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Cooling and changing seasonality in the Southern Alps, New Zealand during the Antarctic Cold Reversal

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Abstract

A comprehensively ¹⁴C AMS dated pollen and chironomid record from Boundary Stream Tarn provides the first chironomid-derived temperature reconstruction to quantify temperature change during Lateglacial times (17,500–10,000 cal yr BP) in the Southern Alps, New Zealand. The records indicate a ca 1000-year disruption to the Lateglacial warming trend and an overall cooling consistent with the Antarctic Cold Reversal (ACR). The main interval of chironomid-inferred summer temperature depression (~2–3 °C) lasted about 700 years during the ACR. Following this cooling event, both proxies indicate a warming step to temperatures slightly cooler than present during the Younger Dryas chronozone (12,900–11,500 cal yr BP). These results highlight a direct linkage between Antarctica and mid-latitude terrestrial climate systems and the largely asynchronous nature of the interhemispheric climate system during the last glacial transition. The greater magnitude of temperature changes shown by the chironomid record is attributed to the response of the proxies to differences in seasonal climate with chironomids reflecting summer temperature and vegetation more strongly controlled by duration of winter or by minimum temperatures. These differences imply stronger seasonality at times during the Lateglacial, which may explain some of the variability between other paleoclimate records from New Zealand and have wider implications for understanding differences between proxy records for abrupt climate change.

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1. Introduction

The Antarctic Cold Reversal (ACR, 14,500–12,500 cal yr BP; EPICA Community Members, 2006) interrupted an otherwise progressive deglacial warming trend in most Antarctic ice core records (Petit et al., 1999; Blunier and Brook, 2001; Jouzel et al., 2001; EPICA Community Members, 2006). The ACR occurred during a time of marked climate fluctuations in the Northern Hemisphere that preceded the strong cooling in the North Atlantic region of the Younger Dryas (YD) chronozone (12,900–11,500 cal yr BP; Alley, 2000). However, the extent to which ACR patterns occurred through-

out the Southern Hemisphere beyond the Antarctic region remains unclear (Heusser, 1998; McCulloch et al., 2000, 2005; Moreno et al., 2001; Hajdas et al., 2003; Mortyn et al., 2003; Lamy et al., 2004; Sugden et al., 2005; Barrows et al., 2007). In New Zealand, the debate as to the existence of an ACR-like cooling has not been resolved (Alloway et al., 2007). For example, recent New Zealand evidence has been interpreted as showing a southern (ACR-like) pattern (Turney et al., 2003; McGlone et al., 2004), a northern (Younger Dryas-like) signal (Denton and Hendy, 1994), or a climate signal intermediate between the ACR and the YD (Newnham and Lowe, 2000; Williams

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et al., 2005; Hajdas et al., 2006). Differences among these records may be attributed to various factors including the extent to which these climate events influence sites that extend across a broad latitudinal range, chronologic control of the stratigraphic sequences, and sampling resolution. However, these differences may also involve the individual response of proxies to climate change. Fluctuations in climate parameters such as precipitation, temperature minima and maxima, and changes in the seasonal variability of climate patterns (seasonality) are all likely to influence biota in different ways. A multi-proxy approach, involving more than one biotic element, therefore, can potentially provide greater insight into the response of different proxies to climate change; the detail of past climate change; and hence promote understanding of their causes.

This paper presents a ^{14}C Accelerator Mass Spectrometry (AMS)-dated, high resolution, multi-proxy paleoecological record (pollen and chironomid) that makes use of a recently developed chironomid-inferred temperature model (Dieffenbacher-Krall et al., 2007) to derive and quantify Lateglacial climate change at Boundary Stream Tarn in the Southern Alps, New Zealand. The results, indicating significant differences in seasonal climate patterns superposed on a predominantly Antarctic-like signal, provide new perspectives on the Lateglacial transition in the southern mid-latitudes and may help explain some of the complexity observed in Lateglacial records from the southern mid-latitudes.

2. New Zealand climate reversal during the Lateglacial

Evidence for climate reversal during the Lateglacial transition in New Zealand has emerged only recently. The Southern Alps have a widespread system of Lateglacial moraines (e.g. Waiho Loop/Canavans Knob, Birch Hill, Prospect Hill, and Arthur's Pass). Limiting minimum ages of as much as 12,400 (Be^{10}) years come from both the Birch Hill moraines at Lake Pukaki and the Lake Misery moraines at Arthur's Pass (Ivy-Ochs personal communication, Ivy-Ochs et al., 1999). Numerous radiocarbon (^{14}C) dates of wood from the base of a till sequence at Canavans Knob suggest ice advance toward the Waiho Loop moraine at 11,100 ^{14}C yr BP (13,000 cal yr BP; Denton and Hendy, 1994) while Be^{10} ages from the moraine crest are interpreted as indicating ice retreat from this feature during the early Holocene ($10,480 \pm 240$ cal yr BP; Barrows et al., 2007). From these ages, it is not yet clear if the Lateglacial advance occurred at the beginning of the Northern Hemisphere YD or the end of the ACR in the south. Recent modelling of the Franz Josef Glacier (Anderson and Mackintosh, 2006) suggests that a 3–4 °C cooling is required to advance the glacier to its Lateglacial limits at the Waiho Loop moraine. High-resolution $\delta^{18}\text{O}$ records from speleothems also indicate environmental change between ca 13,800 and 11,140 cal yr BP with a positive $\delta^{18}\text{O}$ excursion that provides supporting evidence

for regional ice advance (Hellstrom et al., 1998; Williams et al., 2005). A Lateglacial temperature reversal of $\sim 2^\circ\text{C}$ between ca 14,000 and 13,000 cal yr BP is also indicated in a marine record off the east coast of the South Island (Pahnke et al., 2003; Pahnke and Zahn, 2005).

