Antarctic Warming during Heinrich Stadial 1 in a Transient Isotope-Enabled Deglacial Simulation

CHENYU ZHU,a,b JIAXU ZHANG,c,d ZHENGYU LIU,e,f BETTE L. OTTO-BLIESNER,g CHENGEFEI HE,e,h ESTHER C. BRADY,g ROBERT TOMAS,g QIN WEN,i QING LI,j CHENGUANG ZHU,k

ABSTRACT: Heinrich Stadial 1 (HS1) was the major climate event at the onset of the last deglaciation associated with rapid cooling in Greenland and lagged, slow warming in Antarctica. Although it is widely believed that temperature signals were triggered in the Northern Hemisphere and propagated southward associated with the Atlantic meridional overturning circulation (AMOC), understanding how these signals were able to cross the Antarctic Circumpolar Current (ACC) barrier and further warm up Antarctica has proven particularly challenging. In this study, we explore the physical processes that lead to the Antarctic warming during HS1 in a transient isotope-enabled deglacial simulation iTRACE, in which the increased meridional heat transport alone, that lead to the Antarctic warming during the early stage of HS1 without notable changes in the strength and position of the Southern Hemisphere midlatitude westerlies. In particular, when a reduction of the AMOC causes ocean warming to the north of the ACC, increased southward ocean heat transport through the ocean and then through the atmosphere, can explain the Antarctic warming during the early stage of HS1 through the ocean and then through the atmosphere, can explain the Antarctic warming during the early stage of HS1.

KEYWORDS: Antarctica; Paleoclimate; Climate models; Atmospheric circulation; Ocean circulation; Greenhouse gases

1. Introduction

The deglacial warming of Earth from the Last Glacial Maximum (LGM) was punctuated by a few abrupt events, notably the Heinrich Stadial 1 (HS1; \(18–14.7 \text{ ka, i.e., } 18000–14700 \text{ years ago} \)). Heinrich Stadial 1 (HS1) was the major climate event at the onset of the last deglaciation associated with rapid cooling in Greenland and lagged, slow warming in Antarctica. Although it is widely believed that temperature signals were triggered in the Northern Hemisphere and propagated southward associated with the Atlantic meridional overturning circulation (AMOC), understanding how these signals were able to cross the Antarctic Circumpolar Current (ACC) barrier and further warm up Antarctica has proven particularly challenging. In this study, we explore the physical processes that lead to the Antarctic warming during HS1 in a transient isotope-enabled deglacial simulation iTRACE, in which the increased meridional heat transport alone, that lead to the Antarctic warming during the early stage of HS1 without notable changes in the strength and position of the Southern Hemisphere midlatitude westerlies. In particular, when a reduction of the AMOC causes ocean warming to the north of the ACC, increased southward ocean heat transport through the ocean and then through the atmosphere, can explain the Antarctic warming during the early stage of HS1 through the ocean and then through the atmosphere, can explain the Antarctic warming during the early stage of HS1.
This delayed thermal seesaw has been found to be a robust feature between abrupt North Atlantic climate events (Dansgaard–Oeschger events) and Antarctic isotope maxima events during the last glacial (e.g., WAIS Divide Project Members 2015). The timing of the interpolar phasing has recently been evaluated by comparing the WAIS (West Antarctic Ice Sheet) Divide ice core (WDC) with the Greenland NGRIP (North Greenland Ice Core Project) core, suggesting a precise time lag of 208 ± 96 years (2σ) for all the Greenland cooling events leading the corresponding Antarctic warming onsets (WAIS Divide Project Members 2015). The slow nature of oceanic adjustment and the heat reservoir effect of the Southern Ocean has been hypothesized to account for the substantial time lag for Antarctica temperatures relative to its northern counterpart (Stocker and Johnsen 2003). Using transient hosing experiments conducted with the Community Climate System Model version 3 (CCSM3), Pedro et al. (2018) further pointed out such a heat reservoir is located at the global ocean interior to the north of the Antarctic Circumpolar Current (ACC).

