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Response of the East Antarctic Ice Sheet to Past and Future Climate Change

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Preface: The East Antarctic Ice Sheet (EAIS) contains the vast majority of Earth's glacier ice (~52 metres sea-level equivalent), but is often viewed as less vulnerable to global warming than the West Antarctic or Greenland ice sheets. However, some regions of the EAIS have lost mass over recent decades, prompting the need to re-evaluate its sensitivity to climate change. Here we review the EAIS's response to past warm periods, synthesise current observations of change, and evaluate future projections. Some marine-based catchments that underwent significant mass loss during past warm periods are currently losing mass, but most projections indicate increased accumulation across the EAIS over the 21st Century, keeping the ice sheet broadly in balance. Beyond 2100, high emissions scenarios generate increased ice discharge and potentially several metres of sea-level rise within just a few centuries, but substantial mass loss could be averted if the Paris Agreement to limit warming below 2°C is satisfied.

1. Introduction

Over recent decades, ice loss from Antarctica has exceeded mass gains and its contribution to sea-level rise has accelerated[1-9]. The largest imbalances are found in the West Antarctic Ice Sheet (WAIS: Fig. 1d), which holds 5.3 m sea-level equivalent (SLE)[10] and lost over 2,000 Gt of ice between 1992 and 2017, adding ~6 mm to global mean sea level[1]. This imbalance is attributed to warm ocean currents - modified Circumpolar Deep Water (CDW) - melting the underside of floating ice shelves, causing marginal ice thinning, grounding line retreat and increased ice discharge[11-18]. Furthermore, the WAIS is marine-based, resting on topography below sea level that deepens inland (Fig. 1e)[10]. In the absence of buttressing ice shelves[16], retreat could rapidly propagate inland via a feedback known as 'Marine Ice Sheet Instability'[19].

The vulnerability of the WAIS was first recognised in the 1970s[20], prompting a huge growth in research[21]. In comparison, much less work has focussed on the vulnerability of the East Antarctic Ice Sheet (EAIS), which is an order of magnitude larger (52.2 m SLE)[10] and generally viewed as less sensitive to ocean-climate warming. This view stems from the fact that large parts of the EAIS have persisted for millions of years[22], recent mass balance estimates tend towards equilibrium or show modest mass gains[1,2,5,23,24] (Fig. 2), and most model projections show low sensitivity to climate change over the next century[25]. However, some recent observations suggest the EAIS may be more sensitive than previously thought. Although uncertainties are large and often obscure even the sign of any change, the latest efforts to reconcile its mass balance from multiple methods[1,2] have raised the possibility of overall mass loss since ~2014 (Fig. 2). Furthermore, numerous studies[4,5,6,8], including those reporting overall mass gain[7,9,23,24], detect clear signals of regional mass loss from some marine-based catchments (e.g. Wilkes Land: Fig. 1e). Like the WAIS, mass loss has been attributed to modified CDW proximal to major outlet glaciers[4,26-30] that may also be susceptible to Marine Ice Sheet Instability[10,31], driving ice sheet thinning[5,7,11,32], grounding line retreat[17,33-35], and the retreat[30,36] and disintegration[37,38] of floating ice tongues/shelves.

Of further concern is that marine-based sectors of the EAIS lost mass during past warm periods[39-42] and some numerical modelling predicts significant sea-level contributions from them over the coming centuries[31,43-46]. In contrast, 'terrestrial' regions of the EAIS, grounded on land well above sea level, are gaining mass through increased accumulation (e.g. Dronning Maud Land: Fig. 1e), albeit with large inter-annual variability[5-9,47,48]. A key issue is that observational time-series of accumulation or ice discharge are generally too short to elucidate whether recent trends are significant or represent natural variability in the ocean-climate system[49-51], prompting a question of huge scientific and societal importance: what will happen to the East Antarctic Ice Sheet? With an emphasis on marine-based versus terrestrial sectors, we address this question by reviewing how the EAIS changed during past warm periods and during deglaciation from the Last Glacial Maximum; synthesise current observations of change; and then evaluate future projections through to 2500.

2. Response to Past Warm Periods

Since widespread glaciation of Antarctica at the Eocene-Oligocene Transition[52] (34–33.5 Ma: Fig. 1g), climatic changes have caused substantial advance and retreat of the EAIS[53,54]. Early to Mid-Miocene (24–14 Ma) sediment records in the Ross Sea, for example, provide evidence for multiple orbitally-paced fluctuations in EAIS extent[55,56], recorded by erosional hiatuses representing expansion[57], and sediment provenance and vegetation changes indicating parts of East Antarctica were ice-free[58,59].

