JOURNAL OF QUATERNARY SCIENCE (2007) **22**(5) 517–534
Copyright © 2007 John Wiley & Sons, Ltd.
Published online in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/jqs.1137

WILEY InterScience® DISCOVER SOMETHING GREAT

Vegetation and climate of Auckland, New Zealand, since ca. 32 000 cal. yr ago: support for an extended LGM

REWI M. NEWNHAM,[1]* DAVID J. LOWE,[2] TERESA GILES[3] and BRENT V. ALLOWAY[4]†
[1] School of Geography, University of Plymouth, Plymouth, UK
[2] Department of Earth and Ocean Sciences, University of Waikato, Hamilton, New Zealand
[3] The Renwick Centre, North Rocks, Sydney, Australia
[4] Institute of Geological and Nuclear Sciences, Lower Hutt, New Zealand

Newnham, R. M., Lowe, D. J., Giles, T. and Alloway, B. V. 2007. Vegetation and climate of Auckland, New Zealand, since ca. 32 000 cal. yr ago: support for an extended LGM. *J. Quaternary Sci.*, **Vol. 22** pp. 517–534. ISSN 0267-8179.

Received 25 January 2007; Revised 23 March 2007; Accepted 26 March 2007

ABSTRACT: Auckland occupies a climatically sensitive position close to a major biogeographic boundary in the southern mid-latitudes. A new pollen record from Kohuora maar crater, Auckland, displays vegetation and climatic changes for the past ca. 32 000 years. Of particular interest are the inferred climatic patterns for the first part of the interval, encompassing the Last Glacial Maximum (LGM). The Kohuora record corresponds closely with pollen records from other Auckland sites indicating that the patterns observed are at least regional in extent. It is also broadly consistent with a variety of palaeoenvironmental evidence from across New Zealand, including the glacial record from Westland, other palynological records from North Island, other palaeoecological records from the South Island, and aeolian quartz sequences from western North Island. These records show that glacial conditions prevailed across most, if not all, of New Zealand during the interval ca. 29–19 k cal. yr BP, longer and earlier than the LGM *sensu stricto*. We suggest that the term extended LGM (eLGM) may be more appropriate for the New Zealand region. Within this predominantly cold interval, the Auckland pollen records indicate a climatic amelioration for the interval ca. 26–24 k cal. yr BP, also consistent with other palaeocological data from Canterbury, that fall within a period of climate amelioration recognised between the first two eLGM glacial advances in Westland. We refer to this warming interval as the eLGM Interstadial. The ca. 27 k cal. yr BP Kawakawa/Oruanui tephra is instrumental in most of these inter-site comparisons and occurs after the first peak of eLGM cooling in a short-lived comparatively mild phase. A subsequent return to apparently colder climate in the Auckland records may indicate a volcanic cooling effect or, more likely, widespread landscape disturbance following this major eruption event. Strong correspondence between biotic responses, glacial fluctuations and aeolian quartz deposition linked to major shifts in strength and latitudinal extent of the southern westerlies suggest that both the eLGM and eLGM Interstadial may be more widely registered, at least across the Southern Ocean. Support for this assertion comes from parallel investigations in western and southernmost South America and isotopic and palaeoecological records from Southern Ocean marine cores. Recent reconstructions of the globally averaged ice-equivalent sea-level history are in line with this evidence from the Southern Hemisphere, suggesting that the eLGM may have a global registration. In light of these observations, we suggest a re-examination of the defined timing of the LGM along with renewed effort to establish climatic patterns during this period and understand their causes. Copyright © 2007 John Wiley & Sons, Ltd.

KEYWORDS: LGM; eLGM; palynology; climate change; Kawakawa/Oruanui tephra; maars; tephrochronology; Kohuora; Auckland; New Zealand.

Introduction

For several decades, considerable effort has been invested in investigation of environmental patterns during the Last Glacial Maximum (LGM), broadly defined as the last major cold period on earth. It is widely recognised that these patterns are important both for determining sensitivity of global

*Correspondence to: R. M. Newnham, School of Geography, University of Plymouth, Plymouth PL4 8AA, UK. E-mail: rnewnham@Plymouth.ac.uk
†Present Address: Aurelian Ecuador SA, Reina Victoria 1539 y Av. Colon, Edificio Banco de Guayaquil, Piso 13, Suite 1306, Quito, Ecuador.

environmental systems to change and for understanding processes of climate change (Mix *et al.*, 2001).This paper concerns the record of LGM environments from New Zealand in the mid-latitudes of the Southern Hemisphere, which appears to be partly out of step with records from the continental Northern Hemisphere as well as with presumed forcing factors. For example, a recent pollen record from Okarito (Fig. 1) in south Westland, New Zealand, shows some notable climate changes during this period that are more consistent with local (austral) insolation forcing than with the Milankovitch northern insolation 'pacemaker', including an early LGM onset and intra-LGM warming, (Vandergoes *et al.*, 2005; Newnham *et al.*, 2007a). A very similar pattern of LGM climate change was developed independently from a comprehensive review of glacial morphologic and stratigraphic evidence from the same region, one of the most prominent areas of mid-latitude Quaternary glaciation in the Southern Hemisphere (Suggate and Almond, 2005). Explanations for these climate patterns are reliant upon replication in other parts of the New Zealand region and further afield. Here we present a new pollen-climate record from Auckland, ca. 800 km and 6.5° of latitude to the north of Westland that shows a similar LGM structure but in a very different climatic and biogeographic setting. We show that the same detailed patterns are discernible

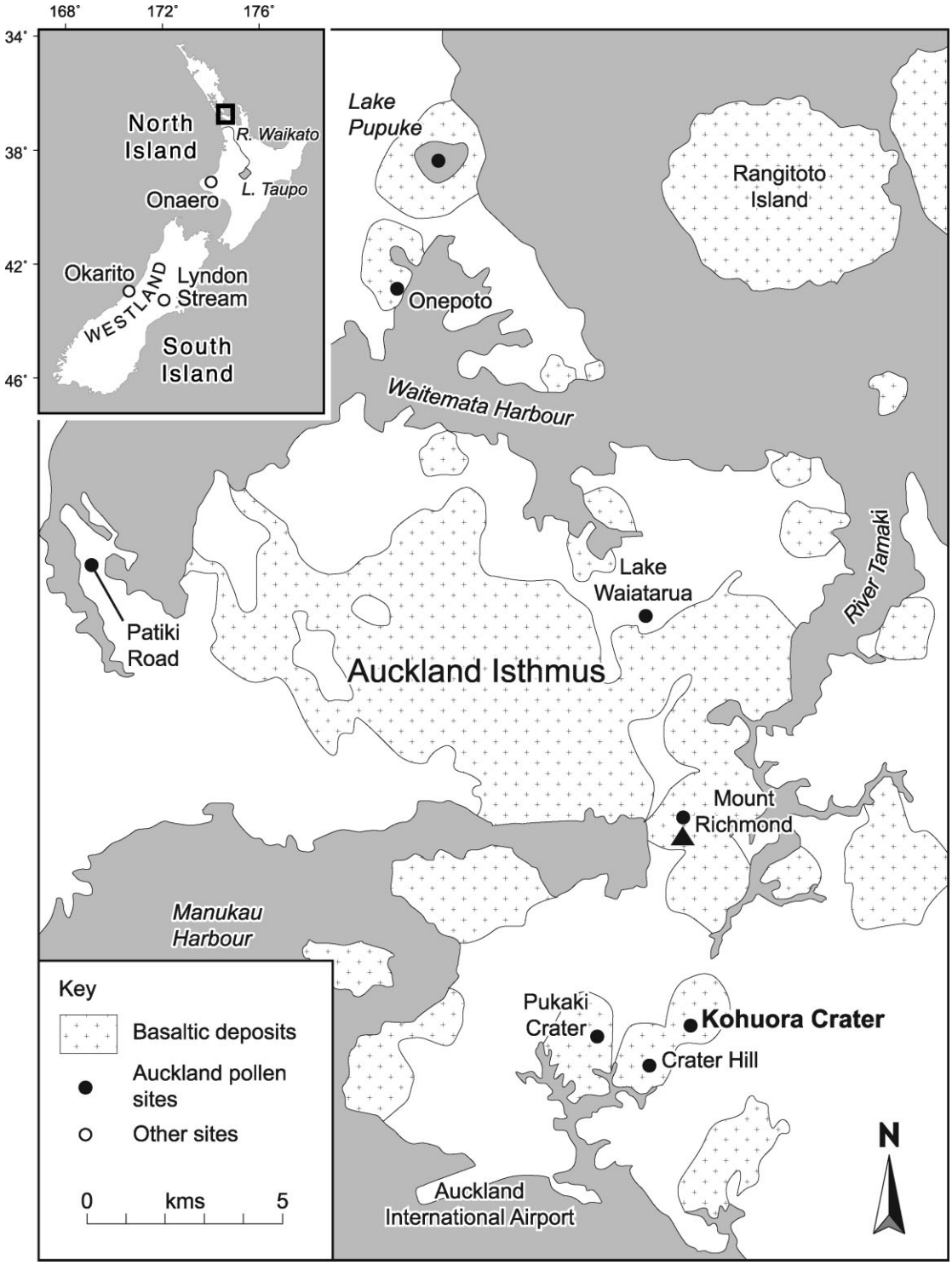

Figure 1 Auckland showing sites mentioned in the text. Inset: New Zealand with further sites discussed in the text

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

in previous pollen records from Auckland and in recently developed climate proxy records from elsewhere, suggesting they are likely to be representative of the wider New Zealand region. Although precise correlation is difficult because of broad chronological error margins, the widespread Kawakawa/Oruanui tephra (ca. 27 100 calibrated [cal.] yr BP; see Table 1) provides an unequivocal tie-point across many of these records. The new record presented here also provides further insight into vegetation dynamics under changing climatic regimes at a sensitive phytogeographic boundary today.

One outcome of this paper is to reinforce recent observations that the term LGM as currently defined is not directly appropriate to New Zealand. It is necessary therefore to use a different term to denote the glacial maximum in New Zealand, although we argue below that this should be linked in some way to the globally defined LGM. The current authoritative definition for the LGM is that recommended by the EPILOG (Environmental Processes of the Ice Age: Land, Oceans, Glaciers) group (Mix *et al.*, 2001). They redefined the LGM as 'the time of the most recent maximum in globally integrated ice volume' whilst recognising that it is unlikely that all ice sheets and glaciers on Earth reached a synchronous maximum. They critically reviewed methods for determining globally integrated ice volume and concluded that the best measure is based on the lowstand of sea-level caused by the lock-up of water on land-based ice sheets. Based on the best sea-level constraints available at the time, they defined the LGM chronozone as the interval between 19 000 and 23 000 cal. yr BP but suggested a second option of 18 000–24 000 cal. yr BP to better accommodate dating uncertainties (Mix *et al.*, 2001). The latter range has been adopted more widely.

Recent improved reconstructions of the LGM sea-level lowstand using both observations and geophysical modelling suggest that a review of this LGM definition may be required. Using sea-level data from several different regions, Lambeck *et al.* (2002) suggested that 'ice volumes approached their maximum values 30 000 (calendar) years ago and remained nearly constant until 19 000 years ago' and that 'this defines the period of maximum global glaciation'. More recently, Peltier and Fairbanks (2006), using an extended set of sea-level observations from Barbados together with a refined model of continental deglaciation, concluded that the LGM must have occurred 5000 years earlier than the assumed [EPILOG] age. Using the methodology proposed by EPILOG but applied to improved sea-level data it is clear that the LGM commenced significantly earlier than currently defined, and thus extended over a much longer period. This earlier, extended LGM is much more closely aligned with the palaeoclimatic evidence from New Zealand, although it has yet to be universally adopted by the global palaeoclimate community.

To avoid confusion with the current [EPILOG] definition of the LGM, hereafter in this paper we use the term 'extended Last Glacial Maximum' (eLGM) in relation to the New Zealand records and follow Suggate and Almond (2005) in suggesting this glacial phase occurred from ca. 29 000 to 19 000 cal. yr BP. We emphasise that, rather than showing incongruence with globally integrated sea-level evidence for the LGM, the eLGM is strongly coherent with the most recent determinations.

Unless stated otherwise, all ages in the text hereafter refer to calibrated ^{14}C ages and we use the abbreviation 'k' for thousands of years before present. Calibrated ^{14}C ages were determined using CalPal2005_SFCP (online) (Weninger *et al.*, 2005) based on Shackleton *et al.* (2004) after correction for Southern Hemisphere offset (-41 ± 14 yr; Hogg *et al.*, 2002). Note that beyond ca. 25 ka, 'calibrated' (more strictly 'comparison') ages remain imprecise as error terms may be

Table 1 Sources and selected properties of tephras in Kohuora crater sediments

Tephra (core and depth)[a]	Source[b]	Thickness (cm)	Colour and grain size in field	Ferromagnesian minerals[c]
Tuhua	Tuhua VC (MI)	3	Pale greenish grey coarse ash	Aeg
Rotoma	Okataina VC (HR)	1–3	Cream–white fine and medium bedded ash	Cgt
Uncorrelated AVF (KoA 475 cm)	Auckland VF	<0.5	Sparse medium ash disseminated in sediment	Olv, Cpx (mod. large crystals)
Rotorua	Okataina VC (OB)	2–3	White or cream fine and medium bedded ash (speckled black with free biotite crystals)[d]	Hbe, Bio, Hyp, Cpx
Rerewhakaaitu (?) and uncorrelated EgV	Okataina VC (TA) and Egmont	0.5	Pale mauve fine ash	Hbe, ?Cgt, Cpx
Kawakawa/Oruanui	Taupo VC	2	White fine to medium ash with sparse white pumiceous medium ash	Hyp, Cpx, Hbe
Uncorrelated AVF (KoB 716 cm)	Auckland VF	<0.5	Very sparse ash grains disseminated in sediment	Olv (mod. large crystals)
Uncorrelated AVF (KoB 735 cm)	Auckland VF	<0.5	Very sparse pumiceous coarse ash grains disseminated in sediment	Cpx, Olv, Hbe (large crystals)

[a] See Fig. 2. Tephra names from Froggatt and Lowe (1990); EgV, uncorrelated andesitic tephra derived from Egmont volcano. Kawakawa Tephra as defined by Froggatt and Lowe (1990) comprises Aokautere Ash (ash-fall) and Oruanui Ignimbrite (pyroclastic density-current deposits) members. However, the eruptives are also known collectively as Oruanui deposits as defined by Wilson (2001) (after Vucetich and Pullar, 1969). Wilson (2001) divided them into ten phases on the basis of nine mappable fall units and a tenth, poorly preserved but volumetrically dominant fall unit, all interspersed with numerous pyroclastic density-current deposits. Consequently, we use the term Kawakawa/Oruanui, following Alloway *et al.* (2007) and Lowe *et al.* (2007).

