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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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 (14C) during ice-rafted debris event Heinrich Event 3 (HE3) in the North Atlantic and Antarctic Isotope Maximum 4 (AIM4) in the Southern Hemisphere. Using 14C in the marine Cariaco Basin and 10Be preserved in Greenland ice, the kauri 14C sequence allows us to precisely align sequences across this period. We observe no significant difference between atmospheric and marine 14C 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 14C in ancient kauri offers a powerful new (and complementary) approach to polar ice core CH4 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 11.6 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, 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 (14C) 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 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, 2015). Excitingly, the use of 14C for high-precision correlation of terrestrial and marine sequences can also be extended to ice core records using 10Be (Adolphi and Muscheler, 2016; Muscheler et al., 2014b, 2008). Both cosmogenic radionuclides (14C and 10Be) 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) 10Be and 14C production rates during times of low (high) solar activity and/or geomagnetic field intensity. Hence, the atmospheric production rates of 14C and 10Be vary simultaneously globally, allowing for the synchronization of cosmogenic radionuclide records across different archives (Adolphi and Muscheler, 2016; Muscheler et al., 2014a, 2014b; Raisbeck et al., 2007).

Here we report a bidecadally-resolved 14C 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 10Be 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 14C 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 (Fig. 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, 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 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 (14C) 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 14C 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 + 0.06 N HCl) at 70 °C, repeated 4 times over 24 h. There followed a rigorous acid-base-acid (ABA) stage, consisting of successive HCI (4%/1.12 N, for 20 min at 70 °C), NaOH (17.5% wt/vol, for 1 h at room temperature, with ultrasonication and under a constant N2 environment), and HCI (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 (∼5 mg of alpha-cellulose weighed into Sn capsules), graphitization involving H2 reduction of CO2 to pure C, graphite, over an Fe catalyst (Vogel et al., 1984) for 6 h at 560 °C (Dee and Bronk 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 (Bronk Ramsey et al., 2004). Samples of alpha-cellulose from the Waikato IS7 kauri background standard (Hogg et al., 2006) were prepared and measured with the unknown age samples. Results on the IS7 blank ranged from 0 to 0.00077 F 14C (>50,300±54,900 kyr BP) with a mean of 0.00022 ± 0.00055 F 14C (>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 plateau’) The kauri radiocarbon ages were therefore calibrated against two high-resolution radiocarbon datasets: a terrestrial record of...
atmospheric $^{14}$C 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 $^{14}$C 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 $^{14}$C series was calibrated using a Poisson process deposition model (P_Sequence) (Bronk Ramsey and Lee, 2013) with the General Outlier analysis option (Bronk 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 ($\Delta R$ of 420 $^{14}$C years) through the Cariaco Basin sequence, a Delta_R with the prior 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 $^{14}$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 and the calibrated age solutions are given in 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, 2014; Seierstad et al., 2014; Svensson et al., 2008, 2006), we aligned variations in atmospheric radiocarbon concentration ($\Delta^{14}$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); $\Delta^{14}$C is defined as $^{14}$C/$^{12}$C after correction for fractionation and decay relative to a standard. We placed the kauri $\Delta^{14}$C sequence on the GICC05 chronology using the methodology outlined by Adolphi and Muscheler (2016). Due to carbon cycle effects the atmospheric $\Delta^{14}$C signal is dampened and delayed compared to $^{14}$C production rate variations. Hence, we modelled $\Delta^{14}$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 $\Delta^{14}$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 $\Delta^{14}C$ variations that have been used for $^{10}\text{Be}/^{14}C$ synchronization by Adolphi and Muscheler (2016) during a more recent period. Hence, we high-pass filtered the $^{10}\text{Be}$-based $\Delta^{14}C$ record with a cutoff frequency of $1/2000$ years$^{-1}$ and linearly filtered the $^{10}\text{Be}$-based $\Delta^{14}C$ record with a cutoff frequency of $1/2000$ years$^{-1}$ and linearly.

Fig. 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 SG06,012 timescale; filled black circles) (C.) (Bronk Ramsey et al., 2012). Error bars denote 2$\sigma$ errors (95% confidence limits). The timing of North Atlantic cooling during Heinrich Event 3 (HE3) in Cariaco Basin is shown by a grey column. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article).

