Kauri tree-ring stable isotopes reveal a centennial climate downturn following the Antarctic Cold Reversal in New Zealand

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Key Points:

- Dual isotope ($\delta^{18}O_{cel}$, $\delta^{13}C_{cel}$) tree-ring stable isotope chronologies (KAU-ISO) were developed from New Zealand kauri trees, covering 13 020 – 11 850 cal BP
- A simultaneous downturn in all tree-ring proxies was established covering 250 years (Kauri Downturn = KD)
- The records presented provide new high-resolution evidence of hydroclimate conditions in NZ at the end of the Late Glacial

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Abstract

The dynamics of the Late Glacial (LG) have been demonstrated by numerous records from the Northern Hemisphere (NH) and far fewer from the Southern Hemisphere (SH). SH paleoclimate records reveal a general warming trend, interrupted by a deglaciation pause (ACR: Antarctic Cold Reversal, ~14,700 – 13,000 cal BP). Here we present decadal tree-ring stable isotope chronologies ($\delta^{18}O$, $\delta^{13}C$) from New Zealand (NZ) subfossil kauri trees (n=6) covering the post-ACR millennium from 13 020 – 11 850 cal BP. We find a distinct, simultaneous downturn (~12 625 – 12 375 cal BP) in all tree-ring proxies paralleling regional tree growth declines, suggesting a widespread climate deterioration. This downturn was characterised by sustained high precipitation, low temperatures and high relative humidity in NZ with incoming weather fronts from the South Ocean. Despite these promising results, questions remain about what drove the Kauri Downturn and how the hydroclimatic conditions were altered during this time period.

1 Introduction

1.1 The Late Glacial

Significant attention in paleoclimate research has focussed on the transitional period between the last glaciation and the current Holocene interglacial, known at the Late Glacial (~14,700 – 11,600 cal BP). In the Northern Hemisphere (NH), the gradual warming was interrupted by numerous fast transitions to cold episodes of various lengths. The most prominent millennium-long cold reversal to near glacial conditions is evidenced in Greenland ice core $\delta^{18}O$ and deuterium excess (GS-1; Figure 1h). This led to significant palynological changes in continental Europe characterizing the so-called Younger Dryas period (YD) (Brauer et al. 2000; Steffensen et al. 2008; Rach et al. 2014). The Southern Hemisphere (SH) however tells a very different story, with the gradual warming continuing until the warm Holocene, with the exception of a brief warming pause during the Antarctic Cold Reversal (ACR, ~14,700 to 13,000 cal BP; Blunier et al. 1997; Pedro et al. 2011; WAIS 2015) (Figure 1g) and various decadal oscillations demonstrated in New Zealand (NZ) records (Barrell et al. 2013). Similar hemispheric divergences have been discovered over long-term Glacial-Interglacial cycles across the Quaternary period (Crowley 1992), evidenced as SH warming leading the NH during glacial terminations; due in part to slower melting of NH ice sheets, greatly impacting hemispheric temperatures (Kawamura et al. 2007).

The abrupt NH cold reversal (Younger Dryas onset / beginning of GS-1) at the end of the Late Glacial has been hypothesised to be a result of weakening of the Atlantic meridional overturning circulation (AMOC)(Carlson and Clark 2012; Renssen et al., 2015); an Atlantic ocean current system characterised by a northward flow of tropical surface warm waters, which cool and sink in the North Atlantic and eventually reroute back south along the coast of North America. AMOC weakening can occur as a result of freshening of the North Atlantic (due to ice melt), decelerating the thermosaline-modulated oceanic circulation system. Such weakening would theoretically block warm tropical waters from entering the North Atlantic allowing for reestablishment of sea ice (and GS-1/YD glacial conditions), while concurrently impeding cool deep-water upwelling in the south Atlantic causing SH warming and thus a bipolar divergence in climate. Changes in AMOC activity are hypothesised to be further compounded in the SH by shifts in the southern westerly wind belt and associated ocean currents as the Intertropical Convergence Zone (ITCZ) oscillated across latitudes (Denton et al., 2010). Thus, climate change induced by variations in AMOC and related ocean circulation can lead to a complex and divergent set of conditions between regions and hemispheres (Figure 1), also demonstrated in the coupled atmosphere-ocean general circulation model, TraCE-21ka (Supplementary Figure 1); referred to as the the “bipolar seesaw” (Stocker et al. 1998)

1.2 LG climate downturns. Southern vs. Northern hemisphere

While the triggers of LG cold episodes are still under debate, numerous climate records in the NH have been established to understand the impact and regionality of cold episodes across the Late Glacial. These records (Lauterbach et al. 2011; von Grafenstein et al. 1999; Brauer et al. 2000; Merkt & Muller 1999) demonstrate long-term and low frequency temperature variability similar to Greenland ice cores (Steffensen et al. 2008; Rasmussen et al. 2014), with some regional differences in the rate, magnitude and timing of climate events (e.g. Pauly et al., 2018; Rach et al. 2014). Conversely, the southern hemisphere has far fewer paleoclimate records available covering this interesting climate period, with only a handful of relevantly high (decadal) temporal resolution (e.g. Hajdas et al., 2006; De Deckker et al., 2012, etc; Figure 1).

