

High-precision dating and correlation of ice, marine and terrestrial sequences spanning Heinrich Event 3: Testing mechanisms of interhemispheric change using New Zealand ancient kauri (*Agathis australis*)

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Highlights:

- **Ancient kauri provides a measure of late Pleistocene atmospheric ^{14}C**
- **^{14}C allows precise alignment of ice, marine and terrestrial sequences**
- **We observe weakened ocean circulation during the glacial period**
- **We find no evidence for shutdown in Atlantic overturning during Heinrich Event 3**
- **Antiphase interhemispheric temperature trends do not appear to be driven by the bipolar seesaw**

Keywords: Abrupt climate change; Antarctic Isotope Maximum-4 (AIM4); atmospheric radiocarbon (^{14}C); bipolar seesaw; extreme events; tree-ring; Heinrich-3 (H3)

Abstract

Robustly testing hypotheses of geographic synchronicity of abrupt and extreme change during the late Pleistocene (60,000 to 11,650 years ago) requires a level of chronological precision often lacking in ice, marine and terrestrial sequences. Here we report a bidecadally-resolved New Zealand kauri (*Agathis australis*) tree-ring sequence spanning two millennia that preserves a record of atmospheric radiocarbon (^{14}C) during ice-rafted debris event Heinrich Event 3 (HE3) in the North Atlantic and Antarctic Isotope Maximum 4 (AIM4) in the Southern Hemisphere. Using ^{14}C in the marine Cariaco Basin and ^{10}Be preserved in Greenland ice, the kauri ^{14}C sequence allows us to precisely align sequences across this period. We observe no significant difference between atmospheric and marine ^{14}C records during HE3, suggesting no stratification of surface waters and collapse in Atlantic Meridional Overturning Circulation (AMOC). Instead our results support recent evidence for a weakened AMOC across at least two millennia of the glacial period. Our work adds to a growing body of literature confirming that Heinrich events are not the cause of stadial cooling and suggests changes in the AMOC were not the primary driver of antiphase temperature trends between the hemispheres. Decadally-resolved ^{14}C in ancient kauri offers a powerful new (and complementary) approach to polar ice core CH_4 alignment for testing hypotheses of abrupt and extreme climate change.

1. Introduction

Contrasting Greenland and Antarctic temperature trends during the late Pleistocene (60,000 to 11,650 years ago, hereafter 60 to 11.65 ka) are thought to be driven by imbalances in the rate of formation of North Atlantic and Antarctic deep water (the ‘bipolar seesaw’) (Bond *et al.*, 1993; Broecker, 1998; Stocker and Johnsen, 2003), with abrupt, millennial-duration warming Dansgaard-Oeschger (D-O) events in the north leading to cooling in the south (Blunier and Brook, 2001; EPICA Community Members, 2006; WAIS Divide Project Members, 2015). Central to understanding abrupt and extreme change and their global transmission are Heinrich Events (HE), massive discharges of ice into the North Atlantic over at least the last 60 ka (Lynch-Stieglitz *et al.*, 2014) that appear to coincide with periods of maximum southern warming (Bond *et al.*, 1993; Ganopolski and Rahmstorf, 2001; Hemming, 2004; Kageyama *et al.*, 2013; Menviel *et al.*, 2014). Importantly, however, high-resolution analysis of North Atlantic marine cores has demonstrated that surface cooling preceded the deposition of ice rafted debris (Barker *et al.*, 2015; Bond and Lotti, 1995; de Abreu *et al.*, 2003) and the release of freshwater during HE2 and HE3 had little impact on the Atlantic Meridional Overturning Circulation (AMOC) (Lynch-Stieglitz *et al.*, 2014; Parker *et al.*, 2015); with both still represented in Antarctic ice cores by extreme warming events (Antarctic Isotope Maximum (AIM) 2 and AIM4, approximately 24 and 30 ka respectively). These works suggest iceberg delivery of freshwater was not the cause of the cold periods surrounding Heinrich events (known as Heinrich Stadials or HS) and the origin of the cooling may reside outside the North Atlantic region. Some workers have therefore argued that changes in the south may have played a leading role in driving global change via atmospheric and/or ocean teleconnections (Kanfoush,

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2013; Landais *et al.*, 2015; Pedro *et al.*, 2011; Turney and Jones, 2010). Testing hypotheses of synchronous abrupt climate changes and the mechanism(s) by which they are transmitted globally is a considerable challenge given the relatively large chronological uncertainties associated with what are relatively short transitions (on sub-centennial timescales) (de Abreu *et al.*, 2003; Lynch-Stieglitz *et al.*, 2014).

Inflections in atmospheric radiocarbon (^{14}C) content offer considerable potential for high-precision correlation between terrestrial records (and in suitable settings, marine sequences) (Palmer *et al.*, 2015; Turney *et al.*, 2010; Tzedakis *et al.*, 2007), but significant chronological differences exist between published datasets (Bronk Ramsey *et al.*, 2012; Hughen *et al.*, 2006; Turney *et al.*, 2006). While a contiguous, absolutely-dated record of changing atmospheric radiocarbon concentration across the full timescale is currently unavailable, annually-resolved sub-fossil kauri (*Agathis australis*) trees recovered from bogs in northern New Zealand provide a unique opportunity to capture multi-millennial periods of time (Palmer *et al.*, 2006; Palmer *et al.*, 2015). Excitingly, the use of ^{14}C for high-precision correlation of terrestrial and marine sequences can also be extended to ice core records using ^{10}Be (Adolphi and Muscheler, 2016; Muscheler *et al.*, 2014b; Muscheler *et al.*, 2008). Both cosmogenic radionuclides (^{14}C and ^{10}Be) are produced via a nuclear cascade that is triggered when galactic cosmic rays collide with atmospheric atoms (Lal and Peters, 1967). The flux of galactic cosmic rays is in turn modulated by the strengths of the heliomagnetic and the geomagnetic field, shielding the Earth's atmosphere from galactic cosmic rays. This leads to increased (decreased) ^{10}Be and ^{14}C production rates during times of low (high) solar activity and/or geomagnetic field intensity. Hence, the atmospheric production rates of ^{14}C and ^{10}Be vary simultaneously globally, allowing for the synchronization of cosmogenic radionuclide records across different archives

(Adolphi and Muscheler, 2016; Muscheler *et al.*, 2014a; Muscheler *et al.*, 2014b; Raisbeck *et al.*, 2007).

