

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 fundamentally altered global ocean–atmosphere climate dynamics¹. Model simulations and palaeoclimate records suggest that glacial boundary conditions affected the El Niño–Southern Oscillation^{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 a 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 Last Glacial Maximum as during the ensuing Holocene epoch (the past 11,700 years). 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 approximately 16,000 years and 15,000 years ago, 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 sheets¹. The state changes are observed in palaeoclimate 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 affects the generation and communication of climate anomalies between the low and high latitudes^{4,5}.

Whether the nature of the El Niño–Southern Oscillation (ENSO) changed between the Last Glacial Maximum (LGM) and the Holocene is an important question^{3,6}. Proxy studies of ENSO variability during the LGM are contradictory^{7,8}, and the results of modelling studies are similarly ambiguous². One obstacle to understanding past ENSO variability is disentangling changes in ENSO itself from changes in its influence on climate outside the tropics, that is, the strength of ENSO-related teleconnections, which can vary over time⁹.

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 sheet¹² and for the ventilation of carbon dioxide from the Southern Ocean¹³.

The influence of tropical variability on the climate of West Antarctica is reflected in ice core records from the WAIS. In particular, the oxygen

and hydrogen isotope ratios of water are sensitive to temperature and atmospheric circulation anomalies associated with large ENSO events¹⁴. 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 use a high-resolution hydrogen isotope (δD) record from the WAIS Divide ice core (WDC) to investigate about 31,000 years (31 kyr) of high-frequency climatic variability. The data were produced using laser absorption spectroscopy coupled with a continuous flow analysis system¹⁵. 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 (Extended Data Fig. 1e).

We use spectral techniques (see Methods) on the WDC δD record to analyse West Antarctic climate variability (Fig. 1). Periods longer than 3 yr 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 ice¹⁶, we apply a diffusion correction on consecutive 500-yr data windows (see Methods; Extended Data Figs 1b, 2c). We isolate the amplitude of 3–7-yr 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 (Extended Data Fig. 1c), while the 4–15-yr range is less affected by diffusion (Extended Data Fig. 1a) and captures the decadal variability observed in the Southern Oscillation Index (Extended Data Fig. 1c), tropical Pacific coral isotope records¹⁷, and in modern observations in West Antarctica^{5,14}.

We find that the mean amplitudes of the diffusion-corrected variability in the 3–7-yr and 4–15-yr bands are elevated in the Late Glacial (16–31 kyr ago) period relative to the Holocene (0–10 kyr ago) (Fig. 1c, d). An approximately 50% decrease in amplitude occurs from about 16 kyr to 10 kyr ago in both time series. This decline occurs in two steps: an abrupt decrease about 16 kyr ago followed by a second but less pronounced decline during the period 13–10 kyr ago. Neither time series reveal other substantial changes in variability during the Holocene or during the Late Glacial. We use an objective algorithm (see Methods) to detect an initial statistically significant ($P < 0.05$) 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 16.44 ± 0.30 kyr ago (all ages are given relative to the present, 1950 AD). 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 subset of 100-yr data windows for the diffusion-corrected 4–15-yr time series (Extended Data Fig. 2c, d), the decline occurs 16.24 ± 0.17 kyr ago. The timing of the decline is not an artefact of the diffusion correction. Using the raw data in the 4–15-yr band for 500-yr data windows (Fig. 1d), an approximately 40% decrease in amplitude occurs from about 16 kyr to 10 kyr ago, with an initial decline occurring

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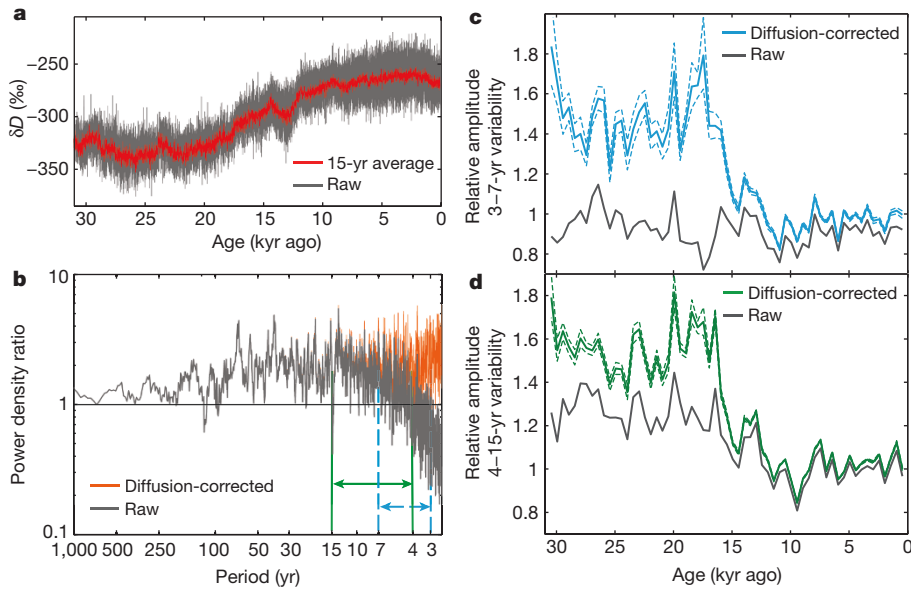


Figure 1 | WDC high-frequency signal strength. **a**, The raw, high-resolution WDC δD water isotope record (grey) and a 15-yr average (red). **b**, Power density ratio for 15 kyr before and after the primary shift in WDC variability (that is, 16–31 kyr ago relative to 0–15 kyr ago); raw data in grey 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-yr and 4–15-yr variability, respectively. The raw data ratio is lower than one at <3 yr owing to increased mean diffusion in the glacial period relative to the Holocene¹⁶. Periods >4 yr 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 3–7-yr variability (**c**) and 4–15-yr variability (**d**). Dashed lines show 1σ uncertainties (see Methods).

15.94 ± 0.30 kyr ago. Conservatively, this places the decline in WDC variability between 16.74 kyr to 15.64 kyr ago, with a central estimate of 16.24 kyr ago.

The climate shift about 16 kyr ago coincides with large-scale geologic events observed in proxy records across the globe. A substantial

lowering of the Laurentide Ice Sheet surface is estimated to have occurred 16 kyr ago (from ^{14}C dating)¹⁸, corresponding to the Heinrich Event 1 iceberg discharge event in the Northern Hemisphere (Fig. 2). The Heinrich Event 1 ice-rafted debris is of Laurentide origin¹⁹, indicating abrupt changes in ice mass and the topography of the Laurentide Ice

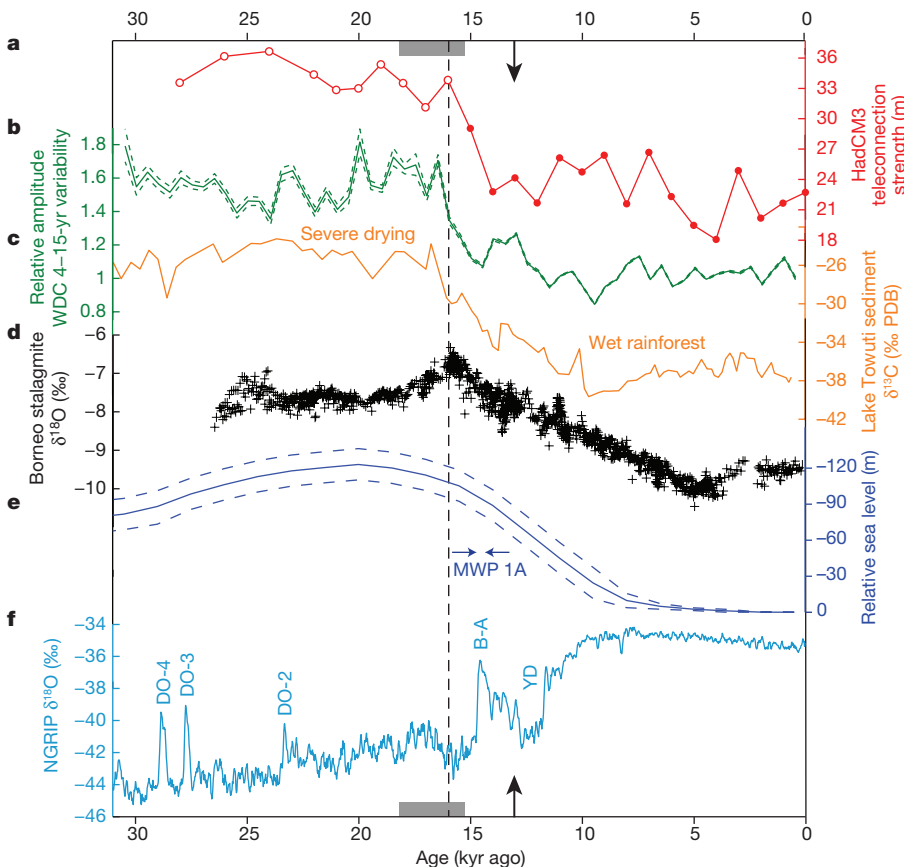


Figure 2 | Indicators of oceanic and atmospheric variability. **a**, Tropical Pacific–West Antarctic teleconnection strength from HadCM3 (see Fig. 3). The open red dots indicate when the teleconnection strength is 95% significantly different from the pre-industrial period using a two-tailed Student’s *t*-test. **b**, WDC diffusion-corrected 4–15-yr variability; dashed lines are 1σ uncertainties (see Methods). **c**, **d**, Hydrologic variability recorded in an Indonesian lake sediment core²¹ (**c**) and Bornean stalagmites²⁶ (**d**). **e**, Relative sea level (solid line) and confidence interval (dashed lines)²⁸, with the estimated duration of melt water pulse (MWP) 1A²⁴. **f**, The NGRIP ice core $\delta^{18}O$ record²⁷ from central Greenland, showing the Younger Dryas (YD), Bølling–Allerød (BA), and Dansgaard–Oeschger (DO) events 2–4. The dashed black line represents the maximum about 16 kyr ago in North Atlantic ice-rafted debris, corresponding to a massive freshwater discharge during Heinrich Event 1¹⁸. The grey block is the estimated duration of Heinrich Event 1²⁹. The black arrow is the estimated timing of the flooding of the Sunda Shelf^{21,23,24}.

