Does the bipolar seesaw extend to the terrestrial southern mid-latitudes?

Rewi M. Newnham a,*, Marcus J. Vangergoos b, Elisabeth Sikes c, Lionel Carter d, Janet M. Wilmhurst e, David J. Lowe f, Matt S. McGlone g, Anna Sandiford g

Article history:
Received 1 April 2010
Received in revised form 11 April 2011
Accepted 15 April 2011
Available online 31 May 2011

Keywords:
Bipolar seesaw
Extended LGM
LGT
Lateglacial reversal
Antarctic isotope maxima
ACR
YD
NZ-INTIMATE
Palynology
Tephrochronology

ABSTRACT

High precision comparison of Greenland and Antarctic ice cores, suggesting a pervasive antiphased temperature relationship between the polar hemispheres during the last glaciation, lends strong support to the bipolar seesaw model (EPICA, 2006). The extent to which reorganisation of ocean-heat transport during abrupt climate change events affected the southern mid-latitudes remains unclear, however, owing to a paucity of well-dated records with robust climate proxies, variability between some records, and varying interpretations of their significance. Here we present temperature reconstructions for three key pollen records recognised by the NZ-INTIMATE (NZ-I) group which, along with the preliminary NZ-I climate event stratigraphy (Alloway et al., 2007) and published marine records, are compared with polar ice core records for the interval 30–10 cal. ka. We focus on key events within the context of Dansgaard Oeschger cycles 4–1 and The Antarctic Cold Reversal/Younger Dryas intervals. The New Zealand records are broadly consistent with an extended bipolar seesaw whereby the oceanic southern mid-latitudes are warmed at times of MOC weakening or cessation in the North Atlantic, and vice versa. Variability between records indicate that other factors must be involved, however, and nor do these records refute alternative models that predict an antiphased inter-hemispheric pattern. Nevertheless an extended bipolar model may explain an early onset of LGM conditions in New Zealand and elsewhere in the Southern Hemisphere at a time when interstadials GI3 and GI4 kept Greenland warm. Similar inter-hemispheric dynamics have been invoked to explain an earlier termination of the LGM in Antarctica than in Greenland (Wolff et al., 2009) which is also evident in the New Zealand records. A prominent mid-LGM interstadial complex observed in several New Zealand records, connected by tephrochronology may represent another antiphased event although stronger chronological control is needed to support this assertion. By comparison, the variability evident between New Zealand Lateglacial records cannot be explained simply by chronological imprecision and there may be latitudinal controls on the extent to which the Antarctic pattern is registered, as has been suggested from other records from the southern mid-high latitudes. Consolidation of these patterns is needed from more precise climate reconstructions at sites spread across latitudinal gradients in both the marine and terrestrial realms.

© 2011 Elsevier Ltd. All rights reserved.

1. Introduction

Ice core records from Greenland show repeatedly that the last glacial period was characterised by recurrent abrupt climate changes (Bond et al., 1993; Alley et al., 1999). Most striking are a series of stadials and interstadials, termed Dansgaard–Oeschger (D–O) cycles, with amplitudes and patterns that mimick glacial–interglacial cycles but on much shorter, millennial timescales (e.g., Johnsen et al., 2001). Similar patterns have been observed in other climate-proxy records from the Northern Hemisphere (Voelker, 2002) and, with subdued amplitudes, in some Holocene marine records from the North Atlantic (Bond et al., 1997, 2001). Considerable investigative effort has been applied to these abrupt changes and to understanding their relevance in the context of contemporary climate (e.g., Alley et al., 1999; Wunsch, 2006). Much
of this work has been directed towards clarifying the global extent of key events and their geographical pattern and timing which provides potential avenues for distinguishing between putative causal mechanisms (Broecker, 2003; Clement and Peterson, 2008).

This approach is highlighted in the recent revelation that Greenland abrupt climate change events during the last glaciation are consistently manifested in the EPICA ice cores from Antarctica, with what is referred to as a one-to-one antiphased coupling (EPICA, 2006). Building on earlier work by Blunier and Brook (2001), the EPICA Community showed that the prominent D–O interstadials/stadials of the Greenland ice cores are consistently matched by more subdued cooling/warming intervals in Antarctica. Their findings provide strong evidence to substantiate the currently-favoured “Bipolar Seesaw” mechanism for abrupt climate change, which calls on the reorganisation of ocean heat transport connected to Meridional Overturning (MOC) in the polar hemispheres (Broecker, 1998). The bipolar seesaw model predicts an antiphased relationship between the high latitudes of the northern and southern hemisphere due to net changes in ocean heat flux (Alley et al., 1999; Knutti et al., 2004). Away from the polar regions however, evidence for the antiphased inter-hemispheric pattern is less consistent and robust suggesting that other physical mechanisms were involved. Postulated mechanisms include tropical processes (Cane, 1998; Cane and Clement, 1999; Seager and Battisti, 2007), sea ice feedbacks (Timmermann et al., 2003; Denton et al., 2005), and changes in the strength of southern westerlies linked to latitudinal displacement of the Intertropical Convergence Zone (Barker et al., 2009; Whittaker et al., 2011) or to changes in the seasonality of atmospheric heat transport (McClune et al., 2010).