During the YD phase, when the NH plunged into cold conditions, Antarctica experienced a warming trend. Antarctic ice core $\delta^{18}O$ (e.g. Pedro et al. 2011), reconstructed Southern Ocean sea surface temperature (Correge et al., 2004) and New Zealand glacier retreat (Kaplan et al., 2010) argue for steadily increasing temperatures between ~12 900 – 11 800 cal BP in the SH. Southern Australia ocean core data suggests this warm phase was regionally modulated by an oscillating Subtropical Front (STF), permitting a millennial flickering of the Leeuwin Current (LC) strength across southern Australia past Tasmania (De Deckker et al. 2012). This STF activity ultimately impacted regional temperature and precipitation patterns during distinct phases. During one such interval (~12 500 – 12 380 cal BP), Tasmanian huon pine (*Lagarostrobus franklinii*) and New Zealand kauri (*Agathis australis)* trees show significant centennial-long growth depressions (Hua et al., 2009; Palmer et al., 2016; Figure 1c) which may be related to a changed state of the mid-latitude Southern Ocean and related El Niño-Southern Oscillation (ENSO) activity (Palmer et al. 2016). Locally, this interval parallels a conversion from a dry ACR to relatively wet conditions at Lake Hayes in South Island NZ (Hinojosa et al. 2019; Figure 1e). In North Island, this short period represents a transition from the late-glacial cool episode (NZce-3) to the pre-Holocene amelioration (NZce-2) according to Kaipo bog stratigraphy (Figure 1i; Newnham & Lowe 2000; Hajdas 2006; Barrell et al. 2013) as well as wet, cool conditions recorded in Ruakuri Cave speleothems during Heinrich Event 0 (Whittaker et al. 2008; Figure 1f). Regionally, a short-term depletion in Talos ice core $\delta^{18}O$ (Figure 1g; ~12 580 – 12 380 cal BP), and Law ice core $\delta^{18}O$ (Pedro et al. 2011; ~12 540 – 12 320 cal BP) may provide hints of the climate deterioration in Eastern Antarctica; although it generally appears to be independent of the Atlantic and West Antarctic regions, with no significant signal visible in Western Antarctica (EDML, Siple and Bryd ice cores; Pedro et al. 2011). Palmer et al. (2016) hypothesised this tree growth downturn was followed by a strengthening of ENSO expression in the region and/or solar variability. However, due to the limited climate records existing over this period and the limited age control of New Zealand records, the climatological characteristics of this centennial-duration event remain under debate.

1.3 Kauri trees as a climate archive

New Zealand kauri (Figure 2ab) trees are southern hemisphere conifers which can grow for multiple millennia (Palmer et al., 2016) and thrive in warm-temperate lowlands regions. Due to their long lifespan, annual rings and sensitivity to summer climate indices, Kauri trees are an important bioarchive. Studies include living kauri trees (Fowler et al., 2004; Ogden and Ahmed 1989), archaeological timber (Boswijk et al., 2006), as well as ancient material (e.g. Lorrey and

Ogden, 2005; Turney et al., 2010; Palmer et al., 2016) dating as far as 60,000 years before present. Tree-ring width chronologies of kauri have demonstrated a connection with ENSO activity in modern times (Fowler et al., 2007, 2000) as well as during the Holocene (Fowler et al., 2012) and the Late Glacial (Palmer et al., 2016).

The sensitivity of kauri to summer season (November – February; NDJF) hydroclimate indices have been further explored through stable isotope analysis of modern kauri cellulose (e.g. Lorrey et al., 2016; Brookman 2014; Text S1). Brookman (2014) demonstrated the concurrent sensitivity of $\delta^{18}O$ and $\delta^{13}C$ from kauri tree-ring cellulose ($\delta^{13}C_{cel}$, $\delta^{18}O_{cel}$) to various climate parameters in the summer, including positive correlations to mean temperature (r_o = 0.70, r_c = 0.89), solar radiation (r_o = 0.64, r_c = 0.89) and soil moisture (r_o = 0.71, r_c = 0.84), as well as negative correlations to rainfall (r_o = -0.33, r_c = -0.22) and relative humidity (r_o = -0.41, r_c = -0.65). Lorrey et al. (2016; 2018) also found a correlation between $\delta^{18}O_{cel}$ and the SOI indicating that kauri $\delta^{18}O_{cel}$ may be a useful proxy of past ENSO conditions. While both $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ in kauri tend to be similarly sensitive to the abovementioned climate variables, age-related offsets as well as divergences in microclimate conditions can introduce complexities into the $\delta^{13}C_{cel}$ record (Brookman 2014) as described in other Late Glacial records (Pauly et al. 2020).

By using kauri tree-ring stable isotopes, we aim to investigate the hydroclimate state of North Island during the post-ACR millennium (13 020 – 11 850 cal BP) and identify the radiation of ocean-atmospheric dynamics and expression of ENSO in the SH by exploring connections amidst other regional records. Using a dual stable isotope modelling approach, we analyse $\delta^{18}O$ and $\delta^{13}C$ to interpret the growing season climate conditions present before, during and after a distinct growth depression (~12 625 – 12 375 cal BP, Figure 1c).

2 Materials and Methods

2.1 Subfossil kauri chronology

Here we present a dual stable isotope chronology (carbon and oxygen) developed from tree-ring cellulose extracted from subfossil trees discovered near Towai in Northland, New Zealand (35°30.3930S, 174°10.3760E; Figure 2c; Text S2). A subset of trees (n_{trees} = 6) has been chosen (based on a minimum required to develop a strong climate signal from $\delta^{18}O_{cel}$; Lorrey et al. 2016) from a floating tree-ring chronology (Towai), which covers 13 134 – 11 694 cal BP (n_{trees} = 37, 1σ error of ±7 years; Hogg et al., 2016).

This period spans the Greenland Isotope defined GS-1 (Rasmussen et al., 2014) and pollen defined NH Younger Dryas (e.g. Björck et al. 1998), immediately following the SH Antarctic Cold Reversal (ACR) (Figure 1) and spanning the SH 'Late-glacial cool episode' into the 'pre-Holocene amelioration (Barrell et al. 2013) defined by NZ bog pollen (Newnham and Lowe, 2000; Hajdas et al. 2006; Lowe et al. 2008, 2013).

