb taxa (McLea, 1996; Moar and Suggate, 1996; Vandergoes *et al*, 2005; Marra and Thackray, 2010). There are only a limited number of vegetation records that cover the LGCP, with the evidence suggesting the onset, represented by a grass peak, began 30.5-28 ka. This is followed by a period of milder climate recognised in the pollen record by an increase in shrub taxa at the expense of grass. Nevertheless, grassland was still the dominant vegetation throughout the majority of New Zealand during this relatively milder interval (Wright *et al*, 1995; Moar and

Suggate, 1996). A rise in grass abundance, and decrease shrub abundance, marks the LGM around 24-18 ka (Vandergoes *et al* 2005; Newnham *et al*, 2007a).

Pollen evidence for the LGM indicates forest cover throughout the period was sparse, with all trees, except *Nothofagus*, rare in many areas (McGlone, 1988; Sandiford *et al*, 2003). The exception is the Northland region of the North Island which remained fully forested throughout the LGM (Newnham, 1992; McGlone *et al.*, 2010). McGlone *et al* (2010) note a regional pattern in vegetation composition during the LGM splitting New Zealand into four zones. A north to south decline in tree abundance was noted with east and west South Island containing only trace amounts of tree pollen. Conversely, east and west South Island contained the highest abundance of herb taxa. Sub-alpine taxa were characteristic of the majority of New Zealand with grassland and open herb fields dominating the landscape. Shrub species were particularly prolific in coastal areas and much of the North Island (Newnham *et al*, 1999; Horrocks *et al*, 2010; McGlone *et al*, 2010). Moar *et al* (2008) found an isolated site in northern Westland, located near the current coastline which contained persistent shrub and tree taxa throughout the last glacial period. This suggests lowland areas of the Westland region were capable of supporting forest growth during the LGM.

Figure 2.5: Summary diagram of LGM vegetation in New Zealand taken from McGlone *et al* (2010).

The late glacial period was characterised by the rapid retreat of ice from its LGM position. This was accompanied by rapid southward afforestation beginning c. 18 ka in the Waikato Basin (Newnham *et al*, 1989) and in southern South Island around 10-9.5 ka (McGlone, 1988). This pattern does not represent a southward migration in vegetation but rather a stepwise southward amelioration of the climate. The increased temperature induced expansion of local forest refugia that had survived the cold conditions during the LGM. The generalised pattern of vegetation succession is: replacement of grassland by shrubs, colonisation by lowland forest, culminating in the growth of tall conifer/broadleaved or beech forests (McLea., 1996; Sandiford, *et al*, 2003; Vandergoes and Fitzsimons, 2003; McGlone *et al*, 2004; Newnham *et al*, 2007a). This transitional period of climatic amelioration is also known as the late glacial/interglacial transition (LGIT) and this term will be used throughout the rest of the text.

Recent evidence, from several records, indicates an interruption in the LGIT warming trend, although this is contested. Pollen reconstructions containing evidence of a climatic reversal are characterised by a stop or reversal in forest development combined with an increase in alpine shrub and herb species e.g. *Coprosma*, *Phyllocladus* and Poaceae. This is indicative of a cooling event (Vandergoes *et al*, 2005; Turney *et al*, 2003; McGlone *et al*, 2004; Newnham *et al*, 2007a; Vandergoes *et al*, 2008). However, other records spanning the same period show no evidence of a reversal with a continued increase in forest development (Singer *et al*, 1998; McGlone, 1995; Sandiford *et al* 2003; Vandergoes and Fitzsimons, 2003).

The Holocene epoch, c. 11.7 ka onwards (Walker *et al*, 2009), shows a continued vegetation change. The nature of change varied from region to region however all changes relate to an increase in biomass. The North Island experienced a generalised increase in the podocarp, *Dacrydium cupressinum*, which became the dominant species. In Wellington and the South Island, reoccupation by tall forest reached its vegetative climax by 9.5 ka (McGlone, 1988). An exception to this is central Otago which remained unforested until 7.5 ka (McGlone *et al*, 1995). From 7.5 ka up to the time of initial human occupation the vegetative pattern was characterised by the spread of *Nothofagus* (beech) which quickly occupied areas

of high rainfall especially in upland areas. However, beech was unable to colonise some isolated areas (McGlone, 1988, see Section 2.5.2). The driving force for vegetation changes during the early Holocene was consistent climatic amelioration. However, edaphic changes and in particular development of soils became a dominant influence on vegetation change from the mid-Holocene until human occupation of New Zealand (McGlone, 1988).

2.6 South Island proxy records spanning the late Quaternary for selected regions

This section will provide a review of previous research carried out in the three regions studied in this project. These regions include Westland, Nelson and the Mackenzie Basin. For each region the geological and proxy evidence spanning the LGCP to early Holocene will be concentrated upon.

2.6.1 Westland

This section will begin by exploring glacial sediment deposits, including successional outwash aggradational surfaces, loess and terminal moraines. Vegetation and inferred climate reconstructions based on pollen and beetle evidence will then be examined. Reconstructions from marine cores will also be discussed, although refer to Section 2.2.2 for more detail. A summary with points of divergence and/or convergence between the records will conclude this section.

Figure 2.6: Location map of the Westland sites mentioned in the text. The light gray circle is a speleothems record site and black circles are pollen records used to produce an overview of the pattern of vegetation change. The stars represent OSL dating of loess by Preusser *et al* (2005).

2.6.1.1 Glacial deposits

Suggate (1965, 1990) identifies three glacial advances during the last glacial period, each with an associated outwash surface formation (see Table 2.1). Since Suggate (1990) a number of studies in the region have been performed, improving upon the earlier glacial surface chronologies. Luminescence dating of river terraces at Raupo, south of the Grey River valley (see Figure 2.6) dates a sand layer lying directly below an outwash gravel associated with the Larrikins Formation. The sand deposit was dated twice, producing the ages 33.4 ± 4 ka and 31 ± 2 ka. Overtopping the Larrikins formation is a fine-silt layer which was also dated and producing the following ages: 22.8 ± 2.2 ka; 20.6 ± 2.2 ka; 23.5 ± 2.1 ka. These ages bracket glacial derived material suggesting a glacial persisted from

~33.4 ± 4 ka to 20.6 ± 2.2 ka based upon luminescence dating (Preusser *et al*, 2005). A change to fine silt overflow deposits, overlaying the Larrikins Formation, indicates a phase of reduced sediment source and river flow attributed to a period of glacial retreat between the Larrikins glacial event and the beginning of the LGM (Hormes *et al*, 2003). This pattern is supported by loess deposits from two southern Westland loess sheets. The oldest was deposited after 36 ka and before 24 ka, supporting an earlier LGCP onset. The second was deposited after 24 ka but before 16 ka, deposited during the LGM. This indicates two periods of cooling during MIS 3 and 2 (Almond *et al*, 2001). Suggate and Almond (2005) date three periods of glacial advance in the Westland region the first between 34-28 ka, again suggesting an earlier LGCP, 24.5-21.5 ka and at 20.5-19 ka. The geological data points towards significant climate variability throughout the LGCP.

The Waiho Loop moraine is the most famous and controversial late glacial feature in New Zealand. Located 10 km below the terminal face of the Franz Josef glacier, it is a substantial terminal moraine and arguably represents a late glacial ice advance and associated climatic reversal (Denton and Hendy, 1994). Wardle (1978) initially dated wood which was directly overlain by glacial till from Canavan Knob, a *rouche moutonnee* located 1.6 km upstream from Waiho Loop. The wood was aged between 13.5-12.7 cal ka. Further dating by Denton and Hendy (1994) constrained the age to 12.9 ka. The timing of both studies indicates a glacial advance beginning synchronously with the beginning of the Northern Hemisphere YDC (12.9-11.5 ka, Alley, 2000), a period of sustained cooling. However, recent dating by Turney *et al* (2007) from the same site, provides an age of c. 13.1 ka, corresponding to a more Southern Hemispheric, ACR event (12.2-13.8 ka, EPICA Community Members, 2004). Cosmogenic nuclide ages collected from boulders forming the moraine, consistently age the feature 2300 years younger than Denton and Hendy (1994). Barrows *et al* (2007) conclude that this moraine is evidence of gradual glacial retreat opposed to a climate induced glacial advance. This conclusion was supported by a $\delta^{18}\text{O}$ marine record (see Section 2.6.1.4.). Other conflicting arguments for the origins of the Waiho Loop add to the controversial nature of this feature. Recent research has attributed the moraine to have been caused by a rock avalanche inducing an ice advance and subsequently depositing the moraine in the existing location (Tovar *et al*, 2008; Shulmeister *et al*, 2009).

However, modelling by Vacco *et al* (2010) disagrees with the rock avalanche theory above. Their models suggest rock avalanches and the debris covering the glacial does reduce glacial melt, but also causes the glacier to stagnate. Under such conditions, they argue a moraine morphologically similar to the Waiho Loop Moraine could not be formed. The foregoing discussion highlights the importance of clarifying the age and occurrence of a LGR in the Westland region

2.6.1.2 Palynological records

Compared with other regions of the South Island, pollen records from Westland are relatively abundant. Extensive research was initially carried out by Moar in the mid-late twentieth century (e.g. Moar, 1971; Moar and Suggate, 1996). However, records spanning back beyond the LGM are rare, and tend to come from river terrace risers, producing a fragmentary record of vegetation and inferred climate change. The few continuous records analysed have a limited resolution and are poorly dated in comparison to recent pollen records. Nevertheless, this initial work brings important and valuable insight of changing environmental conditions during the LGCP and LGIT. More recent work, e.g. Vandergoes and Fitzsimons (2003); Vandergoes *et al* (2005) and Newnham *et al* (2007a), have expanded upon Moar's earlier work, producing increased resolution and well dated continuous records spanning back to the LGM and even as far back as MIS 6 (Vandergoes *et al*, 2005). A summary of this earlier work combined with the recent studies will be discussed below.

Prior to the last glacial period, vegetation was dominated by shrub and sub-alpine taxa (Moar and Suggate, 1996; Vandergoes *et al*, 2005; Newnham *et al*, 2007a). At Okarito, initial LGM cooling, represented by an increase in grass, peaked between 30.5-28.5 ka. This was followed by a milder interval lasting 2-3 ka and straddling the KOT layer (c. 27 ka). This milder interval is characterised by a limited and temporary expansion of trees and shrubs. A second grass peak representing a return to colder climates and the LGM spanned 21-17.3 ka (Vandergoes *et al*, 2005; Newnham *et al*, 2007a). Lake Mudgie and Grahams terrace sites (Figure 2.6) also indicate full glacial conditions beginning prior to the KOT layer, and continuing until the end of the LGM, characterised by an open grassland dominated landscape (Moar and Suggate, 1996).

Termination I (defined as the end of the LGM and, for this project, aged between 18-17 ka, Alloway *et al*, 2007) is marked by sustained decline of grass accompanied by an abrupt increase in shrubs. The timing of this climatic amelioration occurred between 18.3 ka (Vandergoes and Fitzsimons, 2003) to 17.3 ka (Vandergoes *et al*, 2005), and is characterised by a succession of progressively larger peaks of shrub taxa culminating in podocarp/broadleaved forest with *Dacrydium cupressinum* becoming the dominant species by 11.4 ka. Other sites from across the Westland region record a similar pattern of climate change (e.g. Gillespie Beach Road (Moar, 1973; Vandergoes and Fitzsimons, 2003), Maria's Mire, Nan's Kettle and Lake Mudgie (Moar and Suggate, 1996))

A possible climatic reversal spanning 14-11 ka was recognised in the Okarito pollen record (Vandergoes *et al*, 2005; Newnham *et al*, 2007a). A decline in tree taxa and coeval small increase in sub-alpine shrubs was concluded to represent a small and temporary cooling corresponding to both the ACR and YDC. No climatic reversal is reported from other records in the Westland region (Moar, 1971, 1973; Moar and Suggate, 1996; Vandergoes and Fitzsimons, 2003).

2.6.1.3 Beetles

A limited number of beetle studies have been undertaken in the Westland region (Burge and Shulmeister, 2007a, 2007b) with only one study by Burge and Shulmeister (2007b) representing the LGCP, producing a record spanning 37-21.3 ka from Wilson Lead Road. A pollen record was also taken from the same site (refer to Section 2.6.1.2). Similarly to the pollen evidence (Moar and Suggate, 1996), vegetation cover prior to the LGM, 37-34 ka, was dominated by closed canopy, forest vegetation. A transition to a more open landscape, with tussock grasses and alpine/scrubby taxa occurred after 34 ka and continued until 21.3 ka. However, the dominant signal throughout this period is still closed canopy environment. It is therefore concluded by Burge and Shulmeister (2007b) that the LGM experienced milder conditions than evident in the pollen records.

2.6.1.4 Marine records

There is limited marine core evidence taken from off shore Westland coastline. However, during the LGM the ice extent reached near to or beyond the current shoreline. Glacial outwash would have been deposited in the oceans within close proximity to the glaciers. Therefore, marine cores would provide a sensitive record of glacial activity in the region. Marine record SO163-GH3 (Figure 2.6) has two peaks in the sedimentation rate at 28-27 ka and 18.4 ka, the latter representing the LGM. These pulses of sediment have been attributed to periods of glacial advance and associated increased erosion and sediment load in rivers that source this site. The $\delta^{18}\text{O}$ record indicates Termination I began 19 ka and was characterised by gradual, continuous warming, climaxing at 11.4 ka and showing no evidence of a climatic reversal (Barrows *et al*, 2007). A pollen record from core site TAN0513-14 (Figure 2.6) corresponds closely to the Okarito record with the LGM dominated by grass and alpine herb and shrub species including a dominance of Asteraceae. Termination I is marked but an abrupt transition with a rapid decline in grass around 18 ka. Climate amelioration was consistent and culminated with *Dacrydium* and podocarp dominated forest by c. 11 ka (Ryan *et al*, 2009).

2.6.1.5 Summary

Overall, there is general consensus between the different records in the Westland region. The geological, pollen and marine evidence all suggest three glacial advances were experienced during the LGCP. The first advance began c. 30-28 ka and at 25 ka (proxy record dependent), indicative of an extended LGM. The second occurred 24-18 ka, representing the LGM. Relatively milder conditions were experienced in the period between these two advances and straddle an important chronostratigraphic tie marker, the KOT layer (c. 25.4 ka; refer to Section 3.5.2). However, the limited number of continuous records covering this period prevents accurate conclusions of climate change. Inconsistencies between the beetle evidence, where milder conditions were experienced uniformly throughout this period, compared to the pollen results also adds to the uncertainties of this period. Also, the extent and timing of the LGCP warming interval is not clearly defined.

The Westland records also provide general consensus on the timing of Termination I occurring between 18-17 ka. This is marked by sustained and abrupt warming and associated vegetation response. However, whether or not a LGR occurred in the region is debated. Geological interpretations, pollen evidence and chronology of records do not rule out the possibility of such an event. This period would therefore value from more high resolution, continuous records which are well dated and span back to 30 ka.

2.6.2 Nelson and Marlborough regions of northern South Island

This section provides a review of previous work carried out in the Nelson and northern South Island area with an emphasis upon the inland, south-eastern Nelson region. Figure 2.7 shows the location of the sites mentioned in the text. During the last glacial period Nelson and northern South Island were extensively glaciated. Glacial deposits laid down during this period provide evidence of the ice extent and nature of deglaciation. There are also numerous lakes, tarns and river terrace riser containing peat lenses and located outside the LGM ice extent. The deposits from these non-glacial sites have been utilised for pollen, beetle and other proxy reconstructions related to vegetation and climate change during the LGCP and LGIT. Speleothems, also provide useful information about climate change in the region during this period. All of these proxies will be discussed in this section.

Figure 2.7: Location map of the north South Island sites mentioned in the text and the pollen record locations overviews in section 2.6.2.2. Light gray are speleothem records, medium/light grey are pollen records used to produce an overview of the pattern of vegetation change; medium/dark grey are the beetle records (Howard Valley contains both pollen and beetle records), and black circles locate the moraine exposure age sites.

2.6.2.1 Geological evidence

The Spencer and St Arnaud Mountain ranges, which extend the Southern Alps into this region, were extensively glaciated during the last glacial period. The ice extent is marked by numerous terminal moraines which have been mapped and described by Suggate (1965) (see Table 2.1). The LGM terminal moraines are found at the highest elevations on the eastern side of the ranges. For example at Lake Tennyson the terminal moraine is 1100 m.a.s.l (McCaplin, 1992a; Figure 2.7) and 1200 m.a.s.l in the upper Wairau Valley (McCaplin, 1992b; Figure 2.7). Whereas, on the western side of the ranges, in the Nelson Lakes National Park, the terminal moraines are limited to 600 m.a.s.l and 400 m.a.s.l at lake Rotoiti and Rotorua, respectively (Suggate, 1965; Figure 2.7). Therefore, an east to west lowering of the ice extent was experienced during the LGM and was most likely caused by precipitation differences. In the north-west, the Cobb valley terminal

moraine was restricted to 850 m.a.s.l (Shulmeister *et al*, 2005; Figure 2.7) indicating a latitudinal gradient in LGM ice limit as well.

The morainic evidence suggests rapid deglaciation at the end of the LGM, which started, at the earliest, 20 ka with extensive ice wasting and retreat by 18 ka. This retreat was not constant, but stepwise with standstills and minor advances (Shulmeister *et al*, 2005). McCaplin (1992b) located a moraine which could potentially be evidence of a late glacial readvance in the upper Wairau Valley at the confluence of the Wairua River and Island Gully (~1100 m.a.s.l).

2.6.2.2 Palynological records

LGCP pollen records from south-east Nelson are predominantly derived from peat and other organic units interbedded with fluvio-glacial boulders, gravels and silt deposits in river terrace risers. The pollen evidence from these peat sections is indicative of an open landscape environment, dominated by grass and herb taxa (e.g. Apiaceae, Asteraceae). High abundances of Cyperaceae indicate low energy, marshy environments (Moar, 1971, 1980; McLea, 1996). Continuous pollen records also shared a similar vegetation composition during the LGCP (e.g. Mabin, 1983). Quantities of woody/shrub taxa were low, if at all present, with *Phyllocladus* the most consistently represented shrub from the sites in the region (Moar, 1971, 1980; Mabin, 1983; McLea, 1996). Tree species are represented in some pollen records, with *Nothofagus* representing 10% at Howard Valley (McLea, 1996) and Moar (1980) also finding low percentages of *Nothofagus* from other LGCP sites. Therefore, during the LGCP the vegetation of the region was dominated by grassland with refugia of beech and podocarp/broadleaved trees, indicated by trace amounts of this tree pollen at various sites (see Figure 2.7 for the pollen site locations producing this overview).

Termination I is characterised by a rapid transition from grassland and herbland to shrubland dominated vegetation. McLea (1996) found shrub taxa, particularly *Phyllocladus* and *Halocarpus*, beginning to colonise grassland as early as c. 13.7 ka at Lake Tennyson with shrubs becoming the dominant vegetation by 10.4 ka at Wairau. Overall in south-east Nelson, partial colonisation by shrubs occurred by 11.5 ka and this element became dominant at all sites by 9.3 ka. Thus, the

vegetation was becoming increasingly woody in nature (McLea, 1996). Climate-controlled succession continued with the development of forest. Moar (1971) found beech forest already established at Tophouse by c. 8.7 ka with beech forest expanding from Tophouse up the Wairau Valley (McLea, 1996). McLea's (1996) pollen records indicate that expansion of beech from refuges in Murchison and Moutere was initially slow. However, by 8.9 ka beech forest had developed in the upper Buller River after which the forest expanded quickly and by approximately 6.8 ka beech was well established in the Nelson Lakes region. There was little change in vegetation composition until anthropogenic activity (Moar, 1971; McLea, 1996). In the Cobb valley, north-west Nelson (see Figure 2.7), Singer *et al* (1998) also found a similar transition from a dominant grassland environment during the LGCP, followed by rapid reforestation, starting c. 15.9 ka, by shrubs species and culminating in beech forest. Vegetation records from Dew Lakes, in the Bryant range, north-east Nelson (see Figure 2.7), show shrub taxa, including *Coprosma*, *Myrsine* and *Phyllocladus*, were already the dominant vegetation type by c. 12.5 ka with low abundance of podocarp and beech taxa. By c. 9.5 ka beech forest was dominant and remained so until European clearance (Dodson, 1978).

The evidence above shows close agreement between sites. LGCP was dominated by grassland and herb taxa, and a rapid deglaciation. An increase in shrub species, particularly *Halocarpus* and *Phyllocladus*, followed with vegetation culminating in beech forest. The timing of these events is not completely synchronous across the region with the north-eastern sites becoming fully forested prior to the southern sites. This is probably due to the latitudinal and altitudinal differences between these sites.

2.6.2.3 *Speleothem records*

Hellstrom *et al* (1998) produced $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ speleothem records spanning the past 31 ka from a cave site below Mt Arthur (Figure 2.7). Later work by Williams *et al* (2005) produced another $\delta^{18}\text{O}$ records from cave sites in Paturau (Figure 2.7) and Punakaiki (Figure 2.6) also shown in Figure 2.7, spanning the past 23.4 ka and complementing Hellstrom *et al* (1998). The results from both are reviewed below.

During the LGCP three negative $\delta^{18}\text{O}$ excursions are recognisable and occurred between 27-25 ka, 24.5-22 ka and reaching minima between 22-18 ka. Each of these excursions can be associated with a glacial advance providing evidence of extended cold conditions. Altitudinal lowering of the treeline by 600-700 m was experienced at Mt Arthur during this period (Hellstrom *et al*, 1998). Both records indicate the coldest part of the LGCP. The LGM occurred between 23.4-18 ka. Rapid climatic amelioration marks the beginning of the deglaciation at 18 ka (Williams *et al*, 2005), with decreased $\delta^{13}\text{C}$, indicative of increased plant productivity and rapid reforestation in the region (Hellstrom *et al*, 1998). Another negative $\delta^{18}\text{O}$ excursion, representing a cooling period, interrupted the LGIT warming trend between 13.8-11.1 ka. This cooling event spans the later part of the ACR and the YDC and therefore cannot be attributed to just one of these events (Hellstrom *et al*, 1998; Williams *et al*, 2005). There was not a similar response by the $\delta^{13}\text{C}$ possibly due to a limited, if any, vegetation response (Hellstrom *et al*, 1998).

2.6.2.4 Beetle records

A fragmentary, river terrace record from Howard Valley, south-east Nelson (Figure 2.7), provides a beetle and macrofossil record spanning the LGCP (34-18.5 ka). *Nothofagus menziesii* leaves were found between 34.8-33 ka suggesting potentially more widespread forest occurring prior to the onset of the LGCP. Forest beetle species were still present, although in rare amounts, until 27 ka. Grassland taxa were dominant during this period. From 27-8 ka the area was dominated by an open grassland environment, with no evidence of lowland tree species. However beetle taxa suggest the site possibly represents an altitudinal tree line limit during the LGM (Marra and Thackray, 2010). Like the Howard Valley site, LGM beetle assemblages spanning 21-19 ka from Awatere Valley, Marlborough (Figure 2.7) were dominated by tussock grassland species. However, there was also a large forest influence and the record is interpreted as representing a forest refugia site (Marra and Leschen, 2004). Temperature reconstructions, inferred from the beetle evidence, suggests a maximum summer reduction range of 2.5-5 °C and a winter temperature reduction of 3.5-6 °C compared with current average summer temperatures in the same region (Marra *et al*, 2004).

2.6.2.5 Summary

As with the Westland sites, there is good consensus between the different proxies. Both the vegetation and speleothem records provide evidence for the cold period extending beyond the LGM. However, there are few records which span the LGCP and the majority provide a fragmentary reconstruction of past climate and/or vegetation change. Therefore, there is limited insight of climate variability in the region during the LGCP.

Termination I is represented by rapid climatic amelioration and quick vegetation succession. Pollen evidence suggests a southward warming trend with climate amelioration starting in north Nelson region and spreading southwards. The speleothem records (Hellstrom *et al*, 1998; Williams *et al*, 2005) indicate a late glacial climate reversal interrupting the warming trend between 13.8-11.14 ka and McCaplin (1992b), found a moraine that may be attributed to this cooling. However, the pollen records show no evidence for a LGR. The Nelson region contains a variety of records but there is a paucity of sites extending beyond the LGM, limiting firm conclusions of the pattern of climate change during the LGCP. An increase in records spanning this period is necessary for an improved understanding of vegetation and climate change in the region.

2.6.3 Central Canterbury – Mackenzie Basin

This section will concentrate on last glacial evidence from the Mackenzie Basin and surrounding area. However, in comparison to the other regions, this area has considerably fewer records spanning the last glacial. Therefore, only two subsections will be included in this section, geological evidence and proxy records.

Figure 2.8: Location map of the pollen sites used in the description of past vegetation change. The shaded area is a rough location of ice in the region.

2.6.3.1 Geological evidence

The geomorphology of the Mackenzie Basin is dominated by Quaternary aged glacially derived deposition features, including late Quaternary moraines and extensive outwash aggradation surfaces. Gair's (1967) geological mapping of the region recognises five moraine formations evidence of at least five glacial advances; Wolds, Balmoral, Mt John, Tekapo and Birch Hill formations. Subsequently extensive research dating the moraines has been achieved producing a detailed chronology of glacial activity in the region. The Wold formation, the oldest glacial advance, is assigned to MIS 8 (Suggate, 1990) and

both clast specific gravity research (McGregor, 1981) and geomorphological and loess covered distribution (Maizel, 1989) are consistent with this assignment. A period of glacial retreat activity separates the Wolds and Balmoral advance. Radiocarbon dates from wood fragments located in soil on the Balmoral moraine ages this advance at 40.2 ka (Porter, 1975b). Another glacial retreat period follows until the Mt John, Tekapo and Birch Hill Formations (in chronological order) were deposited during the LGM (Porter, 1975b; Suggate, 1990; Table 2.1). Equilibrium-Line Altitudes (ELAs) were estimated using the above formations which represent the former ice limits. ELAs were 1050 m lower during the Balmoral advance and 850 m lower during the LGM (Porter, 1975b).

Schaefer *et al* (2006) dated the inner terminal moraines from around Lake Pukaki's lake margin. The average age for the end of the LGM is 17.4 ka and a maximum age of 19.3 ka. These ages correspond with other ages of the termination from sites spanning the whole of New Zealand (e.g. Alloway *et al*, 2007). Putnam *et al* (2010a) aged the LGM Boundary Stream moraines at 18 k, agreeing with Schaefer *et al* (2006). Numerous postglacial moraines have been deposited by the glacial still stands during glacial retreat. Putnam *et al* (2010b) age a glacial resurgence culminating at 13 ka corresponding with the ACR, and providing evidence of a LGR east of the main divide. Evidence of asynchronous timing of glacial advance in New Zealand, coinciding with Northern Hemispheric warming episodes continues throughout the Holocene (Schaefer *et al*, 2009).

2.6.3.2 Proxy records

Compared to the other regions, a limited number of vegetation reconstructions have been produced from the Mackenzie Basin and surrounding area. No record extends beyond the LGM. Boundary Stream Tarn is a kettle hole located on a LGM lateral moraine bounding Lake Pukaki. The site became ice free and sedimentation began by 17.5 ka, suggesting the onset of deglaciation occurred around this time. Between 17.5-15.5 ka the vegetation was dominated by mixed grassland and *Coprosma* dominated shrub, representative of an alpine environment and indicating the climate was colder and harsher than contemporary regimes. An expansion of tall tussock grass occurs at 14.1 ka and is combined by an estimated mean summer temperature decline of 1.5-2 °C between 14.1-13.2

ka. This late glacial cooling period corresponds with the timing of the ACR, and supports an Antarctica driven cooling and asynchronous interhemispheric pattern of climate change. This cold period is followed by warming and vegetation change from grassland/shrubland to *Halocarpus* podocarpus forest. After 11.5 ka *Phyllocladus* rapidly becomes the dominant taxon in the record (Vandergoes *et al*, 2008).

Other evidence from the region shows a similar pattern of vegetation change from grassland and bare ground to a grassland-*Myrsine-Coprosma* shrubland followed by first *Halocarpus* and then *Phyllocladus* scrub or low lying forest (Moar, 1971). The early Holocene continued to be dominated by *Phyllocladus*. There is a lack of tall tree taxa in the pollen record and only trace amounts of *Prumnopitys taxifolia* suggest the source was either long distance dispersal or local small stands (McGlone and Moar, 1998). All the records suggest limited forest growth consisting of tall tree taxa during the LGIT and early Holocene.

2.6.3.3 Summary

A detailed morainic history of the region shows a complex glacial pattern of retreat and advance throughout the last glacial period also pointing towards a complex pattern in climate change. The available proxy data correspond with the geological data but the paucity of records extending into the LGCP makes it impossible to know how vegetation responded during this period. Only a handful of records span the LGIT. Moraine mapping indicates that a large proportion of this region was glaciated during the LGM (e.g. Suggate, 1990) limiting the areas where sediment records could have been preserved. Also, the regions climate is very dry, which is not the ideal conditions for proxy record preservation, again making it difficult to find any sites which could preserve a record extending back into the LGCP.

2.6.4 Summary

Section 2.6 has examined broadly equivalent-age records from a number of regions across New Zealand which are relevant to this project and also highlights some issues within the existing records. These include the paucity of records extending into the LGCP from all three regions, different climate reconstructions

produced by different proxy records and whether or not a LGR registered in New Zealand. The results from this project aim to build upon the existing knowledge and tackle some of these issues in a bid to enhance the understanding of climate change in New Zealand.

Chapter 3: Methodology

This chapter will describe each method used in this project and provide a rationale, and if necessary, a literature review of the method. Site selection and field sampling will be described first, followed by palynological methods, including the preparation process, pollen identification and statistical analysis. The other laboratory based methodologies (Loss on Ignition (LOI) and Particle Size Analysis (PSA)) will then be described. Finally the methods used to produce the chronology at each site will be explained.

3.1 Site selection and field sampling

Four new sites that collectively span the interval 30 – 8 ka have been identified (Figure 1.2). The four sites represent locations that today are characterised by strong contrasts in temperature and precipitation. The rationale for site selection is explained in more detail in Section 1.4.

Three sites are moraine-impounded lakes or tarns with associated peat deposits. The moraine complex provides geological time constraint on the age of the deposits. Lake Mudgie and Manks Tarn were sampled in 2002 by Prof. Rewi Newnham, Dr Marcus Vandergoes, Dr Chris Hendy and Prof. Tom Lowell. The sites were cored using a Livingstone piston corer directly on the peat surface (Figure 3.1). In February 2008 the author participated in preliminary coring at Forks Lagoon. Further coring continued in February 2009 with a further two overlapping cores collected. This site was cored from a small boat platform (Figure 3.1) using a Livingston piston corer. At each of the above sites an 8 cm diameter barrel, collecting 1 m long sections, was used cored. Coring continued until either the substrate was too difficult to penetrate or the base of the deposit was reached. Whilst on site, each 1 m section was extruded into half cut PVC piping, logged and wrapped in cling-film. All the cores collected were refrigerated at 4 °C until used for analysis.

Howard valley, the fourth site, has a number of river terrace risers with exposed section, three of which were sampled for this project (in February 2008, January 2009). In 2008 the organic bands were sampled at a 1 cm resolution for pollen analysis. Samples were cut directly from the riser face and placed in labelled, sealable bags. In 2009 monoliths were collected from the thickest terrace riser. These were logged then wrapped in foil and boxed for transport. Two cores were collected from the base of two terrace risers using the Livingstone Piston corer with the 8 cm barrel. The cores were extruded into half cut PVC piping, logged and wrapped in cellophane. All the samples collected from Howard Valley were refrigerated at 4 °C until used for analysis.

Prior to subsampling in the laboratory, the cores and monoliths were split in two and one half was kept as an archive. These samples were then photographed, logged and finally subsampled.

Figure 3.1: Examples of the coring technique used on a) the peat surface and b) the boat setup used for the lake coring.

3.2 Pollen procedure

3.2.1 Sampling Interval

The sampling interval used varied between sites and within sequences depending upon core length and extent of change between pollen assemblages. Initially a skeleton technique, i.e. a coarse sampling interval, was used to sample down the cores. Subsequent higher resolution sampling was carried out for the sections of interest. For all pollen samples 0.5 cm³ of material spanning 0.5 -1 cm depth (depending upon site) was taken.

3.2.2 Pollen preparation

The methodology for the laboratory pollen preparation followed the standard method outlined in Moore *et al* (1991) although modifications to procedures were needed depending on the sediment composition (Figure 3.2). To aid the estimation of pollen concentrations a known volume of sediment was used with exotic *Lycopodium* marker tablets added to each sample at the beginning of the preparation. The samples were soaked in 10% potassium hydroxide (KOH) overnight then boiled in fresh KOH for 20 minutes. These were sieved through a 185 µm sieve to separate the coarse material from the sample. Any macrofossils were collected and stored at 4° C for possible radiocarbon dating. If silicate minerals were present, the samples were soaked overnight in 60% hydrofluoric acid (HF) then heated for 20 minutes in 10% hydrochloric acid (HCl). If any silicate minerals were still present the samples were again heated in 60% HF for 45 minutes then heated in 10% HCl for 20 minutes. The peat samples did not undergo the HF stage. Glacial acetic acid dehydrated the samples which then underwent acetolysis for three minutes at 95-97 °C. A 9:1 acetic anhydride ([CH₃CO]₂O) to concentrated sulphuric acid (H₂SO₄) mixture was used for the acetolysis step. Glacial acetic acid was added to the samples and then washed with distilled water, and sieved through 5 µm removing the fine material. If samples retained a high organic content then the acetolysis process was repeated. The samples were stained using aqueous safranin, washed in water then dehydrated using ethanol followed by tertiary-butanol. Finally the samples were mounted in silicon oil.

Some experimentation with heavy liquid density separation was attempted but the large variation in the sediment matrix from sample to sample required considerable adjustments and so this methodology was not applied routinely.

Figure 3.2: Methodological flowchart for pollen sample preparation. Modified from Moore *et al* (1991).

3.2.3 Pollen counting and identification

Slides were analysed using a light microscope under x400 magnification with x1000 oil immersion used on difficult grains. The identification of the pollen was aided by using New Zealand pollen type slide collections at the University of Plymouth and in Landcare Research from Lincoln, New Zealand. The dicotyledonous pollen types were identified using Moar (1993) and the ferns and fern allies were identified using Large and Biggins (1991). A sum of at least 250 dryland pollen grains was counted for each sample where possible.

Certain pollen types were difficult to identify due to deterioration or ambiguous affinities. The following are the procedures used to handle some of these difficulties:

Podocarpoid: This taxonomic term refers to the bi-saccate (reticulate) grains that have subsequently lost their sacs and only the main body of the grain has survived. These species include *Prumnopitys taxifolia*, *P. ferruginea*, *Largarostrobos*, *Lepidothamnus*, and *Podocarpus*. The taxon podocarpoid is still an important part of the pollen count as it represents the Podocarpaceae species, a tall tree family indicative of a forest environment. The deteriorated nature of these grains may be indicative of a high energy environment during grain deposition, or possible reworking of the grain, which in itself provides further insight of the environment these grains were deposited in.

***Prumnopitys*:** This group was split into 3 headings during counting: *Prumnopitys taxifolia*, *P. ferruginea* and *Prumnopitys* for the grains which couldn't be differentiated between the two species. The counts were then adjusted with the ratio of *P. taxifolia* and *P. ferruginea* used to divide the *Prumnopitys* group between the two.

Spores: Species which represent a distinct environment, i.e. trees ferns (Cyathea and Disksonia species) were identified to species level. Those which were less significant or difficult to identify to species or even family level were placed in one of two groups: monolete spores and trilete spores. The monolete spore group includes the following: Psilotaceae, *Marattia*, *Schizaea* spp., Thelypteridaceae,

Lindsea viridis, Davalliaceae, Dryopteridaceae, *Blechnum* spp., *Anarthropteris lanceolata*, *Sticherus* spp, *Pyttosia eleagnifolia*. The trilete spore group encompasses the following: *Ophioglossaceae*, *Osmundaceae*, *Loxsoma cunninghamii*, *Ctenopteris* spp, Pteridaceae, *Lindsaea* spp. (except *L. viridis*) *Pillularia novae-zelandiae*, *Azolla rubra*.

Poaceae: The pollen grains of this family are very similar and almost impossible to identify to species level or even genus. However, tall tussock grasses (*Chionochloa*) produce larger pollen grains than other grass species in New Zealand, tending to be greater than 40 µm. Based upon McGlone and Moar (1998) the taxon Poaceae was split into two groups: > 40 µm and < 40 µm, the later including all other grass species.

Unidentifiable: This class includes any whole grain which was obscured or deteriorated to such an extent that it could not be identified. The number of unidentifiable grains gives an indication of the quality of the slide but also of the environment during deposition or whether possible reworking has occurred.

3.3 Pollen analysis

A number of calculations, statistical analyse and computer programs have be used to analyse the pollen data. The methods are explained below.

3.3.1 Pollen percentages and concentrations

Pollen data can be represented in a number of different ways. For this project both percentage data and pollen concentration are presented. Relative or percentage data are the most commonly used format for communicating pollen information in diagrams. Percentage data express a given pollen type in sample as a proportion of a specific pollen sum for that sample. This removes the effect of differing amounts of pollen grains in each sample, making comparisons between samples easier (Birks and Gordon, 1985). Pollen concentrations differ from percentage data as the concentrations are independent of changes in other pollen types. The percentage of pollen removes the effect of changing total grain amounts between

samples, however this produces a constraint on the data. As one species increases in relative frequency other species must decrease to preserve the 100% total. This statistical artefact could be interpreted as a negative relationship between the two species. In contrast, pollen concentrations can be used to distinguish between actual decline and relative decline in a species presence (Birks, 1985). However, pollen percentage data provide a good overview of the pollen assemblage and have been deemed a reliable format for showing the pollen results at all sites. In this project only Forks Lagoon uses pollen concentrations as well as percentage data due to the over dominance of one aquatic taxa; *Isoetes*. This is explained further in Chapter 7, Section 7.5.1.

Pollen concentrations can be calculated in a number of ways. Direct estimations occur when all pollen and spores are counted from a sub-sample from a known volume/weight of material. By knowing how much this fraction of subsample relates to the original sample, direct estimations of pollen concentration can occur (Jórgensen, 1967). Alternatively, indirect estimates can occur by adding a known amount of exotic marker pollen or spore, i.e. pollen or spore grains types which are not native to the region and morphologically different to the other pollen and spore that may be observed (Bonny, 1972). For this project a *Lycopodium* tablet with a known number of grains was added to each sample at the beginning of the pollen preparation. Once the preparation has been completed, both the pollen and exotic marker are counted. The indigenous pollen concentrations are then estimated through a simple proportional calculation (Maher, 1972; Birks and Birks, 1980). As mentioned above, only the Forks Lagoon chapter presents concentration results.

3.3.2 Pollen diagram and cluster analysis

The programs TILIA and C2 were used to produce the pollen diagrams (e.g. Figure 4.5). Stratigraphically constrained cluster analysis, a multivariate analysis technique for classifying stratigraphic zones, was used to group similar samples together in clusters. The program CONISS was used to perform the analysis and was used in conjunction with TG View. A dendrogram representing the entire set of amalgamations was sectioned, enabling objective zoning of the pollen

diagrams. Stratigraphically constrained cluster analysis differs from ordinary cluster analysis, as only stratigraphically adjacent samples will be considered for grouping. Zoning the sequences helped with the description and correlation of the pollen assemblages. These zones are defined by the presence or absence of fossil taxa and variations in the abundance of selected taxa. Therefore zones of similar pollen assemblage associated with depth are objectively identified. For this project the dissimilarity coefficient using squared root transformation is the Edwards and Cavalli-Sforza's chord distance. This method up weights rare variables relative to abundant variables (Grimm, 1987).

3.3.3 Principle Component Analysis (PCA)

Principle Components Analysis (PCA) is a numerical ordination technique used to simplify multidimensional datasets, and thus enable relationships within the dataset to be objectively analysed in a low-dimensional space. Each pollen type is represented as a coordinate axis. If 40 pollen taxa were identified through a sequence then each would produce its own coordinate axis providing 40 coordinate axes. This is too highly multidimensional to assimilate and therefore PCA is used to reduce the data to a lower dimension. Factor reduction occurs by constructing new components that are linear combinations of existing variables by coalescing highly correlated variables. The first component lies along the axis of greatest variability within the dataset. Subsequent PCs are then chosen to account for the largest amount of residual variability, although this is restrained by the need to have all the principle axes orthogonal to each other. The purpose is to provide a number of linear relationships (PC) from a set of variables which preserve as much of the information from the original variables as possible. Reducing the dimensional complexity of the results aids the observer to analyse the relationships and communicate the results in an objective way. A disadvantage is that some data will be excluded in the process of reducing the output to only one or a few dimensions. However, if there is a strong structure within the dataset then the first few PCs tend to account for the majority of variability within the data set and the remaining PCs can be discarded with a minimal loss of information (Gauch, 1982; Birks and Gordon, 1985).

Eigenvectors and eigenvalues are at the core of any PCA analysis. Eigenvectors represents the weighting of each individual taxon within each component. Eigenvector scores range from 1 to -1. The more weighting a taxon has on the component the nearer to 1/-1 its score will be. Taxa with similar eigenvector scores would exhibit similar distributions down through the core, thus indicating which taxa occur together. Eigenvalues represent the relative contribution each component has in explaining the total variation in the data. This is calculated by summing the square root of each taxon eigenvectors in each component. In PCA analysis, the components are extracted in descending order of importance in regards to their contribution in explaining the total variation of the data. Therefore, component two will have a lower eigenvalue and represent less of the variance compared to component one, and so on. A dataset containing few variables needs only two-three components to represent the majority of the data whilst four-five components are needed for datasets containing a large number of variables. As a guideline, axes are extracted until the component's eigenvalue is less than 1. Eigenvalues are the exact measure of proportion of the total variation that the axis represents and are often expressed as a percentage.

In Sections 4.4.5.2 and 5.4.4.2 eigenvector scores of axis 1 and axis 2 are plotted. The species which plot together in the graph represent a relationship in their appearance in the stratigraphy. Component scores are calculated to enable relationships between samples to be seen. These are calculated by multiplying the original taxa counts in a sample by the eigenvector values for each variable (taxon abundance). These are then summed to produce a component score. This reduces the data down to each sample with orthogonal component scores for each axis. An ordination plot is used to display the data. Samples which plot close together in the graph have similar species composition.

The program CANOCO v 4.5 was used to for the PCA analysis, which was carried out on the pollen data from Forks Lagoon and the two Westland sites. The particle size analysis and loss on ignition results were also be added to the statistical analyses to assess objectively if there are between relationships these factors and the pollen assemblage.

3.3.4 Environmental Lapse Rate (ELR)

Environmental Lapse Rate (ELR) is the rate at which temperature changes with altitude (Meyer, 1992). It is used frequently in paleoecology, in particular to infer MAT depression from elevation lowering of vegetation zones recorded in palaeovegetation records. For instance, if the pollen record is interpreted as showing a 300 m lowering of the alpine zone from its contemporary elevation, and an ELR of 0.5 °C/100 m is applied, then an estimated MAT depression at the time would be 1.5 °C. However, there are some limitations with this approach. ELRs are a semi-qualitative method which relies on analyst interpretation of the palaeovegetation community which can be more prone to bias. There are also limitations associated with the methodology used to calculate ELR which are discussed below.

Meyer (1992) identified a number of different factors associated with both temporal and spatial changes which can modify the ELR calculated. These factors include changing atmospheric conditions (although over long timescales these are assumed to average out), the surface albedo, moisture, local and regional topography, the time of day, and the nature and source of the dominant air mass to name a few. For example in areas of strong relief such as New Zealand, mountain barriers force air masses to rise causing a significant difference in precipitation and temperature between the leeward and windward sides of the mountains (see Section 2.3). Moisture levels, in particular, have a significant influence on the ELR with the drier leeward side of the mountain ranges experiencing a greater terrestrial lapse rate than the windward slopes. Dry adiabatic lapse rates are typically around 1 °C/100m whereas wet adiabatic lapse rates are around 0.5 °C/100 m (Meyer, 1992). Research by Norton (1985) and Meyer (1992) also show large ELR changes on a number of timescales (e.g. daily, seasonally, and annually). Therefore an ELR based on a single temperature measure, from a limited number of sites and over a short time span would not produce a representative ELR for a region (Meyer, 1992). Meyer (1992) advised to use ELR calculated from a number of climate stations.

ELR has been used widely in New Zealand, including estimating temperature depression (e.g. Burrows *et al*, 1977; Wardle, 1998; Hales and Roering, 2005). However, the rate applied has varied between and within studies. Mark *et al* (2000) used data loggers from three sites in the South Island and found a large difference in ELR during the warmest month of the year for different altitudes. Between 1200-1600 m.a.s.l an ELR of 0.45 °C/100 m was recorded whilst the lapse rate between 1600-2000 m.a.s.l was larger at 0.65 °C/100m. However, research by Norton *et al* (1985) suggests a large seasonal difference in ELR occurs and so a rate derived from only one month or one season would not be appropriate for reconstructing MAT. Using data from 301 temperature stations across New Zealand spanning the period 1950-1980, Norton (1985) calculated a mean annual ELR of 0.5 °C/100 m for New Zealand. Therefore, for this project the ELR calculated by Norton (1985) will be used to estimate MAT depressions from the pollen records. The same ELR has also been applied to a number of other records e.g. Soons (1979).

3.3.5 Quantitative pollen-climate modelling

Quantitative methods for reconstructing climate for a number of biological proxies have been developed over the past 30-40 years. Quantitative pollen-climate modelling aims to express the value of an environmental variable, such as mean annual temperature, as a function of biological data (e.g. pollen assemblages). Before a fossil assemblage can be converted into an environmental variable a modern training set is needed. A training set comprises taxa abundances from various sites with known associated modern environmental variables. The current relationship between taxa and their environment is statistically modelled producing a function. This function is then applied or transferred to the fossil data to infer estimates of past environmental variables (Birks, 1995). The function and process of applying it to reconstruct past environments is broadly referred to as a transfer function.

The use of quantitative pollen-climate reconstructions functions has been successfully applied around the world (Kay and Andrews, 2005; Finsinger *et al*, 2007). Traditionally, to produce a modern training set or analogue, modern moss pollsters and surface sediment sites are analysed. The relationship between the

pollen assemblage and modern climate regime at each site is then statistically modelled. In New Zealand, Norton *et al* (1986) were the first to use modern pollen assemblages to infer the climate from the fossil pollen spectra. However, the study failed to produce a robust transfer function due to a number of factors. A variety of pollen site types were used to produce the modern training set and to perform the transfer function. These included an emphasis on sites from the wetter western regions of New Zealand, where remnants of native forest still remained. In contrast, representation of vegetative communities from other regions was inadequate. Also, frequent steep terrain aids the wide dispersal of pollen and spore grains, blurring relationships between climate and pollen taxa. Most significantly the transfer function based on modern surface samples produced an anthropogenic signal that masked climate influences (Norton *et al*, 1986). With these limitations in mind the pollen transfer function was further developed by Wilmshurst *et al* (2007).

Wilmshurst *et al*, (2007) used exclusively a pre-deforestation pollen database to produce a 'modern' training set. Cores from lakes, bogs and swamps were used with the pollen assemblage from the horizon directly underlying the first signal of human interference used in the training set. This signal includes a sustained upwards trend in amounts of large ($>50\ \mu\text{m}$) charcoal, a decrease in tree taxa and an increase in seral species (McGlone, 1989; McGlone and Basher, 1995; Ogden *et al*, 1998). For a few sites that have remained forested and/or lack evidence of Polynesian deforestation, the 750 cal yrs BP horizon was used. The training set comprises a total of 135 sites (Figure 3.3) spanning the Polynesian deforestation period, 750-550 cal yrs BP. For each site climatic variables were applied from the modern climatic data. This introduces an important assumption and possible limitation of this method; that climate 750 cal. yr BP was the same as it is today. Dendrochronological evidence by Cook *et al* (2006) indicates that the average temperature between 13th – 17th centuries was not significantly different than during the mid-twentieth century. However, there are intervals of sustained warming and cooler excursions within the Polynesian deforestation period. The quantitative pollen-climate analysis performed on the training set indicate that mean annual temperature (MAT) can be quantitatively and objectively inferred from fossil records (Wilmshurst *et al*, 2007), although with comparatively broad

errors. Additionally some limitations of the training set are recognised. These limitations may have a significant influence on the inferred climate reconstruction and therefore it is important that these are noted, as discussed below.

3.3.5.1 The training set

A major requirement for any quantitative climate reconstruction includes a large, high quality training set of modern pollen samples and associated climate data. The training set needs to span a wide climate gradient, but also cover the climate range expected to be represented by the fossil pollen record (Birks *et al*, 2010). Analyst consistency in identification of pollen, and nomenclature use combined with detailed taxonomy is a requirement for both the training set and fossil assemblages (Finklestein *et al*, 2006). It is also preferable that the training set samples should come from the same sedimentary environment as the fossil samples for comparable taphonomies (Goring *et al*, 2010). Although the above points are preferable it is not always possible to fully achieve them all with the limited resources and data available. The following highlights the established limitations of the current New Zealand pollen training set used by Wilmshurst *et al* (2007).