Against the backdrop of investigating inter-hemispheric patterns of climate change, the New Zealand INTIMATE project has developed a preliminary climate event stratigraphy for the interval 30–8 calendar (cal.) ka (NZI-CES), based on selected key terrestrial and marine records from the New Zealand region (Alloway et al., 2007). The NZI-CES was supported by a revision of tephrostratigraphical data to help synchronize many of the records and to improve age models (Lowe et al., 2008a). This effort follows a parallel initiative undertaken by the North Atlantic INTIMATE group (Bjöck et al., 1998; Lowe et al., 2008b). On the whole the New Zealand records, from locations which straddle the mid-southern latitudes (c. 34°–47°S), show a broadly coherent relationship with many Antarctic ice core records for this period. The NZI-CES is based on a small number of well-resolved and securely-dated records, however, and uncertainty surrounds the precision of some of the inferred proxy-climate relationships and their potentially variable response times. These limitations may have promoted a tendency to favour one particular alignment over others that may also be plausible, within margins of dating errors.

In this paper we seek to build on the work of NZ-INTIMATE and EPICA by examining the extent to which terrestrial records from New Zealand indicate a likely extension of the bipolar seesaw pattern into the southern mid-latitudes. We present new quantified estimates of mean annual temperature (MAT) from three of the key New Zealand pollen records recognised by NZ-INTIMATE. Within current limits of chronological uncertainty, we examine if the D–O cycles manifest in the New Zealand evidence and, if so, with an Antarctic or Greenland template. If ocean heat transport is the primary mechanism responsible for the antiphased bipolar pattern it should be registered in similar fashion in terrestrial records from the New Zealand region. Based on the preliminary findings of NZ-INTIMATE, we hypothesise that there is a consistent, though likely complex, in-phase correspondence with most Antarctic records and hence support for a bipolar-mid latitude seesaw connection.

2. New Zealand climate setting

One of only a few landmasses athwart subtropical and subantarctic climatic regimes in the Southern Hemisphere, New Zealand is positioned to respond sensitively to regional climatic fluctuations (Fig. 1; Newnham et al., 1999). Its weather is largely the product of interaction between subpolar and subtropical atmosphere—ocean circulation systems played out across a strongly dissected, topographically-diverse landscape. Central and southern regions of New Zealand are strongly influenced by the circum-Antarctic westerly vortex of atmospheric circulation, whereas the subtropical anticyclonic belt extends over northern regions (Fig. 1). Offshore, cool sub-Antarctic and polar surface waters, delivered mainly by the Antarctic Circumpolar Current, bathe southern parts of the region while northern areas are influenced by Subtropical Inflow. At depths >2000 m, the New Zealand micro-continent steers the local sector of the global thermohaline circulation into the Pacific Ocean (e.g. Warren, 1981; Schmitz, 1995; Carter and Wilkin, 1999). As a result of these marine influences, the latitudinal spread of mean annual temperatures on land ranges from a low of 10 °C in the temperate south to 16 °C in the tropical

Fig. 1. (a) Location of New Zealand and elements of the regional oceanic and atmospheric circulation. (b) Location of pollen sites.
north. Temperatures also drop on average about 0.7 °C for every 100 m of altitude with full alpine conditions found in the most mountainous regions. In combination with the prevailing westerlies, the northeast–southwest trending mountains also create a strong rain shadow effect and a marked precipitation gradient from western rainforests to eastern semiarid-plains.

3. Method and approach

We consider two distinct time periods in which D–O events figure prominently. The first, the Last Glacial–Interglacial Transition (LGIT: ca 18–10 cal. ka), has more records, better chronologies and climate-inferences generally are more robust relative to pre-18 cal. ka. Nevertheless, considerable controversy surrounds both and climate-inferences generally are more robust relative to pre-

which includes 2

2006; Tovar et al., 2008). In this study, we focus on three key phased Antarctic events (EPICA, 2006). For this interval, uncer-

2007). Of the seven climate variables correlated with pollen taxa,

settlement (pre-deforestation) pollen database (Wilmshurst et al.,

2007). Previous published studies predated the development of the New Zealand pollen-climate calibration model (Wilmshurst et al., 2007) and only qualitative pollen-climate inferences were available. Here we present new quantitative temperature reconstructions for these records.

Second, we consider the NZI-CES for the interval 30–20 cal. ka, which includes 2–3 prominent Greenland D–O cycles with anti-

phased Antarctic events (EPICA, 2006). For this interval, uncer-

tainities in dating and climate-inference impose more stratigraphic limitations and therefore demand greater caution in developing cross correlations. For both intervals, comparisons are made with other key records from the New Zealand region as well as other sectors of the mid-high latitude Southern Hemisphere.

