Response of the Wilkes Subglacial Basin Ice Sheet to Southern Ocean Warming During Late Pleistocene Interglacials

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Article

Keywords:

Posted Date: January 12th, 2022

DOI: <https://doi.org/10.21203/rs.3.rs-1174987/v1>

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Version of Record: A version of this preprint was published at Nature Communications on September 10th, 2022. See the published version at <https://doi.org/10.1038/s41467-022-32847-3>.

Abstract

The response of the East Antarctic Ice Sheet to past intervals of oceanic and atmospheric warming is still not well constrained but critical for understanding both past and future sea-level change. Furthermore, the ice sheet in the Wilkes Subglacial Basin, which is characterized by a reverse-sloping bed, appears to have undergone thinning and ice discharge events during recent decades. By combining new glaciological evidence on ice sheet elevation from the TALDICE ice core with offshore sedimentological records and ice sheet modelling experiments, we reconstruct the ice dynamics in the Wilkes Subglacial Basin over the past 350,000 years. Our results indicate that the Wilkes Subglacial Basin experienced an extensive retreat 330,000 years ago and a more limited retreat 125,000 years ago. These changes coincided with warmer Southern Ocean temperatures and elevated global mean sea level during those interglacial periods, confirming the sensitivity of the Wilkes Subglacial Basin ice sheet to ocean warming and its potential role in sea-level change.

Introduction

The growth and decay of polar ice sheets exert important controls on regional and global climate, while their future behaviour is a key uncertainty in predicting sea-level rise during and beyond this century². Over the last decade, it has been observed that excess basal melting in Antarctica, arising from ocean heat supply, has increased the dynamic mass loss of grounded ice shelves bordering the Southern Ocean (SO)³. This observation has implications for future ice sheet stability, and also suggests that ocean warming may have played a role in controlling past ice sheet dynamics in Antarctica.

The largely marine-based West Antarctica Ice Sheet, which could contribute up to 4-5 m to Global Mean Sea Level (GMSL)⁴, is well known for its past and future vulnerability to a warming climate^{5,6}. In contrast, past stability of the much larger East Antarctic Ice Sheet (EAIS), which is characterized by a total potential contribution to GMSL of 53 m¹, of which around one-third is marine-based ice, is still under debate⁷⁻¹⁰. The Wilkes Subglacial Basin, which contains 3 to 4 m sea-level equivalent⁴, is characterized by a reverse-sloping bed with an elevation below sea level (Figure 1). Ice that is grounded below sea level is vulnerable to intrusions of warm modified Circumpolar Deep Water (CDW) across the continental shelves into ice shelf cavities^{4,11,12}. An initial grounding line retreat into deeper water may then lead to a marine ice sheet instability condition, which would be followed by increased ice discharge, inland thinning, and a rapid contribution to GMSL¹³.

Satellite altimetry and images reveal that the Cook Glacier, which drains a large proportion of the Wilkes Subglacial Basin, has experienced thinning of 33 ± 12 cm/a (year) over the past 25 years¹⁴, following near-complete loss of the Cook West Ice Shelf between 1973 and 1989 that resulted from intense oceanic warming during the middle of the 20th century^{15,16}. Model simulations suggest a modest sensitivity of the Wilkes Subglacial Basin ice sheet to oceanic warming^{17,18} and margin retreat controlled by the presence of a coastal “ice plug”¹⁹. However, projected atmospheric and oceanic warming could soon lead

to the crossing of tipping points in Antarctica, and hence the destabilization of marine-based sectors of the ice sheet^{20,21}. In light of this potential vulnerability, future predictions of the Wilkes Subglacial Basin ice dynamics should be refined by studying previous occurrences of instabilities during past warm climatic periods when temperatures were comparable to, or warmer than, modern conditions.

Here we explore the past ice dynamics of the Wilkes Subglacial Basin during the recent interglacial Marine Isotopic Stages (MIS) 5.5, 7.5, and 9.3 of the last 350 ka. These warm periods can be considered similar, in terms of atmospheric warming and GMSL increases, to a range of near-future climate projections^{20,22}. Our approach is based on interrogation of the new isotopic data ($\delta^{18}\text{O}$ and d -excess) from the TALDICE ice core at Talos Dome, which is sensitive to grounding line retreat in the Wilkes Subglacial Basin^{9,23}, in comparison to the EDC ice core record at Dome C, which is representative of East Antarctic plateau conditions under the influence of the Southern Indian Ocean (Figure 1). We compare estimated elevation changes at Talos Dome with simulations of local ice thickness variations and Wilkes Subglacial Basin ice dynamics from experiments conducted with the GRISLI ice sheet model²⁴ in order to identify past instability events. To provide a comprehensive picture of the ice sheet behaviour in the Wilkes Subglacial Basin during past warm interglacials, we also integrate our glaciological data and simulation results with late Pleistocene sedimentological and geochemical records from the marine core U1361A offshore Wilkes Subglacial Basin⁷ (Figure 1).

Our results suggest that neither changes in air mass trajectories nor variations in sea-ice extent can explain the unique TALDICE $\delta^{18}\text{O}$ signal recorded during late MIS 5.5, 7.5, and 9.3. Instead, we propose that the interglacial anomalies in the isotopic record have been produced by lowering of the site elevation at Talos Dome due to ice loss and inland retreat of the Wilkes Subglacial Basin grounding line in response to intrusion of warmer ocean waters. The GRISLI ice sheet simulation that best fits with our elevation data suggests a 10% reduction of the Wilkes Subglacial Basin ice volume during MIS 5.5, and a loss of up to 25% for MIS 9.3. Hence, these findings depict a highly dynamic ice sheet in the Wilkes Subglacial Basin and provide insights into the future response of the EAIS in a warmer world.