The stable isotope chronology (KAU-ISO) is of decadal resolution with a sample replication of 4-6 trees for the majority of the chronology (56%; 670 years), with some periods having a replication of 3 trees (28%; 280 years) and even less with a replication of 1-2 (15%; 160 years) (Figure 1d). The subset of trees were chosen for this work as an initial investigation into the potential to reconstruct climate during the Late Glacial with kauri trees.

The tree-ring width chronology (Palmer et al., 2016) with which this dendroisotope data was developed, hypothesised a connection between a distinct growth depression (~12 625 – 12 375 cal BP) and subsequent increased intensity in ENSO activity. Trees used here were growing before, during and after this peculiar 250 year-long growth depression (Palmer et al., 2016), which we aim to further explore in this study.

2.2 Cellulose extraction and stable isotope analysis

Cellulose was extracted from wholewood material of 10-year blocks of tree rings (Wieloch et al. 2011; Schollaen et al. 2017) from individual trees covering 13 020 – 11 850 cal BP, which is a subset of the Towai tree-ring chronology. Annual stable isotopes were also measured from an individual tree, covering mid-point of the growth depression, between 12 520 – 12 400 cal BP (Figure 2d). The samples were homogenized and freeze-dried prior to being weighed and packed (silver capsules (ø3.3x4mm) for stable isotope measurement (Delta V, Thermofisher Scientific Bremen; coupled with TC/EA HT at 1400°C). The measurements were not pooled between trees, only within the 10-year blocks. Results were compared against international and lab-internal reference material (IAEA-CH3, IAEA-CH6 and Sigma-Aldrich Alpha-Cellulose) using two reference standards with widespread isotopic compositions for a single-point normalisation (Paul, Skrzypek, and Forizs 2007). Final isotope ratios are given in δ value, relative to VSMOW ($\delta^{18}O$) and VPDB ($\delta^{13}C$), with replication reproducibility of ±0.3‰ ($\delta^{18}O$) and ±0.15‰ ($\delta^{13}C$).

2.3 TraCE-21ka climate model outputs

The time interval covered by KAU-ISO was investigated using TraCE-21ka, a coupled atmosphere-ocean general circulation model - which simulates global climate evolution between the Late Glacial Maximum (LGM: 21,000 years before present) to modern times - using the Community Climate System Model version 3 (CCSM3). Three periods of time were examined: (1) Pre-downturn (12 800 cal BP), (2) Kauri Downturn (12 500 cal BP), and (3) Post-downturn (12 200 cal BP), and four climate variables: annual (1) relative humidity, (2) precipitation, (3) temperature, and (4) sea level pressure. The modelling was completed over the SH summer season (November - February), when kauri trees exhibit growth (Ecroyd 1982). Climate model outputs were visualised using PaleoView v1.5.1 with a user defined geographic region (covering Australia, New Zealand and a portion of the Southern Ocean) and changes relative to 10,000 cal BP.

3 Results

3.1 Stable isotope & tree-ring chronologies

3.1.1 Decadal stable isotope record (KAU-ISO)

Stable oxygen ($\delta^{18}O_{cel}$) and carbon ($\delta^{13}C_{cel}$) records of tree-ring cellulose average ~30.6‰ and ~ -22.0‰ with ranges of 4.3‰ (28.3 to 32.6‰) and 3.8‰ (-20.3 to -24.1‰) and inter-decadal variability of 0.66‰ and 0.37‰, respectively. Inter-tree correlation for $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ is not consistent; ranging between -0.76 to +0.88 for $\delta^{18}O_{cel}$ and -0.44 to +0.80 for $\delta^{13}C_{cel}$ (Table S1, Figure S2).

Decadal mean $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ records were developed from the individual tree-ring records with an average standard deviation of 0.89 for $\delta^{18}O_{cel}$ and 0.70 for $\delta^{13}C_{cel}$. Correlations between mean $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ are strongly positive during much of the dataset, with 38% of the dataset demonstrating correlation coefficients above +0.4, a high of +0.82 and absolute average of 0.36 (Figure S2).

During the first 350 years of the chronology (~13 000 - 12 650 cal BP), mean $\delta^{18}O_{cel}$ exhibits an increasing trend with a peak at ~12 650 - 12 640 cal BP (Figure 1b), while mean $\delta^{13}C_{cel}$ demonstrates a plateau within a relatively negative anomaly with a similar (albeit less extreme) peak at 12 640 cal BP (Figure 1a). Similar to the previously reported tree-ring width downturn (Palmer et al. 2016; Figure 1c), $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ from KAU-ISO show concurrent and significant depletions between approximately 12 630 – 12 380 cal BP. This set of depletions begin with a peak (maximum of the entire sequence for $\delta^{18}O_{cel}$) around 12 630 cal BP, followed by a steady decline, with deepest point centred ~12 450 cal BP. This event sequence will hereafter be referred to as the "Kauri Depression" (KD). All tree-ring parameters (width, $\delta^{18}O_{cel}$, $\delta^{13}C_{cel}$) recover, peaking at ~12 380 cal BP and then continue slow decline until the end of the chronology at 11 850 cal BP.

Mean inter-tree correlations for $\delta^{18}O_{cel}$ are higher during the downturn (average = 0.22, range = -0.07 to +0.55), compared with pre- and post- downturn; $average_{pre}$ = -0.23, $range_{pre}$ = -0.76 to +0.07, $average_{post}$ = +0.07, $range_{post}$ = -0.22 to +0.28. Conversely, inter-tree correlations for $\delta^{13}C_{cel}$ are equivalent pre- and during downturn ($average_{pre}$ = 0.13, $range_{pre}$ = -0.44 to +0.80, $average_{downturn}$ = 0.13, $range_{downturn}$ =-0.26 to +0.48, respectively) and relatively higher post-downturn (average = +0.28, $range_{post}$ = -0.27 to +0.64). Slightly higher inter-tree correlations have been found modern kauri stable isotope chronologies of $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$, with a correlation of 0.21-0.64 for $\delta^{18}O_{cel}$ (Brookman 2014, Lorrey et al. 2016) and 0.35 for $\delta^{13}C_{cel}$ (Brookman 2014).