Sheet. The Cordilleran ice sheet also declined rapidly at this time²⁰, and the lowering 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 Indonesia²¹ (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 kyr to 0 kyr ago²² (see Methods). We assess variability in both the Amundsen Sea region and the tropical Pacific Ocean. Since the isotopic composition of ice at WDC is indicative of atmospheric circulation in the Amundsen Sea region¹⁴ (Fig. 3a), we examine the mid-level flow in the model, represented by the 500-hPa geopotential 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 16 kyr ago (Fig. 4a). We examine the tropical Pacific influence (Z_{pac}) on the modelled 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 (Z_{local}) is essentially constant over the past 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 teleconnection between the tropics and the high latitudes, 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 simulated teleconnection over the past 28 kyr (Figs 2a, 4b) shows that the teleconnection is stronger through the LGM until 16 kyr ago, at which time it rapidly decreases. This is in agreement with the modelled 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 models of the Paleoclimate Model Intercomparison Project (PMIP2/3) (Extended Data Fig. 7). This is remarkable given the diversity of the response of the tropical climate of these models to LGM forcing²³.

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 greenhouse gas 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 greenhouse gas concentrations and orbital forcing cause a negligible change, whereas the influence of continental ice sheets is statistically significant ($P < 0.05$). 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 ($P < 0.05$) only to the north of the ASL region (Fig. 3e). Since flooding of West Pacific shelf seas is thought to have occurred about 13 kyr ago²¹, this may be a possible cause of the decline in WDC variability between about 13–10 kyr ago, but cannot explain the larger step change about 16 kyr ago.

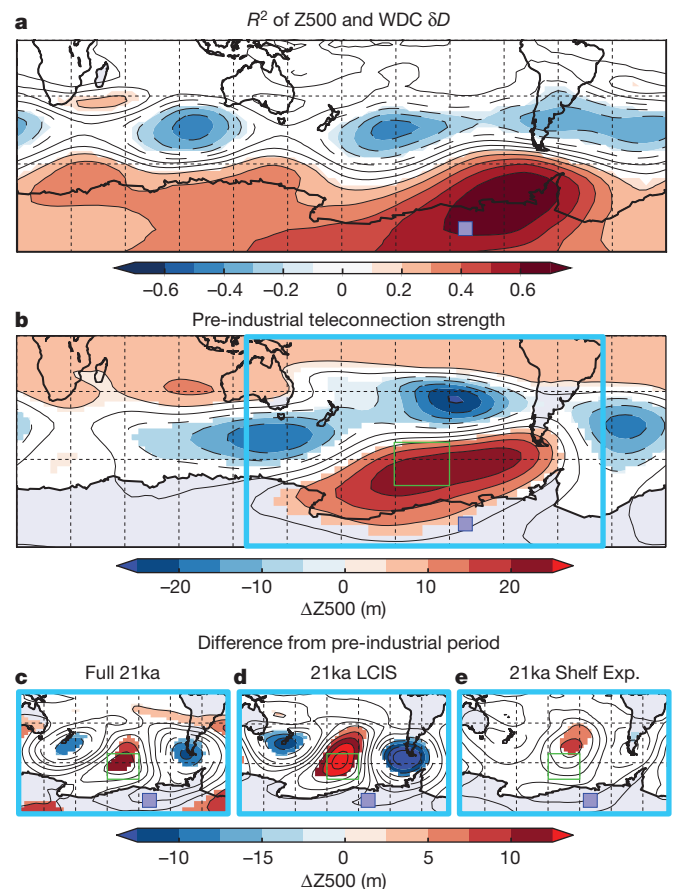


Figure 3 | HadCM3 teleconnection strength. **a**, Map of the correlation coefficient R^2 for the 500-hPa geopotential height ($Z500$) and WAIS Divide δD , using the isotope-enabled HadCM3 model for the pre-industrial period. Contours are filled (not white) when the statistical significance exceeds 95%. **b–e**, Composite maps of the annual average 500-hPa height anomaly ($\Delta Z500$) for ENSO events, including the pre-industrial (**b**), and the differences between three different model experiments and the pre-industrial (**c–e**). Contours are filled (non-white) when statistical significance exceeds 95% 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 blue box in **b** indicates the area shown in **c–e**. The contour interval is 5 m in **b** and 2.5 m in **c**. Negative contours are dashed.

The topography and albedo of the LCIS force a large change to the teleconnection strength about 16 kyr ago, which is statistically significant ($P < 0.05$) across the ASL region (Fig. 3d). This timing is consistent with major deglacial changes to the LCIS¹⁸. To separate the effects of topography and albedo, we perform two tests. First, using pre-industrial settings, we introduce the LGM–LCIS albedo and topographic effects 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 statistically 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 models²⁵. This weakens the mean winds in the West Pacific, causing the warm pool to expand eastward, as occurs during an El Niño event.

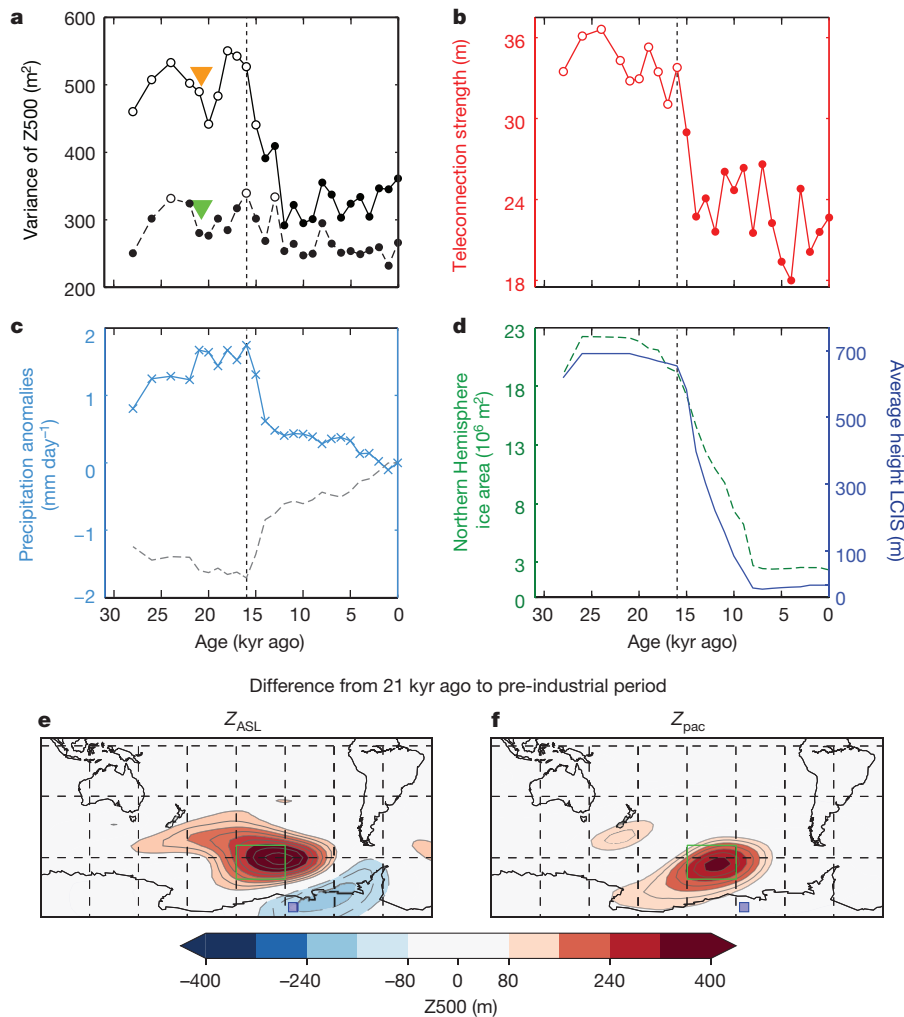


Figure 4 | HadCM3 mechanistic attribution. **a**, The variance of the height of the 500-hPa pressure surface (Z_{500}) for the Z_{ASL} (solid line) and Z_{local} (dashed line) indices, computed from monthly mean data that is filtered with a 4–15-yr band pass filter. The markers are unfilled if the variance of Z_{ASL} at each time slice is 95% significantly different (F -test) to that of 0 kyr ago. At 21 kyr ago, the orange triangle shows Z_{ASL} for the ‘21ka LCIS only’ scenario, and the green triangle shows Z_{ASL} for the ‘21ka Shelf Exp.’ scenario (that is, the effect of sea level change in the Maritime Continent region). Each of these scenarios is defined in Supplementary Data. **b**, The modelled teleconnection strength between the tropical Pacific and West Antarctica (solid red line; open red dots indicate statistically significant differences from pre-industrial). **c**, The simulated difference in mean annual rainfall between the central Pacific and the Maritime Continent (average of 20° N–20° S, 145°–190° E minus