Results

TALDICE isotopic records over past interglacial periods

The EDC and TALDICE water isotopic records are both influenced by precipitation originating mainly from the Southern Indian Ocean²⁵. The site of EDC on the East Antarctic plateau is believed to be predominantly representative of past climatic variations at a hemispheric scale²⁶, while the TALDICE isotopic record is also sensitive to localised sea-ice extent in the Ross Sea^{27–29} and local elevation changes^{9,26,30}. During the current and last interglacial (LIG) periods, TALDICE and the plateau ice cores (EDC, Vostok, Dome F) share common isotopic maxima ($\delta^{18}\text{O}$ and δD), between 12 and 9 ka (thousands of years ago) in the Holocene and at 128 ka during the LIG³¹. The LIG isotopic maxima indicate Antarctic atmospheric temperatures 2–4.5°C warmer than the Holocene^{32–34}, probably arising from the operation of

the bipolar seesaw mechanism during deglaciation of the Northern Hemisphere ice sheets^{31,35,36}. However, during the late stage of MIS 5.5, the TALDICE isotopic record has a $\delta^{18}\text{O}$ peak at 117 ka followed by an abrupt decrease towards the glacial inception, and such features are not recorded in the cores from the plateau³¹.

Here we compare our new TALDICE $\delta^{18}\text{O}$ data (measured at 5 cm resolution) from MIS 5.5³¹, MIS 7.5 and 9.3 with the published EDC $\delta^{18}\text{O}$ record^{37,38}, which is representative of a common Antarctic signal³⁹, in order to investigate TALDICE isotopic patterns during previous late Pleistocene interglacials (Methods). Interestingly, the TALDICE $\delta^{18}\text{O}$ record from MIS 9.3 exhibits a double-peak shape that is similar to the one observed for MIS 5.5, which is not seen in EDC (Figure 2c). Both cores display a common interglacial isotopic peak at 335 ka, but from 331 ka the TALDICE signal diverges from the EDC signal, showing a sustained increase of about 1.5‰ until 326 ka followed by a steep decrease, while the EDC record declines gradually across this entire interval. During MIS 7.5, there is also some divergence of smaller magnitude between the two records (Figure 2c). The TALDICE $\delta^{18}\text{O}$ record exhibits a second late and muted rise of ~0.8‰ between 240 ka and 237 ka, rather than the well-defined late peak observed for MIS 5.5 and MIS 9.3, while the EDC $\delta^{18}\text{O}$ record decreases towards the glacial inception.

We explore three main hypotheses that could explain the discrepancies between the EDC and TALDICE water isotopic records, namely (i) differences in moisture sources, (ii) changes in sea-ice extent in the Ross Sea, and (iii) a local decrease of elevation at Talos Dome.

Differences in moisture sources between EDC and TALDICE can be explored using the deuterium excess ($d\text{-excess} = \delta\text{D} - 8 \cdot \delta^{18}\text{O}$) thanks to the new TALDICE $d\text{-excess}$ profile (5 cm resolution) for MIS 5.5, 7.5 and 9.3. This second-order parameter is an indicator of climate conditions in the vapour source regions, and is therefore sensitive to changes in the source regions or changes in air mass trajectories towards the sites^{40–42}. Our comparison shows no clear differences between the TALDICE and EDC³⁹ $d\text{-excess}$ profiles, which exhibit coherent patterns during all interglacials (Figure 2b). In particular, the absence of peculiarities in the $d\text{-excess}$ during the intervals with $\delta^{18}\text{O}$ anomalies at TALDICE suggests that there are no significant variations of moisture sources^{42,43} that could explain the TALDICE isotopic behaviour.

The hypothesis of a change in sea-ice extent can be addressed using published sea-salt sodium (ssNa^+) fluxes in TALDICE for MIS 1²⁷ and 5.5⁴⁴, and new data for MIS 7.5 and 9.3, as a proxy for sea-ice coverage in the western sector of the Ross Sea and in the southern Indian Ocean facing the Wilkes Subglacial Basin²⁷. The high-resolution (7-8 cm) TALDICE ssNa^+ flux record for MIS 5.5⁴⁴, 7.5 and 9.3 agrees well with the EDC record^{37,38} and no site-specific differences are identifiable (Figure 2a). Note that the peak in the TALDICE ssNa^+ record during MIS 7.5 at 240 ka is not interpreted as being climate-driven, but rather as artifacts from chemical weathering processes in the deep ice layers^{45,46}. The overall coherence between the EDC and TALDICE ssNa^+ records indicates that sea-ice variations in the Ross Sea or southern Indian Ocean cannot explain the double-peak shape of the $\delta^{18}\text{O}$ record at Talos Dome, and hence the second hypothesis can also be discounted.

In summary, our multi-proxy comparison (Figure 2) indicates that neither changes in air mass trajectories nor in sea-ice extent can explain the unique interglacial $\delta^{18}\text{O}$ excursions recorded at Talos Dome. In the following sections we therefore interpret the isotopic anomalies as an indication of elevation changes at this site, and estimate the magnitude of changes that are required to explain the anomalies.

