Southern Hemisphere Climate Variability Forced by Northern Hemisphere Ice Sheet Topography

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The presence of large Northern Hemisphere ice sheets and reduced greenhouse gas concentrations during the Last Glacial Maximum (LGM) fundamentally altered global ocean-atmosphere climate dynamics[1]. Model simulations and paleoclimate records suggest that glacial boundary conditions affected the El Niño-Southern Oscillation (ENSO)[2,3], a dominant source of short-term global climate variability. Yet, little is known about changes in short-term climate variability at mid-to high latitudes. Here, we use an ultra-high resolution water isotope record from West Antarctica to demonstrate that interannual to decadal climate variability at high southern latitudes was almost twice as large at the LGM compared to the Holocene. Climate model simulations indicate that this increased variability reflects an increase in the teleconnection strength between the tropical Pacific and West Antarctica, owing to a shift in the mean location of tropical convection. This shift, in turn, can be attributed to the influence of topography and albedo of the North American ice sheets on atmospheric circulation. As the planet deglaciated, the largest and most abrupt decline in teleconnection strength occurred between ~16 and 15 ka, followed by a slower decline into the early Holocene.

During glacial-interglacial transitions, the coupled ocean-atmosphere system shifts between stable states, driven by large climate forcings, including Milankovitch orbital cycles, greenhouse gas concentrations, and the decay of continental ice sheets[1]. The state changes are observed in paleoclimate proxy archives such as ice cores at high latitudes, and lake sediments and speleothems in the tropics and mid-latitudes. The state of the tropical Pacific is particularly important because this region plays a role in the generation and communication of climate anomalies between the low and high latitudes[4,5].

Whether the nature of ENSO changed between the Last Glacial Maximum (LGM) and the Holocene is an important question[3,6]. However, proxy studies of ENSO variability during the LGM remain contradictory[7,8], and the results of modeling studies are similarly ambiguous[2]. One obstacle to understanding past ENSO variability is disentangling changes in ENSO itself from changes in its influence on climate outside the tropics, i.e. the strength of ENSO-related teleconnections, which can vary over time[9].

Tropical Pacific climate variability is an important driver of interannual to multidecadal climate variability in West Antarctica[4,5]. Warm sea surface temperatures (SSTs) in the tropical Pacific drive the propagation of atmospheric

Rossby waves towards the high latitudes, affecting weather systems that develop over the Amundsen Sea[4,5]. Variability in the Amundsen Sea region dominates climate variability over the adjacent West Antarctic Ice Sheet (WAIS)[5,10,11]. Over longer timescales, the influence of tropical climate variability on the Amundsen Sea may have implications both for the stability of the ice sheet[12] and for the ventilation of CO_2 from the Southern Ocean[13].

The influence of tropical variability on West Antarctica climate is reflected in ice core records from the WAIS. In particular, oxygen and hydrogen isotope ratios of water are sensitive to temperature and atmospheric circulation anomalies associated with large ENSO events[14]. Such records might therefore yield important insights about changes in tropical climate variability and tropical-to-extratropical teleconnections during glacial-interglacial transitions. In this study, we utilize a high-resolution hydrogen isotope (δD) record from the WAIS Divide ice core (WDC) to investigate ~31 kyr of high-frequency climatic variability. The data were produced using laser absorption spectroscopy coupled with a continuous flow analysis system[15]. The record is the longest continuously-measured water isotope record ever recovered, with an effective sample resolution of 5 mm and a maximum difference in age between consecutive samples of 0.33 years.

We use spectral techniques (see Methods) on the WDC δD record to analyze West Antarctic climate variability (Fig. 1). Periods greater than 3 years are preserved throughout the 31 kyr record (Extended Data Fig. 1a). To account for attenuation of the δD signal by diffusion in the firn column and solid ice[16], we apply a diffusion correction on consecutive 500-yr data windows (see Methods; Extended Data Fig. 2c). We isolate the amplitude of 3-7 and 4-15 yr δD variations by averaging power density over these frequency ranges. The 3-7 yr range is typical of a high-frequency spectral peak seen in observations of the Southern Oscillation Index (SOI), while the 4-15 yr range is less affected by diffusion (Extended Data Fig. 1a) and captures the decadal variability observed in the SOI (Extended Data Fig. 1c), tropical Pacific coral isotope records[17], and in modern observations in West Antarctica[5,14].

We find that the mean amplitudes of the diffusion-corrected variability in the 3-7 and 4-15 yr bands are elevated in the Late Glacial (16-31 ka) relative to the Holocene (0-10 ka) (Fig. 1c,d). A ~50% decrease in amplitude occurs from ~16 to 10 ka in both time series. This decline occurs in two steps: an abrupt decrease at ~16 ka followed by a second but less pronounced decline from ~13-10 ka. Neither time series reveal other significant changes in variability during the Holocene or during the Late Glacial. We use an objective algorithm (see Methods) to detect initial significant decline in the diffusion-corrected 4-15 yr time series based on 500-yr data windows (Extended Data Fig. 2a,b). We find that the decline began at 16.44 ka ±0.30 kyr (all ages are given relative to present = 1950 C.E.). Identical results are obtained using the 3-7 yr time series. The timing of the initial decline is robust among different detection techniques. Using a sub-set of 100-yr data windows for the diffusion-corrected 4-15 yr time series (Extended Data Fig. 2c,d), the decline occurs at 16.24 ka ±0.17 kyr. The timing of the decline is not an artifact of the diffusion correction. Using the raw data in the 4-15 yr band for 500-yr data windows (Fig. 1d), a ~40% decrease in amplitude occurs from ~16 to 10 ka, with initial significant decline occurring at 15.94 ka ±0.30 kyr. Conservatively, this places the decline in WDC variability between 16.74 to 15.64 ka, with a central estimate of 16.24 ka.

The climate shift at ~16 ka coincides with large-scale geologic events observed in proxy records across the globe. A substantial drawdown of the Laurentide Ice Sheet (LIS) surface is estimated to have occurred at 16 ka BP (from ^{14}C dating)[18], corresponding to the Heinrich Event 1 (H1) iceberg discharge event in the Northern Hemisphere (Fig. 2). The H1 ice rafted debris is of Laurentide origin[19], indicating abrupt changes in ice mass and the topography of the LIS at this time. The Cordilleran ice sheet also declined rapidly at this time[20], and the drawdown of the combined Laurentide-Cordilleran ice sheets (LCIS) has been suggested as a possible forcing mechanism for the transition from dry to wet conditions in Indonesia[21] (Fig. 2c). Taken together, the similarity in timing of these various records suggests a common link.

To investigate the mechanisms responsible for changing high-frequency variability at WDC, we use the HadCM3 fully-coupled ocean-atmosphere model to simulate climate in thousand-year steps from 28 to 0 ka[22] (see Methods). We assess variability in both the Amundsen Sea region and the tropical Pacific. Since the isotopic composition of ice at WDC is indicative of atmospheric circulation in the Amundsen Sea region[14] (Fig. 3a), we examine the mid-level flow in the model, represented by the 500 hPa height field. The amplitude of the simulated variability in the mean 500 hPa geopotential height field (Z_{ASL}) of the Amundsen Sea Low (ASL) region (*55-70°S, 195-240°E*) follows the same trend as the observations from WDC: more variability during the LGM, less variability during the Holocene, and a decline in variability at 16 ka (Fig. 4a). We examine the tropical Pacific influence (Z_{pac}) on the modeled Z_{ASL} by linearly removing the effect of the basin-wide (*150-270°E, 5°N-5°S*) tropical Pacific SST (SST_{pac}) from the Z_{ASL} time series (Fig. 4e,f). The amplitude of the residual ASL variability is essentially constant for the last 28 kyr (Fig. 4a), demonstrating that the change in the ASL is connected to tropical variability.