The largest reduction in EAIS volume during the past 20 million years occurred during the Mid-Miocene Climatic Optimum (17.0–14.8 Ma), when average atmospheric CO_2 concentrations were around 600–800 ppm (ranging from 300–1400 ppm)[60] and sea surface temperatures peaked at ~11–17°C off the Adélie Coast[61] and ~6–10 °C in the Ross Sea[56]. Under these conditions, ice sheet modelling[57] can simulate tens of metres of SLE contribution from East Antarctica, with mass loss focussed in the three main subglacial basins: the Aurora (ASB), Wilkes (WSB) and Recovery (RSB) (Fig. 1a). Terrestrial sectors are also likely to have retreated, but recent sediment provenance analysis from the central Ross Sea[62] reveals that far-field sea-level records[63] can be reconciled without substantial loss of terrestrial ice in East Antarctica, consistent with coupled climate and ice sheet modeling[57]. Notably, average Mid-Miocene CO_2 concentrations could be reached by 2100[64]; although orbital forcing was stronger than present and global atmospheric temperatures (7–8°C)[65] were significantly warmer than projected for 2100.

The most recent epoch when atmospheric CO_2 concentrations last exceeded 400 ppm was the Pliocene (5.33–2.58 Ma)[66]. Mid-Pliocene (~3.3–3.0 Ma) atmospheric temperatures were ~2–4°C warmer than present[67] and global mean sea level was around 10–25 m higher[68-71]. Given the combined volume of the WAIS and Greenland Ice Sheet (~12 m), and that their mid-Pliocene minima were likely asynchronous[72], most sea-level estimates require an EAIS contribution[21,54,71].

Early work on the EAIS response to mid-Pliocene warming focussed on marine diatoms in subglacial diamictites of the Sirius Group in the Transantarctic Mountains[22,73]. Ambiguity regarding the transport mechanism of these diatoms made it difficult to locate which regions lost mass, but recent work[74] supports at least partial retreat of both the ASB and WSB. Marine sediment records provide more direct evidence of substantial retreat of the WSB, inferred from a shift in the provenance of fine-grained detrital sediment[39], contemporaneous with a shift in marine productivity, indicating a reduction in local sea-ice coverage[75] and a southward migration of the Southern Ocean Polar Front[76]. Substantial retreat of the ASB is also inferred from records of ice-rafted debris[40,77], and from erosional signatures beneath the catchment of Totten Glacier, which could have contributed over 2 m SLE[78]. Elsewhere, evidence of elevated fjord temperatures and vegetated landscapes in the Transantarctic Mountains suggests significant retreat of marine-terminating glaciers[79]; and the Lambert Glacier-Amery Ice Shelf system was highly sensitive to Southern Ocean warming[80].

In contrast, mid-Pliocene retreat of terrestrial sectors of the EAIS and/or the RSB is largely unknown, due to a lack of empirical evidence. Some ice sheet modelling is able to simulate retreat and thinning of the RSB[44,46,81], alongside the WSB and ASB (Fig. 1b), but it has generally proved challenging to simulate mid-Pliocene retreat of marine-based sectors[54,82]. The amount of modelled retreat is sensitive to assumed pre-Pliocene ice sheet configurations[81], climate model forcing[83], and ice sheet model parameters[84], with those simulating the most retreat (e.g. Fig. 1b) often requiring additional processes to enhance mass loss[44,54,81,85], some of which are debated (e.g. Marine Ice-Cliff Instability, discussed below)[86].

Marine-based sectors were also vulnerable during the warmest interglacials of the Pleistocene (2.58–0.017 Ma), with offshore evidence of mass loss in the WSB[41]. During Marine Isotope Stage 11 (~400 ka), subglacial precipitates of opal and calcite[42] suggest the ice margin was ~700 km inland of its current position, potentially contributing ~3–4 m SLE when global atmospheric temperatures were only 1–2°C warmer than present. Terrestrial records from the Transantarctic Mountains[87] also indicate ice surface elevation fluctuations of hundreds of metres during the Pleistocene, similar in magnitude to those during the Pliocene.

Mass loss during the last interglacial (Marine Isotope Stage 5: ~130–115 ka) is equivocal[21]. Ice cores and glacio-isostatic adjustment modelling[88] suggest ice-surface lowering is plausible in the WSB and ASB, with other work[89] placing an upper sea-level contribution of 0.4–0.8 m from the WSB. Sea-level records[71] require no more than a few metres from Antarctica, which is more likely from the WAIS, but the EAIS cannot currently be ruled out.

3. The Last Deglaciation

During the Last Glacial Maximum (27–20 ka), marine-based sectors of the EAIS expanded to near the continental shelf edge[90] (Fig. 1c). Evidence of ice margin extent and subsequent retreat is only available from some regions, but typically indicates deglaciation commencing at ~19–18 ka (e.g. the Lambert-Amery system), with grounding lines reaching the mid-continental shelf in some locations at ~15 ka (e.g. Ross Sea sector)[90-92]. The maximum extent in the Weddell Sea sector is less clear[93], with recent evidence[94] of an oscillating grounding line position on the outer continental shelf until ~12 ka. A rapid rise in global sea level occurred at ~14.6 ka (Meltwater Pulse 1a), but Antarctica's contribution was limited (≤1.3 m)[95]; with little direct evidence (e.g. ice surface elevation changes) for substantial changes in EAIS geometry[90,92,93], although increased ice-rafted debris is recorded in the vicinity of the Weddell Sea[96].