Interglacial elevation changes at Talos Dome from $\delta^{18}\text{O}$ records

The relationship between $\delta^{18}\text{O}$ values and ice sheet elevation has recently been investigated to reconstruct EAIS dynamics during the Last Glacial Maximum (LGM) and LIG^{9,30,47}. To estimate the imprint of elevation changes at Talos Dome during the LIG, Sutter et al. (2020) applied a $\delta^{18}\text{O}$ -elevation relationship of $-0.53\text{‰}/100\text{ m}$, based on multiplying the present day lapse rate of $-0.8^\circ\text{C}/100\text{ m}$ by the local $\delta^{18}\text{O}$ isotope-temperature relationship of $0.66\text{‰}/^\circ\text{C}$ estimated from the atmospheric general circulation model ECHAM5-wiso equipped with an isotope module⁴⁷. In contrast, Goursaud et al.³⁰ obtained a relationship of $-0.93\text{‰}/100\text{ m}$ for the LIG, based on the simulated isotopic response to idealised changes in Antarctic ice sheet elevation in the isotope-enabled coupled ocean-atmosphere-sea-ice general circulation model HadCM3⁴⁸. As a third approach, we directly compute the isotope-elevation relationship from present-day snow-pit $\delta^{18}\text{O}$ data collected along the traverse from GV7 site to Talos Dome⁴⁹, which takes into account the provenance of air masses reaching Talos Dome (see details in Methods). This method leads to an estimate of the modern isotopic lapse rate for Talos Dome of $-1.35\text{‰}/100\text{ m}$.

We use the three different $\delta^{18}\text{O}$ lapse rate estimates for TALDICE to calculate the elevation changes required to explain the observed anomalies in the isotopic signal ($\Delta\delta^{18}\text{O}$) (see Methods) for MIS 5.5, 7.5, and 9.3 (Table 1). The periods of MIS 5.5 and MIS 9.3 were apparently subjected to the largest elevation changes, with similar magnitude for the two periods, while only around half of this elevation decrease is invoked to explain MIS 7.5 signal. Applying the isotopic lapse rates of $-1.35\text{‰}/\text{m}$ ⁴⁹ and $-0.93\text{‰}/\text{m}$ ³⁰, we obtain a decrease in elevation of about 100-200 m during MIS 5.5 and 9.3. Using the lapse rate of $-0.53\text{‰}/\text{m}$ ⁹ would imply larger elevation variations ($\sim 300\text{ m}$) for those intervals. Hence, while subject to uncertainty from the choice of lapse rate, we conclude from the TALDICE isotopic record that the Talos Dome site was subjected to elevation changes on the order of 100-300 m during MIS 5.5 and 9.3 (Table 1). We further note that, due to signal smoothing at the centennial resolution of the record in the deep portion of the ice core^{50,51}, the $\Delta\delta^{18}\text{O}$ anomaly during MIS 9.3, and consequently the ice thickness variation inferred for this interglacial, could be under-estimated.

Table 1

Isotopic anomalies and elevation changes calculated and modelled for Talos Dome during interglacial MIS 5.5, 7.5, and 9.3. Isotopic anomalies are calculated from the $\delta^{18}\text{O}$ record resampled at 200 years intervals. Elevation changes are calculated from the isotopic anomalies using the lapse rate estimates of Sutter et al.⁹, Goursaud et al.³⁰, and Magand et al.⁴⁹. IS and GS elevation changes are modelled for 6 sensitivity tests performed with the GRISLI ice sheet model, varying the Antarctic ice sheet initial conditions and the SO temperature forcing. IS simulations are initialized with Antarctic interglacial initial conditions at 400 ka and GS simulations are initialized with Antarctic glacial initial conditions at 400 ka. Elevation variations at Talos Dome are calculated for interglacial time intervals when the simulated GRISLI ice thickness variations are maximized.

	MIS 5.5	MIS 7.5	MIS 9.3
Time interval of $\delta^{18}\text{O}$ anomaly (ka) at TALDICE	117-127	237-240	326-331
$\delta^{18}\text{O}$ max and $\delta^{18}\text{O}$ min (‰)	$\{-35.45\}$ - $\{-37.13\}$	$\{-37.19\}$ - $\{-37.87\}$	$\{-35.46\}$ - $\{-36.89\}$
$\Delta\delta^{18}\text{O}$ (‰)	1.68	0.68	1.42
Lapse rate (‰/100m)	Elevation changes (m)		
-0.53 (Sutter et al. 2020)	-317	-128	-268
-0.93 (Goursaud et al. 2020)	-180	-73	-153
-1.35 (Magand et al. 2004)	-124	-50	-105
	MIS 5.5	MIS 7.5	MIS 9.3
Ocean forcing	GRISLI elevation changes at Talos Dome (m)		
Time interval for max elevation anomaly at TALDICE (ka)	115-128	233-241	321-332
IS Quiquet et al. (2018)	-132	-103	-749
IS-5 Quiquet et al. (2018) +5%	-126	-101	-720
IS-10 Quiquet et al. (2018) +10%	-750	-87	-714
GS Quiquet et al. (2018)	-126	-85	-134
GS-5 Quiquet et al. (2018) +5%	-116	-89	-473
GS-10 Quiquet et al. (2018) +10%	-152	-103	-754

Sensitivity tests with the GRISLI ice sheet model

To further explore possible elevation changes at Talos Dome, as well as Wilkes Subglacial Basin grounding line displacements, we perform numerical experiments of the Antarctic ice sheet dynamics over the last 400 ka with the GRISLI ice sheet model²⁴, using the same setup as Quiquet et al.²⁴. The model is forced by near-surface air temperatures over Antarctica deduced from the EDC δD record³² and by ocean temperatures derived from the ODP 980 benthic temperature record from the North Atlantic Ocean⁵². Due to the lack of a continuous proxy record of sub-oceanic conditions around Antarctica covering the past 400 ka, changes in the temperature of North Atlantic Deep Water are assumed to influence the temperature of upwelled waters in the SO²⁴, which affects the sub-shelf melting rate at the grounding line. This simplified approach does not account for the influence of local sea-ice changes²⁸ or freshwater capping^{36,53} on sub-surface temperatures around Antarctica, or for variability in Circumpolar Deep Water upwelling or cross-shelf transport⁵⁴. Considering the simplified SO conditions employed, we modify the original oceanic forcing to test the response of the Wilkes Subglacial Basin ice sheet to variations of the oceanic conditions.