How the heat that stored in the ocean interior north of the ACC was released and warmed up the Antarctic, however, remains unclear, especially considering the fact that the Antarctic land surface was elevated on average 2500 m above the sea level. Atmospheric teleconnections have been hypothesized to provide a heat pathway, notably the Southern Annular Mode (SAM) pattern, which is the leading mode of the Southern Hemisphere (SH) atmospheric variability (Hartmann and Lo 1998; Thompson and Wallace 2000; Thompson and Solomon 2002; Lorenz and Hartmann 2001; Sen Gupta and England 2006). As revealed in paleo-records and idealized hosing experiments (Rahmstorf 2002; Stouffer et al. 2007), in response to the interhemispheric surface temperature asymmetry forced by a weakened AMOC, the intertropical convergence zone (ITCZ) shifted southward. The midlatitude westerly winds in the SH also shifted southward and/or got intensified, forming a positive phase of the SAM (Dahl et al. 2005; Broccoli et al. 2006; Krebs and Timmermann 2007; Markle et al. 2017). The positive SAM could have promoted wind-driven upwelling of relatively warm Circumpolar Deep Water (CDW) that released heat stored in the deep Southern Ocean, providing an atmospheric pathway that relayed the bipolar seesaw response to the Antarctic (Lee et al. 2011; Toggweiler and Lea 2010). Moreover, it has also been hypothesized that the strengthening of CDW ventilation could have released CO2 preserved in the deep ocean during the glacial time, inducing a globally in-phase warming (Ahn and Brook 2008; Toggweiler et al. 2006; Anderson et al. 2009; Shakun et al. 2012; Parrenin et al. 2013; Meniel et al. 2018). Applying the atmospheric teleconnection mechanisms drawn from the idealized hosing experiments to the Antarctic warming in HS1, or more generally the Antarctic isotope maxima events during the last glacial, however, turns out to be challenging, because these mechanisms could be dependent on both the background state and time scale. First, an atmospheric response to a weakened AMOC under glacial climate conditions could differ substantially from that under interglacial climate conditions, under which most hosing experiments are performed. In the present day, the extratropical atmospheric circulation in the Southern Hemisphere is overall zonally symmetric, but significant asymmetries associated mostly with quasi-stationary zonal wavenumbers 1 and 3 (ZW1 and ZW3) are embedded in the zonal flow (Raphael 2004). The ZW3 pattern, in particular, has a strong impact on Antarctic sea ice and poleward heat transport (Raphael 2004, 2007). For atmospheric variability in the SH, the leading mode is the SAM while the second one is characterized by a Pacific–South American (PSA) teleconnection (Thompson and Wallace 2000; Thompson and Solomon 2002; Timmermann et al. 2010). In some models, the weakening of AMOC under glacial conditions induces a zonal wavenumber-2 (ZW2) atmospheric response and a net warming of the polar/subpolar Southern Ocean, instead of a PSA teleconnection detected under present-day conditions (Timmermann et al. 2010). In another transient hosing experiment under glacial conditions, there is no evidence of enhanced Southern Ocean upwelling in response to the poleward shift of the SH westerlies (Pedro et al. 2018). Moreover, the atmospheric response to a North Atlantic cooling could be time dependent, and the features related to the phase of the interhemispheric response (i.e., the ~200-yr lag in Antarctica) may be missed or distorted severely in short simulations that last less than 100 years (e.g., Lee et al. 2011). In particular, when the SH warms up gradually in response to the northern cooling, anomalously negative meridional sea level pressure (SLP) gradients would arise between middle and high southern latitudes, reversing the initial positive SAM pattern that induced by the southward shift of the ITCZ.

The purpose of the current study is to investigate the lagged Antarctic warming relative to the North Atlantic cooling during HS1 in a transient simulation with more realistic glacial boundary conditions that lasts for several thousand years. Furthermore, the simulation uses an isotope-enabled Earth system model, which enables a more direct comparison between model results and proxy records. The isotope-enabled Transient Climate Evolution (iTRACE) simulation is described in section 2. In section 3, we first show that iTRACE faithfully captures the millennial-scale variability and the interpolar phasing as recorded in the ice core δ18O in Greenland and Antarctica. We then investigate the physical processes that lead to the Antarctic warming in a coupled atmosphere–ice–ocean perspective. In particular, we propose a mechanism for the Antarctic warming without invoking the shifting and/or strengthening in SH westerly winds. A summary and discussion are given in section 4.