Most regions of the EAIS had retreated prior to the Holocene (~11.7 ka), but some experienced a delayed response (e.g. Adélie Basin, Mac. Robertson Land, Prydz Bay)[90], or even slightly thickened and advanced during deglaciation (e.g. Transantarctic Mountains)[97,98]. Furthermore, data from the Lambert-Amery system and Transantarctic Mountains indicate most ice surface lowering occurred during the Early-Mid Holocene, continuing in some locations into the last few millennia[99-101].

Bed topography influenced the retreat of marine-based sectors across the inner-continental shelf[91,102,103] and may explain some regional asynchronicity. Geomorphological evidence on the sea-floor of the Mertz Trough, Prydz Channel and western Ross Sea, for example, indicates marked accelerations in grounding line retreat across over-deepened basins[90,91,104]. Rapid retreat across over-deepenings in the Ross Sea was also associated with hundreds of metres of ice surface lowering over several centuries[99,101]. In contrast, retreat was slowed by elevated bed topography[102]. Isostatic rebound in the Weddell and Ross Seas may also have exerted a stabilising effect[105], but this process has not been explored elsewhere in East Antarctica.

Ice sheet modelling indicates the greatest ice losses occurred in the major marine embayments during deglaciation (e.g. Ross Sea, Weddell Sea and, to a lesser degree, Prydz Bay), with mass loss elsewhere varying in both magnitude and timing depending on the model[106-108]. Initial retreat was likely triggered by a combination of ocean, atmosphere and sea-level changes[21]. In at least the Ross Sea, atmospheric conditions controlled the timing and spatial pattern of early deglaciation[109]. Meanwhile, terrestrial sectors of the interior EAIS, and areas of the Transantarctic Mountains, likely thickened due to increased snowfall following the Last Glacial Maximum[97,106]. Oceanic warming became an increasingly important control on marine-based retreat during deglaciation[90], driving a positive feedback whereby ice loss freshened surface ocean waters, possibly reducing the formation of dense Antarctic bottom waters, and facilitating incursions of modified CDW onto the continental shelf[108]. Holocene ice loss in the Ross Sea, for example, corresponds with ocean warming and the development of ice shelf cavities and modified CDW intrusion[102,109].

Palaeo-data of grounding line retreat and ice sheet thinning during deglaciation provide important context for modern-day observations (Fig. 3). The highest rates of grounding line retreat exceeded 100 m yr^{-1}, possibly up to 800 m yr^{-1} (Fig. 3a). Deglacial thinning rates from the flanks of outlet glaciers were typically 0.06–0.76 m yr^{-1}, possibly up to several metres per year (Fig. 3c). These time-averaged rates were sustained over several centuries and may mask more extreme rates, but reveal that some present-day thinning (>1 m yr^{-1}) and retreat rates (>200 m yr^{-1}) are comparable to the last deglaciation.

4. Recent Ocean Conditions and Ice Dynamics

Evidence and modelling from past warm periods clearly points to the enhanced sensitivity of East Antarctica's three major marine basins (RSB, WSB, ASB). Terrestrial regions respond mainly to atmospheric forcing, but these marine-based sectors respond to both atmospheric and oceanic processes, potentially involving Marine Ice Sheet Instability. Hence, ocean conditions and bathymetry on the continental shelf are important to understand in relation to EAIS dynamics.

Around most of East Antarctica, strong easterly winds drive onshore Ekman transport of cold fresh Antarctic Surface Water, yielding a 'fresh shelf' regime (Fig. 4a). This wind-driven flow piles up cold fresh water over the continental shelf, inducing a down-welling circulation and, via geostrophic adjustment, a strong Antarctic Slope Current[110]. In these locations, weak cross-slope exchange across the Antarctic Slope Current yields a strong front separating cold fresh waters from warm salty CDW offshore (Fig. 4c), limiting CDW intrusion. Elsewhere, a 'dense shelf' regime prevails in the Ross Sea, Adélie Coast, and around Prydz Bay (Fig. 4d). Here, the overflow of dense shelf water (also referred to as High Salinity Shelf Water in some sectors) is balanced in part by onshore CDW transport, although strong water-mass transformation over the shelf cools these regions during winter[111]. Poleward Ekman transport of cold fresh surface water still occurs, but the Antarctic Slope Current is weaker and less of a barrier to cross-shelf exchange. Finally, along the coast of Wilkes Land a limited 'warm shelf' regime exists (Fig. 4e), where weaker easterly winds enable modified CDW intrusions closer to the ice margin. Recent evidence of a localised warm shelf regime close to Shirase Glacier (Fig. 1d), Dronning Maud Land, has also been detected[112], again enabled by weaker polar easterlies.

Observations of shelf-water temperature trends are extremely sparse around East Antarctica, with few multi-decadal measurements available (e.g. in the Ross Sea)[113,114]. Evidence for long-term shelf-water warming is therefore limited, but warm waters have been detected close to several major outlet glaciers[26-29,112,115], coinciding with high basal melt rates beneath ice shelves (Fig. 4b)[12-15,116]. This can lead to ice shelf thinning and reduced buttressing, increasing ice discharge[18]. Warm water entering ice shelf cavities can also form basal channels, causing localised incision and further structural weakening[117], including transverse fractures associated with calving[118].