The increased Z_{ASL} variability at the LGM could be the result of either increased tropical variability or a change in efficiency of the tropical-high latitude teleconnection, forced by a change in the tropical mean state. To separate these processes, we compute the strength of the teleconnection between the tropics and ASL by constructing composite maps. These composites show the average response of the 500 hPa geopotential height field to ENSO events in the tropical Pacific (see Methods). We use a composite that imposes both an upper and lower limit on the size of SST_{pac} (± 1.25 to 2.5 times the pre-industrial standard deviation) to eliminate the possible influence of stronger or weaker ENSO events during the glacial period. Composites with or without this upper limit are indistinguishable. The results demonstrate that, in our HadCM3 experiments, the change in Z_{ASL} variability is the result of a more efficient teleconnection, rather than a change in ENSO variability (see Methods). A time series of the strength of the teleconnection over the last 28 ka (Fig. 2a, 4b) shows that the teleconnection is stronger through the LGM until 16 ka, at which time it rapidly decreases. This is in agreement with the modeled change in variability in Z_{ASL} and the observed timing of change in the WDC record. A change in the LGM teleconnection strength is also evident in eight out of eleven of the PMIP2/3 models (Extended Data Fig. 7). This is remarkable given the diversity of the response of these models' tropical climate to LGM forcing[23].

The stronger teleconnection in the glacial period in our simulations suggests a change in the tropical Pacific mean state. Such a change could result from the lowered GHG concentrations, the altered orbital forcing, or the presence of continental ice sheets. In HadCM3, we test each of these boundary conditions individually to evaluate their impact (see Methods). The GHG concentrations and orbital forcing cause a negligible change, whereas the influence of continental ice sheets is significant. Two processes associated with continental glaciation may be important for changing the teleconnection: the lowering of sea level leading to the exposure of the shelf seas in the West Pacific[23,24], and the topographic and albedo forcing of the LCIS. When considered separately, the lowered sea level causes a small change in the teleconnection that is statistically significant only to the north of the ASL region (Fig.

3e). Since flooding of West Pacific shelf seas is thought to have occurred at ~13 ka[21], this may be a possible cause of the ~13-10 ka decline in WDC variability, but cannot explain the larger step change at ~16 ka.

The topography and albedo of the LCIS force a large change to the teleconnection strength at ~16 ka, which is statistically significant across the ASL region (Fig. 3d). This timing is consistent with major deglacial changes in the LCIS, including its drawdown[18]. To separate the effects of topography and albedo, we perform two tests. First, using pre-industrial settings, we introduce only the LGM-LCIS into the model. In this case we find that almost the full LGM teleconnection strength change is realized (Fig. 3d). Second, we remove the LGM-LCIS topographic effect, while keeping the albedo effect. Here, the teleconnection pattern changes, but is not significant in the ASL region. Our results therefore demonstrate a link between the topography of the LCIS and multi-year to decadal climate variations in West Antarctica – a previously undocumented inter-polar teleconnection mechanism.

When the LCIS surface is high, the circulation in the North Pacific changes, with a persistent annual mean Aleutian Low that is deeper and farther south, a response that is consistent amongst climate models[25]. This weakens the mean winds in the West Pacific, causing the warm pool to expand eastward, as occurs during an El Niño event. The atmospheric convection moves with the warmer waters away from the Maritime Continent (Fig. 4c); hence, rainfall decreases in this region, consistent with the LGM drying inferred from lake sediments in Indonesia[21] (Fig. 2c) and stalagmites in Borneo[26] (Fig. 2d). The shifted convection then causes a change in the location of the atmospheric heating that occurs during ENSO events. Changing the location of this heating alters the structure of the extra-tropical Rossby waves that are excited during ENSO events, so that, when the LCIS is present, there are additional circulation anomalies in the Southern Hemisphere high latitudes. There are large changes in how the Rossby waves are forced from the tropics, but no change in how they propagate to the Southern Hemisphere. Since Rossby waves are the primary mechanism responsible for the tropical Pacific-West Antarctic teleconnection[4,5], the LCIS can force a change in West Antarctic climate variability. Hence, when an ENSO event occurs during the LGM – even if it is no bigger than during the modern era – it has a larger impact on West Antarctic climate than it does during the Holocene.

Our results identify a novel influence of Northern Hemisphere ice sheet topography on the climate system. By altering the coupled ocean-atmospheric circulation, the decay of the LCIS affected the strength of interactions between the tropical Pacific and high southern latitudes, reducing interannual and decadal variability in West Antarctica by nearly half. While several abrupt climate events occur during the last 31 kyr, including Dansgaard-Oeschger events, the Bølling-Allerød, the Younger Dryas, and melt water pulse 1a[24,27] (Fig. 2e,f), the interannual and decadal climate variability in West Antarctica and the patterns of rainfall in the tropical west Pacific do not appear to be affected. Instead, the initial reduction in West Antarctic variability at ~16 ka corresponds to a maximum in North Atlantic ice rafted debris layers during H1[18]. It was this ice sheet purge that likely reduced LCIS topography beyond a critical threshold, altering the inter-hemispheric climate dynamics of the Pacific basin, even as separate abrupt climate events continued to occur in the Atlantic basin and further afield.

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Supplementary Information is available in the online version of the paper.

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Author Contributions

T.R.J., W.H.G.R., and E.J.S. designed the project and led the writing of the paper. T.R.J., J.W.C.W., E.J.S., and B.R.M. contributed water isotope measurements. W.H.G.R. conducted HadCM3 simulations and led model analysis. T.R.J., K.M.C., E.J.S., and J.W.C.W. developed the diffusion-correction calculations. B.R.M. contributed change point detection algorithms and power density ratio calculations. All authors discussed the results and contributed input to the manuscript.

Author Information

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Methods

1 Water isotope data

The WDC water isotope record (Fig. 1a) was analyzed on a continuous flow analysis (CFA) system[15] using a Picarro Inc. cavity ring-down spectroscopy (CRDS) instrument, model L2130-i. The data are reported in delta notation relative to Vienna Standard Mean Ocean Water (VSMOW, $\delta^{18}O$ = δD = 0‰), normalized to Standard Light Antarctic Water (SLAP, $\delta^{18}O$ = -55.5‰, δD = -428.0‰) scale. WDC is annually dated, with accuracy better than 0.5% of the age between 0-12 ka, and better than 1% of the age between 12-31 ka[31,32].

2 Frequency domain analyses

2.1 Spectral conversion

We use the MultiTaper method fast Fourier transform technique to calculate spectral power densities[33,34] of the measured water isotope time series. Similar to other paleoclimate studies[35], we use the pmtm.m routine of P. Huybers (http://www.people.fas.harvard.edu/~phuybers/Mfiles/). Before spectral analysis, the isotope data are linearly interpolated at a uniform time interval of 0.05 yr.

2.2 High-frequency signal attenuation

High-frequency water isotope information in ice cores is attenuated by diffusion in the firn column and deep ice[16,36,37,38,39,40]. Frequency spectra reveal the amount of signal attenuation as declines in the amplitude of a given frequency through time, relative to lower frequencies (Extended Data Fig. 1a). For WDC, the annual signal (1 yr) is indiscernible at ages >14 ka. The 2 yr signal is indiscernible from 17-19 ka. Signals >3 yr are detectable throughout the last 31 kyr, while signals >4 yr are not substantially attenuated by diffusion (Extended Data Fig. 1a).