2. Model and experiments

iTRACE is a set of transient simulations of the evolving global climate and water isotopes during the last deglaciation (He et al. 2021a). It was conducted with the isotope-enabled Community Earth System Model version 1.3 (iCESM1) (Brady et al. 2019). The resolution of atmosphere and land is nominally 2° (1.9° in latitude and 2.5° in longitude) with 30 vertical levels in the atmosphere; the resolution of ocean and sea ice is nominally 1° with 60 vertical levels in the ocean. The iCESM1 model has been shown to capture the late-twentieth-century structure of δ18O and δD over the global oceans reasonably well compared with proxies (Brady et al. 2019).
Starting from the LGM (20 ka), four forcing factors, namely the continental ice sheets (ICE), solar insolation associated with orbital configuration (ORB), greenhouse gases (GHGs), and meltwater fluxes (MWF), were applied additively to isolate the climate impact of individual deglacial forcings. The continental ice sheet configuration followed the ICE-6G reconstruction (Peltier et al. 2015) and was modified every 1000 years in the simulations. GHG concentrations (CO₂, CH₄, N₂O) were prescribed based on ice-core reconstructions (Monnin et al. 2001; Lüthi et al. 2008). MWF was applied based on sea level reconstructions, largely following the scheme used in TruCE-21ka (Liu et al. 2009). The baseline simulation was integrated with changing continental ice sheets and ocean bathymetry (ICE). Orbital forcing was then additively applied (ICE+ORB), followed by greenhouse gases (ICE+ORB+GHG). MWFs were introduced finally to generate a full-forcing simulation (ICE+ORB+GHG+MWF). This set of experiments allows us to investigate the effects of individual forcings on deglacial climate change. The simulation set has been used to study the LGM Antarctic temperature (Buizert et al. 2021), deglacial pan-Asian monsoon hydroclimate (He et al. 2021a), and deglacial Greenland oxygen isotopes (He et al. 2021b). In the current study, the decadal annual mean output from 19 to 11 ka is used for analysis, unless otherwise specified.

3. Results

a. Deglacial evolution of δ¹⁸O in Greenland and Antarctica

The simulated Greenland δ¹⁸O evolution in the full-forcing experiment closely resembles the NGRIP δ¹⁸O records. The annual mean δ¹⁸O largely follows the annual surface temperature (TS) as expected, notably an increase of δ¹⁸O by 4‰ for ~10°C warming during the BA and toward the end of the YD, as well as a 4‰ decrease in δ¹⁸O corresponding to a similar cooling into the YD (Fig. 1b). The simulated changes in δ¹⁸O and temperature roughly follow a temporal δ¹⁸O-temperature slope of 0.36‰ °C⁻¹ and is also consistent with independent calibrations (Figs. 1b,c; Buizert et al. 2014; He et al. 2021a). The simulated millennial variability in δ¹⁸O is associated with the millennial AMOC variability triggered primarily by meltwater fluxes, in agreement with the ²³¹Pa/²³⁰Th proxy in ocean sediment core GGC5 (Figs. 1e and 2; McManus et al. 2004). By comparing the full-forcing experiment with the other three experiments, we find the northern cooling during HS1 is dominantly attributed to the meltwater forcing, while the increased GHGs heat both hemispheres evenly, opposing the cooling in the north, and the retreat of continental ice sheets and the increase of insolation have negligible contributions (Fig. 2). We note the simulated annual mean δ¹⁸O at the onset of HS1 remains stable in NGRIP and GISP2 (Fig. 3b) as observed (Buizert et al. 2014; He et al. 2021b), while the simulated surface temperatures show a strong, systematic cooling of ~8°C in Greenland (Fig. 3a). The muted δ¹⁸O records have been attributed to a loss of winter precipitation associated with the North Atlantic cooling and sea ice expansion, which enriches the δ¹⁸O and compensates the depletion due to cooling (He et al. 2021b).

Fig. 1. Model-data comparison. (a) Forcing: June solar insolation (SOLIN) at 60°N (orange), atmospheric CO₂ concentration (green; Lüthi et al. 2008), and meltwater fluxes in the Northern Hemisphere (brown) and Southern Hemisphere (gray). (b) Simulated deglacial change in surface temperatures (°C) at WDC (red) and NGRIP (blue) cores. Also shown in (b) are the simulated δ¹⁸O anomalies (‰) for WDC (pink) and NGRIP (light blue). Red diamonds represent an optimal deglacial West Antarctica temperature reconstruction from the WDC borehole (Cuffey et al. 2016). Anomalies are relative to 19 ka means. (c) Observed (black) and simulated (light blue) δ¹⁸O at NGRIP. (d) As in (c), but for the WDC core. Note that model values in (c) and (d) are offset by −8‰ and −4‰, respectively, to better match the records. (e) Simulated NADW transport (Sv; 1 Sv ≡ 10⁶ m³ s⁻¹) in the North Atlantic (purple) and South Atlantic (green) and the observed ²³¹Pa/²³⁰Th in sediment core OCE326-GGC5 (33°42′N, 57°35′W; 4.55 km) as a proxy for AMOC intensity (McManus et al. 2004). (f) Simulated AABW strength (cyan; Sv) and SH mean sea ice concentration (gray). LGM, HS1, BA, and YD intervals are indicated by black vertical lines while the gray dashed vertical line around 17.8 ka marks the onset of Antarctic warming during HS1.