Here we report a bidecadally-resolved ^{14}C dataset from a 2000-year long New Zealand kauri that spans HE3 and AIM4, which we precisely align to the marine Cariaco Basin and polar ^{10}Be records, allowing us to explore climate-ocean linkages during a period of abrupt and extreme change.

2. Materials and methods

2.1. Ancient kauri and ^{14}C dating

Multiple trees of different ages across the late Pleistocene were extracted from a swamp on Finlayson Farm (35° 50' S, 173° 39' E) in 1998 (Palmer *et al.*, 2015; Turney *et al.*, 2010). Included in this cohort was a 2000-year long kauri log (henceforth 'Finlayson 8'). The site is near the west Northland coast at about 80 m above sea level and is close to several other known collection locations (Figure 1) (Turney *et al.*, 2007). A cross-section (or 'biscuit') was cut and transported to the laboratory where a radial strip was removed for research and the residual offcuts archived. The radial strip was first dried and then sanded, progressively stepping from coarse to fine grit paper until a highly polished surface was obtained. After this step, the radii were studied under a binocular microscope and the annual rings identified and counted. Following this, the rings were measured along the different radial strips using a standard dendrochronology travelling stage with an internal linear encoder. The time series were archived for future cross-dating exercises, while bidecadal samples were then marked on each of the radii. These bidecadal samples were subsequently removed by first making additional finer radial cuts with a band saw and

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then removed using a chisel along the annual boundary (with duplicates). As such, the internal age incorporation for each sample is 20 years. Chronological error due to missing rings is estimated to be <1% and is not considered significant given the timescale involved (Palmer *et al.*, 2006; Turney *et al.*, 2010).

For radiocarbon (^{14}C) dating, chemical pretreatment of the bidecadal wood samples was carried out to extract alpha-cellulose – the wood fraction deemed the most reliable for minimizing potential contamination and providing the most robust ^{14}C ages required for such high-precision study (Hogg *et al.*, 2006). Alpha-cellulose extraction followed the methodology described by Staff *et al.* (2014) at the Oxford Radiocarbon Accelerator Unit (ORAU) (pretreatment code ‘UA’). Specifically, this methodology involved sampling ~75 mg of wood from each bidecadal block with a fine-bladed microplane, shaving across rings, followed by a bleaching stage (1.5% wt/vol. NaClO_2 + 0.06 N HCl) at 70°C, repeated 4 times over 24 hours. There followed a rigorous acid-base-acid (ABA) stage, consisting of successive HCl (4%/1.12 N, for 20 min at 70°C), NaOH (17.5% wt/vol, for 1 hour at room temperature, with ultrasonication and under a constant N_2 environment), and HCl (5%/1.4 N, for 10 min at 70°C) applications. After each stage, samples were rinsed with ultrapure (MilliQ™, Millipore) water (typically, 3 times, 5 times, and 5–6 times, until a neutral pH was achieved). Samples were subsequently freeze-dried, prior to combustion in an Elemental Analyser at 1000 °C (\approx 5 mg of alpha-cellulose weighed into Sn capsules), graphitization (involving H_2 reduction of CO_2 to pure C, graphite, over an Fe catalyst (Vogel *et al.*, 1984) for 6 hr at 560 °C (Dee and Ramsey, 2000) and accelerator mass spectrometry (AMS) stages. AMS measurement at ORAU was performed on the HVEE tandem accelerator mass spectrometer at the University of Oxford (Ramsey *et al.*, 2004). Samples of alpha-cellulose from the Waikato OIS7

kauri background standard (Hogg *et al.*, 2006) were prepared and measured with the unknown age samples. Results on the OIS7 blank ranged from 0 to 0.00077 F¹⁴C (>50,300–>54,900 kyr BP) with a mean of 0.00022 ± 0.00055 F¹⁴C (>53,370 kyr BP), consistent with previous work (Hogg *et al.*, 2006). The quoted uncertainties are compiled from uncertainties in the Background and Modern standards, as well as from the variability in the repeated runs on each sample and from counting statistics.

2.2. *Radiocarbon Calibration and Correlation*

To investigate changes in atmospheric and ocean circulation across HE3 and AIM4, it is crucial the ‘floating’ bidecadally-resolved 2000-year radiocarbon series obtained from Finlayson 8 is compared to independently dated ¹⁴C comparison curves representative of these environments. Unfortunately, whilst the International Calibration 2013 dataset (IntCal13) spans the period of interest here (Reimer *et al.*, 2013), relatively short fluctuations in ¹⁴C concentration have largely been removed during its development, limiting the opportunity to exploit temporal changes in the calibration curve that would allow high-precision alignment of the series (e.g. using so-called ‘radiocarbon plateaux’). The kauri radiocarbon ages were therefore calibrated against two high-resolution radiocarbon datasets: a terrestrial record of atmospheric ¹⁴C generated from the Japanese varved Lake Suigetsu sequence (using the SG06₂₀₁₂ timescale) (Bronk Ramsey *et al.*, 2012), and the Venezuelan Cariaco Basin sedimentary sequence which preserves a record of surface water ¹⁴C and changes in the Intertropical Convergence Zone (ITCZ) that parallel D-O events in Greenland (Cooper *et al.*, 2015; Heaton *et al.*, 2013; Hughen *et al.*, 2006). The ¹⁴C series was calibrated using a Poisson process deposition model (P_Sequence) (Bronk Ramsey and Lee, 2013) with the General Outlier analysis option (Ramsey,

2009) in OxCal 4.2. Using Bayes theorem, OxCal identifies possible age solutions for the kauri series of ^{14}C ages in the Cariaco Basin and Lake Suigetsu calibration curves. Taking into account the ‘deposition’ model and the actual ^{14}C age measurements, the posterior probability densities quantify the most likely age distributions. To accommodate a possible collapse in the marine reservoir age (ΔR) through the Cariaco Basin sequence (420 ^{14}C years), a `Delta_R` with the prior $\text{U}(0,420)$ was used, allowing ^{14}C measurements to assume atmospheric values if required. Whilst neither of the ^{14}C calibration curves used here are bidecadally-resolved across the period spanning HE3 and AIM4, there is sufficient structure in both series to precisely align the Finlayson 8 ^{14}C dataset, providing a chronology for the kauri and in the case of Cariaco Basin, investigate possible changes in marine reservoir ages. The OxCal code used to model the radiocarbon series is provided in Appendix A and the calibrated age solutions given in Appendix B (Supplementary Data). Calibrated ages are reported here as calendar years Before Present (BP).