3.1.2 Annual stable isotope record (KAU-ISO)

Dual isotopes of a single tree were analysed covering the latter half of the KD (12 520 – 12 400 cal BP) to provide information on the inter-annual variability of $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ during this climate "event" (Table S2). This tree displayed a range of $\delta^{18}O_{cel}$ values of only 2.9‰ (28.9 to 31.8‰), which is much more limited than modern kauri $\delta^{18}O_{cel}$ values which demonstrate an 8‰ range (28.8 to 36.8‰) according to Brookman (2014). Based on similarities between the absolute values of the isotopic records, tree species and location, we assume similar seasonal trends identified in Brookman (2014) would also impact the annual dataset presented in this study.

3.2 Climate Model

Whilst precipitation and relative humidity can be estimated from the kauri records in this study, information on temperature and sea level pressure, as well as spatial trends of all parameters, were further investigated using Paleoview software to visualise TRaCE21ka data. The TRaCE21ka model outputs confirm that a phase of high relative humidity and high precipitation in SH summer (NDJF) occurred in NZ during the downturn (~12 500 cal BP) compared to pre- and post-downturn (12 800 cal BP and 12 200 cal BP, respectively), matching the results of KAU-ISO.

4 Discussion

4.1 Stable Isotope Chronologies

The decadal stable isotope chronologies (KAU-ISO, 13 020 – 11 850 cal BP) from kauri tree-rings reflected similar means and ranges compared to modern studies (Brookman 2014). Inter-tree correlations varied throughout the record (Figure S1), potentially due to the pooling of decadal tree-rings within individual trees; pooling has been shown to hide bias of individual trees which deviate from the population signal (Liñán et al. 2011). Trees showed more significant correlations between individual trees for $\delta^{13}C_{cel}$ measurements over $\delta^{18}O_{cel,}$ in contrast to other subfossil (Pauly et al. 2018, 2020) stable isotope studies, which generally show stronger population signals in $\delta^{18}O_{cel}$ data. Despite individual tree differences, mean $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ demonstrate strong correlations (38% of chronology >0.4; Figure S2), reflecting a set of interdependent variables within the climate system and similar long-term trends.

Annual values of $\delta^{18}O_{cel}$ tend to be more enriched (mean = 30.5‰) than concurrent decadal $\delta^{18}O_{cel}$ values (mean = 29.7‰). This is likely due to the temporal bundling of depleted rainfall (low $\delta^{18}O$) conditions as a result of persistent, widespread and low frequency (decadal) atmospheric oscillations occurring over the climate downturn. Similar statistics have been reported in relation to modern rainfall extremes and related long-term climate modes in the region (e.g. Aryal et al., 2009; Grimm and Tedeschi 2009; Willems 2013).

4.2 Climate downturn conditions in New Zealand

The simultaneous downturn (KD) in multiple tree-ring proxies (tree-ring width, $\delta^{18}O_{sw,}$ $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$) from our kauri chronology suggests that they are all sensitive to the variability of a single (or set of interacting) climate parameters. Based on modern intra-annual calibrations of the kauri trees from NZ (Brookman 2014), we suspect a decline in growing season temperature, concurrent with an increase in relative humidity and change in precipitation source are responsible for this downturn within the tree-ring and stable isotope signals.

Inter-tree correlations for $\delta^{18}O_{cel}$ are highest during KD (Figure S2) compared to drier and warmer pre- and post- KD periods. Previous studies of Late Glacial subfossil trees (Pauly et al. 2018, 2020) have demonstrated stronger inter-tree $\delta^{18}O_{cel}$ correlations during phases of anomalously low $\delta^{18}O_{cel}$ (assumed to be higher precipitation and/or precipitation from a depleted source), likely due to the reduced influence of stomata-driven fractionation on $\delta^{18}O_{cel}$ (Pauly et al. 2020). Indeed, the $\delta^{13}C_{cel}$ demonstrates the lowest inter-tree correlation during the humid KD, suggesting the trees are less sensitive to stomata dynamics controlling $\delta^{13}C_{cel}$ over this interval. Whilst we assume atmospheric conditions are the main driver in kauri stable isotope variability, we cannot rule out groundwater as being another factor modulating sourcewater uptake.

At annual resolution, during the deepest depression in the tree-ring data, $\delta^{18}O_{cel}$ shows low inter-annual variability (12 520 – 12 450 cal BP, Figure 2d) and low tree-ring growth. Trees then show an abrupt increase in growth and the inter-annual variability of $\delta^{18}O_{cel}$ increases considerably (12

450 – 12 400 cal BP), with instances of inter-annual $\delta^{18}O$ changes >1‰ increasing by more than 3-fold. Modern NZ precipitation exhibits extremes every ~2.9 years (Ummenhofer and England, 2007), whereas data in this study show KD extremes at a rate 1 every 12 years and recovery extremes of 1 in every 3-4 years (Figure 2d). Within the Late Glacial context, dendroisotope records from southern France have revealed similar increases in inter-annual tree-ring $\delta^{18}O$ (+0.2‰ absolute) as a result of the oscillating movement of the polar front at the onset of the Younger Dryas (Pauly et al. 2018).

The similar variability exhibited in modern and ancient kauri tree $\delta^{18}O_{cel}$ and lack of annual or decadal $\delta^{18}O_{cel}$ extremes during the downturn argue against intermittent, flooding as a driver of the depressed tree growth in the region. Furthermore, the tree proxy records do not contain any evidence of flood conditions, which would result in a cessation of growth (rather than a growth depression) as trees would be unable to take up water in anoxic, flooded conditions (Schöngart et al., 2002). Such circumstances have resulted in the destruction of other kauri forests and shortened tree lifespans, leading to swamp preservation of kauri stumps throughout Northland and the Waikato lowlands following extreme storm events (e.g. Green et al. 1985; Ogden et al. 1992; Boswijk et al. 2005). In the case of the kauri in this study, they continued to grow following the depression.