Evidence of a cooling reversal during the Lateglacial transition also comes from pollen records located in climatically sensitive areas such as at high elevation, or at an ecotonal boundary where a relatively minor climatic change has significant impact. These records, including those from Kaipo Bog (Newnham and Lowe, 2000; Hajdas et al., 2006), Otamangakau and Kettlehole Bog (Turney et al., 2003; McGlone et al., 2004), show an interval of about 1000 years when percentages of tree/shrub pollen declined and herbaceous pollen increased. Constraining radiocarbon ages place these events between 13,800 and 12,400 cal yr BP at Kaipo Bog, 14,700 and 13,000 cal yr BP at Otamangakau, and 14,600 and 13,300 cal yr BP at Kettlehole Bog. The records show discernable shifts in pollen assemblages that largely antedate or barely overlap the early YD; hence they may be comparable to cooling initiated during the ACR. At other New Zealand pollen sites not situated at high altitude or ecotonal boundaries, Lateglacial reversal of deglacial warming is either minor (e.g. Newnham et al., 2007a,b) or not apparent (e.g., McGlone, 1995; Vandergoes and Fitzsimons, 2003).

3. Site significance and description

The Pukaki Basin in the central South Island (Fig. 1) lies directly south of the highest ranges in the New Zealand Southern Alps. These mountains intercept the prevailing westerly air flow to produce a strong west-to-east precipitation and temperature gradient and a characteristically dry climate to the east of the ranges. These climate gradients are sensitive to subtle changes in the circumpolar westerlies, which are directly linked to Antarctic air masses (Chinn, 1999; Lamont et al., 1999; Clare et al., 2002). Paleoclimate proxies from sites within this area should be highly sensitive to shifts in precipitation and temperature associated with past changes in intensity and latitude of the westerly wind systems.

The eastern central South Island generally has warm summers and mild winters. Mean annual air temperatures are about 8.8 °C (Mt. Cook township and Tekapo climate stations), with February (summer) and July (winter) means ranging between 15.1 and 1.7 °C, respectively (New Zealand National Climate Centre, 2007). Extreme temperatures are between 33.3 and -15°C in the area. Mean annual precipitation ranges from 4293 mm close to the Southern Alps (Mt. Cook township) to 600 mm some 30 km to the south east of the Southern Alps at Tekapo.

The geomorphology of the Pukaki Basin is dominated by glacial moraines and outwash plains of the last glaciation and by Holocene alluvial fans. Both Last Glacial Maximum (LGM) and Lateglacial moraines are common

Fig. 1. Location of study area in South Island, New Zealand and other sites mentioned in the text. Boundary Stream Tarn (star) in the Pukaki Basin; 1, Mt. Cook township; 2, Tekapo; 3, Birch Hill; 4, Canavans Knob; 5, Prospect Hill; 6, Kettlehole Bog; 7, Arthur's Pass; 8a–b, South Island speleothems. For moraine classification see Schaefer et al. (2006).

(Fig. 1; Ivy-Ochs et al., 1999; Schaefer et al., 2006). The present-day vegetation of the basin has been highly modified by natural and recent anthropogenic fires. Pollen evidence indicates that the area was dominated by podocarp scrubland, primarily *Phyllocladus alpinus*, from ca 8000 yr BP until 5000 yr BP, after which *Halocarpus bidwilli* and grassland communities became prolific in the landscape as a result of drought and fire (McGlone and Moar, 1998). The podocarp scrubland has been further reduced to grassland by fire and land clearance by Maori and Europeans over the past 800 years. The present vegetation is tussock grassland (*Chionochloa rigida*, *Poa colensoi*) above the treeline in the alpine zone, and grassland–shrubland (*Festuca novae-zelandiae*, *Discaria tomatou*, *Dracophyllum*, *Hebe*, and *Coprosma* spp.) on the lower slopes and valley floor. Small stands of *Halocarpus*, *Phyllocladus*, and *Nothofagus* forest survive as remnant patches (McGlone and Moar, 1998).

Boundary Stream Tarn (44°02'S, 170°07'E; 830 m elevation) is one of a cluster of small kettle lakes within a lateral moraine complex of LGM age (Schaefer et al., 2006). Aside from ephemeral drainage input from a small neighboring tarn, Boundary Stream Tarn is devoid of major sedimentary or fluvial influence. The kettle lakes are located in the present montane-to-subalpine vegetation zone and should be ideally placed to provide evidence of past fluctuations in tree line and temperature. Mean annual precipitation at the site, estimated from climate surface models, is 1268 mm (Leathwick et al., 1998). Mean annual air temperature is estimated to be 7.7 °C and mean summer and winter temperatures 13.4 and 2.8 °C, respectively.