Spatial and climate range of the training set:

Currently 135 samples are included in the training set, providing large coverage of New Zealand and a wide range of environments (Figure 3.4). This training set is an improvement upon that of Norton *et al* (1986). However there are still some regions, e.g. central South Island east of the Southern Alps, which are poorly represented. Some important environments and vegetation communities are also lacking representation, in particular the alpine zone and treeless vegetation sites. During the last glacial period vegetation records suggest a large proportion of New Zealand was dominated by grass and herbland communities that contained few, if any, tree taxa. To achieve reliable climate reconstructions for this period a representative training set containing sufficient alpine and treeless sites is necessary. However, collecting representative samples of the alpine vegetation is difficult due the effects of wind transporting lowland pollen into the alpine zone.

Level of taxonomic identification:

To produce a reliable training set for climate reconstruction, pollen and spore identification should be made at as low a taxonomic level as possible. Ideally identification to species level would produce the best training set, but some pollen types are difficult to distinguish, whilst the analyst may not consistently make a distinction with other types. In these cases pollen and spore types are combined into higher taxonomic groups. For the New Zealand pollen training set, around 40% of the pollen and spore types represent individual species and a number of these are major dominant species in the modern vegetation e.g. *Dacrydium cupressinum*, *Nothofagus menziesii*. However, 55% of the pollen and spore types represent one or more species within a genus, and the remaining 5% consist of family groups containing two or more closely related genera or a large group representing a range of species from a number of families e.g. monolete spore grouping fern spores with similar morphologies together. This grouping can reduce the effectiveness of the training set to produce realistic climate reconstructions as some genera or groups may represent a wide range of growth forms and cover a broad environmental gradient e.g. *Coprosma* and *Myrsine*. When analysing the model reconstructions it is important to know the limitations of using certain pollen and spore types in the training set.

Site selection:

For an accurate reconstruction Birks *et al* (2010) suggest the sedimentary environment of the modern and fossil samples should be the same. Basin, lake or bog size, are potential sources of variation in pollen taphonomy associated with site types. However, to cover a broad climate gradient and to represent a large proportion of New Zealand it was necessary to use sites from a range of sedimentary environments with cores taken from bogs, swamps and lakes. Wilmshurst *et al* (2007) concede it would have been preferable to use only small lakes to produce the training set, but also noted that the sedimentology of a site may have changed over time e.g. terrestrialisation of a lake. Ideally, a sufficient number of samples which are fully representative of all major site types which would allow the segregation into site categories would be favourable. However, the limited dataset available does not permit this and may reduce the power of the models to reconstruct past climates.

Summary:

The above points explain the limitations of the current training set used for quantitative pollen-climate reconstructions. However, as Wilmshurst *et al* (2007) has demonstrated, this new pre-deforestation database has significantly improved upon that of Norton *et al* (1986). Also, they have shown that meaningful quantitative MAT reconstructions can be produced for sites in New Zealand and the estimated pattern of MAT are consistent with previous qualitative reconstructions. Quantitative pollen-climate reconstructions also allow for a more objective climate reconstruction which should complement the qualitative results.

3.3.5.2 Methodology

The training set developed by Wilmshurst *et al* (2007) was used to infer MAT from the fossil pollen records produced in this project. Wilmshurst *et al* (2007) performed Detrended Correspondence Analysis (DCA) to the training set to assess the main gradient of the data and whether pollen and spore appearance have a linear or unimodal response to the environment. The calculated gradient was short suggesting linear methods of ordinations would perform better than unimodal methods (ter Braak, 1987). Two models were run with the performance of each compared: the linear response model Partial Least Squares regression (PLS) and the modern analogue technique. These statistical approaches are discussed below.

PLS is a multivariate calibration-function approach which contains the prior assumption that the data behave in a linear manner. This method uses the 'global' estimations of parametric functions for all the taxa present allowing some degree of extrapolation (ter Braak, 1995). Before the training set can be used to reconstruct past climates from the fossil data the relationship of individual species to climate is modelled using regression analysis. This initial stage produces the climate function which is then applied to transform the whole fossil assemblage into a quantitative estimate of past climate (Birks *et al*, 2010). This method successively selects a number of PLS components derived during the cross-validation process of 1000 bootstrap cycles. The PLS component chosen to reconstruct past climates should be the model providing the best statistical results whilst being the lowest component possible (ter Braak, 1995).

The modern analogue technique numerically identifies the closest modern analogues to the fossil assemblage using chi-squared distance of dissimilarity measure. Once the closest modern analogue samples are identified, the model takes a weighted mean of the modern environmental variables for the samples to infer the environmental conditions represented by the fossil assemblage. This process is repeated for each fossil sample. Climate reconstructions can be based on one modern sample which contains the closest pollen assemblage as the fossil assemblage. However, for greater reliability the climate reconstruction should be based on the mean or weighted mean of number of closest analogues. For this project a weighted mean of the 10 closest analogues for each fossil assemblage (Birks, 1995; Birks *et al*, 2010). Cross-validation of the dataset is necessary to assess whether enough analogues have been used to provide a reliable climate reconstruction and test the training set. The cross-validation method of bootstrapping on the modern samples using 1000 bootstrap cycles was used to provide sample-specific errors for each fossil sample.

The main difference between the two models is that the PLS model allows for extrapolation of climate for fossil assemblages which are lacking a close modern analogue. However, this method also incorporates species which represent a broad climate gradient and thus may reduce the climate signal produced by the model (see Section 8.2.4). In contrast, W-MAT can be very accurate when the fossil samples have closely matching modern analogues, but becomes inaccurate when a close match is not available (Birks *et al*, 2010). W-MAT is also more susceptible to autocorrelation, the tendency of sites located near each other to resemble one another compared to randomly selecting sites. Therefore variables from one site can be partially predicted from its neighbouring site. This can cause a lack of sample independence and cause spurious assumptions between environmental variables and species assemblage (Telford and Birks, 2005). To decide which model would be the most suitable the statistical parameters of each model needs to be assessed. The software C2 is used to perform the calibration and the outputs are presented in Table 3.1.

3.3.5.3 *The model's performance (Table 3.1)*

For all the models a number of parameters were calculated to assess each models performance. Each model's parameter outputs are compared and the best performance is the model used to reconstruct MAT. These parameters are briefly explained in the following. The root mean square error (RMSE) estimates the accuracy of the training set to predict temperature (Birks, 1995) whilst the r^2 (coefficient of determination) represents the strength of the correlation between the observed and inferred MAT. The root mean square of predictions error (RMSEP) is a more robust measure of estimated precision of the predictability of the modern training set. This is calculated using a cross-validation method to test the training set and as mentioned already, 1000 bootstrapping cycles was used. The RMSEP is the square root of the sum of squares of the difference between the observed and predicted temperatures (ter Braak and Juggins, 1993; Birks, 1995). The r^2 (boot) is calculated for the RMSEP, and is thus a measure of the correlation between observed and inferred MAT after the cross-validation. Finally, max bias (boot) is a measure of the accuracy of the prediction (Birks, 1995). All these parameters are used to assess the robustness of the models. For the PSL model, which produces five components, the component chosen has the smallest number of useful components needed to produce a reliable calibration (Birks, 1995).

The parameter values for the models and components used to calibrate the New Zealand pre-deforestation dataset are displayed in table 3.1. The two models which performed the best and are the most suitable models to use for the MAT reconstruction is the weighted modern analogue technique (W-MAT) and PSL component 3 (PLS-C3). Both models performed similarly. PSL-C3 marginally outperforming W-MAT with the highest precision of ± 1.4 °C compared to ± 1.5 °C. However, W-MAT gives the highest r^2 (boot) of 0.82, (PSL-C3 is 0.81), and a significantly lower maximum bias (boot) of 1.95 °C compared to 2.53 °C for PSL-C3. Statistically, the W-MAT produces the most robust model and therefore W-MAT is the model favoured in this project and by Wilmshurst *et al* (2007). However, there are limitations with both models which are discussed in Section 8.2.4 and so both reconstructions are displayed in the pollen diagrams and compared in the results sections of the individual site chapters (Chapters 4-7).

Model	RSME (°C)	r^2	RMESP (°C)	r^2 (boot)	Max bias (°C)	(boot)
Modern Analogue Technique	1.46	0.82	1.55	0.82	2.12	
Weighted Modern Analogue Technique	1.38	0.83	1.51	0.82	1.95	
PSL component 1	1.77	0.66	1.85	0.65	3.6	
PSL component 2	1.35	0.81	1.5	0.78	2.39	
PSL component 3	1.2	0.85	1.4	0.81	2.53	
PSL component 4	1.1	0.87	1.42	0.82	1.9	
PSL component 5	1.1	0.88	1.46	0.81	1.84	

Table 3.1: Parameter values for the different models used to calibrate the pre-deforestation training set (source: Wilmshurst pers comm).

Figure 3.3: Location map of the current modern sites used in Wilmshurst *et al* (2007).

3.4 Loss on Ignition and particle size analysis (PSA)

Sedimentological analyses of the sites consisted of Loss on Ignition (LOI) and particle size analysis (PSA). The content of organic carbon in a sample can be estimated by LOI. This provides a means of analysing the biological productivity and relative stability of the catchment. In previous research, high LOI has occurred during periods of forest expansion that is related to climate amelioration after glacial periods (Bos *et al*, 2007). In contrast, low total LOI typically occur during glacial periods where there was reduced productivity, vegetation cover and increased erosion rates. This indicates that climatic change can have a profound influence on carbon content (Beer *et al* 2007; Bos *et al*, 2007; Nesje *et al*, 2006). However, other factors can influence organic content and therefore the LOI results should be interpreted with the pollen results. The methodology for LOI was based upon Heiri *et al* (2001). Approximately 2cc of material for each sample was taken from the core. A crucible was weighed using analytical balance and weighed again with the sample added to the crucible. This was carried out for each sample and was then placed into a drying oven at 100 °C for 12 hours. The samples were left to cool in a dessicator and then reweighed. The samples were then transferred to a furnace at 550 °C for four hours after which the samples were again placed in the dessicator to cool. Finally the crucible and the now combusted sample was weighed for a final time.

The analytical balance was connected to a computer which automatically calculated the LOI. However, the formula used in the calculation is:

$$\text{LOI} = (W3-W4)/(W2-W1) \times 100$$

Where:-

W1 = Weight of empty crucible

W2 = Weight crucible and sample

W3 = Weight crucible and dried sample

W4 = Weight crucible and combusted sample

Particle size analysis measures the variations in the particle size of the sediment within different samples. Previous studies have suggest larger particle size is

related to increased erosion and sediment transport caused by reduced vegetation cover producing bare slopes, increased freeze thaw activity, seasonal snow melt leading to high fluvial discharge, glacial erosion in the mountainous regions and a number of other factor (Digerfeldt *et al*, 2000). A reverse to the above may therefore result in smaller particle sizes, although this is not always the case (Digerfeldt *et al*, 2000; Lawson *et al*, 2004; Shichi *et al*, 2007) as changes in particle size can be influenced by a number of other factors and therefore the results will need careful interpretation.

Samples from the core were divided into three or five sub-samples and placed in vials. Hydrogen peroxide (H_2O_2) treatment was used to remove the organics from the samples. The removal of the organics using H_2O_2 is common practice for particle size analysis (Gray *et al*, 2010), however the exact method used varies. In this project approximately 5 ml of 3% H_2O_2 was added to each vial and heated in a water bath at 95 °C. After a few hours fresh 3% H_2O_2 was again added to each sub-sample. Again, after a few hours 6% H_2O_2 was added to each sample. The addition of H_2O_2 continued until there were no longer visible signs of a reaction (e.g. bubbling). When it appears that all the organic material has been removed the vials remained in the hot water bath until the majority of H_2O_2 has evaporated off. It is important throughout this process that the samples do not dry up entirely. Deionised water is added to each sample and any visible organics, which did not react with the H_2O_2 were removed. This part of the process should take roughly 1 to 2 days to complete it is important that fresh H_2O_2 is continually added until a reaction was no longer taking place.

Particle size analysis was carried out using a Malvern Mastersizer-2000 laser granulometer, with Hydro-G automated dispersal unit, and set for detecting grain size range from 0.02-2000 μm . The particles were passed through the particle sizer using the liquid suspension method. For each sample 3000 blue and red laser measurements were recorded. This was replicated five times for each sub-sample. The Mie theory is used to calculate the particle size distribution in each sample (Rawle, 2011) from which further measures of particle size distribution such as the proportion of sand, clay and silt is made. As mentioned earlier the

PSA and carbon content will be combined with the pollen data and used in the PCA. Using a multiproxy approach should strengthen the vegetation interpretation.

3.5 Chronological method

3.5.1 Radiocarbon Dating

Since American chemistry professor, Willard Libby, thought up the application of radiocarbon dating, it has become one of the most popular methods for aging late Quaternary sediments (Walker, 2005). Subsequently, a multitude of ages have been published and the method for all parts of the process have evolved and refined e.g. Vandergoes and Prior (2003); Newnham *et al* (2007b). Accelerator Mass Spectrometry (AMS) is used to calculate the sample age throughout this project and radiocarbon ages will provide the majority of chronological control at the four sites.

3.5.1.1 Sample preparation

The sample strategy used for pollen and sediment analysis consisted of 1cm and 0.5 cm thick samples taken at regular intervals down the core. In sections of interest, such as an apparent climatic oscillation or climatic transition, sampling was at a higher resolution. As a result, the number and location of dates for each site is dependent upon; 1) the temporal span of the core or section; 2) the amount of climatic transitions and millennial scale events evident in the core; and 3) the resolution and location of the pollen samples, which provide the climate signal. The aim of the project is to constrain the timing of the major climate transitions and also the beginning and end of any millennial scale climate events, e.g. LGR. However, the cost of dating and funding availability imposes limits on the amount of possible ages.

Macrofossils are the preferred medium of dating, when available, due to their general reliability compared to bulk sediment samples. The high rainfall Westland region, the location of Lake Mudgie and Manks Tarn, is known to potentially produce serious dating errors (Moar and Suggate, 1996) caused by younger carbon contamination by humic and fulvic acids. This is most notable with peats

older than 10 ka (Hammond *et al*, 1991). Vandergoes and Prior (2003) established that pollen concentrates consistently produced older ages compared to bulk sediment for dating. Newnham *et al* (2007b) further developed the pollen concentrate method but got slightly disappointing results. Contamination, believed to be sourced from the University of Plymouth, produced younger than expected ages. However, the results pointed to pollen concentrates providing more reliable ages. Recent evidence suggests the source of contamination at the University of Plymouth was the sodium polytungstate liquid caused during factory processing (Vandergoes pers comm).

This project does not use heavy liquid separation. Instead numerous sieving steps, dividing the sample into size fractions, were used. After observing each size fraction under the microscope the most suitable sample fraction, believed to produce the least contamination, was used for dating. The preference was upon the smallest size fraction which contained the highest concentration of pollen. During the sieving process each sample and any standards underwent an acid-alkali-acid (AAA) treatment (refer to Figure 3.4). Samples from Lake Mudgie and Manks Tarn were sent to NERC radiocarbon in East Kilbride for the final stages of the sample preparation and the dating. Forks Lagoon and Howard Valley samples were sent to Rafter Radiocarbon Facility in Lower Hutt.

3.5.1.2 Age calibration and age-depth models

The program OXCal was used to calibrate the ages using IntCal09 calibration dataset set (Reimer *et al*, 2009). This is coupled with simple linear interpolation to produce the age-depth models. Age-depth models are used to provide an age gradient up a stratigraphy allowing sediment deposition rates to be calculated and intermediate samples to have an interpolated age. The most commonly used age-depth models include linear regression, linear interpolation and splines (Bennett, 1994). Refer to Telford *et al* (2004) for an overview of the different age-depth models. For this project linear interpolation models is used. This method connects the age estimates to each other by a line of constant gradient (Bennett, 1994). This model is known to be incorrect as sediment accumulation does not change at the location chosen for radiocarbon dating. However, it is also considered to be the method that provides a reasonable general solution. As stated by Bennett and

Fuller (2002) linear interpolation is “...never right but rarely unacceptably wrong” (Bennett and Fuller, 2002; pp427).

Fig 3.4: Sieving separation methodology used to extract pollen for radiocarbon dating.

3.5.2 Tephrochronology methodology

Tephra is the airborne, unconsolidated pyroclastic material of an explosive volcanic eruption (Lowe and Hunt, 2001). Tephrochronology is the identification of tephra (using geochemical and mineralogical properties) and the use of these tephra deposits/layers as chronostratigraphic markers. The assumption is that a tephra layer will have the same age wherever it is deposited, unless the material has been reworked, and therefore tephra layers are valuable for linking disparate terrestrial records and providing ages (Lowe *et al*, 2008). The KOT eruption is the

largest known eruption from the Taupo Volcanic Zone (TVZ), located in central North Island, New Zealand (Self 1983; Figure 2.1) and ejected a ca 1200 km³ bulk volume of material from the volcano (Alloway *et al*, 2007). The KOT deposit is the most widespread late Quaternary tephra layer in New Zealand (Pillans *et al*, 1992) and is found throughout the South Island, which has no Quaternary volcanic centres (Mew *et al*, 1986; Nicol *et al*, 1994 McLea, 1996; Carter *et al*, 1995; Vandergoes *et al*, 2005). Mineralogical and geochemical analysis of the tephra layer confirms the source of the material and therefore the age. This analysis includes assessing the visual characteristics of the glass shard morphology and microprobing of the tephra to calculate whether the major and trace element glass shard chemistry matches the reference material (Lowe *et al*, 2008).

KOT samples from Howard Valley and Forks Lagoon were sent to Victoria University of Wellington for geochemical analysis. Electron probe microanalysis of the major elements was carried out using a JEOL 733 superprobe. The mean major element compositions for all samples are displayed in Appendix G. Comparison with reference samples confirm the tephra layer at both sites as the KOT. The Lake Mudgie tephra layer did not undergo geochemical analysis but is assumed also to be KOT. This seems reasonable from the stratigraphic position and because the KOT is the only late Quaternary tephra layer yet reported from this region. The KOT is believed to have been deposited prior to the LGM (Lowe *et al*, 2008) and therefore provides an important time-stratigraphic marker (Pillans *et al*, 1992). However, determining the age of the KOT has proved problematic. The inconsistent ages dating the KOT have been attributed to the variable abundance of carbonaceous material directly associated with the eruption (Froggatt and Lowe, 1990). A number of radiocarbon ages, ranging from 20-25 ¹⁴C ka BP, related to KOT deposition has been published (e.g. Wilson *et al*, 1988; Hammond *et al*, 1991; Pillans and Wright, 1992; Newnham *et al*, 2008b; Lowe *et al*, 2008). The recommended age by Lowe *et al* (2008), is 22,590 ± 230 ¹⁴C y BP (ca 27,097±957 cal. yr BP; 95% probability). However, high rainfall regions of New Zealand have produced anomalously young ages, in comparison to the age of Lowe *et al* (2008), and these have been attributed to *in situ* contamination by younger carbon (e.g. Hammond *et al*, 1991; Gillespie *et al*, 1992; Moar and Suggate, 1996; Newnham *et al* 2007b). Other methods for dating the tephra have added to the debate.

Optical luminescence dating has provided both consistent ages, e.g. Shepherd and Price (1990) and Lion *et al* (2002), and also conflicting younger ages, e.g. 19 ka by Grapes *et al* (2010), although the younger age has been considered unlikely (Lowe *et al*, 2009).

Thus the precise age of KOT remains somewhat problematic. The current adopted age of the KOT is $22,590 \pm 230$ ^{14}C y BP (ca $27,097 \pm 957$ cal. yr BP; 95% probability) was produced by an activity-weighted pooled mean age derived from four small carbonised wood fragments embedded within an ignimbrite emplace during the eruption. (Lowe *et al*, 2008). The four fragments, from different sites, represent wood from short-lived plants killed during the eruption and their ages have narrow error ranges (Wilson *et al*, 1988), unlike other ages which are derived from airfall tephra sites and organic material less directly associated with the eruption and which have typically produced wider age ranges. However, recent research by Vandergoes *et al* (2011) has evaluated the age of the KOT in the context of recently obtained ages from pyrolised plant material within ignimbrites, intact and in-situ plant remains which were rapidly buried by distal airfall tephra, and organic material from undisturbed lake sediments bracketing the tephra deposits. Different preparation processes were also considered. On the basis of their evaluations, Vandergoes *et al* (2011) recommended a revised age of $25,400 \pm 400$ cal y BP (2σ age-range) for the KOT.

The recommendation of this group, which included a number of New Zealand Quaternary specialists with considerable experience with this particular problem including Lowe & Wilson, has been followed in this thesis. Nevertheless, the validity of both ages (Lowe *et al*, 2008; Vandergoes *et al*, 2011) in relation to the radiocarbon ages in this study will be discussed in chapter 8. Also, comparisons between reconstructions from this project, which use the younger KOT age, and published records, that use the older age, will be based upon the stratigraphic location of the KOT layer, independent of age. See Section 8.2.1 for further discussion of the KOT age and the chronologies of the pollen records in this site.

Chapter 4: Lake Mudgie

Lake Mudgie is the first of two pollen sites from central Westland. It is a shallow lake located inland, 17 km north-east of Hokitika. Geological mapping of the area indicates the region was extensively glaciated during the LGM (Figure 4.1). However Lake Mudgie appears to be outside this ice margin. It was therefore assumed, and later confirmed, that the sediment accumulation at the site began prior to the LGM. This provided a rare opportunity to produce a continuous pollen record extending beyond the LGM, spanning a large proportion of the LGCP. This chapter begins by providing background information for the region and the field site. The core stratigraphy, radiocarbon dates, pollen and sediment analysis are presented and then discussed. The chapter concludes with a summary section of the findings from Lake Mudgie. Chapter 8 will provide a summary discussion chapter combining the evidence gathered from all four sites investigated in this project.

4.1 Regional setting

In this project central Westland is defined as the narrow lowland coastal plain spanning 42°40'S to 43°40'S. It is bordered by the Tasman Sea to the west and by the Southern Alps mountain range to the east, both of which heavily influence the climate of the region. The mountain ranges support the largest of New Zealand's glaciers, which are climatically sensitive and lie in close proximity to the lowland coastal area. Numerous glacial advances, occurring throughout the Pleistocene, extended out from the mountain ranges onto the coastal plains creating large piedmont glaciers. In most instances the glacial tongues reached close to, and even extending beyond, the present coastline (Suggate, 1990; Figure 4.1). For more information of Westland ice advances refer back to Section 2.6.1.1.

Central Westland lies within the Western South Island (WSI) regional climate zone (Kidson, 2000; Lorrey *et al*, 2007; see Section 2.3 for more information) and is dominated by the westerly airflows crossing the Tasman Sea. The region is therefore characterised by an oceanic/maritime climate, experiencing warm

summers, mild winters and high precipitation. Mean annual rainfall in the coastal areas ranges from 2750 mm in south Westland to 2150 mm in north Westland. Mean annual temperature 11.7 °C and ranges from 10-13 °C (New Zealand National Climate Centre, 2010). Average humidity is 80-90% with dry periods rare. However, the topography and distance from the coastline have a significant influence on the climate. The forcing of the moisture laden westerly winds over the axial ranges significantly increases precipitation levels with altitude. A level of 12000 mm/yr has been recorded in the high altitudes west of the main divide (Figure 4.1; McSaveney and Griffiths, 1983). Altitude and aspect also affect temperature and inland areas experience considerably cooler winters than coastal sites (Moar and Suggate, 1996).

4.1.1 Geological setting

Tectonic uplift, caused by the collision of the oceanic Australian Plate into the continental Pacific Plate, has caused intense folding and faulting of the Alpine Fault, forming the Southern Alps (Katz, 1979). The Southern Alps are formed from a combination of Mesozoic and Palaeozoic quartz-feldspathic schist and Mesozoic greywacke. Quaternary deposits dominate the lowland plains and valleys comprising glacial outwash gravels and tills from the mid-late Pleistocene glacial advances. The main river valleys contain Holocene sand, silts and swamp deposits. Cenozoic sequence of folded greywacke, granite and marine siltstone form outcrops within the Quaternary deposits. (Warren, 1967).

Large moraines dominate the lowland areas, providing evidence of the successive ice advances of the mid-late Pleistocene. In the lowland areas, impeded drainage caused by the hummocky terrain and kettle hole depressions are commonly infilled by swampy deposits. Likewise, in the higher altitudes kettle holes and depressions located on top of terminal and lateral moraines contain small ponds, tarns and swamps which have subsequently been infilled since their creation (Moar and Suggate, 1996).

4.1.2 Present vegetation

A number of vegetation studies in the region have described the floristic composition and plant communities (Wardle, 1975; 1977; 1991; Moar 1970; Reif and Allen, 1988). This section will provide a brief overview of the main native vegetative communities and their contemporary distribution including altitudinal restrictions (also refer to Section 2.4).

Central Westland is located in the beech gap (refer to Section 2.4.1.2), spanning 42°40' S to 43°40' S, and therefore the lowland forest comprises podocarp/broadleaved species dominated by *Dacrydium cupressinum*. The forest canopy also contains a scattered presence of *Prumnopitys ferruginea* forming part of the overstory with *Weinmannia racemosa*, *Podocarpus totara* and *Metrosideros umbellata* typically. Trees ferns, including *Cyathea smithii* and *Dicksonia squarrosa*, are also often important in the small tree tier. Soil development has an important successional influence on the lowland vegetation composition in the Westland region. *Dacrycarpus dacrydioides* dominated communities typically grow on newly formed or frequently rejuvenated soils, such as river terraces that experiences episodic flooding and nutrient replenishment. *D. dacrydioides* performs well in high nutrient environments (Wardle, 1974). However, the high regional rainfall causes rapid leaching, nutrient loss and soil development. Within 1000 years nutrients which are important for *D. dacrydioides*, such as phosphorous, are no longer readily available and *D. dacrydioides* is replaced by *Dacrydium cupressinum* and *Prumnopitys ferruginea* dominated vegetation (Norton and Leathwick, 1990).

Above 500 m.a.s.l, the montane forest zone begins characterised by a *W. racemosa* and *M. umbellata* dominated forest canopy and a significant decline in *Dacrydium cupressinum* and *Prumnopitys ferruginea*. Other important species include *Cyathea smithii*, *D. squarrosa*, *Coprosma*, *Griselinia littoralis*, *Pseudopanax* and *Schefflera*. On soil experiencing leaching, *Phyllocladus alpinus* is also common. The *W. racemosa*/*M. umbellata* montane forest remains present up to 850 m.a.s.l, where the vegetation develops into sub-alpine scrub dominated by the shrubs *Coprosma*, *Myrsine*, *Pseudopanax*, *Hebe* and *Asteraceae*.

At c. 1200 m.a.s.l the alpine zone develops with grassland/herbfield dominating the vegetation community. This zone includes the herb varieties of *Coprosma*, *Hebe* and Asteraceae and the herbs *Gentiana*, Apiaceae and *Kelleria*. A variety of grasses grow in this zone including the tall tussock taxa *Chionochloa* and the small tussock grasses.

Figure 4.1: Location map of Lake Mudgie in central Westland with a) with Lake Mudgie site location in context of the LGM ice extent in the Westland region (redrawn from Suggate, 1990) and b) a zoom in site location with Lake Mudgie core site marked by a red dot. Moraine deposits are marked on the map and are colour coordinated relating to the glacial advance which deposited the features. Moraine location is based upon Barrell *et al* (unpublished). Red- Moana moraines; Purple; Larrikins moraines; Green – Loopline moraines; Deep purple – Waimea moraines

Lake Mudgie is a moraine-impounded lake bounded by a Loopline associated terminal moraine (Figure 4.1; refer to Section 2.2 for terminology definitions used in this section), and developed within Loopline formation terrace (Moar and Suggate, 1996). The terminal moraine and hummocky landscape has impeded drainage to form the wetland comprising the lake. Lying at 143 m.a.s.l, Lake

Mudgie is elevated above the subsequent Larrikins ice extent and associated outwash deposits (Moar and Suggate, 1996), and is confirmed by the presence of the 25.4 ka old KOT layer.

East of Lake Mudgie the Taramakau River flows in a north-west direction (Figure 4.1) and enters the Tasman Sea south of the small settlement Camerons. Throughout the Quaternary the Taramakau River valley was extensively glaciated with glacial advances reaching within a few km of the current coastline. An LGM lateral moraine separates the Taramakau River and Lake Mudgie and the river flows through the terminal moraine north-east of the study site (Figure 4.1b). The area between the lateral moraine and Lake Mudgie consists of numerous hummocky moraines and depressions, deposited by the pre-LGM glaciers and in which Lake Mudgie is formed.

Lake Mudgie is a small lake with much of the area, including the core site, terrestrialised. The core was taken on the bog surface which was composed of wetland vegetation. The bog grades into a fringing scrub and then lowland podocarp forest (Figure 4.2). A pollen record from Lake Mudgie was previously produced by Moar and Suggate (1996; mentioned in Section 2.6.1.2). The core site of this early research was located in an infilled embayment to the east of the southern end of Lake Mudgie (Figure 4.1). It was initially connected to the lake by a small shallow channel, which has subsequently become blocked by peat. The pollen record produced had a comparatively low sample resolution and was not dated. However, it provides a good opportunity to compare the two records and confirm the main vegetation phases at the site. The pollen record by Moar and Suggate (1996) clearly shows a transition from Poaceae dominated vegetation, assumed in the paper to occur during the LGM, followed by a successive pattern of shrub then tree species culminating with *Dacrydium cupressinum* dominated forest marking the start of the Holocene.

Figure 4.2: Coring on the surface of Lake Mudgie. In the background the higher vegetation marks the bounding moraine.

4.2 Stratigraphy

The core location was on the north-eastern bank of the Lake Mudgie, on top of a swampy, terrestrialised section of the site (Figure 4.2). The method used for coring is described in more detail in Section 3.1. The top 2.5 m of the core consisted of sloppy fibrous restiad peat and presumed to be late Holocene and was therefore not recovered. Coring began at 2.5 m down and seven 1 m sections were collected. The primary aim of this project, as stated in the introduction (Section 1.3), is to produce a palynological record spanning 8-30 ka. This interval is represented by the bottom three core sections as described in the section below and in Figure 4.3.

The lowermost section, 647.5-621 cm, is dominated by blue/grey silts and sands which become laminated towards the base of the section. Between 621-559 cm the silts become greyer in colouration and a pebble layer is located at the top

boundary of this section (559-560 cm). Within this section, a 2 cm thick orange-brown layer was deposited at 593-595 cm. This is believed to be the KOT based upon the high abundance of volcanic glass shards in the matrix. However, this has not been confirmed by geochemical analysis, and is a best guess assumption. The silts become increasingly browner and darker up-sequence. The top 66 cm, 516-450 cm, is composed of reddish brown fibrous peat. These changes in the stratigraphy are characteristic of gradual terrestrialisation and shallowing of a lake. The cores above the sections analysed in this project consist of reddish brown fibrous peat, containing visible plant macrofossils. This is of a similar composition as section 450-516 cm. It is important to note that a possible earthquake crack beginning at 536 cm and ending c. 608 cm is visible in the core. This could be a cause of contamination and therefore during the sampling process any material associated with this crack was avoided.

When coring at the site, Moar and Suggate (1996) did not find, or record, the presence of the KOT layer. As it is present in the core used for this project, it seems likely that this record extends further in time than the previous record by Moar and Suggate (1996).

Figure 4.3: Stratigraphic illustration of the three thrusts analysed with the core images beside each corresponding illustration. Calibrated ages are added to the figure. *KOT age (Vandergoes *et al*, 2011).

4.3 Chronology

Three samples provided three AMS ages to establish the chronology for this sequence (Table 4.1). All the ages were derived from pollen concentrates (the method is described in Section 3.5.1). A further age, produced by the chrono-stratigraphic marker the KOT, provides the age c. 25.4 ka for the horizon 593-595 cm, the same horizon as SUERC-28872.

Core Depth	Fraction	¹⁴ C age BP	δ ¹³ C ‰	Reference code	% carbon yield	Calibrated age BP (2σ)
525-526	< 63µm	11908 ± 49	-29.1	SUERC-28870	13.2	13762 ± 165
555-556	< 63µm	14527 ± 62	-26.4	SUERC-28871	-	17604 ± 346
593-594	< 63µm	17418 ± 85	-25.2	SUERC-28872	0.14	20784 ± 423
*Wood standard		11727 ± 50	± 25.0	SUERC-28880	51.9	
*Coal bit standard		54160 ± 350	± 23.2	SUERC-28881	88.6	

Table 4.1: Radiocarbon age details for Lake Mudgie. Ages were calibrated using the online OXCal program
 *Standards used to assess if any contamination occurred during the preparation process.

The table above indicates the three AMS ages lie in stratigraphic order. However, when the established KOT age is added to the group, the oldest AMS age, SUERC-28872, is significantly younger than the tephra layer which this age dates. The two standards that underwent the same laboratory process as the other ages, reproduced ages that were expected, indicating that any contamination by both older and younger carbon introduced during the analytical procedure was minimal. The top uppermost ages, at 555-556 cm and 525-526 cm, are consistent with the pollen stratigraphy (see Section 4.4.1 and Figure 4.5) and with the chronologies for the other pollen records in the Westland (Vandergoes and Fitzsimons, 2003; Vandergoes *et al*, 2005; Newnham *et al*, 2007a). However, the radiocarbon age for sample SUERC-28872 appears younger than expected, both in regards to its depth in the stratigraphy in relation with the KOT Layer and the pollen assemblage

at the same horizon (see Section 4.4.1). The following section discusses this age in particular.

4.3.1 Chronology discussion

Sample SUERC-28872 was taken at 593-594 cm in the core, the horizon directly above the chronostratigraphic marker, the KOT layer. The recently refined age of c. 25.4 ka has been applied to the KOT and is currently regarded as a reliable age for this deposit (Vandergoes *et al*, 2011). Therefore, the date for SUERC-28872 is considerably younger than would be expected for a sample located directly above the tephra layer. The standards indicate that contamination during the preparation procedure was minimal and certainly not significant enough to produce such a young age. The remaining possibilities are that the “accepted” age of the KOT layer is wrong (refer to Section 3.5.2), or contamination from younger carbon occurred prior to the preparation process (see Section 3.5.1). The core stratigraphy (Figure 4.3) shows another potential source of contamination. A probable earthquake induced crack is visible from 536-608 cm. This crack could have opened a pathway for younger carbon to go down the sediment column, and thus causing contamination. Although every effort was made to prevent incorporating material associated with the crack with the dated sample, the low carbon content and the expected age of the sample means that even a small amount of contamination can have a large influence on the age. The <90 µm sediment fraction was used to produce a high pollen concentration sample for dating. It is possible that pollen travelled down the earthquake intrusion causing the contamination. However, this is unlikely as contamination from the younger sediment would contain different taxa and would be noticeable in the pollen assemblage. The most likely cause of contamination is from the more mobile humic and fulvic organic acids. It is therefore likely that contamination at the site is the cause for the young age. However, it is still important to ascertain whether the accepted age of the KOT age is a realistic age.

Figure 4.4 presents an age-depth model for Lake Mudgie. SUERC-28872 AMS is included in the figure but not in the model. If the KOT age is ignored and the SUERC-28872 age used, an initial sediment accumulation rate of 84 yrs/cm is calculated between SUERC-28872 and SUERC-28871. This is followed by a

decrease in accumulation rate of 128 yrs/cm between samples SUERC-28871 and SUERC-28870. This would suggest higher rates of sediment accumulation occurred during the LGCP than the LGIT. When the KOT is used instead the sediment accumulation rate reduces to 205 yrs/cm, over twice as slow. This rate is also an average 70 yrs slower per cm deposited than the LGIT estimated sedimentation rate. Due to the significantly younger age produce by SUERC-28872 compared to the established compared to the KOT age and possible contamination at the site it was decided to exclude SUERC-28872 from the age depth model.

Figure 4.4: Age depth model for Lake Mudgie using the calibrated ages. The KOT age is in red. The dotted lines are extrapolated using the 129 yrs/cm sedimentation rate estimated between SUERC-28870 to SUERC-28871. This provides an extrapolated age of 4130 yrs BP for the top of the section analysed.

4.4 Results

4.4.1 Pollen stratigraphy

The pollen diagram from the Lake Mudgie core is shown in Figure 4.5. This figure includes only the main pollen types, see appendix C for the whole pollen data set. The core stratigraphy and radiocarbon dates are summarised alongside the diagram. The LOI and PSA results and temperature reconstructions are also shown next to the pollen diagram. The sedimentological results will be described later in sections 4.4.2 and 4.4.3. The temperature reconstructions are described in Section 4.4.4. As mentioned above, the time frame of 30-8 ka was chosen for this project and therefore only the bottom 3 cores of the site were analysed. Initially a 10 cm preliminary sample interval was used down the core. This produced a skeleton diagram which was used to decide areas where the higher resolution pollen analysis would take place. Below 600 cm there were insignificant amounts of pollen in the samples suggesting the site is of Larrikins I age, as if it were older then pollen would continue to have been deposited. Due to the low pollen concentrations the analysed was restricted to the depths 450-600 cm.

CONISS stratigraphic cluster analysis, available in the TG View program, was used to group the pollen percentage data (see appendix A for the cluster analysis results). The cluster analysis was only carried out on the total dryland pollen (TDP). The pollen assemblage zones for the Lake Mudgie sequence generally correspond to the biostratigraphic zones defined by the cluster analysis. However, the zones chosen were restricted to assemblage changes in the pollen composition that are considered to be of palaeoecological and palaeoclimatic significance. The main pollen zones are numbered LM1-LM4. Within these zones subsequent changes in pollen composition are defined as a sub-zone (Figure 4.5).

4.4.1.1 Pollen assemblage zones

There are four main pollen assemblage zones splitting the stratigraphy with a number of sub-zones contained within these.

Zone LM1: 600-555 cm (c. 26.6-17.6 ka):

LM1 is dominated by non-arboreal pollen; in particular there are high percentages of Poaceae. There are also significant amounts of shrub species, including *Hebe*, Asteraceae, *Coprosma* and *Myrsine*, although the proportion of shrubs declines towards the top of this zone. *Fuscospora* tree pollen was consistently present throughout the zone averaging 5-10% abundance. The high resolution pollen analysis indicates a small, but significant, composition change with an increase in herb dominance and a reduction in the shrub component from 586-559 cm. These changes are distinctive in the pollen record and the results of the cluster analysis support the chosen zones. This forms the basis for splitting the zone into three sub-zones (Figure 4.5)

Sub-zone LM1a (600-583 cm: c. 26.6- c. 23.2 ka):

This sub-zone is dominated by grass species with a higher percentage of small Poaceae pollen grains in comparison to the Poaceae >40 µm. Other herb species are also present, albeit in small percentages at the base of this zone, including Apiaceae, Brassicaceae and *Kelleria*. A significant amount of *Hebe*, reaching a maximum abundance of 25%, is present throughout this sub-zone, although levels decrease towards the top of the boundary. Other shrub elements include *Coprosma*, Asteraceae and, to a lesser extent, *Myrsine*. These species contribute to a shrub abundance ranging 40-15% in this zone. *Fuscospora* is also present in low percentages of around 5%. There is also a low amount of monolete spores, *Lycopodium* type and some wetland species (*Gonocarpus*, *Empodisma* and Cyperaceae).

Sub-zone LM1b (583-559 cm: c. 23.2- c. 18.4 ka):

This zone is still dominated by Poaceae, with grains under 40 µm consistently above 65% abundance and increased amounts of Poaceae >40 µm pollen averaging between 10-15%. However, Poaceae peaks at the top of this sub zone with 29%. The tall tree component, mostly *Fuscospora* remains at a similar

abundance as in the previous sub-zone. However, the percentage of overall shrub composition decreases to 10-15%. The most significant reduction is the abundance of *Hebe*, declining to 0%. *Coprosma* and *Myrsine* also decline to negligible amounts, although *Coprosma* abundance starts to increase towards the top of this sub-zone. The percentage and composition of fern spore and wetland species is similar to the previous sub-zone.

Sub-zone LM1c (559-555 cm: c. 18.4-17.6 ka):

Sub-zone LM1c marks the transition between the distinctive pollen compositions of zone LM1 and zone LM2. This zone is distinguished from sub-zone LM1b by the dramatic decrease in herb taxa, particularly Poaceae from 70% abundance down to only 10% towards the top of this zone. *Fuscospora* abundance also declines to trace levels. In contrast the proportions of *Coprosma* shift from 15% to 65% and *Myrsine* increases from 7% to 26% towards the top of this sub-zone. Asteraceae abundance remains the same. Coinciding with this increase in shrubs abundance is an increase in monolete spores from 5% to 18%

Zone LM2 (555-520 cm: 17.6 - c. 13.1 ka):

In zone LM2 the TDP is dominated by shrub species, in particular *Coprosma* and *Myrsine* especially in the lower to mid zone location. A pattern of successive introduction of other shrub species occurs up through the zone. The top of this zone begins to receive a large input of tall tree species; *W. racemosa* and to a lesser extent *Prumnopitys ferruginea*. This is the start of a transition from shrub to hardwood/podocarp vegetative communities. This zone also experiences a significant increase in fern, tree fern and wetland components.

Sub-zone LM2a (555-544 cm: 17.6 - c. 16.2 ka):

This sub-zone is dominated by *Coprosma* and *Myrsine* with both genera reaching a maximum abundance of 50-60%. *Coprosma* is the dominant species throughout the majority of this sub-zone, although *Myrsine* increases in abundance towards the top. Asteraceae remains at a similar abundance as in the previous zone. There is a small amount (10%) of Poaceae at the bottom of this sub-zone, a remnant of the alpine dominated vegetation which was prevalent in LM1, but it quickly disappears from the record. The shrub *Pseudopanax* also begins to make an

appearance in this sub-zone. The abundance of monolete spores continues from the previous sub-zone with a similar abundance of 15%, but declines towards the top of this zone coinciding with an increase in *Histiopteris*.

Sub-zone LM2b (544-530 cm: c. 16.2 - c. 14.4 ka):

The TDP abundance is very similar to the previous sub-zone with a high abundance of *Coprosma* which reduces up through the sub-zone, giving way to dominance of *Myrsine*. The presence of *Pseudopanax* continues with the addition of *Pseudowintera* and small amounts of *Griselinia*. The largest noticeable change in the sub-zone relates to the aquatic, tree fern and fern components. *Cyathea smithii* and *Leptospermum type* make an appearance at the base of this sub-zone. The *C. smithii* increase is more significant rising from 0-14% by the top of this zone. Similarly, *Isoetes* also make an appearance in this zone, and increases markedly from 0-86% at the top of this zone. A small peak in *Dacrycarpus dacrydioides* abundance occurs at the top of this sub-zone (532-530 cm) reaching a maximum 5% at 532 cm followed by a decrease to below 1% for the remainder of pollen zone LM2.

Sub-zone LM2c (530-520 cm: c. 14.4 - c. 13.1 ka):

This zone shows the continuation of the upwards decrease in abundance of *Coprosma*, joined by reductions in *Myrsine*, *Pseudopanax* and *Pseudowintera*. However, this zone also marks a transition of new shrub/short tree species reaching either their maximum abundance (*Griselinia*) or first consistent appearance in the record (*Elaeocarpus*). This zone also marks a significant change via an increase in tall tree species. *Libocedrus*, present in the previous zone in small amounts, increases its abundance and is consistently between 5-10% and *Weinmannia* reaches its maximum abundance of 38% at 523 cm, but decreases to 22% at the sub-zone boundary. *Prumnopitys ferruginea* also reaches a peak abundance of 20% at 522 cm.

The wetland taxon, *Leptospermum type*, reaches a peak abundance of 26% and the macrophyte, *Isoetes* continues from the previous sub-zone, with a high abundance of 80%, decreasing to 66% at 523 cm. A dramatic decrease in *Isoetes* follows and by 520 cm it has an abundance of only 6%. *Cyathea smithii* is

persistent throughout the sub-zone, although at slightly lower amounts compared to the previous sub-zone.

Zone LM3 (520-503 cm: c. 13.1 - c. 10.9 ka):

This zone represents a transition from shrub and early tree appearance to a dominance of the podocarp *Dacrydium cupressinum*. *Coprosma*, *Myrsine*, *Griselinia* and *Elaeocarpus* all decrease in abundance to near 0% by the top of this zone. Likewise *Libocedrus*, *W. racemosa* and *P. ferruginea* also decrease significantly to 5%. In contrast, *D. cupressinum* increases dramatically from near 0% to above 80% at the top of this zone. However there isn't a consistent or smooth increase in *D. cupressinum*, rather the pattern of increase is interrupted. Associated with this 'interrupted' *D. cupressinum* increase is a reoccurrence, albeit in small abundances, of Chenopodiaceae, *Phyllocladus* and *Fuscospora*, which were previously present together in Zone L1. This zone is marked by a decrease in *Leptospermum* type but an increase in *Cyathea smithii* and small abundances of *Dicksonia squarrosa* and *Cyathea dealbata*.

Zone LM4 (503-450 cm: c. 10.9 - c. 4.2 ka):

The TDP is dominated by tall trees and in particular by the tree species *Dacrydium cupressinum* that has an abundance ranging from 85% to above 90%. *W. racemosa* and *Prumnopitys ferruginea* are present throughout this zone, albeit in low percentages (5-2%). All the shrub species have very low abundance if present at all. The tree ferns are still present, although *Cyathea smithii* reduces in abundance. *D. squarrosa* and *C. dealbata* are also present in similar percentages as in the previous zone. There is a very small presence, if at all, of fern spores.

Figure 4.5: Lake Mudgie pollen percentage diagram, including core stratigraphy, summary diagram, total carbon content, particle size percentage and pollen-climate reconstructions. Radiocarbon dates are summarised in Table 4.1.

Figure 4.5: (continue)

4.4.2 Particle Size Analysis

Sampling for the particle size analysis (PSA) occurred every other half cm, producing a 1 cm resolution with a sample width of 0.5 cm thick. Sampling began at 490 cm and continued down through the core to 600 cm, covering the section of detailed pollen analysis. Throughout this section the silt fraction was the most dominant particle size, mostly 70-80%, however there are variations within the clay and sand component up the core and, of course, this influenced the percentage of silt in the sediment. The maximum sand abundance occurred, between 600-555 cm peaking at 592 cm (pollen assemblage zone LM 1a) and between 583-555 cm (pollen assemblage zone LM1b) and consistently ranging 17-22%. Between 555-523 cm the abundance of sand decrease to 12-15%. At 523 cm the abundance of sand again increases to near previous maximum levels of 20% although the pattern appears to be more erratic. The sand abundance then decreases back to 15% toward the top of the section analysed. In contrast the amount of clay throughout the core is relatively low, consistently having an abundance of 2% or lower from 600-523 cm. However, between 523-510 cm the proportion of clay increases and has a bimodal peak in abundance. The first of the modes is at 521 cm reaching 20% abundance and then reducing to 8% then rising back up to 15 % at 510 cm before again lowering back to 4% where it remains consistently around this abundance to the top of the section analysed. The abundance of silt therefore decreases to its lowest levels of 65-60% between 523-510 cm.

4.4.3 LOI

As with the PSA, sampling for LOI analysis occurred every other half centimetre at a 1 cm resolution, from 600-490 cm. Three main patterns in the record are evident, with minor variations superimposed upon this main pattern. The first noticeable pattern is the stable low LOI present in the sediment from 600-555 cm, corresponding to pollen assemblage zone LM1. The amount in this section increases from 2% up to a small peak of 10% (590 cm) reducing back to 5% and followed by another peak to 15% at 580 cm. After this second peak a consistent, 5-6% LOI is present sediment up to 555 cm. Between 555-518 cm the LOI increases markedly from 8% to 84% (pollen assemblage zone LM2) with a few minor reversals. This section corresponds to LM2. From 518-490 cm the LOI

reaches the maximum content of 87% and levels off after 520 cm with slight decreases in abundance to 80% towards the top of this record. This corresponds to pollen assemblage zones LM3 and LM4. One sample, at 496 cm, has a carbon content of only 39% but the next samples increase in carbon content back to 80-82% suggesting that this is an erroneous result.

4.4.4 Temperature reconstructions

Fig 4.5 displays the MAT reconstruction comparing the two models, PLS-C3 and W-MAT. Overall the pattern of temperature change is similar between models although W-MAT consistently produces higher temperature estimates than PLS-C3 which exhibits greater variability. The initial predicted MAT in the PLS-C3 model is 5.6 °C increasing to 6.8 °C remaining around this level until the 583 cm. Temperature decreases and ranges 5.4-6.3 °C with an average of 5.7 °C between 583-555 cm. Temperatures increase rapidly after 555 cm reaching a maximum of 10.8 °C at 548 cm. The temperature slowly declines down to 9.5 °C at 523 cm followed by a rapid decline in temperature between 523-510 cm reaching a minimum 6.7 °C and oscillating between 6.7-8.2 °C. At 510 cm temperatures reach 8.9 °C and then rise consistently above 9 °C, the exception being at 459 cm where a temporary MAT dip to 8.5 °C occurs.