2.3 Gaussian determination of diffusion lengths

The quantitative effects of diffusion can be represented by the convolution of a Gaussian filter with the original water-isotope signal deposited at the surface and subsequently strained by ice deformation and firn compaction[36,39] (Extended Data Fig. 1b). The power density spectrum observed in the ice core record (P(f)), after diffusion, is $P(f) = P_o(f)\exp[-(2\pi f \sigma_z)^2]$, where $P_o(f)$ represents the power spectrum of the undiffused signal, f is the frequency $\frac{1}{\lambda}$, λ the signal wavelength, z the depth, and σ_z the diffusion length. Fitting a Gaussian to P(f) defines a standard deviation, σ_f , with units of 1/meters. The conversion $\sigma_z = \frac{1}{2\pi\sqrt{2}} \cdot \frac{1}{\sigma_f}$ then yields the diffusion length σ_z in units of meters[16]. The diffusion length expressed in units of time is $\sigma_t = \frac{\sigma_z}{\lambda_{avg}}$, where λ_{avg} is the mean annual layer thickness (m/yr) at a given depth (Extended Data Fig. 2c). The diffusion length quantifies the statistical vertical displacement of water molecules from their original position in the ice sheet. In the present study, we use our previously published WDC diffusion lengths, calculated for 500-yr data windows through the interval 0 – 29 ka[16].

2.4 Natural log determination of diffusion length

The σ_f variable can also be determined by using the slope of the linear regression of ln(P(f)) vs. f^2. This provides a means of estimating diffusion-length uncertainty[16]. Here, $\sigma_f = \sqrt{\frac{1}{2\cdot|(m_{ln})|}}$, where m_{ln} is the slope of the linear regression over the interval from 0.01 cycles^2/m^2 to the value at which systematic noise from the ice core analysis system overwhelms the physical signal. The point where noise dominates appears as a "kink" or "bend" in the decay of ln(P(f)). A maximum and minimum slope is fit within the standard deviation of the linear regression to determine an uncertainty range for σ_f.

2.5 Power density diffusion correction

Diffusion of a water isotope signal in an ice sheet reduces the power of high-frequencies, so P(f) takes the form of quasi-red noise. Given that WDC Holocene spectra show constant power density in the frequencies largely unperturbed by diffusion (periods >4 yrs), we use a white-noise normalization to estimate the original, pre-diffusion power density spectrum. Specifically, we calculate $P_o(f) = \text{P} \exp\,(4\pi^2 f^2 \sigma_t^2)$, where P is the observed power density (per mil^2·yr), f the frequency (1/yr), and σ_t the diffusion length (yr). We report the average power density for the diffusion-corrected 3-7 and 4-15 yr bands calculated as the integral of power density divided by the frequency range. The uncertainty on these power density diffusion corrections is determined using the uncertainty range for diffusion lengths discussed in section 2.4.

3 Change point detection

We fit linear regressions to blocks of data centered on every possible point within the 3-7 and 4-15 yr relative amplitude time series, using block sizes of 2,500 to 10,000 yrs. We calculate the *p*-value for the *F*-test to determine whether the slope of the regression of each block is statistically different than zero (Extended Data Fig. 2b). We make no *a priori* assumptions about the timing or size of significant change and take an *a posteriori* confidence level equivalent to 95%: $\alpha = 1 - 0.95^{1/n}$, where *n* is the number of statistical test realizations[31]. The first significant change occurs at 16,440 ka for the 4-15 yr diffusion-corrected data (all ages are given relative to present = 1950 C.E.). The 3-7 yr diffusion-corrected data yields identical results. The first significant change in the 4-15 yr raw data (not corrected for diffusion) occurs at 15,940 yrs.

Uncertainty in the timing of initial change includes the spectral window resolution of 500-yrs (±250 yrs) and the WDC14 age scale (±1% for ages >12 ka[32]). For initial change centered at 16.4 ka and 15.9 ka, the age scale imposes an uncertainty of ±164 yrs and ±159 yrs, respectively. Adding the above in quadrature yields initial timing uncertainties of just less than ±300 yrs. We also estimate the change point visually using a sub-set of 4-15 yr diffusion-corrected data, calculated using 100-yr non-overlapping windows between ~16.64 and 15.54 ka. The change occurs at 16.24 ka, with uncertainty on the window length of ± 50 yrs and a dating uncertainty of ±162 yrs. Added in quadrature, this gives an uncertainty value of ± 170 yrs.

4 HadCM3 model simulations

4.1 Model setup

We use the fully coupled ocean-atmosphere model HadCM3[41,42]. This model has been shown to simulate the climate in the tropical Pacific very well, including in its response to glacial forcing[23]. We simulate the climate over the last 28 ka in a series of snapshots run every 1 kyr[22]. For each snapshot, we prescribe the orbital forcing[43], greenhouse gas (GHG) concentration[44,45], and ice sheet topography and sea level[30]. We use a suite of simulations to test the sensitivity of the model to individual boundary conditions, including *Full 28-0 ka, Full 21 ka, 21 ka orbit + GHG, 21 ka ice sheets, LCIS albedo + topography* (where LCIS is the combined Laurentide-Cordilleran ice sheets)*, 21 ka West Pacific shelf exposure,* and *LCIS albedo.* Each of these simulations is defined in the Supplementary Information Data spreadsheet. All simulations are run for at least 500 years with analysis made on the final 200 years of each simulation.

4.2 Amundsen Sea Low variability

4.2.1 Long-term evolution

The variability in the WDC water isotope record on interannual and longer timescales is related to changes in the large-scale atmospheric circulation, particularly in the Amundsen Sea Basin[14] (Fig. 3a). These changes in the circulation can be characterized by the Amundsen Sea Low (ASL)[46,47]. To assess the amount of variability in the ASL, we compute an *ASL index* of monthly mean 500 hPa height averaged from 55°-70°S and 195°-240°E (Z_{ASL}). We use a 4-15 yr band-pass filter to isolate variability on the same timescale as the WDC record. Extended Data Figure 3 shows the variance of Z_{ASL} within each 1 kyr snapshot, using the *Full 28-0 ka* simulations. From 28 ka until around 15 ka, the ASL is more variable than during the Holocene (95% significant, F-test). Both the magnitude and timing of the decrease are comparable to those observed in the WDC water isotope record.

4.2.2 Internal *vs.* forced variability

To assess how much of the change in the simulated Z_{ASL} is due to the tropics, we linearly remove the average of SST across *5°N-5°S* and *150-270°E* (SST_{pac}) from the Z_{ASL}. This operation is performed by regressing Z_{ASL} onto SST_{pac}, as shown by the equation $Z_{pac} = SST_{pac} * \mathrm{regress}(SST_{pac}, Z_{ASL})$. We then subtract Z_{pac} from Z_{ASL} to obtain the component of Z_{ASL} that is unrelated to the tropical Pacific: $Z_{local} = Z_{ASL} - Z_{pac}$. The variance of Z_{local} is plotted in Extended Data Figure 3. The results show that there is no statistically significant change in the ASL variability that is unrelated to tropical Pacific SST over the deglaciation (95% significant, F-test). We compare these results, obtained with 4-15 yr band pass filtered data of monthly mean model output, to annual mean data with no filtering (Extended Data Fig. 3a,b). We obtain very similar results for both outputs. The tests described below use the annual mean data.

4.3 The tropical Pacific-to-ASL teleconnection

The tropical Pacific-to-ASL teleconnection has been well documented[48,49]. To show this teleconnection in HadCM3, we construct composite maps of 500 hPa height for years when the annual SST_{pac} anomaly exceeds ± 0.81 °C (Extended Data Fig. 4). This threshold is chosen so that SST_{pac} anomalies exceed 1.25 standard deviations in a pre-industrial simulation. We include cold ENSO events by multiplying the 500 hPa height field by (–1). Note that computing SST_{pac} over the entire tropical Pacific basin eliminates biases that may arise from changes in the pattern of SST variability, either in response to forcing or as a result of intrinsic ENSO variability.