[b] Source volcanoes in named volcanic centre (VC) or volcanic field (VF): MI, Mayor Island or Tuhua; HR, Haroharo linear vent zone; OB, Okareka basin/embayment; TA, Tarawera linear vent zone.

[c] Undertaken using a petrologic microscope; sparse amounts precluded quantification. Most crystals fine-grained except as noted for proximal eruptives. Ferromagnesian minerals noted as present: Aeg, aegirine; Cgt, cummingtonite; Olv, olivine; Cpx, clinopyroxene; Hbe, hornblende; Bio, biotite; Hyp, hypersthene.

[d] In core B, ash within the lake sediment is disseminated above the visible layer over a distance of about 10 cm.

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

substantial and, moreover, differences of up to ca. 1000 years may result simply by using different comparison curves and datasets (see van der Plicht *et al.*, 2004; Lowe *et al.*, 2007).

Geophysical, biogeographical and climatic setting of Auckland

Auckland (latitude approximately 37° S), New Zealand's largest city (population about 1.4 million), is situated within a distinctive geophysical and biogeographic setting that provides good opportunity to develop late Quaternary terrestrial records of climate change. The city straddles a narrow isthmus across a landscape dominated by the products of late Quaternary basaltic volcanism. The Auckland Volcanic Field , active in part since ca. 180–200 ka (Allen and Smith, 1994; Cassidy *et al.*, 1999; Marra *et al.*, 2006a), comprises about 50 small, mainly monogenetic tuff and scoria cones and maar craters associated with a veneer of volcanic deposits (Fig. 1; Kermode, 1992). Local volcanism has facilitated palaeoenvironmental investigation of this region by creating numerous depositional situations for the accumulation of uninterrupted sedimentary sequences, some of considerable length (Sandiford *et al.*, 2001, 2003; Shane and Sandiford, 2003; Pepper *et al.*, 2004). Some depositional basins have formed in the crater depressions of volcanic landforms, whereas others have been created through lava blocking drainage routes (Newnham and Lowe, 1991). In many of these basins, lakes or swamps have formed, providing anaerobic conditions suitable for palaeoecological investigation.

The products of Quaternary volcanism also provide a mechanism for developing chronological frameworks and an unequivocal way to correlate between sedimentary sequences from separate sites. Although local basaltic volcanism imposes some limitations (Shane and Smith, 2000), the recent discovery in Auckland of a suite of silicic tephras of non-local provenance (Newnham *et al.*, 1999a; Shane and Hoverd, 2002; Alloway *et al.*, 2004) has created a new opportunity to apply the advanced tephrochronologic framework established for the central North Island (Froggatt and Lowe, 1990; Lowe *et al.*, 1999, 2007) to the Auckland area.

The Auckland climate is windy, with mild winters and warm humid summers, reflecting its maritime mid-latitude setting. Air temperature means range from 11°C in mid-winter to 21°C in mid-summer and averages ~16°C annually with an average diurnal range of ~6°C (New Zealand Meteorological Service, 1980; Tomlinson and Sansom, 1994a). Snowfall is exceedingly rare. Around 2100 hours of bright sunshine are received annually, about 50% of the total possible. Rainfall is generally moderate, but variable; mean annual rainfall in recent decades has been 1119 mm (Tomlinson and Sansom, 1994b), but in many years may exceed 2000 mm in the nearby Waitakere and Hunua hills. Apart from this orographic effect, topography exerts only a minor influence on the climate of Auckland, although in autumn and winter, cold air drainage from the slopes of the volcanic cones or nearby hills may produce local frost pockets. The prevailing windflow is from the west or southwest reflecting the prevalence of the southern westerlies across most of the Southern Ocean from the Subtropical Front to the Antarctic Trough. This zonal airflow extends into lower latitudes during the winter and thus has a direct seasonal influence on the Auckland climate. In summer, subtropical airstreams from the north and northeast typically exert a stronger influence on Auckland's climate. These seasonal patterns are in part modulated by the El Nino-Southern Oscillation cycles which typically result in strengthened westerlies relative to subtropical airstreams during El Nino phases and the reverse during the La Nina phases. During the eLGM, the belt of westerly airflow shifted northwards by at least 5° latitude (Nelson *et al.*, 1993; Wright *et al.*, 1995) and therefore is likely to have dominated Auckland climate for much more of the annual cycle than is the case today.

The natural vegetation of Auckland that remains today has been depleted and diminished relative to the natural extent prior to anthropogenic modification. This modification occurred in two phases, beginning with permanent Polynesian settlement in the area (after ca. AD 1300), and then extended by European arrival in the colonial era (Hayward *et al.*, 2006). Despite the relatively recent transformation of the Auckland landscape, isolated remnants of natural vegetation together with Holocene palynological records indicate that the region supported extensive conifer-angiosperm forest with some distinctive communities characteristic of volcanic landscapes (Cranwell, 1981; Newnham and Lowe, 1991). These remnants and records include several species that are confined to the northern North Island phytogeographic province of Cockayne (1928) with natural distributions that do not extend south of 38–39° S. This 'northern' warm temperate group includes *Agathis australis* (kauri, New Zealand's largest tree), *Beilschmiedia taraire, Vitex lucens, Metrosideros excelsa, Halocarpus kirkii, Ixerba brexioides, Weinmannia silvicola* and *Ackama rosaefolia*. Auckland's forests also include several woody species which are absent from the southern half of North Island, but which reappear in northern South Island, e.g. *Libocedrus plumosa, Phyllocladus trichomanoides, Quintinia serrata*. Conversely, a number of common 'southern' cool-temperate species attain their northern limits just to the south of Auckland, usually at the highest altitudes. This group includes *Libocedrus bidwillii, Nothofagus menziesii, Phyllocladus aspleniifolius, Halocarpus bidwillii*, and *Podocarpus nivalis*. Beech forests, which characterise southern and montane areas in New Zealand, are generally absent from Auckland northwards, with the exception of isolated stands of *Nothofagus truncata* and *N. solandri* var. *solandri*.

Auckland therefore sits astride an important modern phytogeographic boundary between these major northern and southern components of the New Zealand forest flora. On the basis of climatic inferences drawn from these biogeographic patterns, several authors have suggested that environmental changes north of Auckland during Quaternary climatic oscillations were likely to have been less severe than for all other regions of New Zealand. Fleming (1979) argued that Pleistocene cooling north of Auckland (Northland) was probably not sufficient to eliminate forest, although the composition of forests may have varied. Others (e.g. Wardle, 1963; Clayton-Greene, 1978) have argued that the Northland region may have acted as a refugium for warm-temperate species during glacial stages. Consistent with these views are palynological records from Northland indicating survival of forest, albeit with expansion of beech and other compositional changes, during the last glacial (Elliott, 1998; Newnham, 1999; Newnham *et al.*, 2004; Elliott *et al.*, 2005; Horrocks *et al.*, 2007; see also Wright *et al.*, 1995). This situation contrasts strongly with records from sites south of Auckland, where, apart from at sheltered coastal localities (Lees *et al.*, 1998), near-complete forest cover during warm (interglacial) stages was reduced to grassland or shrubland with at most patchy forest during cold (glacial) stages (McGlone, 1988; Newnham *et al.*, 1989, 1999b). Previous pollen records from Auckland imply a Quaternary history intermediate between the patterns recorded to the north and to the south (Sandiford *et al.*, 2003; Byrami *et al.*, 2005). Thus the modern biogeographic boundary straddled by Auckland also marks an important distinction between regions with contrasting histories. Crucially, however,

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

prior to this study, no previous pollen record from this pivotal region continuously covers the last glacial–interglacial cycle to provide a definitive record of vegetation change at one locality.

Kohuora volcano study site

Kohuora volcano, a small tuff ring-crater complex, is located in Papatoetoe in Manukau City, southern Auckland (Fig. 1). Together with nearby Crater Hill and Pukaki, it completes the small Papatoetoe volcanic group where activity was predominantly explosive and ranged from wholly phreatic to phreatomagmatic, generating scoria, basaltic ash and lapilli, and lithic tuff (small amounts of lava were erupted from Crater Hill) (Searle, 1981; Kermode, 1992). Lithic tuff deposits, as mapped at Kohuora, comprise thin graded beds of grey mud- to sand-sized fragments of comminuted country rock (mainly sandstone and mudstone chiefly of Miocene age and in places calcareous) together with basalt and basanite fragments (Kermode, 1992). The Kohuora crater floor is 5 m a.s.l., with the top of the tuff ring reaching 33 m a.s.l. No scoria deposits have been observed. The only previous information constraining the age of Kohuora volcano are two ^{14}C ages obtained on wood within Kohuora tuff: (NZ540) 29 000 ± 700 ^{14}C yr BP (33 681 ± 1141 cal. yr BP) and (NZ488) 29 700 ± 2000 ^{14}C yr BP (34 530 ± 2255 cal. yr BP) (Grant-Taylor and Rafter, 1971). The pollen and sedimentary record presented here indicates that a shallow lake (maar) developed in Kohuora crater, which eventually became infilled and the subsequent succession of wetland vegetation gave rise to peat accumulation during the mid-late Holocene (Fig. 2). The lacustrine and peat deposits extend beyond 9.5 m depth. The wetland area within Kohuora crater has undergone extensive disturbance in recent years owing to local residential development and industrial waste disposal. Apart from occasional patches of relatively undisturbed wetland, nothing remains of former vegetation. Nevertheless, the New Zealand Department of Conservation designated Kohuora as a protected site for palaeoenvironmental research in 1995 in recognition of the scientific importance of this site, and the wetland was acquired by Manukau City.

Kawakawa/Oruanui tephra

The caldera-forming Kawakawa or Oruanui eruption of Taupo volcano, one of the largest volcanic events of the late Quaternary (Wilson *et al.*, 2006), resulted in the widespread distribution of an important chronostratigraphic marker for eLGM sequences in New Zealand (Pillans *et al.*, 1993). Several different names have been applied to the various pyroclastic products of this eruption (Table 1). Here we follow recent usage by Alloway *et al.* (2007) and Lowe *et al.* (2007) and use the double name Kawakawa/Oruanui tephra. We use the calibrated age of ca. 27 100 cal. yr BP (Table 2), which was derived from ^{14}C age determinations on small carbonised branch fragments pyrolysed within the encasing pyroclastic density current (ignimbrite) deposits (Wilson *et al.*, 1988; Lowe *et al.*, 2007).

Methods

Stratigraphy and chronology of Kohuora cores

Two sediment cores were taken from Kohuora crater for this study (Fig. 2). The cores can be correlated on the basis of common tephra layers and radiocarbon ages. Core A was taken from the centre of the crater basin in January 1995, using a hand-held 5 cm width Russian peat corer, and provides a 9.5 m sequence spanning from ca. 17.6 ka to near present. Eleven ^{14}C dates were obtained for it, and six tephra layers were recognised or identified (Newnham *et al.*, 1999a; Fig. 2). Core B was taken from a former landfill site on the southern flanks of the tuff ring in November 1996, using a piston corer powered from a mobile drilling rig operated by ProDrill (Auckland). The ~10 cm diameter core, extracted in 1 m lengths, extended to a depth of ~7.5 m depth at which point contact with the underlying tuff deposits was reached. Above this contact, ~1.6 m of olive-grey to dark brown muds and silts, with four identified tephra layers, were retained for this study. Overlying this *in situ* organic sediment were ~4.4 m of typically dark brown lake sediment grading upwards into dark brown peaty sediment and peat to a depth of 1.5 m, at which point no core had been obtainable (too sloppy). Tephrochronology and six radiocarbon dates show that the *in situ* sediments from this core extend from the underlying tuff ring, formed ca. 33 ka, to the Tuhua Tephra (ca. 7 ka). Analytical results from the younger sequence (Core A) were presented by Giles (1999). This paper primarily concerns the analyses undertaken on the lower part of Core B from the base upwards to the Rotorua Tephra (ca. 15.5 ka).

Core B contains in total six visible (macroscopic) tephra layers. Four of these, Kawakawa/Oruanui, Rotorua, Rotoma and Tuhua tephras, were identified through major element analysis (by electron microprobe) of constituent glass shards

Table 2 Ages on tephra layers identified in Kohuora cores

Tephra	Core/depth (cm)	Composite depth (cm)	Age ^{14}C yr BP (±1 sd)	Cal.age cal.yr BP (±2 sd)[c]	Cal.age cal.yr BP (±1 sd)[d]
Tuhua	KoA/384	384	6130 ± 30[a]	7005 ± 155	
Rotoma	KoA/442	444	8530 ± 10[a]	9505 ± 25	
Rotorua	KoA/781	782	13 080 ± 50[a]	15 425 ± 325	
Rotorua	KoB/622	782	13 080 ± 50[a]	15 425 ± 325	
Rerewhakaaitu (?)	KoA/948	950	14 700 ± 95[a]	17 626 ± 425	
Kawakawa/Oruanui	KoB/711	871	22 590 ± 230[b]	27 097 ± 957	27 098 ± 586

[a] From Hajdas *et al.* (2006).
[b] From Wilson *et al.* (1988) and Lowe *et al.* (2007).
[c] Calibrated by wiggle-match on peat via OxCal using IntCal04 (Hajdas *et al.*, 2006) except for Kawakawa/Oruanui tephra which was calibrated via OxCal using Hughen *et al.* (2006) after correction for Southern Hemisphere offset (−41 ± 14 yr; Hogg *et al.*, 2002) (2σ range, 95.3%).
[d] Calibrated via online CalPal2005_SFCP (Weninger *et al.*, 2005; accessed January 2007) using Shackleton *et al.* (2004) after correction for Southern Hemisphere offset (−41 ± 14 yr; Hogg *et al.*, 2002) (1σ range, 68.2%).