W-MAT initially shows temperatures oscillating between 7.3-8.4 °C between 599-583 cm followed by a period of consistently cooler temperatures ranging 7.2-7.8 °C and averaging 7.5 °C between 583-559 cm. Temperature begins to increase at 559 cm but less rapidly than in the PLS-C3 model, reaching a maximum 10.4 °C at 544 cm and plateaus until 523 cm. A minor decline in temperature occurs between 532-510 cm declining to 9.3-9.7 °C. This is followed by an increase in temperature reaching 11.3 °C at 500 cm and remaining above 10 °C for the majority of the top section of the core. However, the MAT changes reconstructed by both modelled lie within the error margin and indicating a limited temperature, if any, can be predicted for this site.

4.4.5 Principal Component Analysis (PCA)

PCA, using CANOCO version 2, was initially carried out on just the pollen data, with subsequent runs including the LOI and the % of sand, silt and clay particle sizes in each sample to explore relationships between the pollen assemblages and changes in core sedimentology. Although pollen and PSA samples were at the same depths, LOI samples were consistently offset by 0.5 cm, due to the sampling strategy. Therefore an average was taken of the two LOI samples which bordered the sampled depth of the PSA and pollen analysis.

The results are presented in two sections with and without the sedimentological data. The eigenvalues for each species are summarised (Figure 4.6 and Table 4.2) followed by a description of the sample eigenvalues (Figure 4.7 and Table 4.3). The results are then considered in context with the pollen diagram and pollen assemblage zones.

4.4.5.1 Palynological PCA results

Figure 4.6 is a biplot of species eigenvalues for axis 1 and axis 2, illustrating the relationships of different pollen types. Three groups are apparent, with one group sub-divided further into two sub groups.

Group 1 contains taxa with positive eigenvalue on axis 1 and negative to near 0 values on axis 2. *Phyllocladus* and *Gonocarpus* are the two exceptions having more positive values along axis 2. This group represents vegetation present in zone LM1 including both the >40µm and <40 µm Poaceae taxa and a further four herb taxa (see Table 4.2). The shrub taxa Asteraceae, *Hebe*, *Phyllocladus* and the tall tree species *Fuscospora* are included in this group as well as four wetland taxa (Table 4.2).

Taxa in group 2 have negative eigenvalues along both axes. This group comprises seven shrub taxa, including *Coprosma*, *Myrsine*, *Griselinia*, *Pseudopanax* and *Pseudowintera*, and four fern species. These taxa are more prevalent in pollen assemblage zone LM2.

Group 3 contains taxa with negative eigenvalues on axis 1 and positive values on axis 2. This group can be further divided into 3a and 3b, where sub-group 3b have more positive eigenvalues on axis 2. Sub group 3a contains species which were prevalent in pollen assemblage zone LM2c and LM3. Four tall tree species *W. racemosa*, *Libocedrus*, *Metrosideros* and *Elaeocarpus* are included in this group along with the tree fern *Cyathea smithii* and wetland taxon *Leptospermum type*.

Sub-group 3b is restricted to just five taxa including three tall tree species; *Dacrycarpus dacrydiodes*, *Dacrydium cupressinum* and *Prumnopitys ferruginea*, and two tree fern taxa. These taxa are restricted to the top of the core in pollen assemblage zone L3 and L4. These species are also representative of the lowland temperate forest that is prevalent in the contemporary Westland region.

Group	Tall tree	Short tree/ shrub	Herb	Wetland	Tree fern	Fern
1	<i>Fuscospora</i>	<i>Hebe*</i> <i>Asteraceae*</i> <i>Phyllocladus</i>	Poaceae <i>Kelleria</i> Chenopodiaceae Brassicaceae <i>Gentiana</i>	Cyperaceae <i>Gonocarpus</i> <i>Empodisma</i> <i>Gunnera</i>		
2		<i>Coprosma</i> <i>Myrsine</i> <i>Griselinia</i> <i>Muehlenbeckia</i> <i>Pseudopanax</i> <i>Pseudowintera</i> <i>Schefflera</i>				<i>Histiopteris</i> <i>Hypolepsis</i> <i>Isoetes</i> Monolete spore
3a	<i>W. racemosa</i> <i>Libocedrus</i> <i>Metrosideros</i>	<i>Elaeocarpus</i> <i>Rubus</i>		<i>Leptospermum type</i>	<i>C.smithii</i>	<i>Hymenophyllum</i> <i>Phymatosorus</i>
3b	<i>D.dacrydiodes</i> <i>Dacrydium cupressinum</i> <i>P.ferruginea</i>				<i>C.dealbata</i> <i>D.squarrosa</i>	

Table 4.2: Taxa comprising the Lake Mudgie PCA groups with their represented vegetation communities.

Figure 4.6: Lake Mudgie PCA results for selected taxa plotted for axis 1 and axis 2. Circles indicate arbitrary chosen group clusters. **Group 1;** P<40 - Poaceae <40µm, N.f - Fuscospora, P>40 - Poaceae >40µm, Gu - Gunnera, Em - Empodisma, As - Asteraceae, Ge - Gentiana, He - Hebe, Ph - Phyllocladus, Go - Gonocarpus, Cy - Cyperaceae, Br - Brassicaceae, Ap - Apiaceae, Ch - Chenopodiaceae, Ke - Kellera. **Group 2;** Mp - Myriophyllum, Mo - monolete spore, Sc - Schefflera, Mu - Muehlenbeckia, Hi - Histiopteris, Hy - Hypolepsis, Pw - Pseudowintera, Gr - Griselinia, Is - Isoetes, Ps - Pseudopanax, My - Myrsine. **Group 3a:** Pt - P. taxifolia, Hm - Hymenophyllum, Me - Metrosideros, Ph - Phymatosorus, Ru - Rubus, El - Elaeocarpus, Le - *Leptospermum* type, Li - Libocedrus, We - W. racemosa, C.s - C. smithii. **Group 3b:** Da - Dacrycarpus dacrydiodes, C.d - C. dealbata, P.f - P. ferruginea, D.s - D. squarrosa, D.c - D. cupressinum.

4.4.5.2 Sample PCA results

Figure 4.7 is a biplot of sample PCA scores for axis 1 and axis 2. Lake Mudgie samples have been split into six sample groups with some samples appearing to be 'transitional' between groups. Arrows on the figure indicate the direction the

results have been described in and indicate the linkage between the groups. The groups are described below.

Group 1 contains samples with positive eigenvalues on axis 1 and negative values on axis 2. These samples span the depth 597-585 cm and are located at the base of the core. The samples correspond to pollen assemblage zone LM1a.

Group 2 samples have a similar values to Group 1 on axis 2 and higher positive values on axis 1. The sample depths span 584-559 cm and correspond to pollen assemblage zone LM1b, although there is one anomalous sample which is the bottom most core sample at 600 cm. Three transitional samples, 558-555 cm, join group 2 to group 3. These samples occur in stratigraphic order and has the pattern were the shallower the sample is in core a lower PCA score on axis 1 it received. These transitional samples correspond to pollen assemblage zone LM1c.

Group 3 samples have negative eigenvalues ranging between -0.7 to -0.5 on both axes. This group of samples span the depths 554-527 cm and corresponds broadly to pollen assemblage zone LM2a.

Group 4 contains samples with the lowest eigenvalues (-1 to -1.5) along both axes. The samples in this cluster range the depths 535-526 cm and closely match the pollen assemblage zone LM2b. Three transitional samples join group 4 to group 5, spanning the depths 524-522 cm. As with the previous transitional samples, these lie in stratigraphic order receiving increasing PCA scores along both axes, particularly along axis 2, with decreasing sample depth. These samples correspond to pollen assemblage zone LM2c.

Samples in group 5 have positive eigenvalues on axis 2 and negative values on axis 1. The samples align in a horizontally linear cluster with similar axis 1 values and PCA scores spanning from 0.8-1.7 on axis 2. The samples included in this group range from 521-509 cm and match the pollen assemblage zone LM3.

The final group, group 6, comprises the top six samples of the core. This cluster has similar negative eigenvalues on axis 1 but larger positive axis 2 values. The

cluster aligns in a horizontal linear fashion along axis 2 that appears to extend the trend established by group 5. The group covers the depths 499-450 and corresponds to pollen assemblage zone LM4. Comparisons between the CONISS and PCA (Table 4.3) show a close similarity between the depth of the groups and the pollen zones. This gives greater weight to the cluster analysis and justifies the placement of the chosen pollen zones.

Cluster Analysis		PCA	
Pollen zone	Depth (cm)	Group	Depth (cm)
LM1	600-555		
LM1a	600-583	1	597-585
LM1b	583-559	2	584-559 (600)
LM1c	559-555	Transition	558-555
LM2	555-520		
LM2a	555-544	3	554-537
LM2b	544-530	4	535-526
LM2c	530-520	Transition	524-522
LM3	520-503	5	521-509
LM4	503-450	6	499-450

Table 4.3: A table comparing the depths of the pollen assemblage zones, carried out by the CONISS stratigraphic cluster analysis, to the depths of the sample groups, identified using PCA statistical method.

Figure 4.7: Biplot of PCA sample scores for axis 1 and axis 2. Each of the 6 groups has been labelled and the depths displayed indicate the depth range or core depth that each sample or group represents. These groups were chosen arbitrarily. The arrows indicate the stratigraphic order of the groups with the arrows pointing in the direction of decreased core depth.

Figure 4.8 plots the sample PCA score for axis 1 and axis 2 along a depth gradient. Unlike the biplot which plots the two axes scores for each sample (Figure 4.7), plotting the axis sample scores against depth can indicate what factors are influencing the individual axis over time, via comparisons with other results. The results are described below.

Axis 1 initially contains high PCA scores oscillating between 1.5-0.9 from the core base to 560 cm. This is followed by a large decrease in eigenvalues to 0.5 at 552 cm. A plateau of 0.5 values continues until 535.5 cm, followed by a further decline in value to -1.5 to -1 between 535.5-525.5 cm. A small increase in sample value to -0.5 occurs after 525.5 cm and this value remains constant to the top of the core. The pattern in the PCA score at the base of the core and the timing of the

reduction in values is very similar to the Poaceae curve, however the similarities between the two records stop here.

Axis 2 initially has a consistent value of 0.5 from the core base until 558 cm. The pattern in the PCA score is then a decrease to -1 until 536 cm, followed by a further depression to a minimum of -1.5 at 530 cm. The values slowly increase to 0.5 at 525 cm. The pattern and values between 558-525.5 cm is the same as that recorded in axis 1 and is the section of the core dominated by shrub taxa. A dramatic increase in PCA score occurs above 525.5 cm, reaching a value of 1 by 520.5 cm. From 520.5 cm to the top of the core is a section of high PCA score which lie consistently between 1-2. The pattern in the upper section of the core (from 525-450 cm) is similar to the pattern in total tree abundance, particularly the pattern of *Dacrydium cupressinum* abundance.

Figure 4.8: Rough diagram with the PCA sample scores of axis 1 and 2 plotting with depth. TDP, LOI and PSA have been added to show if there are any similarities between the records and the axes scores.

4.4.5.3 PCA results combining the pollen and sediment data

The addition of the sedimentology to the PCA enables comparisons between species composition and sediment characteristics (Figure 4.9). Sand size particles lie in group 1 and are associated with alpine vegetation communities. Fine particle size and LOI have similar PCA score as the tall tree and tree fern species. In the contemporary environment these are species located in the lowland temperate rainforest. Silt size particle appear to be an anomaly and are ubiquitous, showing no particular association with any vegetation group. This is probably because silt levels remained between 70-80% throughout the section analysed and therefore do not relate to any vegetation groups in particular.

Figure 4.9: Lake Mudgie PCA results for selected taxa and sedimentology results plotted for axis 1 and axis 2. Circles indicate arbitrary chosen group clusters. Refer to Figure 4.6 for species names.

4.5 Discussion

This section interprets the pollen and sedimentological results in terms of inferred vegetation communities and then climatic reconstruction. The chronology, derived from the three radiocarbon dates and the tephra layer, is discussed and then applied to the vegetation record. The discussion concludes with a summary of Lake Mudgie's results and interpretations where comparisons with other Westland and New Zealand records are made.

4.5.1 Vegetation and climatic interpretation

The vegetation and climatic reconstruction were inferred using contemporary vegetation ecological characteristics (Wardle 1975, 1977, 1991; Moar 1970), palaeoenvironmental interpretations of Westland pollen data (McGlone 1995, Moar and Suggate, 1996; Vandergoes and Fitzsimons, 2003; Newnham *et al*, 2007a), and the temperature reconstruction produced by the ELR. The transfer function results are also tentatively mentioned, but limitations in the method limits the interpretation (see Section 8.2.4). The reconstruction is dominantly based on the dryland flora and tree fern communities. However, some wetland and fern spores were included in the interpretations as these species can also provide insights into the climatic and ecological changes on a more local scale. The pollen transfer function also used a limited number of non-dryland pollen types (Wilmshurst *et al*, 2007), which are also included in the discussion to aid the interpretations. The inferred ages produced by the age-depth model have been applied to each pollen zone discussed.

The base of Lake Mudgie sequence records a change from grassland dominated flora with a mosaic of shrubs (LM1a), indicative of cold climates, to a further increase in grassland dominated vegetation community (LM1b). This suggests a cooling trend leading to a period of extreme cold. This is followed by sharp transition from alpine dominant to shrubland dominant taxa and a successional pattern from shrubs to mixed lowland broadleaved hardwood forest follows (LM1c-LM4). These developments are inferred to represent a change to warmer, wetter conditions at the site although still cooler and drier than the contemporary climate. Finally, the top of the sequence indicates establishment of tall tree communities

associated with warm, wet, temperate climate regimes, similar to that experienced on the west coast today.

The following provides a more detailed interpretation of the vegetation and climate interpretations made from the pollen assemblage and sedimentology of the sequence. The sample PCA result (Table 4.3) indicates a good correspondence between the two statistical methods and provides greater validity to the zonation of the pollen diagram (Figure 4.5).

4.5.1.1 Zone LM1 vegetation reconstruction (c. 26.6 - c. 17.6 ka)

The vegetation in zone LM1 is dominated by grassland and herbfield communities indicating a persistence of alpine vegetation throughout. However, the base of this zone, sub-zone LM1a, has an increased abundance of shrub species including the frost tolerant *Hebe*, Asteraceae, and to a lesser extent *Coprosma*. However, it should be noted that these taxa include a wide range of plant forms including herb types. In the contemporary alpine landscape these plants occupy sheltered sites among the grassland-herb field and this seems a likely habitat in this zone as well (Wardle, 1991). In addition there are low amounts of *Phyllocladus* and *Myrsine* which were probably confined to favourable sites which in the contemporary environment tend to be north facing and areas with poor drainage (Wardle, 1977). The dominance of grass species is characteristic of the alpine area, which in the modern landscape is restricted to 1200 m.a.s.l and above (Moar, 1970; Wardle, 1977). However, the presence of the shrub species, occasional tree pollen suggests that a diverse vegetation community existed during this cold period.

A further increase in grassland vegetation composition occurs in subzone LM1b suggesting a further increase in alpine vegetation and a period of extreme cold. Coinciding with the increase of grassland is the decline of *Coprosma*, *Myrsine* and *Hebe*, indicating that local conditions were no longer suitable for extensive shrub growth. Asteraceae abundance remains at a similar level as in the previous subzone, and this taxon probably comprised herbs growing amongst the grassland as it does today (Wardle, 1991). The LOI is at its lowest abundance during this sub zone and indicates a lack of biomass and productivity of grassland communities. Likewise the maximum abundance of sand size particle during this sub-zone is

indicative of reduced vegetative cover and an increase in erosion. An increase in inorganic material to the sites, linked to erosion and the transport of inorganic material to the site, can also dilute the organic carbon and thus producing low LOI results.

Towards the top of this zone, LM1c, a significant increase in *Coprosma* and *Myrsine* is accompanied by a sharp decline in grassland abundance. This indicates a transition in vegetation composition from an alpine dominated landscape to an abundant shrubland community and is suggestive of a warming in the climate. This vegetation transition is accompanied by an increase in total carbon content of the sediment suggestive of an increase in biomass contribution by the shrub taxa.

4.5.1.2 Zone LM1 climate interpretation

The vegetation composition of zone LM1 is characteristic of the high alpine zone in the contemporary landscape. The lack of any significant shrub or tall tree species indicates insufficient heat to enable these woody taxa to grow in this environment. In the south Westland region the contemporary lower limit of the alpine zone is at c.1200 m.a.s.l (Wardle, 1991), c.1050 m above the current Lake Mudgie core site located at 143 m.a.s.l. This implies the alpine zone descended to at least this altitude during the LGCP. Using the environmental lapse rate of 0.5 °C/100m estimated MAT depression during zone LM1 is >5.4 °C. However the higher abundance of shrub species, in particular *Myrsine*, which is prevalent in the contemporary sub-alpine/montane zone could represent a period of milder conditions in LM1a. Marine records from sites offshore from the Westland and estimations from other records suggest the maximum temperature depression around 5 °C and support the estimated MAT depression at Lake Mudgie (McGlone and Bathgate, 1994; Pelejero *et al*, 2006; Barrows *et al*, 2007).

The existing temperate rainforest and montane species in the Westland region are dependent upon high precipitation and are intolerant to water stress caused by drier periods (Wardle, 1991). The lack of species representing these vegetation communities may represent a reduction in precipitation at the timing of zone LM1. However, the modern alpine environment has high precipitation values, although a

large proportion of this is received in the form of snow. It is probable that the alpine conditions experienced at the site, which today lies considerably below the modern alpine zone, during the much colder LGCP would have had a greater proportion of precipitation in the form of snowfall compared to modern conditions. The short growing season caused by the cold climates and greater snowfall as a proportion of total precipitation would have inhibited the growth of shrub and tree species (Wardle, 1991).

The increase in shrub species *Coprosma* and *Myrsine* in sub-zone LM1c combined with a significant decline in grassland abundance, point towards a sudden increase in temperature causing an expansion of shrub species into the site. Vegetation that is likely to respond to an increase in precipitation, such as *Weinmannia* or *Dacrydium*, made no appearance in this sub zone and implies that the main climatic change was an increase in temperature or reduction in frost severity.

The pattern of inferred temperature based on the quantitative pollen/climate modelled reconstructions suggests the LGCP can be split into two periods corroborating with the qualitative evidence. The initial period spanning 26.6-23.2 ka is characterised by 'milder' (approximately 1 °C warmer) temperatures compared to the period 23.2-18.4 ka which has consistently colder temperatures. The estimated MAT depression predicted by the models during the LGM range 6.1-5.5 °C in the PSL-C3 model and 4.2-3.8 °C in W-MAT, the latter of which is less than the minimum estimated temperature reduction based upon the ELR. The predicted modelled results will be discussed further in Section 8.2.4. Climate warming begins at 18.4 ka and is marked by a period of rapid warming in both records although the temperature increase is less in W-MAT model compared to PLS-C2 model (refer to Section 4.4.4 and Section 8.2.4).

4.5.1.3 Zone LM2 (c. 17.6 - c. 13.1 ka)

Zone LM2 is characterised by a continuously increasing shrub component and the disappearance of grassland leading to the evidential expansion of shrubland becoming the dominant vegetation community in the landscape. This is indicative of climate warming. A vegetation successional pattern occurs with the shrubland giving way to low lying broadleaved/hardwood forest and an expansion of tree ferns being an important component of the sub-canopy. This transitional zone experiences continued rise in LOI in the sediment indicating a consistent increase in plant productivity and biomass, confirming the changes from grassland to, eventually, a multi tiered forest community. The pattern of vegetation change in zone LM2 implies a continuous increase of temperature and the climate becoming warmer and wetter towards the top of this zone. However the continued lack of *D. cupressinum* in the assemblage suggests the climate was still not wet or warm enough for the presence of this species. However, towards the top of LM2 soil development also influenced some successional vegetation patterns. For instance, the appearance of *Dacrycarpus dacrydioides* suggests newly emerged, high nutrient, soils at the site were present. And possibly relates to the terrestrialsation of the lake. However it does not reach a high abundance and its peak appearance is short-lived, maybe representing a reduction in soil nutrients combined with climatic change which producing conditions more preferable to *Dacrydium cupressinum* (refer back to section 4.1.2)

4.5.1.4 LM2 vegetation reconstruction

The high values of *Myrsine* and *Coprosma* in sub-zone LM2a and the small abundance of Poaceae implies that shrubland had taken over grassland as being the dominant vegetation community in the region by the start of this sub-zone. Grassland probably became restricted to higher altitude locations. The lack of any increase in Poaceae after zone LM1 indicates that once grassland disappeared from the landscape it was not a significant component in the local vegetation again and implies a continuous climate warming during this sub-zone.

Asteraceae remains at similar levels as it occurred in the previous zone suggesting that it was still a significant component in the vegetation. Like *Coprosma* and *Hebe*, Asteraceae includes both herb and shrub forms which are

difficult to differentiate in the pollen diagram. However, its continued appearance during the shrub zone, when no other herb species are present including grass, suggest predominantly Asteraceae shrubs are represented in LM2a. During this zone a significant increase in Monolete spore abundance indicates ground ferns became more prevalent in the landscape and may indicate disturbance of the shrubland.

Sub-zone LM2b continues to be dominated by the shrub taxa *Coprosma* and *Myrsine*, however *Coprosma* abundance is declining and *Myrsine* increasing to become the overall dominant species. This change in composition may indicate an increase in temperature and/or precipitation. However, in the contemporary successional pattern of shrub development, *Coprosma* tends to be an important primary stage taxon with *Myrsine* developing and often becoming the dominant shrub after the shrubland has matured (Wardle 1991). Therefore the vegetation change recorded in LM2b may represent a successional vegetation pattern opposed to a climatically induced change.

Sub-zone LM2b assemblage also include *Pseudopanax* and *Pseudowintera* which are small tree/shrub taxa indicating a more diverse floristic composition. Concurrently, there is a sudden increase in *Cyathea smithii* abundance which remains persistent throughout the rest of this zone. The appearance of *C. smithii* in a shrubland environment rather than a forested landscape points towards its persistence as a pre-forest successional element. In the contemporary Westland landscape *C. smithii* occurs in gullies and grows around the bushland perimeter and these are likely to be the areas it inhabited during LM2b (McGlone and Bathgate, 1983). The appearance of *Isoetes* indicates the presence of open water, however, it also occurs during a stratigraphic transition in the core where an increase in organic material and a change from grey to brown silts occurs. This suggests the lake was becoming more terrestrialised with water level shallowing and eventually, in this zone, providing the optimum conditions for *Isoetes* growth. Therefore the occurrence of *Isoetes* here points to a shallowing of the lake (see Section 7.6.2 for more detail of the ecological limitations of *Isoetes*).

Towards the top of zone LM2 (sub-zone LM2c) a significant decrease in *Coprosma* and *Myrsine* abundance occurs coincidently with a peak in the broadleaved, small tree taxa *Griselinia* and *Elaeocarpus*, and a rise in the tall tree taxa *W. racemosa*, *Libocedrus* and *P. ferruginea*. These changes indicate expansion of low lying broadleaved hardwood/podocarp forest in the area. *Coprosma*, *Myrsine* and the low abundance of *Pseudopanax* and *Pseudowintera* pollen in the record suggests these species were forming an understory to the main forest canopy. These inferred communities are similar to montane forest vegetative composition in the modern Westland environment today which lies between an observed altitude of 850-650 m.a.s.l (Wardle, 1977, 1991). The increase in *W. racemosa*, a drought and frost sensitive taxa, indicates that temperature and precipitation levels were high enough (>1000 mm/yr) to allow this species to become established. The consistent presence of *Cyathea smithii* now probably part of the small tree tier is also indicative of moist environments. The small rise in *Dacrycarpus dacrydioides* reflects a successional pattern in reforestation of the site at the top of sub-zone LM2b.

The presence of *Libocedrus*, which has a high freezing tolerance of -13 °C and is often associated with cooler climates, may be indicative of cold air pooling which can occur in the valley basins along river banks. *Libocedrus* in the contemporary landscape is shade intolerant and unable to grow in dense canopy (Veblan and Stewart, 1982), and therefore needs to be in a canopy free zone. This may represent a catastrophic disruption in the landscape such as a landslide or earthquake in the area causing tree felling and leading to canopy gaps. However, and more likely, sinking cold air will inhibit the growth of the frost intolerant species, including the canopy producing trees species, thus allowing *Libocedrus* growth. *Libocedrus* is also found to be more prevalent on poorly drained surfaces in the lowland and montane environments. This indicates the site probably was poorly drained. The presence of *Leptospermum type*, that has peak abundance in sub-zone LM2b and c, is also indicative of poorly drained areas as it commonly forms a shrubby tier on boggy sites (Wardle, 1991).

The increase in *Prumnopitys ferruginea* is likely to represent the start of the development of tall podocarp forest in the area. However, the relatively low levels

of this species suggest it was restricted in the landscape. This species was probably restricted to lowland terrace or hill slopes where it is commonly found today (Reif and Allen, 1988; Wardle, 1991). The disappearance of *Isoetes* indicates there is no longer open water at the site and coincides with a visual stratigraphic change from light brown silts, still probably representing lake deposits, to dark brown organic silts suggesting terrestrialisation of the lake environment has occurred.

4.5.1.5 Climatic interpretation

Temperature increase is indicated at the base of this zone by the increased dominance of *Coprosma-Myrsine* shrubland and the introduction of other shrub species over the previously dominant alpine communities. The shrub communities are representative of the contemporary sub-alpine zone which lies between 800-1200 m.a.s.l (Moar, 1970; Wardle, 1977). The lack of tall podocarp or broadleaved trees and species characteristic of moist environments could represent a reduction in precipitation compared to the present. However, the lack of tall forest trees more likely represents a temperature control.

Towards the top of this zone, increases in tall forest and broadleaved taxa *Griselinia*, *Elaeocarpus*, *Weinmannia racemosa* and *P. ferruginea* indicate that temperatures and precipitation were now high enough (at least 1000 mm/yr) to support the development and expansion of a low hardwood-podocarp forest. The increase of species characteristic of mild moist equable climates such as *Weinmannia racemosa*, which demands high precipitation levels and is sensitive to both water stress and severe frost (McGlone *et al*, 1994), provides support for this interpretation. However the lack of abundance of *Dacrydium*, the species which is dominant in the contemporary forest, suggests that temperature and precipitation may not have been as high as at present. Therefore it is likely that this low hardwood-podocarp forest grew in wetter and warmer conditions than the previous sub-zones, but it was drier than present conditions in the region. The composition of sub-zone LM2b and c is representative of the montane vegetative zone, which in the contemporary landscape is restricted to 600-800 m.a.s.l (Moar, 1970; Wardle, 1977). Using the ELR of 0.5 °C/100 m, a minimum temperature depression of 2.3 °C is implied. The average MAT depression predicted in the

PLS-C2 model for the early LGIT is 1.8 °C and in the W-MAT 1.5 °C. The two modelled temperatures are broadly similar to the ELR estimates when the temperature error of ± 1.5 °C is taken into account. However, the PLS-C3 estimated MAT depression predicted during the early LGIT is less than predicted during the Holocene (see section 4.4.4), which is unexpected. The methods used to reconstruct past MAT are discussed in more detail in Section 8.2.4.

4.5.1.6 Zone LM3 and LM4 (c. 13.1 - c. 4.2 ka)

This part of the discussion section interprets the results from both LM3 and LM4 together. The zone is characterised by the transition from shrub dominated vegetation communities to podocarp dominated lowland forest, primarily dominated by *D. cupressinum*, although *W. racemosa*, and to a lesser extent *Prumnopitys ferruginea* remain significant. This transition reflects the development of vegetation cover similar to contemporary vegetation in the lowland forest today (Wardle, 1977). However, the increase in *D. cupressinum* dominated forest is interrupted by a disruption in *D. cupressinum* expansion. A significant decrease in *Isoetes* in the previous zone and the increased organic content and brown colouration of the core indicates the near terrestrialsation of the lake and core site and the lack of an open water environment. This could also be illustrated by the decline in *Leptospermum* type, which in the contemporary environment inhabits wetter, boggy environments. The continued dominance of *D. cupressinum* in zone LM4 indicates the development of the temperate rainforest similar to that present in the contemporary lowland Westland environment (Moar, 1970; Norton *et al*, 1986; Wardle 1991).

4.5.1.7 LM3 and LM4 vegetation reconstruction

The abrupt increase of *D. cupressinum* in zone LM3 to become the most abundant species indicates the beginning of podocarp temperate rainforest development. It is likely that this species was more dominant in lowland areas on ridge slopes and terraces below 600 m.a.s.l where it is prevalent in the modern Westland landscape (Franklin, 1968; Wardle, 1977; 1991; McSweeney, 1982). However, the pattern of increasing *D. cupressinum* abundance is not continuous and minor reversals indicating a disruption in the development of the *D. cupressinum* forest. This occurs between 13.2-11.9 ka and is characterised by a reversal in *D. cupressinum*

abundance coinciding with a small increase in *W. racemosa*, *Elaeocarpus* and *Libocedrus*. There is also a small increase in *Fuscospora*, *Phyllocladus*, *Chenopodiaceae* and the wetland taxon *Gonocarpus*, which were more prevalent in zone LM1. The PCA results indicate a close relationship between these species, which were also present in zone LM1 in conjunction with alpine species. A disturbance is also seen in the particle size analysis where a sharp influx of clays enters the lake environment although this begins earlier at 13.4 ka and continues until 11.9 ka. The coincidence of these species with the increase in clay particle size and a disrupted increase in *D. cupressinum* abundance may illustrate a disturbance in the local environment, inhibiting the continued growth of *D. cupressinum* and enabling seral successional plants to colonise the area around the site. Although, it may not necessarily represent climatic cooling.

The low levels of tree ferns, *Cyathea smithii*, *C. dealbata* and *Dicksonia squarrosa* indicates these species form an important component of the small tree tier and persisted amongst the tall podocarp forest. However, *C. smithii* reaches maximum abundance between 13.4-11.9 ka, potentially indicating wetter conditions were experienced at the site coinciding with the disruption in forest expansion producing forest gaps again promoting *C. smithii* growth. This also corresponds with a rapid reduction in *Isoetes* indicating there is no longer a water body at the core site, which does not necessarily support increased precipitation. This is indicative of terrestriation of the lake and is combined with a change in stratigraphy to brown organic silts. The increased terrestrial area at the site would produce a larger area suitable for tree ferns allowing these species to locally expand into the site resulting in increased abundance of tree fern pollen in the pollen record. The LOI in zone LM3 reaches a maximum amount indicating the high biomass and plant productivity of the podocarp dominated forest. The core stratigraphy has an increased organic peat content again indicating the increase in biomass and the suitability of the climate for peat production and points towards the continued terrestriation of the site.

Zone LM4 experienced a continued increase in podocarp dominated lowland forest with *D. cupressinum* remaining the dominant species throughout this part of the sequence. *W. racemosa* and *P. ferruginea* are still persistently present in this

zone albeit in small abundance suggesting they contribute significantly to the canopy vegetation. Small amounts of *Coprosma* and *Myrsine* towards the top of this zone indicates presence of these shrubs in the subcanopy or forest margins. Contemporary Westland forest has *Myrsine australis* and *Coprosma lucida* which forms the main shrub tier under the forest canopy. This vegetation assemblage is very similar to contemporary forest communities in the Westland region. Together these assemblages represent a full development of interglacial temperate rainforest (Wardle, 1977). This idea is further supported by the high organic content of the sediment and further increase in visible organic material in the core representing greater biomass and productivity of the environment.

The consistent, low levels of tree ferns, *C. smithii* and *C. dealbata* suggest that these species formed an important component of the small tree tier and persisted amongst the tall podocarp forest. The relative abundance of tall forest and fern species represent a vegetation and pollen assemblage characteristic of the contemporary lowland forest zone that reaches a maximum altitude of c. 400-600 m.a.s.l (Wardle, 1977) and extends down to sea level.

4.5.1.8 Climatic interpretation

The dominance of *Dacrydium* and the presence of *P. ferruginea* and *Weinmannia* in the regional vegetation indicate the establishment of a stable climate during zone LM3. These species occupy the contemporary warm and wet montane and lowland forests in the area (Franklin, 1968; Wardle, 1977) which strongly suggest the climate at this time remained wet and warm. However, the presence of *Dacrydium* with alpine/sub-alpine zone taxa and the influx of clay particles in between 13.4-11.9 ka suggest some form of environmental disturbance to the progress of lowland podocarp forest growth. Whether or not this represents a period of climate cooling is hard to establish, however the timing of this event, 13.4-11.9 ka, coincides with a LGR timing (Alloway *et al*, 2007) and will be discussed further in sections 4.5.2 and 8.1.51. However, there are other possible scenarios for this disruption which is discussed below.

The rapid expansion of *Dacrydium* dominated forest towards the end of LM2 reflects a marked climatic amelioration. The dominance of *Dacrydium* suggests

that precipitation was high enough to allow the species to expand into the area while temperature became consistently mild enough to enable the establishment of tall podocarp forest. However, it is not possible to interpret the vegetation change solely on the basis of temperature and precipitation rise. As previously stated, the area was suitable for the development of some podocarp forest and wet enough to support extensive tree fern groves since the middle of sub-zone LM2b. One explanation for the late appearance of podocarp forest is that the area may have experienced cold air ponding during LM3. This could be supported by the abundance of *Libocedrus* in LM2c and LM3 which would normally be representative of sub-alpine conditions unless there are cool frost prone areas in the lowland region (Wardle 1977). The presence of this tree, which in turn represents more frosty conditions, could also help explain the more disturbed development of *Dacrydium* in the area. *Libocedrus* is also shade intolerant and grows in recently disturbed areas where gaps in the canopy occur. Again this may be indicative of a disturbance in the substrate, limiting *Dacrydium* growth in these disrupted areas. Likewise the appearance in small abundance of species associated with sub-zone LM1b could also indicate opening of the canopy. Catastrophic events including landslides and earthquakes are likely in the Westland region due to the steep topography and fractured lithology in the presence of the active Alpine fault. Wells *et al* (2001) found evidence that large but infrequent earthquakes have a significant impact on forest vegetative composition causing large scale tree mortality and thus opening up the landscape for seral and shade intolerant species. However, the Westland region is prone to large earthquakes and similar pattern in vegetation would be recognisable throughout the pollen record. It is therefore unlikely that an earthquake was responsible for the disruption in forest development.

The continued dominance of *Dacrydium* and the presence of *Prumnopitys ferruginea* and *Weinmannia* in the regional vegetation indicate the establishment of stable climate during zone LM4. These species occupy the contemporary warm and wet montane and lowland forests in the area (Franklin, 1968; Wardle, 1977) which, along with peat development, strongly suggest the climate at this time remained wet and warm. The dominance of *Dacrydium* and reduction of *Libocedrus* may indicate an abatement of cold air ponding and increased stability

of the substrate. The contemporary *Dacrydium* dominated temperate rainforests of Westland are prevalent up to 400-600 m.a.s.l (MAT depression ranging 0-2.3 °C base on ELR) and experiences a mean annual temperature of c. 11 °C and precipitation c. 5000 mm/yr (Wardle, 1977; 1991). It is likely that the climate experienced at the site during zone LM4 was similar to the contemporary Westland lowland climate and indicates that full climatic amelioration was reached.

The W-MAT MAT reconstruction indicates MAT remained relatively stable throughout the LGIT. The PLS-C3 model suggests MAT was getting cooler during the LGIT with MAT depression increasing up core. The PLS-C3 also predicts warmer climates during the early stages of the LGIT compared to the temperatures experienced during the Holocene. This is the opposite pattern implied by the vegetation record and explanations for this will be discussed in Section 8.2.4. The modelled results also suggest a period of cooling climate between 13.2-11.9 ka, however, the degree of cooling varied between the records. PLS-C3 predicts an average MAT depression of 3.6 °C whilst W-MAT predicts only 1.9 °C. This may suggest a period of cooling representing the LGR which is inconclusive in the pollen record alone (see above and Section 4.5.2). However, there are limitations of the modelled reconstructions which are discussed in Section 8.2.4.

4.5.2 Summary Discussion

Pollen assemblage zone	Vegetation reconstruction	Climatic reconstruction
LM4 (503-450 cm: c. 10.8 - c. 4.1 ka)	<i>D. cupressinum</i> dominated temperate rainforest	Stable warm wet climates with mean annual temperatures c. 11 °C and precipitation c. 5000 mm/yr
LM3 (520-503 cm: c. 13.1 - c. 10.8 ka)	Increase in podocarp dominated lowland forest with episodes of disturbance and poor drainage	Warm, wet climate with the site potentially experiencing cool air ponding and localised frost.
LM2c (530-520 cm: c. 14.4 - c. 13.1 ka)	Broadleaved hardwood montane forest with tree fern forming a significant part of the sub-canopy Areas of open water and poor drainage	Warmer climate (a minimum 2.3 °C temperature depression) and precipitation >1000 mm/yr
LM1c-LM2b (559-530 cm: c. 18.4-c. 14.4 ka)	Sub-alpine shrub dominated vegetation	Climate amelioration following glacial maximum
LM1b (583-559 cm: c. 23.2 - c. 18.4 ka)	Alpine vegetation dominated by grassland and herb field species	Extreme cold (estimated up to 5.3 °C temperature depression) probably dry
LM1a (600-583 cm: c. 26.6 -c. 23.2 ka)	Alpine vegetation dominated by grassland with low abundance of shrubs	Cold and probably dry conditions

Table 4.4: A summary table of the vegetation and climate reconstruction for the Lake Mudgie pollen assemblage zones.

The vegetation and climatic changes identified in the Lake Mudgie record are summarised in Table 4.4. The base of this sequence represents cold and possibly drier conditions represented by alpine dominated vegetation, in particular grassland. The initial appearance of shrub species, a minor reduction in grain size and sand component and slight rise in LOI in the sediment suggest increased vegetation cover and milder conditions experienced at the site. The subsequent increase in herb taxa indicates an expansion of the high alpine zone and an associated decrease in MAT, relative to today, by a minimum of 5.3 °C producing extremely cold conditions. A subsequent transition into shrub dominated vegetation indicates a cool but warming climate. An increase in warming and wetter conditions is indicated by the appearance of broadleaved, hardwood low-

land forest. The vegetation composition is representative of the montane floristic zone and points towards MAT which were a minimum of c. 2.3 °C lower than at the site today. The warmer and wetter conditions continue with the expansion of podocarp forest toward the top of the sequence, albeit with some minor disturbances. The presence of *Libocedrus* may indicate presence of cool frost hollows in the valley which may have slowed the expansion of *D. cupressinum* into the region and caused the apparent disruption of its growth in the record. However, by the top of the sequence *D. cupressinum* became persistently the dominant species indicating a stable climate regime had been reached and suggesting that warm and wet conditions prevailed at the site by 11.9 ka.

Zone LM1a contains the KOT layer near the core base with an extrapolated age of c. 26.6 ka dating the base of the sequence. The presence of the KOT confirms the record extends beyond the LGM and covers a large proportion of the LGCP as defined by Alloway *et al* (2007). At 26.6 ka the region was already dominated by alpine vegetation, therefore the onset of the LGCP could not be dated. The Okarito record (Vandergoes *et al*, 2005; Newnham *et al*, 2007a) suggest that the LGCP could have started around 30.5-30 ka with cold conditions already prevalent by the beginning of the Lake Mudgie record. Likewise, the NZ-I CES estimates the onset of the LGCP to have commenced c. 29 ka. The Lake Mudgie record doesn't extend back to 29 ka but it does show that cold conditions were already prevalent in the Westland region by c. 26.6 ka and confirms an extended LGM occurred in southern New Zealand, beginning much earlier than the Northern Hemispheric equivalent (Mix *et al*, 2001).

The bottom 22 cm, subzone LM1a, contains a larger shrub influence than subzone LM1b, potentially indicating milder but still cold climate at the site. This corresponds to other records spanning this period and the NZ-I CES which recognises a milder more variable period during the LGCP prior to the ice advance during the LGM. At Lake Mudgie, this milder cold period spans an estimated c. 26.6 - c. 23.2 ka and corresponds to the NZ-I CES, which dates the milder and more variable climate phase between 27-21 ka (Alloway *et al*, 2007). The Okarito record shows a decline in grass abundance beginning just prior to the deposition of the KOT and continuing to decline afterwards coinciding with small revival of *D.*

cuppressinum, indicative of lowland forest (Newnham *et al* (2007a). Stratigraphic comparison to the Lake Mudgie reconstruction shows the reconstruction does not extend far enough back to record a decline in grass abundance, marking the onset of the LGCP milder interval, however, the vegetation assemblage also indicates the site was experiencing milder conditions just prior and after the KOT deposition. This is represented by a large proportion of shrub taxa in the record, but does not include lowland tree taxa, suggesting the signal is more subtle at Lake Mudgie.

Following this milder period is a period of extreme cold represented in the pollen record by the increased abundance of grassland including the tussock grasses which reach a peak abundance at the top of sub-zone LM1b. LOI is at its lowest abundance, whilst sand size particle abundance is at its highest during zone LM1b. The low organic carbon content indicates that either there was low vegetation productivity at the site due to the type of vegetation and sparse vegetation cover or large transportation of inorganic material to the site diluting the organic content. However, it is likely a combination of the two occurred as the decreases vegetation cover would not only decrease organic content but lead to increased erosion and therefore inorganic sediment input. Therefore, evidence for the start of LGM at Lake Mudgie began at c. 23.2 ka. The continued sustained increase in grassland species up the zone indicates minimum MAT was reached by c. 18.4 ka. The onset of this cold period appears to be earlier than at other sites, particularly compared to the speleothem record by Williams *et al* (2010) that dates the start of the LGM at 21.7 ka. The NZ-I CES shows an overlap between the timing of the LGCP and the onset of the LGM (Figure 1.1), with the latter spanning 25-18 ka, and represent a large uncertainty in the NZ-I CES. The Lake Mudgie record supports an earlier onset of the LGM cold period.

Termination I and the start of deglaciation, began c. 18.4 ka. This period is characterised by an abrupt decrease in Poaceae coinciding with an increase in the shrubs *Coprosma* and *Myrsine*, indicative of milder but still cool climates. By 17.6 ka, Poaceae has nearly been removed from the vegetation composition, emphasising the exceptionally rapid pace of vegetation response to climate amelioration. The timing of this event is similar to other pollen records from the Westland region e.g. Vandergoes and Fitzsimons, 2003 and Newnham *et al*,

2007a. The NZ-I CES estimates the deglaciation occurred between 19-17.9 ka supporting the timing suggesting in the Lake Mudgie record (Alloway *et al*, 2007). The Lake Mudgie record thus indicates that Termination I was well underway by 17.6 ka and the timing of deglaciation is broadly synchronous throughout New Zealand. The age of deglaciation in the South Island will be discussed further in Section 8.1.4

The pattern of vegetation change following Termination I is indicative of continuous climatic amelioration and possibly increased precipitation. This period is characterised by successive peaks in vegetation evolving from sub-alpine shrubs to montane broadleaved hardwood forest and culminating in podocarp temperate rainforest, marking the start of the Holocene. This vegetation successional pattern identified at this site is similar to other Westland sites such as Gillespies Beach Road and Okarito (Moar, 1973; Vandergoes and Fitzsimons, 2003; Newnham *et al*, 2007a). After 13.8 ka an abrupt increase in the lowland tree species, *D. cupressinum*, is apparent. However, this initial growth is not a smooth transition into dominance but contains minor reversals in *D. cupressinum* growth between 13.4-11.9 ka. At the same time a small abundance of species associated with the pollen assemblage zone LM1b return illustrating a potential disturbance in the vegetation growth at the site. This is further compounded by the sedimentology at the site which shows a small increase in sand size particles but more significantly two pulses of clay deposited as well. This indicates that there was possibly some disruption in the environment extending 13.4-11.9 ka although whether this was climate induced or a slow response to catastrophic event, such as an earthquake cannot be determined. The presence of *D. cupressinum* in ever increasing abundance suggests that any climatic deterioration could not have been severe.

The limited chronological control indicates this disturbance coincided with the LGR and corresponds closely with the timing of this event in the Williams *et al* (2005) speleothem record (13.5-11.6 ka). Yet, this disturbance is very minor when compared to evidence of the LGR in other records, such as the Okarito vegetation reconstruction (Newnham *et al*, 2007a) and other causes, such as cold air ponding (see Section 4.5.1.8) cannot be ruled out. The increase in *Cyathea smithii*

between 13.4-11.9 may indicate an increase in precipitation at the time which could also influence the vegetation composition. These disturbances aside, there is no unequivocal evidence in the Lake Mudgie pollen record for a LGR as has been reported in other regions of New Zealand (Alloway *et al*, 2007; Kaplan *et al*, 2010).

By 510cm, 11.9 ka, stable *D. cupressinum* abundance dominated temperate rainforest, indicative of the lowland vegetation communities in the contemporary central Westland region. Likewise at 515 cm total soil carbon abundance reached its maximum amount, indicating peak vegetation productivity had been reached by c. 12.8 ka. This indicates that Holocene conditions were reached at the site by 11.9 ka and represents early or pre-Holocene warmth that occurs during the YDC (12.9–11.5 ka; Alley, 2000), a period of cooling in the Northern Hemisphere. This early or pre-Holocene warming is also recorded at other sites in New Zealand including Okarito (Newnham *et al*, 2007a), Kaipo bog (Newnham and Lowe, 2000), Auckland Maars (Sandiford *et al*, 2003) and the speleothem record (Williams *et al*, 2005). Again Lake Mudgie supports the timing of main events set out in the preliminary NZ-I CES (Alloway *et al*, 2007).

4.5.3 Summary

The overall pattern of vegetation and climate change at Lake Mudgie is as follows. The LGCP began sometime before c. 26.6 ka. From c. 26.6 – c. 23.2 ka a milder period within the cold LGCP was recognised by the presence of shrub species. At c. 23.2 ka an extreme cold period was felt at the site with the area dominated by alpine vegetation which today is limited to altitude above 1200 m.a.s.l (Wardle, 1991). This extreme cold period climaxed towards the top of pollen assemblage zone LM1b. An estimated MAT depression of >6.1 occurred at 18.4 ka and deglaciation was evident by the transition of the alpine vegetation to shrubs. Climatic amelioration continued without significant interruption until a vegetation climax at c. 11.9 ka indicating the start of Holocene conditions. A possible minor climatic reversal occurred between 13.4-11.9 ka although this is tentative and the disruption in the vegetation and sediment record may be caused by some other environmental factor. The pattern and timing of change at Lake Mudgie are

broadly consistent with other records from New Zealand and will be discussed further in Chapter 8.

Chapter 5: Manks Tarn

Manks Tarn, the second site from the central Westland region, is located due south-east of Hokitika (see Figure 5.1) at 130 m.a.s.l. Manks Tarn is therefore altitudinally lower than Lake Mudgie and located further south. Geological mapping of the area suggest Manks Tarn is located within the LGM ice limit (Suggate, 1990). In planning the present study, it was therefore assumed that sediment assimilation commenced at the start of deglaciation and this was later confirmed by a radiocarbon age from the base of the sediment sequence. The site therefore provides a palaeoenvironmental record extending from Termination I to the Holocene. Whilst coring it became apparent that the sedimentation rate at the site was unusually high, allowing for a high resolution pollen record spanning the entire LGIT, producing a detailed palaeoenvironmental reconstruction from the site. As Manks Tarn is located in the same geographical region as Lake Mudgie, the reader is advised to refer to the regional setting description written in the previous chapter.

5.1 Manks Tarn site description

Manks Tarn is a small water body formed in a kettle hole nestled within a LGM lateral moraine (Figure 5.1 and 5.2). This moraine is one of many that step down and sweep along the western edge of Mt Upright, Mt Graham and Reef Knob. They were formed in the later stages of the LGM by a glacier extending down from the Southern Alps into the Kokatahi river valley and then northwards down the Hokitika River valley towards the coast. To the east of Manks Tarn lies the glacially formed Lake Kaniere which is dammed at the northern lake shore by a LGM terminal moraine. Lateral moraines also run along the eastern edge of Mt Graham and spill over the depression between Mt Graham and Mt Upright, eastwards, from the main Kokatahi River valley into Lake Kaniere. The Kokatahi River, today, flows north westwards around the base of Mt Graham and is a major tributary to the Hokitika River, joining it near the Kokatahi township. The Hokitika River continues to flow north westwards, entering the Tasman Sea at Hokitika Township.

Manks Tarn is surrounded by native forest that is dominated by podocarp temperate rainforest and particularly the lowland podocarp *Dacrydium cupressinum*. The edge of the tarn is skirted by swamp and peat land vegetation. Coring took place on the peaty margin by the shoreline on the south-eastern end of the tarn.

Figure 5.1: Location map of Manks Tarn in central Westland with a) with Manks Tarn site location in context of the LGM ice extent in the Westland region (redrawn from Suggate, 1990) and b) a zoom in site location with Manks Tarn circled in black. . Moraine deposits are marked on the map and are colour coordinated relating to the glacial advance which deposited the features. Moraine location is based upon Barrell *et al* (unpublished). Red- Moana moraines; Purple; Larrikins moraines; Green – Loopline moraines

Figure 5.2: Image of Manks Tarn; a) Looking eastwards across the tarn. The foreground is dominated by wetland. The background, on the slopes, is native lowland podocarp forest; b) a bird's eye view of Manks Tarn, the black dot showing the core location.