The slow Antarctic warming (cooling) that followed the rapid Greenland cooling (warming) is also reproduced in the model (Figs. 1d and 3c,d). The simulated WDC δ¹⁸O first increases by 8‰ with a warming of 8°C during HS1, then decreases by 2‰ with a cooling of 2°C during the BA [or
Antarctic cold reversal (ACR), and increases again by 4% with a warming of 4°C during the YD, qualitatively consistent with observations (Fig. 1d). The Southern Hemisphere warming phase during HS1 and the YD coincides with the two main intervals of increasing atmospheric CO₂ whereas the cooling during the BA corresponds to a plateau in atmospheric CO₂ (Figs. 1a,b; Shakun et al. 2012; Monnin et al. 2001; Lüthi et al. 2008). Note the simulated strong oscillation in Antarctic climate around 14 ka is triggered primarily by the abnormally strong meltwater forcing from the Southern Hemisphere (Fig. 1a). The simulated δ¹⁸O values at different Antarctica core sites increase coherently through HS1 (Figs. 3c,d), with a positive bias (∼4‰) against paleo-records (Fig. 1d; EPICA Community Members 2006; Barker et al. 2009; Parrenin et al. 2013). Likely due to topographic change, the warming at the West Antarctic sites (WDC) is generally larger than (by 2–3°C) that at the East Antarctic sites [EPICA (European Project for Ice Coring in Antarctica) Dronning Maud Land (EDML), EPICA Dome C (EDC), and Dome Fuji (DF); Figs. 3c and 6o–a], consistent with multiple reconstructions (Buizert et al. 2021).

Like Greenland, the temperature change in Antarctica in the model is insensitive to changes in continental ice sheets and insolation, but a ∼4°C warming (∼3‰ enrichment in δ¹⁸O) is forced by GHGs between 17.3 and 15 ka (Figs. 2b–d). The bipolar seesaw is reproduced only when meltwater flux forcing is applied (Fig. 2a). The simulated WDC warming of 10.5°C from the LGM to the early Holocene (11 ka) and warming of ∼8°C during HS1 are in good agreement with an optimal deglacial West Antarctica temperature reconstruction from the WDC borehole (Fig. 1b; Cuffey et al. 2016). Therefore, we speculate that the Antarctic warming during HS1 (8°C) is initially triggered and contributed by the AMOC weakening (4°C) associated with meltwater flux forcing and later reinforced by the GHG increase (4°C).

Interestingly, the observed interpolar phasing of the ∼200-yr lag between Antarctic warming and Greenland cooling during the cold phase of the Dansgaard-Oeschger cycle is reproduced very well in the model (Fig. 1; WAIS Divide Project Members 2015; Buizert et al. 2018). In iTRACE, the onsets for Greenland cooling and Antarctic warming during HS1 are 18 ka and ∼17.8 ka, respectively. This modeled 200-yr lag has been found at all Antarctic core locations relative to their Greenland counterparts (Fig. 3), indicating a spatially homogeneous oceanic bipolar seesaw mode. Previous studies suggest that another spatially heterogeneous atmospheric mode associated with the poleward shift of SH westerlies (a positive SAM) should be superposed over the oceanic mode to explain the delayed Antarctic warming (Pedro et al. 2018; Buizert et al. 2018). In the current simulation, however, the SH westerlies remain stable in both the position and strength from the LGM until much later (17.3 ka) in both annual and wintertime means (Fig. 4), suggesting that the shifting of the SH westerlies may not be a necessary ingredient in the delayed Antarctic warming during HS1.
we will investigate the governing mechanism for Antarctic warming in detail.