4. Methods

To reconstruct and quantify Lateglacial climate change, fossil chironomid larval remains were analyzed from overlapping sediment cores recovered from Boundary Stream Tarn. Pollen analysis was also conducted to provide an independent proxy of Lateglacial vegetation and climate change and to evaluate the structure and timing of changes produced by the chironomid record. Four sediment cores were recovered with a square rod, piston corer from the deepest area of the lake in proximity to each other. All four cores show a consistent stratigraphy. Two cores (0201 and 0408), that represent the entire Lateglacial sequence, were sampled for pollen and chironomids and used to establish chronology. Samples were taken at 8 and 4 cm intervals throughout the cores to provide the basic pollen and chironomid stratigraphy for the site. Sections showing marked stratigraphic change, radiocarbon dated to the time of the ACR, were re-sampled at 1 and 2 cm intervals for detailed analyses.

Pollen extraction was carried out using standard laboratory techniques (10% hot KOH, 40% HF, and acetolysis; Faegri et al., 1989). Samples were spiked with exotic *Lycopodium* spores for calculation of pollen concentrations. Pollen identifications were made using standard texts (e.g. Pocknall, 1981a–c; Large and Braggins, 1991; Moar, 1993) and the New Zealand pollen reference collection held at the University of Plymouth. Grass pollen grains (Poaceae) were divided by size, greater than and less than 40 µm, to distinguish between tall tussock *Chionochloa* grassland (>40 µm) components and short

tussock (<40 µm) grassland components (McGlone and Moar, 1998; McGlone, 2001). For each sample, pollen taxon values were expressed as percentages of a minimum pollen sum of 250 grains. This sum included pollen from all dryland plants except pteridophytes.

Chironomid samples were prepared using standard laboratory techniques (5% warm KOH, sieved through 90 µm mesh; e.g. Walker et al., 1991; Brooks and Birks, 2000). Chironomid remains were picked from water using a modified Bogorov tray, placed in water drops on glass slides, and fixed using Pro-Texx mounting medium following the evaporation of water. Chironomid taxon values were expressed as percentages of a minimum sum of 50 head

capsules from each sample. Chironomid identification followed Forsyth (1971), Boothroyd (1994, 1999, 2002, 2004, unpub.), Stark and Winterbourn (2006) and our own identification guide (<http://www.climatechange.umaine.edu/Research/facilities/perl/nzguide.html>). All *Chironomus* specimens had fewer than 60 striae on the ventromental plates (Martin and Forsyth, 2006) and are consistent with the “*Chironomus* spp.” of Woodward and Shulmeister (2006).

We applied the weighted average partial least squares (WA-PLS 3 component) and partial least squares (PLS 3 component) inference models of Dieffenbacher-Krall et al. (2007) to infer mean summer temperature values (Fig. 2). The models have coefficients of determination (r^2_{jack}) of 0.80

Fig. 2. Observed versus predicted temperatures and plots of residual errors for (A) the WA-PLS and (B) the PLS models. A LOWESS smooth (0.45) and rectangle boxes in the residual plots highlight the ranges of best prediction for each model and show the number of lakes used to derive temperature predictions through this range (from Dieffenbacher-Krall et al., 2007).

and 0.79, and root mean square errors of prediction (RMSEP_{jack}) of 1.27 and 1.30 °C, respectively. Both models were applied to the fossil chironomid data from Boundary Stream Tarn to assess the reconstructive power of each model.

4.1. Radiocarbon dates and age model

Seventeen AMS ¹⁴C dates were obtained on organic material in 1-cm-thick slices of bulk sediment (Table 1). All ages have been reported in calibrated years, unless otherwise noted. The age model (Fig. 3a and b) was developed using the Bayesian software Bpeat (Blaauw and Christen, 2005) and contains the following five assump-

tions. The first two assumptions relate to the depositional setting at the site and indicate that: (i) the depositional setting dictates that the dated depths must be ordered chronologically; (ii) the deposition rates of ca 20–80 yr/cm are considered most likely, but other (positive) accumulation rates are possible (priors mean and range 50 respectively 25). Two further assumptions relate to the occurrence of hiatuses and indicate that: (iii) the variable stratigraphy suggests that rapid accumulation rate changes and hiatuses could have occurred (prior variance 2); (iv) the hiatuses were most likely to be very short, but could at times have lasted millennia (priors skewedness and width both 0.0005). The final assumption deals with outlying dates and indicates that: (v) the dates had a prior

Table 1
Radiometric dates of cores 0201 and 0408 from boundary stream

Core	Sub-bottom depth (cm) core 0201	Sub-bottom depth (cm) relative to 0201 from 0408	¹⁴ C lab #	Carbon combusted (mg)	CO ₂ (μmol)	δ ¹³ C	¹⁴ C age	±1σ	Calibrated age range 2σ
0201	308		Wk-10719	Unavailable	Unavailable	−29.9	9991	77	11,239–11,805
0201	316		OS-37971	7.62	119.5	−24.42	10,100	40	11,404–11,968
0201	317		OS-42553	Not weighed	39.2	−28.15	9870	65	11,180–11,601
0201	333		OS-42661	12.7	110.5	−20.73	10,200	45	11,718–12,078
0201	334		OS-37972	13.37	129.1	−21.09	10,400	50	12,077–12,595
0408		360	OS-48412	109.15	247.6	−22.36	10,350	55	11,985–12,665
0408		380	OS-46017	114.22	137.9	−21.9	11,100	70	12,890–13,139
0408		380	OS-45812	42.61	179.8	−22.11	11,150	45	12,939–13,154
0201	382		OS-38191	25.73	116.6	−21.43	11,050	65	12,876–13,093
0201	394		OS-42662	18.74	122.8	−21.86	10,800	60	12,734–12,893
0201	410		OS-38192	34.64	157.3	−22.72	12,150	70	13,827–14,164
0408		410	OS-45813	261.78	275	−22.99	12,300	45	14,012–14,546
0408		410	OS-46018	124.83	144.8	−23.07	12,300	60	13,999–14,611
0408		450	OS-48413	27.2	220.5	−21.63	12,600	55	14,473–15,126
0201	483		OS-42663	21.45	143.1	−20.13	13,200	60	15,288–16,012
0201	528		OS-38193	30.54	195.5	−19.71	14,100	65	16,379–17,183
0201	532		OS-42664	27.19	97.1	−18.93	14,800	65	17,661–18,486