5.2 Sediment stratigraphy

The method used for coring at this site is explained in Section 3.1. The core was taken in January 2002. The top 5 m of the sequence consisted of liquid detrital mud presumed to be late Holocene and was therefore not retained. Coring began at 5 m below the ground surface and a total of eight 1 m sections were collected. Due to the time frame of interest in this project only the bottom six sections starting at 7 m were analysed. The age period of study is 18-8 ka (see Section 5.3) Figure 5.3 displays the core stratigraphy and description. A summary is described below.

The bottom 38 cm of the core was not analysed due to insufficient amounts of pollen for a statistically robust count. This section consists of blue-grey laminated sandy silts indicative of a lake environment. Pollen analysis began at 1170 cm and the stratigraphy between 1170-1160 cm consisted of grey-brown micaceous silts

changing to light brown organic silts between 1160-1118 cm. The remainder of the core analysed was massive brown detrital mud which had limited visual changes. This section is highly organic, containing high abundance of partly decomposed, unidentifiable plant fragments. The texture of this section becomes increasingly siltier down the core whilst becoming increasingly organic up core, indicative of lake shallowing and gradual terrestrialisation of the lake.

Figure 5.3: Core images from Manks Tarn in depth order and labelled with T representing each core section. Calibrated radiocarbon ages are added alongside the depth scale in the red block. The red dots represented sample location for radiocarbon dating.

5.3 Chronology

Prior to this study a bulk sediment sample provided an age for the base of the core and was used to confirm that sediment accumulation in the tarn began soon after Termination I. Subsequently, in 2010, a further five ages, derived from pollen concentrates (method described in Section 3.5.1) were obtained. Thus, a total of six AMS ages were used to establish the overall chrono-stratigraphic framework. An age-depth model is shown in Figure 5.4.

Core Depth	Fraction	¹⁴ C age BP	δ ¹³ C ‰	Reference code	% carbon yield	Calibrated date BP (2 sigma)
714-715	> 63µm	7908 ± 38	-29.2	SUERC-28875	17.9	8788 ± 190
883-884	> 63µm	10367 ± 44	-29.1	SUERC-28876	13.1	12233 ± 173
996-997	> 63µm	12068 ± 50	-29.1	SUERC-28877	14.3	13914 ± 145
1094-1095	>63	12910 ± 54	-29.2	SUERC-28878	10.8	15555 ± 516
1169-1170	>63	14746 ± 64	-27.5	SUERC-28879	2.4	18055 ± 416
1161-1169	Bulk sediment	14350 ± 270		WK10716		17646 ± 786
*Wood standard		11727 ± 50	± 25.0	SUERC-28880	51.9	
*Coal bit standard		54160 ± 350	± 23.2	SUERC-28881	88.6	

Table 5.1: Radiocarbon age details for Manks Tarn. Ages were calibrated using OxCal program. *The standards used to assess possible contamination during the preparation process.

All six ages lie in stratigraphic order and there is no evidence of any age reversals in the sequence. The two standards, which were prepared at the same time and using the same methodology as the other samples, produced reliable and expected ages. This indicates that little contamination from modern carbon occurred during the preparation process. The chronology indicates that 455 cm of sediment accumulation occurred over a period of less than 10,000 years and

confirms the high sedimentation rates at the site. This is important as it suggests the sequence will produce a high resolution and detailed palaeoenvironmental record of the LGIT and early Holocene period.

Figure 5.4: An age-depth model for Manks Tarn.

The age-depth model (Figure 5.4) shows that sediment accumulation remained high throughout the sequence, however, there are minor variations in the rate of accumulation through time. The slowest rates of deposition occur at the base of

the sequence (1170-1094 cm) accumulating 1 cm/ 33 years. The rate of deposition nearly doubles between 1094-883 cm and averages 1 cm/15-16 years. Finally the rate of accumulation reduces towards the top of the section analysed, accumulating 1 cm/20 years. The slowest period of sediment accumulation occurred during the inorganic siltier section of the core.

5.4 Results

5.4.1 Pollen stratigraphy

The pollen diagram is represented in Figure 5.4 along with the core stratigraphy and radiocarbon ages. This figure includes only the main pollen types, see appendix D for the whole pollen data set. The LOI and PSA results are also shown next to the pollen diagram and will be described later in Sections 5.4.2 and 5.4.3. The quantitative MAT reconstructions are also presented in Figure 5.4 and will be described in Section 5.4.3. Initially a skeleton sampling technique was used down the core to produce a preliminary results diagram. This was used to decide where the higher resolution pollen analysis would occur in the sequence. Subsequently a 2 cm resolution was applied to the section spanning 882-1052 cm. Based upon the age-depth model (Section 5.3) the section of detailed pollen analysis would have provided an average 30 year resolution.

CONISS cluster analysis, available in the Tilia program TGview, was used to group the total land pollen percentage data (see appendix B for the cluster analysis results). The pollen assemblage chosen for Manks Tarn tend to correspond to the biostratigraphic zones defined by the cluster analysis. The pollen assemblage zones used on the diagram were restricted to assemblage changes that represent palaeoecological and palaeoclimatic significance on a regional scale opposed to local scale variations. The main pollen zones are labelled MT1-4 (Figure 5.5).

5.4.1.1 Pollen assemblage zone

Zone MT1: 1170-1140 cm (18.1-17.1 ka):

The TDP is dominated by shrub taxa, in particular *Coprosma*, *Myrsine*, *Elaeocarpus* and *Asteraceae* and to a lesser extent *Pseudopanax*. This zone is the only section of the core where significant (10%) *Poaceae* has been counted. There is high fern spore abundance with monolete spores initially receiving 50%, reducing to 30% towards the top of this zone. *Hypolepis* also has a significant abundance, rising from 15% to a maximum of nearly 60% at the top of this zone. To a lesser extent *Histiopteris* and *Phymatosorus* are also present. The wetland taxon *Cyperaceae* also makes its only significant appearance in this zone reaching a peak abundance of 20% towards the top.

Zone M2: 1140-990 cm (17.1-13.8 ka):

This zone is still dominated by the shrub component. *Coprosma* and *Myrsine* share co-dominance at the base of the zone, however both taxa decrease in percentage up the zone. Conversely, *Griselinia* has a low abundance at the base but increases and becomes the dominant taxon at the expense of *Coprosma* and *Myrsine*. *Griselinia* reaches a consistent 30-40%. To a lesser extent *Pseudowintera* and *Pseudopanax* also increase in abundance. The tree taxa, *Libocedrus* and *Weinmannia*, are also present with consistent but low abundance of c. 5%.

This zone contains a large transition in wetland, tree ferns and fern spore component. *Cyperaceae* decreases to near 0% coincidently with the rise in *Leptospermum* type, which steadily increases from 0% to 20%. A large increase in tree fern abundance occurs with *Cyathea smithii* abundance increasing from 0% at 1140 cm to 50% at 1080 cm with its abundance remaining above 45% throughout the rest of the zone. A significant change in fern spore composition also occurs. The monolete spores consistently reduce in abundance concurrently with *Histiopteris* and *Hypolepis*. Conversely the abundance of *Phymatosorus* and *Hymenophyllum* increase from near 0% to 5-8%. More significantly, the aquatic taxon *Isoetes* increases from 0% to near 30% at 1180 cm and remains at a high percentage throughout.

Zone MT3 990-885 cm (13.8-12.3 ka):

Zone MT3 is characterised by an increase of tall tree taxa at the expense of the shrubs. *Coprosma*, *Griselinia* and *Myrsine*, all decline in abundance and the overall shrub composition reduces from 80-90% down to only 20-30%. Conversely, the abundance of *Weinmannia racemosa* increases abruptly from 5% at 990 cm to 70% by 965 cm and continues to increase to 80% remaining, consistently high for the rest of this zone. *Cyathea smithii* and *Leptospermum type* remain at a similar abundance as in the previous zone. The fern spore composition decreases in abundance with all species present in the previous zone decreasing. The aquatic *Isoetes* reduces to near 0% at the MT2-MT3 zone boundary.

Zone M4: 885-714 cm (12.3-8.8 ka):

Like the previous zone, the TDP is dominated by tall tree taxa, with the overall composition increasing to 80-85 % at further expense to shrub species. *W. racemosa* is dominant at the zone base but decreases to only 30% at 818 cm. Conversely the podocarps *Prumnopitys ferruginea* and *P. taxifolia* and more significantly *Dacrydium cupressinum* increase in abundance. *D. cupressinum* increases from 0% at 885 cm to 45% by 810 cm becoming the dominant taxon. It continues to increase reaching a maximum abundance of 70% at 760 cm. The tree fern, *Cyathea smithii* abundance remains nearly consistently above 35% throughout, except for one sample at 828 cm. *Dicksonia squarrosa* is present at low abundance (2-3%). *Leptospermum type* decreases to 10% and then 5%.

Figure 5.5: Manks Tarn pollen percentage diagram, including core stratigraphy, summary diagram, total carbon content, particle size percentage and pollen-climate reconstructions. Radiocarbon dates are summarised in Table 4.1.

5.4.2 Particle Size Analysis

The core was sub-sampled at a 2 cm resolution spanning 790-1170 cm. The results are presented in Figure 5.5. Throughout the section analysed the silt component is consistently dominant in the sequence, mostly between 70-80%, and clay size particle abundance remains between 2-3%. The pattern of the particle size composition can be split into three zones and is primarily based upon the changes in the sand abundance and associated silt variations. PSA results will be put into context with the pollen data. The section from 1170-1090 cm contains the highest abundance of sand in the sequence, decreasing from 25-30% at the base to around 17-20% for the majority of this zone. The silt component ranges from 70-80%. A reduction in sand abundance begins at 1090 cm and remains consistently at c. 7% until 810 cm whilst the silt content increases to 90%. At 810 cm the sand content increases to 20% and remains at that abundance for the remaining section.

5.4.3 LOI

As with the PSA, sampling for LOI analysis occurred at a 2 cm resolution spanning 791-1169 cm. Four main patterns are evident in the core, with minor variations superimposed on the main patterns (see Figure 5.5). The base of the sequence contains 12-10% carbon between 1170-1145 cm. This low carbon content corresponds with pollen assemblage zone MT1 and the section containing the highest amount of sand. After 1145 cm the LOI begins to consistently increase reaching 30% by 1082 cm and remains at this abundance until 1020 cm. This is followed by a second, albeit more subtle increase in carbon content to 40%. The carbon content remains stable at 40% until 900 cm. The pattern of increased carbon content between 1145-900 cm corresponds closely to pollen assemblage zones MT2 and MT3 and the section contains relatively low amounts of sand. The LOI remains high between 900-791 cm, however, the abundance becomes more variable and oscillates between 35-42%. This relatively variable section coincides with the rise in *D. cupressinum* and pollen assemblage zone MT4. This section also includes the rise in sand abundance towards the top of the sequence analysed.

5.4.4 Temperature reconstructions

Figure 5.5 presents the pollen-temperature reconstructions for PLS-C3 and W-MAT. Overall the W-MAT consistently gives higher temperature than the PLS-C3 model. Initially the PLS-C3 model predicts increasing temperature from 9.7 °C to 10.4 °C followed by gradual and continuous cooling reaching a minimum 7.2 °C at 10.8 ka. Temperature started to increase at 10.8 ka reaching 10 °C by 9.1 and decreasing slightly at 8.7 ka. The W-MAT model predicts a less varied pattern in temperature which initially increases from 9.1 °C to 10.1 °C and remains between 9.3-10.1 °C until 9.8 ka. Thereafter temperature increases significantly, reaching a maximum of 11.4 °C at 9.1 ka then reducing to 10.9 °C at the top of the core.

5.4.5 Principle Component Analysis (PCA)

PCA, using CANOCO version 2, is first applied on the pollen data. Subsequent runs with the addition of LOI and the % of sand, silt and clay particle sizes followed to explore the relationships between the pollen assemblage and changes in core sedimentology. Pollen and PSA samples were taken at the same depth horizon, but LOI was consistently 1 cm offset. Therefore an average was taken of the two % carbon samples which bordered the sample depth of the PSA and pollen analysis to enable more meaningful comparisons. At the top and the base of the core, pollen analysis occurred at a lower resolution and therefore the PSA and % carbon sample lying between the pollen samples were excluded from the PCA dataset. The taxa included in the dataset are those with percentages over 1%.

The results are presented in two sections; with and without the sedimentological data. The taxa eigenvalues are summarised in Figure 5.6 and Table 5.2, followed by a description of the sample eigenvalues presented in Figure 5.7 and Table 5.3. The results are considered in context with the pollen diagram and pollen assemblage zones (Figure 5.5).

5.4.5.1 Palynological PCA results

Figure 5.6 is a biplot of taxa eigenvalues for axis 1 and axis 2, illustrating the stratigraphic affinities among the pollen taxa. Three groups are apparent with one group able to be divided into two sub-groups (Figure 5.6 and Table 5.2).

Group 1 contains species with high positive eigenvalues on axis 1 and negative values on axis 2, except for *Coprosma*, which has a positive eigenvalues on axis 2. This group represent taxa that are dominant in zone MT2 and include the shrubs *Coprosma*, *Griselinia*, *Myrsine*, *Psuedopanax* and *Pseudowintera*. This group also includes *Isoetes*, *Hymenophyllum*, and *Phymatosorus*.

Group 2 contains a large number of taxa and has subsequently been split into two sub-groups: 2a and 2b. Overall the group has positive eigenvalues on axis 1 and both positive and negative values on axis 2. Sub-group 2a contains tall tree taxa *Libocedrus*, *Meterosideros* and *Fuscospora* and the shrub/short trees *Elaeocarpus*, *Rubus* and *Schefflera*. This group also contains the tree fern species *Cyathea dealbata* and wetland taxa *Myriophyllum* and *Gunnera*. Sub-group 2b contains taxa with higher eigenvalues on axis 2 and comprises taxa which are most prominent in MT1. Group 2b includes Poaceae and *Kelleria*, and the shrub taxa *Hebe* and Asteraceae. The wetland taxa Cyperaceae, *Gonocaropus* and *Empodisma* and fern spores from *Histiopteris*, *Hypolepsis*, and monolete spores are also present in this group.

Group 3 contains taxa with positive scores on axis 2 and negative scores on axis 1. This group contains a large proportion of tall tree taxa when compared to the other group and includes *Dacrycarpus dacrydioides*, *Dacrydium cupressium*, *Prumnopitys ferruginea* and *P. taxifolia*. This group also includes the shrub *Muehlenbeckia* and the tree fern *Dicksonia squarrosa*.

There are three outliers in this figure which do not appear to relate to any of these groups: *Cyathea smithii*, *Leptospermum type* and *W. racemosa*.

Figure 5.6: Manks Tarn PCA results for selected taxa plotted for axis 1 and axis 2. Circles indicate arbitrary chosen group clusters. **Group 1;** Co – *Coprosma*, Gr – *Griselinia*, My – *Myrsine*, Ps – *Pseudopanax*, Pw – *Pseudowintera*, Py – *Phymatosorus*, Is – *Isoetes*, Hm – *Hymenophyllum*. **Group 2a;** Ap – *Apiaceae*, C.d – *C. dealbata*, Ch – *Chenopodiaceae*, Em – *Empodisma*, El – *Elaeocarpus*, Gu – *Gunnera*, Li – *Libocedrus*, Me – *Meterosideros*, Mp – *Myriophyllum*, N.f – *Fuscospora*, Ru – *Rubus*, Sc – *Schefflera*. **Group 2b:** As – *Asteraceae*, Cy – *Cyperaceae*, Go – *Gonocarpus*, He – *Hebe*, Hi – *Histiopteris*, Hy – *Hypolepsis*, Ke – *Kelleria*, Mo – *monolete spore*, Po – *Poaceae*. **Group 3:** Da – *Dacrycarpus*, D.c – *D. cupressinum*, D.s – *D. squarrosa*, Mu – *Muehlenbeckia*, Ph – *Phyllocladus*, P.f – *P. ferruginea*, Pt – *P. taxifolia*. **Outliers:** Le – *Leptospermum type*, We – *W. racemosa*, C.s – *C. smithii*.

Group	Tall tree	Short tree/ shrub	Herb	Wetland	Tree fern	Fern
1		<i>Coprosma</i>				<i>Isoetes</i>
		<i>Myrsine</i>				<i>Hymenophyllum</i>
		<i>Griselinia</i>				<i>Phymatosorus</i>
		<i>Pseudopanax</i>				
		<i>Pseudowintera</i>				
2a	<i>Libocedrus</i>	<i>Elaeocarpus</i>	Chenopodiaceae	<i>Gunnera</i>	<i>C.dealbata</i>	
	<i>Meterosideros</i>	<i>Rubus</i>				
	<i>Fuscospora</i>	<i>Schefflera</i>	Apiaceae	<i>Empodisma</i> <i>Myriophyllum</i>		
2b		Asteraceae	Poaceae	Cyperaceae		<i>Hypolepsis</i>
		<i>Hebe</i>	<i>Kelleria</i>	<i>Gonocarpus</i>		Monolete spore <i>Histiopteris</i>
3	<i>D.dacrydiodes</i>	<i>Phyllocladus</i>			<i>D.squarrosa</i>	
	<i>D.cupressinum</i>					
	<i>P.ferruginea</i>	<i>Muehlenbeckia</i>				
	<i>P.taxifolia</i>					
Outliers	<i>W. racemosa</i>			<i>Leptospermum</i> type	<i>C.smithii</i>	

Table 5.2: Species list for each group recognised by the PCA results and split into different vegetative components.

Figure 5.7 is a biplot of sample eigenscores for axis 1 and axis 2. There are four sample groups identified with a few samples appearing to be transitional sample points between groups. Arrows on the figure show the stratigraphic order of groups beginning with the deeper samples and decreasing in depth. This is also the order the results are written in.

Group 1 contains the bottom 8 samples which lie in a horizontal linear pattern spanning eigenscores of 3.8-1.2 on axis 2, but only ranging 1.0-1.5 on axis 1. Scores on both axes reduce with a decrease in sample depth with the samples lying in stratigraphic order and ranging 1170-1084 cm. However, sample 1004 cm, lying at the end of the group, appears to be an outlier and is out of stratigraphic order. Group 2 comprises 32 samples clustered together again in a horizontal linear pattern, but not in stratigraphic order. These samples have negative scores on axis 2. The sample cluster spans 1072-992 cm. Four transitional samples join

group 2 to group 3. A general stratigraphic pattern is followed with a decrease in eigenvalues on axis 1 corresponding with a decrease in core depth. Group 3 is the largest cluster containing 52 samples spanning 988-874 cm. This group corresponds to pollen zone MT3. Another two transitional samples link group 3 and group 4. Group 4 contains the top 8 samples, excluding 78 cm which is a transitional sample. This group covers the top 828-714 cm of the core and corresponds to pollen zone M4.

Figure 5.7: The PCA results for each sample plotted for axis 1 and axis 2. The circles around the sample clusters indicate which samples belong to a group and are chosen arbitrarily. Each of the 6 groups has been labelled and the depths displayed indicate the range of core depth that each group represents. The arrows indicate the stratigraphic order of the groups with the arrows pointing in the direction of decreased core depth.

CONISS		PCA	
Pollen zone	Depth	Group	Depth
M1	1170-1140	1	1170-1086
M2	1140-990	2	1072-992
		Transition	990-982 (excl 988)
M3	990-874	3	988-874
		Transition	844,778
M4	874-714	4	828-714

Table 5.3: A table comparing the depths of the pollen assemblage zones, carried out by the CONISS stratigraphic cluster analysis, to the depths of the sample groups, identified using PCA.

Comparison between the CONISS and PCA (Table 5.3) indicates a close similarity between the depth of the groups and pollen zones. Minor differences are mostly attributed to the different means of delineating boundaries: PCA recognises transitional values whereas cluster analysis does not. These differences are accentuated where a comparatively low sampling resolution coincides with a zone boundary, e.g MT4/3. A notable exception is pollen zone MT1, which only includes the bottom 30 cm of the section analysed, whereas PCA group 1 spans a large range on axis 2, more similar to the transitional samples, and could be representative of a continually changing pollen assemblage up core until 1086 cm. However, the PCA groups were arbitrarily chosen which would also have an influence on the samples included in each group. The group classification method can also be used to explain the differences between the CONISS zones and PCA groups. Despite the minor discrepancies, overall the two methods produce similar results.

Figure 5.8 plots the sample PCA scores for axis 1 and axis 2 along a depth gradient. Unlike the biplot (Figure 5.7), which plots the position of each sample in a two dimensional space defined by the first two principles, plotting an axis individually shows the variation of PCA scores in a one dimensional, stratigraphic, space. This allows for stratigraphic comparisons between the PCA scores and

pollen record and makes it easier to assess which species are influencing the PCA score at different depths. The results are described below.

Axis 1 PCA scores have a stable and high value ranging 1.4-0.9 from the core base until 1000 cm. This is followed by a steep decline in values to -1.3 at 920 cm. The values remain low but steadily increase up the core reaching a maximum 0.14 at 810 cm. This is followed by a minor oscillation until 762 cm where the values become stable at 0.03. The stratigraphic pattern of Axis 1 shows that samples with high scores contain high abundance of shrub taxa (pollen zones MT1 and MT2), which have high PCA scores (see fig 5.6.). Samples with low score on Axis 1, contain a high abundance of tree taxa, which have low PCA scores (MT4), and samples containing high abundance of *Weinmannia racemosa*, which has a very low PCA scores, also have a low axis score (MT3).

PCA scores PCA scores contain a much wider range in values and vary comparatively more up the core. Initially, axis 2 has a high value at the core base of 3.7, followed by a large, steep decline to -0.8 at 1072 cm. From 1072-892 cm is a period of low but oscillating values. This group can be split into two with the bottom section, 1072-978 cm, having an overall lower PCA scores compared to the upper section. The PCA scores steadily increase to 1.78 by 828 cm and oscillates with higher PCA scores between 1.78-1 up the remaining core section. Axis 2 scores appear to be sensitive to fern spores and tree fern loading scores. High axis scores, at the core base, coincide with high abundance of monolete spores and ferns which have high axis 2 loading scores. The decline in axis 2 scores is attributed to a decline in fern abundance, and an increase in the *Cyathea smithii*, which has a negative eigenvalue. Axis 2 values increase in MT3 and MT4 due to an increase in tree abundance, which contains positive eigenvalues on axis 2. However, the high abundance of *C. smithii* keeps the axis scores lower than at the base of the core.

Figure 5.8: Manks Tarn pollen and sediment results with the addition of the sample eigenvalues for axis 1 and axis 2.

5.4.5.2 Pollen and Sediment PCA results

Figure 5.9 is a biplot of taxa and sediment eigenvalues of axis 1 and axis 2, illustrating the stratigraphic relationship of the pollen taxa and sediment characteristics. Sand size particles appear to relate to sub-group 2b, which are dominated by herb species, alpine flora and taxa represented in MT1. Silt size particles lie in subgroup 2a and are associated with tall tree species and tree ferns. Clay size particles lie in group 3 and are associated with podocarp forest and tree ferns. LOI is an outlier and is not associated with any taxa. *W. racemosa* remains an outlier but *Cyathea smithii* moves into group 2a and *Leptospermum* type into group 3 (refer back to Section 5.4.1.1 and Figure 5.6).

Figure 5.9: Manks Tarn PCA results for selected taxa and sediment characteristics plotted for axis 1 and axis 2. Circles indicate arbitrary chosen group clusters. (Refer to Figure 5.6 caption for sample codes).

5.5 Discussion

This section interprets the pollen and sedimentological results in terms of inferred vegetation communities and then climatic reconstruction. The same principles and reference material used to interpret the Lake Mudgie results are used for Manks Tarn (refer to Section 4.5). The timing and pattern of vegetation and inferred climatic change will then be put in context with palaeoenvironmental records from other sites in New Zealand and from the Southern Hemisphere mid latitudes.

5.5.1 Vegetation and climatic interpretation

5.5.1.1 Zone MT1 vegetation reconstruction (18.1-c. 17.1 ka)

Zone M1 is dominated by shrubland taxa with grassland and alpine associated flora including *Hebe* and Asteraceae present. The vegetation composition is very similar to that recorded in Lake Mudgie zone LM1c/LM2a. The presence of grassland taxa indicates a restricted alpine vegetation community that is possibly limited to the high altitude locations. *Coprosma* and *Myrsine* are the dominant shrubs in this zone, and are indicative of a sub-alpine/montane environment in the contemporary Westland environment (Wardle, 1977). Asteraceae, which contains both shrub and herb varieties, is also present, reflecting either shrubland or alpine herb environments, or both. The abundance of monolete spores and *Hypolepis* implies a prevalence of ground ferns and may be indicative of a disturbance in the shrubland or instability of the substrate limiting plant growth. The high sand component may also be indicative of instability and of an actively eroding landscape. The presence of Cyperaceae indicates a local environment that is either waterlogged or open water bodies skirted by sedgeland.

The LOI in this zone is at its lowest indicating that vegetation productivity and cover were limited. This is supported by the presence of high sand abundance suggestive of areas with limited vegetative cover and therefore increased erosion bringing larger particles into the tarn. Transport of sand size particles to the tarn basin would be limited due to the small catchment Manks Tarn and limited inflow to the site.

5.5.1.3 MT1 climate reconstruction

The abundance of *Coprosma* and *Myrsine* contained in zone MT1 combined with a significant although low percentage of Poaceae, point towards a cool climate. The timing of deposition coincides with the start of deglaciation and a period of climatic amelioration as represented in other records (see Section 5.5.2). In the contemporary Westland environment, the plant assemblage represented in MT2 is indicative of the sub-alpine/alpine vegetation. This vegetation community is restricted to a minimum 800 m.a.s.l (Wardle 1991). This is 670 m above the Manks Tarn core site at 130 m.a.s.l. The pollen assemblage at 18.1 ka implies the sub-alpine zone descended to at least this altitude. Using an environmental lapse rate of 0.5 °C/100 m, an estimated minimum temperature depression during MT1 is c. 4.2 °C.

Palaeothermometer records from a site offshore of the Westland coastline and estimations from other records suggest a maximum sea surface temperature depression of c. 5 °C was experienced in New Zealand during the LGM (Barrows *et al*, 2007). The estimated temperature depression for MT1 is less than that during the LGM and implies temperature had already started to increase, although moderately, before the start of the record. This adds validity to the assumption that Manks Tarn was ice free soon after the LGM and the start of Termination I.

5.5.1.3 MT2 vegetation reconstruction (17.1-13.8 ka)

Zone MT2 is characterised by the disappearance of grassland and an expansion of shrubland community in the region. The disappearance of grassland and alpine taxa suggest climatic conditions became warmer and more suitable for shrub species, such as *Coprosma* and *Myrsine*. It is likely the alpine flora migrated and became restricted to the upland areas. Poaceae does not reappear with significant amounts after its disappearance in the record, indicating climate did not deteriorate, or if it did, was not great enough for regrowth of grassland in the region.

Coprosma and *Myrsine* remain the dominant taxa at the base of MT2 but begin to decline up the zone. A vegetational successional pattern occurs, similar to that recorded in Lake Mudgie, with a transition from cold sub-alpine dominated

shrubland to montane *Griselinia* dominated forest. In the contemporary Westland environment *Griselinia* forest is restricted to <800 m.a.s.l and is characterised as montane. *Myrsine divaricata*, *Pseudopanax simplex*, *P. colensio*, and *Coprosma wallii* tend to form part of the sub canopy with *C. astonii* being part of the shrub tier. It is likely that a similar species composition was present in this pollen zone. At the site *Griselinia* probably dominated the lowland alluvial terraces and gullies, overtopping the *Myrsine* and *Coprosma* (Wardle, 1991). Coinciding with the *Griselinia* rise is an expansion of the tree fern *Cyathea smithii* which is likely to have been significant component in the sub-canopy. *Pseudopanax* and *Pseudowintera*, also a small tree/shrub taxon, is present in the regional vegetation suggesting an increase in plant diversity and a more complex vegetation structure. There is low abundance of *Libocedrus*, a shade intolerant species. Its presence may indicate there was not a dense forest canopy in the area enabling *Libocedrus* to grow. The LOI increases significantly from the previous zone, however, it is still limited to 30%, suggesting that plant productivity and coverage had not reached its maximum potential at the site.

The abundance of Asteraceae remains at a similar level as in the previous zone suggesting it is still a significant component in the vegetation. Like *Coprosma* and *Hebe*, and as for MT1, Asteraceae includes both herb and shrub forms and it is difficult to differentiate these in the pollen diagram. However, its continued appearance during the shrub zone, when no other herb species are present, including grass, indicate that Asteraceae represents a shrub component in this zone.

Concurrent with the rise in *Griselinia* is an increase in the tree fern *Cyathea smithii* and *Isoetes*, a macrophyte. *C. smithii* abundance increases significantly and remain persists throughout the rest of this zone. The presence of *C. smithii* in a shrub community rather than a forested landscape points towards its persistence as a pre-forest successional element, although it also occurs in gullies and in the bushland perimeter in the contemporary landscape (McGlone and Bathgate, 1983) and these are likely to be the areas it inhabited during MT2. However, it could also be an important component in the small tree tier, dominated by *Griselinia*. *Isoetes*, representing open water in the local environment, was not present in the previous

zone, which can either indicate the climate was unsuitable for its growth or the water level was not shallow enough to support *Isoetes*. *Isoetes alpinus* today grows in the alpine tarns and its presence in the Manks Tarn record suggests that temperature was not the limiting factor, but water level. This species has a limited water depth niche of 0.6-4.5 m (Brown, 1975) and would not be present in deeper water. It is likely Manks Tarn was becoming increasingly shallower due to continued sedimentation of the lake and leading to eventual terrestrialisation.

The base of MT2 continues to have a high abundance of sand indicating the *Coprosma/Myrsine* dominant shrubland had gaps in the canopy enabling erosion to persist at a similar rate as in the previous zone. The small tarn catchment and limited inflow into the tarn suggests the sand size material is most likely sourced from local hillslope erosion. However, coinciding with the rise in the broadleaved montane forest and increase diversity is a decrease in sand abundance and overall average particle size up core. This indicates an increase in plant density in the region protecting the soil surface from erosion. The % carbon increases dramatically in this zone and corresponds with the disappearance of Poaceae and the increase in shrub dominance indicates an increase productivity and biomass input to the sediment. The carbon content maintains a plateau at 30% throughout the section dominated by *Griselinia* and only begins to increase towards the top of MT2, indicating plant productivity was stable throughout this section but had not reached its full potential. A climate interpretation for MT2 is presented along with that for MT3 in Section 5.5.1.5.

5.5.1.4 MT3 vegetation reconstruction (13.8-12.3 ka)

Pollen assemblages in zone MT3 shift from a low lying *Griselinia*-dominated broadleaved forest to a hardwood/podocarp forest dominated by tall tree species, in particular *W. racemosa*. The increase in *W. racemosa*, a drought and frost sensitive taxa, indicates that temperature and precipitation levels were high enough (>1000 mm/yr) to allow this taxon to become established. The presence of *Griselinia*, *Myrsine* and *Coprosma*, although in reduced abundance, indicates broadleaved forest was still widespread in the *W. racemosa*-dominated forest and was probably part of the shrub understory (Wardle, 1977). The consistent presence of *Cyathea smithii*, which is probably part of the small tree tier, is also

indicative of moist environments. *Griselinia* may have also provided a nursery and cover for the podocarp taxa as it does in the contemporary landscape (Wardle, 1977). The increase in *Prumnopitys ferruginea*, *P. taxifolia* and *Dacrydium cupressinum* is likely to represent the start of the development of tall podocarp forest in the area. However, *D. cupressinum* abundance is low and was probably restricted to lowland terrace or hill slopes where it is commonly found today (Rief and Allen, 1988; Wardle, 1991). The disappearance of *Isoetes* is likely to indicate there is no longer open water at the site. However, unlike Lake Mudgie, the reduction in *Isoetes* is not combined with a visual stratigraphic change in the core. *Libocedrus* also reduces in abundance suggesting a denser canopy limiting its growth.

The LOI abundance increases to a stable 40% consistent with an increased biomass in the environment caused by the introduction of tall tree species. Grain size distribution remains similar to that recorded in the previous zone indicating the substrate was at a stable state as in pollen assemblage zone MT2.

5.5.1.5 MT2 and MT3 climate interpretation (17.1-12.3 ka)

Zone MT2 shows a shift from sub-alpine shrubland communities which today lie at 800-1200 m.a.s.l (Moar, 1970; Wardle, 1991) to montane broadleaved forest with a restricted upper altitudinal limit of 800 m.a.s.l. This indicates an increase in temperature that is characteristic of the LGIT.

The increase in *W. racemosa* and the pace of its rise to dominance suggest that temperature and increased precipitation was high enough to allow broadleaved/podocarp forest to grow at the site. *W. racemosa* in particular demands high precipitation levels and is sensitive to water stress and severe frost (McGlone *et al*, 1994). Furthermore, the expansion of *Prumnopitys ferruginea* and *P. taxifolia* provide greater evidence to support an increase in precipitation and temperature during this zone. These species form an important component of the contemporary lowland forest. However, the lack or small abundance of *D. cupressinum*, which is dominant in the contemporary lowland forest suggest the temperature and precipitation may not have been as high as it is at present. Therefore it is likely that this low hardwood/podocarp forest grew in wetter and

warmer conditions than in the previous zones, but the climate was cooler than present condition in the region. The species composition of zones MT2 and MT3 is representative of montane vegetative zone, which in the contemporary landscape is restricted to 600-800 m.a.s.l (Moar 1970; Wardle, 1977). Using the ELR of 0.5 °C/100 m, a minimum temperature depression of 2.4 °C is inferred.

5.5.1.6 MT4 vegetation reconstruction (12.3-8.8 ka)

This zone is characterised by the transition from *W. racemosa* dominated broadleaved/podocarp forest to a *D. cupressinum* dominated podocarp/hardwood lowland forest with *D. Cupressinum* being the dominant species throughout the zone and *W. racemosa* co-dominant. *Prumnopitys taxifolia* and to a lesser extent *P. ferruginea* are still persistent in this zone, albeit both in small abundance, suggesting they contribute significantly to the canopy vegetation. Small amounts of *Coprosma* and *Myrsine* occur towards the top of this zone and would be present in the shrub tier. Contemporary Westland forest has *Myrsine australis* and *Coprosma lucida* which forms the main shrub tier under the forest canopy. This vegetation assemblage is very similar to contemporary forest communities in the Westland region. Together these assemblages represent a full development of interglacial temperate rainforest (Wardle 1977). This idea is supported by the high organic carbon content of the sediment and further increase in visual organic material in the core representing an increase in biomass and productivity of the vegetation.

The high levels of tree ferns, in particular *Cyathea smithii*, but also *Discksonia squarrosa*, suggest that these species formed an important component of the small tree tier and persisted amongst the tall podocarp forest. The relative abundance of tall forest and fern species represent vegetation and pollen assemblage characteristic of the contemporary lowland forest zone that reaches a maximum altitude of c. 400-600 m.a.s.l and extending down to sea level (Wardle, 1977).

5.5.1.7 MT4 climate reconstruction

The dominance of *Dacrydium cupressinum* and the presence of *Prumnopitys ferruginea*, *P. taxifolia* and *W. racemosa* in the regional vegetation indicate the establishment of stable climate conditions during zone MT4. These species

occupy the contemporary warm and wet montane and lowland forest in the area (Franklin, 1968; Wardle, 1977) which strongly suggests the climate at this time remained warm and wet. The contemporary *D. cupressinum* dominated temperate rainforests of the Westland region is prevalent up to 400-600 m.a.s.l and experiences MAT of c. 11 °C and precipitation c. 5000 mm/yr (Wardle, 1977; 1991). It is likely that the climate experienced at the site during zone MT4 was similar to the contemporary lowland Westland climate and indicates that full climatic amelioration was reached. However, when compared to Lake Mudgie, the abundance of *W. racemosa* appears to be much higher during the early Holocene period, co-dominant with *D. cupressinum*. Other tall tree species are also present in high abundance suggesting greater forest diversity at this site than at Lake Mudgie. The establishment of full interglacial vegetation occurs between c. 11.2-10.8 ka.

5.5.1.8 Quantitative pollen/climate reconstructions

The climate changes described above are characterised by a consistent pattern of climate amelioration whereby MAT increases progressively. This pattern of amelioration is consistent with other late glacial records in New Zealand (e.g. Vandergoes et al, 2003; Newnham et al, 2007). However, the estimated temperature changes predicted by the two models do not follow with the above pattern. The PLS-C3 model suggests a continuous and gradual decline in MAT occurred between 17.2-10.8 ka followed by an increase in temperature to near contemporary conditions during the Holocene. The minimum MAT reached was 7.2 °C, this is 4.3 °C colder than current conditions, and a similar temperature depression as expected to have occurred during the LGM (e.g. Barrows et al, 2007).

In contrast the W-MAT model does not show a cooling pattern during the LGIT. Instead this model suggests MAT remained stable between 18.1-10.8 ka with temperatures near to contemporary conditions. MAT increases after 10.8 ka to near contemporary levels during the Holocene. The patterns of predicted MAT change in both models do not correspond with the inferred pattern of climate change from the vegetation record and using ELR. The discrepancies between the patterns of MAT change inferred by the different methods may be attributed to the

limitations with the quantitative methodology. This is discussed in more detail in Section 8.2.4.

5.5.2 Summary discussion

Pollen assemblage zone	Vegetation reconstruction	Climatic reconstruction
MT4; 874-714 cm (12.3-8.8 ka)	<i>D. cupressinum</i> dominated temperate rainforest with <i>W. racemosa</i> being the co-dominant species.	Stable warm wet climates with mean annual temperatures 11 °C and precipitation 5000 mm/yr. Climate amelioration culminated at 10.8 ka representing the start of the Holocene.
MT3; 990-874 cm (13.8-12.3 ka)	<i>W. racemosa</i> dominated broadleaved/podocarp forest.	Warm, wet climate with precipitation at least 1000 mm/yr, probably much higher.
MT2; 1140-990 cm (17.1-13.8 ka)	Broadleaved hardwood montane forest with tree ferns forming a significant part of the sub-canopy. Areas of open water and poor drainage.	Warmer climates (a minimum 2.4 °C temperature depression).
MT1; 1170-1140 (18.1-17.1 ka)	Sub-alpine shrub dominated vegetation. Alpine grassland present in small abundance. Vegetation cover was sparse.	Climate is cooler than at present.

Table 5.4: A summary table of the vegetation and climate reconstruction for the different pollen assemblage zones.

Table 5.4 is a summary table of vegetation and inferred climate reconstructions. The climate reconstruction is based upon the assumption that vegetation change is climatically controlled. However, the pattern of vegetation transition in the core may not solely represent climatic change but could be factor of time lags associated with primary vegetation succession and will be discussed below. Wardle (1980; 1991) documents the successional pattern of primary forest development in Westland forests and concluded that full succession takes a period ranging 500-1000 years. The initial seral stages are characterised by the occupation of *Coprosma* and *Mrysiine* for the first 130 years. Tree ferns, *Griselinia*,

Metrosideros umbellata and *Weinmannia racemosa* then overtop the shrubland and produce a tree canopy. The podocarp *Prumnopitys ferruginea* and *Dacrydium cupressinum* may also be present in the environment but only as small seedlings and do not become canopy trees for another 400-450 years during which time the *Metrosideros/W. racemosa* tree communities are collapsing. By 1000 years *D. cupressinum* becomes dominant.

The successional vegetation pattern described is very similar to the temporal changes in the pollen record which suggest that *Coprosma* and *Myrsine* shrubland was overtopped by *Griselinia* followed by tall tree species *W. racemosa*, *P. ferruginea* and *P. taxifolia* culminating in a *D. cupressinum* dominated forest. However, this process takes a period of 6000 years with a prolonged abundance of the shrubs and *Griselinia* combined with an abrupt increase in *W. racemosa* and *D. cupressinum*, which cannot be explained by successional pattern of vegetation change solely. The PCA sample eigenvalues confirm a rapid vegetation change as there are few transitional samples lining one group to another. This suggests the pollen assemblage changed significantly and abruptly from each zone/PCA group. The abrupt increase in these taxa from previously low representation in the previous pollen assemblage zone suggest increases in precipitation and temperature are important forcing factors for the pattern of vegetation recorded at this site.

The age of sediment accumulation and the first pollen assemblage analysed in the core indicates the minimum age of deglaciation at c. 18.1 ka. This indicates that deglaciation must have begun prior to 18.1 ka. The pollen assemblage in MT1 is dominated by *Coprosma/Myrsine* shrub and a small abundance of Poaceae, providing a very similar assemblage as found in the top of section LM1c in Lake Mudgie (see Section 4.4.1). LM1c clearly marks the transition from alpine dominated grassland to sub-alpine shrubland and the end of the LGM. Before shrubland could grow at Manks Tarn the site needed to become ice free, lithologically stable and have soil development. This suggests deglaciation had begun sometime prior to 18.1 ka. Evidence from Lake Mudgie indicates that vegetation response to climate amelioration was rapid and the same process is

likely at Manks Tarn too and suggests that deglaciation could have started not that long before 18.1 ka. This will be discussed further in Section 8.1.4.

Okarito pollen site, located in south Westland, has a similar pattern of deglaciation characterised by a sustained decrease in grass and its replacement with shrub species. The age given at this site for the onset of deglaciation is 17.3 ka and supports the age recorded at Manks Tarn (Newnham *et al*, 2007a). Furthermore, this age of glacial retreat from its LGM maximum extent is supported by Suggate (1990) and pollen proxy records from other sites in the Westland region (e.g. Vandergoes and Fitzsimons, 2003). Likewise the records used in the NZ-I CES, including the proxy records from the North Island, marine core MD97-2121 and the speleothem records, all consistently age the Termination I at occurring between 19-17.9 ka (refer to Section 1.2). Although there is evidence of large glaciers in the Westland region up until ~19.5-19 ka (e.g. Abut Head; Moar, 1980; Denton *et al*, 1995; Suggate and Almond, 2005), the timing of deglaciation (between 19-17.9 ka) is broadly synchronous throughout New Zealand.

Deglaciation began before 18.1 ka, as discussed above, with climate amelioration culminating in the development of *D. cupressinum* dominated podocarp-hardwood forest around c. 10.8 ka. The pattern of vegetation change shows consistent climate amelioration characterised by a progression of successive peaks in firstly shrub taxa *Coprosma* and *Myrsine*, followed by *Griselinia* to tall tree taxa *W. racemosa*, *Prumnopitys* and finally *D. cupressinum*. The pattern of vegetation change is similar to that recorded at Okarito (Newnham *et al*, 2007a), Lake Mudgie (Chapter 4.6; Moar and Suggate, 1996), Gillespie Beach Road (Vandergoes and Fitzsimons, 2003) and other records from the region (refer to Section 2.6.1.2).

Unlike the Okarito record and the NZ-I CES there is no evidence for a LGR. Instead Manks Tarn shows progressive forest development, increased precipitation and continued climatic amelioration up until the Holocene. However, there is strong evidence of a glacial advance in the Westland region in the form of the Waiho Loop moraine. There is also evidence of a LGR in a number of other records from the South Island (e.g. Newnham *et al*, 2007a; Vandergoes *et al*, 2008; Kaplan *et al*, 2010). Vandergoes and Fitzsimons (2003) argue that in the high

rainfall Westland region lowland vegetation may not provide the best proxy for short term and minor temperature cooling. Also, Vandergoes *et al* (2008) suggest that pollen sites which are not located at either high altitude or at ecotonal boundaries would most likely record only minor cooling, such as at Okarito (Newnham *et al*, 2007a) or no reversal at all (McGlone 1995; Vandergoes and Fitzsimmons, 2003). Manks Tarn is only 130 m.a.s.l and in a high precipitation region, away from any vegetation boundaries and was dominated by lowland forest taxa towards the end of the LGIT. The conditions at this site are therefore probably not conducive for recording a LGR. This event, or lack of, will be discussed further in Section 8.1.5.1.

The Manks Tarn record shows climatic amelioration, inferred by the rise in *D. cupressinum* beginning at c. 12.2 ka, culminating by 10.8 ka after which a stable podocarp-hardwood forest persisted and is similar to that found in the contemporary landscape. This early Holocene warmth is not restricted to Manks Tarn but has also been recognised by the NZ-I CES. Records included in that project dated the onset of the early Holocene warmth occurring between 10.6-11.8 ka. This age is not only broadly synchronous regionally and nationally but also both hemispheric and interhemispheric suggesting global synchronicity for the onset of the Holocene warm period (refer to Section 8.1.6 for more detail).

5.6 Conclusion

The Manks Tarn record is consistent with other records in the region both in the timing of particular climate transitions and in the pattern of vegetation change. Deglaciation, marking the end of the LGM began 18.1 ka and is characterised by reduction in grassland and an increase in sub-alpine shrubland. This is followed by vegetation successional pattern from sub-alpine to montane communities and culminating with podocarp-hardwood temperate rainforest. Climate is the primary forcing factor with these vegetation changes reflecting increase in precipitation and temperature at the site over time.

The record does not show a LGR occurring either at an ACR or YD timing. This may suggest the area did not experience a climatic reversal. Alternatively the site

is situated in a climatically insensitive area due to its comparatively low altitudinal location away from an ecotonal boundary, high precipitation and insensitive vegetation. Holocene conditions were reached around 11.2-10.8 ka and are represented by the dominance of *D. cupressinum* and the stable podocarp-hardwood forest which occupies the lowland areas in Westland today.

Chapter 6: Howard Valley

Howard Valley, the third site to be discussed, lies in the Nelson Lakes region, northern South Island. Located between two glacially formed lakes, Lake Rotorua and Lake Rotoiti (Figure 6.1) this site provides a fragmentary record constructed from three river terrace risers that cumulatively span c. 37.7- 18.9 ka. Sediment accumulation beneath the oldest terrace began prior to the onset of the LGCP. This site provides a unique insight into the vegetation and climate spanning from pre LGCP and the entire LGCP including the LGM. As for the Westland region, geological mapping reveals this area was extensively glaciated during the LGM. A site where sediment accumulation that begins pre-LGM and is still preserved is therefore rare. However, due to the low elevation of the river terraces (500 m.a.s.l.) sediment accumulation continued without interference by ice. As a result, the vegetation and inferred climate signal was captured. A recently published beetle and macrofossil proxy record from the same three terrace risers (Marra and Thackray, 2010) provides further scientific interest and opportunity to compare those proxies with the pollen record presented here.

6.1 Regional description

Howard Valley is located in the Nelson Lakes National Park, south-east Nelson region, at the northern limit of the Southern Alps mountain range. It is in an upland area with the mountain ranges reaching 2000 m.a.s.l and valley floors ranging 1200-600 m.a.s.l. The remaining native vegetation is restricted to the upland and steep mountain sides and is dominated by beech forest. *Nothofagus fusca* is the dominant tree species up to 900 m.a.s.l. *N. menziesii* then becomes dominant until 1100 m.a.s.l above which *N. solandri* var. *cliffortioides* dominates up to a clearly defined treeline at 1400 m.a.s.l. The treeline is characterised by a sharp transition from forest to alpine tussock grassland (Wardle, 1985). The lowland and flat valley bottoms have been heavily modified by human activity, predominantly forest clearance for agriculture use.

6.1.1 Climate

Howard Valley is located within the North South Island (NSI) climate zone (Kidson, 2000; Lorrey *et al*, 2007). The prevailing wind regime is from the north-west bringing moisture-laden air masses resulting in high annual precipitation, which averages 1600 mm/yr at Lake Rotoiti. Precipitation falls evenly throughout the year. The region experiences a cool climate with mean annual temperatures of 9 °C at Lake Rotoiti and average summer temperatures of 14 °C and winter temperatures of 4 °C (New Zealand National Climate Centre, 2010).

6.1.2 Glacial history

The contemporary landscape is dominated both by depositional and erosional features; evidence of the glacial history in the region. Throughout the Quaternary, glaciers flowed down from the St Arnaud and Spencer Mountain ranges into the Rotoiti and Rotorua valleys (Suggate, 1965). Glaciers also spilled over a small divide at the southern end of Lake Rotorua between Mt Cedric and the Muntz Range into the Howard Valley. During earlier glacial events glaciers extended down the Howard Valley into the Buller Valley and its two main tributary valleys (Maud and Maggie creek; see Figure 6.1). Large sections of the upper Howard Valley have been smoothed by the ice producing a valley oversized for the river it contains today. Successive lateral moraines step up the eastern and western sides of the valley representing the progressively shorter glacial tongues caused by incising of the valley via both glacial and interglacial erosion processes (Suggate, 1988; Challis *et al*, 1994).

The mid and lower sections of the Howard Valley were not glaciated during the last glacial period and contain extensive outwash aggradation surfaces known as the Speargrass formation (Challis *et al*, 1994). The origin of this material is from the small cirques and main valley glacier at the head of the Howard Valley. Multiple episodes of aggradation followed by incisions has produced a number of aggradation river terraces and exposed terrace risers, which flank both sides of the Howard River and up its tributaries. In the main Howard River Valley the risers reach up to 30 m high (Challis *et al*, 1994). Near the Howard Valley/ Buller River junction the deposits are described as poorly bedded coarse gravels (Suggate, 1988).