The atmospheric convection moves with the warmer waters away from the Maritime Continent (the part of Southeast Asia comprising Indonesia, the Philippines, Papua New Guinea and nearby countries) (Fig. 4c); hence, rainfall decreases in this region, consistent with the LGM drying inferred from lake sediments in Indonesia²¹ (Fig. 2c) and stalagmites in Borneo²⁶ (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. Although there are large changes in how the Rossby waves are forced from the tropics, there is no change in how they propagate within 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

average of 20° N–20° S, 100°–145° E), shown as an anomaly relative to the pre-industrial period (light blue line). Also shown are the precipitation anomalies over the Maritime Continent only (dashed grey line). **d**, Northern Hemisphere ice area³⁰ (dashed green line) and average height of the LCIS³⁰ (blue line). In all panels, the dashed vertical line shows 16 kyr ago, as in Fig. 2. **e**, Map of the change in the 500-hPa height field variance between 21 kyr ago and the pre-industrial period. **f**, As in **e**, except for the Z_{pac} variability that is linearly related to ENSO (that is, the part removed from Z_{ASL} to yield Z_{local}). The green box is the Amundsen Sea region over which we calculate the indices in **a**. The purple square is the WAIS Divide ice core site. The contour interval is 40 m², with colours changing every 80 m². Negative contours are dashed. Colours are plotted for 95% statistical significance using a Monte Carlo test.

event occurred during the LGM—even if it were no bigger than during the modern era—it had a larger impact on West Antarctic climate than during the modern era.

Our results identify an influence of Northern Hemisphere ice-sheet topography on the climate system. By altering the coupled ocean–atmosphere 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. Although several abrupt climate events occur during the past 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 about 16 kyr ago corresponds to a maximum in North Atlantic ice-rafted debris layers during Heinrich Event 1¹⁸. It was this ice-sheet purge

that probably reduced LCIS topography beyond a critical threshold, altering the interhemispheric climate dynamics of the Pacific basin, even as separate abrupt climate events continued to occur in the Atlantic basin and further afield.

Online Content Methods, along with any additional Extended Data display items and Source Data, are available in the online version of the paper; references unique to these sections appear only in the online paper.

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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.

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METHODS

Water isotope data. The WDC water isotope record (Fig. 1a) was analysed on a continuous flow analysis system¹⁵ using a cavity ring-down spectroscopy instrument (Picarro Inc. model L2130-i). The data are reported in delta notation relative to Vienna Standard Mean Ocean Water (VSMOW, $\delta^{18}\text{O} = \delta D = 0\text{‰}$), normalized to the Standard Light Antarctic Precipitation ($\delta^{18}\text{O} = -55.5\text{‰}$, $\delta D = -428.0\text{‰}$) scale. WDC is annually dated, with accuracy better than 0.5% of the age in the range 0–12 kyr ago, and better than 1% of the age in the range 12–31 kyr ago^{31,32}.

Frequency domain analyses. 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. As in other palaeoclimate studies³⁵, 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.

High-frequency signal attenuation. High-frequency water isotope information in ice cores is attenuated by diffusion in the firn column and deep ice^{16,36–40}. 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 kyr ago. The 2-yr signal is indistinguishable from 17–19 kyr ago. Signals >3 yr are detectable throughout the past 31 kyr, while signals >4 yr are not substantially attenuated by diffusion (Extended Data Fig. 1a).

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 $P(f) = P_0(f) \exp[-(2\pi f \sigma_z)^2]$, where $P_0(f)$ represents the power spectrum of the undiffused signal, $f = 1/\lambda$ is the frequency, λ is the signal wavelength, z is the depth, and σ_z is the diffusion length. Fitting a Gaussian to $P(f)$ defines a standard deviation σ_f with units of per metre. The conversion $\sigma_z = 1/2\pi\sqrt{2} \times 1/\sigma_f$ then yields the diffusion length σ_z in units of metres¹⁶. The diffusion length expressed in units of time is $\sigma_t = \sigma_z/\lambda_{\text{avg}}$, where λ_{avg} is the mean annual layer thickness (in metres per year) at a given depth (Extended Data Figs 1f, 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 throughout the interval 0–29 kyr ago¹⁶.

Natural logarithm determination of diffusion length. The variable σ_f can also be determined by using the slope of the linear regression of $\ln[P(f)]$ versus f^2 . This provides a means of estimating diffusion-length uncertainty¹⁶. Here, $\sigma_f = \sqrt{1/[2|(m_{\ln})|]}$, where m_{\ln} is the slope of the linear regression over the interval from 0.01 (cycles per metre)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 fitted within the standard deviation of the linear regression to determine an uncertainty range for σ_f .

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 yr), we use a white-noise normalization to estimate the original, pre-diffusion power density spectrum. Specifically, we calculate $P_0(f) = P \exp(4\pi^2 f^2 \sigma_z^2)$, where P is the observed power density (in units of $\text{‰}^2 \text{yr}$), f the frequency (in units of per year), and σ_z is the diffusion length (in units of years). We report the average power density for the diffusion-corrected 3–7-yr, 4–15-yr, and 3–30-yr bands (Fig. 1c, d, Extended Data Fig. 1d) 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 the section ‘Natural logarithm determination of diffusion length’.

Change point detection. We fit linear regressions to windows of data within the 3–7-yr and 4–15-yr relative amplitude time series, using window sizes of 2,500 yr to 10,000 yr. We calculate the P value for the F -test to determine whether the slope of the regression in each window is statistically different from zero (Extended Data Fig. 2b). We make no a priori assumptions about the timing or size of statistically 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 realizations³¹. The first statistically significant change occurs 16,440 yr ago for the 4–15-yr diffusion-corrected data (all ages are given relative to the present, 1950 AD). The 3–7-yr diffusion-corrected data yields identical results. The first statistically significant change in the 4–15-yr raw data (not corrected for diffusion) occurs 15,940 yr ago.

Uncertainty in the timing of initial change includes the spectral window resolution of 500 ± 250 yr and the WDC14 age scale ($\pm 1\%$ for ages >12 kyr ago³²). For the initial change centred at 16.4 kyr ago and 15.9 kyr ago, the age scale imposes an uncertainty of ± 164 yr and ± 159 yr, respectively. Adding the above in quadrature yields initial timing uncertainties of just less than ± 300 yr. We also estimate the change point visually using a subset of 4–15-yr diffusion-corrected data, calculated using 100-yr non-overlapping windows between 16.64 kyr ago and 15.54 kyr ago. The initial decline occurs 16.24 kyr ago, with uncertainty on the window length of ± 50 yr and a dating uncertainty of ± 162 yr. Added in quadrature, this gives an uncertainty value of ± 170 yr.

HadCM3 model simulations. 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 forcing²³. We simulate the climate over the past 28 kyr in a series of snapshots run every thousand years²². For each snapshot, we prescribe the orbital forcing⁴³, greenhouse gas concentration^{44,45}, and ice-sheet topography and sea level³⁰. We use a suite of simulations to test the sensitivity of the model to individual boundary conditions, including ‘Full 28–0ka’, ‘Full 21ka’, ‘21ka Orbit + GHG’ (where GHG is greenhouse gas), ‘21ka Ice Sheets’, ‘21ka LCIS’ (where LCIS is the combined Laurentide–Cordilleran ice sheets), ‘21ka Shelf Exp.’ and ‘21ka LCIS albedo’. Each of these simulations is defined in Supplementary Data. All simulations are run for at least 500 yr with analysis made on the final 200 yr of each simulation. (The term ‘ka’ refers to ‘kyr ago’.)

Long-term evolution of ASL variability. 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 Basin¹⁴ (Fig. 3a). These changes in the circulation can be characterized by the 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 $55^\circ\text{--}70^\circ\text{S}$ and $195^\circ\text{--}240^\circ\text{E}$ (Z_{ASL}). We use a 4–15-yr band pass filter to isolate variability on the same timescale as the WDC record. Extended Data Fig. 3 shows the variance of Z_{ASL} within each snapshot of a thousand years, using the ‘Full 28–0ka’ simulations. From 28 kyr until around 15 kyr ago, 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.