3.1. Pollen-temperature reconstructions

Quantification of modern pollen–climate relationships in New Zealand has been complicated by the destruction of >75% of the original forest cover during ca 750 years of settlement history (Norton et al., 1986). This obstacle has been circumvented recently by using a pollen-climate calibration model based on a pre-

settlement (pre-deforestation) pollen database (Wilmshurst et al.,

2007). Of the seven climate variables correlated with pollen taxa, the most statistically robust relationships were obtained with MAT and Wilmshurst et al. (2007) showed that useful reconstructions of this temperature parameter could be made using the modern analogue technique, with error margins of ca 1.5 °C. This error is large compared with the amplitude of temperature changes considered here. As a result our discussion of these data centres on the timing and direction of temperature changes rather than their amplitude.

In this study, we have applied this pollen-climate model to derive estimates of MAT from previously published pollen records for the interval 18–10 cal. ka at three key sites: Pukaki, a maar crater situated at sea level in Auckland (36° 59′S) (Sandiford et al., 2003); Kaipo, a montane bog situated near the treeline in east-

central North Island (38° 40′S) (Newnham and Lowe, 2000); and Okarito, a lowland bog that has leveed within glacial landforms in southwestern South Island (43° 14′S) adjacent to the Southern Alps (Vandergoes et al., 2005; Newnham et al., 2007a). All three records have age models developed by tephrostratigraphy and radiocarbon dating that supported their recognition as key records by NZI. For this study, we used the Kaipo age model of Hajdas et al.

(2006), which was supported by Bayesian analysis using Bpeat reported in Lowe et al. (2007). We have developed new age models for Pukaki and Okarito by re-calibrating the original 14C ages and tephra ages using INTCAL09 with an assumed inter-hemispheric offset of 56 ± 24 yr (McCormac et al., 2004).

3.2. Terminology of events and timescales

During the period 30–8 cal. ka, four prominent abrupt climate change cycles are widely recognised and provide specific targets for this investigation (Table 1). The generally accepted scheme of terminology used to describe these events reflects their marine and terrestrial expression. In Greenland ice cores, the D–O abrupt cycles consist of a prominent cooling phase followed by abrupt warming culminating in an interstadial peak. The coolest phases of the D–O cycles, immediately preceding the abrupt warming phases, coincide with deposition of ice-rafted debris in the North Atlantic Ocean, termed Heinrich (H) events. In the ocean, the H events are accom-

panied by significant reduction or cessation in MOC (McManus et al.,

2004). During the time frame of this study Heinrich events 1, 2, 3 and the Younger Dryas occur.

After synchronising the NGRIP and Antarctic ice core records using global atmospheric changes in CH4 concentrations, the EPICA Community (2006) correlated the H events of the North Atlantic with warming intervals of similar length but more subdued amplitude in Antarctica, referred to as the Antarctic Isotope Maxima (AIM). In most cases, the cooling that terminates an AIM coincided with the abrupt warming of a D–O Interstadial peak in Greenland. The amplitude of the AIM were linearly related to the duration of D–O cold stadials (EPICA, 2006). Following the nomenclature introduced by the North Atlantic INTIMATE group (Björck et al., 1998; Walker et al., 1999; Lowe et al., 2008b), we use the terms D–O interstadials/stadials and Greenland interstadials/stadials synonymously.

Comparing terrestrial and marine based chronologies with ice core chronologies is not precise, because the latter are referenced to a ‘present’ baseline of 2000 AD, whereas the former, which here are largely based on the calibrated radiocarbon timescale, are referenced to a ‘present’ baseline of 1950 AD. We have adjusted the NZ chronologies by 50 yrs when comparing with ice core chronologies (e.g., Figs. 2 and 3). Unless otherwise stated, we use the term ‘cal. ka’ to refer to thousands of calendar years before 1950 AD in the context of calibrated terrestrial and marine records and the term ‘yr b2k’ to refer to before 2000 AD in the context of ice core records.

Table 1

<table>
<thead>
<tr>
<th>Greenland</th>
<th>North Atlantic</th>
<th>Antarctica</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>D04/G14</td>
<td>H4</td>
<td>AIM4</td>
<td>29–31</td>
</tr>
<tr>
<td>D03/G13</td>
<td></td>
<td></td>
<td>29</td>
</tr>
<tr>
<td>D02/G12</td>
<td>H2</td>
<td>AIM2</td>
<td>25–23</td>
</tr>
<tr>
<td>D01/G11</td>
<td>H1</td>
<td>AIM1</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ACR</td>
<td>17–15</td>
</tr>
<tr>
<td>GS1</td>
<td>Younger Dryas</td>
<td></td>
<td>15</td>
</tr>
</tbody>
</table>

a ka b2k for onset of Greenland events (except CS1) to nearest ka (from Andersen et al., 2006 and EPICA, 2006). Cal ka BP for North Atlantic events (estimated to nearest ka). Estimated age of Antarctic events based on assumed correlation with North Atlantic or Greenland event.

1 from Carter et al. (2008).