4.3 Combining climate model outputs with dual-isotope theory

Kauri trees thrive in cool and dry summer conditions; yet the trees in this study reveal a prolonged period of unfavourable conditions (moist, cool summers) for kauri growth. Such basic hydroclimate conditions from a modern perspective would be equivalent to a phase of increased La Niña frequency (cold phase of ENSO). However, the climate model (TrACE21ka, Figure 3) output during the downturn offers additional climate variables, providing strong evidence against a prolonged phase of La Niña activity (Table S3).

Kauri tree-rings show reduced growth (low tree-ring width) with high humidity (depleted $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$) and increased precipitation (depleted $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$) during the downturn based on dual-isotope theory (Scheidegger et al. 2000). The TRaCE21ka model provides additional evidence of low temperatures and low sea level pressure conditions being prevalent during the downturn. Together, the model outputs and KAU-ISO reconstructions support the theory of a regional-scale climate deterioration (Palmer et al. 2016).

The model-predicted low sea level pressure, increased southerly airflow (via incoming air masses) and cooler temperatures spanning the downturn are uncharacteristic of La Niña, leaving the period without a clear modern analogue. Rather, it is more likely that sustained low-pressure conditions (Figure 3k) occurred in New Zealand over the downturn. Areas of low pressure (depressions) in the region would lead to high precipitation, low temperatures and high relative humidity compared to pre- and post- downturn periods, as evidenced in Figure 3. In particular, weather fronts from the South Ocean (Antarctica) would bring about isotopically depleted rain, creating the continuous depletion in $\delta^{18}O_{cel}$, reflecting sourcewater.

Mean $\delta^{18}O_{cel}$ and $\delta^{13}C_{cel}$ show higher correlations (Figure S2) during and after the downturn compared to before, signifying that the interacting climate variables (temperature, relative humidity and precipitation) are picked up more strongly in the kauri trees during the extreme climate interval compared to the preceding average conditions. Pauly et al. (2020) showed stronger dual isotope correlations in subfossil trees during wetter conditions (higher precipitation, higher humidity) in the NH due to reduced leaf level fractionation (^{18}O enrichment) resulting from partial stomata closure. We expect similar results here as the pre-downturn period represents average dry conditions in NZ compared to cooler, wetter conditions during and following KD (as recorded in KAU-ISO and TRaCE21ka).

Modern hydroclimate models suggest that anomalous wet years across New Zealand correspond to below average sea surface temperatures and low sea level pressure across and south of New Zealand (latitude 30-60º band; Ummenhofer and England 2007), similar to the downturn model output. These conditions have been shown to occur as a result of an alteration in atmospheric circulation, with equatorial shifts in and weakening of the westerlies as well an increase in incoming (south eastern) polar winds (Ummenhofer and England 2007). Given the similarity between these modern model results and those from this study (in terms of temperature, sea level pressure and precipitation), we hypothesise the downturn was a result of a similar change in atmospheric circulation. As downturn subsided, westerlies would have theoretically strengthened and moved poleward, warming the 30-60º latitude band region.

4.4 Proxy landscape during the Late Glacial

4.4.1 Proxy records in New Zealand

During KD, coral records in the South Pacific reveal a period of relatively low SST (Corrège et al., 2004), concurrent with a flickering of the Leeuwin Current (De Deckker et al. 2008) impacting flow of westerlies across NZ. Furthermore, a sediment record Lake Hayes in southisland NZ (Hinojosa et al. 2019) recorded a short-term drop in Ca/Ti around 12 500 cal BP (Figure 1e), representing an increase in detrital input and relatively humid/wet conditions (decreased evapouration); compared to the generally dry conditions during the post ACR interval locally (12 900 – 11 600 cal BP).

This period of cool, wet conditions occurs at the transition between the local "Late-glacial cool episode" (NZce-3; 13 740 – 12 550 cal BP) and the "pre-Holocene amelioration" (NZce-2; 12 550 – 11 880 cal BP), according to NZ INTIMATE climate event stratigraphy (Barrell et al. 2013; Figure 1i), as described across North Island (Lowe et al., 2008, 2013). Synchronous downward trends in pollen and SST at the onset of NZce-2 - reconstructed from an east Tasmanian Sea core (MD06-2991) and Okarito Bog, respectively - suggest an ocean-atmosphere coupling of conditions occurred in the region (Ryan 2017).

A short-lived, mild depletion in Antarctic ice core $\delta^{18}O$ (Figure 1g) is evident at the time of the KD – particularly in east Antarctic (e.g. Law and Talos; Pedro et al. 2011) – but is difficult to tie to NZ proxy records due to potential dating uncertainties, coarse (multi-decadal) temporal resolution as well as the fact that the proxies record different seasons. Given the more complex climate parameters influencing NZ (e.g. polar and subtropical air masses, zonal westerlies) compared to Antarctica, one would expect a higher quantity of climate oscillations occurring in this region compared to those recorded in Antarctic ice cores (Barrell et al. 2013). While a general climate transition between the cool Late-glacial and warm Holocene is clear from the available NZ records, the KAU-ISO record suggests this climate conversion involved a significant and prolonged (multi-centennial) hydroclimate shift in NZ. However, the climate drivers and high-resolution dynamics of this interval are difficult to interpret.

4.4.2 Ocean-atmosphere teleconnections

An atmospheric $\Delta^{14}C$ rise in the SH has been associated with the onset of the Younger Dryas (increased ^{14}C at ~12 740 cal BP; Hua et al. 2009). This is followed by a few centuries of high variability, including a short-lived peak at ~12 600, concurrent with the onset of KD. Hua et al.