One region where ocean forcing is impacting ice dynamics is Wilkes Land, overlying the ASB and referred to as East Antarctica's 'weak underbelly'[30,119]. A signal of mass loss has emerged over the last three decades[3-9,23], with one study[9] suggesting mass loss (-51 ±80 Gt yr^{-1}: 2016–2020) may be ten times higher than a decade ago. Observations[26-28] have confirmed modified CDW proximal to Moscow University and Totten glaciers (Fig. 4b). Totten's catchment has been thinning and losing mass since the late-1970s, with numerous studies attributing this to ocean forcing and wind-driven upwelling of modified CDW[4-6,11,24,27,32-34,50,120-123]. Its grounding line has been retreating since at least the 1990s[17,33-34] (Fig. 3a) and, given Totten's large catchment (3.9 m of SLE)[33] and high discharge (~70 Gt yr^{-1})[4], these observations are concerning. However, ice discharge may have slowed recently (2008–2017)[4,123], and some variability in basal melting may reflect intrinsic oceanic processes[50]. Furthermore, the grounding line of both Totten and Moscow University glaciers sit on prograde slopes extending 50-60 km up-ice[10], suggesting that imminent Marine Ice Sheet Instability is unlikely.

Elsewhere in Wilkes Land, glaciers entering Porpoise Bay have received much less attention, but have experienced pronounced thinning[5-7] and are sensitive to buttressing from landfast sea-ice[37]. Frost Glacier has the largest catchment (0.84 m SLE)[4] and underwent moderate thinning (<0.5 m yr^{-1}) over recent decades[5,7,11], concomitant with grounding line retreat[17] (>200 m yr^{-1} from 2010–2016) (Fig. 3b). Holmes Glacier is smaller (0.12 m SLE)[4], but the surface thinning (>1 m yr^{-1}) is greater[7,11], perhaps driven by enhanced ice shelf basal melting[12,116] (Fig. 4b). Both glaciers merit monitoring given their large catchments, high discharge, and sensitivity to ocean processes[37]. Likewise, glaciers draining the ASB into Vincennes Bay are largely unstudied, but lie proximal to some of East Antarctica's warmest intrusions of modified CDW[29]. Bond and Underwood glaciers have increased in flow speed (2008–2016)[3,4]; and the grounding line of Vanderford (Fig. 4b) retreated >800 m yr^{-1} from 1996–2017[4], the highest reported rate for East Antarctica (Fig. 3a).

Further west, Denman Glacier (Fig. 4b) holds ~1.5 m SLE[4] in the ASB. Its grounding line sits atop a deep canyon extending >3,500 m below sea level[10,35]. Both its grounded (17 ±4%) and floating (36 ±5 %) portions accelerated[124] from 1972–2017, accompanied by surface thinning since at least the 1990s[5,24,32]. Denman lost a lateral pinning point during its last major calving event (1984)[124] and, from 1996–2017, the western part of its grounding line retreated 5.4 km along a deep trough[10,17,35]. Ice shelf melt rates of >45 ±4 m yr^{-1} (2011–2014) occur near the grounding line[35] (Fig. 4b), comparable to the highest rates in West Antarctica[12]. One study[4] estimated mass loss from Denman's catchment equivalent to 0.5 mm of sea-level rise from 1979–2017, second only to Totten (0.7 mm) in East Antarctica, but the drivers of any imbalance are unclear given large uncertainties in mass input and limited changes in ice discharge[3,4].

Whilst palaeo-records indicate that the neighbouring WSB retreated during past warm periods, current observations provide limited evidence of change. Cook Glacier has attracted attention due to its large size (~1.6 m SLE)[4] and proximity to a retrograde bed-slope[31]. Its western outlet lost its ice shelf between 1973 and 1989 and subsequently doubled in speed; while the eastern outlet has accelerated since the 1970s[38]. Observations reveal ice surface thinning since at least the 1990s[5,11,120-122], and a small dynamic imbalance (0.2 mm to SLR: 1979–2017)[4], albeit with large uncertainties. Periodic calving events have occurred at the neighbouring Ninnis Glacier[125] (Fig. 2b), also deemed vulnerable to Marine Ice Sheet Instability[10,4], and at Mertz[126]; but without any dynamic response due to negligible buttressing. Limited evidence of current change in glaciers draining the WSB is consistent with low basal melt rates[12-14,116] and a dense shelf regime (Fig. 4a), but ice shelf retreat/calving[30,36,38] in the 1940s to 1980s is suggestive of warmer than present conditions. Hence, recent ocean forcing in this region (difficult to measure due to high volumes of sea-ice/mélange), may not capture the full range of inter-decadal variability.