The pre-industrial composite pattern that we obtain for the atmospheric response to tropical Pacific SST shows greatest amplitude in the Amundsen Sea region, with a positive height anomaly for warm ENSO events (Extended Data Fig. 4a). A similar pattern is obtained from reanalysis data using either regression analysis[48] or compositing[48,50]. The composite is statistically significant across the Southern Hemisphere.

Using HadCM3, we find a similar composite pattern for the pre-industrial (Extended Data Fig. 4a) and *Full 21 ka* (Extended Data Fig. 4b). The difference between the two patterns is shown in Extended Data Figure 4c; a Rossby-wave response emanating from the western tropical Pacific is evident. This response is consistent with other studies that show the tropical Pacific-to-ASL teleconnection[4,5,48,51]. Using each snapshot in the *Full 28-0 ka* simulations, we evaluate the amplitude of the tropical Pacific-to-ASL teleconnection as the mean of the composite averaged over *55-70°S* and *195-240°E* (Extended Data Fig. 5). There is a dramatic decrease in the strength of the teleconnection around 16 ka.

4.4 Teleconnection changes resulting from ENSO changes

4.4.1 Results from HadCM3 experiments

We find that the simulated variance of SST_{pac} anomalies at 21 ka (σ=0.78°C) is increased by 40% compared to the pre-industrial (σ=0.65°C). To determine the effect of SST variability on the teleconnection strength, we re-compute the composites using SST_{pac} anomalies in the range $\pm$ 0.81°C < SST_{pac} < $\pm$ 1.63°C, corresponding to SST_{pac} at 0 ka of $\pm$ 1.25σ to $\pm$ 2.5σ. By imposing both an upper and lower bound, the atmospheric response in the 500 hPa height composites is unaffected by the amplitude of simulated ENSO events. Furthermore, any change in the frequency of ENSO events is rendered unimportant by taking an average composite of events.

The composite patterns for all simulations from 28 ka to 0 ka, as well as their difference with the pre-industrial, are unchanged by imposing the upper and lower bound, while the amplitude of the composites is slightly reduced. For the *Full 21 ka simulation*, the average amplitude of the Z_{ASL} is 32 m without the upper bound, and 30 m with the upper bound imposed (Extended Data Fig. 6). Therefore, even though HadCM3 simulates an increase in the strength of ENSO during the LGM, this is not the primary cause for the change in Z_{ASL} variability. Rather, the increased Z_{ASL} variability occurs because the atmospheric response to SST_{pac} during the glacial period is stronger than during the pre-industrial.

4.4.2. Results from Paleoclimate Model Intercomparison Project

For a subset of PMIP2/3 simulations (those for which sufficient data was available[52]), we follow the same procedure as outlined above to construct composites of 500 hPa height response to SST_{pac} anomalies. We include both the lower and upper SST_{pac} bounds. The composites are computed for pre-industrial and the *Full 21 ka* simulations (Extended Data Fig. 7). A change in the tropical Pacific-to-ASL teleconnection is apparent in many of the models. Since each model has a different teleconnection in the pre-industrial, the response of the models to 21 ka boundary conditions is also different.

At 21 ka, PMIP2/3 models with enhanced SST_{pac} variance do not display a consistently different teleconnection compared to models with reduced SST_{pac} variance. This confirms that changes in the tropical Pacific-to-ASL teleconnection strength are not driven by changes in the amplitude of SST_{pac} anomalies (i.e. ENSO). Indeed, in comparison to HadCM3, the most similar change in the teleconnection occurs in FGOALS-1.0g simulations. FGOALS-1.0g shows a large decrease in ENSO strength at 21 ka, compared to the HadCM3 increase. The FGOALS-1.0g positive geopotential height anomaly is shifted eastward, consistent with the location of the regional low-pressure center in the pre-industrial simulations.

4.5 Boundary condition changes responsible for the teleconnection change

In *Full 21 ka* simulations, a number of the climate forcings are changed significantly compared to the pre-industrial. These include insolation, the greenhouse gas (GHG) concentrations, and the size of the ice sheets (Extended Data Fig. 8a). We investigate the impact of each of these boundary condition changes on the teleconnection. The *21 ka orbit + GHG* simulation does not cause a significant change (Extended Data Fig. 8b). In the *21 ka ice sheets* simulation, which includes reduced sea level, enhanced albedo in the higher latitudes, and altered topography over North America and Europe, there is a large change in the amplitude of the teleconnection (Extended Data Fig. 8c).

Prior modeling simulations have explored other aspects of the topographic effects of Northern Hemisphere ice sheets[53], including their influence on glacial climate[54], abrupt glacial climate change[55], the Atlantic Meridional Overturning Circulation[56], and Heinrich Events[57].

The *LCIS albedo + topography* simulation displays a significant change in the teleconnection strength, similar in magnitude to the *Full 21 ka* simulation (Extended Data Fig. 8e). The *LCIS albedo* simulation (i.e. albedo only) displays a statistically significant change in the teleconnection, but its effect is much smaller than that of topography and albedo combined (Extended Data Fig. 8f). Although we cannot isolate the effect of topography alone, our results suggest topography is the dominant mechanism acting to change the teleconnection strength. This agrees with the WDC shift in variability at ~16 ka, which is concurrent with the abrupt lowering of the LCIS[18].

It has been proposed that lowered sea level in the West Pacific at the LGM can have a large impact on tropical Pacific climate by exposing the continental shelves[23,58]. We test this by changing only the land-sea mask in the West Pacific in the *21 ka West Pacific shelf exposure* simulation (Extended Data Fig. 8d). This results in a change to Z_{ASL} that is 3 to 4 times smaller than that seen in *LCIS albedo + topography* simulation. There is also a statistically significant change in the composite map to the north of the ASL. In the WDC diffusion-corrected variability (3-7 and 4-15 yr bands), there appears to be a step-change that occurs at ~13 ka. This timing is consistent with the flooding of the continental shelves in the West Pacific[24,59]. Therefore, it is possible that the WDC data is detecting flooding of the continental shelves, but this effect is secondary to the abrupt change at ~16 ka.

4.6 Climatic changes associated with the teleconnection change

The tropical Pacific-to-ASL teleconnection can be considered as a Rossby Wave response to diabatic heating in the western tropical Pacific[60]. The location of this diabatic heating will depend upon the mean state. To understand how the mean state changes, we analyze how the patterns of precipitation and SST change in the 21 ka simulations relative to the pre-industrial. Extended Data Figure 9a shows the precipitation difference between the *Full 21 ka* simulation and the pre-industrial. There is a prominent decrease in precipitation over the Maritime Continent, and an increase in precipitation over the Indian Ocean and the west and central tropical Pacific. This is consistent with proxy evidence[21,23,61,]. In Extended Data Figure 9b, the 2*1 ka ice sheets* simulation accounts for the majority of the precipitation changes in the *Full 21 ka* simulation. When we decompose this effect, the *LCIS albedo + topography* simulation accounts for most of the precipitation changes in the west and central tropical Pacific (Extended Data Fig. 9c), while the *21 ka West Pacific shelf exposure* simulation accounts for most of the precipitation changes over the Maritime Continent and in the Indian Ocean (Extended Data Fig. 9d). We find that there is a much larger change in West Pacific diabatic heating (as shown by the precipitation field) in the *LCIS albedo + topography* simulation than in the *21 ka West Pacific shelf exposure* simulation. Since the ASL responds to diabatic heating anomalies in the West Pacific, rather than the Indian Ocean, this is consistent with our interpretation that the LCIS dominates the tropical Pacific-to-ASL teleconnection strength change. The timing of the change in the precipitation pattern is shown in Fig. 4c. At 16 ka, the location of the tropical precipitation moves westward from its glacial position to its pre-industrial position in the far west Pacific.