(2009) hypothesised that the lack of uniformity between SH terrestrial (Hua et al. 2009), Pacific (Edwards et al. 1993; Bard et al. 1998, 2004; Burr et al. 1998, 2004) and Atlantic (Hughen et al. 2004; Fairbanks et al. 2005) $\Delta^{14}C$ datasets during the early YD peak imply this period initiated as a result of ocean circulation changes rather than ^{14}C (solar) production rate. This delayed peak in tree-ring $\Delta^{14}C$ (compared to marine $\Delta^{14}C$; Hua et al. 2009) is concurrent with the KD onset (this study and Palmer et al. 2016), suggesting the ocean circulation changes may have driven this climate deterioration recorded in kauri trees on land. This theory is corroborated by the modelled incoming polar winds and related drops sea surface temperature (Figure 3).

Whilst ^{10}Be-modelled $\Delta^{14}C$ (representing atmospheric $\Delta^{14}C$ unaltered by ocean reservoirs; Hua et al. 2009) underestimates tree-ring $\Delta^{14}C$ at the onset of the YD, it closely follows tree-ring $\Delta^{14}C$ along the KD timeline a couple centuries later (~12 600 – 12 400 cal BP). This hints that either (1) solar variability and/or (2) the release of ^{14}C-depleted oceanic CO_2 from the Southern Ocean (Marchitto et al. 2007), likely acted to sustain the climate downturn in New Zealand after the initial (global) ocean circulation trigger, similar to solar variability modulated climate during the Little Ice Age (Grey et al. 2010 and references therein). While the available SH atmospheric CO_2 records from ice cores (e.g. Marcott et al. 2014) are of low temporal resolution (multi-decadal) during this interval, they do indicate an increase following the Antarctic Cold Reversal plateau, substantiating the CO_2 release theory. Furthermore, other studies have suggested that climate deteriorations may be amplified through anomalous sea surface temperatures, thereby driving persistent atmospheric circulation states lasting for decades to centuries (van Geel et al. 2003).

5 Conclusions

The subfossil kauri tree-ring width and stable isotopes, in addition to climate model outputs from this study provide further evidence of spring/summer hydroclimate conditions during the YD/GS-

1 in the SH, complementing the previously constructed kauri tree-ring width record (Palmer et al. 2016). While the millennial-length Younger Dryas cold reversal is not strongly demonstrated in SH paleoclimate records, variability in tree-ring ^{14}C (Hua et al. 2009) and the Kauri Downturn (this study and Palmer et al. 2016) imply that ocean circulation changes triggered a shorter climate deterioration lasting for at least two and a half centuries (~12 625 – 12 375 cal BP) over this time interval. Such conditions are generally reflected in regional lake and ocean sediment records, albeit at lower temporal resolution. Despite these promising results, questions remain about what factors drove and modulated the Kauri Downturn and how the NZ hydroclimate regime progressed into the early Holocene climate amelioration.