2 from Rasmussen et al. (2006).
4. Results and discussion

4.1. The LGIT (20–10 ka)

MAT reconstructions for our three pollen sites are presented in Fig. 2, alongside the CO₂ and deuterium records from the ice core of EPICA Dome C (EDC), Antarctica (Jouzel, 2004) and the δ¹⁸O record from NGRIP Greenland. Deuterium (δD) and δ¹⁸O serve as proxies for temperature and the atmospheric greenhouse gas (CO₂) serves as a climatic amplifier. At EDC, the δD curve shows pronounced accelerated warming from ca 18–17 cal. ka following the end of glacial conditions in Antarctica (ca 19 cal. ka). Deglacial warming culminates in a peak immediately prior to the ACR/AIM1 at ca 14.5 cal. ka. In Greenland, the early deglacial warming seems at first to be in step with that in Antarctica but from ca 17 cal. ka, accelerated warming in Antarctica coincides with temperature decline back to glacial levels in Greenland. This phase of cooling in Greenland is correlated by EPICA with the H1 event during which MOC in the North Atlantic was nearly, or completely, shutdown (McManus et al., 2004). Climate proxy records from the North Atlantic region indicate highly variable climate prevailed during this period, prompting Denton et al. (2006) to refer to it as a ‘mystery interval’ (see also Broecker and Barker, 2007). They suggested that the observed variability may be explained by changes in seasonal climate with both warmer summers and harsher winters in the North Atlantic region. The New Zealand pollen records suggest deglacial warming commenced ca 18 cal. ka and they show progressive warming through to 14.5 cal. ka at least (Fig. 2). In the mid-high latitudes of the Southern Hemisphere this period was a time of reduced windiness indicated by reduced dust flux or polewards contraction of the westerly vortex relative to the preceding glacial (Alloway et al., 1992; Hesse, 1994; Röthlisberger et al., 2002; Newnham et al., 2003) and reduced upwelling in the Southern Ocean (Anderson et al., 2009). This warming phase broadly coincides with progressive increase in atmospheric carbon dioxide and methane (Minnin et al., 2001) which lagged the initial deglacial warming indicated in most Antarctic records by ca 1–2 kyr (Fig. 2). The apparent correspondence between the MAT estimates derived from the New Zealand pollen record and atmospheric carbon dioxide is essentially maintained throughout the LGIT (Fig. 2). Subtle offsets in the initiation of warming from south to north in New Zealand reflects the progressive development of more woody vegetation throughout the landmass and in particular a marked expansion of podocarp forest in northern and central regions (Newnham et al., 2003).

The period ~17–15 cal. ka, corresponding in time to Heinrich event H1 and the so called “mystery interval” (Broecker and Barker, 2007; Denton et al., 1999), is referred to as ‘variable’ in the NZI-CES. The evidence is based largely on marked fluctuations in speleothem isotopic records that are not reflected in the pollen-temperature reconstructions presented here (Williams et al., 2005, 2010; Alloway et al., 2007). Possibly the speleothem isotopic records were detecting seasonal differences akin to the North Atlantic records (Denton et al., 1999) that are concealed by the annually averaged temperature estimates derived from the pollen. Alternatively the variability in the speleothem isotopic records may be unrelated to temperature (for example, a wide scatter of points around a general trend in a composite speleothem record may be an artefact of the method rather than a reflection of real climatic variability: Williams et al., 2010).

The recent emergence of new multi-proxies that are sensitive to seasonal climate in the terrestrial realm (Vandergoes et al., 2008) or may provide seasonality information in the marine realm (e.g., Sikes et al., 2002, 2009; Carter et al., 2008) offer potential opportunity to resolve this question. Southern Hemisphere marine records typically show a progressive sea surface warming through this period (e.g., Pahnke et al., 2007) with no obvious response to H events, although there appears to be a tightly coupled productivity response which may reflect MOC (Sachs and Anderson, 2005) and wind speeds

Fig. 2. Reconstructed mean annual temperature (MAT ± 1.5 °C) for three pollen sites compared with CO₂ and deuterium (δD) from EPICA Dome C core (EDC), Antarctica (Jouzel, 2004) and δ¹⁸O from NGRIP (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006). Age models for Okarito and Pukaki have been revised by calibrating with INTCAL09 and converted to yr b2k. Kaipo age model is that of Hajdas et al. (2006) converted to yr b2k. Arrows on pollen record depict interval of reversal of lateglacial warming inferred for each site from the temperature curves presented here and the original pollen diagrams. For comparison, horizontal bands indicate events from the North Atlantic INTIMATE event stratigraphy (Lowe et al., 2008b). GI-1 is broadly equivalent to Antarctic Cold Reversal (ACR).
(Anderson et al., 2009) more strongly than temperature. The variability in the speleothem isotopic records provides a tantalizing terrestrial counterpart but the sum of the data is clear: during H1, climate trends in New Zealand were aligned very closely with Antarctica and were largely out of phase with Greenland.

The H1/AIM1 event is succeeded by a short rapid warming event in Greenland (GI1/DO1) and in Antarctica by a longer and more subdued cooling (the Antarctic Cold Reversal; ACR). Because the ACR signal varies between different Antarctic records and displays a relatively low-amplitude signal, it has been difficult to define its boundaries and to identify possible correlates to the considerably more distinctive fluctuations in Greenland climate that follow GI1, culminating in the Younger Dryas (YD) or Greenland Stadial (GS) 1 (ca 12.9–11.7 cal. ka). Thus the structure of the GI1/DO1 event and subsequent temperature oscillations do not reflect a simple anti-phased relationship with events in Antarctica.