Internal versus forced variability of ASL. 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 $5^\circ\text{N--}5^\circ\text{S}$ and $150^\circ\text{--}270^\circ\text{E}$ (SST_{pac}) from the Z_{ASL} . This operation is performed by regressing Z_{ASL} onto SST_{pac} as shown by the equation $Z_{\text{pac}} = \text{SST}_{\text{pac}} \times \text{regress}(\text{SST}_{\text{pac}}, Z_{\text{ASL}})$. We then subtract Z_{pac} from Z_{ASL} to obtain the component of Z_{ASL} that is unrelated to the tropical Pacific: $Z_{\text{local}} = Z_{\text{ASL}} - Z_{\text{pac}}$. The variance of Z_{local} is plotted in Extended Data Fig. 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% significance, F -test). We compare these results, obtained with the 4–15-yr band pass filtered data of monthly mean model output, to the 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.

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 $\pm 0.81^\circ\text{C}$ (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 analysis⁴⁸ 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 period (Extended Data Fig. 4a) and for the ‘Full 21ka’ simulation (Extended Data Fig. 4b). The difference between the two patterns is shown in Extended Data Fig. 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 28–0ka’ simulations, we evaluate the amplitude of the tropical Pacific-to-ASL teleconnection as the mean of the composite averaged over $55^\circ\text{--}70^\circ\text{S}$ and $195^\circ\text{--}240^\circ\text{E}$ (Extended Data Fig. 5). There is a dramatic decrease in the strength of the teleconnection around 16 kyr ago.

Teleconnection changes resulting from ENSO changes using HadCM3. We find that the simulated variance of SST_{pac} anomalies 21 kyr ago ($\sigma = 0.78^\circ\text{C}$) is increased by 40% compared to the pre-industrial period ($\sigma = 0.65^\circ\text{C}$). To determine the effect of SST variability on the teleconnection strength, we re-compute the composites using SST_{pac} anomalies in the range $\pm 0.81^\circ\text{C} < \text{SST}_{\text{pac}} < \pm 1.63^\circ\text{C}$, corresponding to SST_{pac} at 0 kyr ago of $\pm 1.25\sigma$ to $\pm 2.5\sigma$. 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 kyr to 0 kyr ago, as well as how they differ from the pre-industrial period, are unchanged by imposing the upper and lower bound, while the amplitude of the composites is slightly reduced. For the 'Full 21ka' 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 period. *Teleconnection changes resulting from ENSO changes using the PMIP.* For a subset of PMIP2/3 simulations (those for which sufficient data was available⁵²), 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 the 'Pre-industrial' and 'Full 21ka' 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 boundary conditions 21 kyr ago is also different.

In the LGM simulations, 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 (that is, ENSO). Indeed, in comparison to HadCM3, the most similar change in the teleconnection occurs in FGOALS-1.0g simulations, part of the PMIP2/3 model suite. FGOALS-1.0g shows a large decrease in ENSO strength at 21 kyr ago, 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 centre in the pre-industrial simulations.

Boundary condition changes responsible for the teleconnection change. In 'Full 21ka' simulations, a number of the climate forcings are substantially changed compared to the pre-industrial period. These include insolation, the greenhouse gas 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 '21ka Orbit + GHG' simulation does not cause a significant change (Extended Data Fig. 8b). In the '21ka Ice Sheets' simulation, which includes reduced sea level, enhanced albedo in the higher latitudes, and altered topography over North America, Europe, Antarctica, and the Southern Andes, there is a large change in the amplitude of the teleconnection (Extended Data Fig. 8c). Prior modelling simulations have explored other aspects of the topographic effects of Northern Hemisphere ice sheets⁵³, including their influence on glacial climate⁵⁴, abrupt glacial climate change⁵⁵, the Atlantic Meridional Overturning Circulation⁵⁶, and Heinrich events⁵⁷.

The '21ka LCIS only' simulation displays a statistically significant ($P < 0.05$) change in the teleconnection strength, similar in magnitude to the 'Full 21ka' simulation (Extended Data Fig. 8e). The '21ka LCIS albedo' simulation (that is, 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 that topography is the dominant mechanism acting to change the teleconnection strength. This agrees with the WDC shift in variability about 16 kyr ago, which is concurrent with the abrupt lowering of the LCIS¹⁸.

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 '21ka Shelf Exp.' simulation (Extended Data Fig. 8d). This results in a change to Z_{ASL} that is 3 to 4 times smaller than that seen in the '21ka LCIS only' simulation. There is also a statistically significant change in the composite map only to the north of the ASL. In the WDC diffusion-corrected variability (3–7-yr and 4–15-yr bands), there appears to be a step-change that occurs at about 13 kyr ago. 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 are detecting flooding of the continental shelves, but this effect is secondary to the abrupt change at approximately 16 kyr ago.

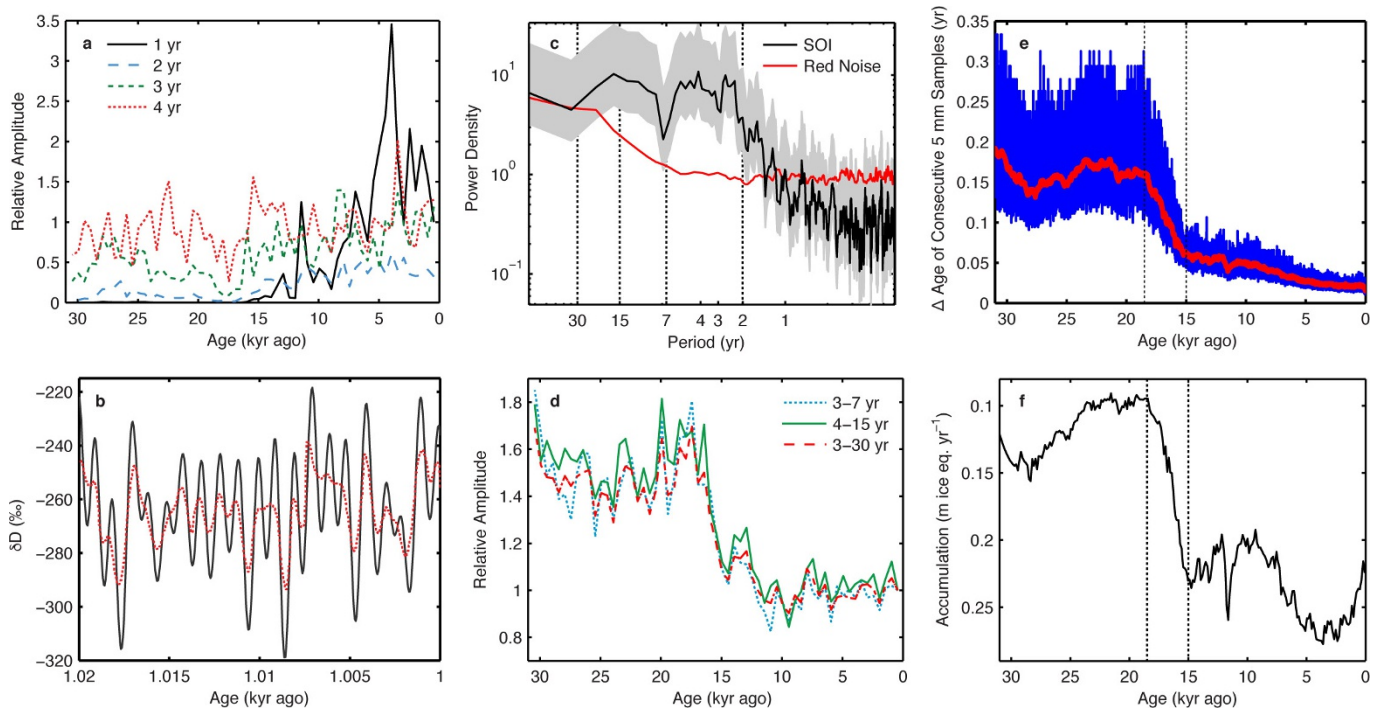
Climatic changes associated with the teleconnection change. The tropical Pacific-to-ASL teleconnection can be considered as a Rossby Wave response to diabatic heating in the western tropical Pacific⁶⁰. The location of this diabatic heating will depend upon the mean state. To understand how the mean state changes, we analyse how the patterns of precipitation and SST change in the '21ka' simulations relative to the pre-industrial period. Extended Data Fig. 9a shows the precipitation difference between the 'Full 21ka' simulation and the 'Pre-industrial' simulation. There is a prominent decrease in precipitation over the Maritime Continent, and an increase in precipitation over the Indian Ocean and the western and central tropical Pacific. This is consistent with proxy evidence^{21,23,61}. In Extended Data Fig. 9b, the '21ka Ice Sheets' simulation accounts for the majority of the precipitation changes in the 'Full 21ka' simulation. When we decompose this effect, the '21ka LCIS only' simulation accounts for most of the precipitation changes in the western and central tropical Pacific (Extended Data Fig. 9c), while the '21ka Shelf Exp.' 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 '21ka LCIS only' simulation than in the '21ka Shelf Exp.' 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. Around 16 kyr ago, 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 21ka', '21ka Ice Sheets', and '21ka LCIS only' simulations are all similar. Although the global mean SST is much cooler in the 'Full 21ka' simulation (Extended Data Fig. 9a), it is the spatial structure of the SST that determines the climatic response⁶². Therefore 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 LGM PMIP2/3 simulations, indicating a robust SST response to an ice sheet. The SST pattern in the 'Full 21ka' simulation is associated with cyclonic flow in the North Pacific that is forced by the presence of the LCIS. Prior modelling 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 Low⁵³.