Calibrated age ranges (2σ) are calculated for all ages using INTCAL04 (Reimer et al., 2004). Samples were submitted to the National Ocean Sciences AMS Laboratory, Woods Hole Oceanographic Institution, Massachusetts, USA (OS) and the Radiocarbon Dating Laboratory, University of Waikato, Hamilton, New Zealand (Wk) for conventional radiocarbon analysis. Bulk organic material from core sediment samples were combusted for this analysis.

Fig. 3. (A) Boundary Stream Tarn ages in relation to INTCAL04 radiocarbon calibration curve. (B) Grayscale age model developed for Boundary Stream Tarn. Areas of best prediction are indicated by darkest gray.

probability of 5% of being an outlier. Outlying dates were not removed manually but were identified and down-weighted automatically by the software. We used the INTCAL04 calibration curve (Reimer et al., 2004) to calibrate all ^{14}C dates and corrected for a Southern Hemisphere offset of 41 ± 14 ^{14}C years (McCormac et al., 2002). We applied 12 section breaks to obtain an age model based on linear interpolation between the dated levels. The best age model, with associated uncertainties was constructed by running one billion Monte Carlo Markov Chain iterations (Blaauw and Christen, 2005). The results provide a rigorous and continuous age model that spans 17,500–10,000 cal yr BP.

5. Results

5.1. Sediment stratigraphy

The basal sediment between 530 and 520 cm is composed of stiff gray diamicton (Fig. 4). This diamicton is similar to that in nearby moraine outcrops and is likely to represent underlying moraine. The sediment between 520 and 466 cm is comprised of light brown banded and unbanded silts. Most of the core between 466 and 317 cm features brown organic silt. However, significant changes in stratigraphy occur between 410 and 383 cm and between 317 and 334 cm, where gray-brown and light brown silt layers are prominent. The upper sediment of the sequence between 317 and 280 cm is composed of brown fibrous silt.

5.2. Dating and chronology

The AMS ^{14}C ages, calibrated using INTCAL04, show strong internal consistency (Fig. 3a). Multiple dates from the same horizon above and below the major sediment changes at 410, 380, 333–334 and 317 cm provide near-identical or overlapping dates (2σ range; Table 1). The consistency of these ages tightly constrains these sediment changes in the sequence. Two ages (OS 42553, OS 42662) were identified as outliers by the modelling analysis (Table 1, Fig. 3a). These ages were not removed from subsequent analysis but their importance for constructing the age model was down-weighted. The age model shows a strong predictive fit between the ages in the sequence (Fig. 3b) and produces robust estimates for the timing of reconstructed environmental change.

5.3. Pollen and chironomid stratigraphy

Pollen of subalpine shrubland and alpine grassland (Poaceae, Asteraceae, *Myrsine*, and *Coprosma*) dominate the sequence from 17,500 to 14,200 cal yr BP, with grassland pollen being most abundant from 17,500 to 15,500 cal yr BP (Fig. 4). Chironomid analysis reveals a mixture of taxa between 17,500 and 14,200 cal yr BP. Taxa widespread in cool to warm lakes today (*Tanytarsus*

funnebris types A-I and C-I) are common as well as *Chironomus*, which is dominant in high-altitude, low-temperature lakes today (Dieffenbacher-Krall et al., 2007).

The interval between 14,100 and 13,100 cal yr BP shows distinctive changes in the pollen and chironomid records, as well as in the sediment stratigraphy. This interval is marked by the increase in tall *Chionochloa* alpine tussock (Poaceae $>40\mu\text{m}$) and grassland-open herbfield pollen (Poaceae $<40\mu\text{m}$, Asteraceae), and decrease in low shrubland species (*Coprosma*). The increase in alpine tussock pollen and associated grassland-herbfield pollen changes are indicative of expanding alpine environments and a return to conditions similar to those experienced before 15,000 cal yr BP when grassland pollen levels were at similar levels. A change in sediment stratigraphy from organic silt to alternating layers of organic and inorganic silt occurs at this time and coincides with a marked increase in aquatic pollen spores of *Isoetes* and *Botryococcus*. The interval between 14,100 and 13,100 cal yr BP also displays significant change in the chironomid assemblage with elevated levels of taxa more tolerant of low temperatures, notably *Chironomus* and *Tanytarsus funnebris*-type C-s, and wide temperature ranges (*Apsectrotanytus*) which may become more abundant where less tolerant species decline.

The first appearance, decline, and subsequent increase of *Corynocera* and *Paucispinigera* taxa more commonly associated with lower altitude temperate environments (Dieffenbacher-Krall et al., 2007), also occurs between 14,100 and 13,100 cal yr BP.