We present 6 sensitivity experiments, varying (i) the initial state of the Antarctic ice sheet at 400 ka, and (ii) the sub-shelf melting rate. The GRISLI experiments labelled IS (Interglacial Start) adopt an Antarctic ice sheet simulated for the present day as the initial condition at 400 ka, while the GS (Glacial Start) experiments use a Last Glacial Maximum (21 ka) initial state²⁴. These initial ice sheet states are simulated as in Quiquet et al.²⁴. For both IS and GS initial states, we present three GRISLI simulations forced by different ocean sub-surface conditions (see Methods): (i) the sub-shelf melting rate signal applied by Quiquet et al.²⁴ and derived from the ODP 980 benthic record⁵², (ii) the signal defined by Quiquet et al.²⁴ increased by 5%, and (iii) the signal defined by Quiquet et al.²⁴ increased by 10% over the past 400 ka (Table 1). The GRISLI experimental results include elevation changes at Talos Dome, ice volume variations in the Wilkes Subglacial Basin, grounding line displacements, and the contribution of grounded ice above flotation to GMSL.

The ice thickness results from the GS and IS sensitivity experiments are compared to ice thickness variations calculated from the isotopic record at Talos Dome (Table 1 and Figure 3), with simulation GS-5 (i.e. Glacial Start and +5% oceanic forcing) showing the best agreement with the elevation changes deduced from the ice core record. For MIS 5.5, simulation GS-5 predicts an elevation decrease at Talos Dome on the order of 100 m (Figure 3c, Table 1), an average of ~100 km grounding line retreat into the Wilkes Subglacial Basin between 133 ka and 115 ka (Figure 4a), and loss of 10% of its ice volume (Figure 3d).

The main grounding line retreat during MIS 5.5 is seen earlier in the model (128 ka) than what is suggested by TALDICE isotopic signal (117-127 ka) (Table 1). We however do not expect a perfect synchronicity between the two series because of uncertainties in relative timescale and in the ice sheet time response. Moreover, the elevation change at Talos Dome may not be completely connected to the grounding line retreat, and could also be influenced by a change in accumulation rate during interglacials at the site.

For MIS 7.5, GS-5 predicts no grounding line displacement and only a very limited reduction in ice thickness at Talos Dome, in agreement with the TALDICE isotopic record (Figure 3b-c, Table 1). For MIS 9.3, GS-5 simulates a more dynamic ice sheet response in the Wilkes Subglacial Basin, with an average of ~330 km grounding line retreat between 339 ka and 318 ka (Figure 4b) leading to the loss of ~25% of its ice volume (Figure 3d). The modelled elevation at Talos Dome exhibits an abrupt decrease of ~470 m between 332 ka and 321 ka (Figure 3c), synchronous with the TALDICE isotopic anomaly (Figure 3b). However, the ice thickness variations simulated at Talos Dome during MIS 9.3 are much greater than those calculated from the isotopic record (~100-250 m) (Table 1). The TALDICE lower temporal resolution (~200 years/5 cm) below 1530 depth probably leads to smoothing of the isotopic signal^{50,51}, which could be responsible for a muted $\Delta\delta^{18}\text{O}$ signal during MIS 9.3. Overall, the modelled contribution from grounded ice above flotation in the Wilkes Subglacial Basin to GMSL increase is estimated to be +0.5 m during MIS 5.5 and +0.9 m during MIS 9.3 for the GS-5 experiment. The former value is supported by the changes in the TALDICE ice core isotopic record, while the latter value may be an upper estimate, depending on the extent of signal smoothing in the TALDICE record during MIS 9.3.

The results of these IS experiments depict a highly-unstable ice sheet in the Wilkes Subglacial Basin, which is inconsistent with the elevation changes deduced from the TALDICE $\delta^{18}\text{O}$ record and with evidence from previous studies^{8,10}. A ~700 m elevation reduction at Talos Dome is simulated over MIS 9.3, while a Wilkes Subglacial Basin collapse is simulated in the IS-10 experiment but not the other IS simulations for MIS 5.5 (Table 4.1). An increase of 10% in the oceanic warming forcing seems to represent a tipping point that triggers deglaciation of the Wilkes Subglacial Basin ice sheet in both IS and GS experiments during 9.3 and in IS simulations during MIS 5.5 (Table 1). We suggest that this tipping point was not passed during those intervals (Figure 3).

Comparison to the U1361A marine sediment core record

The Antarctic ice sheet loses the majority of its mass via iceberg calving and sub-ice-shelf melting⁵⁷. Past retreat of the ice shelves and ice sheet in the vicinity of the Wilkes Subglacial Basin is therefore expected to have left geochemical and sedimentological signatures in the nearby marine sediments. The sediment core U1361A (64.41°S, 143.89°E, 3,454 m water depth)⁵⁸, recovered from the continental rise adjacent to the Wilkes Subglacial Basin (figure 1), provides a near-continuous archive of Pliocene to Pleistocene variability of this marine EAIS margin^{7,59,60}. Specific events of iceberg-rafted debris (IBRD) discharge can indicate dynamic ice loss, while Nd isotopes (ϵ_{Nd}) provide a provenance indicator for changes in subglacial erosion and transport that could reflect ice sheet retreat^{7,59,60} (figure 4). To enable a direct comparison of the TALDICE isotopic record and GRISLI simulations with the U1361A record, we refined the original U1361A chronostratigraphy to achieve consistency with the AICC2012 framework (see Methods).