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

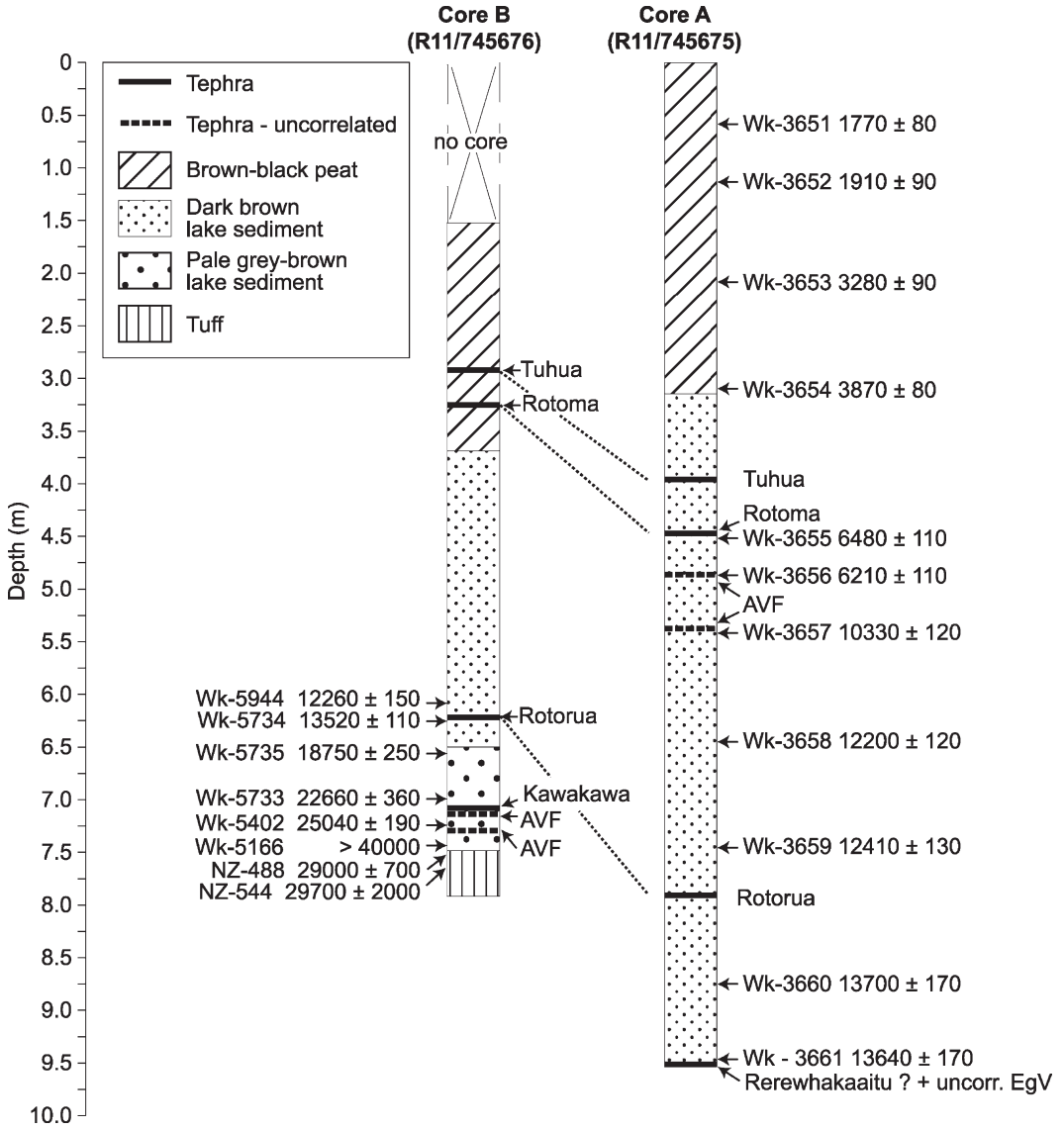

Figure 2 Stratigraphic columns for Kohuora crater cores A and B showing sediment stratigraphy, tephrostratigraphy, and sampling points and laboratory numbers for ^{14}C dates (NZ-544 was additionally obtained from Kohuora tuff) (see Table 4). Dashed lines connect identified tephra layers observed in both cores

(Table 3) supplemented with qualitative ferromagnesian mineralogical identifications (Table 1). The correlations were facilitated primarily by comparison of the glass major element compositions with recently published analyses on glass reported by Smith *et al.* (2005) and Lowe *et al.* (2007). The other two tephras, erupted just before Kawakawa/Oruanui tephra, are basaltic and, on the basis of their olivine- and clinopyroxene-dominated ferromagnesian mineralogy and coarse grain size, were derived from the local Auckland Volcanic Field (Fig. 1). In addition to ages provided by correlation of the identified tephra layers (Lowe *et al.*, 2007), six bulk organic sediment samples (2 cm thick or 3 cm thick slices) were extracted for radiocarbon dating at the University of Waikato Radiocarbon Dating Laboratory (Fig. 2; Table 4). An age–depth model is plotted in Fig. 3. The record appears to be continuous with no apparent sedimentary breaks, except for the deposition of the tephra layers. Kawakawa/Oruanui tephra occurs as a 2 cm thick, white, fine–medium ash at 710–712 cm depth in Core B (Fig. 2; Table 1). A 2 cm thick bulk sediment sample from immediately above, and contiguous with, the tephra was radiocarbon dated as 22 660 ± 360 ^{14}C yr BP (Wk-5733), which is statistically identical to the established ^{14}C age for the tephra (Lowe *et al.*, 2007) (Tables 2 and 4).

The basal radiocarbon date, on olive-grey mud, of >40 000 ^{14}C yr BP (Wk-5166) is seemingly at odds with the previously determined age for Kohuora crater of ca. 33 ka (Grant-Taylor and Rafter, 1971) and with the overlying ^{14}C ages presented here (Fig. 3). This apparent discrepancy and the infinite age for sample Wk-5166 may be explained by the presence in the dated sample of comminuted pre-volcanic Miocene-aged shell material brought to the surface during the eruption and formation of the lithic tuff ring (e.g. Kermode, 1992). We estimated an age of ca. 32 ka for the base of the sedimentary sequence by extrapolation from the overlying calibrated ^{14}C ages (Fig. 3). This estimate is consistent with the previous estimates by Grant-Taylor and Rafter (1971) for the age of Kohuora crater's formation if allowance is made for a small lag between the eruption and commencement of the overlying sedimentary sequence, the strong possibility that the wood dated by Grant-Taylor and Rafter included an inbuilt age component, the very wide laboratory counting errors associated with the wood dates, and their uncertain calibration.

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

Table 3 Electron microprobe analyses[a] of glass from tephras in Kohuora crater sediments

Tephra	Tuhua		Rotoma		Rotorua			Rerewhakaaitu (?) + EgV		Kawakawa/Oruanui
Core-subpopulation	KoA	KoB	KoA	KoB	KoA-T1 (low K)[c]	KoB-T1 (low K)[c]	KoB-T2 (high K)[c]	KoA-T1 (medium K)[c]	KoA-EgV[b]	KoB
Depth (m)	3.84–3.87	2.91–2.94	4.44–4.47	3.60–3.61	7.81–7.83	6.19–6.22		9.48–9.49		7.10–7.12
SiO_2	74.01 (0.26)	74.25 (0.51)	77.76 (0.34)	77.10 (0.80)	76.47 (0.33)	76.00 (1.15)	77.08 (0.64)	77.01 (0.24)	61.22 (0.43)	77.31 (0.90)
Al_2O_3	0.33 (0.05)	9.63 (0.17)	12.51 (0.24)	12.81 (0.58)	13.01 (0.14)	13.01 (0.24)	12.55 (0.39)	12.75 (0.10)	17.76 (0.23)	12.89 (0.17)
TiO_2	9.59 (0.10)	0.31 (0.07)	0.16 (0.05)	0.22 (0.07)	0.27 (0.05)	0.26 (0.08)	0.23 (0.09)	0.18 (0.05)	0.81 (0.05)	0.27 (0.08)
FeO[d]	5.82 (0.24)	5.66 (0.21)	0.93 (0.12)	1.00 (0.23)	1.38 (0.12)	1.43 (0.33)	1.18 (0.43)	1.12 (0.08)	4.72 (0.25)	1.24 (0.17)
MnO	0.35 (0.09)	na	0.23 (0.08)	na	0.25 (0.07)	0.19 (0.14)	0.19 (0.06)	0.18 (0.04)	0.36 (0.05)	na
MgO	0.06 (0.01)	0.04 (0.04)	0.14 (0.02)	0.10 (0.03)	0.27 (0.05)	0.22 (0.06)	0.10 (0.03)	0.16 (0.06)	1.65 (0.12)	0.16 (0.05)
CaO	0.26 (0.05)	0.27 (0.05)	0.81 (0.04)	0.86 (0.11)	1.31 (0.09)	1.29 (0.13)	0.75 (0.10)	0.92 (0.07)	3.49 (0.25)	1.14 (0.15)
Na_2O	5.11 (0.11)	5.38 (0.25)	3.93 (0.14)	4.17 (0.25)	3.84 (0.15)	4.07 (0.27)	3.35 (0.49)	3.83 (0.20)	5.07 (0.23)	3.39 (0.75)
K_2O	4.26 (0.13)	4.18 (0.15)	3.42 (0.11)	3.54 (0.40)	3.02 (0.17)	3.19 (0.29)	4.39 (0.26)	3.66 (0.27)	4.48 (0.26)	3.31 (0.33)
Cl	0.24 (0.03)	0.28 (0.07)	0.15 (0.04)	0.20 (0.02)	0.17 (0.03)	0.24 (0.08)	0.29 (0.07)	0.19 (0.04)	0.39 (0.14)	0.28 (0.03)
Water[e]	3.07 (1.07)	1.72 (1.75)	1.62 (0.96)	4.04 (3.69)	4.03 (1.11)	4.51 (1.52)	6.14 (2.60)	3.62 (0.62)	1.06 (1.44)	6.36 (1.00)
n	12	10	11	10	12	16	4	5	7	10

[a] Means and standard deviations (in parentheses) of *n* analyses (individual glass shards) normalised to 100% loss-free basis (wt%). Analyses by wavelength-dispersive Jeol JXA-733 Superprobe at the Analytical Facility, Victoria University of Wellington, undertaken by Drs Dave Manning and Paul Froggatt and the late Jeanette Gillespie using various glass and elemental standards to correct for machine drift, defocused beam diameter 20 μm, current 8 nA, and accelerating voltage 15 kV; analyses calculated from 11 × 2 s counts across the peak, curve integrated. na, not available.

[b] This andesitic tephra from Egmont volcano is possibly a correlative of sample AT203 in Pukaki (Sandiford *et al.*, 2001) and/or sample 133 in Onepoto (Shane and Hoverd, 2002) (see also Shane, 2005).

[c] T1 and T2 refer to separate magma types (early to late eruptive phases, respectively) as documented by Smith *et al.* (2005) and Lowe *et al.* (2007). That type T1, low in K_2O, predominates in Kohuora is consistent with findings by Sandiford *et al.* (2001), Shane and Hoverd (2002), and Shane *et al.* (2003) who demonstrated that type T1 glass was dispersed primarily to the northwest.

[d] Total Fe as FeO.

[e] Water by difference from original analytical total.

Copyright © 2007 John Wiley & Sons, Ltd.

Table 4 Radiocarbon ages obtained at Kohuora crater[a]

Laboratory no.[b]	Core/depth (cm)	Composite Depth (cm)	Sample material[c]	Age ^{14}C yr BP (1 sd)	δ^{13}C[d]	Age cal. yr BP OxCal (±2 sd)[f]	Age cal. yr BP CalPal (±1 sd)[g]
Wk-3651	KoA/40	40	Peat	1770 ± 80	−28.5	1615 ± 205	1655 ± 95
Wk-3652	KoA/100	100	Peat	1910 ± 90	−29.3	1775 ± 215	1806 ± 102
Wk-3653	KoA/200	200	Peat	3280 ± 90	−30.1	3465 ± 225	3485 ± 95
Wk-3654	KoA/300	300	Lake sed.	3870 ± 80	−30.5	4220 ± 230	4243 ± 124
Wk-3655	KoA/442	440	Lake sed.	6480 ± 110	−32.2	7300 ± 300	7355 ± 93
Wk-3656	KoA/470	480	Lake sed.	6210 ± 110	−30.9	7000 ± 300	7062 ± 138
Wk-3657	KoA/550	550	Lake sed.	10 330 ± 120	−27.9	12 200 ± 550	11 999 ± 479
Wk-3658	KoA/650	650	Lake sed.	12 200 ± 120	−26.4	14 200 ± 450	14 212 ± 279
Wk-3659	KoA/750	750	Lake sed.	12 410 ± 130	−25.2	14 500 ± 500	14 523 ± 365
Wk-3660	KoA/860	n/a	Lake sed.	13 700 ± 170	−25.2	16 350 ± 600	16 869 ± 370
Wk-3661	KoA/940	950	Lake sed.	13 640 ± 170	−25.8	16 250 ± 600	16 784 ± 408
Wk-5944	KoB/592	n/a	Lake sed.	12 260 ± 150	−28.0	14 300 ± 550	14 309 ± 301
Wk-5734	KoB/623	783	Lake sed.	13 520 ± 110	−30.0	16 100 ± 500	16 692 ± 411
Wk-5735	KoB/652	812	Lake sed.	18 750 ± 250	−24.3	22 150 ± 800	22 401 ± 409
Wk-5733	KoB/709	869	Lake sed.	22 660 ± 360	−26.0	27 316 ± 1200	27 178 ± 658
Wk-5402	KoB/733	893	Lake sed.	25 040 ± 190	−26.7	30 074 ± 310	29 948 ± 308
Wk-5166	KoB/748	898	Lake sed.	>40 000	−29.0	>40 000	>40 000
NZ540[e]		In tuff	Wood	29 000 ± 700	n/a	33 547 ± 1433	33 681 ± 1141
NZ488[e]		In tuff	Wood	29 700 ± 2000	n/a	35 924 ± 4952	34 530 ± 2255

[a] Ages calibrated using both OxCal and CalPal and different calibration/comparison datasets (see footnotes f and g).
[b] Wk-numbers, Waikato Radiocarbon Dating Laboratory, Hamilton, New Zealand; NZ-numbers, Rafter (New Zealand) Radiocarbon Dating Laboratory, Lower Hutt; n/a, not applicable or not available.
[c] All Wk- samples were washed in hot 10% HCl, rinsed and dried.
[d] All measurements ± 0.2‰.
[e] From Grant-Taylor and Rafter (1971).
[f] OxCal calibration/comparison (Bronk Ramsey, 2001) using SHCal04 (McCormac *et al.*, 2004) or IntCal04 (Reimer *et al.*, 2004) and Hughen *et al.* (2006) after correction for Southern Hemisphere offset (-41 ± 14 yr; Hogg *et al.*, 2002) (2-σ range, 95.4%).
[g] Online CalPal2005_SFCP calibration/comparison (Weninger *et al.*, 2005; accessed January, 2007) using Shackleton *et al.* (2004) after correction for Southern Hemisphere offset (-41 ± 14 yr; Hogg *et al.*, 2002) (1-σ range, 68.2.%).