Figure 6.1: Map showing the ice limits during glacial periods in the Lake Rotorua and Howard Valley region during the late Quaternary period. Copied and adapted from Challis *et al* (1994). The red dot marks the location of the site area used for this project. The list of ice limits are in age order with the Speargrass limit being the youngest and Manuka the oldest (refer to Table 2.1).

6.2 Local site description

The middle and lower sections of the Howard Valley and its tributaries contain a series of low river terrace risers, reaching >30 m above the valley floor. These terraces have been incised by the river and streams resulting in exposed sequences (Figure 6.2) of gravels, sands and silt layers that also contain organic-rich lenses. Three exposed sections from three different terraces were sampled. Two of these terraces are located up a small unnamed tributary of the Howard River, upstream from Coleman Creek (Figures 6.3, 6.4). The catchment of this tributary is small and restricted to the east by the Muntz Range. The upland area of this catchment is vegetated by native beech forest which is replaced by forest-cleared grassland in the valley bottom. The tributary enters the western side of the Howard Valley (Figure 6.4).

The lower section of the unnamed tributary contains a complex system of terraces at varied heights which do not necessarily pair across the valley. A recent conifer plantation obscures and has potentially modified the surface of the largest terrace. The Howard Road has also been built on top of the largest terrace which may also have modified the terrace morphology. However, the two terraces sampled are smaller, located away from the road and pine plantations and free from human disturbance. The two terraces are adjacent to the small stream which is actively cutting the riser of the terraces to reveal these sections. The third terrace is located in the main Howard River Valley, set back from the active river channel today.

Sampling for pollen analysis used the same riser sections as analysed by Marra and Thackray (2010) for beetle and macrofossil analysis (refer to Section 6.2.1 for an overview of their findings). Section 1, the youngest deposit aged between 20.5-18.9 ka, is a large terrace riser located on the south-eastern bank of the unnamed tributary. Section 3, the oldest deposit (38.9- 28.5 ka; refer to chronology Section 6.4) is located 30 m across the valley from section 1 and the section base is only 0.72 m below the base of section 1. Section 2 is located 200 m downstream from sections 1 and 3, in the main Howard River valley and is aged between 26.2-21.9

ka (refer to the chronology Section 6.4). The base of section 2 lies c. 22 m below the base of section 1.

All three sections are capped by a coarse fluvially derived cobble layer. The contact between the cobble layer and underlying silts shows evidence of downcutting by stream activity above (Figure 6.3). This provides an explanation for the complex pattern in terrace deposition found in the area and how two terrace risers in the same valley at similar elevations can have such disparate ages. This will be discussed in greater detail in Section 6.4.

Figure 6.2: Image of section 1. The black line shows the erosion contact layer of the capping cobble/pebble and coarse sand layer which overlies the section analysed.

Figure 6.3: Image looking downstream the unnamed tributary valley. Terraces can be seen on both sides of the valley. Vegetation is primarily pastoral, although a *Nothofagus menziesii* tree is located in the middle ground. Pine plantations are located in the background and on both the larger terraces either side of the stream valley.

Figure 6.4: Location map of the Howard Valley site. Numbers 1-3 are the three terrace risers analysed. The dark lines represent the location of the river terraces in this valley.

6.2.1 Previous research

Previous research from the region has already been discussed in Section 2.6.2. This section will instead touch upon the research which has been carried out at the site itself rather than provide a regional overview. McLea (1996) produced a pollen record from two samples, one above and one below the KOT layer in the Howard Valley, either near to or at section 2 of this study (see Section 6.3). Grass taxa dominated the pollen assemblage along with Asteraceae, including a large proportion of *Taraxacum*, and Apiaceae. *Fuscospora* was the dominant tree taxon, providing only 10% of the pollen count whilst only trace levels of broadleaved forest and shrub taxa were counted. The pollen assemblage also contained trace

levels of *Nothofagus menziesii*. The author concluded that during the period of KOT deposition the region was dominated by alpine grass and herbfields with forest refuges in sheltered locations in the region, which probably contained a significant presence of *N. menziesii*.

Subsequent research by Marra and Thackray (2010) expanded the temporal and spatial record at the site by producing a beetle and plant macrofossil record from three river terrace risers in the area. Their proxy record extends from 34.5-18.9 ka producing a fragmentary record during late MIS 3 and the entire LGCP. The stratigraphy at the site is complex indicating a dynamic depositional environment, changing from terrestrial organic accumulation in low energy environments to fluvial gravel deposits. Erosional contacts indicate there has been loss in the stratigraphy providing a discontinuous record. The record is summarised below:

Section 3, the oldest terrace riser, is aged between 34.4-28.5 ka. Grass and moss material dominate the macrofossils and woody fragments reaching >10 cm diameter are also present. Abundant *N. menziesii* leaves, indicative of beech forest, are found in the lower riser section, and decreasing in abundance up the section. The beetle taxa are also indicative of a beech forest habitat with the assemblage representative of forest and forest floor habitats.

Section 2 contains the KOT layer which has an age of 25.4 ka. The macrofossil assemblage is dominated by grass material. Two *N. menziesii* and some podocarpus leaves, potentially *P. hallii*, are found below the tephra layer. The beetle assemblage is dominated by moss-habitat taxa with some forest and shrub species represented.

The youngest section is aged 20.5-18.9 ka spanning the LGM period (Alloway *et al*, 2007). The beetle assemblage represents a diverse habitat and includes forest, shrub, riparian and marshland taxa. The riser base is dominated by grass, marsh and moss taxa and one beetle which live in *N. menziesii* leaf litter habitats. Midway up the riser beetle taxa are indicative of stream environments above the treeline. These samples also contain *Pimelea* and *Podocarpus* seeds and mountain totara leaves and indicate the site was either at or just above the treeline. The upper part

of the section contains a reduced beetle abundance and decline in diversity. The majority of the insects relate to marshland environments. Throughout the whole section *Chenopodiaceae* seeds indicative of marshland environments are present along with diving beetles which inhabit slow moving water or stagnant ponds.

Grass is the dominant plant macrofossil, indicating grass dominated the vegetation at the site. Fossil beetle evidence and leaves from *N. menziesii* suggest forest was present at the site from late MIS 3 to early stages of LGCP. No forest leaves were present in the younger sediment. The record shows the forest composition to change over time from *Nothofagus*, in the older samples, to *Nothofagus/Podocarpus*, commonly a sub-alpine shrub component at the altitudinal tree limit (Wardle, 1980), and is indicative of progressively cooling temperatures.

6.3 Stratigraphy

6.3.1 Section 3 (Figure 6.5)

Unit 1 (core: 262-193 cm): The core base is dominated by an unsorted coarse grey silty sand containing gravels reaching 5 cm diameter. A 6.5 cm-thick organic-rich sandy silt, containing wood pieces (aged 37.7 ka), lies between the base gravel section and another coarse sandy gravel section. There is a sharp transition between the underlying organic-rich deposits and an overtopping silty sand and coarse gravel layer with the gravels reaching up to 2 cm diameter. The top 6 cm returns to a dark brown organic-rich silt layer, which overlaps the base of the river terrace riser section.

Unit 2 (199-150 cm): Two dark brown organic-rich silt layers, located at the top and base of this unit, are interrupted by a 20 cm-thick coarse gravel and sand layer. Both organic sections contain plant macrofossils, and the top section contains a piece of wood c. 6 cm long. This piece of wood produced two ages both dating c. 35.7 ka. The bottom organic layer produced an age of 34.4 ka.

Unit 3 (150-42 cm): This unit consists of grey silt containing gravels <1 cm. A single rounded cobble located in the centre of the unit measures <15 cm diameter. The material becomes finer towards the top of the unit which has a sharp basal contact with the overlying organic unit.

Unit 4 (42-37 cm): This unit is a 5 cm-thick dark brown, organic-rich, silt which contains visual plant macrofossils, but no evidence of wood. This section was dated by Marra and Thackray (2010) and aged 28.5 ka.

Unit 5 (37-0 cm): Section 3 appears to have slumped by c. 1 m. Modern tree roots have disturbed the upper unit and destroyed some stratigraphic detail. The exposed part of this unit is 37 cm-thick and consists of grey silts containing gravels and large cobbles reaching <15 cm.

Figure 6.5: Detailed stratigraphy and photographic image of section 3. The units are described in the text. Calibrated ^{14}C ages are located next to the depth scale. The stratigraphy and image of this section does not match due to a step cut into the stratigraphy, marked by the stone, which distorts the true depth of the sequence

6.3.2 Section 2

Section 2 forms part of the major terrace system in the main Howard Valley. A similar section has previously been described in Challis *et al* (1994). Marra and Thackeray (2010) have also described the same section, however their work was restricted to the base of the terrace riser which contains the KOT layer spanning only 160 cm. The description below covers the whole exposed terrace riser and a core which was taken at the base of the riser (see Figure 6.6). This section will be described in four subsections. Refer to Figure (6.6) for each described section location in relation to each other and for a more detailed description.

6.3.2.1 Subsection 2D – Core 1- 534-465 cm

The core was taken directly from the base of the terrace riser, with the assumption that the top of the core lies directly at the base of the riser. During coring, the corer hit a solid, impermeable layer consisting of fluvial boulders which may indicate we had reached the base of the valley fill and the surface of the palaeo-channel.

Unit 1(534-465 cm): The core consists of brown, organic-rich silts from 534-478 cm, interrupted by a 25 cm-thick coarse grey sand layer containing pebble layers. Some pebbles reached >2 cm diameter. A sharp, and potentially erosional contact occurs at the top of the first organic layer and the base of the interrupting coarse layer. The organic-rich silts stop at 478 cm and the top of the core consists of grey banded silts with a few pebble layers. This is consistent with the stratigraphy of the riser base.

6.3.2.2 Subsection 2C: 465-325 cm

Unit 2 (465-441 cm): This unit covers the bottom 24 cm of the riser. Alternating mm-cm wide sand and gravel layers occur up through the stratigraphy. There is no organic material present in this unit.

Unit 3 (441-329 cm): This unit consist of alternating brown and grey silts which are interrupted by fine sands and a pebble layer between 399-391 cm. Plant macrofossils are evident in the organic brown sections but become notably less visible up section. The organic band at 373 cm is interrupted by a 10 cm-thick fine, well sorted yellowy/brown silty sand. This has been ascribed as the KOT layer

(KOT; Campbell, 1986; Challis *et al*, 1994; geochemical analysis in appendix g). The tephra layer is bordered above and below by organic-rich brown silts. Vegetative material directly below the tephra layer appears to have been flattened by the weight of the depositing tephra on top of the plants. Above the tephra is a continuation of alternating brown to grey silts with sediment becoming coarser towards the top of the section and the organic content reducing.

Unit 4 (329-325 cm): The top 4 cm of this riser section is an unsorted pebbly silt layer which is assumed to be part of a slump.

6.3.2.3 Subsection 2B: 359-162 cm.

Unit 5: This coarse unit makes up the whole of section B. The base contains the top of the tephra layer, providing a stratigraphic overlap with Unit 3 of subsection 2C. Grey silts dominate the section above the tephra and become coarser upwards. At 317 cm the silt layer grades into a sand dominated zone, which also becomes coarser upwards. This is overtopped by a coarse sandy unit containing gravels and large boulders. The material is unsorted but becomes finer towards the top. The unit is capped by a 3 cm-thick iron-stained sandy silt providing a stratigraphic link between subsections 2B and 2A.

6.3.2.4 Subsection 2A – top section of terrace riser 2: 162-0 cm

Unit 6 (162-113 cm): The base of subsection A lies on the contact of an underlying iron stained sandy silt and very likely to be equivalent to the top of subsection B. Unit 6 is dominated by dark brown sandy silts, containing fine banding. The unit is organic-rich and contains visible plant macrofossils. The contact between the organic-rich silts and the underlying iron-stained silty sand is separated by a 2 cm gleyed grey silt layer.

Unit 7 (113-25 cm): This is an inorganic unit spanning 113-25 cm of subsection 2A. The base of unit seven is dominated by grey silts, containing 1-2 cm-thick sand bands. The sediment matrix becomes coarser upward with sands dominating at 93 cm. This sand layer is well sorted and contains alternating coarse to fine bands. Further coarsening follows, with the top 49 cm of unit 2 composed of poorly sorted gravels, sands and large boulders.

Unit 8 (25-0 cm): This unit covers the top 25 cm of the section. Alternating mm to cm-thick silt and sand layers occur throughout. However, the base of the unit is dominated by grey silty sands fining to sandy silts towards the top of the unit. Contained throughout this unit are cm-thick dark brown organic-rich silts containing visible plant macrofossils. The top of this section is capped by a section of large fluvial boulders and was not included in the description.

Figure 6.6: Detailed stratigraphy and photographic image of section 2. The units are described in the text. Calibrated ¹⁴C ages are located next to the depth scale.

6.3.3 Section 1

Terrace 1 is the youngest terrace located on the south-eastern bank of the unnamed stream (Figures 6.2 and 6.7). The analysed section spans 2.80 m and is capped by a c. 1m thick unsorted fluviially derived large boulder layer, containing sands and silts. This section has been previously described by Marra and Thackeray (2010).

Unit 1 (325-215 cm): Unit 1 comprises the bottom 110 cm. This unit consist of a 60 cm-thick yellowy brown sandy silt section at the base interbedded with a 5 cm-thick brown organic-rich organic-richsilt. This grades into a 50 cm coarser grey silty sand section, also interbedded with a 11 cm-thick brown organic-rich layer. The bottom organic layer is dated at 20.7-20.3 ka.

Unit 2 (215-195 cm): Unit 2 consists of a 20 cm-thick matrix of coarse sand and fine gravels up to 5 cm diameter.

Unit 3 (195-105 cm): Unit 3 consists of laminated sand and silts becoming coarser towards the top of the unit. The sands and silts are interbedded by 1-3 cm-thick brown organic-rich silt layers

Unit 4 (105-80 cm): Unit 4 consist of a 25cm-thick coarse gravel and sand layer containing rounded cobbles reaching <12 cm diameter.

Unit 5 (80-0 cm): The top 80 cm consist of grey silts interbedded with cm-thick brown organic-rich silt layers. The bottom 10 cm consists of a continuous organic layer forming a sharp contact with the coarse gravels in unit 4. A large cobble layer caps unit 5 and forms an erosional contact. At this point along the terrace the capping cobbles layer is c. 80 cm-thick, however a little downstream this layer extends further into unit 5, being c. 150 cm-thick. At this location it appears to gouge a channel-shaped incision (Figure 6.2). The top of Unit 5 is aged 19.3-18.5 ka.

Figure 6.7: Detailed stratigraphy and photographic image of section 1. The units are described in the text. Calibrated ^{14}C ages are located next to the depth scale.

6.4 Chronology

The chronology for the site was produced using bulk sediment and plant macrofossil samples. The pollen horizons sampled in this project are the same as that used by Marra and Thackray (2010). Therefore the ages reported in that paper can also be applied to the stratigraphy for this project. The terrace ages confirm the complex depositional and erosional environment of the site. As discussed earlier, the terraces are not topographically paired and apart from a few exceptions, units cannot be stratigraphically traced from one terrace riser to another. However, the records can be chronological related, with the site producing a record, albeit fragmentary, spanning 37.7-18.9 ka.

Sample	Fraction	^{14}C age BP	$\delta^{13}\text{C}$ ‰	Reference code	Calibrated age BP (2 sigma)
HV1-179	>125 μm	15950 $\pm$ 170	-28.6	OS-83035	19033 $\pm$ 282
HV2-150	Wood	32800 $\pm$ 1500	-24.8	OS-83036	35833 $\pm$ 790
HV2-150	Wood	32500 $\pm$ 1500	-26.27	OS-83037	35747 $\pm$ 738
HV2-core1-28-29	Wood	34600 $\pm$ 1900	-25.31	OS-83038	37650 $\pm$ 1026
HV3-M3-8-9	>125 μm	18650 $\pm$ 250	-23.67	OS-83039	21917 $\pm$ 374
HV3-core1-68-69	125-63 μm	21900 $\pm$ 380	-28.31	OS-83040	26243 $\pm$ 697
*How 1-7	Peat	17586 $\pm$ 133	-28.5	WK-21131	^a 18856 $\pm$ 405
*How 1-1	Peat	18902 $\pm$ 106	-30.7	WK-21129	^a 20475 $\pm$ 202
*How 3-3	Peat	24349 $\pm$ 1393	-31.2	WK-21564	^b 28526 $\pm$ 1499
*How 3-1	Peat	30004 $\pm$ 433	-30.1	WK-21130	^b 34424 $\pm$ 436
**H3B-3-1	Plant/wood	19050 $\pm$ 270	-22.45	OS-83041	22670 $\pm$ 548
**H3B-3-2	Plant/wood	21500 $\pm$ 360	-22.44	OS-83042	25237 $\pm$ 420

Table 6.1: Radiocarbon age details for Howard Valley. The ages were calibrated using the online OXCal program. * Ages produced by Marra and Thackray (2010), calibrated using ^a Calib Rev. 5.0.1. and ^b converted to calendar years using GRIP calibration (Bard *et al*, 2004). ** Unpublished ages by Vandergoes (pers comm).

The oldest organic-rich layer in section 3 has an age of 37.7 ka which is 19 ka older than the base of section 1 but lies at a similar elevation. Both of these sections are adjacent to a small but active stream. It is therefore likely that section 1 represents a channel fill within the valley fill described in section 3. Evidence from the cobble layer that caps the terrace risers also indicate that this process of infilling stream cuts has been ongoing in the area. Further evidence is shown by Challis *et al* (1994) who photographed the section containing the KOT layer. This image clearly shows the KOT bearing strata occupying a channel cut into a valley fill. The subsequent infilling of channel cuts into older sections can explain the complex chrono-stratigraphic pattern of the three terrace risers.

The KOT layer has contributed to the site chronology. Challis *et al* (1994) already identified the deposit as being the KOT. Recent geochemical analysis of both the major and minor elements of the deposit from this site confirms the original identification by Challis *et al* (1994) (see appendix g). As mentioned in the previous sections the age of the KOT used is currently estimated at 25.4 ka (Vandergoes *et al*, 2011).

The stratigraphic description of the three terrace risers indicates that this was a dynamic setting during sediment deposition. The environment oscillated from a low energy terrestrial environment allowing the accumulation of organic-rich sediments to a high energy fluvial environment capable of carrying and depositing boulder size objects. The erosional contacts, evident in the stratigraphies, indicate a loss in stratigraphic record and thus a discontinuous record. The periods of organic accumulation, therefore represent periods where the site hydrology and geomorphology was such that back ponding could occur, producing the low energy environment needed for organic accumulation. The multiple episodes of aggradation and incision were driven by base level changes probably caused by bedrock constriction and episodic damming in either the Buller River Valley, Howard River valley, or a combination of both. The rate of sedimentation however cannot be assessed with certainty as the amount of deposit which has been lost through erosion is unknown and therefore a continuous age-depth model for each terrace riser cannot be produced. Figure 6.8-6.10 show the chronology in stratigraphic order for each terrace riser and will be discussed below

6.4.1 Section 3

The chronology in section 3 is based upon five ages which date the terrace riser and the core. Marra and Thackray (2010) produced two ages (WK-21130 and WK-21564) using bulk peat samples dating the top and bottom of the riser. A further three ages, sampled from wood and bark, date the top organic band in unit 2 (refer to Section 6.3) and the core. The chronology is discussed below.

The samples OS-83039, OS-83038 and OS-83037 and WK-21564 all lie in stratigraphic order. The samples OS-83038 and OS-83037 used wood and bark encasing the wood to date the same horizon and both samples provided a similar calibrated age of c. 35.7 ka at 150 cm down the riser. However, there is an age inversion at 180 cm where sample WK-21565 produces an age 1300 yrs younger than the ages above. Bulk sediment was used to produce the age for WK21565 which is of greater risk of contamination than isolating a single macrofossil from the sediment. The wood and bark, used to date the 150 cm horizon, were both carefully cleaned and inspected under a microscope to remove any organic or inorganic material which was not part of the wood and thus reducing the potential contamination. The age cannot be ruled out entirely but the probability that the 180 cm organic horizon could be deposited 1400 years after the overtopping organic horizon is unlikely. It was therefore decided to exclude WK-21130 from the record and thus to apply age-depth model 2 to this section (Figure 6.8).

WK-21130 dates the top organic layer of the terrace riser. This age is considerably younger compared to the underlying organic horizon and suggests only 100 cm of sediment accumulated over 6000 years. This appears to be quite a low sediment accumulation rate, especially when comparisons are made to the youngest terrace, located on the opposite bank which accumulated 270 cm over only 1000 years. This could suggest contamination to the sample occurred resulting in a younger age. However, the younger age may also represent cut and fill of the terrace, whereby the older material was eroded away followed by aggradation of younger material. This would explain the large age difference over 100cm of deposit. Evidence of stream cut erosion followed by aggradation is visible throughout the other terrace risers and is the likely cause of the large age difference.

Figure 6.8: Age model for section 3, including the error bars for each AMS age. The model use two AMS ages produced by Marra and Thackray (2010).

6.4.2 Section 2

The chronology for section 2 is based upon four AMS ages taken from the riser and the core and one tephra marker. The AMS ages are derived from either bulk sediment samples or plant macrofossils (see table 6.1). The AMS dates however do not all lie in stratigraphic order and the two ages from the same sediment horizon directly underlying the KOT layer, are different. OS-83041 is younger than the established KOT age (Vandergoes *et al*, 2011) whilst OS-83042 provides a very similar age to the KOT. Two age model is discussed further below (Figure 6.9).

Sample OS-83041 is 3000 years younger than OS-83042 which dates the same horizon and the established KOT, which immediately overlies the sample. A

discussion on the age of the KOT has been presented in Section 3.5.2. Sample OS83041 appears to be considerably younger than the KOT established age, suggesting that the sample may have been contaminated by younger carbon. This age has been excluded from the record.

The second AMS sample, OS-83042, is within error range of KOT (Vandergoes *et al*, 2011). The age is also similar to those from other sites around New Zealand which date the tephra deposit (refer to Section 3.5.2). For these reasons it was decided to apply the calibrated OS-83042 age of 25.4 ka to the age model of this section.

Figure 6.9: Age models for section 2, using calibrated ages and including the error bars for each age. The KOT age is also included on the diagram.

6.4.3 Section 1

The radiocarbon chronology of terrace 1 is derived from two bulk sediment samples produced by Marra and Thackray (2010) and one bulk sample of the 10-60 μm fraction (Figure 6.10). This shows that the period of valley/channel between 18.9-20.5 ka suggesting 279 cm of material was deposited over 1.5-2.0 ka. All three ages lie in stratigraphic order, however the upper two age errors overlap and therefore both samples produce similar ages and may be representative of rapid deposition at the site.

Figure 6.10: Age model for section 1, using calibrated ages and including the error bars for each age

6.5 Results

The results are split into two sections. Firstly the pollen analysis is described (Figure 6.11-6.13) followed by the results of the temperature reconstructions for each section which are also presented in Figure 6.11-6.13. These figures include only the main pollen types, see appendix E for the whole pollen data set.

6.5.1 Pollen analysis

The pollen diagram for Howard Valley is shown in three figures (Figure 6.11-6.13). Each riser stratigraphy and chronology is summarised alongside the diagram. The sampling strategy used in Howard Valley was adjusted due to the nature of sediment deposition at the site. Firstly comparison with the beetle results of Marra and Thackray (2010) dictated the location of sampling. However, a good representation from the organic units identified in each sequence was needed. Therefore samples from each organic unit identified in the stratigraphy were analysed. The pollen results will be described in chronological sequence beginning with the oldest section.

6.5.1.1 Section 3: Terrace riser 3 and core (37.7- 28.5 ka) (Figure 6.11)

Section 3 consists of 12 pollen samples, four from the core and another eight from the exposed riser section. The top two assemblages of the core overlap the base of the riser and provide an opportunity to link the two sections together. The base of the core is dominated by the shrub component, with the small tree *Phyllocladus* at 38%. The shrubs *Asteraceae* and *Coprosma*, and tall tree taxa *Fuscospora* and *Halocarpus* are present at 10% with minor abundance of *Nothofagus menziesii*. *Poaceae* <40µm has the second highest abundance of 19%.

The abundance of *Poaceae* <40µm consistently increases up core peaking at 38% at the top. Conversely, *Phyllocladus* consistency decreases from 38% at the core base to 14% at the core top. *Fuscospora* follows a similar pattern as *Poaceae* <40 µm, although on a smaller scale, reaching a maximum abundance of 15% at the top of the core. *N. menziesii* abundance increases minutely to 2-3 %. The wetland taxon *Cyperaceae* has consistent high abundance throughout the core, ranging 39-29%.

Three organic units were analysed in the riser spanning 199-180 cm, 159-150cm and 42-37 cm (the youngest sub-section). The unit 199-180 is dominated by herb species, particularly Poaceae <40 µm which increases up the unit from 30%-49%. Apiaceae and Poaceae >40 µm are present in significant abundance of c. 10% each. This unit contains a significant amount of shrub species, particularly *Phyllocladus* which reaches a maximum abundance of 21%. To a lesser extent Asteraceae is present with a percentage of only 3%-6%. A limited tree flora presence is in the record with *Fuscospora* spanning 8%-12% and *Halocarpus* only reaching 2%. The wetland taxon, Cyperaceae, reaches its highest abundance in this unit, increasing from 16% to 21% by the top of this unit.

Poaceae <40 µm continues to dominate the pollen assemblage between 160-150 cm, ranging 47%-57% and Poaceae >40 µm also increases to 12%. The total shrub component decreases with *Phyllocladus* ranging between 7%-17%. Asteraceae, *Fuscospora* and Cyperaceae remain at similar percentages as the previous organic unit. Poaceae <40 µm reaches its maximum abundance (67%) at the 43 cm but subsequently declines to only 33% at the top of this unit (37 cm). Apiaceae increases from 5% to 21% and *Phyllocladus* from 9% to 20%. A small increase in the tree taxa *Fuscospora* and *Halocarpus* coincides with the increase in the shrub component. Cyperaceae remains at a similar abundance to the previous zone.

Figure 6.11: Howard valley section 3 pollen percentage diagram, including core stratigraphy, summary diagram, and pollen-climate reconstructions. Radiocarbon dates are summarised in Table 6.1.

6.5.1.2 Section 2: Terrace riser 2 and core (26.2–21.9 ka) (Figure 6.12)

Section 2 comprises one large river terrace riser spanning nearly 5 m and a core 69 cm long. However, only the organic layers were analysed, producing 13 samples covering the whole section. Although a large amount of sediment accumulation between the organic sections occurred, the vegetation composition remained similar throughout the section and therefore the number of samples appears to be sufficiently representative.

The two core samples are dominated by Poaceae <40 µm with an abundance of 51% and 47%. The herb taxon *Taraxacum* has a 12% abundance and Apiaceae 7%. *Taraxacum* decreases to trace amount at the top of the core. The shrub component is dominated by Asteraceae (4%) and *Phyllocladus* which decreases from 17% to 2%. *Fuscospora* remains between 11%-16%. A large increase in the wetland taxon Cyperaceae occurs up the core (from 10% to 24%).

The base of the riser, section 2c, is again dominated by Poaceae <40 µm, spanning 48%-45% between 430-390 cm. Apiaceae remains at a similar abundance as in the core. Likewise *Fuscospora*, Asteraceae, *Phyllocladus* and Cyperaceae have similar percentage as the last sample in the core. At 381 and 371 cm, Poaceae <40 µm decreases dramatically to only 27% and 22%, respectively. Apiaceae becomes the dominant taxon reaching 34% and 29%, respectively. *Fuscospora*, *Phyllocladus* and Cyperaceae decrease in abundance. Conversely, Asteraceae increases to a maximum 18%. The top three samples of riser section 2c shows a gradual increase in Poaceae from 36% to 46% at the expense of Apiaceae, which declines from 23% to 8%. As the depth decreases the abundance of *Phyllocladus* and *Fuscospora* increases and to a lesser extent so does *Hebe* and *Taraxacum*. Conversely the proportion of Asteraceae decreases up the sequence. Cyperaceae reaches a peak abundance of 55% at 358 cm but subsequently decreases to 32% at the top of this section.

The top of the riser, section 2a, consists of three samples taken from two organic zones. The section is dominated by Poaceae < 40 µm ranging 42-52%. Apiaceae shows a continued decrease from 15% to 6%. Asteraceae, *Phyllocladus* and *Fuscospora* all show a similar pattern whereby the percentage increases from c.

7% or below, at the base of section 2a, to 10% in the middle sample before reducing to a similar abundance as in the first sample, at the top of the section. Cyperaceae shows a continued increase up the section from 33 to 57%.

Figure 6.12: Howard Valley section 2 pollen percentage diagram, including core stratigraphy, summary diagram and pollen-climate reconstructions. Radiocarbon dates are summarised in Table 6.1.

6.5.1.3 Section 1: Terrace riser 1 (20.5-18.9 ka) (Figure 6.13)

Section 1 is the youngest terrace and contains eight organic zones, each of which were sampled and analysed. Throughout the section Poaceae < 40 µm, Asteraceae and Poaceae >40µm are the three dominant taxa and are consistently present throughout. Poaceae is the dominant taxon at the base of the sequence with 47% reducing to 30% at 250 cm. Conversely Asteraceae has a small abundance of only 6% at the riser base but increases to 31%, becoming co-dominant at 250 cm. Apiaceae has a percentage range of 14-26% at the riser base and Cyperaceae increases from 43% to a maximum 83% at 255 cm and then reduces to 65% at 250 cm.

From 244-74 cm Poaceae <40 µm is the dominant taxon ranging 40-62%. Overall, Apiaceae abundance increase, but ranges between 10-30%, whereas Asteraceae decreases in overall abundance to 5-15% with the majority of samples contain less than 9%. *Phyllocladus* remains between 4-7% and *Fuscospora* oscillates between 1-8%. Cyperaceae abundance is highly variable in this section ranging 21-65%.

From 70-0 cm the pollen assemblage becomes increasingly variable with large changes in taxa percentages occurring throughout this section. Poaceae varies between 20%-51%, Asteraceae 6%-33% and Apiaceae 17%-47%. The other land pollen, such as *Fuscospora* and *Phyllocladus*, remain at a consistent, but low abundance in this uppermost section. Cyperaceae shows a similar erratic pattern as Poaceae, ranging 24%-61%.

Figure 6.13: Howard Valley section 1 pollen percentage diagram, including core stratigraphy, summary diagram, and pollen-climate reconstructions. Radiocarbon dates are summarised in Table 6.1.

6.5.2 Temperature reconstructions

The pollen diagrams (figures 6.11-6.13) also present the modelled MAT reconstructions produced by PLS-C3 and W-MAT (refer to section 3.2.7). W-MAT reconstructs warmer MAT (between 1-2 °C) than PLS-C3 for all three sections. In Section 3 (fig 6.11) the PLS-C3 model predicts a MAT ranging 3.5-4.8 °C. Initially MAT values oscillate between 3.5-4 °C and then increase to 4.7 °C at 150 cm. MAT values remains steady around 4.7 °C for the remainder of the section analysed. W-MAT reconstructs a considerably smaller MAT range, between 7-7.4 °C, showing little change in temperature throughout the section. Section 2 (fig 6.12) shows a limited MAT range for both models. The PLS-C3 model initial predicts a decline in MAT from 5.6 °C to 4.6 °C by 430 cm, returning to 5.6 °C at 388 cm. A second decline in MAT reaches 4.8 °C at 382 cm which is followed by a steady increase in MAT reaching a maximum of 5.7 °C at 145 cm. Temperatures decrease at the top of this section reducing to 5 °C in the uppermost sample. Like the previous section, W-MAT predicts a small temperature range between 7.1-7.4 °C throughout the section.

For section 1 (fig 6.12), the PLS-C3 model gives an initial increase in MAT from 5.5 °C at 279 cm to 6.3 °C at 151 cm. Temperatures then reduce to a minimum 4.8 °C. This is followed by an overall increasing temperature trend up section, reaching a peak of 6.6 °C at 15.5 cm, decreasing slightly to 6.3 °C at the top of this section. The W-MAT model reconstructs a small temperature range between 7.1-7.4 °C showing little change in temperature throughout the section. Overall, the pattern of MAT reconstructed by PLS-C3 suggests MATs were warmest during the period when the youngest terrace was formed and coldest at the time the oldest terrace was formed. In contrast the W-MAT indicates MAT remained at a near consistent temperature (between 7-7.4 °C) throughout the period spanning 37.7-18.9 ka.

6.6 Discussion

This section interprets the pollen record produced for Howard Valley. This will be separated into pre-LGCP and LGCP time periods, based upon the chronology discussed in Section 6.4 and the timing of the LGCP as ascribed by the NZ-I CES (Alloway *et al*, 2007). An inferred vegetation community followed by a climate reconstruction will be produced for each period. The timing and pattern of vegetation change and inferred climatic change will be put in context with palaeoenvironmental records from other New Zealand sites and elsewhere in the Southern Hemisphere.

6.6.1 Pre-LGCP: Section 3 (37.7- 28.5 ka)

The base of the core is dominated by the shrub *Phyllocladus* combined with a large abundance of Poaceae and Cyperaceae and moderate abundance of *Fuscospora* pollen. There are three species of *Phyllocladus* in New Zealand, two lowland forms and one; *P. alpinus*, a montane/sub-alpine species (Wardle, 1969; McQueen, 1992) although it is also found in the lowland forest in the Westland region (Patel, 1968). The pollen assemblage in the core represents a montane/sub-alpine vegetation community and *P. alpinus* is prevalent in the region today. It is therefore assumed the *Phyllocladus* pollen in the samples represent *P. alpinus*. In the contemporary environment of the Nelson Lake National Park, *P. alpinus* shrub sparsely covers large areas above the treeline (Wardle, 1965) and the contemporary treeline is characterised by a sharp transition from *Nothofagus solandri* var *cliffortioides* sub-alpine forest to *Chionochloa* grassland (Wardle, 1965). The combination of *Fuscospora*, Poaceae and *Phyllocladus* could lead to an assumption that the site was representative of a sub-alpine/alpine treeline environment. However, the small but significant abundance of *N. menziesii* would suggest otherwise.

N. menziesii is a montane beech species which in the Nelson Lake National Park is the dominant forest tree between 900-1100 m.a.s.l. The *N. menziesii* pollen abundance is low at the site, reaching only 1 % at the core base and a maximum 3% at the top of the core. At first glance this could indicate that *N. menziesii* is not prevalent in the landscape. However, Moar (1970) studied the modern pollen rain

at Top House and St Arnaud, both located in the Nelson Lakes National Park. The sites studied were located within a mixed *Nothofagus* forest but the pollen rain at both sites recorded only 3% *N. menziesii* and suggests this species is severely underrepresented in the pollen rain due to its poor ability to disperse. Therefore the 1-3% *N. menziesii* in the pollen record would probably be representative of a large *N. menziesii* population in the region and suggest the region was not representative of the sub-alpine/alpine treeline limit.

Wardle (1965; 1969; 1991) reports *Phyllocladus alpinus* growing in poorly drained, boggy areas or in wet hollows at elevations below 1000 m.a.s.l, in *Nothofagus fusca* and *N. solandri* forest. In these environments *Coprosma parviflora* and *P. alpinus* form a shrub layer whilst *C. depressa*, sedges and ferns cover the hollow floor (Wardle, 1965). The large abundance of Cyperaceae suggests the site is waterlogged and represents a wetland habitat. Combined with the *P. alpinus* and small abundance of *Halocarpus*, which can also grow in waterlogged environments (Wardle, 1977), indicates that at 37.7 ka the area was characterised by mixed montane beech forest comprising *N. menziesii* and *Fuscospora*, with wetland habitats located near the rivers, which were dominated by *P. alpinus* shrub and sedges. A decreased sample depth in the core and riser corresponds with a decrease in *P. alpinus* and Cyperaceae abundance and an increase in *Fuscospora*, Poaceae and to a considerably lesser extent *N. menziesii*. This vegetation pattern probably represents a drying of the site causing a decline in sedges and other wetland vegetation whilst providing an increase habitat suitable for beech and grass which expanded into the site. Poaceae is the dominant vegetation type in the record between 37.7-34.4 ka.

Unit 4, the top organic layer in the riser, is void of *N. menziesii* and contains only a small abundance of *Fuscospora*. The shrub component is small but *Phyllocladus* remains the most abundant shrub taxon in a Poaceae dominant vegetation community. The *Phyllocladus* abundance is not combined with a large Cyperaceae abundance and suggests the area was not as waterlogged as the period spanning 37.7-34.4 ka. Rather, the pollen assemblages suggest a *Phyllocladus* shrub/grassland mosaic. As mentioned above, the contemporary timberline, east of the Southern Alps, is characterised by a sharp transition from *Nothofagus*

solandri var. *cliffortioides*-dominated sub-alpine forest to *Chionochloa* grassland. Patches of shrubland comprising *Phyllocladus alpinus*, *Hebe*, *Coprosma*, *Dracophyllum uniflorum*, *D. pronum* and *Podocarpus nivalis* grow at this transition point (Wardle, 1965). This community is consistent with the pollen assemblage at the top of the terrace riser and indicates the site is representative of a sub-alpine/alpine community at, or near to, the timberline. The pollen assemblage contains significant amounts of Apiaceae, alpine herb taxa and a comparatively increased amount of tussock grass pollen compared to the underlying organic units and again supports an alpine/sub-alpine/tree limit environment.

The pollen assemblage at the top of the riser also contains trace amounts of broadleaved/podocarp species, including *Weinmannia racemosa*, *Dacrydium cupressinum* and *Prumnopitys*. These taxa typically occur at lower altitudes than the beech forest in the region today and can be located in scattered groves within the beech forest up to 650 m.a.s.l. The trace amounts of these podocarp and broadleaved species and their limited distribution in the modern landscape suggests the nearest podocarp/broadleaved forest to the site would have been close to the present sea level (McLea, 1996) and the pollen grains were transported to the site as opposed to growing in the area.

6.6.2 Pre-LGCP: Climate reconstruction

The pollen assemblage at the base of the core is representative of a boggy wetland hollow surrounded by a mixed beech forest containing *Fuscospora* and *N. menziesii*. In the Nelson Lake National Park today *N. menziesii* is the dominant tree taxon between 900-1100 m.a.s.l. As the site is located at c. 500 m.a.s.l today this implies between 37.7-34.4 ka *N. menziesii* was 400-600 m below its current preferred elevation. By using the environmental lapse rate (ELR) of 0.5 °C/100 m, an estimated minimum 2 °C and maximum 3°C MAT depression occurred at the site, suggesting that although warmer than the LGCP, climates during late MIS3 were cooler than at present.

A vegetation transition from montane beech forest between 37.7-34.4 ka to sub-alpine/alpine dominated vegetation is represented at the top of section 3. The

exact timing of this change was not recorded due to the fragmentary nature of sediment deposition. The top of the section is dominated by Poaceae and the cold tolerant *Phyllocladus* and other elements of the sub-alpine/alpine zone. *Phyllocladus* is restricted in the contemporary environment to the *N. solandri* tree limit, between 1100-1400 m.a.s.l. This suggests that climate was considerably cooler than present day by 28.5 ka, which depressed the treeline down to 500 m. By using the same ELR as above, a MAT depression of 3-4.5 °C is estimated to have caused a 600-900 m lowering of the treeline. This suggests temperatures were getting progressively colder leading into the LGCP.

6.6.3 LGCP: Section 2 and 1 vegetation interpretation (26.2- 18.9 ka)

The dominant vegetation types throughout section 2 and 1 are grassland and herb communities, indicating alpine-dominated vegetation, which is located above the treeline in the contemporary landscape (Wardle 1965). A similar vegetation composition has been reported from various other sites in the region which span the LGCP (Moar, 1980; Mabin, 1983; McLea, 1996). The low pollen abundance of podocarps; *Halocarpus*, *Phyllocladus*, and *Fuscospora* and trace amounts of *N. menziesii* indicates that there were tree refugia in the region.

Significant amounts of Asteraceae remain present in the landscape throughout section 2 and 1, and may be indicative of a shrub element but the prominence of grass and other herb taxa suggest the Asteraceae is more likely to represent herb forms. Moar (1970) carried out a pollen spectra study from a high altitude site (1700 m.a.s.l) between Lake Rotoroa and Lake Rotoiti. The following herb taxa were present in the pollen rain: Asteraceae taxa (incl. *Abrotanella* sp, *Celmisia sessiliflora* and *Craspedia uniflora*), the Apiaceae *Anistoma aromatic*, *Coprosma pumila*, *Gaultheria depressa* and Poaceae taxa (incl. *Chiononchloa australis*, *C. pallens* and *Hierochloe*). The alpine vegetation composition described by Moar (1970) was very similar to the taxa forming the pollen assemblage at the site indicating the regional vegetation during 26.2-18.9 ka is dominated by alpine flora. This would also suggest the Asteraceae in the record represents herb forms of the taxa.

Overall the pollen assemblage shows little change in composition between sections 1 and 2 which are both consistently dominated by Poaceae and other alpine herb taxa including *Hebe*, Apiaceae, Brassicaceae, *Gentiana* and Asteraceae, all indicative of very cold climates. However, *Fuscospora*, varies in abundance between the two sections ranging 18-10 % in section 2 and 8-1% in section 1. Coinciding with the *Fuscospora* decline in section 1 is an increase in herb abundance during the LGM. The relatively high abundance of *Fuscospora* in section 2 could suggest that beech forest is at or near the site and that the treeline limit was close by. However, the lower abundance of *Fuscospora* in section 1 suggests beech forest declined in the region at that time. *Fuscospora* pollen can be transported over extremely large distances, which would be made easier in open landscapes dominated by herb taxa, as the pollen assemblage in the record suggest. During the LGCP, other sites in the Westland region, which are suspected to have been beech free, consistently contained up to c. 10% *Fuscospora* pollen abundance (Vandergoes and Fitzsimons, 2003; Newnham *et al*, 2007a; Lake Mudgie). Modern pollen rain studies have found *Fuscospora* pollen to be the most widely dispersed mainland pollen type and it is even found in remote islands including the Antipodes 770 km away from New Zealand (Moar, 1967, 1970; Dodson, 1978; Mildenhall 1976). The abundance of *Fuscospora* at different sites and the sites' distance from source appear not to share a relationship (Moar, 1969; 1970) although vegetation cover does appear to play a role. A pollen rain study in central Canterbury Plains, which is 60 km downwind from *N. fusca* forest, found 21 % *Fuscospora* pollen in exposed shrubland but only 7% in dense shrubland (Myers, 1973). This illustrates the importance of an open low-lying vegetation community in promoting *Fuscospora* dispersal. Considering the pollen assemblage is dominated by alpine herb and grasses the *Fuscospora* pollen probably came from lower altitude refugia and was blown up to the site. Section 1 has significantly lower abundance than section 2 suggesting the period between ca 20.5- 18.9 ka experienced harsher and colder conditions were causing the beech forest to migrate to more favourable conditions at lower altitudes.

6.6.4 LGCP: Climate reconstruction

The pollen composition changed little between sections 2 and 1 apart from the changes in *Fuscospora* abundance that is significantly lower during the LGM. This suggests climatic conditions between 26.2-21.9 ka were not as cold as the conditions experienced in the region during the later LGM. The fragmentary nature of the record does not reveal conditions during the period preceding 26.2-21.9 ka, but the pollen assemblage is similar to that at the top of section 3 and suggests similar conditions were recorded in both section. However, the reduction in sub-alpine treeline taxa such as *Phyllocladus* and increase in alpine herb taxa suggest the site was not at the treeline during 26.2-21.9 ka.

In the contemporary landscape, the alpine zone is located, at >1400 m.a.s.l in the Nelson Lake National Park. By using the standard ELR of 0.5 °C/100 m (Burrows, 1977), a minimum of 4.5 °C cooling is needed to reduce the alpine zone from 1400 m.a.s.l to 500 m.a.s.l Therefore the climate during the LGCP including the LGM was at least 4.5 °C colder than at present, reducing the MAT at Lake Rotoiti to c. 3.5°C, similar to the current average winter temperature of 4 °C (New Zealand National Climate Centre, 2010). Cyperaceae, indicative of poorly drained, boggy and wetland areas, are consistently present in both terrace risers with a large abundance throughout section 2 and 1. The site therefore remained intermittently waterlogged between 26.2-18.9 ka. This does not necessarily indicate a precipitation signal but rather periods of poor drainage at the site caused by damming and subsequent backponding.

6.6.5 Quantitative pollen/climate reconstructions

The pattern of MAT change predicted differs between the two models (see section 8.2.4) and with the above estimated pattern of climate change using ELR and the vegetation record. The pattern of MAT predicted by the W-MAT model suggests temperature remained relatively unchanged throughout the period spanning 37.7-18.9 ka. The model estimated the MAT to be 1.4-2 °C cooler than present which during some periods is significantly lower than that estimated by the ELR (see sections 6.6.5 and 8.2.4). The pattern of MAT change predicted by the PLS-C3 model suggests temperatures got progressively warmer with the reconstruction

and that MIS 3 was colder at Howard Valley than during the LGM. The pattern of climate change reconstructed by both models is strikingly different to that inferred from the vegetation record and using the ELR which suggest MAT were at least 1.6 °C warmer during the late MIS 3 than during the LGM. Records from elsewhere in New Zealand also support the MAT pattern as estimated by the ELR and inferred from the vegetation record (e.g. Alloway *et al*, 2007; Newnham *et al*, 2007) in contrast to the modelled results. Therefore, the pattern of MAT change reconstructed by both models seems unrealistic. The causes of the different reconstructions are likely to be attributed to the limitations of the quantitative methods and are discussed in detail in Section 8.2.4.

6.6.6 Discussion summary

Pollen assemblage zone	Vegetation reconstruction	Climatic reconstruction
Howard Valley sections 2 and 1 (20.5-18.9 ka)	Alpine vegetation dominated by grassland and herb field species. <i>Fuscospora</i> abundance declined suggesting beech forest moved further away from the site to more favourable conditions.	Extreme cold environment with a minimum 4.5°C cooling experienced at the site.
Howard Valley section 2 (26.2- 21.9 ka)	Alpine vegetation dominated by grassland and herb field species. <i>Fuscospora</i> pollen suggests the site was at or near the upper limit of the beech treeline.	Colder climates than that experienced at the site today, estimated between 3-4.5 °C colder. The section indicates a progressive cooling at the site.
Howard Valley section 3 (28.5 ka)	Transition from mixed beech forest with wetland groves, to an alpine/sub-alpine grassland dominated community at or near the treeline.	Colder climates than that experienced at the site today, estimated between 3-4.5 °C colder. The section indicates a progressive cooling at the site.
Howard Valley section 3 (37.7- 37.4/ 34.4 ka)	Wetland community dominated by sedges and <i>Phyllocladus alpinus</i> shrub which is surrounded by a mixed <i>Fuscospora</i> and <i>N. menziesii</i> beech forest.	Cooler climates than is experienced at the site today estimated between 2-3 °C colder.

Table 6.2: A summary table of the vegetation and climate reconstruction for Howard Valley pollen assemblage for section 1-3.

6.6.6.1 Summary discussion for pre-LGCP (section 3)

Howard Valley is a unique site in the region as it is one of a handful of sites which produces a terrestrial record extending back into MIS 3. MIS 3 is the milder interval spanning 51-24 ka and interrupts the two glacial periods of MIS 4 and MIS 2. The vegetation reconstruction at the site between 37.7-34.4 ka is a *Fuscospora*

and *N. menziesii* beech forest containing *Phyllocladus alpinus* and Cyperaceae-dominated wetland areas along the main river margins and up the side creeks and tributary streams. At the same sample locations used for this study, Marra and Thackray (2010) found abundant *N. menziesii* leaves and beetles associated with forest and forest litter habitats and confirm the pollen based vegetation reconstruction. As mentioned earlier, a modern pollen rain study by Moar (1970) found only 3% *N. menziesii* pollen in an area containing a large *N. menziesii* population and suggest this species is underrepresented in the pollen record. The abundant *N. menziesii* leaves at this also show that *N. menziesii* is underrepresented in the palaeorecord and highlights the significant importance of the low abundance of this species when reconstructing past vegetation and climates. Therefore at this site the pollen, beetle and plant macrofossil records complement each other and indicate the importance of a multiproxy approach.

N. menziesii forest in the region today is limited to 900-1100 m.a.s.l and suggests that this species was 400-600 m lower during MIS 3 than its current altitudinal limit. This represents a temperature depression of 2-3 °C cooler than present, indicating MIS 3 in New Zealand was cooler than present day but not as cold as during the LGCP (see Section 6.6.2). Marra *et al* (2009) reported beetle and plant macrofossil evidence from Taranaki during late MIS 3, indicating that taxa which are now restricted to the cooler north South Island were able to grow at the site during MIS 3. This evidence supports the findings at Howard Valley. Likewise the pollen record from Okarito, in south Westland is described as having milder climates during late MIS 3 than during the LGCP but experienced progressive moderate cooling represented by a sequence of minor grass peaks (Newnham *et al*, 2007a). In central North Island McGlone and Topping (1983) estimated temperatures were up to 3 °C cooler than that experienced in the region today, which corresponds with the MAT depression implied by the elevation of *N. menziesii* growth at Howard Valley site during 37.7-34.4 ka. It is therefore reasonable to estimate MAT would have been around 6.6-5.4 °C in the region, based upon a 2-3 °C reduction from the contemporary MAT of 9 °C observed from Lake Rotoiti, during MIS 3.