b. Mechanisms for Southern Ocean and Antarctic warming during HS1

1) OCEANIC WARMING NORTH OF THE ACC

We first focus on the early stage of HS1, spanning 18–17.5 ka, to investigate the oceanic warming to the north of the ACC. Figures 5 and 6 demonstrate the surface and subsurface climate change relative to the 19-ka climatology (calculated as the mean of 19–18 ka). An AMOC weakening results in the cooling in the North Atlantic surface and a warming in the subsurface (Figs. 5a–j), which has been shown to be a robust feature in both reconstructions (Marcott et al. 2011; Weldeab et al. 2016; Barker and Diz 2014; Dokken and Jansen 1999; Rasmussen and Thomsen 2004) and model simulations (Liu et al. 2009; Mignot et al. 2007; Brady and Otto-Bliesner 2011; Zhang et al. 2017; Pedro et al. 2018; He et al. 2020). The warming signal propagates southward along the Deep Western Boundary Current into the South Atlantic in 100–200 years, in agreement with the advection time scale for a signal to cross the Atlantic basin (Figs. 5f–o; Pedro et al. 2018). Beyond the South Atlantic basin, the warming signal is then brought to surface through wind-driven upwelling, reinforcing the surface warming that has been caused by the weakened northward heat transport. This lagged warming in the Southern Ocean in 100–200 years is further confirmed in a hosing experiment under the LGM boundary conditions using the same model, in which a constant 0.25 Sv (1 Sv ≡ 10^6 m^3 s^-1) freshwater flux was added to the North Atlantic for 100 years (Fig. S1 in the online supplemental material). The warming signal in the Atlantic sector of the Southern Ocean then spreads eastward along the ACC and penetrates northward into subtropics in other basins (Figs. 5a–j). Therefore, a hemispheric-scale warming is established around 200 years after the onset of HS1 (i.e., the onset of North Atlantic cooling; Fig. 5h). Note that the oceanic warming signal cannot easily spread across the ACC (Figs. 5a–o) due to a strong mean zonal flow and the lack of meridional boundary at the depth of NADW (Figs. 5p–t). The oceanic warming is therefore mostly confined to the north of the ACC, with the latter acting as a dynamic barrier (Fig. 5; Pedro et al. 2018; Schmittner et al. 2003). In this case, anomalous poleward heat transport cross the ACC is instead achieved by enhanced eddy fluxes, which will be discussed below.

2) WARMING BETWEEN 50° AND 70°S

Different from the ocean warming that is confined to the north of the ACC (Figs. 6a–e), the strongest surface atmospheric warming is centered south of 60°S, beyond the latitude of the ACC (Figs. 6k–o). Here we use the sea ice edge (defined as 15% annual sea ice concentration) as a
reference to the ACC location, as they generally coincide well with each other in the model (Figs. 5a–j). While the surface warming is most (least) significant in austral winter (summer), all four seasons share patterns of temperature anomalies similar to the annual mean (Fig. S2 in the online supplemental material). Since the surface air cannot be directly heated by the SSTs from below south of the ACC, and the SH westerlies remain stable and have been excluded as an explanation, we seek other atmospheric processes that are required to heat the atmosphere beyond the ACC.

We perform a heat budget analysis over two zonal bands 50°–70°S and 70°–90°S enclosing the Southern Ocean south of the ACC and the Antarctic domain following Pedro et al. (2018), which is shown schematically in Fig. 7. Four primary terms are taken into account: the net radiative flux at the top of the atmosphere (TOA), the air–sea interface heat flux (SHF), the total meridional atmospheric heat transport (AHT; determined as a residual unless otherwise specified), and the total meridional oceanic heat transport (OHT; integrated for the entire water column). The heat budget analysis, including the calculation and decomposition of AHT and OHT, largely follows Yang et al. (2015). We first check the glacial climatology at 19 ka. For the Southern Ocean domain between 50° and 70°S, OHT across its northern boundary contributes to a heat gain of −7 W m⁻², which is balanced by an upward SHF. For the corresponding atmospheric domain, the AHT across the northern and southern boundaries is −93 and −26 W m⁻² (negative heat transport

**Fig. 4.** Hovmöller diagrams of (a) decadal annual mean and (b) austral winter (JJA) mean SH surface zonal wind (units: m s⁻¹). In each panel, the gray solid line indicates the meridional shift of westerly winds maximum while the black dashed line denotes the 19 ka mean position of westerly winds maximum. Note the negligible westerly shift before 17.3 ka.
means southward heat transport), respectively. A net heat outflux of ~7.5 W m⁻² to the outer space occurs at the TOA.

During HS1, a mild increase in southward OHT is detected across the 50°S boundary and initiates the warming to the south of the ACC (Fig. 7). What could lead to the increased OHT across the ACC when the ACC acts as a dynamic barrier? We find the answer lies in the eddy-induced heat transport. In the mean state of 19 ka, the total OHT averaged over the 50°–70°S band is ~6.9 W m⁻², with its Eulerian component being ~5.6 W m⁻² and eddy component being ~1.3 W m⁻². During HS1, the ocean warming to the north of the ACC acts to increase the meridional temperature gradient and steepen the isopycnals (Figs. 5p–t), thus increasing baroclinic instability and mesoscale eddy activities [parameterized by the Gent–McWilliams (GM) parameterization in the model; Gent et al. 2011], and their associated southward heat transport (Fig. 8a). The southward OHT across 50°S at 17.5 ka is increased by ocean eddies while partly compensated by the decreasing Eulerian heat transport (caused by the weakened AMOC), resulting in a net increase of 0.5 W m⁻².