To compare the kauri ^{14}C series with the Greenland Ice Core Chronology 2005 (GICC05) (Rasmussen *et al.*, 2006; Rasmussen *et al.*, 2014; Seierstad *et al.*, 2014; Svensson *et al.*, 2008; Svensson *et al.*, 2006), we aligned variations in atmospheric radiocarbon concentration ($\Delta^{14}\text{C}$) as recorded by the New Zealand tree against the ^{10}Be measurements from the GRIP ice core (Muscheler *et al.*, 2004; Wagner *et al.*, 2001a; Yiou *et al.*, 1997); $\Delta^{14}\text{C}$ is defined as $^{14}\text{C}/^{12}\text{C}$ after correction for fractionation and decay relative to a standard). We placed the kauri $\Delta^{14}\text{C}$ sequence on the GICC05 chronology using the methodology outlined by Adolphi and Muscheler (2015). Due to carbon cycle effects the atmospheric $\Delta^{14}\text{C}$ signal is dampened and delayed compared to ^{14}C production rate variations. Hence, we modelled $\Delta^{14}\text{C}$ from GRIP ^{10}Be fluxes assuming that these are proportional to global production rate variations using a box-

diffusion carbon cycle model run under preindustrial conditions (Muscheler *et al.*, 2004; Siegenthaler *et al.*, 1980). The long term trends in $\Delta^{14}\text{C}$ can be increasingly affected by carbon cycle changes which are not reflected in ^{10}Be and difficult to quantify (Adolphi and Muscheler, 2016; Muscheler *et al.*, 2004). Due to the radioactive decay of ^{14}C and the relatively old ages the kauri, ^{14}C measurement uncertainties are too large to resolve the centennial $\Delta^{14}\text{C}$ variations that have been used for $^{10}\text{Be}/^{14}\text{C}$ synchronization by Adolphi and Muscheler (2016) during a more recent period. Hence, we high-pass filtered the ^{10}Be -based $\Delta^{14}\text{C}$ record with a cutoff frequency of $1/2000 \text{ years}^{-1}$ and linearly detrended the kauri $\Delta^{14}\text{C}$ data. Given the length of the kauri chronology of 2000 years, this detrending is comparable to a 2000-year high pass filter but it avoids edge-effects induced by filtering. It is difficult to estimate an uncertainty to the GRIP ^{10}Be -based $\Delta^{14}\text{C}$ record. Uncertainties arising from the unknown history of the carbon cycle as shown by Muscheler *et al.* (2004) may be systematic (on the time scales considered here), and would be removed by filtering the modelled $\Delta^{14}\text{C}$. On the other hand, a different state of the ocean's deep convection and/or air-sea gas exchange rates would impact on the amplitude of $\Delta^{14}\text{C}$ variations for a given ^{14}C production rate change (Köhler *et al.*, 2006) which would also be present after filtering. We therefore assigned a 1σ uncertainty of 25% to the GRIP ^{10}Be -based $\Delta^{14}\text{C}$ record. This is consistent with simulated millennial $\Delta^{14}\text{C}$ variations that can be induced by carbon cycle changes alone (Köhler *et al.*, 2006). It should be noted, however, that the derived fits of the kauri sequence onto the modelled $\Delta^{14}\text{C}$ record depend very little in practice on the assumed error for the GRIP ^{10}Be -based record.

To obtain a probability estimate for the GICC05-age of the kauri sequence we aligned the detrended tree $\Delta^{14}\text{C}$ sequence to the modelled GRIP ^{10}Be -based $\Delta^{14}\text{C}$

record using the method published previously (Adolphi and Muscheler, 2016; Bronk-Ramsey *et al.*, 2001). This method is summarized in equations 1 and 2:

Equation 1:

$$P_i(t_s + \Delta t_i) \propto \frac{\exp\left(-\left(R_i - \frac{R(t_s + \Delta t_i)}{\delta R_i^2 + \delta R^2(t_s + \Delta t_i)}\right)^2\right)}{\sqrt{\delta R_i^2 + \delta R^2(t_s + \Delta t_i)}}$$

Equation 2:

$$P_s(t_s) \propto \prod_{i=1}^n P_i(t_s + \Delta t_i)$$

where R_i and δR_i are the kauri $\Delta^{14}\text{C}$ determinations and their corresponding uncertainties; δR_i , and R and δR represent the GRIP ^{10}Be -based $\Delta^{14}\text{C}$ values and their uncertainties. The GICC05 age of the kauri sequence is indicated as t_s while Δt_i represents the age difference between successive tree samples derived from the tree ring series. Hence, $P_i(t_s + \Delta t_i)$ represents a likelihood estimate for the GICC05 ages of each kauri sample, while $P_s(t_s)$ is the overall likelihood of the entire sequence of kauri samples. Details regarding the mathematical formulation of the method and its application to ^{14}C and ^{10}Be records have been published elsewhere (Adolphi and Muscheler, 2016; Bronk-Ramsey *et al.*, 2001). To take into account the relative uncertainty of the ice core time scale we repeated the calculations assuming linear stretches from -100 to 100% of the GICC05 time scale along the maximum counting error in steps of 10% for the period of overlap.