East Antarctica's other major marine basin – the RSB – is drained by several large glaciers with retrograde slopes (e.g. Recovery, Bailey, Slessor)[127] and may be highly vulnerable to future ocean warming[45], but there is currently no evidence of changes in ice dynamics[3,4,17]. Elsewhere, few glaciers have been studied in East Antarctica's terrestrial sectors, with no obvious changes reported. In Victoria and Oates Land, for example, numerous glaciers have large, unconfined ice tongues that calve periodically, but with no trends in frontal position or ice velocity since the 1970s[30,36,125,128]. The large region encompassing Mac. Robertson Land to Dronning Maud Land is characterised by ice sheet thickening[5,6,11,24,32] due to increased accumulation[47,48] (Fig. 4a), with some evidence of grounding line advance[17] and most catchments gaining mass[3,4]. Shirase – the fastest-flowing glacier in East Antarctica (~2,500 m yr^{-1}) – experiences relatively high basal melt rates (7–16 m yr^{-1})[112] and may have thinned in the 1990s[5,24], but is currently in balance[129] or gaining mass[4].

In summary, the vast majority of East Antarctic outlet glaciers show no discernible change in velocity or discharge over recent decades[3,4], including those draining two of the three marine basins (WSB, RSB). However, some glaciers draining the ASB in Wilkes Land appear to be losing mass in response to ocean heat forcing, similar to glaciers in the WAIS, and with potential for this to be sustained or even increase.

5. Recent Surface Mass Balance

The large spread in estimates of EAIS mass balance (Fig. 2) is derived largely from uncertainties in mass input (surface mass balance: SMB), rather than ice discharge. The mean annual EAIS SMB (1980-2018) over grounded ice has been estimated at +1,247 Gt yr^{-1} from MERRA-2 global reanalysis data[7] and +1,290 Gt yr^{-1} from the MAR regional atmospheric climate model[130] (Fig. 5a). The SMB is dominated by snowfall, with other components (rainfall, sublimation, blowing snow erosion/deposition, runoff) at least an order of magnitude smaller[131].

While most of East Antarctica is relatively dry with typical (interquartile) annual snowfall ranging from 0.05–0.14 meters water equivalent[131], the area is vast. Thus, atmospheric variability (and snowfall) on time-scales from hours to decades[51] is imprinted on overall mass balance. Indeed, inter-annual variations in SMB (e.g. 1980–2018: σ = 106 and 91 Gt yr^{-1} for MERRA-2 and MAR, respectively; Fig. 5a) are comparable to the signal of overall mass change (Fig. 2), highlighting the sensitivity of SMB to short-lived but extreme events and the need for long observational time-series (>10s of years).

The absence of an EAIS-wide array of direct snow accumulation observations means that assessments of SMB rely on atmospheric datasets, atmospheric reanalyses and regional climate models. However, the lack of observations available for assimilation into global reanalyses, and the regional climate models forced by those reanalyses, means that the representation of atmospheric circulation over the EAIS is poorly constrained. This contributes to the large spread in SMB estimates, exacerbating uncertainty in overall mass change. A recent comparison of Antarctic SMB in eight regional climate models[132] found the range in basin-wide SMB varied from ~3% to ~40% of the model mean. Model differences were greatest in the large, dry basins of Adélie and Victoria Land, whilst high accumulation basins (e.g. Wilkes Land) showed more consensus.

Although the mean SMB of models varies substantially, they broadly agree on the magnitude of inter-annual variability, as well as on recent trends[131,133]. Both MERRA-2[7] and MAR[130] show no significant change in EAIS SMB over the last 40 years (<0.1% per year from 1980–2018: Fig. 5a). Shallow ice/firn core analyses[133,134] indicate this forms part of a century of no significant change (1901–2000 SMB trend = 0.1 ± 0.4 Gt yr^{-2}). Over the period 1800–2000, however, a trend of increased accumulation has been reported[133] (0.3 ± 0.1 Gt yr^{-2}), but substantial low-frequency variability increases the uncertainty, suggesting it may be insignificant[51]. Furthermore, due to their relatively short (<20 years) observational record, altimetry and gravimetry methods do not fully capture these decadal-to-century variations in

SMB, which can complicate the separation of SMB and ice dynamical change. Long-term SMB variations, while relatively unconstrained, are also essential to correct altimetry records for changes in firn air content.

Recently, ponding of meltwater on East Antarctic ice shelves has received considerable attention[135-139] due to its potential role in ice shelf collapse via hydro-fracturing[44,85,140-142]. Surface meltwater (streams, lakes, slush), found in the grounding zone of numerous East Antarctic ice shelves[136], indicates insufficient porosity for drainage into the firn. Where firn air content is high, meltwater drains into the firn and may refreeze[142]. Firn air content can be approximated by a liquid-to-solid ratio, defined by the amount of surface melt (and rainfall, rare over East Antarctica[143]) divided by snowfall. Although subject to uncertainties (particularly the liquid component in the marginal areas of the ice sheet), liquid-to-solid ratios are relatively easy to compute and are <25% on most East Antarctic ice shelves (Fig. 5c), indicating annual snowfall is >4 times larger than liquid water fluxes and that a porous firn layer (10s m) accommodates storage/refreezing of summer meltwater. Notable exceptions include the grounding zones of Amery Ice Shelf, with liquid-to-solid ratios up to 80% (similar to Antarctic Peninsula ice shelves), and some eastern Dronning Maud Land ice shelves (~40%). These ice shelves support high densities of supraglacial lakes[136,137,139], but their physical confinement and thickness (e.g. Amery) protect them from widespread hydrofracture. Indeed, ice shelf collapse via hydrofracturing is critically dependent on stress conditions, with <1% of vulnerable ice shelf areas in East Antarctica currently supporting lakes[141].