Copyright © 2007 John Wiley & Sons, Ltd.

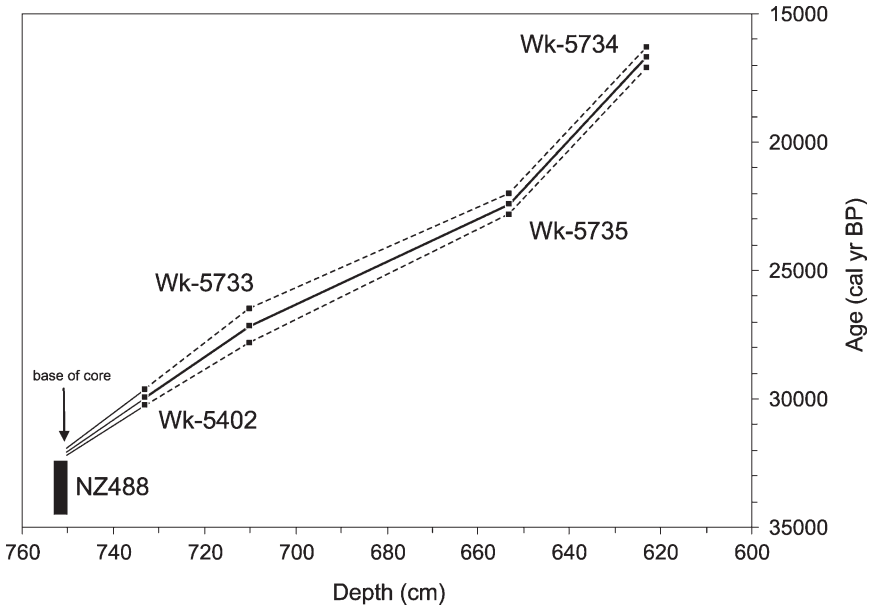

Figure 3 Age–depth model for lower part of Kohuora Core B, derived by linear interpolation between radiocarbon-dated horizons. Dashed lines indicate ±1 standard deviation errors. Solid thick line connects mean calibrated ages. Thin solid lines extrapolate age model to base of sequence (at 750 cm). Solid rectangle (NZ488) indicates mean ± 1 standard deviation cal. age range for previous estimate of age of Kohuora volcano (see Table 4)

Palynology

Samples for pollen analysis were taken at mostly 5 cm intervals from Core B from the base of the sequence to the position of the Rotorua Tephra. This compositionally distinctive tephra provides a robust tie-point between the two Kohuora cores (Fig. 2). Pollen samples for Core A were taken at higher resolution to facilitate a separate investigation into short-term volcanic impacts on vegetation through the Lateglacial and Holocene (Giles, 1999). The two records from cores A and B are spliced here to provide a continuous sequence from before the eLGM to near present but the focus of our discussion is primarily concerned with the results from Core B.

Preparation of pollen slides followed standard procedures (Faegri and Iversen, 1989). Known quantities of exotic *Lycopodium* spores were added during preparations in order to calculate pollen concentrations. Pollen percentages for dryland pollen taxa were calculated as a percentage of the total dryland pollen sum (250 grains counted per slide). Pollen and spores from wetland and fern taxa are excluded from the total land pollen sum. These taxa are expressed as a percentage of the particular group (e.g. wetland) plus the total dryland pollen. *Leptospermum* pollen is over-represented in the record, indicating local on-site pollen influx, and therefore also is excluded from the dryland pollen sum.

Charcoal fragments greater than 10 μm diameter were quantified from the pollen slides by recording the number of tephra and charcoal fragments passing a graticule endpoint (after Clark, 1982). Fragments smaller than 10 μm proved difficult to identify and count accurately. Charcoal concentrations were calculated in the same way as for total pollen concentrations using the marker spore method.

Results

Results for the full sequence of pollen changes in both cores are displayed in Fig. 4 and those for Core B at an expanded scale in Fig. 5. Six pollen zones are recognised with two further divided into subzones as described below.

Zone Ko1, with estimated age of ca. 32–29 ka, exhibits a clear transition between two types of vegetation complex and is divided into two subzones accordingly. The lowermost sediments, in Zone Ko1a, were deposited close to a restiad mire with wire rush (*Empodisma*) and umbrella fern (*Gleichenia*) prominent but in decline. The regional vegetation was an unusual complex of beech and cedar forest with some lowland-montane conifer trees present (e.g. *Agathis, Dacrydium, Podocarpus, Prumnopitys*) along with *Cyathea smithii* tree ferns. As the zone progresses, these latter more thermophilous elements increase at the expense of the beeches and cedar. Forest cover was far from complete at first as grassland and shrubland were prominent, along with the restiad mire, although these elements are also declining upwards in the zone. Pollen concentrations were very low. The estimated age of this subzone is ca. 32–31 ka.

Zone Ko1b records the continued expansion of lowland-montane conifers to a prominent level, at the expense of beech, cedar, grassland herbs and shrubs. This interval is dated to around 31–29 ka. The first appearance of the macrophytic quilwort *Isoetes* indicates the development of lacustrine conditions.

Zone Ko2 exhibits a reversal of many of the trends depicted in Zone Ko1b with the return to prominence of the beeches, shrubland and grassland and the decline to very low levels of lowland-montane conifers. Lacustrine conditions are now well established with *Isoetes* spores recorded in increasingly high levels along with the colonial alga *Botryococcus* and *Pediastrum*. These elements broadly define an extended period of open landscape and cool climate that extended from ca 29–20 ka but with some variation, represented by three subzones. Of these, subzones Ko2a (ca. 29–25.5 ka) and Ko2c (ca. 24–20 ka) are characterised by high levels of grasses, other herbs and various shrubs, notably *Halocarpus*, *Coprosma* and Asteraceae that are typical of subalpine to alpine environments in New Zealand today. Tree pollen is reduced to the lowest levels for the entire sequence in these subzones although beech pollen remains important. The intervening

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

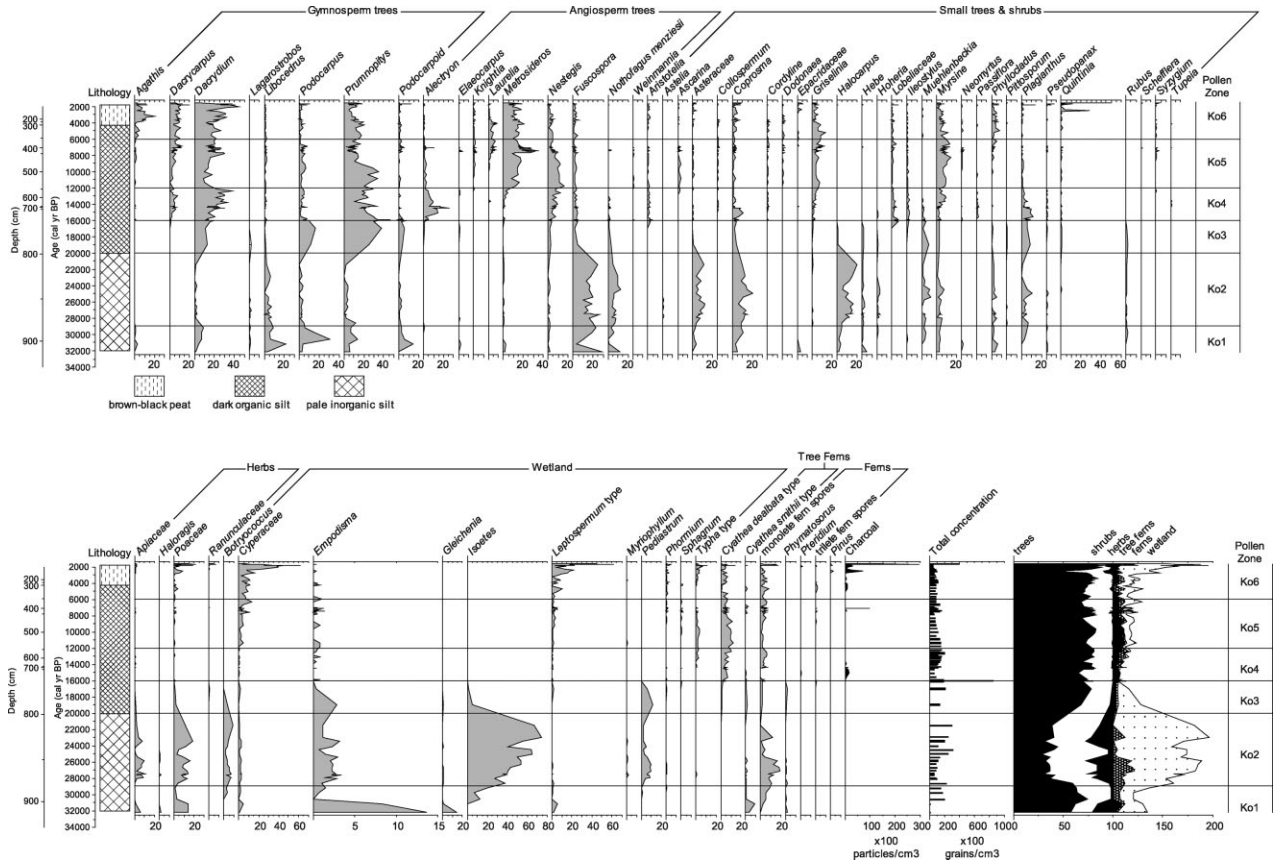

Figure 4 Kohuora full pollen percentage diagram (cores A and B). Depth scale is based on Core B from top down to 781 cm depth (position of Rotorua Tephra). Below 781 cm, depths are based on relative depth below Rotorua Tephra in Core B

subzone, Ko2b (ca. 25.5–24 ka) exhibits a temporary decline in grassland taxa whilst shrubland taxa generally expand along with the cool temperate trees cedar and beech.

Zone Ko3 marks a striking change in the sequence as the cool climate elements of grasses, shrubs and cool temperate trees (cedar and beeches) are replaced by lowland-montane podocarp forest taxa, in particular *Prumnopitys*, *Podocarpus* and *Dacrydium*. This major transition commences sometime between ca. 21 and 19 ka although the precise timing is difficult to establish at this site because of weak dating control for this interval. During this expansion of podocarp forest, the pre-existing grassland communities decline to negligible levels.

Further compositional changes are evident in these podocarp forests during Zone Ko4 (ca. 16–12 ka), marked by expansion to high levels of *Dacrydium*, along with the angiosperm trees *Alectryon* and *Nestegis* and *Cyathea dealbata* type tree ferns. Zone Ko5 (ca. 12–6 ka) is characterised by decline of *Dacrydium*, expansion of *Metrosideros* along with several other angiosperm forest taxa, including *Nestegis* and *Ascarina*, while *Cyathea dealbata* type tree ferns are now prominent. These taxa remain prominent in the uppermost zone (Ko6; ca. 6–2 ka) which is characterised by the expansion of *Agathis*. The uppermost levels in this zone show spasmodic changes which suggest that these near-surface sediments may have been

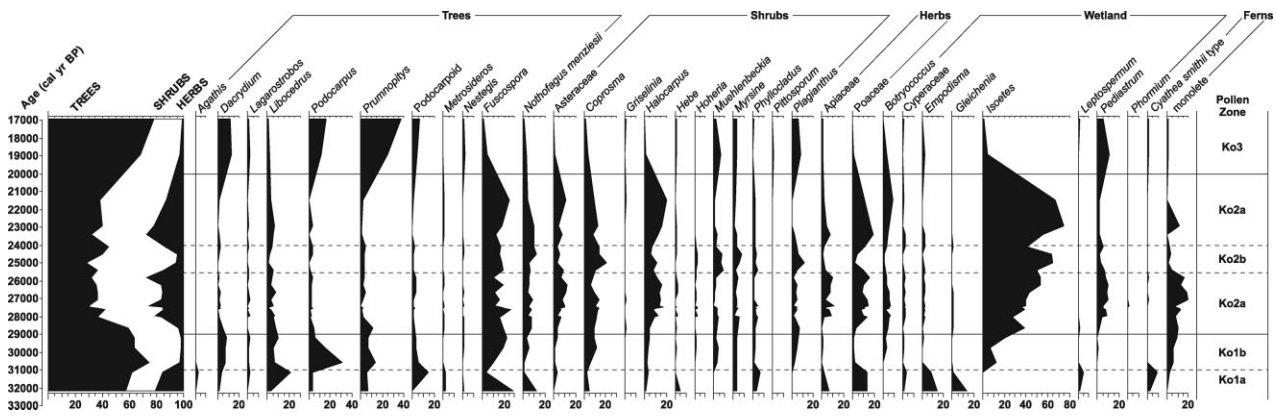

Figure 5 Kohuora Core B pollen percentage diagram for extended LGM

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

subject to post-depositional disturbance as a result of various land-use activities at the site and hence are not discussed further here.

Discussion

Climate change inferred from the Kohuora pollen records

The pollen sequence from Kohuora Core B shows a clear distinction between assemblages more typical of modern vegetation in the region (Zones Ko3–6) and those characterised by 'southern' New Zealand vegetation communities today (zones Ko1 and Ko2). This distinction corresponds broadly to the Lateglacial–Holocene and eLGM, respectively, as discussed below.