The location of diabatic heating is ultimately a response to the winds and SST. Extended Data Fig. 9 shows that the SST patterns for the *Full 21 ka, 21 ka ice sheets, and LCIS albedo + topography* simulations are all similar. Although the global mean SST is much cooler in the *Full 21 ka* simulation (Extended Data Fig. 9a), it is the spatial structure of the SST that determines the climatic response[62]. Thus it is not surprising that all three simulations alter the teleconnection in a similar way. Furthermore, the SST pattern is consistent across most of the 21 ka PMIP2/3 simulations, indicating a robust SST response to an ice sheet. The SST pattern in the *Full 21 ka* simulation is associated with cyclonic flow in the North Pacific that is forced by the presence of the LCIS. Prior modeling has shown that this wind response to the LCIS could result from a summertime weakening of the sub-tropical high and a wintertime deepening of the Aleutian Low[62].

5 Data availability

The WDC water isotope data that support the findings of this study are available at http://gcmd.gsfc.nasa.gov/search/Metadata.do?entry=NSF-ANT10-43167#metadata. Additional supporting data, including model results, are provided in the Supplementary Information Data spreadsheet and the Extended Data.

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Figure Legends

Figure 1 | WDC high-frequency signal strength

a, *The raw, high-resolution WDC δD water isotope record (grey) and a 15-yr average (red).* ***b****, Power density ratio for 15 kyr before and after the primary shift in WDC variability (i.e. 16-31 ka relative to 0-15 ka); raw data in grey and diffusion-corrected data in orange. For centennial periodicities, the mean ratio is 1.3, whereas the mean ratio for raw and diffusion-corrected periods for the 4-15 yr band is 1.9 and 2.5, respectively. The blue and green intervals show 3-7 and 4-15 yr variability, respectively. The raw data ratio is lower than one at <3 years due to increased mean diffusion in the glacial period relative to the Holocene[16]. Periods >4 years are not substantially influenced by diffusion (Extended Data Fig. 1a).* ***c, d,*** *Plots of diffusion-corrected relative amplitudes using 500-yr data windows, normalized to the most recent value, for (c) 3-7 yr variability and (d) 4-15 yr variability. Dashed lines show 1σ uncertainties (see Methods).*

Figure 2 | Indicators of oceanic and atmospheric variability

a, *Tropical Pacific-West Antarctic teleconnection strength from HadCM3 (see Fig. 3). The open red dots indicate where the teleconnection strength is 95% significantly different from the pre-industrial using a two-tailed Student's t-test.* ***b,*** *WDC diffusion-corrected 4-15 yr variability; dashed lines are 1σ uncertainties (see Methods).* ***c, d,*** *Hydrologic variability recorded in (c) an Indonesian lake sediment core[21] and (d) Bornean stalagmites[26].* ***e,*** *Relative sea level (solid line) and confidence interval (dashed lines)[28], with the estimated duration of melt water pulse (MWP) 1a[24].* ***f,*** *The NGRIP $\delta^{18}O$ record[27] from central Greenland. The dashed black line represents the ~16 ka maximum in North Atlantic ice rafted debris, corresponding to a massive freshwater discharge during H1[18]. The grey block is the estimated duration of H1[29]. The black arrow is the estimated timing of the flooding of the Sunda Shelf[21,23,24].*

Figure 3 | HadCM3 teleconnection strength

a, *Map of the correlation coefficient (R^2) for 500 hPa height and WAIS Divide δD, using the isotope enabled HadCM3 for the pre-industrial. Colors are shown for 95% statistical significance.* ***b, c, d, e*** *Composite maps of annual average 500 hPa height anomaly for ENSO events. Colors are plotted for 95% statistical significance using* *a Monte Carlo test. The green box is the Amundsen Sea region over which we calculate the teleconnection strength from the mean 500 hPa height anomaly. The purple square is the WAIS Divide ice core site. The contour interval is 5m in (b) and 2.5m in (c). Negative contours are dashed.*

Figure 4 | HadCM3 Mechanistic Attribution

a, *The variance of the ASL Indices* Z_{ASL} *(solid line) and* Z_{local} *(dashed line), computed from monthly mean data that is filtered with a 4-15 yr band pass filter. The markers are unfilled if the variance of* Z_{ASL} *at each time slice is 95% significantly different (F-test) to that of 0 ka. At 21 ka, the orange triangle shows* Z_{ASL} *for 21 ka LCIS-only, and the green triangle shows* Z_{ASL} *for 21 ka Shelf Exposure (i.e. the effect of sea level change in the Maritime Continent region).* ***b,*** *The modeled teleconnection strength between the tropical Pacific and West Antarctica (solid red line; open red dots indicate statistically significant differences from pre-industrial).* ***c,*** *The difference in mean annual rainfall between the Central Pacific and the Maritime continent (average of 20°N-20°S, 145°E-190°E minus average of 20°N-20°S, 100°E-145°E), shown as an anomaly relative to the pre-industrial (light blue line). Also shown are the precipitation anomalies over the Maritime Continent only (dashed grey line).* ***d,*** *Northern Hemisphere ice area*[30] *(dashed green line) and average height of the LCIS*[30] *(blue line). In all panels, the dashed vertical line shows 16 ka, as in Figure 2.* ***e,*** *Map of the change in the 500 hPa height field variance between 21 ka and the pre-industrial.* ***f,*** *As in (e), except for the* Z_{pac} *variability that is linearly related to ENSO (i.e. the part removed from* Z_{ASL} *to yield* Z_{local}*). The green box is the Amundsen Sea region over which we calculate the ASL Indices in (a). The purple square is the WAIS Divide ice core site. The contour interval is 40* m^2*, with colors changing every 80* m^2*. Negative contours are dashed. Colors are plotted for 95% statistical significance using a Monte Carlo test.*

Extended Data Figure Legends

Extended Data Figure 1 | Signal Detection

a, *Relative amplitudes for 1-yr, 2-yr, 3-yr, and 4-yr periods calculated for 500-yr spectral data windows, normalized to the value for the annual signal in the most recent data window. Both the climate and diffusion affects the amplitude of these high-frequency signals. However, it is mainly the effects of diffusion that cause the loss of the annual signal at >14 ka. Similarly, the 2-yr period is lost between ~17-19 ka, when the rate of diffusion for WDC was highest[16]. Periods >3 yrs survive diffusion throughout the last 31 kyr.* ***b,*** *An example of a deconvolution calculation showing the observed δD water isotope record (i.e raw data; dotted red line), and the diffusion-corrected record (black line). While this calculation was not used for the data presented in this paper, it serves as a visual aide in understanding how the diffusion-corrected water isotope record would look in the time-domain. We performed all diffusion-correction calculations in the frequency domain to reduce uncertainty.* ***c,*** *The power density spectrum for the SOI Index from 1951-2017 (black; 95% confidence intervals in grey), compared with a red noise null hypothesis (red) calculated from the average of 100 power spectrums of synthetic data that have the same autocorrelation and variance as the SOI. The SOI has power greater than the red noise across a broad spectral peak between ~2-17 yrs, which can be subdivided into a ~2-7 yr high frequency peak and an ~8-17 yr peak. Due to the limited temporal span of modern observations, multi-decadal spectral estimates of the SOI cannot be adequately defined.* ***d,*** *Diffusion-corrected relative amplitudes using 500-yr windows of WDC water isotope data.* ***e,*** *The difference in age of consecutive 5 mm WDC water isotope samples (Δage_{5mm}; blue) and a 500-yr sliding average (red).* ***f,*** *The WDC accumulation rate[63], inverted. The accumulation, and by extension the Δage_{5mm} in (e), undergoes large changes during the deglaciation at ~18.5 ka, occurring 2.5 millennia before the change point in teleconnection strength at ~16 ka.*