Despite being mild, the ocean warming to the south of the ACC triggers sea ice melt starting at ~17.8 ka, and by 17.6 ka the melt is significant (Figs. 1f and 6f–j). The sea ice retreat results in a reduction in local surface albedo and an anomalous radiation gain of 3.5 W m⁻² at the TOA (mostly via shortwave radiation) by 17.5 ka (Fig. 7). Although with small change in ocean temperature south of the ACC, the resulting sea ice melting initiates a positive albedo feedback, steadily warming the atmospheric column between 50° and 70°S since 17.8 ka.
3) ANTARCTIC WARMING WITH STRENGTHENING ZW3 PATTERN

The atmospheric mode around Antarctica in the first 100 years after the onset of HS1 is characterized by a deepening of Amundsen Sea low (ASL; Fig. 9a). In response to the North Atlantic cooling, the mean ASL at 18.0–17.9 ka deepens by ∼1.5 hPa (Fig. 9a; more profound in austral winter; not shown). This deepening is likely operated by the PSA teleconnection, associated with the western and/or central tropical Pacific cooling and the reduced deep convection there through Rossby wave dynamics (e.g., Timmermann et al. 2010; Ding et al. 2011; Li et al. 2021). The deepening of the ASL leads to strengthened onshore winds over the Ross Sea and offshore winds over the Bellingshausen Sea, cooling the west of Antarctica Peninsula (Figs. 6k and 9f), consistent with the climate impact of ASL variability observed in the present day (e.g., Raphael et al. 2016).

As the atmospheric column between 50° and 70°S is warmed up, the Antarctic climate is further heated by a strengthened southward AHT across 70°S. For the Antarctic domain south of 70°S, the mean balance at glacial time is mostly between the southward AHT (80 W m⁻²) and the outgoing heat flux (85 W m⁻²) at the TOA, with the surface flux being negligible (∼5 W m⁻²; Fig. 7). The southward AHT is mainly attributable to southward dry static energy (DSE) transport (∼70 W m⁻²) while latent energy (LE) plays a minor role. Since 17.8 ka, the effective warming of the atmosphere between 50° and 70°S leads to increased AHT to both the northern and southern boundaries. By 17.5 ka, the southward AHT across 70°S is increased by more than 2 W m⁻², warming up the Antarctic climate (Fig. 7). The increase in AHF across 70°S is dominated by increase in DSE while the change of LE is minor and can be neglected (not shown).

The strengthened southward AHT is achieved primarily by strengthening stationary waves associated with ZW3. In the mean climatology, there is a major stationary asymmetric component of large-scale SH atmospheric circulation associated with the meridional flow, resembling the present-day ZW3 pattern (van Loon and Jenne 1972; Raphael 2004, 2007), as can be seen from the decadal mean SLP and meridional winds (contours in Figs. 9a–j). When the SLP over lower latitudes is gradually decreased by surface warming, the meridional pressure gradients between the middle and high southern latitudes are decreased, favoring a strengthening ZW3 pattern extending southward to 70°S (shading in Figs. 9a–e).
The variability in meridional flow associated with anomalous ZW3 also explains the heterogeneous patterns of surface warming and sea ice melting across the Southern Ocean. The strengthening northward flow (Figs. 9h–j) is always accompanied by relatively weaker surface warming (Figs. 6m–o) and less sea ice melting (Figs. 6h–j), and vice versa, consistent with the advection effect associated with ZW3 in present-day observations (Raphael 2007). Here we define the ZW3 index as the averaged zonal deviation of the SLP at the center of the decadal mean positions of the three troughs (indicated in Figs. 9a–e; Raphael 2004). The dynamic component of DSE change (DSE_dynamic) dominates the total DSE change ($r = 0.7$ for 19–17 ka; $p < 0.01$; Fig. 8b), which closely resembles the evolution of ZW3 index ($r = 0.8$ for 19–17 ka; $p < 0.01$; Fig. 9c). Moreover, the cumulative anomalous poleward DSE transport associated with the strengthening ZW3 shows a strong increasing trend since 17.8 ka, consistent with the warming of Antarctica (Fig. 9c). We therefore argue that the strengthening stationary waves associated with ZW3 serve as a major contributor to the Antarctic warming without needs for shifting and/or strengthening in the SH westerlies over a longer time scale.