The basal pollen assemblages in Zone Ko1a exhibit an unusual combination of comparatively thermophilous elements such as *Agathis* and *Dacrydium* alongside cooler-climate indicators such as *Fuscospora*, *Nothofagus menziesii* and Poaceae and Apiaceae. There is also good representation of genera whose species collectively encompass a broad climatic range, such as *Libocedrus, Phyllocladus, Halocarpus* and *Metrosideros*. This unusual composition of the basal pollen assemblages is explained if it is assumed that they reflect the early vegetation sequence following soon after the eruption of Kohuora volcano. The basal assemblages include several taxa that respond to disturbance or are prominent in early secondary vegetation succession, including Poaceae, Apiaceae, *Myrsine*, *Coprosma*, Epacridaceae, *Leptospermum* and *Cyathea smithii*. All these pollen curves either decline upwards during Zone Ko1a or early in the overlying subzone, in a pattern resembling successional response to a major disturbance event. The prominence of *Empodisma* pollen and *Gleichenia* spores in these basal sediments may reflect wetland development from impeded drainage associated with the eruption as occurred elsewhere in the Auckland area (Newnham and Lowe, 1991). The pollen-vegetation succession culminates in a peak of ~20% for *Libocedrus*, a tree which is known to colonise freshly disturbed surfaces (Veblen and Stewart, 1982) including those formed by volcanic processes (McGlone *et al.*, 1988; Horrocks and Ogden, 1998; Newnham and Alloway, 2004). It seems likely therefore that the pollen assemblages in Ko1a represent seral succession in the dryland vegetation as well as localised wetland development following the eruption that formed Kohuora volcano. The co-existence of both warm temperate and cool temperate communities points to a moist interstadial climate, intermediate between glacial and interglacial conditions, consistent with a Marine Isotope Stage (MIS) 3 age indicated by the radiocarbon and tephra-based chronologies. A similar vegetation complex has been reported from the Bay of Plenty, northeastern North Island at around the same time (McGlone *et al.*, 1984). The balance between warm temperate and cool temperate elements changes in pollen zone Ko1b, notably characterised by a marked decline in grassland and limited expansion of podocarp forest. This development indicates the completion of secondary succession to 'climax vegetation' following the Kohuora eruption under a climate favouring more thermophilous elements over cool-climate taxa.

The pollen assemblages of Zone Ko2 indicate marked contraction of podocarp forest and expansion of beech forest, shrubland and grassland. These expanding vegetation communities are rare in northern New Zealand today, except in the vicinity of the treeline, and so their prominence in this lowland setting indicates considerable cooling, consistent with onset of glacial conditions at this locality (Sandiford *et al.*, 2003). However, the relative importance of these three cool climate elements in the pollen record and hence the inferred extent of the vegetation they represent vary significantly during this interval. The subdivision of this zone indicates a distinctive oscillation, with a period of expanded shrubland (Ko2b) separating two intervals when grassland communities were more prominent (Ko2a and Ko2c). This pattern indicates that a prominent mild interval occurred within the eLGM although the continued repression of forest development throughout Ko2 suggests that temperatures remained considerably lower than present. This mid-eLGM mild phase is estimated to have occurred from ca. 25.5 to 24 ka although the dating control for this part of the Kohuora record is imprecise. The timing and regional extent of the mid-eLGM mild phase are discussed further below.

Termination of the eLGM is indicated by the marked expansion of warm temperate podocarp forest and contraction of cool climate vegetation communities, indicated by the pollen assemblages in Zone Ko3. Although the timing of the (last) termination is difficult to determine precisely at Kohuora as the sample spacing is comparatively coarse for this interval (Fig. 6), we can be confident that extensive podocarp forest had established in the region certainly by ca. 16 ka with the expansion commencing possibly as early as soon after 19 ka. Further clarification of the timing of this event is provided below through comparison with other records from the region (see also Newnham *et al.*, 2003).

Pollen Zones Ko4–6 indicate further compositional changes in the Lateglacial–Holocene forests, notably the expansion of important angiosperm trees in Zones 4 and 5 and of *Agathis australis* and *Phyllocladus* (presumably *P. trichomanoides*) in Zone 6. It is difficult to confidently relate these compositional changes to climate influence as they involve taxa which co-exist in the Auckland region today with broad ecological and climatic tolerances. For example, the most prominent taxa in Zones 4–6 are *Dacrydium cupressinum*, *Prumnopitys* (both *P. taxifolia* and *P. ferrugineus*) and *Metrosideros*, each of which have a very broad biogeographic distribution today that extends across the lowland-montane zone for most of New Zealand.

Some general patterns are apparent, however, which allow some climate inference. Zone Ko4 (ca.16–12 ka) is characterised by a switch in the dominant podocarp taxa from *Prumnopitys* to *Dacrydium cupressinum* along with expansion of certain angiosperm forest taxa, notably *Nestegis* and *Alectryon*. This change strongly suggests an increase in precipitation and humidity as has been observed in other climate proxy records from northern New Zealand at around this time (e.g. Newnham *et al.*, 1989). There is no obvious reversal of the deglacial warming trend as has been observed elsewhere in New Zealand in a variety of climate proxy records during the interval ca. 15–10 ka (e.g. Denton and Hendy, 1994; Hellstrom *et al.*, 1998; Ivy-Ochs *et al.*, 1999; Newnham and Lowe, 2000; Turney *et al.*, 2003; Samson *et al.*, 2005; Williams *et al.*, 2005; Hajdas *et al.*, 2006). It is noticeable that this 'New Zealand Lateglacial Reversal' (Williams *et al.*, 2005) has only been clearly observed to date at sites located at higher latitude and often considerably higher altitude than Auckland, although the Pukaki (Auckland) pollen record of Sandiford *et al.* (2003) shows a minor expansion of grass pollen relative to forest podocarp pollen, possibly indicative of minor cooling at around 13 ka (Alloway *et al.*, 2007). Also, Pepper *et al.* (2004) observed changes in thickness of laminated sediments at Onepoto maar, Auckland (Fig. 1), during the interval, which they suggested may be linked to enhanced southwesterly

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

airflow at Auckland during the Antarctic Cold Reversal (ca. 14–12 ka).

A further forest compositional change at the transition to Zone Ko5 at around 12 ka is marked by the rise to prominence of *Metrosideros*, along with several other angiosperm trees and shrubs, including *Nestegis*, *Ascarina* and *Laurelia*. Tall podocarp trees remain prominent whilst the beeches decline to the lowest levels observed for the entire sequence. These developments in forest composition are consistent with the concept of an early Holocene 'thermal optimum' during which climate in northern New Zealand was consistently warm, moist and equitable (McGlone,1988; Williams *et al.*, 2004). From ca.7 to 6 ka, broadly increasing to present, the expansion of the northern conifers *Agathis australis* and *Phyllocladus* (probably *P. trichomanoides*) accompanies a decline in some of these warm-temperate angiosperm trees and shrubs, notably *Ascarina* and *Nestegis*. The same broad vegetation patterns can be observed in other Holocene pollen records from northern New Zealand (Ogden *et al.*, 1992; Newnham *et al.*, 1995).

Comparison with other palynological records from Auckland

Comparable Late Quaternary palynological records from the Auckland area have been obtained at Lake Waiatarua (Newnham and Lowe, 1991), Mt Richmond (Sandiford *et al.*, 2002), Pukaki Lagoon (Lancashire *et al.*, 2002; Sandiford *et al.*, 2003), Onepoto (Shane and Sandiford, 2003), Crater Hill (Lancashire *et al.*, 2002), and Lake Pupuke (Horrocks *et al.*, 2005) (Fig. 1). As is the case with Kohuora, all of these depositional sites were formed by volcanism and each contains identified silicic tephra layers derived from central North Island volcanoes that considerably enhance their radiocarbon chronologies and facilitate direct inter-site comparisons for these discrete moments in time. Pukaki and Onepoto are essentially tuff-ring explosion craters or maars that occur close to modern sea level. Mt Richmond, like Crater Hill, comprises a scoria cone with explosion crater and tuff ring, whereas Lake Waiatarua was formed when a pre-existing valley was blocked by lava flow from a nearby vent and scoria cone (Mt Wellington/Maungarei). None of these records spans the entire interval encompassed by the two sequences reported here for Kohuora. The Pupuke record covers the mid-late Holocene; Lake Waiatarua is essentially a Holocene record, spanning the period 12–4 ^{14}C ka; the Onepoto record of Shane and Sandiford (2003) spans most of MIS 4 and MIS 3; the Mt Richmond and Pukaki records commence just prior to deposition of the Kawakawa/Oruanui tephra and provide good coverage of MIS 2, with Pukaki extending into the early Holocene when the basin was inundated by marine transgression. The Crater Hill record spans the interval ca. 18 ^{14}C ka to the late Holocene whilst a Holocene pollen diagram from Pukaki Lagoon was also presented by Lancashire *et al.* (2002).

The Kohuora pollen records substantiate and consolidate the general patterns and timing of vegetation and climate change indicated by these fragmented records, whilst at the same time extending the regional coverage of sites and providing some local detail. For the eLGM, the best dated and highest-resolution records comparable to Kohuora are from Pukaki Lagoon and Mt Richmond. The occurrence of Kawakawa/Oruanui tephra at three sites within 5 km of one another permits direct correlation between these three records at the time of its deposition (Fig. 6). The records at Mt Richmond and Kohuora indicate sustained cooling from around 29–28 ka. Prior to this, both records indicate vegetation cover of predominantly beech and podocarp forest with some limited shrubland and grassland patches and an inferred climate intermediate between glacial and interglacial states. Between 29 and 28 ka a marked period of grassland expansion commenced (Fig. 6), during which podocarp forest was extremely restricted in the region, whilst beech forest and shrubland maintained persistent high levels. The pollen record from Pukaki shows the same pattern, except that the lowermost assemblages were deposited later than at the other two sites, by which time the grassland expansion had already occurred. This period of grassland expansion is not uniform, as discussed below, but persists until ca. 20–19 ka at which time podocarp forest expansion is evident at all three sites. A temperature depression of at least 4°C from present levels is implied for much of this period by the inferred lowering of treeline (Sandiford *et al.*, 2003). This prolonged period of regional grassland expansion and treeline lowering indicates substantial cooling from ca. 29–19 ka which is appropriately termed the eLGM, as noted earlier.

Environmental change around the time of the Kawakawa/Oruanui eruption

It is interesting to make a detailed comparison between these three pollen records for the early stages of the eLGM when the

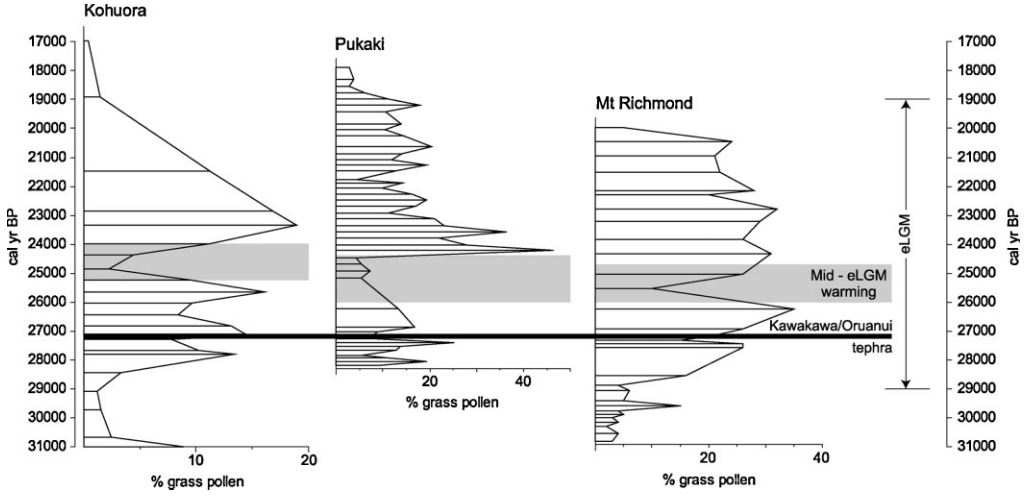

Figure 6 Comparison of grass pollen percentage curves for Kohuora (this study), Pukaki (Sandiford *et al.*, 2003) and Mt Richmond (Sandiford *et al.*, 2002)

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

Kawakawa/Oruanui tephra provides a direct tie-line or isochron (Fig. 6). All three records show that this tephra layer was deposited approximately 1000 years after the commencement of sustained cooling but during a brief interval of less severe climate when shrubland expanded regionally at the expense of grassland. This milder interval probably only persisted for at most a few hundred years and was neither long enough nor strong enough for any discernible forest expansion to occur. The evidence for this brief, weak climate amelioration would probably have been overlooked or attributed to local variation were it not for the occurrence of the Kawakawa/Oruanui isochron which synchronises the three records and shows that the timing and amplitude of this brief climate amelioration were near-identical across the three sites (Fig. 6). Furthermore, all three records show that this brief milder interval ended very soon after deposition of Kawakawa/Oruanui tephra, when the grass pollen curves at all three sites return to high levels culminating in a peak at 27–26 ka. The juxtaposition of evidence for a substantial eruption with evidence for the termination of a comparatively mild phase within the early eLGM raises the possibility of a volcanic cooling effect.

Recent observations of climate following large volcanic eruptions show that when large quantities of volcanic aerosols and ash are injected into the stratosphere, particularly from the equatorial region, complex changes in weather patterns may occur at a regional to global scale (e.g. Robock, 2002). A net temporary surface cooling is often observed, primarily due to the volcanic aerosols causing increased reflection or attenuation of insolation in the stratosphere. Whilst any cooling following historic eruptions has lasted only a few years at most, there is a compelling argument that past eruptions that occurred on a much larger scale may have had a considerably longer and more severe cooling effect. The best example from the late Quaternary record is the ca 74 ka Toba 'super eruption' in Sumatra which coincided approximately with the onset of glacial conditions at the end of MIS 5 (e.g. Rampino and Self, 1993), although the extent of cooling and impact is debated (Zielinski *et al.*, 1996; Oppenheimer, 2002). The Kawakawa/Oruanui eruption is the largest known 'wet' eruption (explosive interaction between silicic magma and external water), generating ~430 km^3 of fall deposits, ~320 km^3 of pyroclastic density-current (mostly ignimbrite) deposits, and ~420 km^3 of primary intracaldera material, approximately 1170 km^3 in total and equivalent to ~530 km^3 of magma (Wilson, 2001; Manville and Wilson, 2004; Wilson *et al.*, 2006). It is reasonable to postulate therefore that the apparent cooling trend inferred by the increasing levels of grass pollen in Kohuora, Pukaki and Mt Richmond pollen records commencing just after deposition of the Kawakawa/Oruanui tephra was triggered by volcanic cooling of the atmosphere.