Extended Data Figure 2 | Change Point Detection

a, *WDC 4-15 yr variability for 500-yr data windows (dashed lines are 1σ uncertainties; see Methods).* ***b,*** *Regression test algorithm to determine the first significant change in the WDC 4-15 yr variability in (a). The first colored data point below the p-value significance threshold occurs at 16.44 ka.* ***c,*** *Example of the diffusion-correction calculation for a 100-yr data window centered on 15.54 ka bp; raw data (black), diffusion corrected data (blue), Gaussian fit* *(red) with dashed 1σ uncertainty bounds (see Methods). The same calculation is made for 500-yr data windows.* ***d,*** *The sub-set of diffusion-corrected 4-15 yr amplitudes (green) calculated at 100-yr resolution. The values are normalized to the amplitude value at 16.24 ka, which represents the change point towards smaller amplitudes.*

Extended Data Figure 3 | ASL variability

Panels (a) and (b) show the variance of the ASL Indices Z_{ASL} (black) and Z_{local} (blue). ***a,*** *The ASL Indices are computed from monthly mean data then filtered with a 4-15 yr band pass filter.* ***b,*** *The ASL Indices are computed from annual mean output. We compute Z_{ASL} as the mean 500 hPa height in the region 55°-70°S and 195°-240°E. The blue lines show Z_{ASL} after linearly removing the SST_{pac} time series from the 500 hPa height field; this is the ASL variability unrelated to the tropical Pacific (Z_{local}; see Methods). The markers are unfilled if the variance of Z_{ASL} at each time slice is 95% significantly different (F-test) to that of the pre-industrial.* ***c,*** *Map of the change in the variance of the 500 hPa height field between 21 ka and the pre-industrial.* ***d,*** *The variability that is linearly related to ENSO, Z_{pac} (this is the part removed from Z_{ASL} to yield Z_{local}).* ***e,*** *The variability with the effect of ENSO linearly removed (see Methods; this the equivalent of Z_{local}). Changes not attributable to ENSO occur to the north of the Amundsen Sea, while changes over the Amundsen Sea are related to ENSO. In (c), (d), and (e) the contour intervals are 40 m^2, with colors changing every 80 m^2. Negative contours are dashed.*

Extended Data Figure 4 | Composites of the 500 mb height field for ENSO events

Composites for ***a,*** *the pre-industrial,* ***b,*** *Full 21ka, and* ***c,*** *the difference between Full 21ka and pre-industrial. Colors are plotted for 95% statistical significance using a Monte Carlo test. Negative contours are dashed. In (a) and (b) the contours are plotted every 5 m and colors saturate at ±25 m. The thin blue box shows where the ASL index is computed. In (c) contours are plotted every 2.5 m and colors saturate at ±12.5 m.*

Extended Data Figure 5 **| Tropical Pacific-to-ASL teleconnection strength**

Computed using the Full 28-0 ka simulations. ***a,*** *Average 500 hPa geopotential height anomaly in the Amundsen Sea region, computed within the blue box shown in Extended Data Figure 4. These values are derived from composites constructed using only the lower limits on the size of ENSO events (see Main Text and Methods).* ***b,*** *The t-score is shown by the red line. Values outside the shaded red region are 95% statistically different from the pre-industrial.*

Extended Data Figure 6 | Composites of the 500mb height field for ENSO events with and without upper bound

a, *The teleconnection without upper limit using the Full 21 ka simulation.* ***b,*** *The teleconnection with upper limit using the Full 21 ka simulation.* ***c,*** *The difference in teleconnection between the Full 21ka simulation and the pre-industrial without upper limit.* ***d,*** *The same difference as (c) with upper limit. In all panels, negative contours are dashed, and colors are plotted for 95% statistical significance using a Monte Carlo test. In (a) and (b), the contours are plotted every 5 m and colors saturate at ±25 m. In (c) and (d), contours are plotted every 2.5 m and colors saturate at ±12.5 m.*

Extended Data Figure 7 | Composite maps of annual average 500mb height inPMIP2/3 models

For each model the top panel shows the 0 ka composite and the bottom panel the difference between the Full 21 ka and 0 ka simulation. Colors are plotted for 95% statistical significance using a Monte Carlo test. In the upper panels, contours are plotted at 5m intervals, with colors saturating at 25m. In the lower panels contours are plotted at 2.5m intervals, with colors saturating at ±12.5 m. Negative contours are dashed. The reduced statistical significance in these panels compared to those shown in Extended Data Figures 4, 6, 8, 9, and 10 is due to the shorter data series available in the PMIP2/3 archives.

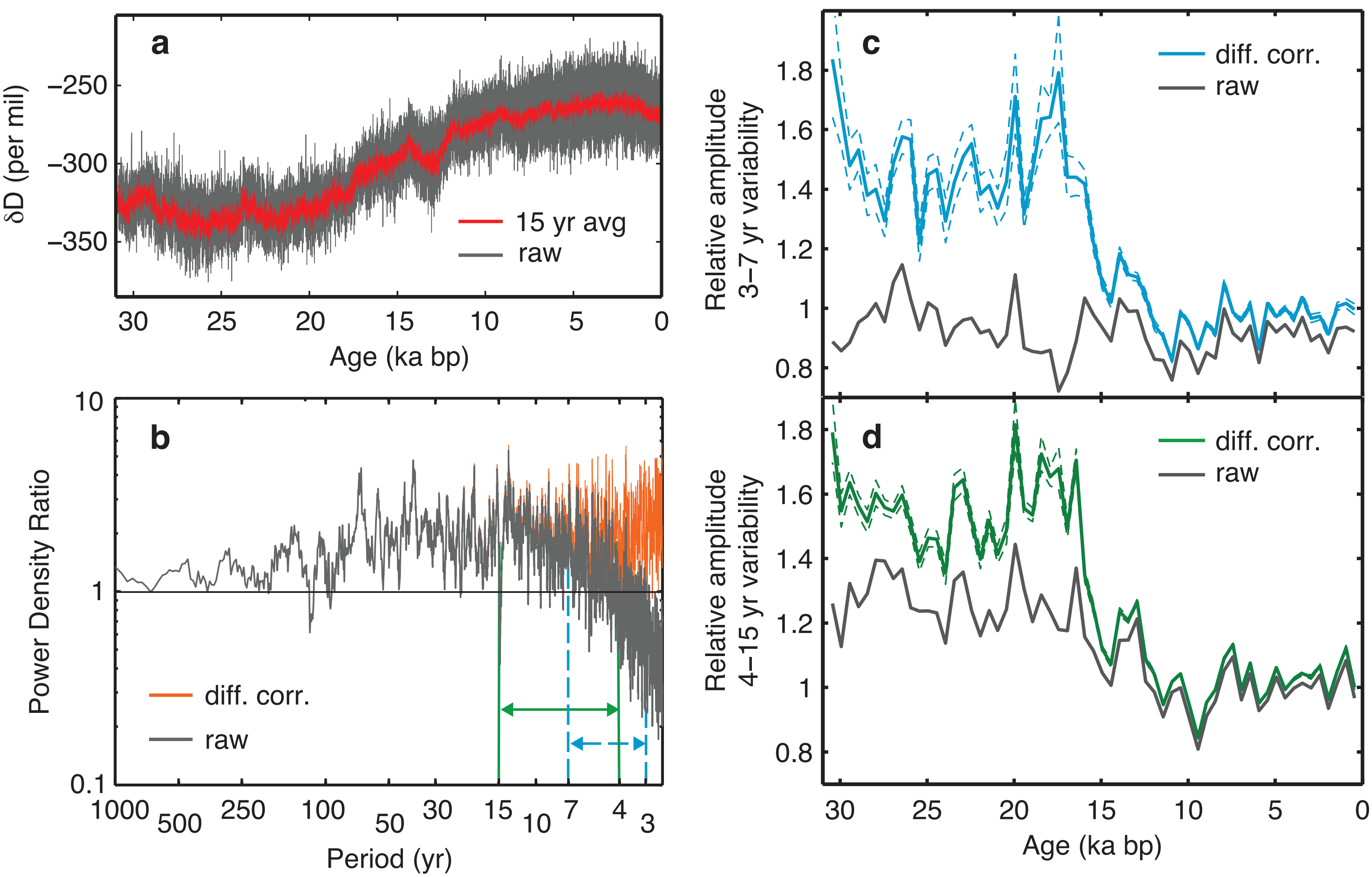
Extended Data Figure 8 | Sensitivity of composite maps to different sets of 21ka boundary conditions