The three key NZ pollen-temperature records have considerable inter-sample and inter-site variability through this period but overall trends are similar. The progressive warming shown during H1 time is attenuated and then reversed, with all three records showing a minor cooling with variable timing during the ACR/YD interval, lasting ca 1.0–1.5 ka. We refer to this minor cooling as the Lateglacial Reversal. By earliest Holocene times (11.7 cal. ka), strong, progressive warming has returned at all three sites. Although the trends are similar there are some subtle yet significant differences in the timing and amplitude of the Lateglacial Reversal between the three sites. This is perhaps not surprising given that they span c. 5° latitude and range from situations near current sea level (Okarito, Pukaki) to ca 1000 m a.s.l (Kaipo). At Okarito, the southernmost of the three sites, deglacial warming culminates in a marked peak at ca 15 cal. ka followed by cooling of MAT by up to 2 °C that persists for ca 1.5 ka. The climate reversal is evident in the pollen-vegetation record as expansion of subalpine shrubs Coprosma, Myrsine and Halocarpus while the montane tree Metrosideros declined. The reversal is also evident in a lithostratigraphic change to coarser, lighter-coloured silts, suggesting associated catchment disturbance (Newnham et al., 2007a). At this southern site which sits within the sphere of Southern Ocean waters, the cooling trend seems closely aligned with the ACR. Warming is indicated for the entire GI1 interval (12.9–11.7 cal. ka) and continues into the early Holocene, although shorter, lower amplitude cooling reversals are evident in this period as well. Recent pollen-summer temperature reconstructions from Campbell Island in the Subantarctic islands to the south of New Zealand show a very similar pattern of temperature change across the LGIT (McGlone et al., 2010).

At Kaipo Bog, on the North Island well within subtropical water influence, some of the inter-sample variability can be attributed to impacts on vegetation following volcanic activity as discussed by Newnham and Lowe (2000) and Hajdas et al. (2006). When these are accounted for, minor cooling in MAT of 0.5–1.0 °C is apparent from ca 13.6 to 12.6 cal. ka and succeeded by progressive warming through much of the YD interval. The temperature reversal is evident in the pollen-vegetation record as expansion of alpine grasses and other herbs relative to forest trees and by a lithostratigraphic change from peat to less organic peaty mud (Newnham and Lowe, 2000; Allaway et al., 2007). Similar palynological changes are evident in another central North Island montane pollen site, Otamangakau (Turney et al., 2003), also recognised as a key pollen record by NZ-INTIMATE (Allaway et al., 2007). At Otamangakau, the cooling signal approximately coincides with deposition of the Waiohau Tephra (erupted ca 13.8 cal. ka) whereas at Kaipo cooling began several centuries after its fall (Lowe et al., 2008a). At Pukaki, the northernmost site considered here, the cooling reversal is the weakest of the three sites with maximum estimated MAT decline of ca 0.5 °C and commences later, at ca 12.8 cal. ka. In the pollen-vegetation record, temperature decline is evident as a minor expansion of grass pollen with no observed accompanying lithostratigraphic change (Sandiford et al., 2003).

In summary, the southernmost pollen site has a temperature profile that corresponds closely with Southern Ocean records that clearly record the ACR (e.g., Pahake et al., 2003; Anderson et al., 2009). In contrast, the northernmost and lowland pollen site has a weakly registered reversal that is more closely aligned in time with the YD chron; whilst the upland North Island site at an intermediate latitude shows a cooling reversal that commences in mid or late ACR time and overlaps with the first few hundred years of YD time. The temperature reversals at the northern sites (Kaipo and Pukaki) are broadly consistent with patterns evident in sea surface temperature reconstructions offshore which show a muted SST response and cooling that lags the initiation of the ACR (Samson et al., 2005; Pahake and Sachs, 2006; Carter et al., 2008). Although the pollen-temperature reconstructions have broad error terms (~1.5 °C), the estimated mean temperature changes are consistent with qualitative interpretations of the pollen records at these sites and with lithostratigraphic change consistent with deteriorating catchment conditions at Kaipo (Newnham and Lowe, 2000) and Okarito (Newnham et al., 2007b). The timing and direction of temperature changes are unequivocal.

Recently the base of the Holocene has been redefined, based on the abrupt warming that followed GS1 in Greenland Ice cores at 11,700 yb2k (Walker et al., 2009). The working group that proposed this redefinition acknowledged that many Pleistocene–Holocene transition records from the Southern Hemisphere mid-high latitudes appear to be antiphased with climate trends from the North Atlantic region. They typically show a progressive warming that begins earlier than 11,700 yr b2k rather than an abrupt warming step at that time. The three pollen-temperature reconstructions from New Zealand illustrate this asynchronous pattern whilst the differences in timing of the lateglacial reversal between the sites additionally demonstrate the problem of determining a global stratotype section and point (GSSP) for the Holocene boundary that is globally applicable (Walker et al., 2009).