The prominent IBRD and Nd isotope peaks during MIS 5.5 and 9.3 are interpreted as evidence of ice loss from the margins of the Wilkes Subglacial Basin⁷. Interestingly, the IBRD peaks, which record transient

events of ice discharge and iceberg calving, and the Nd isotope maxima, representing inland erosion, coincide with the onset of the TALDICE $\delta^{18}\text{O}$ anomalies at ~ 125 ka during MIS 5.5 and at ~ 330 ka during MIS 9.3 (Figure 4b and 4e). In addition, the duration of the TALDICE isotopic anomalies are consistent with the sustained high Nd isotope values during the later stages of MIS 5.5 and MIS 9.3, suggesting a prolonged interval of inland glacial erosion and relative increase of iceberg calving. Furthermore, the highest IBRD content in the U1361A record ($\sim 7.5\%$) occurred at ~ 330 ka, indicating a significant ice loss event and potentially a more dynamic Wilkes Subglacial Basin ice sheet during MIS 9.3 in comparison to MIS 5.5, for which the IBRD peak was less pronounced ($\sim 2\%$) (Figure 4e). A more dynamic behaviour during MIS 9.3 than MIS 5.5 is also supported by the GRISLI modelling results (Figure 4c-d).

In contrast, the interglacial periods of MIS 1 and MIS 7.5 are characterized by only minor IBRD occurrences and more muted Nd isotope maxima in core U1361A (Figure 4e), indicating a more stable ice sheet. Similarly, the TALDICE isotopic anomaly is less pronounced at these times (Figure 4b), and the GRISLI simulations also indicate a muted response in the Wilkes Subglacial Basin during those interglacials (Table 1). As such, evidence from these independent datasets and approaches appears to converge on a consistent picture of the differential response of the ice sheet in the Wilkes Subglacial Basin to the subtly different climate forcing of individual late Pleistocene interglacials.

Discussion

By combining the TALDICE $\delta^{18}\text{O}$ record with GRISLI ice sheet model outputs and records from marine sediment core U1361A, we propose that the TALDICE isotopic anomalies during late Pleistocene interglacials reflect a reduction in elevation at Talos Dome arising from accumulation rate changes, ice loss, and grounding line retreat in the Wilkes Subglacial Basin. Our results from the GRISLI GS-5 simulation suggest that even a small increase of the SO sub-shelf melting rate of 5%, which we attribute to the intrusion of warm water at the grounding line depth was enough to trigger significant margin retreat (although not complete collapse) of the ice sheet in the Wilkes Subglacial Basin during the warmest late Pleistocene interglacials. Notably, our simulated Talos Dome elevation changes during MIS 5.5 are consistent with recent studies^{9,10}, while GRISLI GS-5 depicts a larger retreat at the margin of the Wilkes Subglacial Basin in closer accordance with the U1361A record⁷.

Since the SO Polar Front and Subantarctic Front probably shifted polewards during MIS 5.5, the more southerly and warmer Antarctic Circumpolar Current could have induced a dynamic response in the EAIS sectors with reverse-sloping subglacial basins^{61,62}. A comparable scenario can also be anticipated for MIS 9.3. Interestingly, a new late Pleistocene Subantarctic sea surface temperature reconstruction from the Indian sector of the SO (core DCR-1PC) shows that both MIS 5.5 and MIS 9.3 were characterized by a double warming phase⁵⁶ (Figure 4f). The first warming phase corresponded to the interglacial optimum, which was followed by a cooling phase attributed to feedbacks from the Antarctic Ice Sheet, and then a second late warming phase before the glacial inception⁵⁶ (Figure 4f). Over time, the two long warming phases probably contributed to the destabilization of the ice shelves and outlet glaciers of the Wilkes Subglacial Basin, leading to inland retreat of the grounding line. Our evidence also points to a Wilkes

Subglacial Basin ice sheet susceptible to local SO warming and/or CDW intrusion during MIS 9.3⁵⁶, which suggest that this ice sheet remained in a relatively unstable state following the proposed ice sheet collapse during MIS 11^{7,8}.

Finally, sea-level reconstructions for MIS 5.5 and MIS 9.3 are also consistent with our reconstruction of the Wilkes Subglacial Basin ice dynamics during the late Pleistocene. Data from corals and other sea-level proxies indicate a late peak in GMSL during MIS 5.5 at ~119 ka of ~6 to 9 m above present^{22,63}. For MIS 9, a stacked GMSL reconstruction based on various proxies indicates a sea-level highstand of ~9 m above modern⁶⁴. The GS-5 experiment indicates that the grounding line in the Wilkes Subglacial Basin may have retreated by several hundred kilometres during those interglacials, leading to contributions of approximately +0.5 m and +0.9 m to the GMSL increases during MIS 5.5 and MIS 9.3, respectively (Methods). These instabilities of the Wilkes Subglacial Basin ice sheet appear to have been driven mainly by increases in SO temperatures and/or CDW intrusion during past interglacials. Despite the qualitative agreement between the ice core data, marine sediment core proxies, and the ice sheet modelling results, it will be essential to better assess quantitative differences between the data and models, as well as the relative roles of atmospheric and oceanic warming in triggering ice sheet instabilities¹⁷, in order to improve our understanding of EAIS dynamics during past interglacial intervals and on a future warming Earth.

Declarations

Acknowledgements

This work is part of the TALDEEP project funded by MIUR (PNRA18_00098) and Talos Dome Ice core Project (TALDICE), a joint European program, funded by national contributions from Italy, France, Germany, Switzerland and the United Kingdom. Primary logistical support was provided by PNRA at Talos Dome. This is TALDICE publication no 61. The research leading to these results has also received funding from the European Research Council under the European Union H2020 Programme (H2020/20192024)/ERC grant agreement no. 817493 ERC ICORDA.