All plots show the difference between 21ka and 0ka composites. The composites are constructed using both upper and lower limits on the size of ENSO events. Contours are plotted every 2.5 m (negative contours are dashed) and colors saturate at ±12.5 m. Colors are plotted for 95% statistical significance using a Monte Carlo test. ***a,*** *Full 21 ka simulation.* ***b,*** *21 ka orbit + GHG.* ***c,*** *21 ka ice sheets.* ***d,*** *21 ka West Pacific shelf exposure.* ***e,*** *21 ka LCIS albedo + topography, (where LCIS is the combined Laurentide-Cordilleran ice sheets).* ***f,*** *21 ka LCIS albedo. Each of these simulations is fully defined in the Supplementary Information Data spreadsheet.*

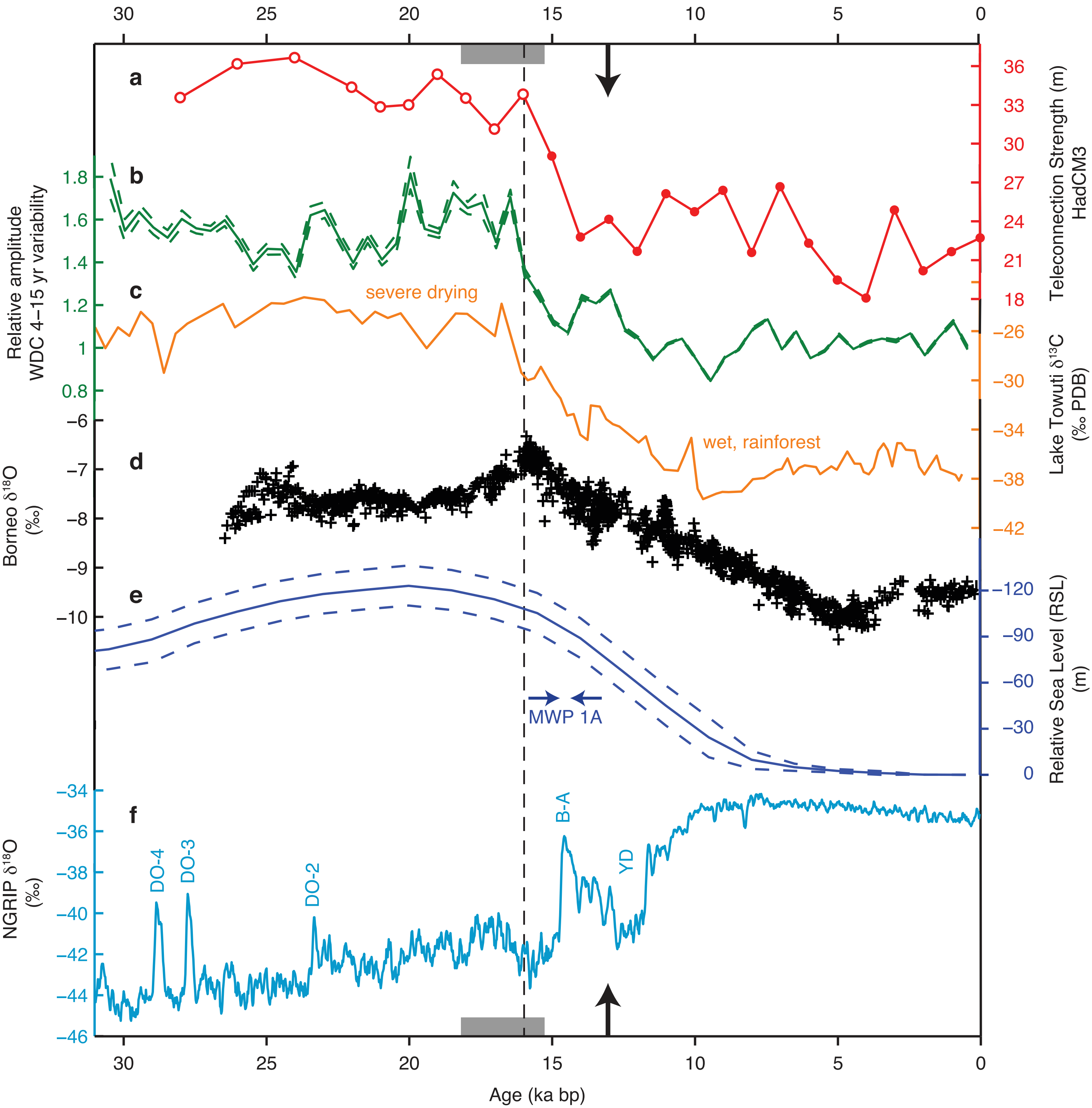
Extended Data Figure 9 | Annual mean anomalies of precipitation and sea surface temperature

Maps of 21 ka anomalies relative to the pre-industrial. ***a,*** *Full 21 ka simulation.* ***b,*** *21 ka ice sheets.* ***c,*** *21 ka LCIS albedo + topography.* ***d,*** *21 ka West Pacific shelf exposure. Annual means are calculated from 100 years of output.*

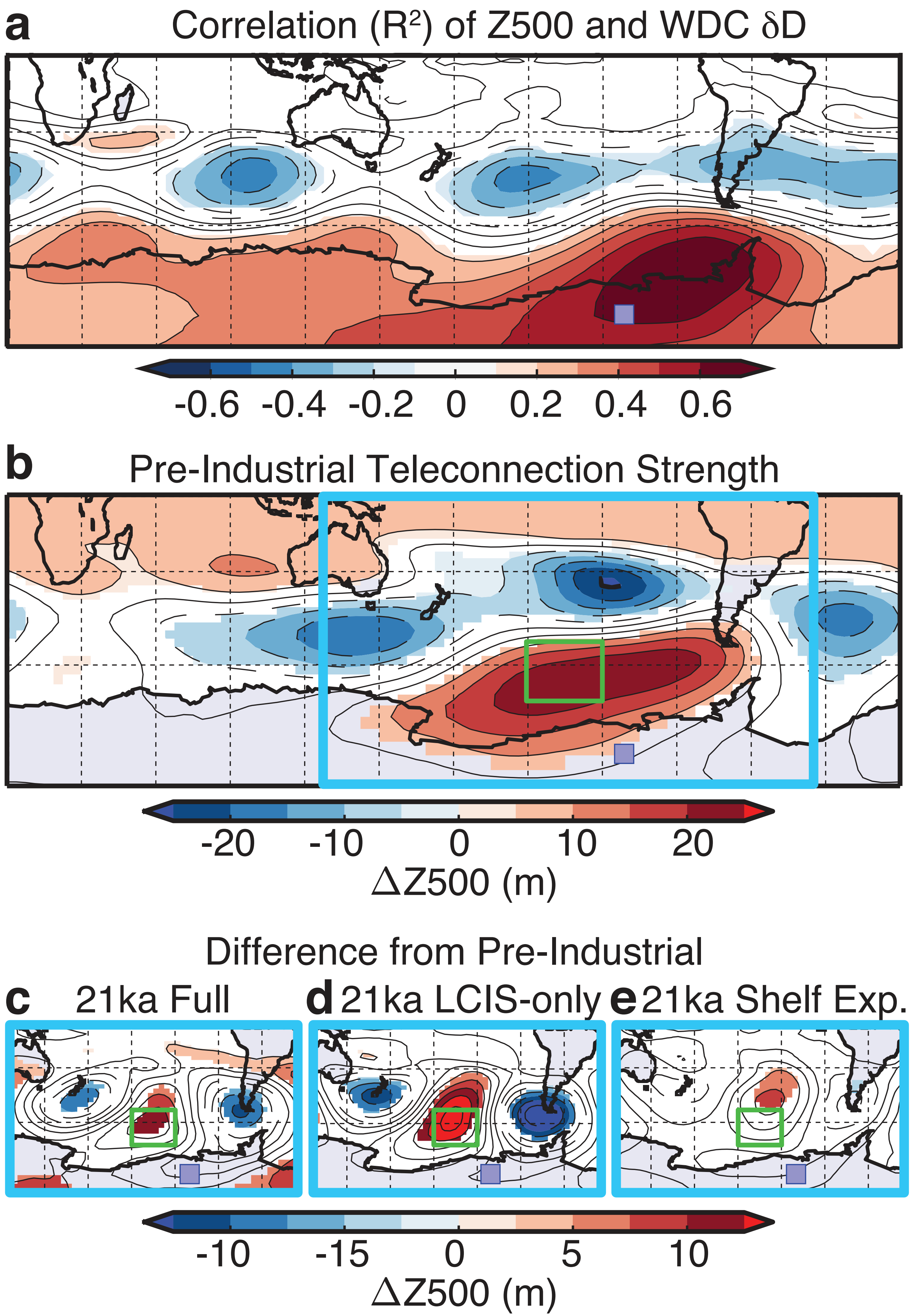
Contour intervals for precipitation are 1 mm day^{-1}, and for sea surface temperature are 0.5°C. Land areas are shown in in grey. Note that the temperature color scale in (a) ranges from -4 to 0 °C. This accounts for the mean GHG cooling that is seen in the Full 21 ka simulation.