Given that surface melt has not significantly increased in any of East Antarctica's drainage basins over the last 40 years (Fig. 5b) and that snowfall has remained broadly the same, and increased over western East Antarctica, we suggest few ice shelves are currently at risk from hydrofracture. However, climate projections indicate surface melt and rainfall (especially on East Antarctic ice shelves), as well as snowfall (over the entire EAIS), will increase in the next century[143-149]. This will increase the vulnerability of the northernmost ice shelves[130,142,147,149], such as West and Shackleton[130,140,141]. Shackleton, partially fed by Denman Glacier, already hosts high densities of supraglacial lakes[136,138], experiences high basal melt rates (Fig. 4b), and is considered most at risk[141,149].

6. Future Projections

Since the 2013 IPCC report[150], there has been significant progress in understanding the uncertainties associated with modelling future ice sheet response in Antarctica. Here, we focus on projections that partition the EAIS-only sea-level contribution at 2100, 2300 and 2500 (Fig. 6).

Using the IPCC (2013)[150] method gives median EAIS sea-level contributions of +0.5 to +0.8 cm at 2100 (Fig. 6a: 'IPCC 2013 updated'). Here, the dynamic response was extrapolated from observations and does not vary with emissions scenario, and the SMB response to warming was derived from climate models (recalculated here for Shared Socioeconomic Pathways (SSPs), using temperature projections from the IPCC (2021)[151]). More recent studies generate a wider range of projections with both negative and positive sea-level contributions from the EAIS by 2100, some of which approach +15 cm or more under very high emissions[43,152,153] (Fig. 6a). The Ice Sheet Model Intercomparison Project (ISMIP6) for the sixth phase of the Coupled Model Intercomparison Project (CMIP6) represents the most comprehensive and up-to-date synthesis of these projections[25,148,152,154], using eleven ice sheet models forced by six CMIP5[149] and four CMIP6[25] climate models. Experiments include high and low emission scenarios (RCP8.5/SSP5-85 and RCP2.6/SSP1-26, respectively), a range of parameter values governing the sensitivity of ice shelf basal melting to ocean temperatures[155], and various scenarios of ice shelf collapse driven by atmospheric forcing[144]. Overall, ISMIP6 gives an EAIS-only sea-level contribution ranging from -7 to +15 cm at 2100 (Fig. 6a).

A major uncertainty is the balance between the SMB input (influenced by the choice of climate model) and dynamic losses (largely influenced by the choice of basal melt sensitivity to ocean warming). A comparison of ISMIP6 simulations driven by two different global climate models under RCP8.5, for example, can change the sign of overall mass balance (Fig. 6a: 'ISMIP6: GCM1' versus 'GCM 2'). A similar effect is seen when comparing ISMIP6 simulations using two distributions of the parameter governing basal melting: one derived from the Antarctic average, and the other from a high-melt region proximal to Pine Island Glacier, WAIS (Fig. 6a: 'ISMIP6: mean melt' versus 'high melt').

The ISMIP6 ensemble was unavoidably relatively small (344 simulations from 14 modelling groups) and unevenly sampled. Recently, statistical emulation was used[152] to resample the uncertainties, giving median projections from the EAIS at 2100 (+1.5 to +2.6 cm) that are 2-3 cm higher than the original ensemble means (Fig. 6a: 'ISMIP: all' versus 'ISMIP6 emulator'). This is partly due to the greater weight given to high basal melt values[152], and partly due to the updated IPCC (2021) projections, which have a mean increase of +1.1 cm arising mostly from the estimated response to pre-2015 climate change[151]. The influence of the basal melt sensitivity can also be seen in the dynamic-only contributions from the Linear Antarctic Response to basal melting Model Intercomparison Project phase 2 (LARMIP-2)[156], recalculated here with IPCC (2021)[151] temperature projections (Fig. 6b). The ISMIP6 emulator projects similar 5th to 95th percentiles to LARMIP-2, but much higher medians (+9 to +10 cm), under a 'risk-averse' scenario[152] (Fig. 6b): a subset of climate and ice sheet models that lead to high mass loss via high basal melting and ice shelf disintegration. The SMB contribution to sea level is not modelled by LARMIP-2 but is expected to be negative, lowering the total sea-level contribution (LARMIP-2's EAIS region is also smaller than other studies). After adding the estimated SMB input (median -2 to -5 cm SLE), the IPCC (2021)[151] found that differences between LARMIP-2[156] and the ISMIP6 emulator[152] could largely be explained by different assumptions about basal melt sensitivity, and combined the two for the main assessment with a 'p-box' approach (mean of the two individual medians gives the assessed median; outer edges of the individual 17-83rd percentiles gives the outer edges of the assessed *likely* (17-83%) ranges). Taking the same 'p-box' approach with LARMIP-2 and ISMIP6 here gives the combined median EAIS contributions of +1 to +3 cm by 2100 across all scenarios, with 5th percentiles of -3 to -5 cm, and 95th percentiles of +15 to +17 cm (SSP1-2.6), +16 to +19 cm (SSP2-4.5), and +20 to +25 cm (SSP5-8.5).