I. Crotti acknowledges funding from Università Italo-Francese/Université Franco-Italienne and the Vinci Scholarship for co-tutorship Ph.D. program. A. Quiquet acknowledges funding from the SCOR foundation project COASTRISK. D. J. Wilson is supported by a NERC independent research fellowship (NE/T011440/1).

Author contributions

I. Crotti performed the isotopic analysis, processed the experimental data, designed the main conceptual ideas, wrote the manuscript, and designed the figures. A. Quiquet performed the model experiments, contributed to the manuscript, and designed the figures. A. Landais and B. Stenni and M. Frezzotti contributed to the interpretation of the results and revised the manuscript. D.J. Wilson refined the

sediment core age model, contributed to the interpretation of the results, and revised the manuscript. M. Severi, F. Wilhelms and R. Mulvaney performed the chemistry analysis and contributed to the interpretation of the results. C. Barbante revised the manuscript.

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Methods

TALDICE $\delta^{18}\text{O}$ and δD records

In this study we focus on the oxygen and hydrogen isotopic composition ($\delta^{18}\text{O}$) and d -excess ($d = \delta\text{D} - 8 \times \delta^{18}\text{O}$) new and published profiles measured at high resolution (5 cm) in the TALDICE ice core during interglacials. Here we present the published $\delta^{18}\text{O}$ bag resolution profile (1 m) for the Holocene³⁵, the published profile for MIS 5.5³¹ and new 5 cm resolution profiles MIS 7.5, and MIS 9.3. For the d -excess, we show the published bag resolution signal (1 m) for MIS 1²⁷ and the new 5cm resolution signal for the oldest interglacials. We draw data from MIS 1 (0-20 ka) and MIS 5.5 (~115-132 ka) on the AICC2012 age scale⁵⁵. The new data for $\delta^{18}\text{O}$ and d -excess during MIS 7.5 (~240-246 ka) and MIS 9.3 (~324-339 ka) are plotted on the TALDICE-deep1 age scale⁵⁰.

The 5 cm samples were analysed in Italy (University of Venice) and France (LSCE) using the Cavity Ring Down Spectroscopy (CRDS) technique. Analyses were performed using a Picarro isotope water analyser (L2130-i version for both laboratories). The data were calibrated using a three-point linear calibration with three lab-standards that were themselves calibrated versus Standard Mean Ocean Water (SMOW). Intercomparisons between the two laboratories have been performed over the analysis period. The average precision for the $\delta^{18}\text{O}$ and δD measurements is 0.1 and 0.7 ‰, respectively.

In this work we compare $\delta^{18}\text{O}$ and d -excess records from TALDICE and EDC cores³⁹, which are characterized by different temporal resolution, mostly due to differences in the sampling interval, snow accumulation and thinning function at different depth (Table 2).

		MIS 1 (0-20 ka)	MIS 5.5 (110-140 ka)	MIS 7.5 (234-252 ka)	MIS 9.3 (320-345 ka)
TALDICE	Sample length (cm)	1 m	5 cm	5 cm	5 cm
	Resolution (years/5 cm)	-	32	95	192
	Resolution (years/m)	24	630	1891	3843
EDC	Sample length (cm)	55 cm	55 cm	55 cm	11 cm
	Resolution (years/sample length - cm-)	22	53	116	33
	Resolution (years/m)	40	96	212	301

Table 2: Comparison of temporal resolution for TALDICE and EDC $\delta^{18}\text{O}$ samples during interglacial periods. We calculated the sample resolution based on the original sample length (5 cm for TALDICE, 55 cm or 11 cm for EDC). We also calculated the average temporal resolution for each metre of the core.

Identification of change-points and calculation of anomalies

To identify differences between the TALDICE and EDC isotopic records, we assessed changes in slope and intercept in the data sets. We first normalized both records (data centred and scaled to have mean 0 and standard deviation 1) and resampled them at the same time step of 200 a. We then searched for changes using the MATLAB task *find change points* with a maximum of 3 or 4 change points⁶⁵. The software identifies points with changes in slope (derivative) and intercept (figure 5). Isotopic anomalies were then calculated as the differences between values in the resampled record between the two change points (Table 1).

Lapse rate calculation for TALDICE

In the modern day, Talos Dome receives 50% of its total precipitation from the west (Southern Indian Ocean), 30% from the east (Ross Sea and Southern Pacific Ocean), and approximately 15% from the interior²⁵. In contrast, at EDC the modern precipitation is mainly from the western Southern Indian Ocean (85%), with a small amount coming from the east via the Ross Sea and Transantarctic Mountains (15%).

The isotopic lapse rate expresses the variation in the oxygen isotopic composition ($\delta^{18}\text{O}$) in permil (‰) for every 100 m change in altitude. Here we calculate the isotopic lapse rate for TALDICE using the ITASE traverse dataset⁴⁹. We take into account the $\delta^{18}\text{O}$ values of snow between GV7 and Talos Dome sites sampled every 5 km and the altitude profile from GV7 site to Talos Dome. This approach seems reasonable, given that the air masses travelling to Talos Dome mainly originate from the Southern Indian Ocean at 60°S and likely follow a similar path to the traverse once they reach Antarctica²⁵. The calculated isotopic lapse rate is -1.35‰/100 m.

TALDICE and EDC ssNa⁺ fluxes

In this study we show the sea-salt sodium (ssNa⁺) flux measured in the TALDICE ice core during MIS 1 (1 m resolution)²⁷, MIS 5.5 (1 cm resolution)⁴⁴, and new data for MIS 7.5 and MIS 9.3 (8 cm resolution). The concentrations of ssNa⁺ were measured by classical ion chromatography on discrete samples collected using a melting device connected to an auto-sampler for the MIS 1, MIS 7.5 and MIS 9.3 whereas Continuous Flow Analysis (CFA) was applied for MIS 5.5 samples⁴⁴.