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 Supplementary Data and Extended Data Figs 1–9.

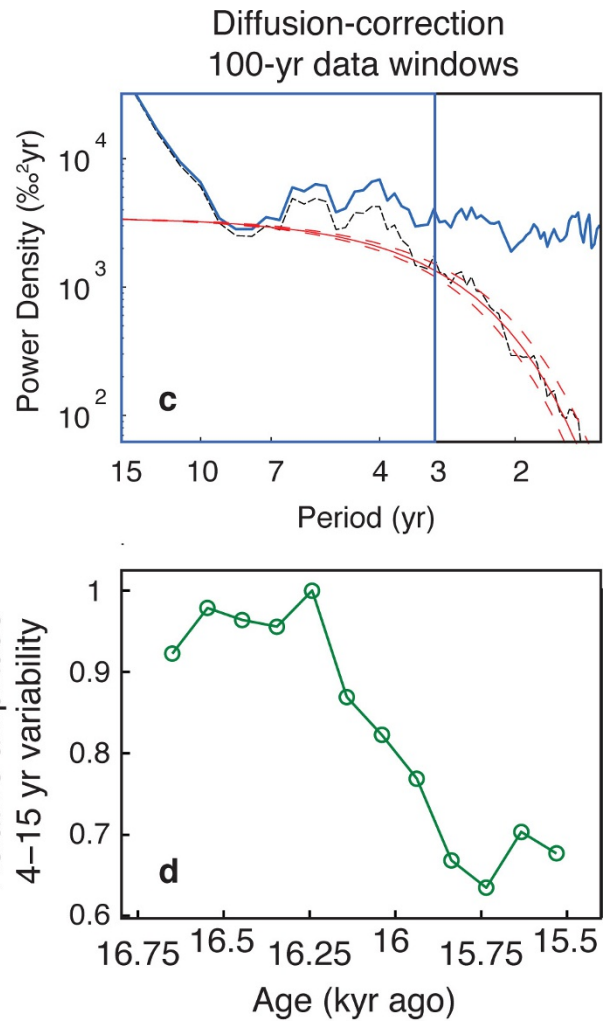
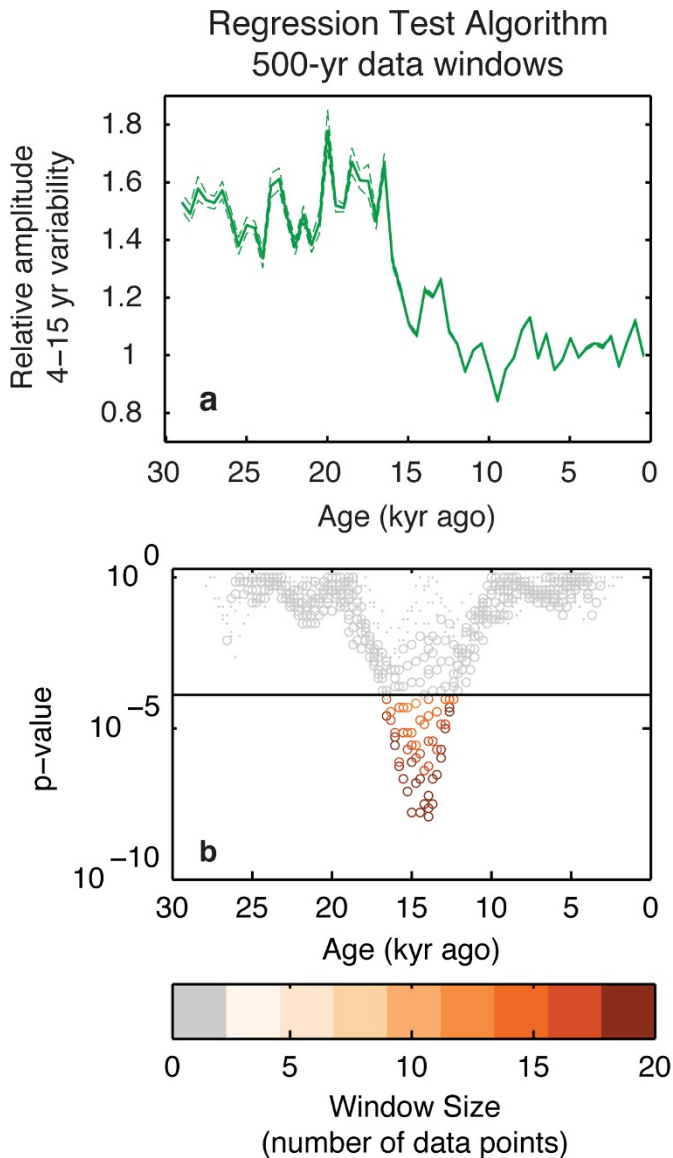
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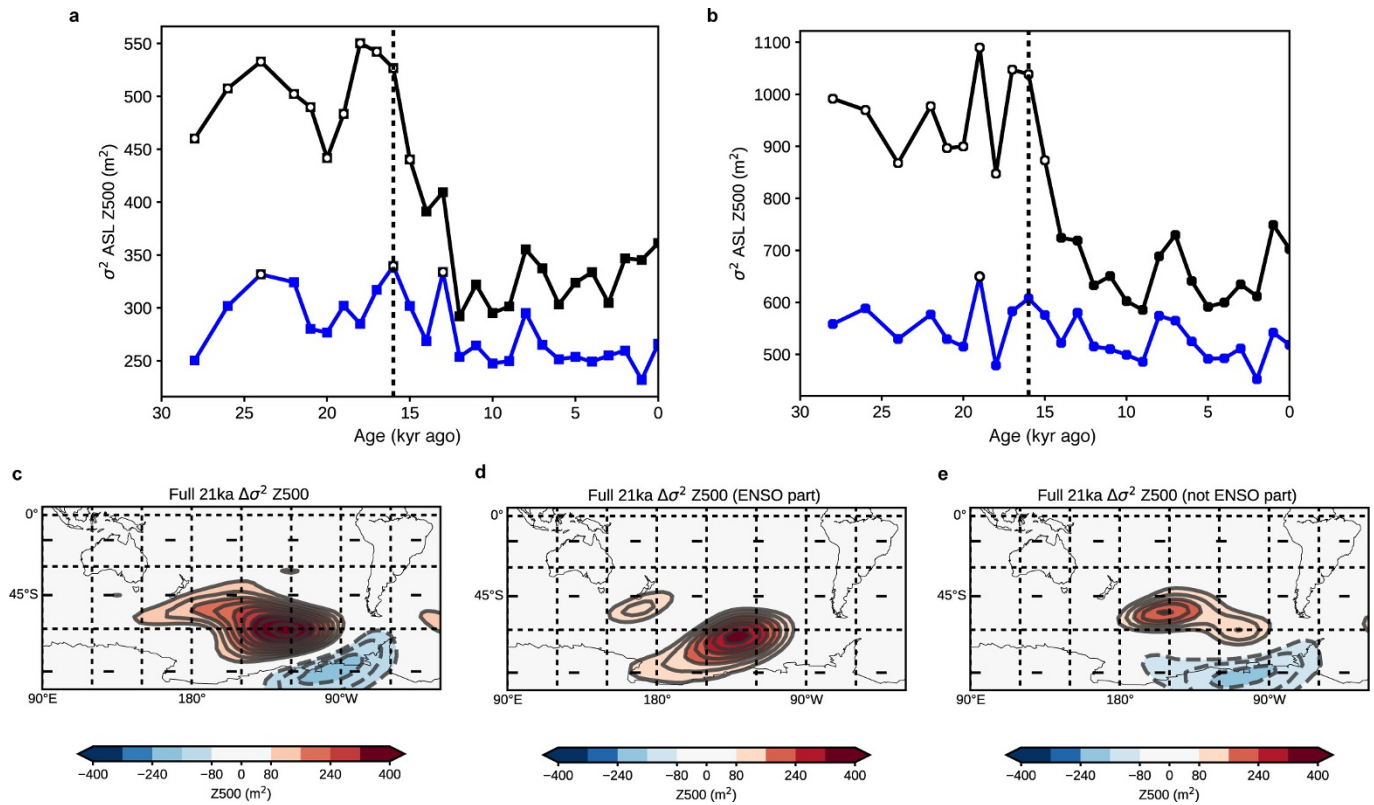
Extended Data Figure 1 | Signal detection. **a**, Relative amplitudes for 1-yr, 2-yr, 3-yr and 4-yr periods calculated for 500-yr spectral data windows, normalized to the value for the annual signal in the most recent data window. Both the climate and diffusion affects the amplitude of these high-frequency signals. However, it is mainly the effects of diffusion that cause the loss of the annual signal at >14 kyr ago. Similarly, the 2-yr period is lost between 17 kyr ago and 19 kyr ago, when the rate of diffusion for WDC was highest¹⁶. Periods >3 yr survive diffusion throughout the past 31 kyr. **b**, An example of a deconvolution calculation showing the observed δD water isotope record (that is, raw data; dotted red line), and the diffusion-corrected record (black line). Although this calculation was not used for the data presented in this paper, it serves as a visual aid 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 Southern Oscillation Index from 1951 to 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 Southern Oscillation Index. The Southern Oscillation Index has power greater than the red noise across a broad spectral peak between 2 and 17 yr, which can be subdivided into a 2–7-yr high-frequency peak and an 8–17-yr peak. Owing to the limited temporal span of modern observations, multi-decadal spectral estimates of the Southern Oscillation Index cannot be adequately defined. **d**, Diffusion-corrected relative amplitudes using 500-yr windows of WDC water isotope data. The difference in age of consecutive 5-mm WDC water isotope samples (blue) and a 500-yr sliding average (red). **e**, The difference in age of consecutive 5-mm WDC water isotope samples in **e**, undergoes large changes during the deglaciation at around 18.5 kyr ago, occurring 2.5 kyr before the change point in teleconnection strength at about 16 kyr ago. **f**, The WDC accumulation rate⁶³, inverted. The accumulation (in metres of ice equivalent per year), and by extension the difference in age of consecutive 5-mm WDC water isotope samples in **e**, undergoes large changes during the deglaciation at around 18.5 kyr ago, occurring 2.5 kyr before the change point in teleconnection strength at about 16 kyr ago.