Alternatively, the pollen assemblages immediately above Kawakawa/Oruanui tephra may reflect expansion of grassland communities in response to regional scale disturbances in the wake of the eruption. Manville and Wilson (2004) showed that in the immediate aftermath of the Kawakawa/Oruanui eruption intense erosion and remobilisation of Oruanui pyroclastic deposits were widespread over much of the central North Island. Tephric loess formed in much of this region (Eden and Hammond, 2003) and there is evidence for massive downstream fluvial aggradation, particularly in the lower Waikato River (Hume *et al.*, 1975; Hogg *et al.*, 1987; Edbrooke, 2001) which currently approaches to within ~50 km of Auckland (Fig. 1), and in eastern North Island (Litchfield and Berryman, 2005; Litchfield and Rieser, 2005). Although no such volcanic impacts have been reported for the Auckland region, it is possible that the pollen records discussed above are in part registering wider regional vegetation disturbance brought about by the eruption. The distinction between these two scenarios—volcanic cooling and volcanic disturbance—cannot be made from the Auckland pollen records alone, but might be achieved by closely examining palaeoenvironmental records containing Kawakawa/Oruanui tephra at localities well beyond the zone of landscape impact. Such records are discussed further below.

Mid-eLGM Interstadial

The palynological records at Kohuora, Mt Richmond and Pukaki are very similar in showing two phases of cold climate which together encompass the eLGM. The first cold phase extends from ca. 29 ka to ca. 26 ka and the second from ca. 24.5 ka to ca. 19 ka. These two coldest periods are separated by a brief milder interval recognised at all three sites by the marked contraction of grassland and expansion of shrubland that occurs after the impact of Kawakawa/Oruanui tephra has been registered in the pollen records. We refer to this mild interval, lasting approximately 1000–2000 years from ca. 26 to 24 ka, as the mid-eLGM Interstadial. Considering the broad error limits of the 'calibrated/comparison' ^{14}C ages and uncertainties surrounding ^{14}C production at this time, the concordance in detail between these three sites is strong. The pollen records also show an earlier, shorter-lived mild period during the first cold phase and within which the Kawakawa/Oruanui tephra was deposited. As discussed above, the climatic signals in this part of the pollen record are complicated by the volcanic impacts that accompanied this eruption.

Comparison with glacial chronostratigraphy from southwestern New Zealand

Suggate (1990) reviewed evidence for glaciation in New Zealand, one of the most important centres of Quaternary ice accumulation in the Southern Hemisphere. A key region for eLGM (Late Otiran) glaciation is western South Island (Westland), where a comprehensive record for glacial advances has recently been established by Suggate and Almond (2005). In north Westland, glacial fluctuations are determined through a combination of stratigraphic and geomorphic evidence integrated with sequences of loess, pollen analysis and radiocarbon dating of peats and buried soils that formed in deglacial contexts. In the south of the region, correlative features are recognised through a framework of soil stratigraphy developed for loess coverbeds formed on glacial landforms (Almond *et al.*, 2001). Throughout the region, three distinct LGM glacial advances of approximately similar extents have been mapped ($la_1/M5_1$, $la_2/M5_2$, mn/M6). These advances are estimated to have culminated at 28, 22 and 19 ka, with the first and last of these used by Suggate and Almond to constrain the duration of the LGM in New Zealand. In addition they reported a period of climatic amelioration, also evident in pollen records (Vandergoes *et al.*, 2005; see also Newnham *et al.*, 2007a), that separates the first two of these glacial advances and is estimated to extend from ca. 28 to ca. 25 ka. Kawakawa/Oruanui tephra has been recorded at several localities in Westland, typically at localities either beyond the LGM ice limits (Newnham *et al.*, 2007b) or within these limits, interbedded with non-glacial sediments (Suggate and Almond, 2005).

This record of glaciation from southwestern New Zealand for the most part shows strong agreement with the sequence of

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

eLGM climatic patterns inferred from the Auckland pollen records, some 6.5 degrees of latitude to the north. First, both records suggest an onset of cooling resulting in a cold or glacial peak at ca. 28 ka. Although the Westland glacial record cannot be used to date the onset of this glacial cooling, the Auckland pollen records suggest it commenced around 1000 years before 28 ka (i.e. ca. 29 ka). Second, both records indicate marked and sustained deglacial warming from around 19 ka. This strong agreement in delimiting the duration of the eLGM, using different proxies from widely separated regions, gives substantial support to the suggestion by Suggate and Almond (2005) that the 'LGM' in New Zealand occurred between ca. 28 and 19 ka and thus commenced during late MIS 3. Third, both records show that the Kawakawa/Oruanui tephra was deposited after this initial eLGM peak but during a subsequent milder period. The two records differ, however, in showing the duration of this early eLGM amelioration. Whereas the Westland glacial sequence implies climate amelioration extending from ca. 28 to 25 ka (Suggate and Almond, 2005), the Auckland pollen record suggests that it lasted just a few hundred years before deposition of the Kawakawa/Oruanui tephra, although a later more substantial climate amelioration is indicated for the period ca. 26–24.5 ka. One explanation for these differences stems from our speculation that the immediate post-Kawakawa/Oruanui tephra palynological records at Auckland reflect volcanic impact, either through a volcanic cooling effect or through widespread landscape disturbances following the eruption. Following this line of argument, this geographical pattern may be useful in distinguishing between the two types of volcanic impact postulated from the Auckland records. Although a volcanic cooling impact would likely have been widespread and affected both regions, it is unlikely that landscape disturbance following an eruption from central North Island would be registered in the sedimentary sequences from southwestern New Zealand, where the Kawakawa/Oruanui tephra layer is observed as a layer only a few millimetres thick at most (Newnham *et al.*, 2007b). This logic favours the assertion that the volcanic impact at Auckland is not cooling but landscape disturbance and implies that the climate amelioration might have commenced as early as ca. 28 ka generally in accordance with the glacial evidence from southwestern New Zealand. Ongoing work aimed at enhancing the chronology and sampling resolution of Westland pollen records containing Kawakawa/Oruanui tephra (M. Vandergoes *et al.*, unpublished) should provide further insight into the patterns and causes of climate change at this time.

Comparison with other palaeoenvironmental climate proxies from New Zealand

An exhaustive review of palaeoecological records for the last glacial cycle is beyond the scope of this paper and past summaries have been provided by Pillans *et al.* (1993) and Newnham *et al.* (1999b). In this section we review a broad range of climate proxy records that can be compared directly with the palynological records from Auckland, in most cases involving the Kawakawa/Oruanui tephra isochron.

Pollen records from multiple sites in the Tongariro region of central North Island (McGlone and Topping, 1983) and the Bay of Plenty–Gisborne regions of northeastern North Island (McGlone *et al.*, 1984) attest to the environmental conditions prevailing before and after the Kawakawa/Oruanui eruption. The Tongariro records show pronounced cooling at ca. 29 ka, consistent with the Auckland records, as well as harsh climatic conditions and high levels of erosion for an extended interval following the Kawakawa/Oruanui eruption. Similarly, the Bay of Plenty–Gisborne records also indicate landscape-scale erosion commencing before the Kawakawa-Oruanui eruption. Further insights come from the pollen record of marine core P69, off southeastern North Island, which shows the Kawakawa/Oruanui tephra was deposited in a relatively warmer period (McGlone, 2001). This marine record also suggests that LGM-like cooling may have commenced as early as 45 ka, at odds with the terrestrial evidence presented here.

Records of aeolian quartz deposition from Taranaki (Alloway *et al.*, 1992), midway between Auckland and Westland (Fig. 1), show similar patterns during the eLGM and provide insight into a possible causal mechanism. The aeolian quartz flux corresponds closely with independent records of climate change with strong increases during glacial stages thought to reflect the expansion of Antarctic sea ice into lower latitudes along with strengthening of the southern westerlies (Alloway *et al.*, 1992). Although imprecisely dated, these records contain Kawakawa/Oruanui tephra and so have a secure tie-point with the Auckland pollen records for the early eLGM. In depositional sequences at Waitui and Onaero, the 'LGM' is marked by considerably increased but variable rates of aeolian quartz flux. The Kawakawa/Oruanui tephra occurs within the early phase of a variable interval of generally lower aeolian quartz flux that separates two peaks, the first of which represents the culmination of 'LGM' onset (Alloway *et al.*, 1992; Fig. 7). Although the timing of these changes cannot be determined directly, this pattern is broadly consistent with the pollen records from Auckland and glacial evidence from Westland and precisely so at the time of deposition of Kawakawa/Oruanui tephra (Fig. 7). The consistent detection of a milder interval within the eLGM using different proxies in three distinct regions of western New Zealand indicates a significant widespread climate phenomenon. If so, then the interval of weakened aeolian flux may represent a moderate reversal in these conditions, in particular polewards contraction of the circum-Antarctic westerlies and sea-ice front.

Further supporting evidence for this early eLGM climate amelioration comes from a variety of palaeoecological climate proxies reported from lake sediments deposited between periods of glacial advance at the Lyndon Stream site in Canterbury, central-eastern South Island (Fig. 1). As with Westland, this region is well beyond, and upwind from, the likely zone of impact of landscape disturbance following the Kawakawa/Oruanui eruption. On the basis of stratigraphic relationships and the occurrence of plant macrofossils preserved in these lake sediments, Soons and Burrows (1978) suggested they represented an interstadial within the LGM with a reduction of February (summer) temperatures of up to 2.5°C from present levels. More recently, Woodward and Shulmeister (2007) presented chironomid-based temperature reconstructions from these lake sediments which suggest summer temperatures during this period were on average only 1°C cooler than present. These temperature reconstructions are in broad agreement with recent beetle evidence from the base of the same sequence suggesting cooling of ~2°C from present in both summer and winter (Marra *et al.*, 2006b). Using two radiocarbon ages reported by Soons and Burrows (1978) that approximately bound the lake sediments, Woodward and Shulmeister (2007) suggested a calibrated age range of ca 26 600 cal. yr BP to 24 500 cal. yr BP for this within-LGM interstadial. Although error margins on these age estimations are broad (1σ errors are ca. 650–1000 yr) this interval coincides with the period of climate amelioration recognised in the Auckland pollen records and falls within the climate amelioration interval recognised from the Westland glacial records. Woodward and Shulmeister (2007) questioned the

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

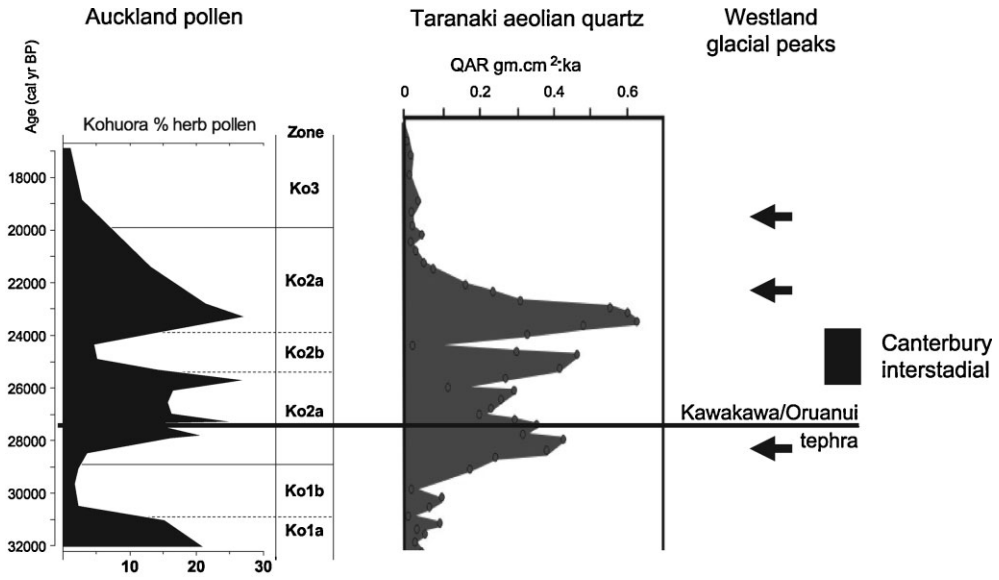

Figure 7 Comparison of Kohuora herb pollen curve (this study) with aeolian quartz flux from Waitui, Taranaki (Alloway *et al.*, 1992). Arrows indicate timing of three 'LGM' glacial advances recognised by Suggate and Almond (2005) and Lyndon stream interstadial (Woodward and Shulmeister, 2007)

validity of qualitative pollen-based temperature reconstructions for the LGM in other New Zealand records, because they appeared to suggest much stronger cooling than the chironomids-based temperature reconstructions from Lyndon Stream. We suggest an alternative explanation: that the Lyndon Stream chironomid-based reconstruction is from a mild interval that is not characteristic of the eLGM as a whole.

Conclusion

In summary, the Auckland pollen records add further insight into the patterns of LGM climate in northern New Zealand and are broadly consistent with a variety of palaeoenvironmental evidence from southern New Zealand. Strong concordance between glacial evidence from the south and the pollen evidence from the north implies that LGM conditions prevailed across most of New Zealand during the interval ca. 29–19 k cal. yr BP, longer and earlier than the LGM *sensu stricto*. We suggest that the term extended LGM (eLGM) may be more appropriate for the New Zealand region. Within this predominantly cold interval, the Auckland pollen records indicate a climatic amelioration for the interval ca. 26–24 k cal. yr BP, consistent with other palaeoecological data from Canterbury, which falls within a period of climate amelioration recognised between the first two LGM glacial advances of Suggate and Almond (2005). We refer to this mild interval as the eLGM Interstadial on the basis of its widespread recognition in a range of proxies. The eLGM Interstadial may have commenced as early as 28 ka, based on the glacial record, although climatic interpretation of this part of the Auckland pollen records is complicated by the possibility of volcanic disturbance or cooling in the aftermath of the Kawakawa/Oruanui eruption (27–26 ka). Nevertheless it is clear that this major eruption occurred during a period of climate amelioration and glacier retreat soon after the first significant cold peak and ice advance of the eLGM. As further palaeoenvironmental sequences containing this tephra layer are developed at localities well beyond the zone of volcanic landscape construction and disturbance, it should be possible to determine whether this amelioration continued uninterrupted into the eLGM Interstadial as implied by the glacial evidence, or was separated from it by a colder interval, possibly associated with volcanic cooling. This distinction is critical to explanations of the possible causes of the eLGM Interstadial.