The total deposition ssNa⁺ flux was calculated multiplying the measured ice concentration of ssNa⁺ by the reconstructed accumulation rate. The accumulation rates were derived from the AICC2012 age scale⁵⁵ for the upper part of TALDICE (until 1438 m depth), while for the deepest part (1438-1578 m) the accumulation rates were obtained from the TALDICE-deep1 age scale⁵⁰.

Refined age model for core U1361A

To compare the sedimentological and geochemical records from core U1361A with the TALDICE ice core isotopic records, we needed to define a common chronostratigraphy. The original chronology for the U1361 composite splice (U1361A and U1361B) was based on biostratigraphic and magnetostratigraphic data and one radiocarbon date^{7,58,66}. The TALDICE chronology is based on the AICC2012 age scale until ~150 ka⁵⁵, built with a multi-site approach including both Greenland and Antarctic ice cores, while for the older part it is based on TALDICE-deep1 chronology⁵⁰.

Here we use the AICC2012 ice core chronology⁵⁵ as a reference curve in order to compare the late Pleistocene sediment core data from U1361A to the TALDICE ice core record. Specifically, we refine the existing U1361A age model through the alignment of barium/aluminium (Ba/Al) ratios from XRF-scanning⁷ with the EDC δD record on the AICC2012 age scale. Glacial-interglacial cycles are clearly expressed in the U1361A Ba/Al record, with higher ratios reflecting warmer conditions and reduced sea-ice extent during interglacials⁷. As such, and similar to the EDC δD record, the Ba/Al record reflects a combination of local forcing and global climate boundary conditions. We apply a conservative tuning strategy to align the two records, using tie points (derived by visual matching) only at the mid-points of the major glacial terminations I-V, where large and rapid signals are recorded in both U1361A Ba/Al and EDC δD records. Such transitions are likely to be synchronous in the two records, at least at the multi-ka sample resolution of the U1361A marine record, whereas there is greater uncertainty in the exact timing and progression of productivity changes during the steps from interglacial to glacial periods.

Over the interval of interest for the present study (i.e. ~100-350 ka), the new age model for core U1361 differs by only 0 to 6 ka compared to the previous age model in which the sedimentation rate was assumed to be constant⁷. The temporal resolution of the U1361A record is on the order of several ka for the time period from 0 ka to 345 ka. Specifically, the mean resolution of the record is 5.9 ka for IBRD, 6.5 ka for Nd isotopes, and 3.4 ka for Ba/Al ratios.

GRISLI ice sheet model

The GRISLI (Grenoble ice sheet and land ice) model is a large-scale three-dimensional thermomechanical ice sheet model. In this work, we use the GRISLI version 2.0²⁴, which is the latest revision, to model changes of the Antarctic ice sheet between 100 ka and 400 ka, since it is mostly designed for multi-millennial integrations. The model combines an inland ice model with an ice shelf model, extended to the case of ice streams considered as dragging ice shelves⁶⁷. The latest release includes a better representation of grounding line migration and a sub-glacial hydrology model. The model uses finite differences on a Cartesian grid at 5 to 40 km resolution depending on the application. Here we use a 40 km grid to take advantage of the model calibration performed in Quiquet et al. (2018)²⁴. Given its low numerical cost at this resolution, Quiquet et al. (2018)²⁴ performed an ensemble of 600 simulations to calibrate the mechanical parameters. We use the ensemble member of Tsai et al. (2015)⁶⁸ for the formulation of the flux at the grounding line, since it best reproduces the glacial-interglacial transitions. This ensemble member is labelled AN40T213 in Quiquet et al. (2018)²⁴. Note that, due to the relatively coarse resolution (40 km) applied here, fine-scale structures such as individual ice streams might not be properly represented. The model setup used to perform the transient paleo ice sheet simulations is identical to that used by Quiquet et al. (2018)²⁴.

Because there are some slow feedbacks that affect the Antarctic ice sheet (e.g. glacial isostasy, internal temperature), the simulated state of the ice sheet for a given interglacial largely depends on the preceding glacial cycles. For this reason, we performed two different families of sensitivity tests in which we changed the initial state of the Antarctic ice sheet (i.e. glacial or interglacial) at 400 ka. The oceanic forcing is the main driver for glacial-interglacial grounding line migration. However, since glacial-interglacial ocean temperature changes around Antarctica are poorly constrained, this forcing represents an important source of uncertainty for the ice sheet evolution. To account for past changes in oceanic forcing, Quiquet et al. (2018)²⁴ used a paleo-index to modify the present-day sub-shelf melt rates. This index is based on a record of deep water temperature in the North Atlantic Ocean⁵², which is assumed to correlate with temperature variations of the waters upwelling in the Southern Ocean, and is therefore applied as a proxy for past Antarctic sub-shelf melt rate changes. We scaled the temperature change so that sub-shelf melting is virtually suppressed during the Last Glacial Maximum (100% reduction but with a minimum melt rate of 1cm/year) but remains unchanged for the present-day. Given that the oceanic forcing represents a large source of uncertainty, we perform sensitivity simulations with the oceanic forcing increased by 5% and by 10% (Figure 6).

In parallel to ice thicknesses and elevation calculations at Talos Dome, we calculate the ice volume evolution of the Wilkes Subglacial Basin over the past 400 ka from the GRISLI IS and GS simulation results. The drainage basin boundaries are from IMBIE2, extended offshore for the ISMIP6 project⁶⁹.