Acknowledgments, Samples, and Data

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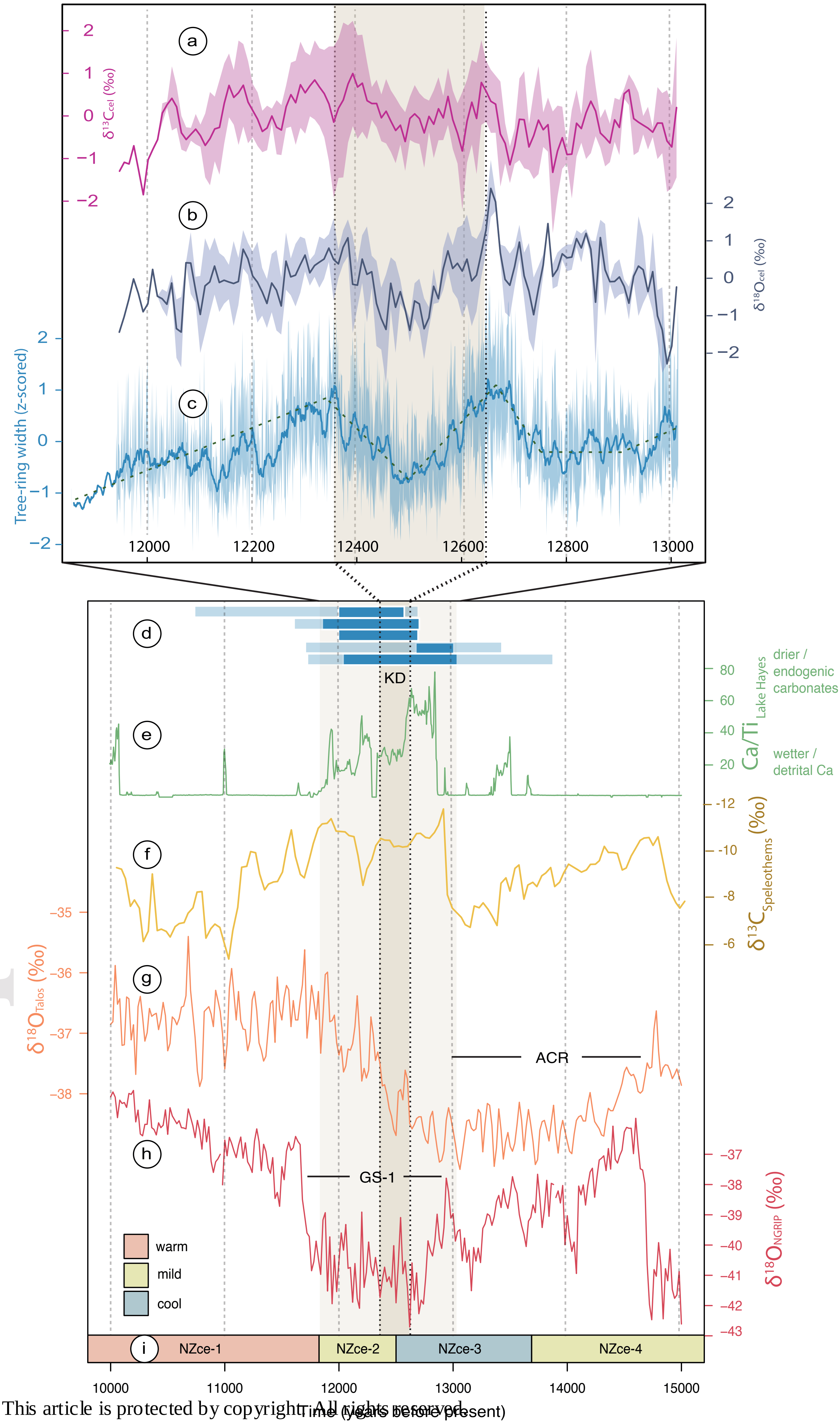
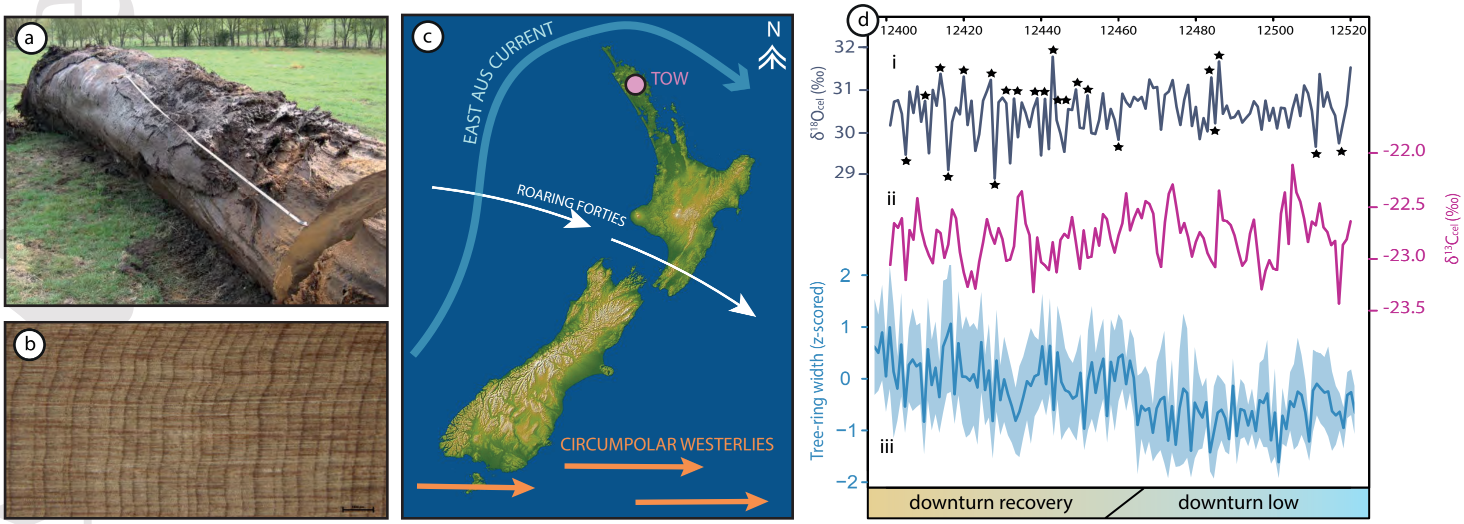
Figure 1. A high-resolution view of the Late Glacial in the Southern Hemisphere. Decadal z-scored chronology (KAU-ISO) of (**a**) $\delta^{13}C_{cel}$ and (b) $\delta^{18}O_{cel}$ and with standard deviation shaded (n_{trees} = 6); (**c**) tree-ring width of full Towai chronology (Palmer et al. 2017) subsetted tolength of KAU-ISO (13 020 - 11 850 cal BP). Kauri downturn (KD) shaded. (**d**) Bars represent individual kauri trees used in this study, with the full length of individual trees in light blue and decadal stable isotope (KAU-ISO) subset in dark blue. (**e**) Lake Hayes sediment record of Ca/Ti (Hinojosa et al. 2019); (**f**) Speleothem $\delta^{13}C$ from New Zealand (Whittaker et al. 2008); (**g**) Antarctic (Talos Dome; Pedro et al. 2011) and (**h**) Greenland (NGRIP, Rasmussen et al. 2013) ice core $\delta^{18}O$ during the Late Glacial with the Antarctic Cold Reversal (ACR) and Greenland Stadial 1 (GS-1; equivalent to the pollen defined Younger Dryas) indicated. Climate stages according to (**i**) Kaigo Bog stratigraphy (Newnham & Lowe 2000; Hajdas 2006), including the Late-glacial mild episode (NZce-4), the Late-glacial cool episode (NZce-3), the Pre-Holocene amelioration (NZce-2) and theHolocene Inter-glacial (NZce-1); Time according to GICC05 (years before 1950).