N. menziesii disappears from the record by 28.5 ka when the vegetation composition is dominated by grass and alpine/sub-alpine vegetation communities demonstrating significant cooling had already commence. An estimated temperature depression of 3-4.5 °C suggests climates were becoming less favourable for *N. menziesii* growth in the region. The plant macrofossil record in Marra and Thackray (2010) also indicates a decline in *N. menziesii* leaf abundance up the riser profile, again, corresponding with the pollen evidence of climate deterioration and migration of trees from the site probably to more suitable conditions at lower elevations. The temperature depression and pollen assemblage is similar to that experienced during the LGCP (see Section 6.6.3).

The onset of the LGCP, according to the NZ-I CES, began between 29-28 ka, the timing of which corresponds with the timing of deposition of unit 4 in section 3. However, due to the record being fragmentary, a precise age for the onset of the LGCP cannot be estimated. At 28.5 ka in Howard Valley herb taxa, particularly grass taxa, were already dominating the vegetation community. This may suggest cooling in the region began earlier, leading to alpine dominated flora becoming established in the region by 28.5 ka. Alloway *et al* (1992) and Nelson *et al* (1993) recorded a peak in aerosol quartz accumulations, which are indicative of increased glacial activity, occurring round 34.4 ka. Likewise, Suggate and Almond (2005) aged a glacial advance in the Westland region also beginning at 34.4 ka. An earlier onset of climate deterioration at the site is therefore feasible and would explain the large transition from montane mixed beech forest to alpine/sub-alpine grassland shrub mosaic. However, due to the fragmentary record of the site, firm conclusions cannot be made.

6.6.6.2 Summary discussion for the LGCP

Marra and Thackray (2010) concluded that the dominant vegetation type throughout the LGCP was grassland with trees and shrubs becoming scarce or absent in the youngest sediments spanning 18.9-20.5 ka. The pollen record shows a similar pattern with sections 2 and 1 dominated by grass and herb taxa indicative of the alpine zone. However the abundance of *Fuscospora* pollen ranged 18-10% in section 2 and only 8.1% in section 1 suggesting a decrease in *Fuscospora* in the region and a two step cooling. The pollen assemblage in section 2 is similar to

that recorded in unit 4 of section 3 and appears to be a continuation of this vegetation community and inferred climate. Section 2 spans 26.2-21.9 ka and corresponds with the LGCP milder and more variable climate. The pollen record from Okarito is characterised by a decline in herb taxa and an increase in tree and shrub abundance marking a mild climate amelioration. At Okarito this milder interval is bounded by two grass peaks, the older representing the onset of the LGCP and the latter the LGM (Vandergoes *et al*, 2005; Newnham *et al*, 2007a). At Howard Valley, a change from very cold conditions during the early LGCP to a milder period of the LGCP was not recorded. Instead the pollen record spanning 28.5-21.9 ka suggest conditions were cold with MAT ranging 3-4.5 °C colder than today. Whether conditions got even harsher between the top of section 3 and the bottom of section 2 is unknown.

Evidence from the plant macrofossil and beetle analysis by Marra and Thackray (2010) concluded that between 26.5-20.7 ka the site was a mixed podocarp beech forest. However, the plant macrofossils contained mainly grass remains and include tussock material. Also, the *Podocarpus* leaves found in the samples could not be accurately identified but were assumed to be *Podocarpus hallii* which spans from the lowland to the sub-alpine zone (Poole and Adam, 1963). Marra and Thackray (2010) suggest the mixed forest community is of a similar composition to sub-alpine scrub located at the altitudinal tree limit in the modern environment (Wardle, 1980). The pollen evidence concurs with this conclusion. Grass pollen dominated the assemblage, combined with herb, *Fuscospora* and *Phyllocladus alpinus* pollen representative of the contemporary subalpine/alpine treeline limit (Wardle, 1980). However the original use of forest to describe this vegetation community by Marra and Thackray (2010) is misleading and this is a disparity between the interpretations. In this instance, to prevent misinterpretation and confusion, a accurate and description classification of the vegetation community, opposed to a vague and ambiguous classification such as forest, should be used instead. McGlone *et al* (2011) have argued that the difference in interpretation of beetle and pollen records can also be attributed to the vocabulary used and the different definitions of particular words such as forest. It is important that the same vocabulary is used consistently between different proxies.

The deposition of section 1 occurred during the LGM period of major ice advance in the region. Suggate (1965) estimated the ice extent reached 650 m.a.s.l at Lake Rotoiti and 450 m.a.s.l at Lake Rotoroa during the LGM. Howard Valley, located between the two was therefore glaciated in the upper valley but the project site, which is located at 500 m.a.s.l, would have been ice free. The pollen record shows an increase in herb taxa and a decline in *Fuscospora*, suggesting the migration of beech and a lowering of the treeline in the region. Again the pollen record corresponds with Marra and Thackray (2010) record where shrubs and forest become absent from the record, indicative of extreme cold environments in the region. The trace amounts of *N. menziesii* suggest there is probably a significant presence of this tree in LGM refugia in the region.

In the contemporary environment, the winter snowline has a large influence on forest growth (Wardle, 1965). Soons (1979) estimated the snowline limit during the LGM along a transect located 42°S, 12 km south of Lake Rotoroa. She concluded that the snowline lowered from east to west and a maximum overall snowline lowering of 800-850 m. In the Nelson Lakes National Park this would have caused a lowering of the snowline from its contemporary position of 1400 m.a.s.l (Wardle, 1965) down to ~ 550-600 m.a.s.l which is close to the Howard Valley site elevation. In the Westland region today forest can grow near to the ice margin and it could be possible that this occurred during the last glacial (Gage, 1945). However, New Zealand shrub and tree taxa are comparatively cold intolerant when compared to Northern Hemisphere continental taxa, and so it is unlikely that these species would grow above the snowline and also cope with periodic extreme cold events, however long these lasted (Sakai and Wardle, 1978). Although, technically, trees could grow in areas up to 650 m.a.s.l, the extreme climate may have limited their growth to areas of more favourable and unexposed conditions. McLea (1996) concluded that considering glaciers did extend below the treeline in some areas in south-east Nelson, 500 m.a.s.l serves as an approximation for the upper limit of trees. At 500 m.a.s.l, the Howard Valley site is located on the edge of the treeline during the LGM. However the pollen beetle and plant macrofossil records evidence suggests that trees were located near to the site but the site itself was forest free.

Wardle (1973) suggested that the altitudinal treeline is controlled by mean summer temperatures. Temperatures reaching an average 10 °C during the warmest months are therefore related to treeline (Wardle, 1965, 1985; Wardle and Campbell, 1976). Summer temperatures are a critical component for *N. fusca* growth as the growing season needs to be long enough to enable the seedlings to harden before the onset of colder weather (Wardle, 1973). The mean annual summer temperature in the region at 600 m.a.s.l is 14 °C (Lake Rotoiti; New Zealand National Climate Centre, 2010). The inferred MAT depression based upon the pollen evidence at the site was 4.5 °C. However, the elevation reached by the alpine vegetation advancing down the valley is unknown and the temperature depression could well be lower. McGlone *et al* (1993) estimated the maximum temperature depression during the LGM was 5 °C, similar to that estimated from the pollen record here, which if subtracted from the contemporary average summer temperature, would place Howard Valley at or very near to the possible limit of tree growth during the LGM. However, it is also important to note that temperature is not the only climatic factor influencing forest growth. It is believed that the LGM also experienced an increase in windiness, frost and reduced precipitation, all of which can limit tree growth (McGlone *et al*, 2010).

6.6.7 Conclusion

Howard Valley uniquely provides a vegetation record spanning from 37.7-18.9 ka. This site indicates that climate during the later stages of MIS 3 in the Nelson region was considerably cooler than current conditions, but milder than during the LGCP. Nevertheless it appears that climate deterioration had already started by 28.5 ka with the region experiencing cold temperatures and alpine dominated vegetation. The onset of the vegetation transition to sub-alpine/alpine vegetation and cooling was not captured in the record. However, significant change had occurred between the two periods of organic accumulation and possibly suggests an earlier onset of the LGCP than in the NZ-I-CES (Alloway *et al*, 2007). The vegetation record confirms that temperatures during late MIS 3 were cooler than present, possibly 2-3 °C colder. The transition into alpine/sub-alpine dominated taxa at 28.5 ka suggest conditions at Howard Valley were significantly colder than today with similar temperatures as that experienced during the LGM.

The LGCP was recorded in section 2 and 1 (spanning 26.2-18.9 ka). The vegetation record indicates the region experienced a period of severe cold with temperatures of at least 3-4.5 °C cooler than at present. The organic section analysed suggests a two step cooling with temperatures possibly milder between 26.2-21.9 ka before the harsh LGM conditions followed. The pollen record produced by this study corresponds closely to others in the region which covers the same time period, including the beetle record produced by Marra and Thackray (2010).

Contrary to findings elsewhere in New Zealand (e.g. Burge and Shulmeister 2007b) the pollen and beetle record are in close agreement throughout the record. The low pollen counts of *N. menziesii* mask the species importance in the environment which has been supported by both the beetle and plant macrofossil analysis in Marra and Thackray (2010) and emphasises the importance of using a multiproxy approach when producing vegetation and inferred climate reconstructions.

Chapter 7: Forks Lagoon

The final site investigated in this project is Forks Lagoon, a shallow lake located in the Mackenzie Basin, central Canterbury (Figure 7.1). This upland plateau lies to the east of the Southern Alps, and, climatically, forms a contrasting site to the previous three sites already discussed. During the Quaternary the area was extensively glaciated, however, geological evidence indicates that Forks Lagoon is located outside the LGM ice margin. Sediment accumulation at the site therefore continued beyond the LGM and throughout the entire LGCP. Due to the dry climate and high altitude of the region, a continuous record extending beyond the LGM is exceptionally rare and makes this an important location. However, the site has its limitations. The dry land pollen counts do not produce statistically significant counts and therefore cannot produce detailed vegetation or inferred climate reconstruction. Also, initially it was thought the section analysed spanned the entire 30-8 ka period. However, after dating, the record was found to only span 31.4-17.2 ka. However, the aquatic macrophyte taxa produced abundant grains, which have proved to be important climate indicators.

Due to the low dryland pollen counts the structure of this chapter will be different to the other three. PCA is not included in the chapter as it was deemed not to provide any useful information due to the low pollen counts. An estimated temperature and treeline depression for the period using ELR was also not carried out as an accurate estimation of both would not be possible. Instead the emphasis in the chapter will be upon the macrophyte pollen concentration. The sedimentology and quantitative pollen/temperature reconstruction will also be discussed.

7.1 Regional description

The Mackenzie Basin is bounded to the north and west by the Southern Alps; to the east by the Two Thumb, Rokesby, Dalgery and Kirkliston ranges; and to the south by the Ohau and Waitaki Rivers. The Grampian Mountains jut into the Mackenzie Basin from the Kirkliston and Dalgery ranges. The geomorphology of the basin is dominated by glacial landforms. Extending from the Main Divide and

considerably southwards of Lakes Tekapo and Pukaki are extensive moraines. The moraines are made up of till and gravels. Beyond the moraines lie an extensive outwash plain that has been dissected by rivers and streams. The material deposited on the plains is well graded greywacke gravels (Mildenhall, 2001).

Figure 7.1: Location map of Forks Lagoon.

7.1.1 Climate

Mackenzie Basin is located in the interior region of New Zealand and is surrounded on all sides by mountains ranges which cause a rain shadow effect and an important influence on the precipitation in the region. This region has low mean annual precipitation averaging ca. 600 mm/yr and can experience regular water deficits and periodic severe drought, producing a fire prone landscape. Mean annual temperature is 8.8 °C, however, there is a large seasonal difference ranging from an average 14.5 °C during the summer to 2.6 °C during winter. Extreme temperatures events also occur in the region with temperatures reaching up to 31 °C and down to -15 °C (New Zealand National Climate Centre, 2010). This area is also susceptible to the dry north-westerly (föhn) winds which increase the level of evapotranspiration, however, the prevailing wind direction is westerly. The climate in the Mackenzie Basin is vastly different to Westland (refer to Section 4.1). The Mackenzie Basin lies on the boundary between East South Island (ESI) and West South Island (WSI) climate regimes with the study site, Forks Lagoon, situated in the ESI (Lorrey pers comm). This site therefore responds differently to the prevailing synoptic conditions compared to the Westland region. During Zonal regimes, maximum and minimum temperatures are above normal, and precipitation is below normal. In Blocking regimes, maximum temperature is above normal, minimum temperature is below normal and so is precipitation. Finally, during Trough regimes maximum and minimum temperatures are below normal and precipitation is above normal (Renwick, 2011). Therefore, Forks Lagoon would receive most of its rainfall during periods dominated by the Trough regimes (see Section 2.3 for more detail).

7.1.2 Vegetation

Human activity has nearly totally transformed the original vegetation cover of the region. Any natural vegetation cover is scattered and restricted to small, and possibly atypical, remnants. Tussock grassland dominates the vegetation and is dominated by *Chionochloa rubra* and *Festuca novae-zealandiae*, a short tussock grass. Podocarp shrubland is dominated by *Halocarpus bidwilli* and includes a number of other shrubs including *Phyllocladus alpinus*, small leaved *Coprosma* spp, *Aristotelia fruticosa*, *Melicytus alpinus*, and other species. The region also

contains a few scattered stands of *Nothofagus solandri* var. *cliffortioides*. However, the majority of the vegetation has been modified by human activity that it is dominated by grassland which is utilised for farming. Many non-native species have also been introduced and thrive in the environment (McGlone and Moar, 1998).

7.2 Site specific description

Forks Lagoon (170°22'56.9"E, 43°56'35.5"S) is a small lake located in the Tekapo Military zone, lying between Lake Pukaki and Lake Tekapo. The lake is impounded by a lateral moraine that is assumed to pre-date the LGM (see Figure 7.2). There is no evidence of groundwater or surface water input to the lake, suggesting the lake is predominantly rainwater fed and therefore sensitive to changes in precipitation. It is evident that lake levels fluctuate, with a dry lake basin and an ephemeral outlet stream south of the lake indicating lake levels have been higher in the past. As there is no overland source of water into the lake. The ephemeral stream (Figure 7.3) controls the maximum lake depth.

Figure 7.2: Map showing the location of moraines in relation to Forks Lagoon (circle). The red marks the location of LGM moraines and green are deposits older than the LGM. Forks Lagoon is located within a red area suggesting the moraine it lies within is an LGM moraine. However, subsequent work in the area, combined with the KOT in the core confirms the bounding moraine was deposited significantly before the LGM (Barrell *et al*, unpublished).

The lake today contains two basins (Figure 7.4). The first basin at the southern end of the site is currently dry. The second basin forms a shallow lake reaching a maximum current depth of 40 cm. The lake itself is heavily dominated by macrophytic freshwater plants, particularly *Isoetes alpinus*. There is a lake edge effect caused by wind-induced waves where the stones bounding the lake are free of vegetation. *Limosella* sp forms a mat covering the stones on the lake edge. Interspersed between the stones, above the wave effect, are a number of herb taxa, the majority of which are non-native. These include Brassicaceae, *Polygonum* spp, *Senecio jacobaea* and *Taraxacum officianale*. The lake is surrounded by *Poa* dominated tussock grassland that is characteristic of this region (Figure 7.4).

Figure 7.3: Aerial photographic image of Forks Lagoon showing the location of the ephemeral stream at the southern end of the Lagoon. Source: Google Earth.

Figure 7.4: Image of Forks Lagoon taken from the south-west/ looking north-east. The black circle illustrates the core location.

7.3 Stratigraphy

The core site was located in the centre of the northern end of the lake, which is the deepest part. Two cores were recovered, the first core, 0901, was taken on the lake bed surface, and the second, 0903, began 50 cm below the lake bed surface. The method for coring is described in Section 3.1. A maximum of 8 m of sediment was cored without reaching the base of the deposit. The description of the stratigraphy is shown in Figure 7.5, and includes only the sections analysed for pollen and sediment analysis. Due to the overlap between the two cores and the distinctive stratigraphy which could be correlated between cores, a continuous analysis down the core could be produced.

Sections from each core were combined to produce a continuous record spanning 206-418 cm. This section of the core consists of silts throughout. The main variation down the core is the sediment colour which changes distinctively and abruptly from light grey to brown and provides distinct colour units. Figure 7.5 shows a detailed core description which is summarised below on a continuous depth scale.

206-245 cm: Brown silts which contain alternating light brown, brown and dark brown bands, usually 2-3 cm thick, although some are thicker. A grey band interrupts the organic bands between 234-238 cm.

245-252 cm: Grey silt layer.

252-302.5 cm: Alternating thick (10-30 cm) bands of grey-brown to dark brown silts. The sediment is fibrous between 254-296 cm.

302.5-357.5 cm: Grey silts containing cm-thick brown organic bands towards the top of the section. A pinky yellow sand layer interrupts the grey silts between 344-346 cm. Visual and geochemical analysis confirms this sand to be the KOT layer (see appendix g).

357.5-366 cm: Light brown silts.

366-373.5 cm: Grey silts.

373.5-384 cm: Brown silts which contain alternating light brown, brown and dark brown bands ranging 3-5 cm thick.

384-400 cm: Grey silt.

400-418 cm: Brown silts with alternating light brown, brown and dark brown silts.

Figure 7.5: Stratigraphy cores analysed for this project with calibrated age and tie point. The black boxes on 0901 and 0903 mark the areas used to produce the continuous core image.

7.4 Chronology

The radiocarbon chronology for the site was produced using bulk sediment and plant macrofossil samples. A few ages are also derived from macrospores of the aquatic macrophyte *Isoetes*. The results of the radiocarbon dating are displayed in Table 7.1. Two to three organic fractions were separated for each sample using different size fractions or material to establish the age. The collection of ages from the two samples was combined to age one horizon depth. In general the ages lie in stratigraphic order. The younger ages from the horizons 242-243 cm (0903 T2 42-42.5, 42.5-43 cm) and 300-302 cm (0903 T3 7-8, 8-9 cm) form tight clusters. However the older ages show greater variability and will be discussed below.

Six AMS ages were collected for horizon 401-403 cm (0901 T4 76.5-77, 78.5-79 cm), the base of the section analysed. A cluster of three ages lie between 30.8-30.6 ka, collectively providing the oldest ages in the core. Three samples from the same horizon, OS-76549, OS-76895, OS-76551, produce younger ages of 28.9 ka, 24.9 ka and 4.7 ka. Likewise, the higher horizon 380-381 cm (0903 T3 86-86.5, 86.5-87 cm) contain three ages clustering between 28.5-29.1 ka and one significantly younger age, sample OS-77227, of 24.9 ka. Both the horizons lie below the KOT layer which has an established age of 25.4 (Vandergoes *et al*, 2011). The younger ages from these deeper horizons therefore appear to be too young and therefore excluded from the age-depth model (Figure 7.6). The age range 28.5-29.1 ka for 380-381 cm appears to be more realistic. The cluster 30.8-30.6 ka at 401-403 cm appear to fit well stratigraphically, however, the younger age, 28.9 ka, is similar to the above dated horizon and therefore appears to lie out of stratigraphic order.

Each of these samples giving anomalously younger ages used sedge or plant fragments from the coarser >90 µm fraction and suggest the plant macrofossils were more prone to contamination by younger carbon. The younger ages from the horizons 242-243 cm and 300-302 cm also have samples which date plant fragments but the ages produced from these samples were similar to those using other material to date. All the samples were processed using the same method. If contamination occurred during the preparation process then it would be expected

that all the samples would produce younger ages. The larger material appears to be more susceptible to contamination in situ, whereas the smaller material, a large proportion of the organic portion is probably made up from pollen, will be less prone to contamination (see Section 3.5). The samples in the deeper horizons produced older ages and therefore even small amounts of contamination can have a large influence on the age. As the larger fraction consistently produced younger ages, it was decided not to use these ages for the age-depth model in Figure 7.6.

Figure 7.6 is the age-depth model developed for Forks Lagoon using simple linear interpolation and is discussed in the following. The ages which have not been included in the models are also shown on the figure, the exception being OS-76863. As discussed above, this sample produce a significantly young age and therefore was not added to the figure. The model includes the revised age of the KOT layer of 25.4 ka (Vandergoes *et al*, 2011). The initial sediment accumulation rate is 1 cm/84 years between 402-380 cm changing to 1 cm/96 years between 380 cm to the tephra layer at 344cm. From 344-301 cm the accumulation rate remains similar to before with a rate of 1 cm /101 years. It then doubles in pace to 1 cm/42 years. There is no visual evidence in the stratigraphy of the site drying out suggesting continuous deposition of sediment occurred throughout the sections analysed.

Sample	Overall depth (cm)	Fraction	¹⁴ C age BP	δ ¹³ C ‰	Reference code	Calibrated age BP (2 sigma)
0903 T2 42-42.5	242-242.5	>90 ⁺	14950 ± 300	-17.46	OS-76936	18126 ± 669
0903 T2 42-42.5	242-242.5	>90 ⁺	15300 ± 70	-15.32	OS-76768	18441 ± 303
0903 T2 42-42.5	242-242.5	90-10 ^{***}	15750 ± 70	-15.87	OS-76861	18981 ± 296
0903 T2 42.5-43	242.5-243	>90 ⁺	15150 ± 100	-18.12	OS-76556	18330 ± 301
0903 T2 42.5-43	242.5-243	>90 ⁺	15450 ± 55	-16.32	OS-76769	18691 ± 141
0903 T2 42.5-43	242.5-243	90-10 ^{***}	15700 ± 75	-15.42	OS-76770	18953 ± 317
0903 T3 7-8	301-302	>90 [*]	17500 ± 100	-16.23	OS-76631	20860 ± 434
0903 T3 7-8	301-302	>90 ^{**}	17800 ± 80	-19.04	OS-76597	21066 ± 469
0903 T3 7-8	301-302	90-10 ^{***}	17800 ± 65	-17.09	OS-76763	21210 ± 322
0903 T3 8-9	302-303	>90 [*]	17800 ± 75	-18.46	OS-76764	21069 ± 464
0903 T3 8-9	302-303	>90 ^{**}	17950 ± 80	-19.8	OS-76765	21403 ± 274
0903 T3 8-9	302-303	90-10 ^{***}	17850 ± 70	-17.27	OS-77120	21283 ± 274
0903 T3 86-86.5	380-380.5	>90 ⁺⁺	23800 ± 110	-13.98	OS-76598	28582 ± 492
0903 T3 86-86.5	380-380.5	90-10 ^{***}	24400 ± 110	-16.42	OS-76766	29099 ± 438
0903 T3 86.5-87	380.5-381	>90 ⁺⁺	20800 ± 390	-15.94	OS-77227	24922 ± 1013
0903 T3 86.5-87	380.5-381	10-90 ^{***}	24100 ± 130	-16.36	OS-76767	28920 ± 432
0901 T4 78.5-79	408.5-409	>90 ⁺	25800 ± 160	-21.18	OS-76553	30636 ± 333
0901 T4 78.5-79	408.5-409	>90 ⁺	4120 ± 40	-16.71	OS-76549	4674 ± 148
0901 T4 78.5-79	408.5-409	90-10 ^{***}	25900 ± 150	-16.55	OS-76860	30690 ± 325
0901 T4 76.5-77	406.5-407	>90 ⁺	24200 ± 230	-25	OS-76895	28986 ± 509
0901 T4 76.5-77	406.5-407	>90 ⁺	20800 ± 120	-19.39	OS-76551	24776 ± 345
0901 T4 76.5-77	406.5-407	90-10 ^{***}	26100 ± 130	-15.72	OS-76863	30818 ± 338

Table 7.1: Radiocarbon age details for Forks Lagoon. Ages were calibrated using the online OxCal program. *Isoetes spores and fine organics, **bulk fine organics, *** organic sediments, +bulk plant fragments, ++ bulk sedge fragments.

Figure 7.6: Age-depth model using calibrated radiocarbon ages for Forks Lagoon. The red dot is the KOT age estimated by Vandergoes *et al* (2011). Sample OS-76549 was excluded from the figure because of its apparent young age.

7.5 Results

7.5.1 Pollen stratigraphy

The pollen results are described in two sections. Firstly the percentage data are described, and includes all taxa counted (See Appendix F for raw data). Secondly, counts from the pollen concentrates for the aquatic and total dry land pollen are presented. The *Isoetes* count in all samples is high and therefore the percentage data masks any important abundance changes in the other aquatic taxa. The total dry land pollen concentrates were also calculated to assess whether these related to core colour changes and/or LOI. The percentage and concentration data are presented in Figures 7.7 and 7.8. For each figure the radiocarbon ages are summarised alongside the diagram. The LOI, particle size and a summary of the core colour changes is also added to both figures. The sedimentology will be described in Section 7.5.2 and 7.5.3. The quantitative temperature reconstructions are also added to Figure 7.7 and are described in Section 7.5.4.

For the pollen analysis a 10 cm sampling strategy spanning 241-401 cm was used. To produce statistically robust data it is general practice to count a minimum of 250 dry land pollen grains for each sample. However, at Forks Lagoon a count reaching 250 grains for any sample was not possible. A maximum 97 grains were counted for one sample whilst the majority of the slides did not exceed 40 grains and one slide only contained 10 dry land pollen grains. A statistically robust dry land pollen count was not achieved and therefore a reliable vegetation and climate reconstruction cannot be achieved. However there is abundant aquatic pollen. It was therefore decided to continue with the dry land pollen analysis but focus on the aquatic pollen for climatic interpretations. The dry land pollen will therefore only produce a broad context of vegetation in the region with the counts being insufficient to go into any further detail.

The sample preparation used on this site was the same method used on the other sites (refer to Section 3.2) and produced very clean slides. The pollen grains that were present were in good condition and tended not to be obscured allowing any visible dry land pollen to be identified relatively unambiguously. However, the over abundance of *Isoetes* slowed the process of counting down considerably. To

speed the process up only one traverse of *Isoetes* was counted. The exotic *Lycopodium* count for one transect was also conducted allowing for the concentrations of *Isoetes* and its percentage for the whole slide to be calculated.

7.5.1.1 FL1: 411-395 cm (c. 31.5-c. 30.1yrs)

The bottom few samples are dominated by herb taxa, in particular Poaceae <40 µm combined with low amount of Asteraceae and other herb taxa. Tree taxa occurred at low abundance and include *Halocarpus*, *Fuscospora*, *N. menziesii*. However, the low pollen counts overemphasise the tree abundance as only 1 grain was counted for each taxon. The wetland taxa are dominated by *Isoetes*.

7.5.1.2 FL2: 395-250 cm (c. 30.1-c. 18.9yrs)

This zone is dominated by herb taxa, with Poaceae <40 µm the dominant taxa throughout the zone. The herb component represents 100% of the TDP at 250 cm. Persistent but low abundances of Apiaceae, Chenopodiaceae *Gentiana*, and Poaceae >40 µm have been derived. A significant but again low shrub component is dominated by Asteraceae and *Hebe*. Only trace amounts of tree taxa are present in the pollen assemblage. *Isoetes* continues to dominate the wetland taxa, an unknown grain, called Type A, possibly aquatic algae, is present in significant abundance, particularly between 370-310 cm.

7.5.1.3 FL3: 250-209 cm (c. 18.9- c. 17.2 yrs)

Herb taxa is still the dominant vegetation component and Poaceae <40 µm the dominant taxon. This zone shows a small increase to low abundance in *Griselinia*, *Myrsine* and *Coprosma* and a small abundance of *Metrosideros*. *Isoetes* remains the dominant aquatic but a significant increase in *Myriophyllum* abundance occurs up the core, peaking in the topmost sample.

Figure 7.7: Relative percentage pollen diagram of Forks Lagoon with the core colour stratigraphy and sediment analysis added to the figure.

7.5.1.4 Pollen concentrations

The pollen concentration data illustrate firstly the low concentrations of TDP present in the core and, secondly, the extremely high abundance of *Isoetes*, which is an order of magnitude greater than any other taxa counted. This is significant, as the large *Isoetes* concentrations mask the changes in other aquatic taxa, in particular *Myriophyllum* which at this site provides an important palaeoclimatical signal which would otherwise become unnoticed in the percentage data. At the base of the core to 369 cm the total dry land pollen abundance consistently increases reaching a peak of 255×10^3 grains/cc at 349 cm. The abundance then decreases down to a low concentration at 319 cm followed by second smaller peak at 290 cm. Again this is followed by very low pollen concentration which increases towards the top of the core and plateauing between $95\text{--}105 \times 10^3$ grains/cc between 229–209 cm.

Although consistently present, the concentrations of *Isoetes* are variable. The bottom 60 cm contains very high concentrations with an overall increasing, stepwise trend peaking with 36×10^5 grains/cc at 351 cm. This is followed by a decrease in concentration to 8×10^5 grains/cc at 321 cm. A small increase to 10×10^5 occurs at 301 cm which is again followed by a decrease to only 2×10^5 grains/cc, the lowest concentration in the record, at 291 cm. In the top 70 cm the *Isoetes* concentration oscillates between high and low concentrations with the low concentrations getting progressively lower up the core with the topmost sample only receiving 4×10^5 grains/cc.

In contrast to *Isoetes*, *Myriophyllum* has a much lower concentration and is more sporadic in appearance. *Myriophyllum* is absent from the record until 381 cm where it has very low concentrations until 350 cm where it rises to a peak of 15×10^3 grains/cc at 341 cm and then decreases back to trace amounts at 321 cm. Another small peak at 301 cm occurs with *Myriophyllum* again disappearing from the record. At 251 cm a rapid increase in concentrations occur peaking with 55×10^3 grains/cc at the top of the core.

Figure 7.8: Pollen concentrations of the two aquatic taxa and total dry land pollen. Total carbon content and particle size is added for comparison between the results.

7.5.2 Particle Size Analysis (PSA)

Sampling for PSA occurred at a 2 cm resolution and a sample width of 0.5 cm thick. Sampling began at 209 cm and continued down through the core to 411 cm covering the section of pollen analysis.

Throughout this section the silt fraction was the most dominant particle size, mostly ranging between 80-90%. The percentage of clay also remained constant, ranging 1-2%. However, the sand abundance and consequently changes in the % of silt abundance varies up the core ranging from 10% to nearly 20%.

7.5.3 LOI

As with the PSA, sampling of LOI analysis occurred every 2 cm from 209.5 to 411.5 cm. The pattern of carbon content is variable and fluctuates throughout the section analysed, and remains consistently below 30%.

The base of the core, the bottom 38 cm, contains two peaks in carbon content at 403.5-411.5 cm and 375.5-385.5 cm, ranging 10-14% and 7-17.5% respectively. Much lower carbon content occurs between the two peaks, ranging only 2.8-6%. After 375.5 cm the carbon content again declines to a range of 2.8-6% and remains stable at these low percentages up through the core until 246.5 cm, after which a more variable carbon content ranging between 4-30% is evident. However, the majority of samples in this top 40 cm contain >10% carbon and the general trend is a continued increase in LOI. The top 15 cm receive the highest carbon abundance pattern.

Figure 7.9 compares the location for the organic and inorganic bands in the core with the total carbon content. A consistent pattern is evident where maximum % carbon occurs with brown or dark brown bands. Conversely, the sections of the core with minimum carbon abundances occur when the core bands are grey. This confirms that the addition of organic material is the cause of the brown sediment colour.

Figure 7.9: A simplified core description showing the location of the colour bands and their relationship with the total % organic carbon.

7.5.4 Temperature reconstruction

Figure 7.7 contains the predicted MAT reconstruction from PLS-C3 and W-MAT. The PLS-C3 model reconstructs an initial increase in MAT from 4.4 °C at 31.5 ka to 7.4 °C by 29.8 ka. Throughout the rest of the core MAT remains between 6.4-7.4 °C and averages 6.8 °C throughout the section analysed. The W-MAT reconstruction shows a small MAT range from 7.1-7.7 °C recorded for the majority of the core except at 17.2 ka where a maximum MAT of 8.3 °C is shown. The average MAT predicted by the W-MAT model is 7.5 °C.

7.6 Discussion

This section interprets the pollen and sedimentological results described in the previous section. As aforementioned the dry land pollen can only produce a limited vegetation reconstruction which is discussed first. The aquatic macrophyte and sediment analysis will follow and will be discussed in more detail.

The vegetation in the Mackenzie Basin has been heavily modified by human activity. Today, grassland is the dominant vegetation type in the region with *Chionochloa* (tall tussocks) sparse but consistently present in the region. Naturally these tall tussocks are restricted to the alpine zone, however in the Mackenzie basin their altitudinal extent has lowered, due to the lack of forest in the lower regions, (Molloy *et al*, 1963; McGlone and Moar, 1998; McGlone, 2001b; Espie and Barrett, 2006). Therefore contemporary vegetation composition and species relationship with various environmental variables, although insightful, may be misleading for interpreting native palaeo-vegetation communities. To produce a broad regional vegetation context a combination of modern vegetation studies and palaeo-records from the central Canterbury region, and neighbouring areas, will be used to produce a tentative vegetation reconstruction for Forks Lagoon during the LGCP.

7.6.1 Vegetation reconstruction

Throughout the 31.4-17.2 ka period the dry land pollen count is exceptionally low indicating a sparsely vegetated region. The pollen that is present is dominated by small tussock and grass taxa, coexisting with tall tussock and other herb taxa. This vegetation assemblage is indicative of alpine grassland/herbfield community. The modern alpine vegetation community in the Mount Cook National Park is dominated by *Chionochloa oreophila*, *C. pallens*, *C. crassiuscula*, *Rytidosperma setifolium* and *Poa colensoi* (Wardle, 1991). It is likely that these species were prevalent in the region during the LGCP. The shrub taxa Asteraceae, *Coprosma* and *Hebe* all have low abundances in the pollen record. These taxa also grow in herb forms which are more prevalent in alpine regions. The dominance of grass and herb taxa would suggest that these 'shrub' taxa represent herb forms.

The top section analysed, 18.9-17.2 ka, contains an increase in *Myrsine*, a sub-alpine shrub taxon along with an increase in *Coprosma*, *Griselinia* and a small abundance of *Meterosideros*. McGlone and Moar (1998) recorded a vegetation composition from grassland dominated vegetation and bare ground to grass-*Coprosma-Myrsine* shrubland marking the end of the LGM. This may suggest that the Forks Lagoon record is experiencing climate amelioration at this time. However, the low counts cannot allow for any firm conclusions on climate change at this time.

Overall the DLP is indicative of an open, sparsely vegetated landscape. Any vegetation that is present represents cold-tolerant alpine community vegetation. Based upon this limited information the regional climate was harsh and cold, although this is a tentative conclusion. The modelled temperature reconstructions appear to underestimate the MAT depression with both models estimating temperatures only 1-2 °C cooler than at present. These modelled results will be discussed further in Section 8.2.4.

7.6.2 Aquatic macrophyte and sediment records

Isoetes spores are consistently present in the record with high concentrations indicating that the site was a lake environment throughout the LGCP. There are two species of *Isoetes* in New Zealand Lakes (Mason, 1975). These are *Isoetes alpinus* common in montane lakes and *I. kirkii*, which is more prevalent in lakes at lower altitudes (Brown, 1975). Due to the elevation of the site, it is assumed that the *Isoetes* spores are *I. alpinus*. This species can live in a water depth ranging from only 0.1 to a maximum of 4.5 m and cannot tolerate lake environments where silting is occurring (Brown, 1975). Therefore, Forks Lagoon sustained a water body, albeit shallow, between c. 31.4 - c. 17.2 ka.

Myriophyllum is also present in the record although its appearance is comparatively more sporadic and is not continuously present in the record. New Zealand contains five native species of *Myriophyllum*, but only two which can grow in sub-alpine lakes and tarns; *M. pendunculatum* and *M. tryphyllum*. In the modern environment *M. pendunculatum* is observed to prefer boggy/peaty or

stony substrates along lake margins (Mason, 1975). The pollen grains from this plant are described by Moar (1993) as being tetra-hexaporate, but dominantly pentaporate (76%) and sometimes have a verrucate pattern on the exine surface. *M. tryphyllum* is observed to grow in mid-depth lake zone on sand-sandy silt substrate (Mason, 1975). The pollen morphology ranges triporate-hexoporate with 64% of grains observed being tetraporate (Moar, 1993). The *Myriophyllum* grains observed at Forks Lagoon are dominantly tetraporate, with some triporate grains. The exine surface contains no verrucate pattern. The substrate of the core based upon the particle size analysis is a sandy silt. Therefore it is assumed that the *Myriophyllum* pollen at Forks Lagoon is *M. tryphyllum* a mid-water depth species.

A modern aquatic plant survey was produced for Lake Alexandrina, located in the Mackenzie Basin c. 2.5 km east of Forks Lagoon providing a modern analogue to apply on the macrophyte record. In this lake *M. tryphyllum* is restricted to the water depths 2-6 m located in the mid-water depth zone. *I. alpinus* was restricted to 0.7-4 m water depth and recognised as a shallow water species (Ward and Talbot, 1984). Other records also show a similar categorisation of *I. alpinus* and *M. tryphyllum* suggesting this water depth pattern is consistent in New Zealand (Brown, 1975). It can therefore be assumed that when *M. tryphyllum* is present in the macrophyte record, lake levels are deeper compared to periods without *M. tryphyllum*.

Forks Lagoon is a rain fed lake, with precipitation being the only input of water to the lake. As discussed earlier, consistent and high abundance of *Isoetes* indicates that a lake was continuously present at the site throughout the LGCP. The presence of *M. tryphyllum* is indicative of deeper water levels and therefore potentially periods of increased precipitation. *M. tryphyllum* occurs in the core between 28.7-23 ka although the core only contains significantly high concentrations between 26-24 ka, at 21 ka and 18.9-17.2 ka with the maximum abundance and concentration occurring at the top of the core. This suggests an increase in precipitation in east South Island occurred during the NZ-I CES assigned LGCP milder interval and at the end of the LGM. This will be discussed further in Section 7.6.3.

The total percent carbon in the record remains low throughout the section analysed, and particularly during the earlier part of the LGCP (Figure 7.9). This confirms a low vegetation biomass in and around the lake as suggested by the TDP record. However, there are distinct fluctuations in LOI which relate closely to core colour changes where increased carbon relates to brown silts and low carbon corresponds with grey silts. The cause for the carbon content changes is not clear. However, two periods of increase *M. tryphyllum* at 21 ka and between 18.9-17.2 ka corresponds with periods of increased carbon content and suggesting a possible relationship between the two proxies. The increase in *M. tryphyllum* concentrations may represent an increase in plant abundance and therefore an increased input of organic material. The maximum carbon content coincides with the increase in *M. tryphyllum* between 18.9-17.2 ka and also coincides with evidence of shrub increase, possibly confirming climate amelioration had started by 18.9 ka. However, the pattern of increased organic content and appearance of *M. tryphyllum* is not consistent, suggesting other actors are influencing carbon content at the site.

The addition of inorganic material to the site is most probably aeolian derived. The particle size analysis suggests the composition of particles remained relatively unchanged, however, the amount of material reaching the site over certain periods may have changed. Suggate and Almond (2005) research on loess deposition in the Westland region recognised periods of increased organic content coinciding with glacial retreat. One explanation for this is a reduction in source material and therefore a relative decrease in inorganic sediment and increase in the proportion of organic material. The periods of increased organic content in the Forks Lagoon record may represent periods of reduced wind power to transport the material to the site. This suggests that prior to the start of the LGCP and at the end of the LGM the area was less windy, or there was less source material than during the LGCP, including the milder interval. However, if there is less inorganic material entering the site we would also expect a decrease in sediment accumulation rate which, based upon the available chronology does not seem to occur (see Figure 7.6 and Section 7.4). The cause of the changing carbon is therefore not clearly understood and requires more investigation.

7.6.3 Summary discussion

Core depth interval	Vegetation interpretation	Climate interpretation
FL3 250-209 cm (c. 18.9-17.2 ka)	Alpine grass and herb taxa dominate the landscape. Possible expansion of sub-alpine/ montane shrubs into the lowland areas. A lake was continually present.	Very cold but possibly beginning to warm. Increase in precipitation which possibly continues to increase up the section.
FL2 395-250 cm (c. 30.1-c. 18.9 ka)	Alpine grass and herb taxa dominate the landscape. Vegetation is sparse in the environment. A lake was continually present.	Very cold and harsh environment. Precipitation fluctuated throughout the period. It is possibly increasing between 26-24 ka and at 21 ka and drier between these periods.
FL1 411-395 cm (c. 31.4- c. 30.1 ka)	Alpine grass and herb taxa dominate the landscape. Very sparse and open vegetation. A lake was continually present.	Very cold and harsh environment with relatively low precipitation.

Table 7.2: Summary table of the vegetation and climate reconstructions for Forks Lagoon.

For the period spanning 31.4-17.2 ka the vegetation remained dominated by alpine herb/grassland with very little, if any, shrub and tree taxa present in the local and regional landscape, indicating a, very cold harsh environments. Drost *et al* (2007) modelled and simulated climates during the LGM in New Zealand. The results indicate that ESI, including inland Canterbury, had a MAT 4 °C colder than at present and increased seasonality in temperature, with most cooling during the winter months. The model also predicts an increase in the number and intensity of southerlies, which bring cold polar air into the region, adding to the already harsh conditions that are expected to have been endured. This modelled climate scenario of eastern South Island is supported by the sparse vegetation cover and evidence of hardy cold-tolerant taxa in the pollen record. The limited evidence suggests that the cold conditions started before the LGCP as defined by the NZ-I

CES, indicating cold conditions were experienced at the site 3-4 ka earlier than the CES.

7.6.3.1 Aquatic plant record

Macrophytes have consistently been used as an open water body indicator and a lake-level change indicator in palaeorecords from around the world (e.g. Grosjean *et al*, 2001; Vandergoes *et al*, 2005; Newnham *et al*, 2007a). However, unlike these other sites, Forks Lagoon is unique as it is possible to interpret the changes in macrophyte abundance from the site as a precipitation signal. This is possible at Forks Lagoon for two main reasons. Firstly *Isoetes* spores are always present with a high abundance in the core, suggesting a shallow lake was consistently present at the site throughout the period analysed. Therefore, when *M. tryphyllum* is present in the core the only interpretation is that lake level increased and when it is not present then lake level has decreased. Secondly, unlike the majority of the lake and wetland palaeorecord sites, Forks Lagoon is solely rain fed and appears to be a sensitive site to precipitation changes due to its small basin size and low rainfall amount in this area of New Zealand. Therefore, the macrophyte record at Forks Lagoon can be assumed to represent lake level changes caused primarily by precipitation changes.

The onset of the LGCP, pre 26 ka, is characterised by drier conditions. Between 26-24 ka and at 21 ka, the period spanning the LGCP milder interval is wetter. The wetter periods are interrupted by a period of drier climate spanning 24-21 ka. The third period of increased precipitation, between 18.9-17.2 ka, corresponds with a large increase in % carbon content and an increase in *Myrsine*, indicating possible climate amelioration and wetter conditions coinciding with the end of the LGM and Termination I. The NZ-I CES ages Termination I between 19-18 ka and again corresponds closely with the period of wetter climate at the site.

When the timing of *M. tryphyllum* appearance is compared to other records from across New Zealand, the following pattern is observed (see Figure 8.5). Suggate and Almond (2005) reported moraine and loess deposition representing periods of ice advance in northern Westland region that coincide with periods of low or absent *M. tryphyllum* and inferred drier conditions. Similarly, peak grass

abundance in the Okarito grass curve record by Newnham *et al* (2007a) and peaks in aeolian quartz dust (Alloway *et al*, 1992) both indicative of colder climates, correspond with periods without *M. tryphyllum*. Conversely, periods of reduced grass abundance (Newnham *et al*, 2007a), reduced aeolian quartz accumulation (Alloway *et al*, 1992) and periods where moraines and loess deposition is not evident in the landscape (Suggate and Almond, 2005), all of which suggest milder climates, coincide with periods of increased precipitation in Forks Lagoon and the surrounding area. The macrophyte record therefore suggests drier conditions in the region prevailed during periods of ice advance whilst the regional climate became wetter during ice retreat in New Zealand.

However, it is important to note that the age of the KOT layer used in the Forks Lagoon record is younger than that used by Alloway *et al* (1992), Suggate and Almond (2005) and Newnham *et al* (2007a). Therefore, the records need to be stratigraphically, not chronologically, compared during the deposition of the KOT layer. Despite this age difference, the stratigraphic comparisons produce the same conclusions as described above. Grass abundance decreases in the Okarito record (Newnham *et al*, 2007a) and *M. tryphyllum* concentrations increase at Forks Lagoon, just prior to, and continuing after, the KOT layer and both coinciding with the LGCP milder interval. This again supports a wetter climate at Forks Lagoon coinciding with warmer conditions and reduced glacial activity.

7.6.3.2 Low pollen counts and lack of tree/shrub pollen

Throughout the Forks Lagoon sequence the TDP abundance and concentration is extremely low. The abundance and preservation of the aquatic pollen taxa show preservation is not a primary factor in low pollen counts. The land pollen which is present in the record is primarily grass and other herb species, indicating the vegetation cover throughout the analysed section was alpine/grassland communities.

Modern pollen rain studies from alpine areas show the pollen spectra to be dominated by pollen derived from forest located downslope. In these situations alpine grasslands tend to have relatively low representation in the pollen rain (Pocknall, 1982; Moar, 1980; Horrocks and Odgen, 1994). McGlone (2001b)

attributes this partially to the relatively small pollen source area of the upland regions compared to the surrounding lowlands. However, during the LGCP the upland vegetation would have significantly expanded, increasing the pollen source area. Therefore the pollen source area cannot be the sole factor for the low pollen counts at this site. McGlone (2001b) also notes that there is lower pollen production in the cool sub-alpine/alpine climates. If the vegetation community is dominated by alpine herbs and grasses, the pollen production will possibly be low and with little pollen being produced the amount deposited would also be low, equating to smaller counts. The prevailing wind direction today, and likely to have been throughout the LGCP, is north-westerly. Wind is a large transporter of pollen to a site but north-west of the site lies the Southern Alps mountain ranges. Therefore wind would have firstly come from the glaciated, vegetation free mountain ranges and then over sparsely vegetated land in the periglacial environments. Under these circumstances only sparse local and extra-local vegetation would grow in this harsh environment which perhaps again help explain the low volumes of dry land pollen in the sequence.

To summarise, a number of factors including low pollen production, relatively small pollen catchment and the prevailing wind blowing any available pollen away from the site would have contributed to the low pollen counts in the samples. The volume and composition of dry land pollen at Forks Lagoon is consistent with cold dry conditions in the Mackenzie Basin throughout the LGCP.

7.7 Conclusion

Forks Lagoon is a unique site in the region as it is the only one of a few records that extends the LGCP in the interior South Island area. Unfortunately the TDP counts were low throughout the section analysed and a statistically robust count was not achieved. However, this in itself provides some insight into the palaeoenvironment at the site between c. 31.4-17.2 ka. There is a significantly larger abundance of macrophytes, in particular *Isoetes* and *M. tryphyllum* which, due to the site's morphology and hydrology, can be used as a precipitation signal. However, the conclusions made for this chapter are tentative and Section 9.2 will suggest the future work needed to develop the record.

The TDP throughout the record is sparse and dominated by alpine grassland suggesting very cold conditions were felt at the site from c. 31.4-17.2 ka. This also suggest that the LGCP began at least 4000 yrs prior to the NZ-I CES timing of this event and suggest an adjustment for the timing of this event may be needed. The core base from 31.4-26 ka is a period of drier climate. The LGCP 'milder' interval can be separated into two periods of wetter climates spanning 26-24 ka and at 21 ka, which are interrupted by a period of drier climate. The timing of this drier period corresponds with evidence from other records around New Zealand of glacial advance and a period of cooling. The period 21-18.9 ka is also characterised by drier conditions in the region which is followed by wetter conditions again from 18.9-17.2 ka. This late wet period is combined with an increase in organic carbon content suggesting increase vegetation biomass and the appearance and increase in the shrub *Myrsine* and other sub-alpine/montane taxa. This possibly indicates the onset of Termination 1 and the end of the LGM in the region.

Chapter 8: Discussion

The main aim of this project is to produce four new vegetation and inferred climate reconstructions from southern New Zealand which can be used to test the NZ-I CES and in particular the periods of uncertainty highlighted in the introduction, Section 1.2.2. In this chapter, the results and conclusions for each site are synthesised. The vegetation reconstructions used to compare and test the NZ-I CES are discussed using a time slice approach. Regional differences between sites are also discussed. The significance of the results in the context of comparable records from Antarctica and elsewhere in the mid-latitude Southern Hemisphere follows. Consideration is given to potential drivers of climate change and the changing climatic and oceanic regimes in the Southern Hemisphere. The final section considers the age of the KOT, reviews the NZ-I CES, discusses the wider implications of climate during the LGCP and the quantitative pollen-climate reconstructions. The discussion is followed by the conclusion chapter which will summarise these findings.