In Figure 7 we show the Wilkes Subglacial Basin ice volume changes and the Talos Dome elevation variations over the past 350 ka for the GRISLI IS simulations. The IS simulations have been disregarded due to the modelled collapse of the Wilkes Subglacial Basin ice sheet during MIS 9.3, which is inconsistent with the TALDICE record. Hence, only the GS simulations are shown in the main paper. We also compute the contribution of the Wilkes Subglacial Basin grounded ice above flotation to GMSL increase during past interglacials for all the GS and IS experiments (Figure 8), which was not shown in the main text.

Data availability

The new data set produced in this study includes $\delta^{18}\text{O}$ data (5 cm resolution for MIS 7.5 and MIS 9.3), the d-excess profiles (5 cm resolution for MIS), ssNa^+ fluxes record at 7-8 cm resolution for MIS 5.5, MIS 7.5 and MIS 9.3) of the TALDICE ice core, the refined U1361A age model and the GRISLI simulations results will be available at the Pangaea repository (<https://www.pangaea.de/>) upon acceptance of the paper.

Figures

Figure 1

Location of the Antarctic ice cores main drilling sites and the U1361A marine sediment core. The map shows the subglacial bedrock elevation above sea level (m) and the Antarctic ice sheet present-day surface elevation above sea level (m)¹. The main ice core drilling sites are Talos Dome (TD), Dome C (DC), Vostok (VK) and the marine sediment core U1361A are indicated with blue dots. The studied area of the Wilkes Subglacial Basin is delimited by the red dashed contour.

Figure 2

Compilation of TALDICE and EDC sea-salt sodium fluxes, d-excess, and $\delta^{18}\text{O}$ records over the last four interglacial periods. For all the proxies, the resampled records at 200 years (blue curves for EDC and red curves for TALDICE) are superimposed on the raw signals (light blue curves for EDC and pink for TALDICE). a. EDC^{37,38} and TALDICE^{27,44} ssNa^+ fluxes on logarithmic scale. TALDICE ssNa^+ fluxes for MIS 7.5 and 9.3 are from this study. b. EDC³⁹ and TALDICE²⁷ d-excess records. TALDICE d-excess profiles for MIS 5.5, 7.5 and 9.3 are from this study. c. EDC^{37–39} and TALDICE^{31,35} $\delta^{18}\text{O}$ records. The TALDICE $\delta^{18}\text{O}$ data for MIS 7.5 and 9.3 are from this study. The anomalies in the TALDICE $\delta^{18}\text{O}$ record are identified by grey shaded bars (marking the start and end of the anomalous increase in $\delta^{18}\text{O}$ values).

Figure 3

Comparison of the TALDICE isotopic record and GRISLI simulations to sedimentological data from sediment core U1361A and Antarctic atmospheric and oceanic temperature records since 350 ka. a. Antarctic ice core temperature difference (ΔT , difference from mean values of the last millennium) derived from δD at EDC³² plotted on the AICC2012 age scale⁵⁵. b. TALDICE normalized $\delta^{18}O$ record (data centred and scaled to have mean 0 and standard deviation 1) (measured, pink; resampled at 200 years intervals, red). Holocene data are from Stenni et al.³⁵ and MIS 5.5 record is from Masson-Delmotte et al.³¹. The MIS 7.5 and MIS 9.3 isotopic data are from this study. c. Talos Dome (TD) and Dome C (DC) elevation from GS simulations, based on the original oceanic forcing used by Quiquet et al. (2018) (blue curve), and the

original forcing increased by 5% (yellow) and by 10% (orange). d. Wilkes Subglacial Basin ice volume evolution from GS simulations. e. Core U1361A Nd isotope record (plotted as ϵ_{Nd} ; green curve with dots) and iceberg-rafted debris (IBRD) % (brown curve with dots)⁷ on AICC2012 age scale (see Methods). f. Diatom-based sea-surface temperatures (SST) from core DCR-1PC (blue curve), and smoothed data with 5-point moving average (red curve)⁵⁶. Grey bars highlight the intervals with a unique isotopic signal in the TALDICE ice core compared to the EDC record.

Figure 4

Interglacial grounding line retreat of the Wilkes Subglacial Basin ice sheet simulated by the GRISLI GS-5 experiment at three different time intervals at 40 km resolution. a. Grounding line displacement during MIS 5.5 between and 133 ka and 115 ka. b. Grounding line displacement during MIS 9.3 between and 339 ka and 318 ka. The map shows the subglacial bedrock elevation above sea level (m), the Antarctic ice sheet present-day surface elevation above sea level (m) and the present day grounding line position¹.

Figure 5

Identification of change points and anomalies in the TALDICE ice core in comparison to EDC. Original TALDICE and EDC $\delta^{18}\text{O}$ records were normalized (pink and light blue, respectively) and then re-sampled at a 200 years time step (red and dark blue, respectively). Change points in the TALDICE record (grey bars) are identified where the record shows a change in slope (dashed black lines). Anomalies are calculated based on the differences in values in the resampled record between the two change points.

Figure 6

Oceanic forcing prescribed in our GRISLI simulations over the past 400 ka. This forcing is used as a percentage change from the present-day sub-shelf melt rate. The original forcing is derived from the North Atlantic benthic temperature reconstruction of Waelbroeck et al. (2002) (blue curve), with the other versions increased by 5% (yellow curve) and by 10% (red curve).

Figure 7

GRISLI IS simulations during the past 350 ka. a. Wilkes Subglacial Basin ice volume evolution. b. Talos Dome elevation. We applied the original oceanic forcing from Quiquet et al. (2018) (blue curve), and the original forcing increased by 5% (yellow) and 10% (red).

Figure 8

Contribution of Wilkes Subglacial Basin grounded ice above flotation to GMSL during the past 350 ka, simulated by the GRISLI model. a. GS experiment. b. IS experiment. We applied the original oceanic forcing from Quiquet et al. (2018) (blue curve), and the original forcing increased by 5% (yellow) and by 10% (red).