The decline in relative abundance of grassland-open herbfield pollen beginning at 13,100 cal yr BP is followed by increases in pollen of the shrub conifers (*Halocarpus*, *Phyllocladus*) at 12,800 and 11,900 cal yr BP, respectively. These pollen changes imply replacement of open grassland and shrubland communities by woody vegetation near the site. A decline of cold tolerant chironomids and an increase of warm temperate types *Corynocera* and *Poly-pedilum* occurred between 13,200 and 13,100 cal yr BP. *Corynocera* becomes a dominant component of the assemblage after 13,000 cal yr BP. A sedimentary change from organic to inorganic silt occurs between 12,000 and 11,500 cal yr BP and corresponds with a slight increase in *Tanytarsus funnebris*-type C-I. Increases in tall podocarp pollen (*Podocarpus*, *Prumnopitys taxifolia*) after 11,000 cal yr BP suggest that the succession of conifer forest continued near the site, with tall podocarp trees increasing in importance in the *Phyllocladus* shrub forest. There is little variation in the chironomid record during this time, aside from a slight increase in *Tanytarsus funnebris*-type B.

5.4. Chironomid inferred temperature reconstructions—model evaluation

Both the WA-PLS and PLS (Fig. 4) models produce a nearly identical structure of paleotemperature change during the Lateglacial transition. However, the temperatures

predicted by the WA-PLS predictions are generally 1–2 °C cooler than those of the PLS model. Temperature estimates range from 13.2 to 8.1 °C (± 1.4 °C sample specific error) for the WA-PLS and 14.1–9.5 °C (± 1.6 °C sample specific error) for the PLS model, thus displaying strong overlap within the error ranges of each model. Relative to the mean modern summer air temperature at this site (13.4 °C), reconstructed temperatures reach a maximum of 2.3 °C warmer and 6.7 °C cooler between 17,500 and 11,000 cal yr BP.

Both reconstructions show generally cool temperatures between 17,500 and 15,500 cal yr BP, followed by warming until 14,200 cal yr BP. A subsequent cooling began after 14,200. By 14,000 cal yr BP the reconstructed temperatures were similar to, or slightly cooler than, those at around 16,000 cal yr BP. This initial cooling was interrupted by a sharp rise in temperature to their highest levels, of Lateglacial times, about 13,900 cal yr BP, followed by the resumption of cooling until 13,200 cal yr BP. A rapid rise in temperature occurred at around 13,200 cal yr BP, after which temperatures increase slightly (WA-PLS model) or cool slightly (PLS model) until ca 11,500 cal yr BP. These differences between model reconstructions for the interval ca 13,200–11,500 cal yr BP are well within the error margins of each reconstruction. Taken together, the two reconstructions suggest a period of comparatively little change in mean summer temperature; however, the individual reconstructed temperature estimates may indicate some degree of variability between 12,000 and 11,500 cal yr BP. A cooling trend is indicated at the top of the sequence. Although age control here is imprecise, this cooling may have occurred after ca 11,200 cal yr BP.

To evaluate the models, we consider how well they predict modern temperature from the original New Zealand lake data set (Fig. 2) over the ranges shown in the paleotemperature reconstruction. The plot of observed against estimated temperatures and residuals errors for each model suggests that the PLS model may provide the best temperature estimate within the range of 8.5–13 °C. The WA-PLS model performs best in the range of 9.5–12 °C, although it may be more likely to underestimate temperatures in this range than the PLS model (Dieffenbacher-Krall et al., 2007). This may be because the PLS model utilizes a larger suite of lakes than the WAPLS model to develop temperature estimates throughout its range of best prediction (8.5–13 °C). Therefore, we adopt the PLS model for the chironomid-derived summer temperature reconstruction here, in conjunction with the pollen record, to determine the paleoclimate history for the site. However, temperature estimates produced by both models between 10 and 12 °C are derived from a small number of lakes in the modern data set, which may affect the robustness of the temperature predictions in this range. Both models tend to underestimate the warm range (>13 °C) and overestimate the cool range (<8 °C), as is typical for such models (e.g. Birks, 2003).

5.5. Paleotemperature reconstructions 17,500–10,000 cal yr BP

The age of ca 17,500 cal yr BP for the basal sediments in this sequence provides a minimum estimate for the onset of deglaciation from glacial maximum limits in the upper Pukaki Basin. This is consistent with exposure age estimates for major ice recession from terminal moraine limits in the lower Pukaki Basin (Schaefer et al., 2006) and with evidence from throughout New Zealand for a major change in climate around 18,000 cal yr BP that terminated the last major glacial episode (Newnham et al., 2003, 2007a; Alloway et al., 2007). Nevertheless, at Boundary Stream Tarn, absence of tall forest and dominance of subalpine shrubland and alpine grassland between 17,500 and 14,200 cal yr BP indicates that climate regimes were cooler and more extreme than at present. Chironomid inferred mean summer temperatures increased from 11 to 12.7 °C during this time, 0.7–2.4 °C lower than the present-day mean summer temperature of 13.4 °C (Fig. 4).