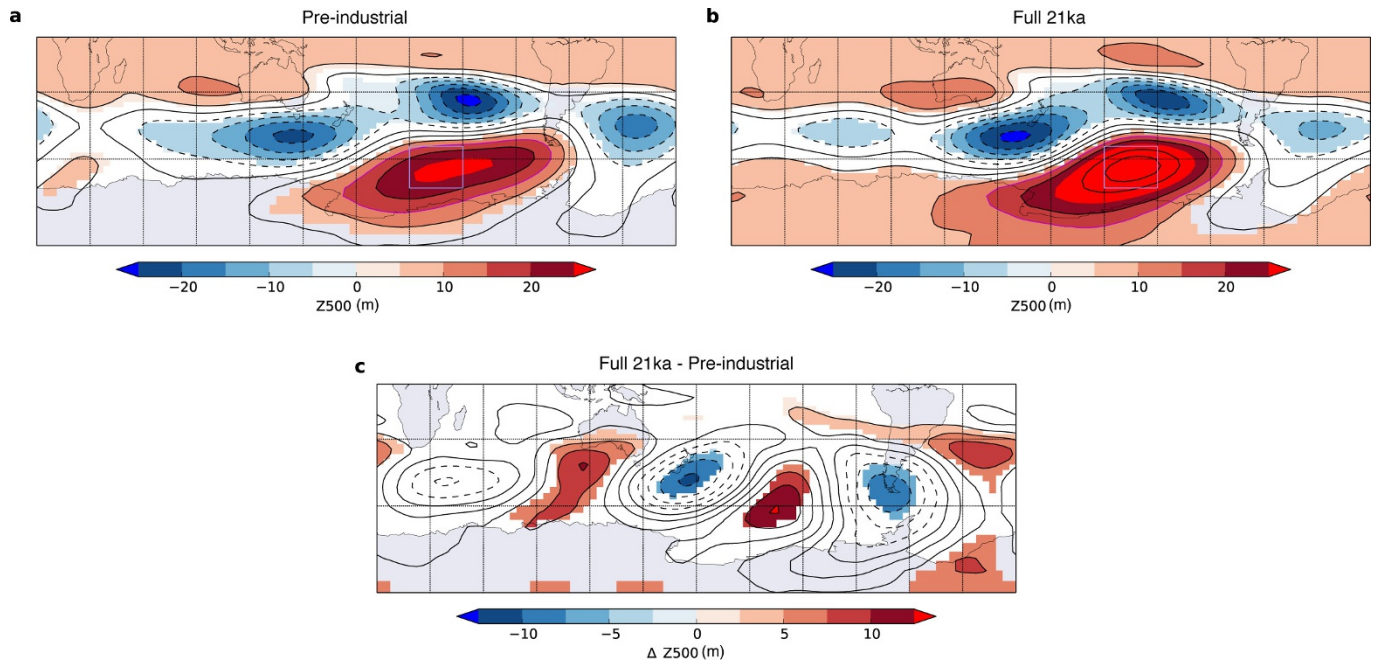
Extended Data Figure 2 | Change point detection. **a**, WDC 4–15-yr variability for 500-yr data windows (dashed lines are 1σ uncertainties; see Methods). **b**, Regression test algorithm to determine the first significant change in the WDC 4–15-yr variability in **a**. The first coloured data point below the P -value significance threshold occurs 16.44 kyr ago. **c**, Example of the diffusion-correction calculation for a 100-yr data window centred

on 15.54 kyr ago; 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 subset of diffusion-corrected 4–15-yr amplitudes (green) calculated at 100-yr resolution. The values are normalized to the amplitude value at 16.24 kyr ago, which represents the change point towards smaller amplitudes.



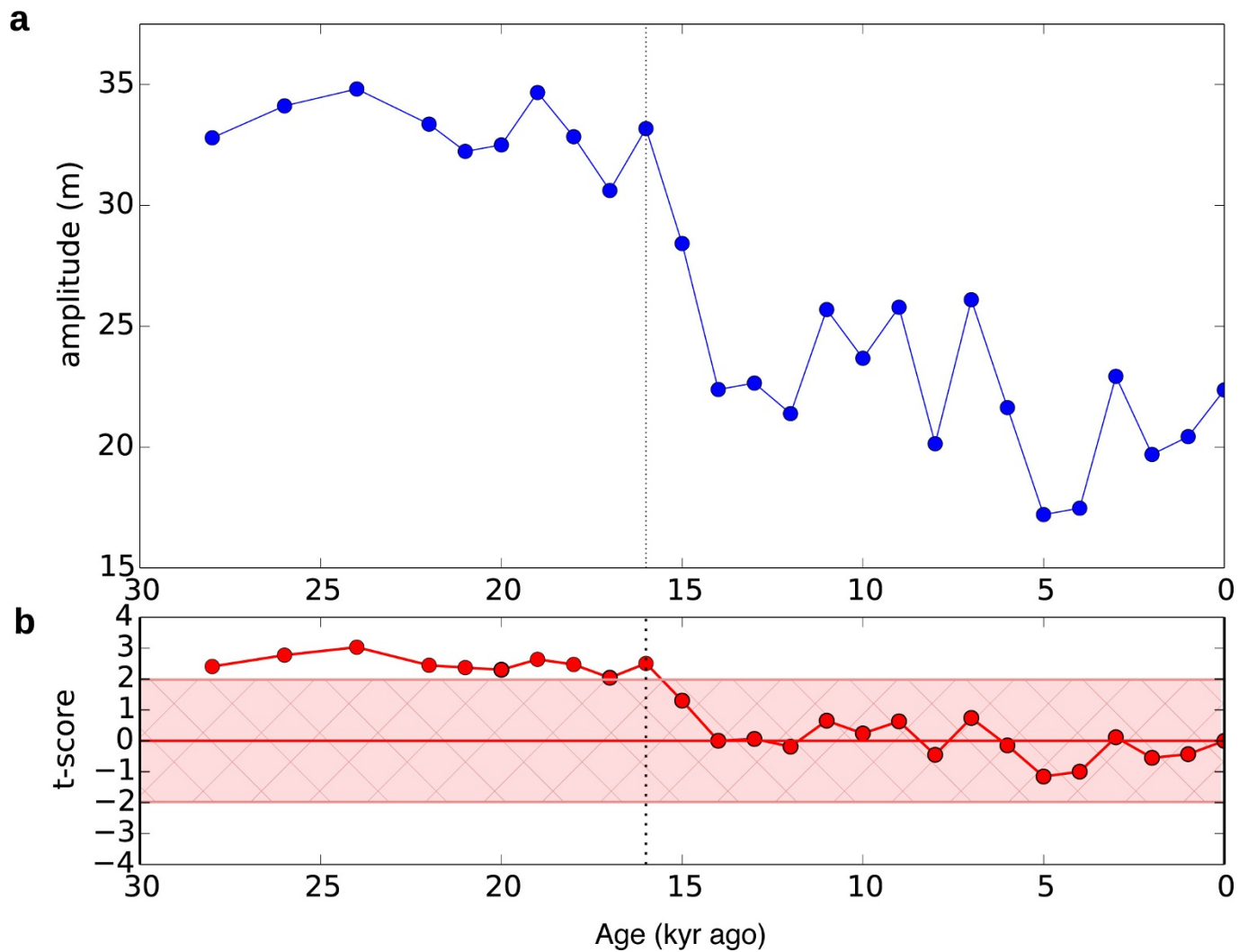
Extended Data Figure 3 | ASL variability. Panels **a** and **b** show the variance σ^2 of the indices Z_{ASL} (black) and Z_{local} (blue). **a**, The indices are computed from monthly mean data, then filtered with a 4–15-yr band pass filter. **b**, The indices are computed from annual mean output. We 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 period. **c**, Map of the change in the variance of the 500-hPa height field between 21 kyr ago and the pre-industrial period. **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–e** the contour intervals are 40 m^2 , with colours changing every 80 m^2 . Negative contours are dashed.