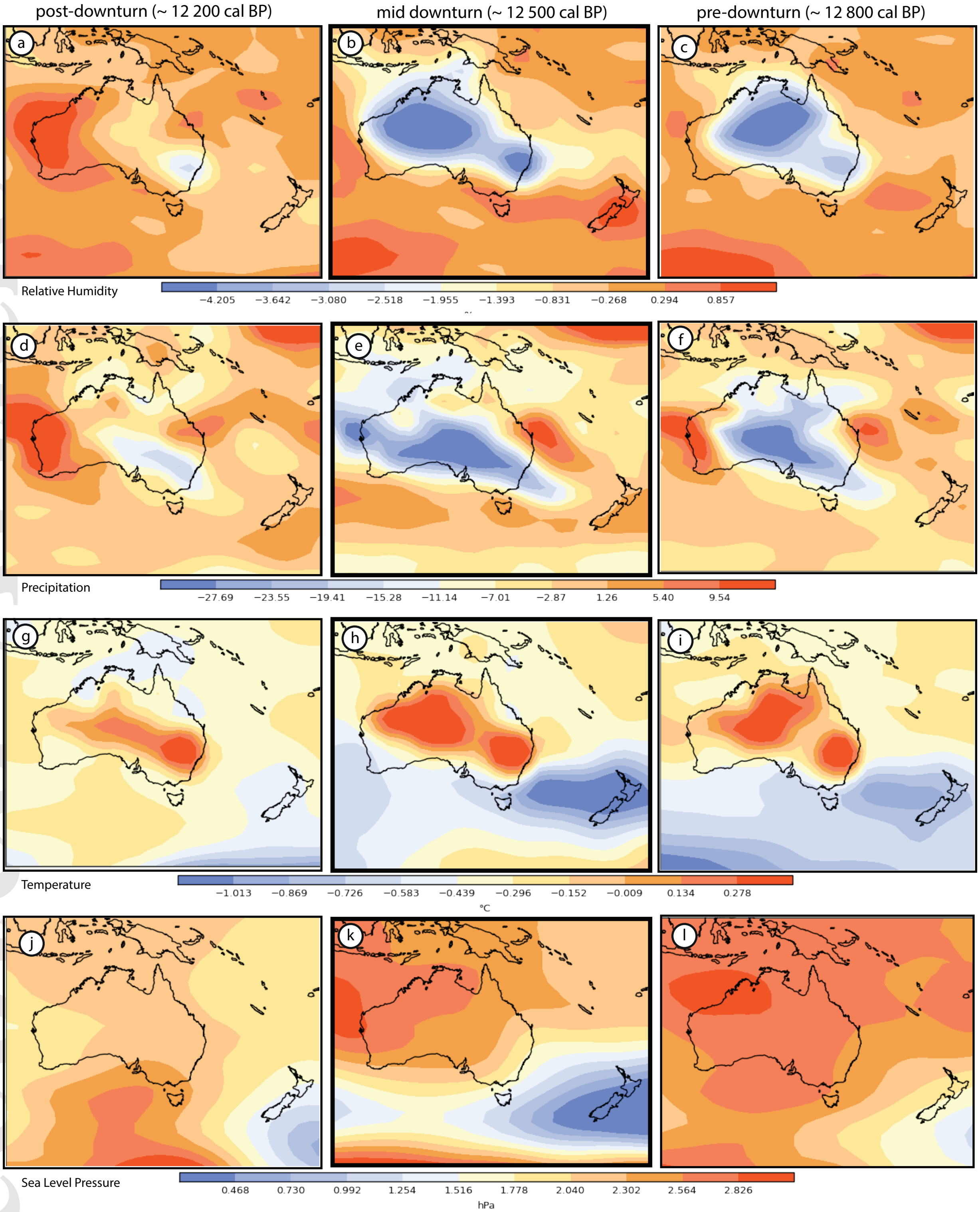
Figure 2. Kauri tree characteristics, site and annual variability. (**a**) Stump of kauri tree found on Towai farm; (**b**) close-up of individual tree-rings; (**c**) climate situation in New Zealand with site location (TOW) indicated; (**d**) annual (**i**) $\delta^{18}O_{cel}$ (**ii**) $\delta^{13}C_{cel}$ (**iii**) and tree-ring width (z-scored) of a single tree during the latter half of the Kauri Downturn; stars represent annual variability >1‰.

Figure 3. The Late Glacial in the South Pacific demonstrated by coupled atmosphere-ocean general circulation TraCE-21ka. (**a-c**) relative humidity, (**d-f**) precipitation, (**g-i**) temperature and (**j-l**) sea level pressure in three periods: pre-downturn (12 800 cal BP), mid-downturn (12 500 cal BP) and post-downturn (12 200 cal BP).



a
b
c
δ13Ccel (‰)
δ18Ocel (‰)
Tree-ring width (z-scored)
12000
12200
12400
12600
12800
13000
d
KD
e
f
g
h
i
drier / endogenic carbonates
wetter / detrital Ca
Ca/Ti Lake Hayes
δ13C Speleothems (‰)
δ18O Talos (‰)
δ18O NGRIP (‰)
ACR
GS-1
warm
mild
cool
NZce-1
NZce-2
NZce-3
NZce-4
10000
11000
12000
13000
14000
15000
Time (years before present)


a
b
c
EAST AUS CURRENT
N
TOW
ROARING FORTIES
CIRCUMPOLAR WESTERLIES
d
12400
12420
12440
12460
12480
12500
12520
32
31
30
29
δ18Ocel (‰)
i
ii
iii
-22.0
-22.5
-23.0
-23.5
δ13Ccel (‰)
2
1
0
−1
−2
Tree-ring width (z-scored)
downturn recovery
downturn low



post-downturn (~ 12 200 cal BP)
mid downturn (~ 12 500 cal BP)
pre-downturn (~ 12 800 cal BP)
a
b
c
Relative Humidity
−4.205 −3.642 −3.080 −2.518 −1.955 −1.393 −0.831 −0.268 0.294 0.857
d
e
f
Precipitation
−27.69 −23.55 −19.41 −15.28 −11.14 −7.01 −2.87 1.26 5.40 9.54
g
h
i
Temperature
−1.013 −0.869 −0.726 −0.583 −0.439 −0.296 −0.152 −0.009 0.134 0.278
°C
j
k
l
Sea Level Pressure
0.468 0.730 0.992 1.254 1.516 1.778 2.040 2.302 2.564 2.826
hPa