3. Results and Discussion

3.1. Alignment of kauri ^{14}C against ^{14}C datasets

Synchronization of the Finlayson 8 and Cariaco Basin ^{14}C datasets allows us to compare bidecadal atmospheric radiocarbon changes during a period of extreme and abrupt climate change in the tropical and North Atlantic (Figure 2). The 2000-year tree series shows an excellent fit with the Cariaco Basin with agreement indices exceeding the 60% threshold (Agreement index_{model}= 207.5; A_{overall}=138) (Supplementary Table 1). Across the full series, we derive a mean ΔR of 354 ± 28 years but could only align the kauri against 1460 years in the Cariaco Basin (between 28,750 and 30,210 Cariaco BP) with an implied decrease in sedimentation during the periods at the start and end of the Finlayson 8 sequence (Figure 3). We find that the Finlayson 8 sequence falls within HS3 during which time the ITCZ experienced sustained but highly variable changes (Figure 2). As the relative dating error of the tree ring series is essentially negligible (<1% dating error) (Palmer *et al.*, 2006), this implies compression of the published calendar timescale of the Cariaco record by ~500 years, not unsurprising given the non-varved nature of the sediments in this part of the marine sequence (Hughen *et al.*, 2006). Regardless of changes in sedimentation rate, the Finlayson 8 tree spans the stadial preceding Dansgaard-Oeschger 4.

In contrast, the 2000-year Finlayson 8 sequence was calibrated against 1792 years (Agreement index_{model}= 618.3; A_{overall}=442.2) on the Lake Suigetsu timescale (SG06₂₀₁₂ modelled) (Bronk Ramsey *et al.*, 2012) but could be slotted into a period of equal duration (*i.e.*, 2000 years) within the errors of the chronology, spanning 29,505 and 31,505 SG06₂₀₁₂ BP (Figure 2). It was therefore possible to splice the tree measurements of atmospheric ^{14}C into the Japanese varve sequence to construct a bidecadally-resolved radiocarbon comparison dataset spanning the full two millennia.

3.2. *Alignment of kauri ^{14}C against ^{10}Be in Greenland ice*

The calibration of the kauri sequence against the Cariaco Basin (Heaton *et al.*, 2013; Hughen *et al.*, 2006) and the Lake Suigetsu (Bronk Ramsey *et al.*, 2012) ^{14}C records leads to calendar age estimates for the start of the radiocarbon dataset at 30,210 BP and 31,505 BP respectively. Including the GICC05 maximum counting error for these ages we, therefore, investigated kauri-GICC05 links between 29,300 and 32,600 GICC05 BP. Constraining the possible kauri age this way, we obtain two possible matched positions of the tree $\Delta^{14}\text{C}$ record on the GRIP ^{10}Be -based $\Delta^{14}\text{C}$ record (Figures 4 and 5).

When investigating both possible GICC05 links of the kauri sequence in a climatic context it can be seen that the younger link would indicate an overlap between the tree sequence and Dansgaard-Oeschger event 4 (DO-4; Figure 5). This is in conflict with the inferred position of the kauri ^{14}C series in the Cariaco Basin sequence whose radiocarbon ages indicate that the kauri-chronology precedes DO-4. It should be kept in mind that the comparison to GRIP ^{10}Be -based $\Delta^{14}\text{C}$ is based solely on $\Delta^{14}\text{C}$ variations and cannot account for the absolute level of $\Delta^{14}\text{C}$ due to uncertainties in the long term evolution of the carbon cycle (Köhler *et al.*, 2006). The Cariaco Basin record indicates significantly younger ^{14}C ages of DO-4 than the tree sequence that are impossible to reconcile with reservoir age changes or measurement uncertainties. Based on this inference, we exclude the younger position (Figure 4A) of the kauri sequence on GICC05 and, thus, can establish a link of the tree record on GICC05 at 31,400 GICC05 BP (95% range = 31,240 – 31,470 BP; Figure 4B). The resulting modelled $\Delta^{14}\text{C}$ is also consistent with the GRIP ^{10}Be concentration (Figure 6) (Wagner *et al.*, 2001b).

3.3. *Testing mechanisms of change*

To robustly test mechanisms of synchronous (or asynchronous) interhemispheric change across the period 31,400 to 29,400 BP it is critical that the Cariaco Basin, Greenland and the West Antarctic Ice Sheet (WAIS) Divide (WAIS Divide Project Members, 2015) records are precisely correlated. Fortunately, the impact of Heinrich Events on North Atlantic sea ice extent has recently been recognized by parallel increases in the $\delta^{18}\text{O}$ difference across NGRIP, GRIP and GISP2 records (Seierstad *et al.*, 2014). Although HE3 was reportedly less clear than other Heinrich events, the divergence in $\delta^{18}\text{O}$ values between all three ice core records can be recognised between D-O events 4 and 5.1, strongly suggesting the expression of HE3 (Figure 7), allowing us to place the timing of this event against radiocarbon ages using the link of the kauri ^{14}C sequence on GICC05 as established by the $^{14}\text{C}/^{10}\text{Be}$ comparison (section 3.2). We therefore date the onset of HE3 in Lake Suigetsu at $29,965 \pm 62$ BP (SG06₂₀₁₂ timescale), coincident with the onset of a multi-centennial duration radiocarbon plateau centred on $\sim 25,400$ ^{14}C BP.

Crucially, the greenscale of the Cariaco Basin sediments preserves a record of marked changes in the location of the ITCZ over the equatorial Atlantic (Figures 2 and 7). We recognise a shift in greenscale beginning around $29,500 \pm 133$ cal. BP on the Cariaco Basin timescale (Heaton *et al.*, 2013; Hughen *et al.*, 2006; Tzedakis *et al.*, 2007) suggesting a sustained equatorward migration of the ITCZ, consistent with an increased latitudinal temperature gradient [in the northern Atlantic](#) during HE3 (Cooper *et al.*, 2015; Rind, 2000; Wang *et al.*, 2004). On closer investigation, the maximum shift in the ITCZ lies stratigraphically above the onset of the radiocarbon plateau at $25,400$ ^{14}C BP (Figure 2) identified by the kauri-Greenland correlation (Figure 5), implying a decadal-duration lag in the response of the climate system to the latitudinal temperature gradient (Buizert *et al.*, 2015; Rind, 2000). As a result, the

bidecadally-resolved kauri ^{14}C allows us to synchronize the Greenland, Cariaco Basin and (via CH_4) the WAIS sequences on a common timescale (Figure 8).