These observations are interesting in light of recent debate over patterns of the lateglacial climate change depicted in other terrestrial records from New Zealand. In broad terms, various records have been interpreted as indicating either progressive uninterrupted warming (McGlone, 1995; Singer et al., 1998) or warming interrupted to some extent by a cooling reversal broadly aligned with the ACR but with timing that varies between records (Newnham and Lowe, 2000; Turney et al., 2003; McGlone et al., 2004; Williams et al., 2005; Vandergoes et al., 2008). What is clear is that any lateglacial temperature reversal was of low amplitude in comparison with the Last Glacial Maximum (LGM) temperature depression and that some of the inter-site variability could be due to differences in the climate sensitivity of vegetation and other proxies (McGlone, 1995; Newnham, 1999) or to changes in seasonal climate (Vandergoes et al., 2008). The quantified temperature reconstructions presented here suggest that latitudinal control may be an important additional explanation for this variability, and the extent to which any New Zealand records ‘senses’ an Antarctic signal.

Debate has also centred on the timing and climate interpretation of various post-LGM moraines in southern New Zealand (Denton and Hendy, 1994; Barrows et al., 2007; Applegate et al., 2008). More recent exposure age dating has indicated a more consistent pattern of ice readvance during ACR time and retreat during YD time, however (Turney et al., 2007; Kaplan et al., 2010; Putnam et al., 2010). The glacial evidence is consistent with the timing, trend and estimated amplitude of lateglacial temperature
change at the southernmost pollen record presented here (Okarito), situated within this formerly glaciated region.

It is interesting to consider these patterns in the context of observations by Carter et al. (2008) based on the comparison of sea surface temperature records (SST) in two marine cores from east of New Zealand (Pahnke and Sachs, 2006). Their analysis of multiple proxies in addition to SST reconstructions indicate a different response from north and south of the Subtropical Front (STF; Fig. 1): at the Southern Ocean site south of the STF, the SST profile shows a distinctive ACR-like pattern, similar to the pollen-based temperature curve at Okarito (Fig. 2). In contrast, in subtropical waters, north of the STF, both SST and the levels of tropical and subtropical forams increased during ACR time with no clear temperature reversal (Carter et al., 2008) although the rate of deglacial warming decreased. Carter et al. (2008) suggested this apparent contradiction may be explained by the interaction of two opposing forces: enhanced delivery of subtropical waters co-occurring with cooling temperatures related to the ACR. Several recent studies suggest that ACR cooling forced a northward expansion of cold atmospheric and ocean temperatures (Russell et al., 2006; Toggweiler et al., 2006) at the same time as equatorial regions continued to warm. As a result, the thermal gradient between pole and equator increased which in turn promoted the northwards migration of more vigorous westerlies and thus brought stronger winds to the New Zealand region (Carter et al., 2008). On the basis of speleothem isotopic records from northwestern South Island of New Zealand, Whittaker et al. (2011) also proposed intensification of westerlies during the ACR due to regional climate cooling forcing northwards migration of the polar and subtropical fronts. They suggested a subsequent change to drier climate within the YD chron resulted from a southwards displacement of the ITCZ, and was shorter (< 400 yr) and smaller in magnitude than is recorded in Northern Hemisphere climate records. Although their argument calls for transmission of interhemispheric climate signals via the atmosphere rather than the oceans, the close coupling of atmospheric and oceanic circulation systems makes it feasible that both processes were involved.

Similar evidence for northwards amelioration or loss of an ACR signal have long been inferred from δ18O planktic foraminiferal records from the Southern Atlantic (Kanfoush et al., 2000) and Southern Pacific (Lamy et al., 2004; Kiefer and Kienast, 2005) (Sachs and Anderson, 2005). The terrestrial New Zealand evidence is also consistent with latitudinal variations in patterns of LIGT climate reported for western Patagonia, whereby a stronger resemblance to Antarctic records are observed in records to the south of the Antarctic Polar Front than in records to the north of that boundary (Moreno et al., 2010).

4.2. Pre-LGIT events (30–20 cal. ka)

We now turn to the interval 30–20 cal. ka, during which two prominent cycles of abrupt climate events are highlighted in the EPICA bipolar ice core comparison (Fig. 3). At ca 30–29 yr b2k, Greenland records consistently show stadial conditions. This interval has been correlated with H3 in the North Atlantic and with AIM4, the last of a series of prominent interstadials evident in Antarctic records (EPICA, 2006). In Greenland, this stadial is followed by two short lived (200–300 year) interstadials: GI3, (commencing 27,780 ± 416 ye b2k) and GI4 (28,900 ± 449 yr b2k) (GICC05 chronology: Andersen et al., 2006). In Antarctica (EDML) the interval 29–27 yr b2k is characterised by prominent temperature decline that essentially marks a shift to stadial conditions. By ca 25 yr b2k, temperatures at Antarctica were at their lowest levels for the entire Last Glaciation.