4. Discussion and conclusions

In this study, we revisit the Antarctic warming during HS1 in a set of isotope-enabled transient deglacial simulations iTRACE. iTRACE successfully captures the interpolar phase as recorded in $\delta^{18}O$ in Greenland and Antarctica during the last deglaciation. In particular, the simulated deglacial warming in West Antarctica matches especially well with reconstructions. During HS1, the oceans in the Southern Hemisphere slowly warm up to the north of the ACC in response to a weakened AMOC. The ACC acts as a dynamical barrier, prohibiting the warming signal from spreading southward either through advection or wave propagation. However, mild southward ocean heat transport anomalies across the ACC are still achieved by enhanced mesoscale eddy activities associated with the steepened Southern Ocean isopycnals. Despite being mild, such oceanic warming decreases the sea ice concentration and leads to more absorption of incident shortwave radiation, further heating the atmospheric column above it.

Fig. 7. Summary of 17.6–17.5 ka mean heat budget anomaly (W m$^{-2}$) south of 50°S. The 19-ka means are given in parentheses. Following Pedro et al. (2018), here the sign convention is defined so that positive anomalies contribute to warming of the atmospheric column in their respective latitude band. TOA: net radiative flux at the top of the atmosphere; SHF: air-sea heat flux; AHT: meridional atmospheric heat transport; OHT: meridional oceanic heat transport. Note the poleward AHT across 70°S is 0.7 W m$^{-2}$ out of the 50°–70°S band and 2.1 W m$^{-2}$ into 70°–90°S due to different areas of the two bands.

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Fig. 7. Summary of 17.6–17.5 ka mean heat budget anomaly (W m$^{-2}$) south of 50°S. The 19-ka means are given in parentheses. Following Pedro et al. (2018), here the sign convention is defined so that positive anomalies contribute to warming of the atmospheric column in their respective latitude band. TOA: net radiative flux at the top of the atmosphere; SHF: air-sea heat flux; AHT: meridional atmospheric heat transport; OHT: meridional oceanic heat transport. Note the poleward AHT across 70°S is 0.7 W m$^{-2}$ out of the 50°–70°S band and 2.1 W m$^{-2}$ into 70°–90°S due to different areas of the two bands.

The variability in meridional flow associated with anomalous ZW3 also explains the heterogeneous patterns of surface warming and sea ice melting across the Southern Ocean. The strengthening northward flow (Figs. 9h–j) is always accompanied by relatively weaker surface warming (Figs. 6m–o) and less sea ice melting (Figs. 6h–j), and vice versa, consistent with the advection effect associated with ZW3 in present-day observations (Raphael 2007). Here we define the ZW3 index as the averaged zonal deviation of the SLP at the center of the decadal mean positions of the three troughs (indicated in Figs. 9a–e; Raphael 2004). The dynamic component of DSE change (DSE_dynamic) dominates the total DSE change ($r = 0.7$ for 19–17 ka; $p < 0.01$; Fig. 8b), which closely resembles the evolution of ZW3 index ($r = 0.8$ for 19–17 ka; $p < 0.01$; Fig. 9c). Moreover, the cumulative anomalous poleward DSE transport associated with the strengthening ZW3 shows a strong increasing trend since 17.8 ka, consistent with the warming of Antarctica (Fig. 9c). We therefore argue that the strengthening stationary waves associated with ZW3 serve as a major contributor to the Antarctic warming without needs for shifting and/or strengthening in the SH westerlies over a longer time scale.

4. Discussion and conclusions

In this study, we revisit the Antarctic warming during HS1 in a set of isotope-enabled transient deglacial simulations iTRACE. iTRACE successfully captures the interpolar phase as recorded in $\delta^{18}O$ in Greenland and Antarctica during the last deglaciation. In particular, the simulated deglacial warming in West Antarctica matches especially well with reconstructions. During HS1, the oceans in the Southern Hemisphere slowly warm up to the north of the ACC in response to a weakened AMOC. The ACC acts as a dynamical barrier, prohibiting the warming signal from spreading southward either through advection or wave propagation. However, mild southward ocean heat transport anomalies across the ACC are still achieved by enhanced mesoscale eddy activities associated with the steepened Southern Ocean isopycnals. Despite being mild, such oceanic warming decreases the sea ice concentration and leads to more absorption of incident shortwave radiation, further heating the atmospheric column above it.