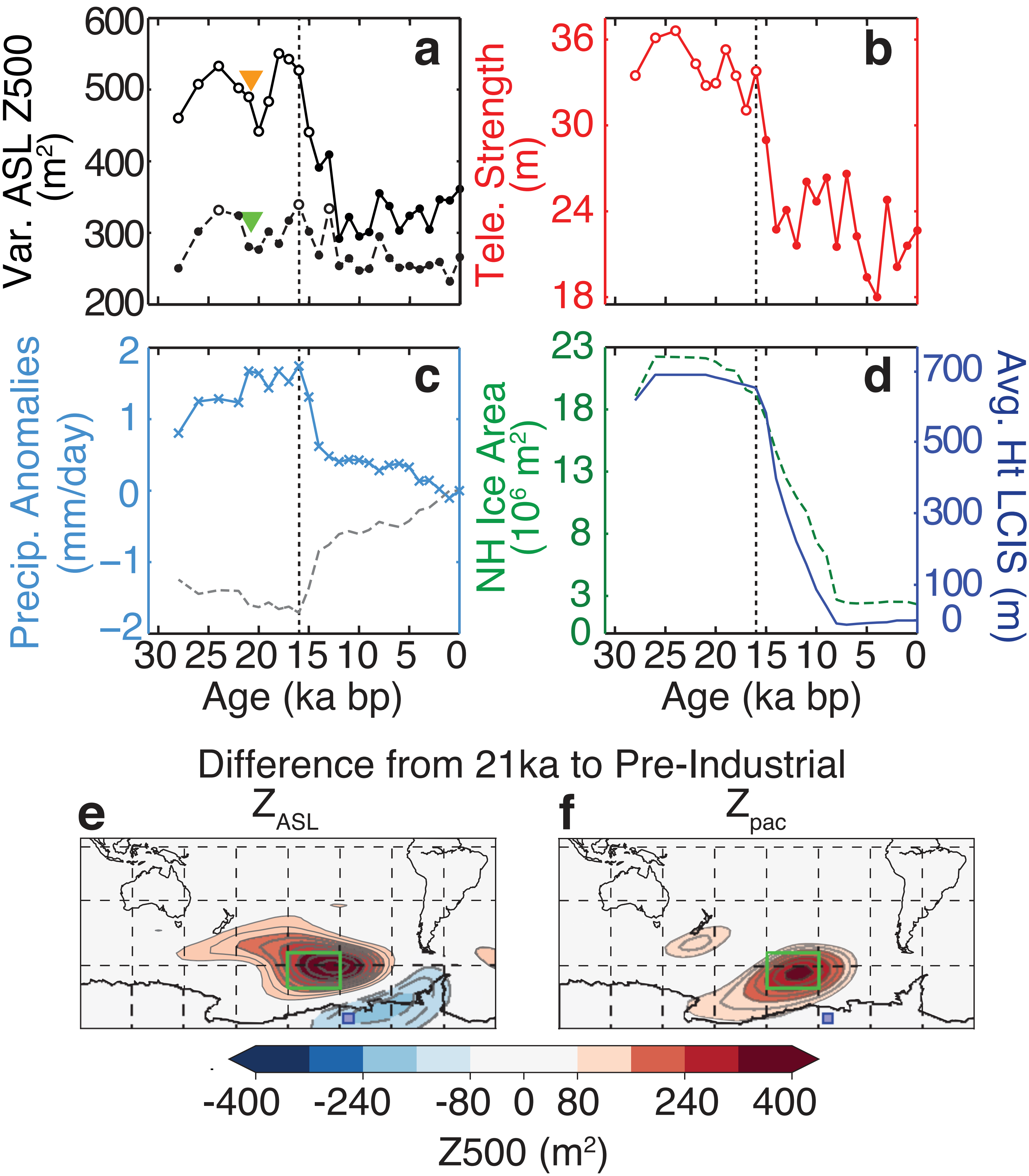
a
δD (per mil)
−250
−300
−350
15 yr avg
raw
30 25 20 15 10 5 0
Age (ka bp)
b
Power Density Ratio
10
1
0.1
diff. corr.
raw
1000 500 250 100 50 30 15 10 7 4 3
Period (yr)
c
Relative amplitude 3–7 yr variability
diff. corr.
raw
1.8 1.6 1.4 1.2 1 0.8
d
Relative amplitude 4–15 yr variability
diff. corr.
raw
1.8 1.6 1.4 1.2 1 0.8
30 25 20 15 10 5 0
Age (ka bp)



a
b
c
d
e
f
severe drying
wet, rainforest
MWP 1A
DO-4
DO-3
DO-2
B-A
YD
Relative amplitude
WDC 4–15 yr variability
Borneo δ18O (‰)
NGRIP δ18O (‰)
Teleconnection Strength (m) HadCM3
Lake Towuti δ13C (‰ PDB)
Relative Sea Level (RSL) (m)
Age (ka bp)



a
Correlation (R^2) of Z500 and WDC δD
-0.6
-0.4
-0.2
0
0.2
0.4
0.6
b
Pre-Industrial Teleconnection Strength
-20
-10
0
10
20
ΔZ500 (m)
Difference from Pre-Industrial
c
21ka Full
d
21ka LCIS-only
e
21ka Shelf Exp.
-10
-15
0
5
10
ΔZ500 (m)



Var. ASL Z500 (m^2)
a
Tele. Strength (m)
b
Precip. Anomalies (mm/day)
c
NH Ice Area (10^6 m^2)
d
Avg. Ht LCIS (m)
Age (ka bp)
Age (ka bp)
Difference from 21ka to Pre-Industrial
e
Z_{ASL}
f
Z_{pac}
-400
-240
-80
0
80
240
400
Z500 (m^2)