In the New Zealand region, although there are several SST reconstructions, few quantitative terrestrially based climate estimates are available and the preliminary NZI-CES for this period (Fig. 3) is largely based on inferred climate reconstruction. Strong emphasis is placed on well-developed but fragmentary records, in particular geomorphic evidence for glacier advance and retreat from southwestern New Zealand (Suggate and Almond, 2005; Alloway et al., 2007). These records consistently show a deteriorating climate commencing around 30–29 cal. ka, evidenced by advancing glaciers, renewed accumulation of loess deposition, fluvial aggradation, declining sea surface temperatures, and replacement of forest with non-forest communities (Alloway et al., 2007). These conditions persisted with some fluctuations into the LGM (24–18 cal. ka) but as the glacial conditions commenced earlier in New Zealand, terms such as Last Glacial Cold Period (LGGP; Alloway et al., 2007) and extended LGM (eLGM; Newnham et al., 2007b) have been used to emphasise the distinction. The onset of

Fig. 3. Reconstructed MAT (±1.5 °C) for Okarito (30–16 cal. ka) with new age model based on INTCAL09 and converted to yr b2k in relation to timing of ice core and ocean events (Lowe et al., 2001; EPICA, 2006); the preliminary New Zealand Intimate Climate Event Stratigraphy (Alloway et al., 2007); deuterium (δD) from EPICA Dome C core (EDC), Antarctica (Jouzel, 2004); and δ18O from NGRIP (Andersen et al., 2006; Svensson et al., 2006).
this extended LGM period at ca 29–28 cal. ka coincided with cooling observed in Antarctica and with interstadial conditions in Greenland, consistent with the bipolar model (Fig. 3). Thus the timing of eLGM onset in New Zealand (and Antarctica) relative to Greenland may have been accelerated by the warming of GI3 and 4 in the North Atlantic region and by an antiphased cooling response in the Southern Hemisphere, analogous in an opposite sense to the early onset of the Last Termination in the Southern Hemisphere discussed earlier. If so, the implication is that D–O cycles may cause an apparent offset in the relative inter-hemispheric timing of glacial onset and termination evidenced in polar ice cores (Wolff et al., 2009) and perhaps in other proxy records.

An additional prominent abrupt climate cycle event occurs during the interval (ca 25–23 cal. ka). This interval is characterised by interstadial conditions (AIM2) in the EDML core and by stadial conditions (H2) generally in the North Atlantic region. In the NGRIP core, the interval is abruptly terminated by GI2 at ca 23 yr b2k. In New Zealand a period or periods of warmer, more variable conditions within the eLGM, has/have been evidenced by sedimentary profiles from glaciated terrain east and west of the Southern Alps that show soil development between glacial advances (Suggate and Almond, 2005). The NZI-CES describes the interval 27.0–21.0 cal. ka as a mid-late Glacial Cold Period ‘warming complex’ with two periods of warmer and more variable climate at ca 27.0–24.5 and 23.0–22.0 cal. ka. More recently, Williams et al. (2010) reported a mid-LGM interstadial at 23 to 21.7 cal. ka from speleothem records. When the preliminary NZI-CES is compared with polar ice core and Heinrich events for the interval 25–22 cal. ka (Fig. 3) an antiphased relationship is not evident. The interval encompassing AIM2/H2, ca 25–23 cal. ka is designated as a cooling phase within the mid-LGCP ‘warming complex’, whilst the onset of GI2 coincides with the onset of a warming interval in the NZI-CES at 23 cal. ka. Considerable variability is evident, however, between key records used to designate the ‘warming complex’ and this, together with imprecise dating of the timing of key events in the earlier part of the preliminary NZI-CES (Alloway et al., 2007; Lowe et al., 2008a) suggests that inter-hemispheric comparisons for this interval should be made with caution.

The Okarito pollen-temperature record illustrates this uncertainty (Fig. 3). As with other records from New Zealand, dating control for this time interval is comparatively imprecise (Vandergoes et al., 2005; Newnham et al., 2007a). For the period 30–26 cal. ka, this record is broadly consistent with the preliminary NZI-CES in showing temperature decline into the LGCP in antiphase with the Northern Hemisphere climate events (Fig. 3). A subsequent warm interval, marked by a phase of lowland forest development from ca 26 to 22 cal. ka, is broadly correlative with the mid-LGCP ‘warming complex’ of the preliminary NZI-CES. This warmer interval encompasses the AIM2/H2 interval of 25–23 yr b2k, but the correspondence is not precise. The comparatively wider error margins in the Okarito chronology make it difficult to argue for either a Greenland or Antarctic template for this event but the latter cannot be discounted.

The Okarito record is complemented by three palynological records from sites around Auckland near the site of the Pukaki core (Fig. 1) (Newnham et al., 2007b). These records, linked with Kawakawa/Oruanui tephra, have dating control from radiocarbon and tephrochronology (Newnham et al., 2007b). Their strong similarity with one another in timing and patterns of change underpin their usefulness as regionally representative records for the period. As well as demonstrating an earlier glacial maximum onset in New Zealand, the Auckland records show two phases of cold climate separated by a milder interval (26–24.5 cal. ka). The latter is broadly correlative with the warming interval coarsely dated to 26–22 cal. ka in the Okarito site. A climate amelioration within the eLGM has also been observed in several palaeoecological records obtained at Lyndon stream, Canterbury (Soons and Burrows, 1978; Woodward and Shulmeister, 2007), as well as in records of aeolian quartz flux from Taranaki (Alloway et al., 1992) and in the glacial moraine record from Westland (Suggate and Almond, 2005). Thus it appears to be widespread in the New Zealand region where suitable proxies have been reported (Alloway et al., 2007; Newnham et al., 2007b). Such a period of warming has been supported by research in eastern Australia (Petherick et al., 2008).