The minor expansion of tall tussock alpine grassland at ca 14,100 cal yr BP is indicative of a return to conditions similar to (or cooler than) those between 17,500 and 15,500 cal yr BP. An increase in inorganic sediment after 14,100 cal yr BP may result from enhanced weathering and erosion due to frost heave and vegetation changes near the site. Increased aquatic plant pollen and spores (e.g. *Isoetes*) and colonial algae (e.g. *Botryococcus*); indicate greater prevalence of both running and ponded water at the site during this time. Chironomid temperature reconstructions indicate that cooling began at ca 14,200 cal yr BP and by ca 14,000 cal yr BP temperatures had decreased to 11.5–12, or 1.5–2 °C below modern summer air temperature. A short-lived interruption to this cooling trend occurred about 13,900 cal yr BP, when mean summer temperature increased to around 13.5 °C, at or slightly above its present-day level (Fig. 4). Within 100 years or so, pronounced cooling resumed for ca 600–700 years. During this interval, average mean summer temperatures dropped to around 10.5–11.4 °C, with extreme cool temperatures reconstructed from one sample approximating 9.5 °C. These temperatures were generally 2.0–2.9 °C below the present day and in the most extreme case as much as 3.9 °C lower.

A marked abrupt warming from ca 11.4 to ca 12.6 °C at ca 13,200 cal yr BP is evident in the chironomid summer temperature reconstruction and is followed by an interval (13,100–11,500 cal yr BP) when the two climate proxies may have responded differently. The pollen record indicates sequential replacement of grassland/shrubland communities by *Halocarpus*-dominated subalpine conifer forest. This vegetation development is evident in other pollen records from the region (e.g. McGlone et al., 2004) and has been interpreted as indicating increasing minimum winter or spring temperatures. In contrast, chironomid reconstructions show no clear temperature trend between 13,000 and 11,500 cal yr BP, but some variability associated with an

Fig. 4. Summary pollen and chironomid diagram for Boundary Stream Tarn. Dark shading on the Poaceae pollen curve represents grains greater than 40 µm, characteristic of tall *Chionochloa* tussock grassland. Both WA-PLS and PLS model reconstructions are shown. A LOWESS smooth (0.1) has been fitted to highlight the major trends and sample specific errors are shown by light shading on each reconstruction. Vertical line on the temperature reconstructions at 13.4 °C represents present mean summer temperature at the site. The pollen and chironomid diagrams represent only those species that occur in abundances > 5% or provide information directly relevant for Lateglacial environmental and climate reconstruction in the area. Taxa excluded from this figure are shown in supplementary Fig. 1. Gray shading represents zone of Lateglacial change in sediment stratigraphy, and pollen and chironomid assemblages.

increase in inorganic silt between 12,000 and 11,500 cal yr BP. During the interval 13,000–11,500 cal yr BP, temperatures fluctuated between 11.6 and 13.8 °C, averaging around 12.5 °C or about 1.0 °C below present (Fig. 4). These differences between chironomid and pollen-vegetation response imply a minor decoupling of summer and winter temperature trends during this interval. Alternatively the differences may be due to variable response times between the two proxies to the abrupt warming evident in the chironomid temperature reconstructions at 13,200 cal yr BP. In this scenario, the chironomids responded rapidly to the abrupt warming while the pollen record showed a more lagged response, which eventually results in successional replacement of grassland–shrubland communities with sub-alpine conifer forest.

After 11,600 cal yr BP, a general rise of temperature to values near those of today is indicated in the chironomid reconstructions. The rapid transition from *Halocarpus*-dominated to *Phyllocladus*-dominated subalpine conifer forest after 11,500 cal yr BP is a feature evident in other South Island pollen diagrams (e.g. Burrows and Russell, 1990; McGlone et al., 2004). The vegetation change appears to be a precursory development of tall forest and is considered to represent a shift to temperatures similar to the present. Marked changes are

also recorded in the sediment stratigraphy at this time with a transition from inorganic to organic fibrous silt, indicative of sustained catchment stability.

In summary, the multi-proxy record from Boundary Stream Tarn offers new insights into patterns of Lateglacial climate change in the Pukaki Basin. Following deglacial warming from ca 17,000 cal yr BP, the pollen and chironomid evidence collectively indicates a period of markedly fluctuating, but overall cooler, temperatures from 14,200 to 13,200 cal yr BP, broadly coinciding with the ACR. This cooler period includes at least one pronounced fluctuation evident in the chironomid record but not the pollen record, when chironomid-derived temperatures increased by 2–3 °C and then fell by a similar or greater amount during the interval ca 14,000–13,600 cal yr BP. Both paleoecological proxies, together with the sediment stratigraphy, strongly suggest that this was a period of relative climate instability.

This period of cooling and instability was terminated by an abrupt warming, accompanied by increased stability in the catchment at ca 13,200 cal yr BP. This warming step was followed by an interval broadly coinciding with the Younger Dryas chronozone, from 13,000 to 11,500 cal yr BP, when the inferred temperature changes from pollen and chironomids may show minor differences.

Gradual changes from the pollen record during this time versus rapid changes in the chironomid record at the beginning of this interval suggest that summer temperatures, indicated by the chironomid reconstructions, remained comparatively stable with only brief periods of variability while minimum temperatures in winter and spring, indicated by pollen-vegetation developments, increased. Alternatively, the differences may be due to variable response times between the two proxies; the chironomids responded rapidly to the abrupt warming whilst the pollen-vegetation showed a lagged response, eventually resulting in the successional development of subalpine conifer forest.

Finally, at ca 11,500 cal yr BP, a further warming step is evident in both the pollen record and chironomid temperature reconstruction. These changes, together with the shift to fibrous organic sediment indicate further increases in summer temperatures, catchment stability and a major transition in subalpine forest composition at the site.