Fig. 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.
detrended the kauri $^{14}$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 $^{14}$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 $^{14}$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 $^{14}$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 $1 \sigma$ uncertainty of 25% to the GRIP $^{10}$Be-based $^{14}$C record. This is consistent with simulated millennial $^{14}$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 $^{14}$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 $^{14}$C sequence to the modelled GRIP $^{10}$Be-based $^{14}$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):

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

(1)

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

(2)
The application to \(^{14}\text{C}\) and \(^{10}\text{Be}\) records have been published elsewhere regarding the mathematical formulation of the method and its likelihood of the entire sequence of kauri samples. Details the tropical and North Atlantic (Fig. 2). The 2000-year tree series sets allows us to compare bidecadal atmospheric radiocarbon shows an excellent fit into account the relative uncertainty of the ice core time scale we (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 the GICC05 ages of each kauri sample, while \(R_i\) is the overall likelihood of the entire sequence of kauri samples. Details regarding the mathematical formulation of the method and its application to \(^{14}\text{C}\) and \(^{10}\text{Be}\) 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 \(^{14}\text{C}\) against \(^{14}\text{C}\) datasets

Synchronization of the Finlayson 8 and Cariaco Basin \(^{14}\text{C}\) datasets allows us to compare bidecadal atmospheric radiocarbon changes during a period of extreme and abrupt climate change in the tropical and North Atlantic (Fig. 2). The 2000-year tree series shows an excellent fit with the Cariaco Basin with agreement indices exceeding the 60% threshold (Agreement index\(_{\text{model}} = 207.5; \text{A}_{\text{overall}} = 138\) (Supplementary Data Table 1). Across the full series, we derive a mean \(\Delta R\) 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 (Fig. 3). We find that the Finlayson 8 sequence falls within HS3 during which time the ITZC experienced sustained but highly variable changes (Fig. 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 stadal preceding Dansgaard-Oeschger 4.

In contrast, the 2000-year Finlayson 8 sequence was calibrated against 1792 years (Agreement index\(_{\text{model}} = 618.3; \text{A}_{\text{overall}} = 442.2\) 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 to 31,505 SG062012 BP (Fig. 2). It was therefore possible to splice the tree measurements of atmospheric \(^{14}\text{C}\) into the Japanese varve sequence to construct a bidecadally-resolved radiocarbon comparison dataset spanning the full two millennia.

3.2. Alignment of kauri \(^{14}\text{C}\) against \(^{10}\text{Be}\) 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) \(^{14}\text{C}\) 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 \(^{14}\text{C}\) record on the GRIP \(^{10}\text{Be}\)-based \(^{14}\text{C}\) record (Figs. 4 and 5).

When investigating both possible GICC05 links of the kauri sequence on 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; Fig. 5). This is in conflict with the inferred position of the kauri \(^{14}\text{C}\) 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 \(^{10}\text{Be}\)-based \(^{14}\text{C}\) is based solely on \(^{14}\text{C}\) variations and cannot account for the absolute level of \(^{14}\text{C}\) due to uncertainties in the long term evolution of the carbon cycle (Köhler et al., 2006). The Cariaco Basin record indicates significantly younger \(^{14}\text{C}\) 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 (Fig. 4A) of the kauri sequence on GICC05 and thus can establish a link of the tree record at 31,400 GICC05 BP (95% range = 31,240–31,470 BP; Fig. 4B). The resulting modelled \(^{14}\text{C}\) is also consistent with the GRIP \(^{10}\text{Be}\) concentration (Fig. 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 recognised by parallel increases in the \(^{18}\text{O}\) difference across NGRIP, GRIP and GISP2 records (Seierstad et al., 2014). Although HE3 was reportedly less clear than other Heinrich events, the divergence in \(^{18}\text{O}\) values between all three ice core records can be recognised between D-O events 4 and 5.1, strongly suggesting the expression of HE3 (Fig. 7), allowing us to place the timing of this event against radiocarbon ages using the link of the kauri \(^{14}\text{C}\) 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 ± 62 BP (SG062012 timescale), coincident with the onset of a multi-centennial duration radiocarbon plateau centred on ~25,400 \(^{14}\text{C}\) BP.

Crucially, the greenscale of the Cariaco Basin sediments preserves a record of marked changes in the location of the ITZC over the equatorial Atlantic (Figs. 2 and 7). We recognise a shift in greenscale beginning around 29,500 ± 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 P (Fig. 2) identified by the kauri-Greenland correlation (Fig. 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 (Fig. 8).

We find that northern cooling associated with the HS3 did not commence with HE3 but occurred earlier (Fig. 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 (Fig. 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.

**Fig. 8.** Comparison between North Greenland \(^{18}\)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) \(^{18}\)O on the WD2014 timescale (C.) (Broecker, 1998; EPICA Community Members, 2006). 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.
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 (Fig. 2A). Importantly, the $\Delta$T 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 (Figs. 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 AM4 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 $^{14}$CH$_2$ analysis 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 Overturining Circulation across the 2000 year long record, including HE3. Alignment with polar records suggests HE3 was not the cause of the onset of stadial cooling (between Dansgaard-Oeschger events 5.1 and 5.4). 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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Supplementary data

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