Fig. 8. (a) Decomposition of OHT change at 50°S (red) to contributions from Euler mean heat transport and eddy heat transport. (b) Decomposition of atmospheric DSE change $[\Delta(\vec{v}T)]$ to contributions from thermal effect (perturbation temperature on 19 ka mean meridional velocity; $\vec{v}\Delta T$ ) and dynamic effect (perturbation meridional velocity on 19 ka mean temperature; $\vec{T}\Delta v$ ) and the nonlinear effect diagnosed as a residual, shown in 50-yr running means. Anomalies are relative to 19 ka means. Given the mean state of southward OHT and AHT at their corresponding latitudes, negative (positive) anomaly indicates increase (decrease) in OHT and AHT (see text for more details). The decomposition method for OHT and AHT largely follows Yang et al. (2015).
The warmed atmosphere over the Southern Ocean further enhances the poleward atmospheric heat flux through strengthened stationary ZW3 pattern and warms up the Antarctic climate. In our simulation, we further find that for the total \( \sim 8 \)C of the West Antarctic warming during HS1, the partition due to AMOC bipolar seesaw accounts for \( \sim 4 \)C while the partition due to background CO\(_2\) increase accounts for another \( \sim 4 \)C after 17.3 ka. Other sites in the Antarctica share similar features but with smaller warming magnitude. Our study therefore suggests that both weakening AMOC and increasing CO\(_2\) could have contributed to the Antarctic isotope maxima events during the last glacial, with the climate
sensitivity to individual forcing likely being different between model and reality, as well as between models.

Our simulations show that the shift of the SH westerlies may not be necessary to initiate the Antarctic climate change as part of the bipolar seesaw. In our simulation, pronounced Antarctic warming starts in 17.8 ka, but there is no significant southward shift and/or strengthening in SH westerly winds until 17.3 ka. Instead, we identify a two-step atmospheric response that leads to the delayed Antarctic warming during HS1: 1) In the first 100 years after the HS1 onset, the PSA teleconnection pattern dominates and deepens the ASL, inducing a mild cooling in the west of Antarctica Peninsula by anomalous offshore winds; then 2) when the ocean warming to the north of the ACC (~50°S) becomes significant 200 years after the HS1 onset, more heat is accumulated in the atmosphere between 50° and 70°S through the sea ice albedo positive feedback; the anomalous heat reduces the SLP gradients between the Southern Ocean and Antarctica and triggers increasing stationary ZW3 activities, which primarily contributes to the Antarctic warming. The permanent anomalous ZW3 pattern also explains the heterogeneous feature of surface temperature and sea ice changes around Antarctica.

The Antarctic warming mechanism proposed in our study also differs from the previous hypothesis of enhanced Southern Ocean deep convection as a driver of the Antarctic warming (Menviel et al. 2015). In that hypothesis, enhanced deep convection over the continental shelves and increased Antarctic Bottom Water (AABW) production could have released heat stored in the subsurface ocean warming in regulating the Atlantic AMOC (Brady, E. C., and B. L. Otto-Bliesner, 2011: The role of meltwater-induced subsurface ocean warming in regulating the Atlantic

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**Data availability statement.** The iTRACE dataset is available at https://www.earthsystemgrid.org/dataset/ucar.egd.csm4.iTRACE.html. Data for the main figures can be found at http://doi.org/10.5281/zenodo.6019711.

**REFERENCES**


Brady, E. C., and B. L. Otto-Bliesner, 2011: The role of meltwater-induced subsurface ocean warming in regulating the Atlantic


Mignot, J., A. Ganopolski, and A. Levermann, 2007: Atlantic sub-
surface temperatures: Response to a shutdown of the over-

Minnunen, E., A. Diedenhofen, J. Flückiger, B. Stauffer, T. F. Stocker, D. Raynaud, and J.-M. Barnola, 2001: Atmospheric CO₂ concentrations over the last glacial termina-


Peltier, W. R., D. F. Argus, and R. Drummond, 2015: Space ge-


Sen Gupta, A. S., and M. H. England, 2006: Coupled ocean–atmosphere–ice response to variations in the southern annu-


Shakun, J. D., and Coauthors, 2012: Global warming preceded by increasing carbon dioxide concentrations during the last deglaciation. Nature, 484, 49–54, https://doi.org/10.1038/nature10915.


Stocker, T. F., and S. J. Johnsen, 2003: A minimum thermody-
namic model for the bipolar seesaw. Palaeogeogr. Palaeocli-


van Loon, H., and R. L. Jenne, 1972: The zonal harmonic stand-


Yang, H., Q. Li, K. Wang, Y. Sun, and D. Sun, 2015: Decompos-