The next section considers the broader implications of these insights into Lateglacial climates facilitated by the Boundary Stream Tarn records, through comparison with other records from the New Zealand region.

6. Discussion

6.1. Lateglacial cooling in the Southern Alps, New Zealand

A Lateglacial cooling event, broadly comparable to that evident in the Boundary Stream Tarn records is observed in

several other high-resolution records from New Zealand (e.g. Newnham and Lowe, 2000; Pahnke et al., 2003; McGlone et al., 2004; Pahnke and Zahn, 2005; Vandergoes et al., 2005; Williams et al., 2005; Newnham et al., 2007b; Fig. 5).

The cooling indicated by the pollen and chironomid records from Boundary Stream Tarn (ca 14,200–13,200 cal yr BP) broadly coincides with other pollen-based evidence of Lateglacial climate reversal from the eastern margins of the Southern Alps (Moar, 1971; McGlone et al., 2004) dated between 14,600 and 13,000 cal yr BP (McGlone et al., 2004). Cooling centered on 13,600 cal yr BP is also shown by mean annual air temperature estimates from a recently developed pollen-temperature calibration model (Wilmshurst et al., 2007). These records provide strong evidence of a regional climate cooling event. Climate cooling between 14,100 and 13,000 cal yr BP is indicated by the marine oxygen isotope and reconstructed sea surface temperature records from MD97 site 2120 (Fig. 5) off the east coast of the South Island (Pahnke et al., 2003; Pahnke and Zahn, 2005). The dating control for cooling between ca 14,200 and 13,200 cal yr BP provided by the Boundary Stream Tarn chronology strengthens the assertion that cooling occurred about this time, as noted in the marine record. The onset of the Boundary Stream Tarn cooling event predates by ca 600 years the cooling signal indicated by high-resolution pollen evidence from Kaipo Bog north-eastern New Zealand (Newnham and Lowe, 2000; Hajdas et al., 2006), and South Island speleothems (Williams et al., 2005; Fig. 5), although it

Fig. 5. (A) Boundary Stream Tarn chironomid temperature records as compared to (B) marine oxygen isotope and SST reconstruction from MD97-2120 (Pahnke and Zahn, 2005), (C) Kaipo Bog lowland pollen: grassland ratio and percentage carbon record (Newnham and Lowe, 2000), (D) composite South Island speleothem record (Williams et al., 2005), (E) EPICA Dome C Deuterium excess record (EPICA Community Members, 2006), and (F) Greenland GISP 2 oxygen isotope record (Alley, 2000). Black bar (1) represents the Lateglacial advance of the Franz Josef Glacier over Canavans Knob (Denton and Hendy, 1994). The ACR and YD are represented by overlapping shaded bars.

is possible that these events may be more closely aligned when chronological error margins are taken into account. As the onset of cooling in the Kaipo Bog and speleothem records shows strong correspondence with the coldest phase of cooling in the Boundary Stream Tarn record between 13,900 and 13,200 cal yr BP, it seems probable that these two events are linked.

The termination of Boundary Stream Tarn cooling (after 13,200 cal yr BP) antedates the end of cooling observed in Kaipo Bog by ca 600 years (approximately 12,600 cal yr BP; Newnham and Lowe, 2000; Hajdas et al., 2006). Cooling is indicated to have persisted for another 1600 years (until 11,600 cal yr BP) in the South Island speleothem record (Williams et al., 2005). The termination of the Boundary Stream Tarn cooling event coincides with well-dated till deposits at Canavans Knob indicating an advanced position of the Franz Josef Glacier at 13,100 cal yr BP (Denton and Hendy, 1994).

The Lateglacial cooling interval at Boundary Stream Tarn, dated at 14,200 to ca 13,200 cal yr BP, falls within the ACR recorded in Antarctic ice cores, broadly dated as 14,500–12,500 cal yr BP (EPICA Community Members, 2006). Thus evidence from Boundary Stream Tarn indicates that the ACR signal extended into southern New Zealand. There is no clear evidence that this climate reversal persisted into the YD chron. However, the PLS chironomid temperature estimates indicate that mean summer temperatures remained about 1 °C cooler than present for the interval 13,100–11,500 cal yr BP, after which summer temperatures became more similar to those of the present day. In contrast, the pollen assemblages are consistent with others from this region and show vegetation changes interpreted as progressive increase in minimum winter or spring temperatures throughout this interval or, alternatively, a warming step at ca 13,200 cal yr BP sufficient to trigger the slightly lagged successional development of low forest. The former explanation implies strong seasonal differences occurred during this interval, whereas the latter interpretation would indicate similar patterns between seasons but with the vegetation developments lagging the chironomid response to abrupt warming at ca 13,200 cal yr BP.

Regardless of which of these interpretations is favored, a slightly cooler than present, climate over New Zealand between 13,100 and 11,500 cal yr BP may help explain evidence of prolonged cooling from other high-resolution records such as the persistence of the cooling reversal at Kaipo Bog to ca 12,600 cal yr BP (Hajdas et al., 2006), and of the 'New Zealand Lateglacial Reversal' identified in South Island speleothems lasting to ca 11,600 cal yr BP (Williams et al., 2005). In both these studies, cooling appears to have been initiated during the ACR but extended into either part or all of the YD chron. This evidence contrasts with interpretations from earlier New Zealand studies which imply that the YD was a period of gradual warming (e.g. McGlone, 1995; McGlone et al., 2004). By indicating that the transition from ACR to YD

may have been characterised by an abrupt warming step, followed in some regions at least by slightly cooler-than-present temperatures during the YD, the Boundary Stream Tarn record provides a potential explanation for this conflict. Clearly the presence at times of variable and slightly cooler climate between 13,100 and 11,500 cal yr BP needs to be corroborated by other multi-proxy investigations from climatically sensitive sites.