We find that northern cooling associated with the HS3 did not commence with HE3 but occurred earlier (Figure 8), in support of the observations that ice rafted debris events were not the cause of stadial cooling (Barker *et al.*, 2015; Bond and Lotti, 1995; de Abreu *et al.*, 2003). Furthermore, comparison of southern warming as preserved in WAIS did not commence with freshwater input into the North Atlantic (Ganopolski and Rahmstorf, 2001; Menviel *et al.*, 2014; Stocker and Johnsen, 2003), but rather the onset of AIM4 preceded HE3 by about 500 years (Figure 8).

Importantly, recent work has suggested that Cariaco Basin ^{14}C is sensitive to Atlantic Ocean stratification associated with a slowdown in the AMOC (Southon *et al.*, 2012); for instance, during HE1 and the onset of the Younger Dryas stadial, surface ocean ^{14}C closely assumed atmospheric radiocarbon values as a result of greater air-sea mixing, leading to a collapse in the 420 year marine reservoir age. In contrast to HE1 and the Younger Dryas, however, we do not observe a sustained shift to significantly younger ages in the Cariaco Basin ^{14}C record that can be reconciled with the above timescale compression during HE3 (Figure 2A). Importantly, the ΔR of 354 ± 28 years derived from comparing Finlayson 8 with Cariaco Basin is close to the ‘modern’ reservoir age of 420 years (Hughen *et al.*, 2006) consistent with a weakened AMOC across the full two millennia and suggesting negligible (if any) change during HE3 (Lynch-Stieglitz *et al.*, 2014; Parker *et al.*, 2015).

The above raises important implications for understanding the timing and transmission of abrupt global change. While recent work has suggested North Atlantic cooling may be related to gradual freshening of surface North Atlantic waters and/or expansion of ice shelves (Barker *et al.*, 2015), this does not readily explain the

contrasting temperature response in the Southern Hemisphere. Given the dampened nature of AMOC in the North Atlantic at this time, the absence of any appreciable change across HE3 (Lynch-Stieglitz *et al.*, 2014; Parker *et al.*, 2015) and the parallel trends in atmospheric and equatorial marine Atlantic ^{14}C (Figures 2 and 7), it seems unlikely high-latitude temperature trends were driven by an ocean bipolar seesaw (Broecker, 1998; EPICA Community Members, 2006; WAIS Divide Project Members, 2015). Instead, the observation of AIM4 warming unambiguously preceding HE3 but apparently synchronous with North Atlantic stadial cooling does appear to be more consistent with a global atmospheric transmission of change (Liu and Alexander, 2007; Pedro *et al.*, 2011; Timmermann *et al.*, 2005). Regardless of the precise trigger(s) and means of global teleconnection, we demonstrate decadal to bidecadal changes in ^{14}C preserved millennial-duration ancient kauri offers a powerful new (and complementary) approach to polar ice core CH_4 alignment for testing hypotheses of abrupt and extreme climate change.

4. Conclusions

Here we report a 2000-year long sub-fossil kauri (*Agathis australis*) tree from New Zealand that provides a bidecadally-resolved record of atmospheric radiocarbon (^{14}C) content across the period of Heinrich Event 3 (HE3). This unique subfossil record of changing atmospheric ^{14}C provides a means of precisely and accurately aligning ice, marine and terrestrial records across two millennia of abrupt change (within 60 years at 1σ). Comparison between the Cariaco Basin and kauri ^{14}C datasets shows no evidence of surface ocean water stratification, implying no collapse or reduction in Atlantic Meridional Overturning Circulation across the 2000 year long record, including HE3. Alignment with polar records suggests HE3 was not the cause of the

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onset of stadial cooling (between Dansgaard-Oeschger events 5.1 and 4). Our results imply that high-latitude temperatures trends around the time of HE3 were not driven by ocean teleconnections and suggest a role for atmospheric transmission of climate signals.

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Figures and Captions

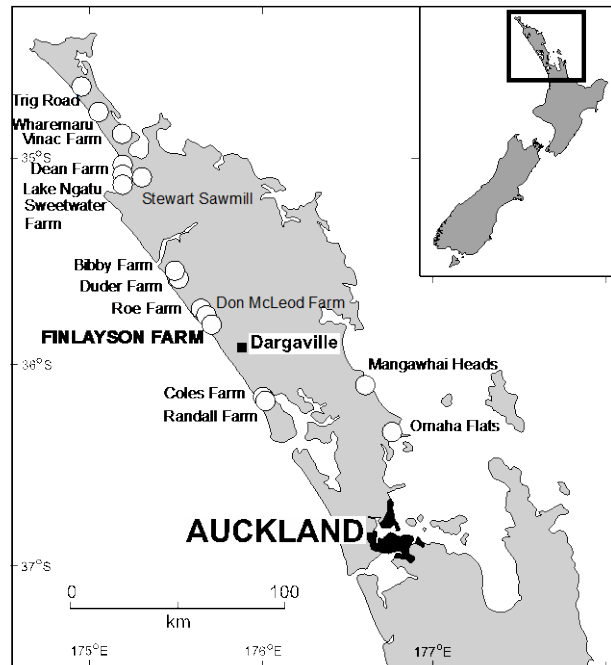


Figure 1: Location of the ancient kauri (*Agathis australis*) site Finlayson Farm, Northland, New Zealand. Other sites with floating radiocarbon dates series from Oxygen Isotope Stage 3 are also shown.