8.1 Time slices

Since the initial event stratigraphy was published the NZ-I group have subsequently met and discussed ways to develop and move the project forward. One outcome is to concentrate on time slices which are predefined in the preliminary NZ-I CES (Almond *et al*, 2010). It therefore seems suitable for the discussion chapter to follow a chronological time-slice approach. Unlike the NZ-I CES 30-8 ka time period, the oldest record here extends to c. 38 ka and therefore the time slices start at this age. Figure 8.1 provides a summary of the four reconstructions compared to the preliminary NZ-I CES. It also shows the time slices which will be discussed. Subsequent sections refer back to this figure.

Figure 8.1: An overview of the timing of climate change recorded in the four records of this project and compared with the preliminary NZ-I CES

8.1.1 Late MIS 3 and an early LGCP onset

The two oldest records, Howard Valley and Forks Lagoon, extend beyond the LGCP into the milder, although still cooler than contemporary climate, of MIS 3. The vegetation at Howard Valley between 37.7-34.4 ka is characteristic of a mixed montane *Nothofagus menziesii* and *Fuscospora* forest. The estimated treeline lowering from contemporary levels was 400-600 m, implied a MAT depression ranging 2-3 °C. This corresponds with other records from New Zealand which estimate temperatures to have been 3 °C cooler than today (McGlone and Topping, 1983; Marra *et al*, 2009). The lowermost part of the Howard Valley sequence therefore implies onset of the LGCP was after 34.4 ka.

The Forks Lagoon record extends back to 31.4 ka, roughly 3000 years earlier than the LGCP as defined by the NZ-I CES (Alloway *et al*, 2007). The low dry land pollen concentrations constrain vegetation interpretations but, in combination with the low total % carbon, suggest an open and sparsely vegetated region, and very harsh conditions. The macrophyte record suggests the period 31.4-26 ka was relatively dry compared to other periods during the LGCP. These dry conditions combined with the low pollen counts and herb dominated taxa suggest cold and harsh conditions were already established in ESI considerably earlier than the onset of the LGCP (Alloway *et al*, 2007). Sutherland *et al* (2007) present exposure ages from lateral moraines in south Westland suggesting a period of glacial advance occurred between 31.9-29.5 ka. Suggate and Almond's (2005) research on moraine and loess deposits from central Westland suggests initial cooling began ~34 ka, culminating by 28 ka, again supporting an earlier start to the cold period.

The Howard Valley record provides an incomplete reconstruction with a gap between 34.4-28.5 ka. Over this period vegetation changed from mixed montane beech forest to an alpine grassland/sub-alpine treeline vegetation community. A further treeline reduction of 600-900 m represents MAT depression of 3.6-5.2 °C which again suggests conditions similar to those expected during the LGCP. The onset of the LGCP in the NZ-I CES is characterised by a transition from relatively milder conditions to cold e.g. increasing herb abundance at Okarito (Vandergoes *et al*, 2005; Newnham *et al*, 2007a) and a decrease in lowland podocarp relative to

grass pollen at Pukaki (Alloway *et al*, 2007). The onset of the LGCP in the evidence above is characteristic of a transition from mild to colder conditions. The Howard Valley and Forks Lagoon records suggest cold and dry conditions were already established by 31.4-28.5 ka. This data is consistent with moraine evidence which suggests an earlier onset for the LGCP (e.g. Almond *et al*, 2001; Suggate and Almond, 2005; Sutherland *et al*, 2007). This indicates a need for the NZ-I to extend the temporal period included in the NZ-I project remit.

The Northern Hemispheric LGM period of extreme cold, spans 24-18 ka (Mix *et al*, 2001). Abundant evidence from the Southern Hemisphere points towards an extended cold period, referred to as the LGCP in this project. The LGCP includes the LGM which is the period of most extreme cold conditions and spans 21-18 ka based on the NZ-I CES. For the remainder of this discussion the LGM will refer to the New Zealand period spanning 21-18 ka unless otherwise specified. As mentioned above the NZ-I CES places the onset of the LGCP between 28-29 ka and other evidence around New Zealand, included in this project points towards an even earlier onset of cold conditions. Exposure ages from moraines in southern Chile suggest initial glacial advance began between 35-32 ka (Denton *et al*, 1999; Fairbanks *et al* 2005) whilst moraines with exposure ages of 31 ka are found in Tasmania (Fink *et al*, 2000). The global sea-level curve also shows a rapid decrease in sea level between 33-30 ka (Lambeck and Chappell, 2001; Lambeck *et al*, 2002) coinciding with evidence of colder climates in the terrestrial records including periods of early ice sheet maximum between 33-26.5 (Clark *et al*, 2009).

8.1.2 LGCP

8.1.2.1 LGCP milder interval

This section concentrates on the LGCP mild interval which interrupts the period of cold between 27-21 ka as defined by the NZ-I CES. The onset of this milder interval was not captured in the total dry land pollen records, however, the evidence does point towards relatively less severe conditions in all three reconstructions compared to the LGM (Figure 8.1). The Howard Valley fragmentary records suggest a similar vegetation composition was present at the site between 28.5-21.9 ka and contained a higher abundance of tree taxa than

present in the record between 20.2-18.9 ka. The Lake Mudgie vegetation reconstruction contains a relatively high shrub component compared to the samples post 23.2 ka and suggest milder conditions at the site compared to those experienced after 23.2 ka at Lake Mudgie and 20.2 ka at Howard Valley, although conditions were still very cold. The Forks Lagoon aquatic macrophyte record supports a postulated increase in precipitation suggesting conditions were relatively less severe between 26-24 ka and at 21 ka than the drier conditions between 21-18.9 ka. The macrophyte record will be discussed in more detail below (and refer to Section 7.6). These sites suggest that the LGCP milder interval occurred sometime between ~28.5-26 ka to 21.9-21 ka.

The milder interval in the NZ-I CES spans 27-21 ka and is broadly consistent with the evidence presented in this project. However, the macrophyte record at Forks Lagoon shows a climatically complex milder interval. Two periods of wetter conditions occur between 26-24 ka and at 21 ka which are interrupted by a period of reduced precipitation. Evidence from loess and moraine deposits from the Westland region indicates a period of reduced glacial activity between 28-24.5 ka and 21.5-20.5 ka (Suggate and Almond, 2005). Speleothem records from north Westland region also show two $\delta^{18}\text{O}$ excursions inferred to represent warmer climates during the LGCP (Hellstrom *et al*, 1998;). Likewise, the Okarito pollen record shows the LGCP milder interval spanning 27-21 ka but interrupted by a period of strong cooling between 25-23 ka (Newnham *et al*, 2007a).

The sites investigated here are located in different geographical regions of the South Island (Figure 1.2) and the pollen records from each contain different vegetation composition. The Forks Lagoon sequence (ESI) contains extremely low pollen concentrations and low carbon content, indicating a sparsely vegetated region and harsh conditions persistent through the LGCP milder interval. At Lake Mudgie, average herb abundance was 65%, shrub abundance 30% and tree abundance was only 5% for the LGCP milder interval. There is a significantly larger abundance of shrubs between 26.6-23.2 ka than 21.3-18.4 ka, which is consistent with milder climates at the site during the LGCP milder interval. Finally Howard Valley contains the highest tree abundance averaging 13%, with shrub abundance only 21% and herb abundance is 65%. The 13% tree abundance

indicates forest existed within close proximity to the site between 28.5-21.9 ka suggesting this period was milder than between 20.5-18.9 ka, the LGM. The higher abundance of trees and shrubs in Howard Valley indicates that climatic conditions were more favourable at this site than at the others. This will be expanded upon in Section 8.1.3.2.

8.1.2.2. LGM: 21 to 18.9 ka

The timing of the LGM in New Zealand, according to the NZ-I CES, occurs between 21-18 ka, although some records used in the CES suggests it extends back to 24 ka. The Lake Mudgie record shows a large decline in shrub abundance beginning at 23.2 ka which would correspond with an earlier LGM. The Forks Lagoon macrophyte record suggests drier conditions also began at ~24 ka to 18.9 ka but were interrupted by wetter conditions at 21 ka coinciding with the timing of milder climates in records elsewhere in New Zealand (Suggate and Almond, 2005; Whittaker *et al*, 2011) and may reflect a minor reversal in cooling (this is discussed further in Section 8.2.2). The Howard Valley vegetation record suggests that at 21.9 ka the climate in the region was milder than during the deposition of section 1, aged 20.5-18.9 ka. Section 1 contains significantly lower tree abundance and higher herb abundance than recorded at 21.9 ka and suggests the period of extreme cold in Howard Valley began after 21.9 ka. Therefore the three records combined suggest a climate transition from milder and wetter to colder and drier climates began ~23.2 ka, but was not the period of greatest cooling. The trend in cooling is instead interrupted at 21 ka which is then followed by a period of prolonged dry and cold climates. Evidence from the records in this project indicates the most prolonged period of harshest conditions occurred between 21-18.9 ka (Figure 8.1).

The LGM is characterised at all sites by an increased dominance of herb taxa, in particular Poaceae. Lake Mudgie contains 83% herb abundance. Howard Valley herb abundance is ~10% lower, but also 10% higher than during the LGCP milder interval. Howard Valley experiences a significant decrease in average tree abundance, reducing from 13% to 5%, suggesting forest had migrated further away from the site during the LGM. Only trace amounts of tree pollen was recorded in Lake Mudgie which contained a similar shrub component as Howard

Valley. All three records suggest very harsh, cold and dry conditions whilst conditions at Howard Valley, which contained comparatively more trees and shrubs and less herbs, were not as harsh as experienced in the other sites.

The estimated MAT depressions, based upon pollen assemblages are 3.6-5.2 °C at Howard Valley and >5.3 °C at Lake Mudgie. Therefore, Lake Mudgie has significantly lower minimum temperatures and the largest altitudinal treeline depression of at least 1050 m, compared to 600-900 m lowering at Howard Valley. However, it should be noted that the exceedingly low pollen concentrations and lack of woody vegetation at Forks Lagoon suggest the region also experienced very harsh conditions. Drost *et al* (2007) modelled a maximum MAT depression of 2.5-4 °C whilst Porter (1975) and Soons (1979) estimated an ELA lowering during the LGM of 800-850 m, which is the equivalent of a 4.5 °C lowering of MAT assuming an ELR of 0.5 °C /100m. Ocean core records from west of the South Island estimated MAT lowering of 4-5 °C during the LGM (Pelejero *et al*, 2006; Barrows *et al*, 2007). Therefore both the proxy and simulated evidence points towards a 4-5 °C cooling during the LGM and complements the estimated temperature and treeline depressions from Howard Valley and Forks Lagoon, but not Lake Mudgie. This will be discussed further below.

McGlone and Bathgate (1983) estimated a 5 °C cooling during the LGM which would lead to an 800 m altitudinal lowering of vegetation zones. Based on temperature alone, montane forest would have been able to grow up to 300-400 m above present sea-level. It is therefore not unexpected that Forks Lagoon, located at 873 m.a.s.l, contains low dry land pollen counts indicative of an open and sparsely vegetated region, suggesting harsh conditions were experienced throughout the LGM. However, Lake Mudgie is only 140 m.a.s.l, 250 m below the estimated maximum elevation for montane forest growth yet the site contains a pollen assemblage resembling alpine herb and grassland. This suggests that reduced MAT was not the sole cause of tree decline in southern New Zealand. Also, the Howard Valley altitudinal location is 300 m above Lake Mudgie but contains relatively higher amounts of tree and shrub abundance during the LGM, which indicates a potential difference in regional vegetation and inferred climate.

Drost *et al* (2007) modelled LGM climate and suggest cooling was greatest in ESI with temperatures reducing by a maximum 4 °C. The modelled results predict conditions to have been coldest in ESI, however Lake Mudgie's inferred temperature depression suggest Westland experienced colder conditions than predicted. Drost *et al* (2007) concede that even the maximum amount of cooling predicted would not be sufficient to cause the deforestation evident in the vegetation records. The model output also suggests there was an increase in seasonality, extreme events and the occurrence and intensity of southerly wind flow over the whole of New Zealand, although particularly ESI. This airflow is capable of bringing very cold polar air to New Zealand leading to severe cold conditions. The July 1996 Southland frost can be used as a modern-analogue for this scenario. Temperatures during this event remained below freezing for over a week and reached a minimum -15°C. This led to a widespread mortality of indigenous vegetation (Banister, 2003). During the LGM such conditions would have been more frequent and for a longer duration which could explain the lack of tree pollen in the records and regional forest, particularly in inland ESI and exposed areas.

McGlone *et al* (2010) produced a compilation of LGM pollen records and found, based upon the average vegetation compositions, that New Zealand could be split into four vegetation regions: north North Island (NNI), south North Island (SNI, including north South Island), west South Island (WSI) and east South Island ESI. The tree pollen shows a north to south and west to east decline. Sites east of the Southern Alps contain the lowest tree/shrub abundance ranging 0%-4.9%. Conversely, grass abundance is higher in the South Island averaging 88% in WSI and 98% in ESI (McGlone *et al*, 2010). A similar regional variation can be seen in the three records of this project with Howard Valley, the most northern site, containing the highest tree and shrub abundance whilst Lake Mudgie (WSI) and Forks Lagoon (ESI) contain the lower tree/shrub abundances. McGlone *et al* (2010) suggest the main cause for the paucity of forest in ESI was an increase in southerly winds causing an increased occurrence of severe frost (McGlone *et al*, 1993; Drost *et al*, 2007). Harle *et al* (2004) noted that when tree dormancy is broken around late spring their tolerance to the cold is lost. A period of freezing at this time would be very effective at impeding tree growth. Cold tolerant plants grew

from current sea-level during the LGM, providing evidence to support the occurrence of frequent cold events (McGlone *et al*, 2010). However, both Drost *et al* (2007) and McGlone *et al* (2010) suggest areas sheltered from these cold southerlies would still be suitable for forest growth. Howard Valley (the most northern site) and the Nelson Lakes region may be an example where pollen-vegetation composition reflects relatively less harsh conditions during the LGM. Forks Lagoon, located in ESI is an exposed site located in a region which is predicted to have experienced the most severe conditions during the LGM (Drost *et al*, 2007; Soons, 1976) and would explain the vegetation composition and very low pollen concentrations at the site. Lake Mudgie would also have been susceptible to these hyper-cold conditions, resulting in the grass dominated pollen assemblage at this lowland site.

It is also important to comment on the method used to estimate MAT depressions at the site. The national mean annual ELR of 0.5 °C/100 m was applied to the estimated vegetation zone depressions to estimate MAT depression. However, as mentioned in Section 3.2.6, ELR's are variable on both spatial and temporal scales. Howard Valley is located further inland, at a higher altitude and experiences lower mean annual precipitation than Lake Mudgie. Research by Meyer (1992) suggests that in drier areas the ELR is generally higher than in wetter regions. Therefore it would be expected that the ELR used for Howard Valley may be lower than one used on Lake Mudgie. This may mean the ELR used at both sites may be overestimating or underestimating MAT depressions. If possible, regional ELR would be useful but there are insufficient observational data and research on ELR. The limitations with the ELR methodology may also have some impact on the different MAT depression estimates for each site.

Estimates of precipitation from the dry land pollen records are difficult to assess as it is difficult to determine a direct relationship between vegetation assemblage and precipitation alone (Wilmshurst *et al*, 2007). Temperature depressions during the LGM would have had a significant influence on vegetation type at the time. However, the lack of high precipitation taxa, such as tree ferns, drought intolerant tree taxa, and a low shrub component could possibly also imply a decrease in precipitation. The Forks Lagoon macrophyte record also implies a decreased

precipitation in the Mackenzie Basin and ESI. Therefore, all three sites point towards drier conditions in the South Island during the LGM.

8.1.2.3 Global context of LGCP

Exposure ages from moraines in Tasmania date two periods of ice advance, the earliest at 31 ka and the second, marking the onset of the LGM, at 24 ka (Fink *et al*, 2000). Vegetation reconstructions show a landscape dominated by herb and grass taxa implying a MAT 4 °C colder than today and ranging 3.7-4.2 °C (Fletcher and Thomas, 2010). Conditions were also drier than contemporary environments, with the period of main ice advance suggested to be restricted to 20-19 ka (MacPhail, 1979). Tasmania therefore appears to show similar timings of glacial advance, vegetation composition and inferred climate as New Zealand during the LGCP. Exposure ages from the Chilean Lake District (CLD) dated glacial advance beginning late MIS 3 at ~35 and 32 ka (Fairbanks *et al*, 2005), whilst Denton *et al* (1999) concluded that full glacial conditions persisted in northern CLD between 29.4-14.55 ¹⁴C ka. Mean summer temperatures were reduced by 6-8 °C and equilibrium line altitudes (ELA) lowered by 1000 m. In Laguna del Maule region snowline depression is evident after 30 ka whilst glacial advance and retreat occurred between 25.6-23.3 ka (Singer *et al*, 2000). In Lago Buenos Aires, Kaplan *et al* (2004) and Douglass *et al* (2006) record two ice advances at 23 and 16 ka with the period of maximum ice advance at 23 ka and ELA lowering by 900m. Also in the strait of Magellan LGM moraines were deposited between 25-17 ka (McCulloch *et al*, 2005; Kaplan *et al*, 2008). Kaplan *et al* (2008) concluded that a cold period, similar to the NZ-I LGCP, occurred between c. 30-16 ka in southern Patagonia, although some ages suggest ice advance occurred earlier, and a period of greatest cooling was sometime between 25-17 ka.

To summarise, in New Zealand the LGCP persisted from at least 31.4 ka -18.9 ka based upon the records used in this project. As discussed in Section 8.1.2 the onset of the cold period appears to be earlier than suggested by the NZ-I CES and is supported by a number of records both in New Zealand and elsewhere in the Southern Hemisphere (e.g. Heusser, 1981; Denton *et al*, 1999; Fink *et al*, 2000; Fairbanks *et al* 2005; Suggate and Almond, 2005). The LGCP is climatically complex with evidence of three drier periods from 31.5-26 ka, 24-21 ka and 21-

18.9 ka coinciding with periods of glacial advance and increased westerly wind intensity (Alloway *et al*, 1992; Suggate and Almond, 2005; Newnham *et al*, 2007a; Anderson *et al*, 2009; Whittaker *et al*, 2011). Wetter conditions interrupted these dry periods and also mark the end of the LGM, again coinciding with periods of ice retreat and reduced westerly wind intensity (Alloway *et al*, 1992; Suggate and Almond, 2005; Newnham *et al*, 2007a; Anderson *et al*, 2009; Whittaker *et al*, 2011).

8.1.3 Termination I (18.9-18.4 ka)

Termination I (Figure 8.1) marks the end of the LGM and the start of climate amelioration which the NZ-I CES places between 18-19 ka. The records used in the NZ-I CES and other records from across New Zealand are broadly synchronous. Likewise the timing of Termination I in the vegetation reconstructions used in this project closely matches the NZ-I CES. At Forks Lagoon climate begins to get wetter and presumably warmer at 18.9 ka whilst the onset of vegetation change at Lake Mudgie starts at 18.4 ka. The pattern of climate change from each site is marked by a decrease in herb abundance and an increase in the shrub and tree component. At Lake Mudgie, average herb abundance decreases by 10% whilst the shrub abundance increases to 21% and tree abundance to 4%.

The Howard Valley record is different from the two above as a clear vegetation change marking the start of Termination I is not recognisable. Therefore, the calculated vegetation composition at Howard Valley was derived from averaging the top samples from section 1. Herb abundance at this site decreases by 20% and shrub abundance reaches an average 34%. The tree component also increases to 8%. However, the pollen assemblage and vegetation pattern at the time was erratic and changes significantly from one sample to the next. The large shrub component was dominated by Asteraceae which comes in both shrub and herb forms, the later being more representative of alpine/sub-alpine vegetation. The increased shrub component may not necessarily represent climatic amelioration. However the vegetation record does show a changing vegetation community suggesting an unsettled environment preventing a stable community establishing at the site. This may represent a disturbance caused by ice melt leading to flooding events causing these disturbances. Therefore, this instability

may be linked to climate change but the age of section 1 may also be too old to capture the event which may explain the lack of a clear Termination I in the vegetation record.

Unlike the other sites, Manks Tarn provides only a postglacial vegetation reconstruction. The pollen assemblage at the core base contains little herb and grass taxa and is dominated by the shrub component by the start of sediment deposition. However, the base of this record is dated 18.1 ka, only 300 years younger than the age of the onset of Termination 1 at Lake Mudgie and given the error of ^{14}C dating the ages are broadly the same. Before shrub vegetation can establish the site first needs to be ice free, become geomorphically stable, been inhabited by primary colonisers and had soil developed, all of which would take some time. This means the area was ice free prior to 18.1 ka and colonisation of the site, once initiated, was rapid.

The pollen assemblage at the base of the Manks Tarn record is similar to Lake Mudgie pollen zone LM1c suggesting the base of Manks Tarn pollen record marks the end of Termination I indicating the onset of deglaciation occurred not long prior to 18.1 ka.. The pattern of deglaciation at Lake Mudgie also suggests a change from herb taxa to shrub-dominated vegetation taking up to ~800 yrs. This is marked by a rapid decline in grass abundance coinciding with *Myrsine* and *Coprosma* shrub increase. A range of evidence from both terrestrial and marine records from New Zealand and surrounding oceans suggests a marked climatic amelioration began sometime between 19-17.3 ka (e.g. Newnham *et al*, 2003; Vandergoes and Fitzsimons, 2003; Alloway *et al*, 2007). A similar age of Termination I is recorded in Pukaki Basin on the eastern side of the Southern Alps (Schaefer *et al*, 2006; Vandergoes *et al*, 2008; Putnam *et al*, 2010a). Given the errors associated with radiocarbon dating the timing of this event recorded in this project is consistent with other records from across New Zealand and corresponds with the NZ-I CES range of 18-19 ka.

8.1.3.1 Termination I: Global context

Termination I or the start of deglaciation is recorded in a number of different proxy records spanning a range of geographic, geologic and climatic regions elsewhere in the mid-latitudes of the Southern Hemisphere. Like New Zealand's records, South America shows a broadly synchronous timing and pattern for this event. Termination I occurs ~18.9-17 ka and is marked by rapid ice retreat and vegetation response to climate amelioration (McCulloch and Davies, 2001; Moreno and León, 2003; Kaplan *et al*, 2004; Douglass *et al*, 2006; Boës and Fagel, 2008; Kaplan *et al*, 2008). Further south the Antarctic ice core records suggest Termination I to have occurred sometime between 19-17 ka (Blunier *et al*, 1998; Petit *et al*, 1999; Jouzel *et al*, 2001; Blunier *et al*, 2001). Termination I is not only hemispherically synchronous but appears to be globally synchronous. Clark *et al*'s (2009) global collation of ^{14}C , ^{10}Be and ^3He ages suggest the deglaciation began between 20-19 ka in the Northern Hemisphere and 19-18 ka in the southern hemisphere. Schaeffer *et al* (2006) also concluded a similar interhemispheric timing for deglaciation based on moraine exposure ages and suggested that the most straight forward explanation for a synchronous timing of the end of the LGM is a global rise in summer temperatures.

8.1.4 LGIT (18.9-10.8 ka)

The LGIT (Figure 8.1) is a period characterised by climate amelioration, glacial retreat, and a change from herb and grass dominated vegetation to a gradual increase in woody vegetation both in richness and diversity. The NZ-I CES ages the period of climate amelioration as spanning 18-11.6 ka with a period of brief cooling interrupting the pattern of warming between 13.5-11.6 ka (Alloway *et al*, 2007). In this project, only two sites record the pattern of deglaciation, both of which are located in the Westland region (Lake Mudgie and Manks Tarn). Both records show a four stage vegetation pattern change that is related to climate warming and potentially precipitation change. Initially the onset of climate amelioration and the early stages of LGIT are dominated by *Coprosma/Myrsine* shrub growth, beginning at 18.4-18.1 ka. Increased warming is followed by the second vegetation stage characterised by the spread of *Griselinia*, *Libocedrus* and *Pseudopanax* joining the *Coprosma/Myrsine* shrub component, increasing

vegetation diversity and producing a low lying broadleaved tree and shrub community at 16.3-15 ka. The increase in *Griselinia* may also indicate an increase in precipitation. The third phase of warming, occurring between 14.3-13.9 ka, is characterised by the increase in the montane/lowland taxon, *Weinmannia*. The rise of *Weinmannia* is greater in Manks Tarn compared to Lake Mudgie but both records suggest a continued warming and increased precipitation. The final stage of warming occurs at 13.2-12.4 ka and is characterised by a rapid increase in *D. cupressinum* dominated podocarp/hardwood forest (Figure 8.2). The vegetation composition change indicates an increase in drought intolerant taxa towards the later stages of the LGIT and continued warming throughout the LGIT. No irrefutable climatic reversals are evident in either record, although a LGR cannot be completely ruled out (the LGR will be discussed in more detail in the following section).

The pattern and timing of the vegetation change as described above (see Figure 8.2) is strikingly similar between the records and shows a strong regional pattern to climate change. The timing of vegetation change and pattern of taxa appearance is also similar to the Okarito vegetation pattern during the LGIT, except instead of *Weinmannia racemosa*, *Metrosideros* is present (Newnham et al, 2007a). These vegetation reconstructions show that once conditions are suitable, a rapid expansion of vegetation occurs and follows a specific vegetation succession. Lake Mudgie has a less well constrained chronology than Manks Tarn, but due to the same distinct vegetation patterns and abrupt transitions evident in both records it is possible to use the Manks Tarn record to better validate the timing of some vegetation transitions.

At both Manks Tarn and Lake Mudgie a sudden increase followed by a persistent abundance of *Isoetes alpinus* occurs between 16.3-12.4 ka and 15.6-13.2 ka, respectively. The timing is broadly synchronous between the records, suggesting conditions to promote *I. alpinus* growth were similar at each site. *I. alpinus* is an aquatic macrophyte restricted to a shallow lake level depth range of 0.4-4 m (Brown, 1975; see Section 7.6.2). Therefore its presence is indicative of an open water, albeit shallow, environment. Its occurrence in both records has been attributed to a lake level lowering unrelated to a direct climate influence but rather

a process of gradual terrestrialisation of the site. This is visually evident in the Lake Mudgie core as a transition from blue-grey inorganic laminated lacustrine silts, to brown silts culminating in organic rich brown mud. The occurrence of *I. alpinus* coincides with the transition from grey inorganic silts to organic brown silts, suggesting increased organic material was entering the site and indicating terrestrialisation was occurring. However, recent research from Fletcher and Moreno (2011) uses the pattern of *I. alpinus* abundance at Okarito, south Westland region, as a precipitation signal during the LGIT. An increase in *I. alpinus* abundance was inferred to relate to increased precipitation. Conversely a decreased *I. alpinus* abundance is representative of drier conditions. Therefore a more rigorous discussion on *I. alpinus* appearance in the Westland sites is necessary.

As mentioned above the stratigraphic changes at Lake Mudgie clearly show a gradual transition from a lake environment to a terrestrialised boggy environment. The Manks Tarn stratigraphy is composed of homogenous brown detrital mud containing no visible evidence of sedimentological change. However, at both sites evidence in the pollen and sediment records also point towards terrestrialisation. The PCA results from both sites indicate *I. alpinus* appearance is in close relationship with *Coprosma*, *Griselinia*, *Myrsine*, *Pseudopanax* and *Pseudowintera*. These taxa inhabit the montane/sub-alpine zone and suggest conditions were milder than the LGCP although still colder than contemporary conditions. The presence of these taxa combined with evidence of increased total % carbon in the sediment indicates an increased organic component was entering the lake environment and contributing to sediment accumulation. The coincidence of *I. alpinus* with warmer climates may represent an indirect response to climate change whereby the increase in temperature led to greater vegetation productivity and organic material entering the lake; an additional contribution to sediment accumulation which resulted in decreased water levels.

The appearance of *Leptospermum type* in both records after the initial rise of *I. alpinus* also indicates a continued lake level lowering. In the contemporary environment, *Leptospermum type* is dominant on peat bog margins and poorly drained areas such as lake margins (Wardle, 1979). Its appearance with *I. alpinus*

suggests the lake fringes were becoming increasingly terrestrialised, which would have enabled *Leptospermum type* to establish progressively closer to the core site. The continued high abundance of *Leptospermum type* after *I. alpinus* disappearance from the record suggests the core site no longer contained an open water body. Instead the area was fully terrestrialised but poorly drained providing suitable conditions for continued *Leptospermum type* growth. The pollen, sediment and stratigraphic evidence all point towards terrestrialisation causing lake levels to lower and promoting *I. alpinus* growth. The results do not appear to represent a precipitation signal.

Other records from the Westland region, e.g. Okarito and Gillespie Beach Road sites (see Vandergoes and Fitzsimons, 2003; Newnham *et al*, 2007a), also show a significant change in *I. alpinus* abundance during the LGIT. Okarito shows a similar stratigraphic change and *I. alpinus* pattern as the Lake Mudgie record with an abrupt increase of *I. alpinus* to nearly 100% at 14.2 ka and dropping to 0% at 12 ka. This coincides with a stratigraphic transition from blue-grey silts to brown organic silt. This is followed by the sustained appearance of *Empodisma*, Cyperaceae and *Leptospermum type*, all of which are representative of peat bog/wetland environments (Wardle, 1979). The disappearance of *I. alpinus* at Okarito, at a similar timing as in Manks Tarn and Lake Mudgie, suggests there is no longer an open water body at the core location. At all these sites, therefore, lake level lowering is caused by gradual terrestrialisation of the site during the LGIT. However, Fletcher and Moreno (2011) suggest the oscillation in the *I. alpinus* abundance during this period represents a precipitation signal. These findings question the interpretation of Fletcher and Moreno (2011) that the *I. alpinus* oscillation at Okarito during this period represents a precipitation signal.

Figure 8.2: Comparison of Lake Mudgie and Manks Tarn LGIT vegetation pattern. The dry land pollen is in order of appearance followed by the wetland, tree fern and fern pollen.

8.1.4.1 Late Glacial Reversal (LGR)

The Manks Tarn vegetation reconstruction points towards a continuous pattern of forest development and inferred temperature and precipitation increase from the start of deglaciation to the early Holocene (18.1-10.8 ka). The Lake Mudgie record indicates a minor reversal in forest development between 520-510 cm. This is characterised by a disturbance and small decline in *Dacrydium cupressinum* growth combined with a minor increase in *Weinmannia*, *Libocedrus* and *Elaeocarpus*. These taxa were more prevalent deeper in the core, while *Fuscospora*, *Chenopodiaceae* and *Gonocarpus* were also present in low abundance during the LGCP. A distinct lithostratigraphic change occurred at the same time, characterised by an influx of clay size particles into the site, suggesting some form of environmental disturbance in the area. The slight increase in colder tolerant taxa at the expense of *D. cupressinum* may indicate a period of minor cooling. Both models used in the transfer function suggest a cooling event is visible although the temperature depressions predicted by each model are very different. There are also a number of limitations with these models which are discussed in Section 8.2.4 and therefore these temperature reconstructions are tentative. On the basis of the Lake Mudgie chronology, this event occurred between 13.4-11.9 ka broadly coeval with other evidence of late glacial climate reversals from elsewhere in New Zealand (Newnham and Lowe, 2000; Turney *et al*, 2003; McGlone *et al*, 2004; Williams *et al*, 2004; Newnham *et al*, 2007a, Vandergoes *et al*, 2008; Kaplan *et al*, 2010).

Although the above cooling inference is plausible, coinciding with the disruption in *D. cupressinum* growth is a small peak in *Prumnopitys ferruginea* and an increase in the tree fern *Cyathea smithii*. Both species indicate increased precipitation and continued warming (Wardle, 1991) and are abundant in significant numbers at a similar timing in Manks Tarn. Some other records from New Zealand also show limited or no evidence of climate cooling but rather suggest there was an increase in precipitation at the time (McGlone, 1995; Singer *et al*, 1998; Vandergoes and Fitzsimmons, 2003). Also, a flux of coarser material is usually representative of a cooling event (Newnham *et al*, 2007a), whereas at Lake Mudgie, an increase in finer particles is recorded. These two scenarios: precipitation forced and temperature forced vegetation disruption will be discussed below.

Controversial evidence for a late glacial ice advance has been suggested from the Waiho Loop Moraine, southern Westland (see Section 2.6.1.1 for more detail). Modelling climatic parameters needed to cause the Franz Josef Glacial to advance as far as the Waiho Loop, Anderson and Mackintosh (2006) estimated a 3-4.7 °C cooling in MAT would be necessary when assuming no change in precipitation. The upper end of this temperature depression range is similar to that experienced during the LGM. This amount of cooling is not represented in proxy records which suggest a maximum of 2-3 °C cooling during this period (Newnham *et al*, 2007a; Vandergoes *et al*, 2008; Newnham *et al*, 2011) and a treeline depression of only 100-200 m (Newnham *et al*, 2007a). The small disturbance in forest development and minor increase in cool tolerant taxa at Lake Mudgie suggest only a very minor cooling would be experienced, if at all, during the LGR period and likewise does not support a 4 °C cooling.

Some records in New Zealand show little or no evidence of a cooling event occurring during the ACR/YDC interval. Instead there is some evidence to suggest an increase in precipitation occurred over New Zealand at this time caused by an increased intensity of the southern westerlies and in turn resulting in a glacial advance (Singer *et al*, 1998; Vandergoes and Fitzsimons, 2003; Robinson *et al*, 2004). Both the Lake Mudgie and Manks Tarn vegetation reconstruction point toward very wet conditions at 13.4-11.9 ka with the Lake Mudgie record suggesting the region experienced an increase in precipitation. However, Anderson and Mackintosh's (2006) model results estimates precipitation would need to increase 330-450% from current levels to induce a glacial advance reaching the Waiho Loop, assuming no temperature depression was experienced as well. This region already experiences exceptionally high precipitation and a 330-450% increase seems to be implausible.

An alternative scenario to consider is a combination of the two postulates above whereby an increase in precipitation is accompanied by a decrease in temperature. Anderson and Mackintosh (2006) showed a 3.2 °C cooling combined with a 39% increase in precipitation from existing levels would be sufficient to advance the Franz Josef Glacier to the Waiho Loop location. However this scenario still appears to be inconsistent with palaeoecological

reconstruction. A MAT decrease of 3.2 °C would cause a 640 m vegetation depression. Under these conditions, at Lake Mudgie, at 143 m.a.s.l, montane forest including *Weinmannia racemosa*, *Libocedrus bidwillii* and a subcanopy of *Myrsine divaricata*, *Griselinia lucida*, *G. littoralis*, *Pseudopanax simplex*, *Pseudowintera colorata*, and *Coprosma foetidissima* (Wardle, 1977) would dominate. Although a small increase in some of these taxa are evident in the Lake Mudgie record, *Dacrydium cuppressinum* remains the dominant species. The upper growing limit of a *D. cuppressinum* dominated forest is 500 m.a.s.l, 350 m above the Lake Mudgie site. A maximum MAT depression, which could still produces a *D. cuppressinum* dominant vegetation community, would be ~1.8 °C and is significantly lower than the Anderson and Mackintosh (2006) estimate. Instead the Lake Mudgie MAT depression is more consistent with other palaeoenvironmental reconstructions (e.g. Newnham and Lowe, 2000). However, *D. cuppressinum* have reasonably well dispersed pollen and therefore a high abundance of this taxon could still be deposited in the pollen record potentially masking the true degree of cooling at the time. Vandergoes and Fitzsimons (2003) also concede that the vegetation characteristics from the lowland Westland region, which experiences high precipitation amounts, may not provide a climatically sensitive enough proxy to pick up a relatively minor short-term temperature decrease. Therefore this minor disruption in the Lake Mudgie pollen record may have underrepresented any cooling experienced in the region.

The Okarito pollen record shares a similar problem to Lake Mudgie, as it also shows a minor disturbance in the vegetation record. This was initially discounted as a cooling episode by Vandergoes and Fitzsimons (2003), but later revised by Newnham *et al* (2007) as representing a LGR. The original paper by Vandergoes and Fitzsimons (2003) contained results from three sites in south Westland but only Okarito showed a possible LGR, with the other two, like Manks Tarn, indicating a continuous pattern of forest development. However, elsewhere in New Zealand LGR is more clearly defined e.g. Kaipo Bog (Newnham and Lowe, 2000), Kettle Hole Bog (Turney *et al*, 2003), Boundary Stream (Vandergoes *et al*, 2008). Again, it should be noted that the lowland vegetation composition from Westland may not be sensitive enough to record a reversal, unlike the regional vegetation

from the other sites mentioned above. From the evidence reported above the status of an LGR in Westland remains inconclusive although as discussed above the data are not inconsistent with a scenario of minor temperature depression accompanied by minor precipitation increase.

Recently, Newnham *et al* (2011) suggested the timing, extent and temporal length of the LGR has a latitudinal pattern. Using quantified estimates of MAT the extent of cooling could be assessed, albeit with broad errors, and the results indicating a northward decrease in cooling. Okarito, in south Westland, shows a 2 °C cooling, Kaipo bog in central North Island reconstructs a 0.5-1 °C cooling and Pukaki, from northern North Island, estimates a maximum MAT cooling of only 0.5 °C. The timing of this cooling phase also differs from site to site with cooling occurring earlier at Okarito, beginning at 14.1 ka, and latest at Pukaki (beginning at 13 ka). Lake Mudgie is located near to Okarito and it can be expected that a similar pattern of cooling would be experienced at the two sites. The qualitative vegetation reconstruction for Lake Mudgie does not clearly show a climate reversal except for the possible disturbance between 13.2-11.9. The quantitative pollen-climate reconstructions propose a different conclusion however.

Lake Mudgie's quantitative pollen-climate reconstructions indicate a cooling trend between 13.2-11.9 ka, although these results are tentative. This is 1 ka later than at Okarito, and more closely aligned with the timing of cooling at Pukaki. However, this could be a factor of the limited chronology for the Lake Mudgie record causing the temporal disparity. The Boundary Stream record located in the Mackenzie Basin, also dates the LGR between 14.2-13.2 ka, again implying the Lake Mudgie age is unexpectedly young. The extent of estimated cooling ranges 1.3-2.2 °C for the W-MAT model and 2.1-4.8 °C for the PLS-C3 model, the latter appearing to be overestimating MAT depressions. The extent of cooling, particularly the W-MAT, is similar to the cooling estimated at Okarito. Lake Mudgie vegetation record, as mentioned above, also suggests an increase in precipitation occurred during the period. This also corresponds with the Anderson and Macintosh (2006) scenario of 2.2 °C cooling combined with a 39 % increase in precipitation. However, it should be stressed that there are limitations with the quantitative pollen-climate models impeding firm conclusions of temperature change at the site (Section

8.2.4). To conclude, the LGR is neither ruled out nor confirmed by the evidence presented here.

8.1.4.2 LGIT global context

Antarctic ice core records suggest a two step warming during the LGIT between 18 ka and 14 ka. The warming pattern is interrupted by a cold reversal with lowest temperatures reached between 13.5-12.5 ka. This is followed by a second warming phase reaching Holocene conditions by 11.5 ka. An overall 9 °C temperature range from the LGM to Holocene was recorded in the Antarctic EPICA Dome C ice core record (Jouzel *et al*, 2001). Evidence from Australia shows a similar pattern as New Zealand, again showing a rapid response to climate amelioration. The Lake Selina record, Tasmania, shows a LGR occurring between 13.5-11.7 ka (Colhoun *et al*, 2010) whilst Turney *et al* (2004) and Muller *et al* (2008), report a period of wetter climate in north-eastern Australia between 13.5-12.5 ka. Both records suggest a southward migration of the ITCZ and zonal movements of mid-latitude storm tracks causing wetter conditions to be experienced in north-eastern Australia coinciding with the ACR and the Northern Hemispheric Bølling-Allerød interval. Southern Ocean marine records also suggest a cooling event occurring coincidently with the ACR (Carter *et al*, 2008). In New Zealand, exposure ages of moraines in ESI date a glacial advance culminating 13 ka, again coinciding with the peak period of cold in the Antarctic ice core records. It was concluded that the most likely cause for cooling at this time was the migration northwards of the subtropical front combined with an expansion of the colder Southern Ocean waters (Putnam *et al*, 2010b).

The pattern of deglaciation in southern South American records appears to pose similar problems of disparity between records as seen in New Zealand. As mentioned above all the records suggest an abrupt and rapid pattern of deglaciation, but there is some difference of opinion as to whether a LGR occurred in South America. Some reconstructions show a continuous trend in forest development and warming culminating at the start of the Holocene at ~11.4 ka (Ashworth and Hogansen, 1993; Haberle and Bennett, 2004; Massaferro *et al*, 2005). However, other records suggest a cooling period with ACR timing interrupting this pattern of climate amelioration (McCullock *et al*, 2005; Douglass

et al, 2006; Boës and Fagel, 2008). Others suggest either an intermediate ACR/YDC timing of cooling (Hajdas *et al*, 2003; Moreno and León, 2003; Sugden *et al*, 2005) or a YDC timing (Ariztegui *et al*, 1997; Heusser, 1998; Strelin and Malagnino, 2000). As in New Zealand, this varied timing could well be due to some sites in southern South America being climatically insensitive compared to others.

8.1.5 Holocene

The NZ-I CES shows a period of early Holocene warmth occurring between 11.6-10.8 ka (Alloway *et al*, 2007). *Dacrydium cupressinum* in the contemporary environment dominates the warmer lowland areas. The dominance of *D. cupressinum* in the records presented here is inferred to represent temperatures that are similar to present. At Lake Mudgie, *D. cupressinum* becomes the dominant taxon at 11.9 ka and at Manks Tarn, later, at 10.8 ka (Figure 8.1), corresponding with the NZ-I CES. Similar timing of this transition has also been identified in other records: Kaipo Bog 11.5-11 ka (Newnham and Lowe, 2000), Okarito ~12-11 ka (Newnham *et al*, 2007a), Boundary Stream, ~11.5 ka (Vandergoes *et al*, 2008), Pukaki crater 11.5 ka (Sandiford *et al*, 2003), Kettlehole bog, Cass Basin 11.3-10.4 ka (McGlone *et al*, 2004). This indicates that the Holocene onset was broadly synchronous for all regions of New Zealand. The Holocene period for both records show a continued dominance of *D. cupressinum*-dominated podocarp lowland forest, indicating warm conditions continued.

The start of the Holocene in New Zealand appears to be broadly similar throughout the Southern Hemisphere. In Tasmania rapid rainforest expansion reached a maximum during the early Holocene, 10.7 ka, and climate was even possibly 1-2 °C warmer than current conditions and wetter (MacPhail 1979). Likewise in South America a broadly synchronous timing of the Holocene is recorded around 11.8-10.6 ka (Moreno and León, 2003; Massaferro *et al*, 2005; Boës and Fagel, 2008; Unkel *et al*, 2008). In Antarctic records similar temperature optima occur between 11.5-9 ka (Masson *et al*, 2000). Walker *et al* (2009) dates the onset of the Holocene, based on multi-parameter annual layer counting in North Greenland Ice core Project (NGRIP) and other auxiliary records, from

around the globe, at 11.7 ka. The global synchronicity of Holocene warming suggest direct global transfer of climatic change via changes in atmospheric greenhouse gas and water vapour concentration in the tropics (Lowell *et al*, 1995; Denton *et al*, 1999).

8.1.6 Summary of climate change in this project

Table 8.1 is a summary table displaying the timing of climatic events and transitions as recorded in this project and compared to the NZ-I CES. Overall the timing of climate change is broadly synchronous between sites and with the NZ-I CES. The timing of Termination I, the possible LGR at Lake Mudgie, and the start of the Holocene are particularly closely matched. However, the timing of climate change during the LGCP is more variable between the sites and the CES. The onset of the LGCP possibly begins earlier than suggested by the NZ-I group and this earlier onset is supported by other records (Section 8.1.2). This suggests an extension of the NZ-I CES timeframe is needed (Section 8.2.3.1). The period of 'milder' and more variable climate appears to be more complex than portrayed by the NZ-I group (see Section 8.2.2). The onset of the LGM is also difficult to define. The Lake Mudgie record shows cooling began ~23.2 ka which corresponds with the earliest timing of this event in the CES. At Howard Valley and Forks Lagoon the LGM, marked by drier conditions at Forks Lagoon, occurred between 21-18.9 ka, closely matching the NZ-I CES. Better chronological control and higher resolution may help to address the difficulties of pinning down the timing of climate change in this period. Also, an increase in records covering this period in New Zealand would also be useful before firm conclusions can be made on the pattern of climate change and the possible forcing factors. Section 8.2.3 will discuss the NZ-I CES in more detail.

Termination I began sometime between 18.9-18.5 ka and closely aligns with the NZ-I CES timing of 19-18 ka. The pattern of deglaciation in the two Westland sites closely matches the pattern of deglaciation in other Westland sites. Manks Tarn does not show a LGR in the record but instead a continuous pattern of climate amelioration. The qualitative evidence for Lake Mudgie suggests a possible vegetation disturbance, interrupting forest development, occurred between 13.2-

11.9 ka and corresponds with the timing of the LGR in the NZ-I CES. Whether or not this is attributable to a cooling event cannot be concluded from the qualitative evidence alone. The quantitative pollen-climate transfer functions calculate a cooling trend at this time, suggesting a possible LGR occurred, but methodological limitations prevent firm conclusions from being made. The Holocene period begins between 11.9-10.8 ka and is characterised by *D. cupressinum*-dominated podocarp forest and conditions which are similar to the contemporary environment.

Event	NZ-I CES	Howard Valley	Forks Lagoon	Lake Mudgie	Manks Tarn
MIS 3	-	>39-<34	-	-	-
LGCP	29-18	>28.5-18.9	>31.7-19	-	-
'milder interval'	27-21	28.5, 26.2-22.2	26-24, 21	>26.6-23.2	-
LGM	21-18	20-18.9	21-19	23.3-18.5	-
Termination I	19-18	-	19-17	18.5-17.7	>18.1
LGIT	18-10.6	-	-	17.7-11.8	18.1-10.8
LGR	13.5-11.6	-	-	13.4-11.9?	-
Holocene	11.6-10.8	-	-	11.9	11.2-10.8

Table 8.1 Summary table comparing the timing of climate transitions and millennial scale events between the four project records and the NZ-I CES. All ages are ka

8.2 Broader context

This section expands upon the first section by discussing the broader themes touched upon above. A discussion of the age of the KOT layer will start this section followed by an assessment and critique of the NZ-I CES. This critique also considers the regional locations of sites as a factor in reconstruction of climate signals. Detailed discussion of the LGM and LGCP climate changes in this broader context will end this section.

8.2.1 Chronology and age models

Chronology and age models are an integral tool used in palaeoenvironmental reconstructions, without which the timing of climate events and/or transitions and the duration of events could not be estimated. Also, comparisons between different palaeoreconstructions would prove difficult and the formulation of a climate event stratigraphy, such as the NZ-I CES, would be impossible. For this project chronology and age models have been extensively used to define the timing of climatic transitions and for comparisons to other records. However, there are important influences and limitations of chronology on age models that also have an influence on the conclusions made in this project. This includes the effect of calibrating radiocarbon years to calendar years and using mean calendar years for age models and for defining temporal boundaries of palaeoenvironmental change in reconstructions. These limitations are discussed further below.

The calibration of radiocarbon ages to calendar ages markedly decreases the age precision and increase the 2σ age-range for each sampled dated, thereby creating greater error margins in age models. This is due firstly to inbuilt uncertainties in the available calibration models (e.g. Reimer *et al* 2009) and secondly to the changing ^{14}C production rate over time, that is still not fully understood (Taylor *et al*, 1992). During the calibration process a single mode, Gaussian structured, radiocarbon age frequently calibrates into a multi-modal calendar age which has a broader 2σ age-range and thus reducing the age precision. This makes producing a defined age boundary for a palaeoenvironmental event/transition difficult as the event/transition can be ascribed to a broad age range. Often, and including this study, palaeoenvironmental reconstructions use mean calendar age when defining climate change boundaries, however, this leads onto a number of other limitations. The radiocarbon ages analysed in this study are all calibrated into calendar years and are initially expressed as mean calendar age with 2σ age-range (see tables 4.x, 5.x, 6.x and 7.x). The 2σ age-range illustrates the 95% probability that the true age of a specific sample lies somewhere within that age-range. The greater the 2σ age-range the lower the age precision. Nevertheless, throughout the thesis the mean calendar age is expressed when reporting the timing of palaeoenvironmental change that is evident in the pollen records. This is common practice among palaeoenvironmental reconstructions, particularly when the age

expressed has been extrapolated from an age-depth model (e.g. Alloway *et al*, 2007; Newnham *et al*, 2007; Lowe *et al*, 2008). However, this can give a false impression of age precision and it is important to refer to the whole 2σ age-range, especially when making comparisons with other reconstructions. An apparent difference in timing of a climate event between records based upon mean calendar age may prove to not be so different when the whole 2σ age-range is taken into account.