The magnitude of the Lateglacial cooling trend in summer temperatures at Boundary Stream Tarn (2.9–2 °C below present, PLS model) is less than the range of likely temperature decreases (3–4.7 °C) and combined precipitation increases indicated in recent modelling of Lateglacial ice advance of the Franz Josef Glacier (Anderson and Mackintosh, 2006). The ice modelled temperature reconstructions are more consistent with the possible extreme cooling value (–3.9 °C) indicated by the chironomid PLS model and the average trend range of the WA-PLS model (–2.7–4.2 °C) reconstructions for this site. The cooling trend at Boundary Stream Tarn is more comparable with an approximate 2.5 °C cooling noted in a marine record off the east coast (Pahnke et al., 2003; Pahnke and Zahn 2005; Fig. 5). Regional temperature depressions below present-day values in southern New Zealand may not have been more extreme than the 2.5 °C derived from marine core proxies. However, if the modern climate-temperature gradients from immediately east of the Southern Alps to the east coast of the South Island persisted in the past, it is likely that inland regions of southern New Zealand experienced more extreme climates and temperature depressions than the lowland and coastal maritime areas. A scenario of more extreme inland climates in the past resulting from a depressed or intensified east–west climate gradient would accommodate the prediction errors associated with the chironomid reconstructions of both PLS and WAPLS models. Consistency between the chironomid PLS model results and other quantitative temperature estimates from New Zealand, such as sea surface temperature reconstructions, supports the current use of this model for Lateglacial temperature reconstructions.

6.2. Lateglacial shifts in seasonality

A notable difference between the Boundary Stream Tarn pollen record and chironomid-based temperature estimates occurs between 14,000 and 13,800 cal yr BP. An abrupt warming event represented in the chironomid temperatures and driven by minor increases in *Corynocera*, *Naonella kimihia* and *Paucispinigera* sp. is not shown in the pollen record. It is possible that this warming event was too short-lived to have impacted on local vegetation, but alternatively it may indicate differences in biotic response to climate change. Chironomids show the strongest relationship to mean summer temperature (Dieffenbacher-Krall et al., 2007). In contrast, New Zealand vegetation, with its low freezing resistance (Wardle, 1991), is likely to be strongly controlled by the duration of winter or

annual minimum temperatures. The difference between the pollen and chironomid results may indicate seasonal climate variability, in which summers were short and warm and winters cold or prolonged, resulting in short vegetation growing seasons. This interpretation implies more extreme seasonality between 14,200 and 13,100 cal yr BP than prior to 14,200 cal yr BP, when both summer and winter temperatures seem to have been generally cool. After 13,100 cal yr BP, seasonal variability or a more parallel relationship between summer and winter temperatures may equally explain the differences between the pollen and chironomid signals.

The warming at Boundary Stream Tarn between 14,000 and 13,800 cal yr BP may be similar to other events recorded in other sites around New Zealand. For example, an increase in lowland pollen and organic carbon in a North Island pollen sequence (Newnham and Lowe, 2000), and a positive excursion in the $\delta^{18}\text{O}$ from South Island speleothems (Williams et al., 2005) and the offshore east coast marine record (Pahnke et al., 2003; Pahnke and Zahn, 2005; Fig. 5) may represent warming in these records. Minor differences in timing of this event among these records are likely to be within the error margins of the dating control. In each case, the warming event appears to be a precursor to the subsequent ACR-like cooling and therefore the two phenomena may be linked. Further investigations are necessary to confirm the timing and distribution of this abrupt warming event.

7. Conclusions

The Boundary Stream Tarn sedimentary sequence shows that deglaciation from glacial maximum limits indicated by the Pukaki lateral moraines commenced prior to ca 17,500 cal yr BP. This is consistent with the timing of major recession from terminal moraines down valley and with comparable evidence from elsewhere in New Zealand. The chironomid temperature reconstruction at this site represents the first of its kind for the Lateglacial in New Zealand, and provides a quantitative assessment of climate change for this time period. While the chironomid-derived reconstruction generally confirms the pattern of temperature change inferred from pollen, differences in the structure and possible magnitude of cooling are indicated. These minor differences between the two proxies are likely to reflect seasonal climate variations although a lagged vegetation response to abrupt climate change may also be a factor. Both seasonal differences in climate and a lagged vegetation-climate response may be contributing factors to the variability thus far observed between other proxy records for Lateglacial climate reversal in New Zealand. The ^{14}C AMS-derived chronology constrains the timing of Lateglacial cooling in the Southern Alps of New Zealand to the ACR. The Boundary Stream Tarn record thus supports an ACR-like climate reversal in southern New Zealand and an antiphased relationship between Antarctic-New Zealand and Northern Hemisphere climate at this

time. A possible abrupt warming event, occurring around the onset of the ACR in the Boundary Stream Tarn chironomid reconstructions, may have correlatives in other New Zealand records and should be a focus of further investigation.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version at [doi:10.1016/j.quascirev.2007.11.015](https://doi.org/10.1016/j.quascirev.2007.11.015).

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