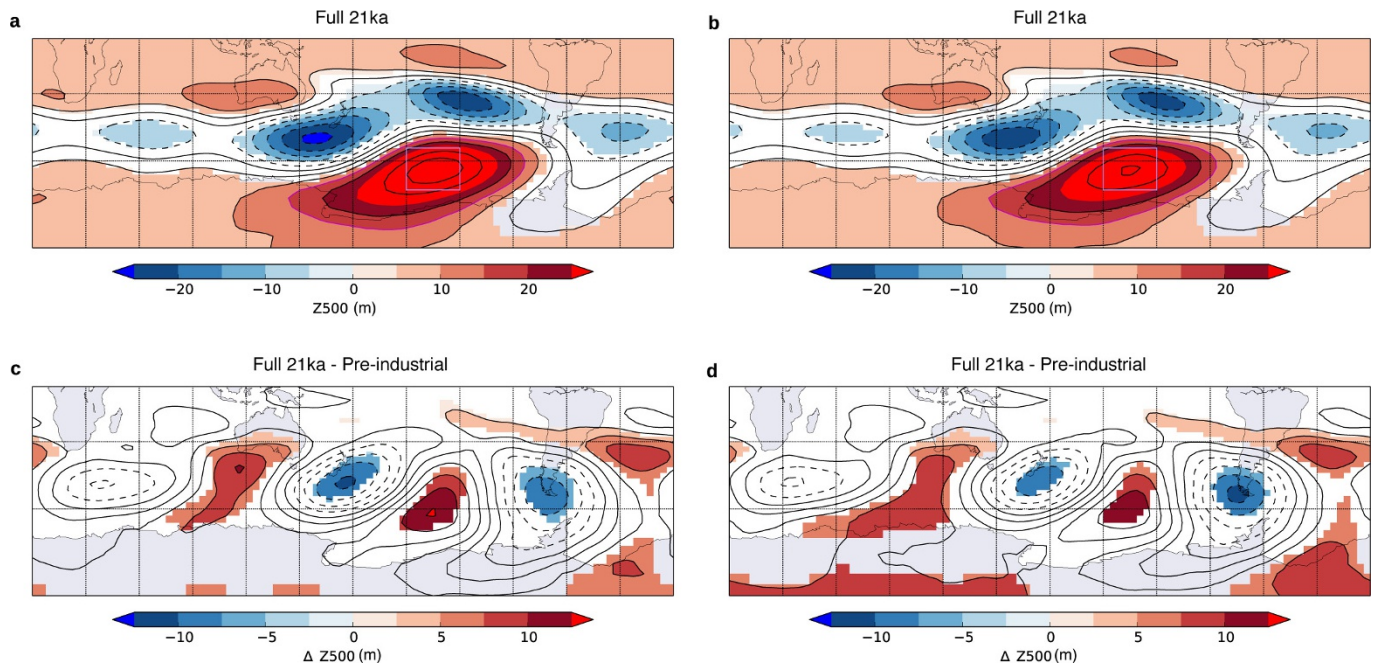
Extended Data Figure 4 | Composites of the 500-hPa height field for ENSO events. a–c, Composite for the ‘Pre-industrial’ (a), ‘Full 21ka’ (b) and the difference between the ‘Full 21ka’ and ‘Pre-industrial’ simulations (c). Contours are filled (not white) when statistical significance exceeds

95% using a Monte Carlo test. Negative contours are dashed. In **a** and **b** the contours are plotted every 5 m and colours 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 colours saturate at ± 12.5 m.



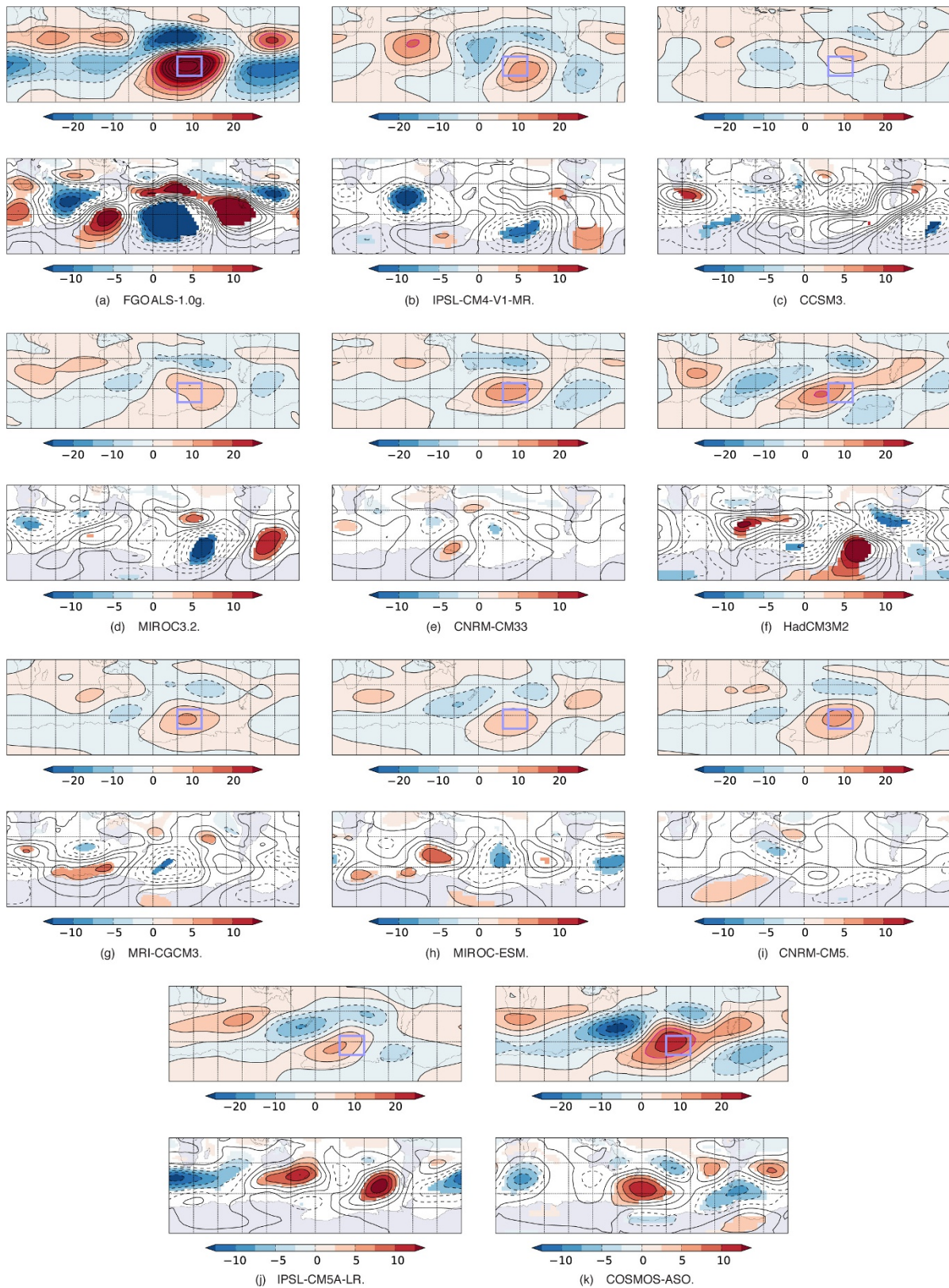
Extended Data Figure 5 | Tropical Pacific-to-ASL teleconnection strength, computed using the ‘Full 28–0ka’ simulations. a, Average 500-hPa geopotential height anomaly in the Amundsen Sea region, computed within the purple box shown in Extended Data Fig. 4. These

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 indicated by the red line. Values outside the shaded red region are 95% statistically different from the pre-industrial value.