Even higher projections at 2100 have been made by incorporating the proposed 'Marine Ice Cliff Instability' [44,85], which involves the collapse of deep ice cliffs at the grounding line, initiated by ice shelf disintegration (Fig. 6b: lower section). This mechanism has been added to one ice sheet model[44,85], motivated by theoretical considerations and observations of ice cliff calving mechanics, and for this model to be able to simulate the highest potential Pliocene sea level contributions (Fig. 1b). The inclusion and representation of an ice-cliff instability are debated[68,86,151,157-159] and the timing of ice shelf disintegration is highly uncertain[151]. Indeed, projections using marine ice-cliff instability[46] are extremely sensitive to warming, with negative contributions under low and medium emissions, but a 95th percentile of +38 cm under very high emissions (Fig. 6b). Expert elicitation[160] does not explicitly define contributing processes, but encompasses the full range of model projections under very high emissions and is narrower for low emissions (Fig. 6b). Both marine ice-cliff instability and expert elicitation were assessed by the IPCC (2021)[151] as *low confidence* projections - that could nevertheless not be ruled out - and were combined with the main projections in a p-box approach. Taking a similar approach here gives a *low confidence* 95th percentile of +47 cm SLE from the EAIS at 2100 under SSP5-8.5.

Few scenario-dependent EAIS projections are available beyond 2100. Maximum contributions under low and medium emissions are +0.6 m SLE at 2300 (Fig. 6c) and +0.7 m at 2500 (Fig. 6d). This suggests the EAIS contribution to sea-level would be well under +1 m over the next few centuries if emissions follow current Nationally Determined Contributions, which are lower than the medium scenario (SSP2-4.5)[161]; and less than +0.5 m under low emissions with a median warming of <2°C (SSP1-2.6 95th percentile at 2300 is 2.2°C)[151].

Under very high emissions, model projections show a wide range, with the EAIS contributing -0.08 to +3.0 m SLE at 2300 (Fig. 6c, upper panel) and +1.0 to +5.4 m at 2500

(Fig. 6d), although the upper bounds would halve (1.3 m at 2300 and 2.3 m at 2500) when excluding the study[44] that the IPCC (2019)[68] deemed to over-estimate ice shelf collapse. *Low confidence* projections using marine ice-cliff instability[46] and from expert elicitation[160] are even higher (Fig. 6c, lower), with 95th percentiles of +4.7 m and +3.9 m at 2300, respectively, although most of the elicited distribution is far lower (83rd percentile 0.2 m SLE). Taking the IPCC (2021)[151] p-box approach to combine these gives a *low confidence* 95th percentile approaching +5 m at 2300. Such high emissions are becoming increasingly less probable, as they would greatly exceed those predicted for Nationally Determined Contributions under the Paris Agreement and other pledges (e.g. net zero emissions by mid-late century)[161].

Spatial patterns of modelled mass loss consistently highlight the vulnerability of East Antarctica's marine-based sectors, but the magnitude and rate of ice loss is model-dependent. Multi-century simulations[43,44,46,148,162,163] typically show grounding line retreat and mass loss in the ASB (Fig. 1f), followed by the WSB and RSB, although the latter shows high sensitivity to ocean warming in some studies[45,163]. Notably, most models do not include recent discoveries[10] of over-deepened subglacial topography that might exacerbate Marine Ice Sheet Instability, such as the deep trough connecting Denman Glacier to the ASB. Given that large parts of East Antarctica remain unsurveyed[10], there may be undiscovered over-deepenings upstream of grounding lines or unknown bathymetric troughs with potential to carry warm waters towards the ice margin: both could increase mass loss beyond current expectations.

Future ocean conditions will exert a critical influence on ice discharge via basal melting and ice shelf buttressing[12-18]. However, coupled global ocean-climate models do not resolve important processes, such as circulation within sub-ice-shelf cavities, tidal flows and eddies, and gradients across the Antarctic Slope Current. This leaves global models with baseline biases in hydrographic properties over the continental shelf and shelf-break, although multi-model means are more realistic[164]. Insights can be gained from examining climate projections in terms of future surface atmospheric warming, sea-ice melt, wind and ocean circulation changes, and shelf-water hydrography. Projected atmospheric circulation changes include an on-going poleward shift and strengthening of the Southern Hemisphere westerly jet across all seasons[165], and a weakening of the coastal easterlies during austral summer and autumn, particularly around East Antarctica[146]. These wind changes are likely to weaken the Antarctic Slope Current, enabling enhanced CDW intrusions onto the shelf[166,167], particularly around Wilkes Land and west to Prydz Bay. Projected surface warming and the addition of meltwater also enhances vertical stratification over the shelf, reducing or shutting down dense shelf water formation[168] and leaving the deep-shelf waters warmer. Meltwater input from ice shelves may also create a positive feedback, with additional freshening driving sub-ice-shelf warming, leading to further melt[169,170], as hypothesised for the last deglaciation[108]. Sea-ice loss also reduces albedo over the ocean, driving further warming[171] and increasing the vulnerability of outlet glaciers to ice shelf/tongue collapse[37]. However, climate models have typically struggled to reproduce Antarctic sea-ice trends[172], which are improved when ice shelf melt[173] and improved representations of sea-ice motion[174] are included.