To conclude, this paper demonstrates the widespread applicability of a new framework for the New Zealand eLGM, visible in a range of proxies and at contrasting sites. There is now a compelling need to explain these distinctive climate changes in the southern mid-latitudes in the context of global patterns and forcing mechanisms. Strong correspondence between biotic responses, glacial fluctuations and aeolian quartz deposition linked to major shifts in strength and latitudinal extent of the southern westerlies, suggest that both the eLGM and eLGM interstadial may be more widely registered across the Southern Ocean. Support for this assertion comes from widespread geomorphic and palaeoecological data from western and southernmost South America (Heusser *et al.*, 1999; Denton *et al.*, 1999; Clapperton, 2000; McCulloch *et al.*, 2005) and from marine isotopic, pollen and foraminifera records from the Southern Pacific Ocean (Nelson *et al.*, 1993; Heusser and van der Geer, 1994; Pahnke *et al.*, 2003; Lamy *et al.*, 2004). All these records show or imply commencement of LGM conditions earlier than the currently defined onset of 24 k cal. yr BP, often by several thousand years in line with the New Zealand evidence. These widespread Southern Hemisphere patterns are congruent with recent reconstructions of the globally averaged ice-equivalent sea-level history (Waelbroeck *et al.*, 2002; Peltier and Fairbanks, 2006), suggesting that the eLGM may have been a global phenomenon. A comprehensive authoritative revisiting of the EPILOG definition of LGM is now required, alongside renewed effort to examine natural archives in the context of an extended LGM.

Acknowledgements We thank Peter de Lange (Department of Conservation) and Calvin Mora for valued help with field work. Drs Alan Hogg (Waikato Radiocarbon Dating Lab) and Tom Higham (now at University of Oxford) are thanked for processing and assaying 17 ^{14}C dates from the Kohuora cores, and for help with calibrations. Funding for the drilling by ProDrill, Auckland, was provided by a grant to B.V.A. from the Auckland Regional Council (ARC), facilitated by Michelle Daly (former manager, Hazards Management Group).

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs

Detailed reviews by Matt McGlone (Landcare Research) and Andrew Lorrey (University of Auckland), and comments by the editor, Chris Turney, were very helpful. We would also like to thank James Scourse for inviting this contribution and for extending great patience over its completion. Funding for the ^{14}C dating was provided by grants to D.J.L. from the ARC (also facilitated by Michelle Daly) and from the Department of Earth Sciences (University of Waikato). Renat Radosinsky (University of Waikato) and Drs Paul Froggatt and Dave Manning (formerly Victoria University of Wellington) are thanked for their assistance in obtaining electron microprobe data.Tim Absolom and Brian Rodgers drafted the figures.

References

Allen SR, Smith IEM. 1994. *Eruption styles and volcanic hazard in the Auckland Volcanic Field, New Zealand*. Geoscience Reports of Shizuoka University **20**; 5–14.

Alloway BV, Stewart RB, Neall VE, Vucetich CG. 1992. Climate of the last glaciation in New Zealand, based on aerosolic quartz influx in an andesitic terrain. *Quaternary Research* **38**: 170–179.

Alloway BV, Westgate J, Pillans B, Pearce N, Newnham RM, Byrami MB, Aarburg S. 2004. A prolific terrestrial record of Mid-Pleistocene distal TVZ silicic volcanism in Auckland Region, New Zealand: stratigraphy, age and correlation. *New Zealand Journal of Geology and Geophysics* **47**: 447–479.

Alloway BV, Lowe DJ, Barrell DJA, Newnham RM, Almond PC, Augustinus PC, Bertler N, Carter L, Litchfield NJ, McGlone MS, Shulmeister J, Vandergoes MJ, Williams PW and NZ-INTIMATE members. 2007. Towards a climate event stratigraphy for New Zealand over the past 30 000 years (NZ-INTIMATE project). *Journal of Quaternary Science* **22**: 9–35.

Almond PC, Moar NT, Lian OB. 2001. Reinterpretation of the glacial chronology of South Westland, New Zealand. *New Zealand Journal of Geology and Geophysics* **44**: 1–15.

Bronk Ramsey C. 2001. Development of the radiocarbon calibration program OxCal. *Radiocarbon* **43**: 355–363.

Byrami M, Newnham RM, Alloway BV, Pillans B, Ogden J, Westgate J, Mildenhall DC. 2005. A long early-mid Pleistocene pollen record from Auckland New Zealand. In *Early–Middle Pleistocene Transitions: The Land–Ocean Evidence*, Head MJ, Gibbard PL (eds). Special Publication, 247 Geological Society: London; 183–208.

Cassidy J, Locke CA, Miller CA, Rout DJ. 1999. The Auckland volcanic field, New Zealand: geophysical evidence for its eruption history. In *Volcanoes in the Quaternary*, Firth C, McGuire WJ (eds). Special Publication 161, Geological Society: London; 1–10.

Clapperton C. 2000. Interhemispheric synchroneity of Marine Oxygen Isotope Stage 2 glacier fluctuations along the American cordilleras transect. *Journal of Quaternary Science* **14**: 435–468.

Clark RL. 1982. Point count estimation of charcoal in pollen preparations and thin sections. *Pollen et Spores* **24**: 523–535.

Clayton-Greene KA. 1978. Aspects of the distribution of certain indigenous woody species in the Waikato District, New Zealand. *Journal of the Royal Society of New Zealand* **8**: 283–291.

Cockayne L. 1928. *New Zealand Plants and their Story*. Engelmann: Leipzig.

Cranwell LM. 1981. *The Botany of Auckland*. Auckland Institute and War Memorial Museum: Auckland.

Denton GH, Hendy CH. 1994. Younger Dryas age advance of Franz Josef Glacier in the Southern Alps, New Zealand. *Science* **264**: 1434–1437.

Denton GH, Lowell TV, Heusser CJ, Schlüchter C, Andersen BG, Heusser LE, Moreno PI, Marchant DR. 1999. Geomorphology, stratigraphy and radiocarbon chronology of Lanquihue drift in the area of the southern Lake District, Seno Reloncaví, and Isla de Chiloé, Chile. *Geografiska Annaler A* **81**: 167–229.

Edbrooke SW (compiler). 2001. *Geology of the Auckland Area. Institute of Geological and Nuclear Sciences 1:250,000 Geological Map3. 1 sheet + 74 p.* Institute of Geological and Nuclear Sciences: Lower Hutt, New Zealand.

Eden DN, Hammond AP. 2003. Dust accumulation in the New Zealand region since the last glacial maximum. *Quaternary Science Reviews* **22**: 2037–2052.

Elliot MB. 1998. Late Quaternary pollen records of vegetation and climate change from Kaitaia Bog, far northern New Zealand. *Review of Palaeobotany and Palynology* **99**: 189–202.

Elliot MB, Neall VE, Wallace C. 2005. A Late Quaternary pollen record from Lake Tangonge, far northern New Zealand. *Review of Palaeobotany and Palynology* **136**: 143–158.

Faegri K, Iversen J. 1989. *Textbook of Pollen Analysis*, 4th edn. Wileys: Chichester.

Fleming CA. 1979. *Geological History of New Zealand and Its Life*. Auckland University Press: Auckland.

Froggatt PC, Lowe DJ. 1990. A review of late Quaternary silicic and some other tephra formations from New Zealand: their stratigraphy, nomenclature, distribution, volume, and age. *New Zealand Journal of Geology and Geophysics* **33**: 89–109.

Giles TM. 1999. *Volcanic emissions and distal palaeoenvironmental impacts in New Zealand*. PhD thesis, University of Plymouth, UK.

Grant-Taylor TC, Rafter TA. 1971. New Zealand radiocarbon age measurements—6. *New Zealand Journal of Geology and Geophysics* **14**: 364–402.

Hajdas I, Lowe DJ, Newnham RM, Bonani G. 2006. Timing of the late-glacial climate reversal in the Southern Hemisphere using high-resolution radiocarbon chronology for Kaipo bog, New Zealand. *Quaternary Research* **65**: 340–345.

Hayward BW, Grenfell HR, Ashwaq TS, Morley MS, Horrocks M. 2006. Effect and timing of increased freshwater runoff into sheltered harbour environments around Auckland City, New Zealand. *Estuaries and Coasts* **29**: 165–182.

Hellstrom J, McCulloch M, Stone J. 1998. A detailed 13,000-year record of climate and vegetation change, from the isotope geochemistry of two New Zealand speleothems. *Quaternary Research* **50**: 167–178.

Heusser CJ, Heusser LE, Lowell TV. 1999. Paleoecology of the southern Chilean Lake District-Isla Grande de Chiloe during middle-late Llanquihue glaciation and deglaciation. *Geografiska Annaler* **81A**: 231–284.

Heusser LE, van de Geer G. 1994. Direct correlation of terrestrial and marine paleoclimatic records from four glacial-interglacial cycles—DSDP site 594 southwest Pacific. *Quaternary Science Reviews* **13**: 273–282.

Hogg AG, Lowe DJ, Hendy CH. 1987. University of Waikato radiocarbon dates I. *Radiocarbon* **29**: 263–301.

Hogg AG, McCormac FG, Higham TFG, Reimer PJ, Baillie MGL, Palmer JG, Stuiver M. 2002. High-precision ^{14}C measurements of contemporaneous tree-ring dated wood from the British Isles and New Zealand: AD 1850–950. *Radiocarbon* **44**: 633–640.

Horrocks M, Ogden J. 1998. The effects of the Taupo Tephra eruption of c. 1718 BP on the vegetation of Mt Hauhungatahi, central North Island, New Zealand. *Journal of Biogeography* **4**: 649–660.

Horrocks M, Augustinus P, Deng Y, Shane P, Andersson S. 2005. Holocene vegetation, environment, and tephra recorded from Lake Pupuke, Auckland, New Zealand. *New Zealand Journal of Geology and Geophysics* **48**: 85–94.

Horrocks M, Nichol SL, Augustinus PC, Barber IG. 2007. Late Quaternary environments, vegetation and agriculture in northern New Zealand. *Journal of Quaternary Science* **22**: 267–279.

Hughen K, Southon J, Lehman S, Bertrand C, Turnbull J. 2006. Marine-derived ^{14}C calibration and activity record for the past 50,000 years updated from the Cariaco Basin. *Quaternary Science Reviews* **25**: 3216–3227.

Hume TM, Sherwood AM, Nelson CS. 1975. Alluvial sedimentology of the Upper Pleistocene Hinuera Formation, Hamilton Basin, New Zealand. *Journal of the Royal Society of New Zealand* **5**: 421–462.

Ivy-Ochs S, Schlüchter C, Kubik PW, Denton GH. 1999. Moraine exposure dates imply synchronous Younger Dryas glacier advances in the European Alps and in the Southern Alps of New Zealand. *Geografiska Annaler* **81**: 313–323.

Kermode LO. 1992. *Geology of the Auckland urban area. Scale 1:50 000. Institute of Geological and Nuclear Science Geological Map 2*. Institute of Geological and Nuclear Sciences Ltd, Lower Hutt, New Zealand.

Copyright © 2007 John Wiley & Sons, Ltd.

Lambeck K, Yokoyama Y, Purcell T. 2002. Into and out of the Last Glacial Maximum: sea-level change during Oxygen Isotope Stages 3 and 2. *Quaternary Science Reviews* **21**: 343–360.

Lamy F, Kaiser J, Ninneman U, Hebbeln D, Arz HW, Stoner J. 2004. Antarctic timing of surface water changes off Chile and Patagonian ice sheet response. *Science* **304**: 1959–1962.

Lancashire AK, Flenley JR, Harper M. 2002. Late Glacial beech forest: an 18,000–5000-BP pollen record from Auckland, New Zealand. *Global and Planetary Change* **33**: 315–327.

Lees CM, Neall VE, Palmer AS. 1998. Forest persistence at coastal Waikato, 24,000 years BP to present. *Journal of the Royal Society of New Zealand* **28**: 55–81.

Litchfield NJ, Berryman KR. 2005. Correlation of fluvial terraces within the Hikurangi Margin, New Zealand: implications for climate and baselevel controls. *Geomorphology* **68**: 291–313.

Litchfield NJ, Rieser U. 2005. Optically stimulated luminescence age constraints for fluvial aggradation terraces and loess in eastern North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* **48**: 581–589.

Lowe DJ, Newnham RM, Ward CM. 1999. Stratigraphy and chronology of a 15 ka sequence of multi-sourced silicic tephras in a montane peat bog, eastern North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* **42**: 565–579.

Lowe DJ, Shane PAR, Alloway BV, Newnham RM. 2007. Fingerprints and age models for widespread New Zealand tephra marker beds erupted since 30,000 years ago: a framework for NZ-INTIMATE. *Quaternary Science Reviews* **26**: (in press).

Manville V, Wilson CJN. 2004. The 26.5 ka Oruanui eruption, New Zealand: a review of the roles of volcanism and climate in the post-eruptive sedimentary response. *New Zealand Journal of Geology and Geophysics* **47**: 525–547.

Marra MJ, Alloway BV, Newnham RM. 2006a. Paleoenvironmental reconstruction of a well-preserved Stage 7 forest sequence catastrophically buried by basaltic eruptive deposits, northern New Zealand. *Quaternary Science Reviews* **25**: 2143–2161.

Marra MJ, Shulmeister J, Smith EGC. 2006b. Reconstructing temperature during the Last Glacial Maximum from Lyndon Stream, South Island, New Zealand using beetle fossils and maximum likelihood envelopes. *Quaternary Science Reviews* **25**: 1841–1849.

McCormac FG, Hogg AG, Blackwell PG, Buck CE, Higham TFG, Reimer PJ. 2004. ShCal04 Southern Hemisphere calibration, 0–11.0 cal kyr BP. *Radiocarbon* **46**: 1087–1092.