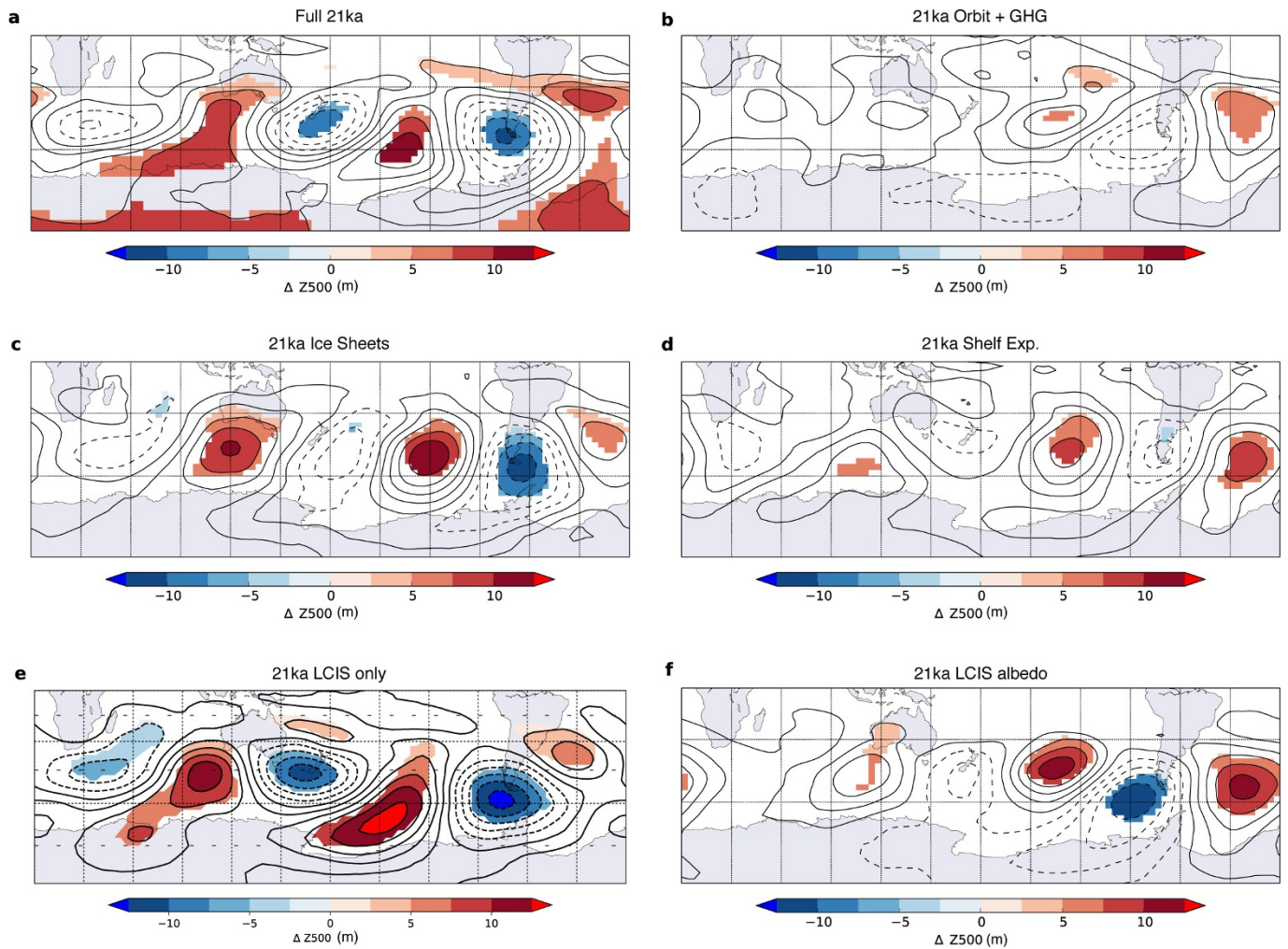
Extended Data Figure 6 | Composites of the 500-hPa height field for ENSO events with and without upper bound. a, The teleconnection without upper limit using the 'Full 21ka' simulation. **b,** The teleconnection with upper limit using the 'Full 21ka' simulation. **c,** The difference in teleconnection between the 'Full 21ka' simulation and the 'Pre-industrial' simulation without upper limit. **d,** The same difference as c with upper

limit. In all panels, negative contours are dashed, and contours are filled (not white) when statistical significance exceeds 95% using a Monte Carlo test. In **a** and **b**, the contours are plotted every 5 m and colours saturate at ± 25 m. In **c** and **d**, contours are plotted every 2.5 m and colours saturate at ± 12.5 m.



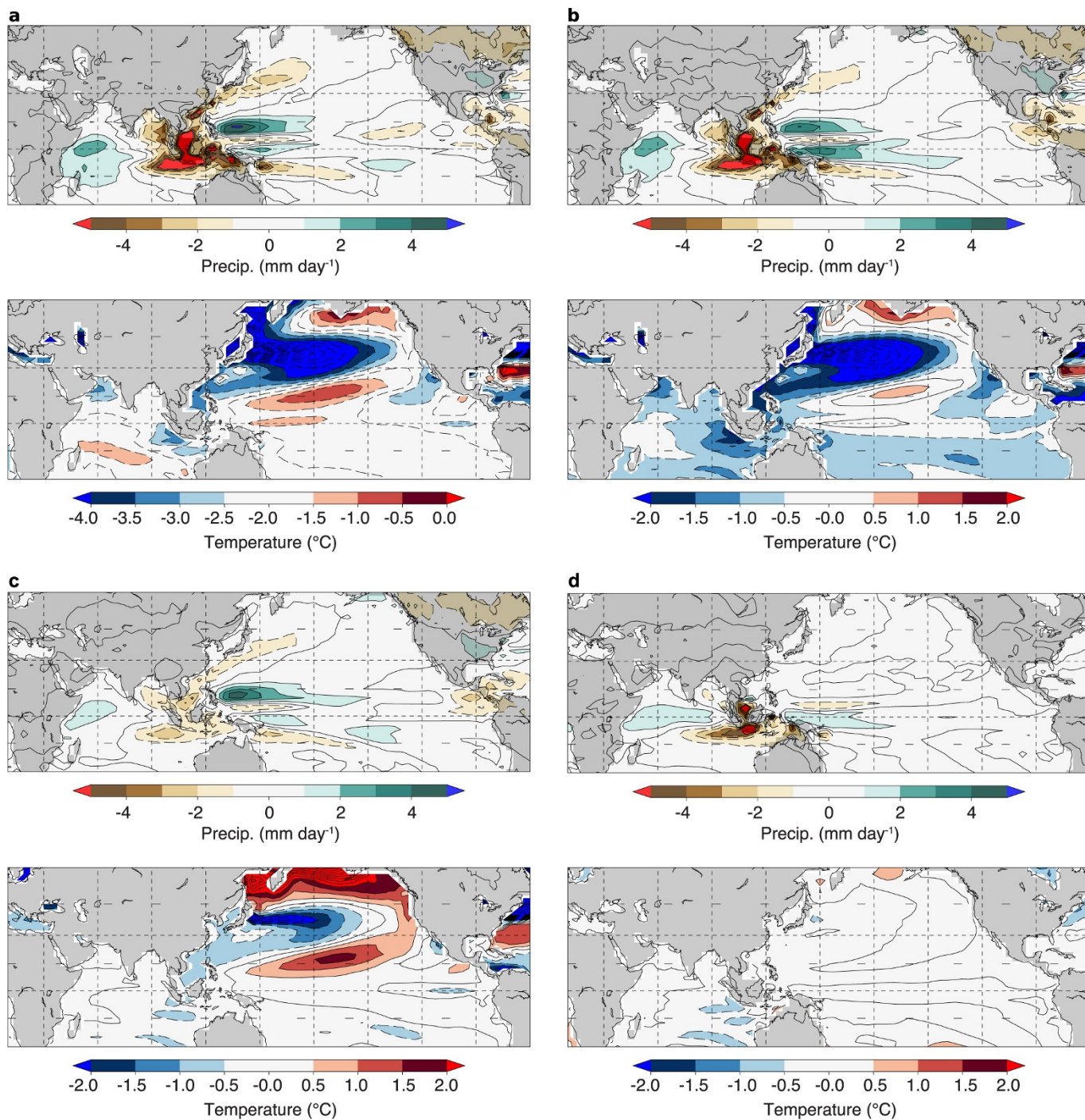
Extended Data Figure 7 | Composite maps of the 500-hPa height field for ENSO events in the PMIP2/3 models. For each model in a–k, the top panel shows the ‘Pre-industrial’ composite and the bottom panel the difference between the ‘Full 21ka’ and ‘Pre-industrial’ simulations. In the bottom panel, contours are filled (non-white) when statistical significance exceeds 95% using a Monte Carlo test. In the top panels, contours are

plotted at 5-m intervals, with colours saturating at 25 m. In the bottom panels, contours are plotted at 2.5-m intervals, with colours saturating at ± 12.5 m. Negative contours are dashed. The reduced statistical significance in these panels compared to those shown in Extended Data Figs 4, 6, 8 and 9 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 boundary conditions 21 kyr ago. All plots show the difference between 21-kyr and 0-kyr 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 colours saturate at ± 12.5 m. Contours are filled (not white) when statistical significance

exceeds 95% using a Monte Carlo test. **a**, 'Full 21ka' simulation. **b**, '21ka Orbit + GHG' simulation. **c**, '21ka Ice Sheets' simulation. **d**, '21ka Shelf Exp.' simulation. **e**, '21ka LCIS only' simulation. **f**, '21ka LCIS albedo' simulation. Each of these simulations is fully defined in Supplementary Data.



Extended Data Figure 9 | Annual mean anomalies of precipitation and sea surface temperature. Maps of anomalies 21 kyr ago relative to the pre-industrial period. **a**, 'Full 21ka' simulation. **b**, '21ka Ice Sheets' simulation. **c**, '21ka LCIS only' simulation. **d**, '21ka Shelf Exp.' simulation. Annual means are calculated from 100 years of output. Contour intervals for

precipitation are 1 mm per day, and for sea surface temperature are 0.5 °C. Land areas are shown in grey. Note that the temperature colour scale in **a** ranges from -4 °C to 0 °C. This accounts for the mean greenhouse gas cooling that is seen in the 'Full 21ka' simulation.