Age-depths models used in this project were created using simple linear interpolation constructed by joining the mean values of individual ages. This assumes a constant sedimentation rate between each dated horizon (refer to section 3.x for more information). However, this method also contains some inherent limitations, the most obvious being that sedimentation rates are rarely constant. However, there are other limitations regarding the chronology and the age-range of individual samples. A coherent age-depth plot, i.e. an age-depth model consisting of ages in chronological order, does not necessarily mean the calendar age of individual samples are reliable indicators of age. If the difference in mean calendar age $\pm 2\sigma$ age-range is of sufficient magnitude between sampled horizons to not overlap, such as the majority of the Manks Tarn ages, then the model may well be valid. However, if the mean calendar age $\pm 2\sigma$ age-range of a sample overlaps an underlying/overlying sample horizon, such as Forks Lagoon, then a reliable age-depth model is more difficult to determine (Lowe *et al*, 2007). In such cases using multiple samples from the same horizon to produce a number of ages can help, especially if these samples reproduce similar ages, as is the case with most duplicate ages for Forks Lagoon (refer to section 7.x for a greater discussion of the ages from Forks Lagoon). Also, using the prior knowledge that sediment accumulation occurs in chronological order and therefore samples taken up the core should produce younger ages, can also increase the validity of the model.

Despite the limitations highlighted above, the chronology and age models used form a good basis for discerning the timing of palaeoclimatic change evident in this project and for allowing broad comparisons between the project's records and published reconstructions. Future work on the chronologies at the sites used for

this study, including an increase in the amount of radiocarbon ages at some sites and the use of Bayesian statistical approaches (refer to section 9.x), will help improve the age models.

8.2.2 The age of the KOT layer

Tephrostratigraphical data were used in the NZ-I CES to aid the synchronising of the records and to improve the age model. Using tephra as an age marker was particularly important in the fragmentary loess records where suitable datable material was at a minimum (Suggate and Almond, 2005). Therefore, reliable ages for these tephra layers are necessary for an accurate chronology and to accurately date the climate transitions and millennial scale events. The KOT is a key tephra marker as it lies near the beginning of the LGCP and within the LGCP milder interval. It is also the only tephra deposit in the South Island stratigraphically tying records from both the South Island and North Island sites. An accurate age for the KOT is therefore very important. However, the conflicting ages for this event have been reported, making it difficult to pinpoint a reliable age, although work has been carried out in an attempt to accomplish this (Wilson *et al*, 1995; Lowe *et al*, 2008; Vandergoes *et al*, 2011). In this project three records contain the KOT layer and discussion now turns to the age of this layer in connection with the pollen stratigraphy and chronology of these sites.

The radiocarbon ages produced for this project can be used to assess the compatibility of the two KOT ages of Lowe *et al* (2008) and Vandergoes *et al* (2011) with this project. Figure 8.x is a compilation of all the calibrated radiocarbon ages used in this project placed into three groups. Group 1 are the ages from samples taken stratigraphically below the KOT layer at each site. Group 2 are ages from samples taken from the sediment horizon immediately above of below the KOT layer. Group 3 are ages stratigraphically above the KOT layer.

In general, and to be expected, the samples taken above the KOT layer are younger than the accepted and revised KOT ages, and the samples below the KOT layer are older than the accepted and refined ages. However, there are some samples from the Forks Lagoon record, which are from the sediment horizons

below the KOT layer and produce younger ages than the Lowe *et al* (2008) age and overlapping the Vandergoes *et al* (2011) ages. This has been discussed in chapter 7.4 and has been attributed to use of plant macrofossils for dating, which are more susceptible to contamination. Of the three ages dating the KOT layer directly, two produce younger mean calendar ages than those of Vandergoes *et al* (2011). The youngest age, SUERC-28872, from Lake Mudgie is considered to be too young (20.8 ka). Research has shown that high rainfall regions are prone to contamination by younger organic sediments. In addition, the identification of other tephra layers in the North Island which overlay and hence post-date the KOT, also have ages which are older than the Lake Mudgie estimated age, further indicating that this age is too young (Lowe *et al*, 2010; See Section 3.5.2) Previous reasons for this have already been discussed in Section 4.4.1, justifying the use of the Vandergoes age of 25.4 ± 0.4 ka instead. Howard Valley is a site used by Vandergoes to refine the age of the KOT. In this project the horizon directly below the tephra dates this deposit as 25.2 ± 0.4 ka (OS-83041) and is similar to the Vandergoes age, corresponding with a younger timing for the eruption. A sample from at least 1 m below the tephra deposit produces an older age of 26.2 ka which is still 1000 yrs younger than the Lowe *et al* (2008) age, suggesting that the KOT was deposited later than previously thought. Nevertheless, the age-range of the Howard Valley samples (OS-83041 and OS-83042) make it difficult to decide with certainty whether the Vandergoes *et al* (2011) age is correct. However, the stratigraphic position of the ages and the OS-83041 mean calendar age $\pm 2\sigma$ age-range aligning more closely with the Vandergoes *et al* (2011) age, suggests this younger KOT age is more appropriate for this site.

Figure 8.3: Calibrated radiocarbon ages for each site (Light blue=Lake Mudgie; Orange= Manks Tarn; Dark Blue=Howard Valley; Green=Forks Lagoon). The ages are split into three groups; Group 1 – samples taken below the KOT layer, Group 2 – samples dating the KOT layer, Group 3 – samples taken above the KOT layer. The accepted age of the KOT layer (Dark gray; Lowe *et al*, 2008) and the refined ages (Light gray; Vandergoes *et al*, 2011) are also plotted on the figure.

At Forks Lagoon the KOT layer was not directly dated, but the core chronology provides a good basis for age estimation. Figure 8.4 is a comparison of three age-depth models for Forks Lagoon. The first model is based solely on the calibrated ^{14}C ages, the second uses the Vandergoes *et al* (2011) age for the KOT and the third the Lowe *et al* (2008) age for the KOT. The extrapolated age of the tephra

horizon in model (a) (Figure 8.4) is ~24.5 ka, 900 years younger than the Vandergoes age but with overlapping errors (Figure 8.4). In comparison the Lowe *et al* (2008) age for the KOT does not correspond well with the sediment deposition rates. In summary the research here provides strong support for the revision of the age of the KOT.

Figure 8.4: Comparison of three Forks Lagoon age-depth models using two KOT ages (b) and (c) with an age-depth model excluding the KOT tephra age (a). The extrapolated age for the tephra later in (a) is outlined by the box

8.2.3 Precipitation variability during the LGCP

A range of evidence from palaeoproxy records suggests New Zealand's climate during the LGCP was dry (McGlone *et al*, 1993; Shulmeister *et al*, 2001; Whittaker *et al*, 2011) with an increased intensity of westerly winds (Alloway *et al*, 1992; Hellstrom *et al*, 1998; Shulmeister *et al*, 2004). Recent climate modelling of the LGM (21 ka) in New Zealand also points towards drier conditions over the whole of New Zealand regardless of an increased intensity of westerlies (Drost *et al*, 2006; Rojas *et al*, 2009). Whether these same conditions persisted throughout the LGCP is unclear due to the paucity of records covering this period. However, Section 7.6.3.1 suggests precipitation was variable throughout the LGCP. An understanding of the pattern of precipitation change in relation to glacial activity and temperature change can provide insight into the possible processes and forcing factors of climate change in New Zealand. The following discusses the pattern of precipitation change during the LGCP, how the pattern of precipitation relates to other evidence of climate change and tentatively suggests possible mechanisms and forcing factors of climate change in New Zealand.

The palaeoprecipitation signal from Forks Lagoon suggests three drier periods occurred between 31.4-26 ka, 24-21 ka and 21-18.9 ka. These dry periods are interrupted by two wetter periods between 26-24 ka and 21 ka. Other paleoclimate records providing precipitation signals are not common in the South Island. An exception is a speleothem record from north-west Westland providing a sensitive precipitation record. Whittaker *et al*'s (2011) $\delta^{13}\text{C}$ speleothem record from Hollywood Cave indicates that overall MIS 2 was drier than MIS 4. However, three millennial-scale wetter periods (Figure 8.5) occurred between 30.5-29 ka, 25.5-24.2 ka and 22.6-21.4 ka, which, like the Forks Lagoon record, suggest the climate was not uniformly dry during the LGCP. The two later wetter phases recognised in Hollywood cave appear to roughly coincide with wetter periods inferred at Forks Lagoon suggesting a near synchronous pattern of wetter and drier conditions between the two regions (Figure 8.5).

Hellstrom *et al* (1998) provides a record of changing $\delta^{18}\text{O}$ isotopes values in meteoric water recorded in speleothems taken from Nettle Bed Cave in north-west

Nelson. Unlike the precipitation record from Hollywood Cave, Nettle Bed Cave does not provide a precipitation signal representing wet and dry conditions. Instead a change in $\delta^{18}\text{O}$ values represents a change of the source of precipitation, in this instance the Tasman Sea, which is related to oceanic fronts and atmospheric circulation changes. Hellstrom *et al* (1998) recognise three $\delta^{18}\text{O}$ excursions between 27-25 ka, 24.5-22 ka and 20-19 ka, (Figure 7.10) which, based on a limited chronology, is broadly synchronous with the dry periods at Forks Lagoon. The $\delta^{18}\text{O}$ excursions were concluded to be caused by the northward migration of the STF into the Tasman Sea (see Section 2.2.2) influencing the New Zealand temperature and $\delta^{18}\text{O}$ values of the source area. The migration of STF northward is also considered to have caused stronger westerly winds in central and southern New Zealand (McGlone *et al*, 1993; Hellstrom *et al*, 1998). The evidence discussed above suggest that dry periods inferred from Forks Lagoon and Hollywood Cave are broadly synchronous with increased westerlies and vice versa during periods of wetter climates.

The timing of westerly winds intensity variations recorded at Nettle Bed Cave, and precipitation changes at Forks Lagoon and Hollywood Cave may indicate a regional response to dominant synoptic climate types. As previously mentioned in Section 2.3, in the contemporary environment Trough regimes, which are characterised by reduced westerly wind dominance, can lead to an increase in precipitation in both the ESI and WSI (Kidson, 2000; Renwick, 2011). Therefore, a dominant Trough regime could explain the increase in precipitation at both Forks Lagoon and Hollywood Cave (Whittaker *et al*, 2011) and the reduced westerly wind intensity (Hellstrom *et al*, 1998). The pattern of regional climate during a dominant Zonal regime (a regime dominated by westerly winds) causes decreased precipitation in ESI and increased precipitation in WSI (Kidson, 2000; Renwick, 2011). Decreases in precipitation at Forks Lagoon are broadly synchronous with increased westerly wind intensity from the Nettle Bed Cave record and correspond with a Zonal regime. However, under this regime the Hollywood Cave precipitation record should show an increase in precipitation opposed to a decrease. During the LGM, climate models suggest an increase in westerly wind strength would not produce wetter conditions due to cold climatic condition (Rojas *et al*, 2009). The cause of drier conditions

during the LGCP at Forks Lagoon and Hollywood Cave (Whittaker *et al*, 2011), and period of inferred increase in westerly wind intensity, may relate to periods of very cold climate. This relationship is explored further below.

The variable pattern of climate during the LGCP is not restricted to precipitation change. The NZ-I CES places the LGCP milder interval between 27-21 ka, but again a complex pattern in temperature is evident with warmer periods between 27-24.5 and 23-22 ka (Alloway *et al*, 2007). At Lake Pukaki a milder interval spanning 26-24.5 ka (Newnham *et al*, 2007c) closely corresponds with the wetter climate phases in Forks Lagoon and Hollywood cave (Whittaker *et al*, 2011). Similar climate amelioration has also been recorded at a number of other sites in New Zealand (Alloway *et al*, 1992; Suggate and Almond, 2005; Woodward and Shulmeister, 2007). A later period of climate amelioration is evident in Williams *et al*'s (2010) speleothem record between 23-21.7 ka and again this is also evident in other records from New Zealand (Alloway *et al*, 2007; Newnham *et al*, 2007a) and broadly corresponds with evidence of wetter conditions at Hollywood Cave and Forks Lagoon. This suggests that wetter conditions in Forks Lagoon, and Hollywood Cave, corresponds with periods of possible glacial retreat, in north Westland (Suggate and Almond, 2005), and milder climates (e.g. Newnham *et al*, 2007a). Conversely, during periods of glacial advance (Suggate and Almond, 2005), and increased herb abundance representing cold environments (Newnham *et al*, 2007a) drier conditions occurred at Forks Lagoon. Drier periods during glacial advance corresponds with the climate modelling for the LGM (Drost *et al*, 2007; Rojas *et al*, 2009) suggesting a continuation of this relationship between cold and dry conditions into the LGCP. However, the Hollywood cave record moves away from this generalised pattern at the onset of the LGCP where evidence of cooling temperature coincides with wetter conditions, whilst dry conditions persist at Forks Lagoon. This could be a factor of the limited chronologically control at either site and/or other factors are at work influencing climate at this point of time. Better chronological control and further climate information is needed to better understand the causes of this disparity.

The paleoclimate reconstructions mentioned above suggest the climate during the LGCP was neither uniformly dry nor cold but variable. However, due to the different

KOT ages used between reconstructions it is important to stratigraphically compare records around this deposit, particularly between records which the chronology is dependent on the KOT age. Pukaki and its comparison to the Forks Lagoon record is an example of why stratigraphically comparing sites are important. The pollen record at Lake Pukaki was originally published by Sandiford *et al* (2003). In this original paper a milder period was not recognised but rather an extended cold period spanning 28-18 ka was concluded. In this record the KOT age used was 26.5 ka, although in some figures it was written as 26 ka. Newnham *et al* (2007c) used the same pollen record and identified a decline in grass abundance, which was ignored in the original paper. Newnham *et al* (2007c) suggest this grass decline represented a period of milder climate spanning 26-24.5 ka. However, the chronology uses the accepted Lowe *et al* (2008) age of 27.1 ± 1 ka. The Forks Lagoon record uses the 25.4 ± 0.4 ka KOT age. Therefore, between the three records mentioned, four different ages are used to date the same tephra. To make comparisons to either record chronologically would be incorrect and confusing. However, the KOT layer is a chronostratigraphic marker and the timing of its deposition anywhere in New Zealand, regardless of the radiocarbon age it receives, is the same. Consequently, records using different KOT ages can, and should, be stratigraphically compared. Stratigraphic comparison between records are discussed further below

At Forks Lagoon *Myriophyllum tryphyllum* concentrations start to increase just prior to the deposition of the KOT layer. *M. tryphyllum* concentrations continue to increase after the KOT deposit and then declines to nothing by 24 ka. The Okarito grass curve shows a similar pattern. Grass abundance begins to decline just prior to the KOT deposition and continues to decrease afterwards, until the start of the LGM where it increases again (Newnham *et al*, 2007a). In the North Island, Pukaki experiences a decline in grass abundance stratigraphically later than at Okarito. Grass abundance is still increasing around the time of the KOT deposition suggesting the climate was getting colder. However, after the KOT layer, a decline in grass abundance coinciding with the mid-LGCP occurs (Newnham *et al*, 2007c). Although this occurs stratigraphically later, the grass decline still coincides with the increase in *M. tryphyllum* concentrations. Therefore, even when stratigraphic comparisons between records occur, a similar conclusion can be made (see fig 8.5).

Increased *M. tryphillum* concentrations, implying wetter conditions at Forks Lagoon, coincide with a decrease in grass abundance and therefore evidence of milder climates at Okarito and Pukaki. However, it appears the onset of the milder conditions at Pukaki occurred later than at Okarito (Newnham *et al*, 2007a; 2007c)

The pattern of changing westerly winds intensity, the precipitation signal from both the ESI and WSI, and evidence of glacial activity suggest Zonal dominated regimes during glacial advance. Conversely, Trough regimes are dominant during glacial retreat and mild climates. However, using current synoptic conditions to explain the precipitation patterns in ESI during the LGCP may not necessarily be correct and may prove to be too simplistic. A number of potential forcing factors influencing climate change during this timeframe are possible and include; the influence of local insolation budget (Vandergoes *et al*, 2005), changing latitudinal position of the ITCZ (Whittaker *et al*, 2011), and an extended bipolar seesaw effect influencing the strength of the southern westerlies during the LGCP (Newnham *et al*, 2011). Although these mechanisms have been put forward more research is necessary before firm conclusions on the pattern, timing and cause of climate change in New Zealand can be established.

Figure 8.5: Comparison of proxy records from around New Zealand and the macrophyte record of Forks Lagoon which contains the *Myriophyllum tryphyllum* concentration curve. The red boxes represent periods of wetter climate. The blue boxes are period of colder climates evident in the proxy records, of glacial advance, or periods of stronger westerly wind flow. The grey line on the Forks Lagoon, Alloway *et al* (1992) and Okarito grass curve shows the chronological location of the KOT layer use by the different studies.

8.2.4 NZ-I CES

The NZ-I CES provides a good framework for the general pattern of climate change in New Zealand for the period 30-8 ka, which is needed for the processes linked to climate change to be better understood, as suggested above. Overall, large scale climate transitions occur broadly synchronously across New Zealand. However, there is greater variability in the timing of millennial scale events. Some of these differences might be attributable to poor chronological constraint, low sampling resolution, and different responses to climate change from different proxy methods used in the CES. However, addressing these limitations is an ongoing process which takes time. There are some other limitations which could be more easily and quickly resolved. The following will discuss these limitations in context of the timing and pattern of climate change represented in this project. The timing of LGCP climate

variability is not well resolved. The previous section has already noted these limitations which also apply to the NZ-I CES and are therefore not repeated here.

8.2.4.1 Temporal remit of the NZ-I CES

The NZ-I CES not only provides a framework for the period spanning 30-8 ka but also has some influence on future research in New Zealand. In this project, sampling strategy was designed to cover the NZ-I CES timeframe. Other new research is also restricted to the NZ-I CES temporal remit (e.g. Williams *et al*, 2010; Augustinus *et al*, 2011). However, this timeframe is restrictive and prevents the addition of important evidence of patterns of climate change which may also help to ascertain potential forcing factors of climate change in New Zealand. For instance, does the upper boundary exclude important information on late glacial cooling? This example will be discussed further below.

The NZ-I CES defines the onset of the LGCP occurring between 29-28 ka. This event is characterised by an increase in herb and cold tolerant taxa and a decrease in forest taxa. This boundary therefore marks the transition from milder to colder conditions. However, this is not necessarily the case. In the Okarito pollen record this boundary lies near to the peak abundance of herb and grassland taxa suggesting significant cooling had already commenced. The 30 ka boundary used by Newnham *et al* (2007a) marks the onset of the LGCP near the top of a rising limb in herb abundance. In this record the 29-28 ka boundary ignores the earlier transition from mild to cold conditions which would be useful when investigating the potential climate change mechanisms. Figure 8.6 is taken from Vandergoes *et al* (2005) and compares the pattern of change in a number of records including the Okarito summary pollen diagram. An increase in herb taxa at the expense of tree and shrub abundance appears to begin at ~35 ka. Not only does this correspond with other evidence of glacial advance in the Westland regions (see Section 8.1.2) but also with cooling in the EPICA Dome C ice core record and a decrease in regional summer solar insolation budget. This shows a clear link between the three records suggesting the onset of herb increase, indicative of climate cooling, corresponds with an Antarctic timing of cooling combined with a decrease in insolation budget, which

is potentially the forcing factor of climate change in the region. This is an important aspect of climatic change that should not be ignored or excluded from the NZ-I CES.

In this project Forks Lagoon and to a certain extent Howard Valley support an earlier onset of the LGCP. Other evidence in New Zealand also points towards an earlier glacial advance which began between 35-28 ka again suggestive of an earlier onset of the cold conditions. Although the CES is restricted to 30-8 ka it would appear that a temporal extension of the NZ-I project would be useful. The North Atlantic INTIMATE group has already extended the remit to 60 ka and recent discussions within the Australasian INTIMATE group suggest a similar development may occur. The results of this project support an extension of the temporal remit of the NZ-CES.

Figure 8.6: Plot of summer insolation for 40° S and 65° N (1)(Berger *et al*, 1991), 2) Vostock temperature curve (Petit *et al*, 1999), 3) Okarito pollen stratigraphy (Vandergoes *et al*, 2005). Figure adapted from Vandergoes *et al* (2005) Solid horizontal lines show a) showing the age of the onset of the LGCP used in the NZ-I CES. b) is a line drawn across the 35 ka marking where herb abundance begins to increase.

8.2.4.2 Region differences in reconstructions

The NZ-I CES combines records from the whole of New Zealand to produce a national CES. However, using records from different latitudinal locations and

different climate regions may not be the most plausible method of creating a CES as different regions experience different climate and oceanic patterns which can influence the signal recorded in each site. McGlone *et al* (2010) recognised different vegetation composition between different regions of New Zealand during the LGM. The vegetation reconstructions produced in this project from the three records during the LGCP suggest that regional vegetation differences are not just restricted to the LGM but continue throughout the whole LGCP. It would also be plausible that the same regional differences would extend to the LGIT and into the Holocene.

Newnham *et al* (2011) have noted a latitudinal variation between records in the timing and extent of temperature depression during the LGR. Okarito, which is the most southern record used in the NZ-I CES has the earliest recognisable LGR beginning at 14.5 ka and consisting of a 2 °C cooling, which persist for 1 ka. This cooling trend appears to be closely aligned with the ACR. Kaipo Bog, in the North Island provides evidence of a minor cooling episode between 13.6-12.6 ka and a MAT depression of 0.5-1 °C. Whilst Pukaki, the northernmost site reconstructs the smallest MAT decline of 0.5 °C commencing at 13 ka and is the shortest cooling period estimated in the three records. Newnham *et al* (2011) conclude that the southernmost site follows Southern Ocean records and an ACR timing (e.g., Pahnke *et al*, 2003). The northernmost site shows a weak reversal that closely aligned in timing with the YDC, whereas Kaipo Bog located in an upland North Island site shows an intermediate timing overlapping both the ACR and YDC. It was postulated that the location of the ocean fronts had an influence on these climate changes with Okarito located within the spheres of the Southern Ocean, whereas Pukaki and Kaipo Bog lie equatorward of the STF (Section 2.2.2). Research by Turney *et al* (2003) and Carter *et al* (2008) in the marine realm found a similar pattern for the same time period again suggesting latitudinal variations in climate signals.

Regional climate regimes may also have an important influence on the pattern of climate change recorded at different sites in New Zealand, and could possibly explain disparities between records. As touched upon in Section 2.3 and 7.6.3.1 the South Island can be separated into three climate regions which experience different climatic responses to the same synoptic conditions. For instance in the

contemporary environment during Zonal dominated regimes the ESI experiences drier conditions whilst the WSI experiences wetter conditions (Lorrey *et al*, 2007). Palaeorecords from these areas would record a different precipitation signal whilst experience the same synoptic conditions. If these regional differences were not taken into account then a different conclusion on the possible forcing factors of climate change may result. Section 8.2.2 and the pattern of precipitation and westerly wind intensity during the LGCP also points towards the importance of regional climate regimes. An understanding of each regional response to the same synoptic climate condition is important to prevent different conclusions on the main process influencing climate. Compiling a regional CES would enable comparisons between regions which may highlight this different regional response to the same synoptic conditions.

8.2.4.3 NZ-I CES Summary

The NZ-I CES provides a good start for synthesising key records from around New Zealand and producing a generalised stratigraphy. This can be used as a framework for comparisons, although there are some limitations with the scheme. Overall the timing of climate transitions and millennial scale events identified in this project occur broadly synchronously with the NZ-I CES (Table 8.1), particularly the LGM, Termination I and start of the Holocene. The records suggest an earlier onset for the LGCP and it is concluded that the temporal remit of the NZ-I project needs to extend further back in time. The limited records which extend into the LGCP and their relatively poor chronological control makes firm conclusions of the pattern of climate change in New Zealand for this period and the possible forcing factors more difficult to ascertain. However, the NZ-I CES provides a good foundation to build upon. It also highlights some periods of uncertainty, such as the timing and pattern of climate change during the LGCP milder interval and the LGR.

Since the NZ-I CES was initially produced by Alloway *et al* (2007), significant advances have been made which would prove useful in the CES. These advances include the production of the New Zealand pollen/climate transfer function (Wilmshurst *et al*, 2007), climate modelling of past New Zealand climates (Rojas *et al*,

2009), broader understanding of regional climate regimes and the possible influences these had on past climates (Lorrey *et al*, 2007) and the improvement of existing and addition of new proxy records (e.g. Whittaker *et al*, 2011; Williams *et al*, 2010; Augustinus *et al*, 2011). Utilising these advances to make improvements to the NZ-I CES are important in its development. However, the limitations discussed above should also be addressed to help refine the NZ-I CES. This includes extending the remit of the NZ-I CES and taking regional differences into account when comparing climate reconstructions between sites.

8.2.5 Quantitative pollen-climate reconstruction

Quantitative pollen-climate modelling using the New Zealand pre-deforestation database is a notable development towards a robust quantitative method for predicting past MAT changes (Wilmshurst *et al*, 2007). It could also be a useful aid to define boundaries of climate change in an objective manner (refer to Section 8.2.4.2). This technique has been applied to all four pollen records in this project providing interesting and insightful MAT reconstructions. However, as noted in each site chapter (Chapter 4-7), there are discrepancies with estimated MAT depression based upon ELR. This section will discuss the benefits and limitations of the quantitative pollen-climate modelling in light of the MAT reconstruction presented in this project. Suggestions of how this effort might be advanced are included.

8.2.5.1 ELR vs. the modelled MAT depression estimates

Table 8.2 compares the estimated MAT depression of each site using the three methods: PLS-C3, W-MAT and ELR. Due to the low pollen counts at Forks Lagoon ELR estimates were not calculated. However, the pollen data were used in the climate calibration as this performed a more objective temperature reconstruction than the qualitative ELR would have. Nevertheless, it should be noted that there are limitations in using the Forks Lagoon pollen record for reconstructing past climates, due to the low pollen count. At Howard Valley the PLS-C3 model reconstruction shows MAT was progressively warming between 37.7-18.9 ka, with the youngest sediments deposited during the LGM showing the warmest climates. The W-MAT model shows no change in MAT during this period. In contrast, the estimated MAT

using ELR predicts increased cooling with the LGM experiencing the coldest conditions. At Howard Valley and Forks Lagoon the modelled estimated LGM temperatures were lower than expected, although the PLS-C3 produces the coldest range of temperatures, which are broadly similar with the ELR estimates at Howard Valley.

The Lake Mudgie record produces a similar pattern of estimated MAT in all three methods although PLS-C3 shows the greatest disparity to ELR. The LGCP is split into two periods; a milder period between 26.6-23.2 ka and the colder LGM between 23.2-18.4 ka. Climate amelioration began at 18.4 ka with the W-MAT and ELR showing a period of warming interrupted by a slight cooling event followed by maximum MAT during the Holocene. PLS-C3 differs from the other methods because it estimates the highest temperatures occurring during the early LGIT. Temperatures slowly declined until 13.4-11.9 ka when a large temperature depression is estimated. This is followed by a period of warming although the MAT remains lower during the Holocene period than at the start of deglaciation. The Manks Tarn estimated MAT depression using ELR shows a consistent increase in temperature throughout the LGIT which is similar to the Lake Mudgie W-MAT and ELR temperature reconstructions. However, the modelled results are similar to Lake Mudgie PLS-C3 reconstructions and show MATs that are continuously declining with the largest temperature depression at 9.7 ka. This is followed by an increase in MAT with the warmest temperatures towards the top of the core. Like the Howard Valley record, both modelled results for Manks Tarns and the PLS-C3 model for Lake Mudgie do not correspond well with the inferred climate change produced in this project and from evidence elsewhere in New Zealand (see Section 8.1.5 and 8.1.5.2). This suggests the modelled results are problematic and is discussed in Sections 8.2.5.2 and 8.2.5.3. However, before the transfer function and Modern Analogue Technique can be discussed the method of ELR should first be mentioned.

As already discussed in Section 3.2.6 there are some limitations of ELRs, including temporal and spatial variation in the rate caused by a number of factors. A mean annual ELR of 0.5 °C/100 m, although a national average, may not be wholly representative of the sites in this project and may account for some discrepancies

between the modelled results and ELR estimates. On long timescales, large changes in climatic conditions (and in particular moisture/humidity) may have a significant influence on the ELR at a site. Using a variable ELR would probably produce a more realistic MAT. However, limited knowledge of past climatic conditions makes it difficult to modify the ELR for particular periods of time. The current national average ELR could possibly be underestimating or overestimating MAT depressions during the late Quaternary. However, until suitable alternative using national annual ELRs is the best option.

Period	Howard Valley (°C)			Forks Lagoon (°C)			Lake Mudgie (°C)			Manks Tarn (°C)		
	MAT today (9 °C)			MAT today (8.8 °C)			MAT today (11.5 °C)			MAT today (11.5°C)		
	PSL-C3	W-MAT	ELR	PSL-C3	W-MAT	ELR	PSL-C3	W-MAT	ELR	PSL-C3	W-MAT	ELR
Late MIS 3	4.2-5.5 (5)	1.7-1.9 (1.8)	2-3	-	-	-	-	-	-	-	-	-
LGCP 'milder interval'	3.3-4.8 (4)	1.6-2 (1.7)	3-4.5	1.4-4.4 (2.3)	1.3-1.5 (1.4)	-	4.5-5.6 (5.2)	3.2-4.4 (3.8)	>5.3	-	-	-
LGM	2.4-4.2 (3.3)	1.5-1.8 (1.7)	3->4.5	1.6-2.4 (2.1)	1.3-1.5 (1.4)	-	5.5-6.1 (5.3)	3.8-4.2 (3.6)	5.3	-	-	-
LGIT early	-	-	-	-	-	-	0.7-5.2 (1.8)	1.3-3.4 (1.5)	3.3-5.3	1.3-3.1 (2.3)	1.3-2.1 (1.7)	3.4-5.4
LGIT late	-	-	-	-	-	-	1.4-2.8 (2.2)	1.5-1.9 (1.7)	2.3-3.3	2.4-3.9 (3.2)	1-2.3 (2)	2.4-3.4
LGR	-	-	-	-	-	-	2.1-4.8 (3.6)	1.3-2.2 (1.9)	?	3-4.4 (3.7)	1.2-2.1 (3.7)	-
Holocene	-	-	-	-	-	-	1.9-3 (2.3)	0.2-1.6 (0.7)	0-2.3	1.7-3.5 (2.4)	0.1-1.9 (0.8)	0-2.4

Table 8.2: MAT depression range at each site inferred from the quantitative pollen/temperature model output and estimations by ELR and treeline depressions. The average MAT depression for each time frame is bracketed for the modelled results. The contemporary MAT for each site is added below the site name.

8.2.5.2 Howard Valley and modelled MAT

As mentioned previously there is a large discrepancy in the pattern of temperature change at Howard Valley between the modelled MAT and ELR estimated MAT. The qualitative methods' inferred climate change at the site is characterised by consistent cooling with the coldest conditions during the LGM. In contrast the PLS-C3 reconstruction shows the opposite pattern. Figure 8.7 shows the abundances of *Phyllocladus alpinus* and *Fuscospora* exhibit a nearly consistent inverse relationship with the PLS-C3 reconstruction, suggesting the abundance of these species has a significant influence over the predicted MAT reconstruction. In the pollen training set used in the pollen transfer functions, both *Phyllocladus alpinus* and *Fuscospora* are prevalent in colder conditions. Therefore, in the Howard Valley record when the abundance of these species declined up the sections the models inferred warming opposed to cooling, which was inferred by the vegetation reconstruction and ELR. The sparse training set for alpine vegetation could be responsible for the low predicted temperature depressions during the LGM. As the environmental gradient is not fully represented in the training set, this would limit the effectiveness of the transfer function and modern analogue technique. The dominant influence of a few taxa on the transfer function may also outweigh the influence of other taxa so that the models respond to changes of these dominant taxa as opposed to the assemblage as a whole.

Another factor which could be limiting the temperature reconstruction is the exclusion of taxa which could provide important insight to past climates in the vegetation reconstructions. This is due to the broad environmental range that some taxa represent. In this instance, Howard Valley contains a large proportion of Apiaceae, which is inferred to represent part of the alpine herbland community and is present in the youngest section at the site. However, due to the poorly modelled relationship this taxon has with temperature it is not included in the training set. Therefore a large proportion of the vegetation data cannot be applied. Better subdivision of these broad taxonomic groups to genus, subgenus or species levels, if possible, could help improve the models by enhancing the training set and providing more appropriate climate analogues than currently exist.

Figure 8.7: Comparison between the PLS-C3 MAT reconstruction and a) *Phyllocladus alpinus* and b) *Fuscospora* pollen abundance at Howard Valley. The samples have been placed in age order with the younger deposits at the top. The order is based upon the assumption that timing of sediment deposition is in stratigraphic order.

8.2.5.3 MAT and MAT depressions during the LGCP

The vegetation composition of the alpine herb/grassland recorded during the LGCP and estimated temperatures from quantitative pollen/temperature reconstructions are similar between sites. However, Lake Mudgie is a lowland site and has a modern MAT of 11.5 °C whilst Forks Lagoon and Howard Valley are at higher elevations and have lower MAT (8.8 °C and 9 °C respectively). When the estimated MAT for the LGCP is subtracted from the contemporary MAT at the sites, the temperature depression is less at the sites located at the higher altitudes in comparison to the lowland sites due to the former having considerably lower contemporary MAT. The training set produced by Wilmshurst *et al* (2007) contains sites reaching a maximum of 1540 m.a.s.l, but the cold alpine zone is still underrepresented which could lead to the MAT and therefore temperature depression during the LGCP to be underestimated.

The MAT reconstructions produced by Wilmshurst *et al* (2007) suggest that both models, but in particular the modern analogue technique, would overestimate the MAT of treeless periods in the records. Within the training set the only samples containing significant grassland and shrubland cover are from the south-eastern South Island. This region experiences moderate MATs, warm and dry summers and cold winters. Wilmshurst *et al* (2007) attributed the overestimated MAT reconstruction to the reliance of the samples from this region. This would have had greater significance on the W-MAT model, which methodologically is dependent upon the closeness of match with a set number of individual sites. A similar reason can be used to explain the overestimated MAT calculated for the LGCP in this project. A need to extend the spatial coverage of the training set is necessary before a reliable LGCP temperature reconstruction is to be achieved.

8.2.5.4 LGIT in Westland

The pattern of vegetation change and inferred temperature change using ELR suggest the LGIT was a period of near continuous climate amelioration with temperatures reaching a maximum at the beginning of the Holocene. A minor cooling episode may have occurred at Lake Mudgie between 13.4-11.9 ka but the amount of cooling cannot be quantified using ELR. This pattern of climate change is supported by a number of records from Westland and elsewhere in New Zealand (sees

Sections 8.1.5 and 8.1.5.1). The W-MAT temperature reconstruction for Lake Mudgie also shows a similar pattern of climate change. However, the other modelled results for both Lake Mudgie and Manks Tarn reconstruct a cooling pattern from initially high temperatures at the start of the LGIT to the start of the Holocene which is followed by increasing temperatures. At Lake Mudgie, the PLS-C3 model predicts higher MAT during the early LGIT than reached in the Holocene, which is unexpected. The pattern of LGIT temperature reconstructions for both sites points to further limitation of the current models which relate to the use of taxa spanning a wide temperature gradient and the use of a national training set opposed to a regional one.

ELR reconstructions are based upon regional vegetation characteristics whilst the transfer function uses a national training set. In some cases using a nationwide training set is useful because the training set covers a wide environmental gradient, but for others it would appear a regional approach may be more appropriate. A large vegetation component in both Lake Mudgie and Manks Tarn during the early LGIT is a mixed *Myrsine/Coprosma* shrubland. In combination these taxa make up 60-90% of the pollen assemblages. These taxa can grow in a wide range of habitats, from the lowland to alpine zone and in both the North and South Island of New Zealand covering a wide environmental gradient blurring the climate signal. In the Westland records the high MAT predicted at the onset of the LGIT can be attributed to the dominance of *Coprosma* and *Myrsine*. The subsequent decline in temperature during the LGIT is a factor of the decreasing abundance of these taxa up the cores. In this case the model's overestimation of MAT is largely due to applying climate envelopes for these taxa that are based on genus-level and nationwide pattern rather than on the actual species likely to be occurring in the region at the time. Regionally, *Coprosma/Myrsine* shrubland dominates the sub-alpine zone in the contemporary Westland environment and therefore are representative of cooler climates. Using a regional training set therefore may be helpful to reduce the wide environmental envelope that these taxa represent and thus produce a more realistic temperature reconstruction at the site.

8.2.5.5 W-MAT and PLS methodology

Deciding which pollen-climate model is preferable for this project was difficult. The statistical parameters suggest W-MAT performs slightly better than PLS-C3 although there was not much between the two models. However, complications arise from methodological limitations discussed above and from the pattern of results presented in Chapters 4-7. The following discusses the differences in reconstruction and methodology between the two models.

For each record W-MAT consistently estimated higher MAT than PLS-C3. Wilmshurst *et al* (2007) noted a similar pattern for two records with modern analogue technique consistently producing 1 °C warmer MAT than the PLS during the LGIT. The authors attributed this to the underrepresentation of treeless sites in the training set. A similar explanation is proposed for this project especially for the LGCP where W-MAT appeared to consistently overestimate the MAT (Section 8.2.5.3). The W-MAT method is more susceptible to poor representation of modern analogues in the training set as reliable reconstruction is dependent on close matches between modern and fossil assemblages. Instead the closest modern analogues for the fossil assemblages result in warmer estimated MAT than expected. Therefore this method is limited during the treeless periods. This evaluation indicates that more treeless sites should be added to the training set.

The PLS-C3 model is also influenced by the lack of modern pollen sites representing the alpine and treeless environments. However, this method allows species and climate data from a number of different samples that are not necessarily the closest match but infer MAT closer to the expected temperature. However, the same benefits in methodological approach for the LGCP become a hindrance during periods which contain large abundance of broad climate ranging taxa such as the *Coprosma/Myrsine*-dominated LGIT. The method's ability to extract climate data from the whole training set results in the inclusion of temperature data that are not representative of the environment represented by the fossil pollen assemblage. This results in warmer (LGIT at Lake Mudgie and Manks Tarn: 8.2.5.4) and colder (late MIS 3 at Howard Valley: 8.2.5.3) MAT reconstructed. Under these circumstances W-MAT method restricts the climate data used to only the most closely related sites which may provide greater accuracy than PLS.

As discussed above both the modelled results and the methods used contain limitations and the reconstructions produced should be treated carefully. Deciding upon the most appropriate model is difficult as PLS-C3 performs better during the LGCP and W-MAT performs better during the LGIT. However, the estimated MAT and calculated MAT depression are not the only important aspect of the climate reconstruction, but the pattern of change also. Based on this criterion the W-MAT performs better, particularly during the LGIT.

8.2.5.6 Summary

Despite the limitations discussed in the previous section, the pollen-climate models have provided some useful information regarding the pattern of climate change at the sites. For instance, for Lake Mudgie both models support a LGR reversal which could only be tentatively suggested using qualitative analysis. Nevertheless, the limitations (and potential solutions) which have been discussed above appear to be significant and are not only applicable to this project but to New Zealand as a whole. There is a need for better identification, including analyst consistency, to allow for other species/taxa to be incorporated in the training set and so reducing the data which is currently being ignored in the models. An increased spatial resolution of modern sites, including colder environment vegetation analogue, needs to be achieved in the training set before accurate reconstruction of LGCP, including LGM, conditions can occur. Associated with this is the use of regional training sets at sites which can refine the climate signal from the vegetation assemblage that is relative for a specific site. Again an increase in spatial resolution would be essential before regional training sets can be applied. Finally, some taxa appear to strongly influence the temperature reconstruction, such as at Howard Valley and during the LGIT in Westland. More work is needed to down-weight the influence of some taxa which may currently be distorting climate reconstructions under certain conditions, especially in regions where there is low representation of the contemporary vegetation and associated environmental variables in the training set.

Chapter 9: Conclusion and Future Work

9.1 Conclusions

Pollen and sediment analysis from three regions in the South Island was undertaken to produce a vegetation and climate history for the area and test the robustness of the NZ-I CES. The project provides new data and a new precipitation proxy method; assesses the use and application of quantitative pollen-climate reconstructions; highlights the strengths and weaknesses of the NZ-I CES and suggests possible refinements.

- The pollen records contain orbital-scale and millennial-scale patterns of vegetation and climate change. The principle vegetation changes identified in the records produce similar results to studies conducted elsewhere in New Zealand. Five main vegetation and inferred climate changes are recorded between c.38-4 ka:
 1. Late MIS 3 (c. 38-34 ka): at Howard Valley the vegetation is characteristic of montane *Nothofagus menziesii*/*Fuscospora* forest implying a MAT depression between 2-3 °C.
 2. The LGCP is a period of variable climate with a two step cooling pattern evident in the pollen record and estimated in the quantitative pollen-climate reconstructions for Lake Mudgie
 - (a) LGCP (c.32 ka to 23-21 ka [boundary indeterminate or transitional]): vegetation is dominated by alpine grass and herbland which contains a significant abundance of tree and/or shrub taxa indicative of cold but milder climates than the LGM. MAT depression range between 5.3-3 °C.
 - (b) LGM (23-21 ka to c.19 ka): dominated by grass and herbland taxa and containing trace amounts of tree taxa. MAT depressions range >5.3-4.5 °C.

3. Termination I and LGIT (18.9 ka to 11.9-10.8 ka [boundary indeterminate or transitional]): initially vegetation is dominated by *Myrsine/Coprosma* shrubland developing into *Griselinia* dominated forest followed by *Weinmannia racemosa* dominated montane forest. The pattern implies consistent climate amelioration and MAT depression change from 5.3- 2.3 °C.
- Holocene (11.9-10.8 ka onwards): characterised by the dominance of *Dacrydium cupressinum* lowland forest at western sites
 - A new precipitation proxy is derived from the comparative abundance of the plant marophytes *Isoetes alpinus* and *Myriophyllum tryphyllum* at Forks Lagoon, ESI. This proxy suggests that the LGCP, including the LGM, was generally drier, but is interrupted by two wetter periods between 26-24 ka and at 21 ka. These wetter periods in ESI correspond with evidence of wetter conditions in WSI, decreased westerly winds, milder conditions and decreased glacial activity. Conversely, the intervening dry period in ESI coincides with dry periods in WSI, increased intensity westerly winds, colder temperatures and increased glacial activity. The relationships imply that dry conditions relate to stadials and wet periods with interstadials.
 - The patterns of climate change between western, eastern and northern South Island sites during the LGCP provide potential evidence of the influence of regional climate regimes. For example, at Forks Lagoon, evidence of drier conditions implies increased zonal flow and increased westerly wind intensity whereas evidence for wetter periods implies increased prevalence of a trough regime and reduced westerly wind intensity. In connection with regional climates, the precipitation pattern supports a northward migration of the Subtropical Front causing an increase in southern westerly wind intensity during stadials and brings colder waters around central and southern New Zealand.
 - Evidence for a late glacial reversal in Westland is ambiguous. On the one hand, the pollen-vegetation reconstructions imply no major disruption to progressive climate amelioration. On the other hand, quantitative pollen-

climate models show a MAT depression ranging 4.8-1.3 °C between 13.4-11.9 ka. During this interval some evidence of possible catchment disturbance is also evident at Lake Mudgie.

- Vegetation reconstructions based on pollen evidence at Howard Valley closely align with previously published vegetation interpretations based on beetle and plant macrofossil evidence. The three proxies complement one and imply a LGCP temperature depression of 3-4.5 °C. The NZ-I group uses the Kawakawa/Oruanui Tephra (KOT) age of 27.1 in the CES. However, the chronologies derived from Howard Valley and Forks Lagoon supports the younger 25.4 ka age for the KOT. It is important to reconsider the KOT age to allow accurate comparisons between records in New Zealand and elsewhere.
- The application of the quantitative pollen-climate models produced conflicting results. Limitations in the training set and the methodology used to transfer modern environmental data to fossil data are likely to be the main reasons. The main limitations in the training set are attributed to a preponderance of taxa with climate ranges and the limited or lack of modern analogues for some of the fossil assemblages. Expanding the modern training set by the addition of more high altitude sites would improve the reconstructions produced.
- The results from this project and the NZ-I CES tend to correlate well, in particular the timing of Termination I, the possible LGR and the Holocene onset. However, there are some disparities, particularly during the LGCP including the complex climate period which is currently over simplified in the NZ-I CES. It should be noted that the disparities in the LGCP records may in part be due to different ages used for the KOT to date key boundaries or events.

9.2 Future work

This section remarks on the limitations specific to this project and provides wider recommendations for future research in this field.

9.2.1 Project specific recommendations

- Lake Mudgie's chronology is constrained by only three AMS ages and one tephra age. The oldest AMS age appears to be too young and was excluded from the record. The stratigraphy at this site is therefore chronologically constrained by only three direct ages. It is clearly evident that more ages are required to produce a well constrained vegetation and climate reconstruction for the site.
- Statistical analysis of the chronology is needed to reduce arbitrary decisions regarding the ages. Age outliers in the datasets will be identified using sequence analysis and the OxCal program. This will refine the individual ages used in the age models.
- There are few precipitation records in New Zealand, particularly from the ESI and during the LGCP, making the Forks Lagoon marocphyte record unique. The palaeoprecipitation signal for this site could be further enhanced by increasing sampling resolution and extending the coverage beyond 31.4 ka and into the LGIT and Holocene period. This could identify whether variations in precipitation can be identified during other key time periods such as the LGIT.

9.2.2 Wider recommendations

- The aim of using the quantitative pollen-climate models was to produce an objective estimate of palaeo MAT. However, limitations to the training set hinder reliable reconstructions. Improvements to the training set to provide better coverage of alpine vegetation analogues in particular are necessary to provide more robust reconstructions for cold intervals.

- Environmental Lapse Rates (ELRs) provide a useful complementary tool to pollen-climate model reconstructions. However, these have often been applied based on generalised rates derived from large scale studies. More accurate reconstructions would be obtained from ELRs derived from the study region itself, particularly if these could be adjusted for possible changes in humidity over time.
- Analysis of modern climate emphasises the distinctive regional response to changing synoptic conditions. The research presented in this study further highlights the importance of taking a regional approach to understanding New Zealand's palaeoclimate record.

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Appendices

Appendix A

Appendix A: Lake Mudgie CONISS results on a depth scale with the sediment stratigraphy and pollen summary diagram.

Appendix B

Appendix B: Manks Tarn CONISS results on a depth scale with the sediment stratigraphy and pollen summary diagram.

Appendix G

Site and site code	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	Cl	Total
Howard Valley	77.4	0.19	12.6	1.25	0.14	1.03	4.03	3.15	0.22	95.43
H3/5/37-38	<i>0.76</i>	<i>0.12</i>	<i>0.46</i>	<i>0.22</i>	<i>0.07</i>	<i>0.25</i>	<i>0.27</i>	<i>0.43</i>	<i>0.06</i>	<i>4.42</i>
Howard Valley	77.3	0.17	12.7	1.21	0.14	1.09	4.16	3.09	0.21	94.52
H3/6/40-41	<i>0.33</i>	<i>0.11</i>	<i>0.19</i>	<i>0.18</i>	<i>0.04</i>	<i>0.15</i>	<i>0.28</i>	<i>0.31</i>	<i>0.06</i>	<i>1.46</i>
Howard Valley	77.6	0.14	12.6	1.19	0.14	1.03	4.05	3.07	0.22	94.33
H3/7/43-33	<i>0.45</i>	<i>0.07</i>	<i>0.27</i>	<i>0.22</i>	<i>0.05</i>	<i>0.17</i>	<i>0.20</i>	<i>0.29</i>	<i>0.09</i>	<i>2.90</i>
Forks Lagoon	77.9	0.15	12.4	1.20	0.13	1.03	3.88	3.12	0.23	96.08
0802/T3/52	<i>0.59</i>	<i>0.09</i>	<i>0.20</i>	<i>0.31</i>	<i>0.05</i>	<i>0.14</i>	<i>0.31</i>	<i>0.34</i>	<i>0.07</i>	<i>1.53</i>
Forks Lagoon	77.6	0.16	12.8	1.21	0.14	1.01	3.82	3.04	0.23	94.07
0802/T3/52.75	<i>0.48</i>	<i>0.07</i>	<i>0.29</i>	<i>0.21</i>	<i>0.06</i>	<i>0.17</i>	<i>0.35</i>	<i>0.27</i>	<i>0.11</i>	<i>1.69</i>
AT-331	77.4	0.15	12.6	1.19	0.14	1.08	4.06	3.08	0.21	94.71
(ODP site 1123)*	<i>0.84</i>	<i>0.07</i>	<i>0.45</i>	<i>0.25</i>	<i>0.05</i>	<i>0.17</i>	<i>0.50</i>	<i>0.33</i>	<i>0.08</i>	<i>3.98</i>
KOT**	77.2	0.16	12.8	1.25	0.14	1.10	4.06	2.97	0.20	94.98
	<i>1.04</i>	<i>0.11</i>	<i>0.77</i>	<i>0.18</i>	<i>0.06</i>	<i>0.22</i>	<i>0.38</i>	<i>0.27</i>	<i>0.06</i>	<i>2.29</i>

Appendix G. Glass geochemistry of mean major element compositions for the KOT layer located in Howard Valley and Forks Lagoon. Analyses normalised to 100% before calculation of the mean in bold and ± 2 standard deviations in weight % in italics. Total refers to the mean total of original electron probe microanalyses before normalisation. ** KOT from Irirangi section, 10 km south of Waioura. *and ** data from Allan *et al* (2008).