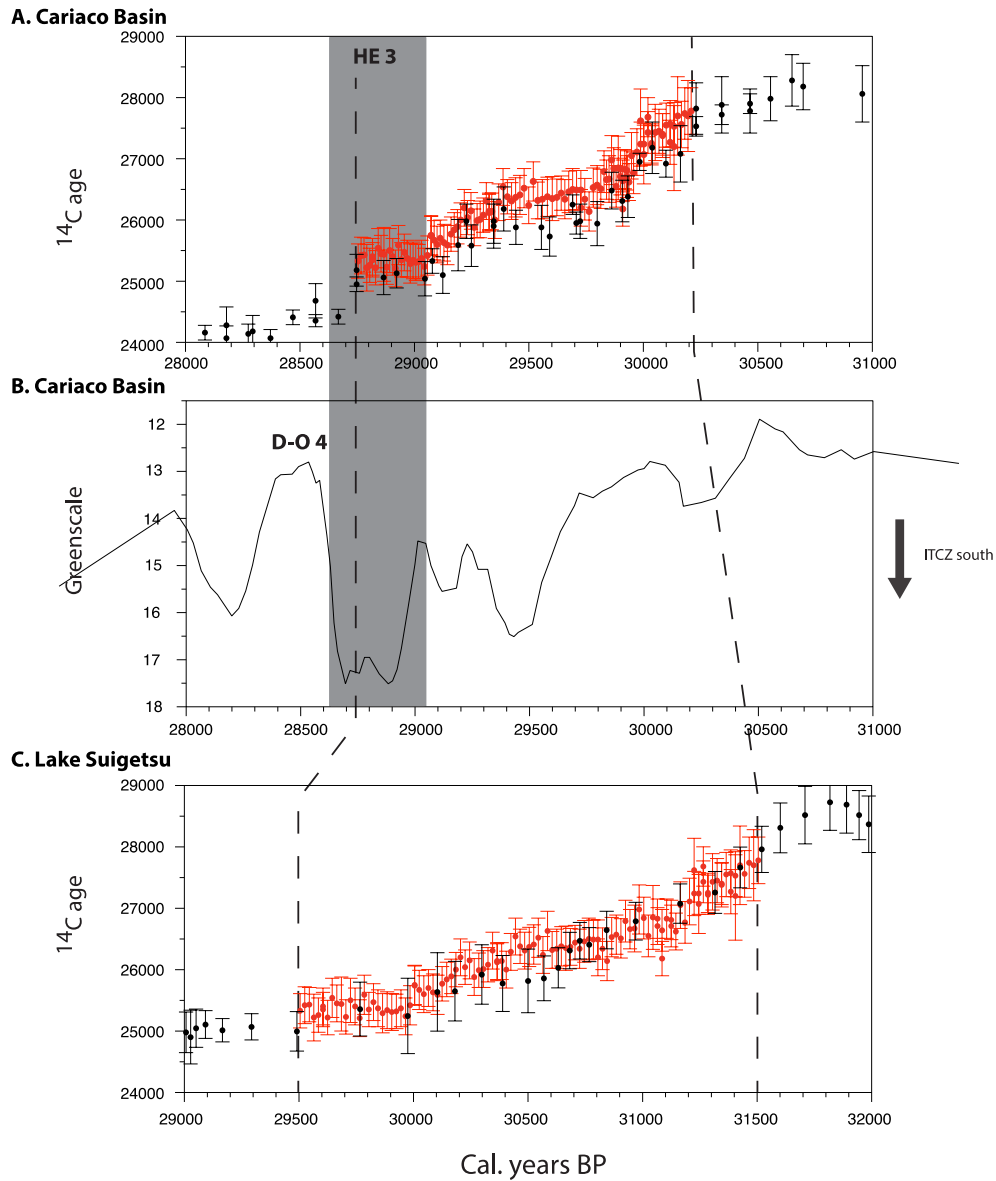


Figure 2: Comparison between Finlayson 8 (filled red circles) and marine-reservoir corrected Cariaco Basin radiocarbon (^{14}C) (Hulu Cave timescale; filled black circles) (Panel A.), associated changes in the Intertropical Convergence Zone (ITCZ) as recorded by Cariaco Basin greenscale (B.) (Heaton *et al.*, 2013; Hughen *et al.*, 2006) and Lake Suigetsu radiocarbon (on the SG06₂₀₁₂ timescale; filled black circles) (Bronk Ramsey *et al.*, 2012) (C.). Error bars denote 2σ errors (95% confidence limits). The timing of North Atlantic cooling during Heinrich Event 3 (HE3) in Cariaco Basin is shown by a grey column.

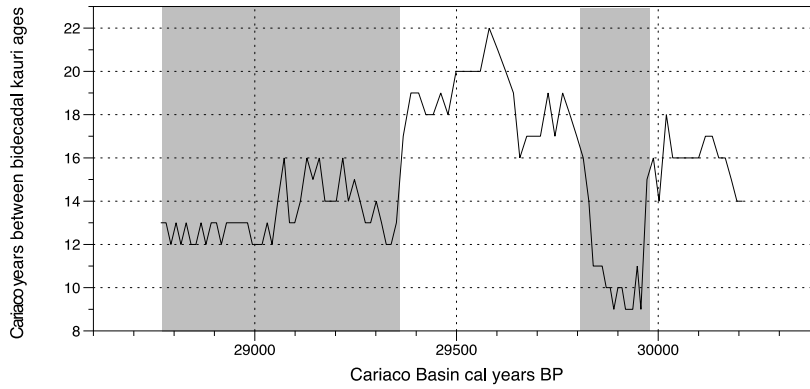


Figure 3: Differences in calendar age across Finlayson 8 calibrated against the Cariaco Basin ^{14}C sediment sequence. Grey columns denote sections in the Cariaco Basin with reduced sedimentation rate compared to the published age scale.