McCulloch RD, Fogwill CJ, Sugden DE, Bentley MJ, Kubik PW. 2005. Chronology of the last glaciation in the central Strait of Magellan and Bahia Inutil, southernmost South America. *Geografiska Annaler* **87A**: 289–312.

McGlone MS. 2001. A late Quaternary pollen record from marine core P69, southeastern North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* **44**: 69–77.

McGlone MS, Topping WW. 1983. Late Quaternary vegetation, Tongariro regions, central North Island, New Zealand. *New Zealand Journal of Botany* **21**: 53–76.

McGlone MS, Howorth R, Pullar WA. 1984. Late Pleistocene stratigraphy, vegetation and climate of the Bay of Plenty and Gisborne regions, New Zealand. *New Zealand Journal of Geology and Geophysics* **27**: 327–350.

McGlone MS. 1988. New Zealand. In *Vegetation History*, Huntley B, Webb T (eds). Kluwer: Dordrecht; 557–599.

McGlone MS, Neall VE, Clarkson BD. 1988. The effect of recent volcanic events and climate changes on the vegetation of Mt Taranaki (Egmont), New Zealand. *New Zealand Journal of Botany* **26**: 123–144.

Mix AC, Bard E, Schneider R. 2001. Environmental processes of the ice age: land, oceans, glaciers (EPILOG). *Quaternary Science Reviews* **20**: 627–657.

Nelson CS, Cooke PJ, Hendy CH, Cuthbertson AM. 1993. Oceanographic and climatic changes over the past 160,000 years at Deep-sea Drilling Project site 594 off southeastern New Zealand, south west Pacific Ocean. *Paleoceanography* **8**: 435–458.

Newnham RM. 1999. Environmental change in Northland, New Zealand during the Last Glacial and Holocene. *Quaternary International* **57–58**: 61–70.

Newnham RM, Alloway BV. 2004. A terrestrial record of Interglacial climate preserved by voluminous debris avalanche inundation in Taranaki, western North Island, New Zealand. *Journal of Quaternary Science* **19**: 299–314.

Newnham RM, Lowe DJ. 1991. Holocene vegetation and volcanic activity, Auckland Isthmus, New Zealand. *Journal of Quaternary Science* **6**: 177–193.

Newnham RM, Lowe DJ. 2000. Fine-resolution pollen record of late-glacial climate reversal from New Zealand. *Geology* **28**: 759–762.

Newnham RM, Lowe DJ, Green JD. 1989. Palynology, vegetation, and climate of the Waikato lowlands, North Island, New Zealand, since c. 18 000 years ago. *Journal of the Royal Society of New Zealand* **19**: 127–150.

Newnham RM, de Lange PJ, Lowe DJ. 1995. Holocene vegetation, climate, and history of a raised bog complex, northern New Zealand, based on palynology, plant macrosfossils and tephrochronology. *The Holocene* **5**: 267–282.

Newnham RM, Lowe DJ, Alloway BV. 1999a. Volcanic hazards in Auckland, New Zealand: a preliminary assessment of the threat posed by central North Island silicic volcanism based on the Quaternary tephrostratigraphical record. In *Volcanoes in the Quaternary*, Firth C, McGuire WJ (eds). Special Publication 161, Geological Society: London; 27–45.

Newnham RM, Lowe DJ, Williams PW. 1999b. Quaternary environmental change in New Zealand: a review. *Progress in Physical Geography* **23**: 567–610.

Newnham RM, Eden DN, Lowe DJ, Hendy CH. 2003. Rerewhakaaitu Tephra, a land-sea marker for the Last Termination in New Zealand, with implications for global climate change. *Quaternary Science Reviews* **22**: 289–308.

Newnham RM, Lowe DJ, Green JD, Turner GM, Harper MA, McGlone MS, Stout SL, Horie S, Froggatt PC. 2004. A discontinuous ca. 80 ka record of Late Quaternary environmental change from Lake Omapere, Northland, New Zealand. *Palaeogeography, Palaeoclimatology, Palaeoecology* **207**: 165–198.

Newnham RM, Vandergoes MJ, Hendy CH, Lowe DJ, Preusser F. 2007a. A terrrestrial palynological record for the last two glacial cycles from southwestern New Zealand. *Quaternary Science Reviews* **26**: 517–535.

Newnham RM, Vandergoes MJ, Garnett M, Lowe DJ, Prior C, Almond PC. 2007b. Test of AMS ^{14}C dating of pollen concentrates using tephrochronology. *Journal of Quaternary Science* **22**: 37–51.

New Zealand Meteorological Service. 1980. *Summaries of climatological observations to 1980*. Miscellaneous Publication 177, New Zealand Meteorological Service, Wellington.

Ogden J, Hogg AG, Wilson AT, Hendy CH, Newnham RM. 1992. The late Quaternary history of kauri (*Agathis-australis*) in New Zealand and its climatic significance. *Journal of Biogeography* **19**: 611–622.

Oppenheimer C. 2002. Limited global change due to the largest known Quaternary eruption, Toba ~74 kyr BP? *Quaternary Science Reviews* **21**: 1593–1609.

Pahnke K, Zahn R, Elderfield H, Schulz M. 2003. 340,000-year centennial-scale marine record of southern hemisphere climatic oscillation. *Science* **301**: 948–952.

Peltier WR, Fairbanks RG. 2006. Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews* **25**: 3322–3337.

Pepper AC, Shulmeister J, Nobes DC, Augustinus PA. 2004. Possible ENSO signals prior to the Last Glacial Maximum, during the deglacial and the early Holocene, from New Zealand. *Geophysical Research Letters* **31**: L15206, DOI:10.1029/2004GL020236.

Pillans BJ, McGlone MS, Palmer AS, Mildenhall DC, Alloway BV, Berger GW. 1993. The Last Glacial Maximum in central and southern North Island, New Zealand: a palaeoenvironmental reconstruction using the Kawakawa Tephra Formation as a chronostratigraphic marker. *Palaeogeography, Palaeoclimatology, Palaeoecology* **101**: 283–304.

Rampino MR, Self S. 1993. Climate-volcanism feedback and the Toba eruption of ~74,000 years ago. *Quaternary Research* **40**: 269–280.

Reimer PJ, Baillie MGL, Bard E, Bayliss A, Beck JW, Bertrand CJH, *et al.* 2004. IntCal04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP. *Radiocarbon* **46**: 1029–1058.

Copyright © 2007 John Wiley & Sons, Ltd.

Robock A. 2002. Pinatubo eruption: the climatic aftermath. *Science* **295**: 1242–1244.

Samson CR, Sikes EL, Howard WR. 2005. Deglacial paleoceanographic history of the Bay of Plenty, New Zealand. *Palaeoceanography* **20**: 1–12.

Sandiford A, Alloway BV, Shane P. 2001. A 28 000–6600 cal yr record of local and distal volcanism preserved in a paleolake, Auckland, New Zealand. *New Zealand Journal of Geology and Geophysics* **44**: 323–336.

Sandiford A, Horrocks M, Newnham RM, Ogden J, Alloway BV. 2002. Environmental change during the Last Glacial Maximum (ca 25,500–16,500 years BP) at Mt Richmond, Auckland Isthmus, New Zealand. *Journal of the Royal Society of New Zealand* **32**: 155–167.

Sandiford A, Newnham RM, Alloway BV, Ogden J. 2003. A 28 000–7600 cal yr BP pollen record of vegetation and climate change from Pukaki Crater, northern New Zealand. *Palaeogeography, Palaeoclimatology, Palaeoecology* **201**: 235–247.

Searle EJ. 1981. *City of Volcanoes—A Geology of Auckland (2nd edn)*. Longman Paul: Auckland.

Shackleton NJ, Fairbanks RG, Chiu T-C, Parrenin F. 2004. Absolute calibration of the Greenland time scale: implications for Antarctic time scales and for $\Delta^{14}C$. *Quaternary Science Reviews* **23**: 1513–1522.

Shane PAR. 2005. Towards a comprehensive distal andesitic tephrostratigraphic framework for New Zealand based on eruptions from Egmont volcano. *Journal of Quaternary Science* **20**: 45–57.

Shane PAR, Hoverd J. 2002. Distal record of multi-sourced tephra in Onepoto Basin, Auckland, New Zealand: implications for volcanic chronology, frequency and hazards. *Bulletin of Volcanology* **64**: 441–454.

Shane PAR, Sandiford A. 2003. Paleovegetation of marine isotope stages 4 and 3 in northern New Zealand and the age of the widespread Rotoehu tephra. *Quaternary Research* **59**: 420–429.

Shane PAR, Smith IEM. 2000. Geochemical fingerprinting of basaltic tephra deposits in the Auckland Volcanic Field. *New Zealand Journal of Geology and Geophysics* **43**: 569–577.

Shane PAR, Smith VC, Lowe DJ, Nairn IA. 2003. Re-identification of c. 15 700 cal yr BP tephra bed at Kaipo Bog, eastern North Island: implications for dispersal of Rotorua and Puketarata tephra beds. *New Zealand Journal of Geology and Geophysics* **46**: 591–596.

Smith VC, Shane PAR, Nairn IA. 2005. Trends in rhyolite geochemistry, mineralogy, and magma storage during the last 50 kyr at Okataina and Taupo volcanic centres, Taupo Volcanic Zone, New Zealand. *Journal of Volcanology and Geothermal Research* **148**: 372–406.

Soons JM, Burrows CJ. 1978. Dates for Otiran deposits, including plant microfossils and macrofossils, from Raikaia Valley, New Zealand. *New Zealand Journal of Geology and Geophysics* **16**: 425–438.

Suggate RP. 1990. Late Pliocene and Quaternary Glaciations of New Zealand. *Quaternary Science Reviews* **9**: 175–197.

Suggate RP, Almond PC. 2005. The Last Glacial Maximum (LGM) in western South Island, New Zealand: implications for the global LGM and MIS 2. *Quaternary Science Reviews* **24**: 1923–1940.

Tomlinson AI, Sansom J. 1994a. *Temperature normals for New Zealand for the period 1961 to 1990*. National Institute of Water and Atmospheric Research Science & Technology Series 4, National Institute of Water and Atmospheric Research, Wellington.

Tomlinson AI, Sansom J. 1994b. *Rainfall normals for New Zealand for the period 1961 to 1990*. National Institute of Water and Atmospheric Research Science & Technology Series 4, National Institute of Water and Atmospheric Research, Wellington.

Turney CSM, McGlone MS, Wilmshurst JM. 2003. Asynchronous climate change between New Zealand and the North Atlantic during the last deglaciation. *Geology* **31**: 223–226.

Vandergoes MJ, Newnham RM, Preusser F, Hendy CH, Lowell TV, Fitzsimons SJ, Hogg AG, Kasper HU, Schluchter C. 2005. Regional insolation forcing of late Quaternary climate change in the Southern Hemisphere. *Nature* **436**: 242–244.

van der Plicht J, Beck JW, Bard E, Baillie MGL, Blackwell PG, Buck CE, Friedrich M, Guilderson TP, Hughen KA, Kromer B, McCormac FG, Bronk Ramsey C, Reimer PJ, Reimer RW, Remmele S, Richards DA, Southon JR, Stuiver M, Weyhenmeyer CE. 2004. NotCal04—comparison/calibration ^{14}C records 26–50 cal kyr BP. *Radiocarbon* **46**: 1225–1238.

Veblen TT, Stewart GH. 1982. On the conifer regeneration gap in New Zealand: the dynamics of *Libocedrus bidwillii* stands on South Island. *Journal of Ecology* **70**: 413–436.

Vucetich CG, Pullar WA. 1969. Stratigraphy and chronology of late Pleistocene volcanic ash beds in central North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* **12**: 784–837.

Waelbroeck C, Labeyrie L, Michel E, Duplessey J-C, McManus J, Lambeck K, Balbon E, Labracherie M. 2002. Sea-level and deep water temperature changes derived from benthids foraminifera isotopic records. *Quaternary Science Reviews* **21**: 295–305.

Wardle P. 1963. Evolution and distribution of the New Zealand flora as affected by Quaternary climates. *New Zealand Journal of Botany* **1**: 3–17.

Weninger B, Jöris O, Danzeglocke U. 2005. The CalPal Online Radiocarbon Calibration [interactive online calibration CalPal_2005_SFCP at http://www.calpal.de/ —incorporates timescale of Shackleton *et al*. (2004), cited above].

Williams PW, King DNT, Zhao J-X, Collerson KD. 2004. Speleothem master chronologies: combined Holocene ^{18}O and ^{13}C records from the North Island of New Zealand and their palaeoenvironmental interpretation. *The Holocene* **14**: 194–208.

Williams PW, King DNT, Zhao JX, Collerson KD. 2005. Late Pleistocene to Holocene composite speleothem ^{13}C and ^{18}O chronologies from South Island, New Zealand—did a global Younger Dryas really exist? *Earth and Planetary Science Letters* **230**: 301–317.

Wilson CJN. 2001. The 26.5 ka Oruanui eruption, New Zealand: an introduction and overview. *Journal of Volcanology and Geothermal Research* **112**: 133–174.

Wilson CJN, Switsur RV, Ward AP. 1988. A new ^{14}C age for the Oruanui (Wairakei) eruption, New Zealand. *Geological Magazine* **125**: 297–300.

Wilson CJN, Blake S, Charlier BLA, Sutton AN. 2006. The 26.5 ka Oruanui eruption, Taupo volcano, New Zealand: development, characteristics and evacuation of a large rhyolitic magma body. *Journal of Petrology* **47**: 34–69.

Woodward CA, Shulmeister J. 2007. Chironomid-based reconstructions of summer air temperature from lake deposits in Lyndon Stream, New Zealand spanning the MIS 3/2 transition. *Quaternary Science Reviews* **26**: 142–154.

Wright IC, McGlone MS, Nelson CS, Pillans BJ. 1995. An integrated latest Quaternary (Stage 3 to present) paleoclimatic and paleoceanographic record from offshore northern New Zealand. *Quaternary Research* **44**: 283–295.

Zielinski GA, Mayewski PA, Meeker LD, Whitlow S, Twickler MS, Taylor K. 1996. Potential impact of the Toba mega-eruption ~71,000 years ago. *Geophysical Research Letters* **23**: 837–840.

Copyright © 2007 John Wiley & Sons, Ltd.

DOI: 10.1002/jqs