In summary, a test of the bipolar model for the interval 30–20 cal. ka is limited by imprecision in the dating of New Zealand records and by the comparative scarcity of climate-proxy records. Nevertheless, two features of this period are consistent among the available records. First, the New Zealand region was thrust into an extended glacial maximum period at ca 29–28 cal. ka when Greenland records are dominated by the interstadial peaks of GI3 & 4. Second, the mid-LGM warming complex evident in many New Zealand records (Alloway et al., 2007; Newnham et al., 2007b; Williams et al., 2010) and possibly eastern Australia (Petherick et al., 2008) coincides broadly with a period of cooling in the North Atlantic region and warming in Antarctica. Although some inter-site variability is evident and chronologies lack precision, the climate records are consistent, widely registered and demand a unifying explanation. With current data, it is difficult to reject the extended bipolar hypothesis although, clearly, there needs to be a focus on improving the chronology of existing and new records through this interval so that a more robust test can be attempted.

At this point it is necessary to acknowledge that we have considered just one postulated mechanism for generating asynchronous inter-hemispheric patterns of climate change, and that other plausible mechanisms may be invoked to generate similar patterns. For example, over 130 years ago, James Croll suggested that an antiphased climate response between hemispheres would have resulted from asynchronous inter-hemispheric patterns of seasonal insolation operating at orbital scales. More recently, several authors have invoked a precession-dominated signal such as local summer insolation to explain patterns of climate change (e.g., Kim et al., 1998; Kale Sniderman, 2007) including divergence between New Zealand and Northern Hemisphere records (Vandergoes et al., 2005). Other proposed mechanisms linking Northern and Southern hemisphere climates at times of abrupt change include the displacement of large scale atmospheric circulation features, in particular the Intertropical Convergence Zone and the southern westerlies (Clement and Peterson, 2008; Whittaker et al., 2011) or changes in the seasonality of atmospheric heat transport, also tied to the shifting westerlies. McGlone et al. (2010) suggest that this latter phenomenon may have played a similar role in both hemispheres which would imply a synchronised rather than a synchronised pattern.

5. Conclusion

Our comparison of key New Zealand and polar ice core records across a series of abrupt climate cycles during the interval 30–10 cal. ka provides general support for a persistent antiphased inter-hemispheric relationship. The patterns of climate change evident in New Zealand thus are broadly consistent with an extended bipolar seesaw whereby the oceanic southern mid-latitudes are warmed at times of MOC weakening or cessation in the North Atlantic, and vice versa, although variability between records indicate that other factors must be involved. Alternative models that predict an antiphased inter-hemispheric pattern cannot be refuted from the New Zealand evidence, whilst mechanisms involving transfer of heat through the atmosphere, implying a synchronised inter-hemispheric response, may also play a role.
Nevertheless an extended bipolar model provides one possible explanation for an early onset of LGM conditions in New Zealand and elsewhere in the Southern Hemisphere at a time when interstadials GI3 and GI4 delayed the onset of glacial maximum conditions in Greenland ice core records – and perhaps in other records from the North Atlantic region. Similarly an earlier termination of the LGM in Antarctica and the Southern Ocean than in Greenland from the North Atlantic region. By comparison, New Zealand records for the LGIT are more numerous and more precisely dated. For this period, there is greater confidence in developing a picture of inter-hemispheric asynchronous supportive of the bipolar model, as has been suggested previously (Tunney et al., 2003). The caveat in this case is that variability in the timing and amplitude of events in the New Zealand climatic record may reflect complex circulation patterns that cannot be simply explained by chronological imprecision. One factor contributing to this variability may be linked to latitudinal controls on the extent to which the Antarctic pattern is registered, as has been suggested from other records from the southern mid-high latitudes (Carter et al., 2008; Moreno et al., 2010). From the limited dataset presented here, we show that the southermost site has a temperature profile that corresponds more closely to an Antarctic pattern than the two North Island sites which show a later and more subdued latitudinal temperature reversal.

Clearly, more precisely-dated climate reconstructions from sites spread across latitudinal gradients in both the marine and terres-

trial realms are needed to test this new hypothesis against alter-

native hypotheses and to explore further the mid-latitude limits for which an extended bipolar model might apply. If this model is upheld, it may in addition go some way towards explaining the observed variability between latitudinal climate records from the New Zealand region.

Acknowledgements

We thank Alan Hogg (Waikato University) for valued assistance with radiocarbon calibrations, Max Oulton (Waikato University) for drawing figures and Chris Tunney and an anonymous referee for insightful comments that helped improve the manuscript.

References


Peterson, R.E.,下面就继续回答您的问题。