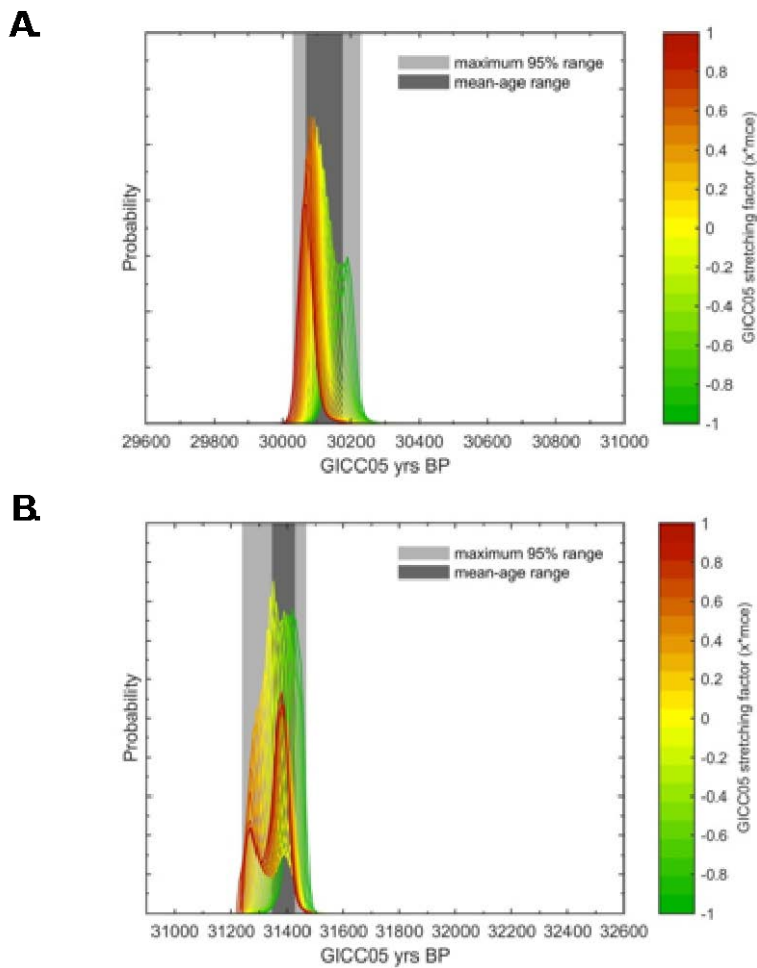


Figure 4: Probability distributions for two different positions of the kauri $\Delta^{14}\text{C}$ sequence on GICC05. Each colour corresponds to a different linear stretching of the

GICC05 time scale across the period of overlap. The light grey bars indicate the lower and upper limits of the 95% probability intervals of all probability distributions (Panel A. = 30,030 – 30,230 BP; Panel B = 31,240 – 31,470 BP). The dark grey bars indicate the upper and lower limits of the derived mean ages (A. = 30,070 – 30,176 BP; B. = 31,345 – 31,430 BP). All ages correspond to the start of the kauri sequence, i.e. the oldest bidecadal sample.

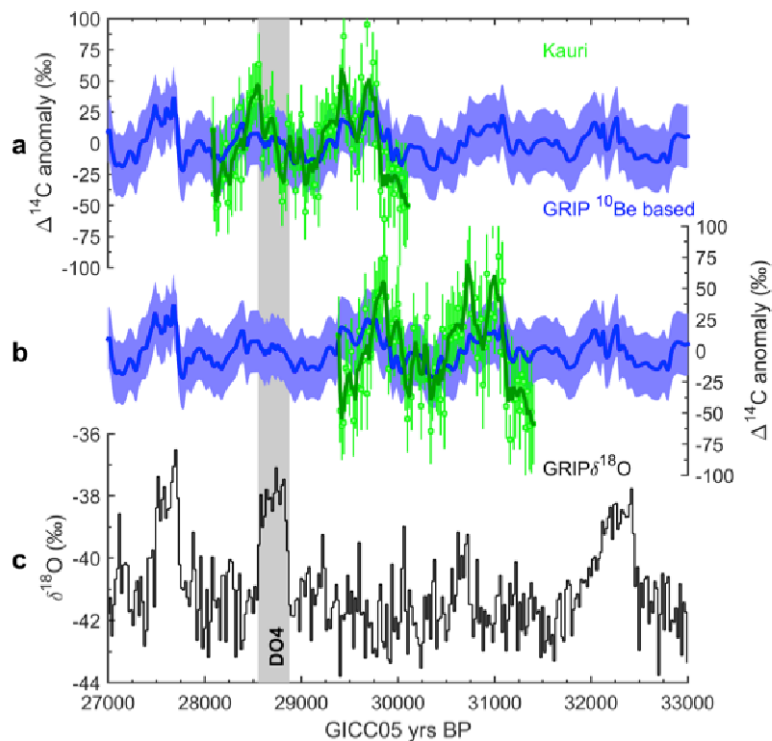


Figure 5: Comparison of the inferred positions of the kauri sequence (Panels a and b) on GICC05 in comparison to Greenland climate (Panel c) (Johnsen et al., 1995). The blue line shows the high-pass filtered GRIP ^{10}Be -based $\Delta^{14}\text{C}$ record. The light green line shows the detrended kauri $\Delta^{14}\text{C}$ data and a 3 point-running mean (dark green line). The black line shows the GRIP $\delta^{18}\text{O}$ record as a proxy for Greenland climate. The vertical grey bar highlights the position of DO-4.

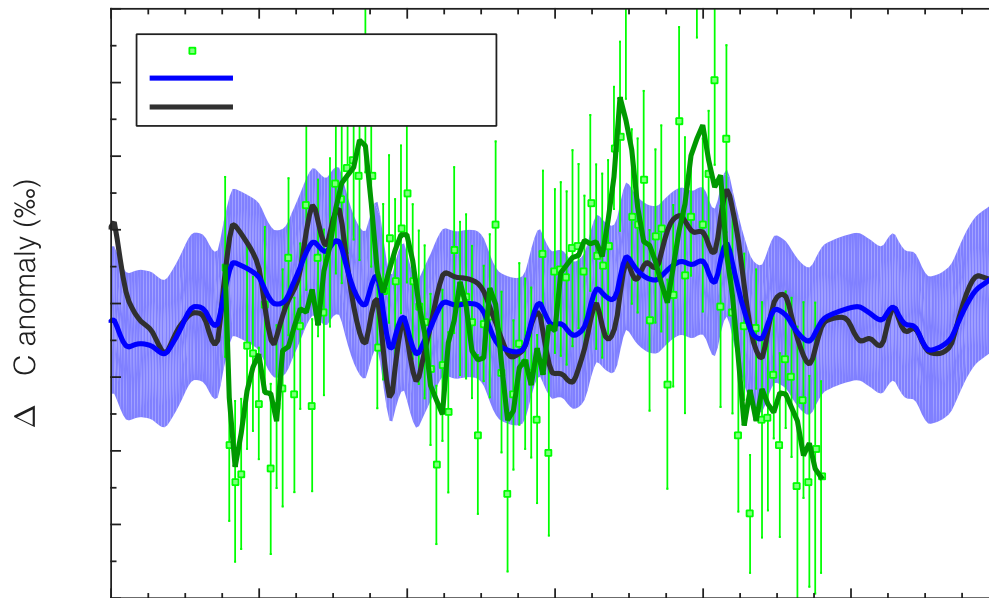


Figure 6: Comparison of ^{10}Be flux (blue) and concentration (black)-derived $\Delta^{14}\text{C}$ anomalies to the kauri $\Delta^{14}\text{C}$ data (green) at its most likely GICC05 age (see text). The ^{10}Be concentrations have been detrended against ice core $\delta^{18}\text{O}$ to correct for obvious climatic influences on ^{10}Be deposition prior to carbon cycle modelling (Muscheler *et al.*, 2004).

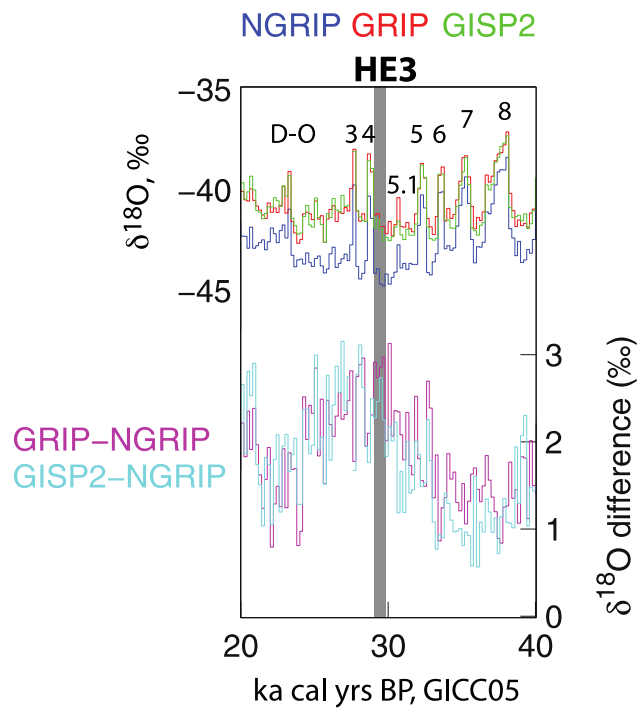


Figure 7: $\delta^{18}\text{O}$ differences between Greenland ice cores (NGRIP, GRIP and GISP2) allowing identification of Heinrich Event 3 (HE3) (Seierstad *et al.*, 2014). Associated Dansgaard-Oeschger (D-O) events are also shown.

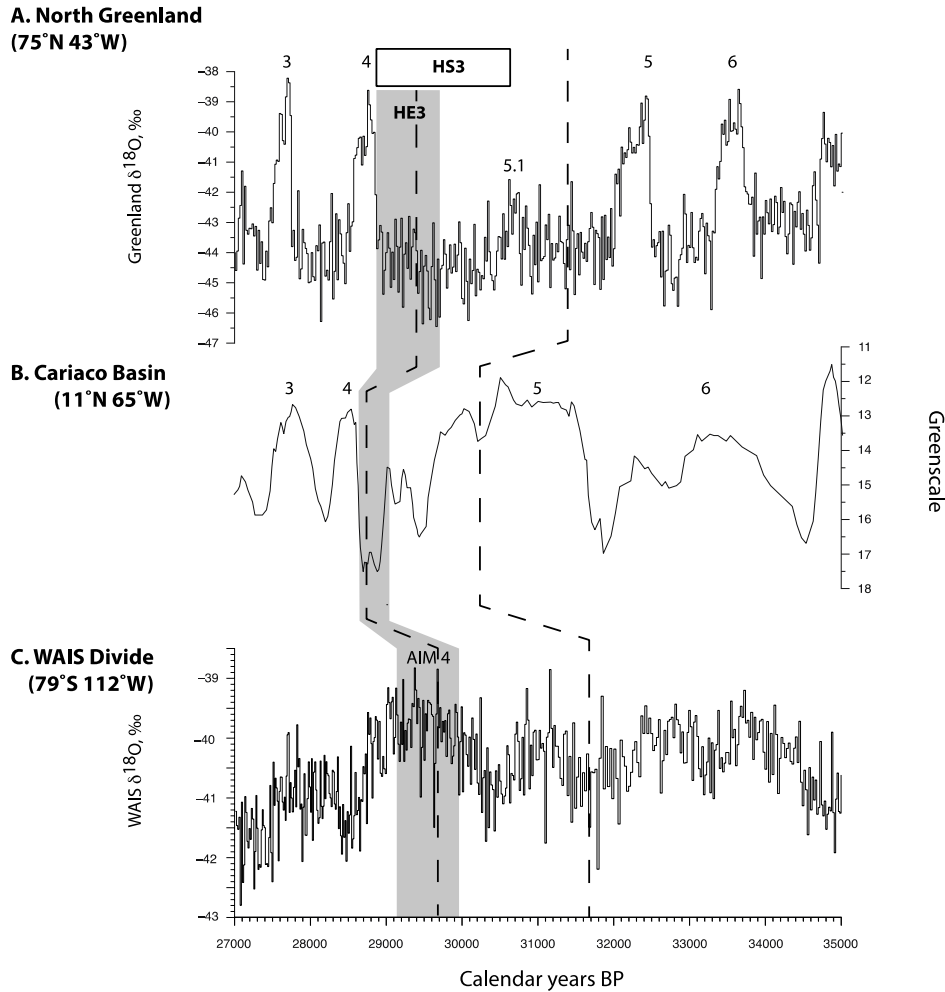


Figure 8: Comparison between North Greenland $\delta^{18}\text{O}$ on the GICC05 timescale (Panel A.) (Svensson *et al.*, 2008), Cariaco Basin greenscale on the modelled Hulu Cave timescale (B.) (Hughen *et al.*, 2006) and the West Antarctic Ice Sheet Divide (WAIS) $\delta^{18}\text{O}$ on the WD2014 timescale (Broecker, 1998; EPICA Community Members, 2006) (C.). The timing of Heinrich Event 3 (HE3) is given by the grey column; the age range of Heinrich Stadial 3 (HS3) is defined above Panel A.; the numbers above the isotope and greenscale curves denote Dansgaard-Oeschger (D-O) and Antarctic Isotope Maximum (AIM) events. The dashed lines denote the time range of the Finlayson 8 kauri ^{14}C record.

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