7. Lessons from the Past Inform the Future

Evidence from the palaeo-record and numerical modelling highlight the sensitivity of East Antarctica's major marine basins (the ASB, WSB and RSB) to past warm periods, including significant ice loss during the early to Mid-Miocene (24–14 Ma), and a multi-metre sea-level contribution during the mid-Pliocene (5.3–2.6 Ma), when atmospheric CO_2 concentrations were comparable to present-day. Retreat of the WSB during more recent interglacials (marine isotope stage 11) further highlights its sensitivity to modest warming scenarios (1.5–2°C). During the last deglaciation, however, there were only limited changes to the ASB and WSB, with grounding line retreat focussed in the marine embayments of the Ross and Weddell Seas that connect the WAIS and EAIS (Fig. 1c). Here, retreat has been linked to a positive feedback driven by ocean warming, whereby meltwater input freshened surface waters, facilitating increased incursions of modified CDW onto the continental shelf, continuing into the Holocene[108]. This mechanism, coupled with evidence of Marine Ice Sheet Instability across major over-deepenings during deglaciation, illustrates a plausible scenario for destabilising some major East Antarctic outlet glaciers over the next few centuries (e.g. Denman). Furthermore, there are signs that some glaciers draining the ASB in Wilkes Land are currently losing mass, with grounding line retreat and ice surface thinning rates comparable to, and sometimes exceeding, millennial-scale rates of change during the last deglaciation (Fig. 3); and raising the possibility that a longer-term dynamic response to ocean forcing is underway. However, the WSB and RSB currently show limited evidence of change; even though the latter is deemed most vulnerable to future ocean warming[45] and may already be exposed to periodic intrusions of modified CDW[175] that could increase later this century[176].

Current understanding is therefore insufficient to determine if and when specific thresholds of instability might be reached in East Antarctica's three marine-based sectors. Indeed, there is no single EAIS response, or time-scale of response, and estimates of overall mass balance may obscure emerging trends of mass loss from some catchments. Furthermore, recent and future trends in SMB, dominated by snowfall, are subject to extreme inter-annual variability and large uncertainties. These uncertainties, together with limited data to inform models of glacio-isostatic adjustment and corrections for firn air content, lead to satellite-based estimates of EAIS mass balance with large uncertainties.

Future work should continue to target early-warning signs of dynamic imbalance in the three major marine basins, such as ice surface thinning propagating upstream from retreating grounding lines, together with ice flow acceleration and ice shelf thinning. There remains an urgent need to understand the sensitivity of basal melting to ocean temperatures, and for more detailed observations of continental shelf bathymetry, bedrock topography proximal to, and up-ice from, current grounding lines, and improved observations of sub-shelf cavities and oceanic processes. These observations should be supplemented with more widespread and systematic palaeo-campaigns on East Antarctic continental shelves to constrain the sensitivity of catchments to past ocean-climate forcing (e.g. the RSB), including rates of change and potential tipping points. Such data can inform numerical modelling, where multi-model and perturbed parameter ensembles are required to improve the robustness of multi-century projections.

Despite current uncertainties, surface melt and rainfall (particularly on ice shelves), and snowfall (over the entire EAIS), will increase this century. By combining the 'ISMIP6 emulator' with 'LARMIP-2 updated’ (Fig. 6b) and an intermediate IPCC (2021) SMB estimate, we find the EAIS makes a small positive contribution to sea level (+2 cm) at 2100, but with a wide range depending on scenario (5^{th} to 95^{th} percentiles: -4 cm to +16-22 cm), and with upper bounds driven by high basal melt sensitivity to warming. *Low confidence* projections, due to limited evidence, reach +0.47 m at 2100 under very high emissions (Fig. 6b). If warming continues beyond 2100, sustained by high emissions, evidence from the palaeo-record, recent observations, and numerical modelling projections (albeit derived from a small number of studies) point to significant potential contributions to global mean sea level from marine-based sectors, reaching ~1-3 m or more by 2300 (Fig. 6c) and ~2-5 m by 2500 (Fig. 6d). Catchments most at risk drain the ASB in Wilkes Land (Frost, Holmes, Totten, Vanderford, Denman, Moscow University), and the WSB in George V Land (Cook, Ninnis), with the RSB also potentially vulnerable. Crucially, if the Paris Agreement to limit warming to well below 2°C above pre-industrial is satisfied, significant mass loss could be reduced or averted (Fig. 6c, d: SSP1-2.6/RCP2.6), with the EAIS sea-level contribution remaining below +0.5 m at 2500. Even under emissions similar to Nationally Determined Contributions which exceed this temperature target (Fig 6c, d: SSP2-4.5/RCP4.5), East Antarctica’s contribution to sea-level rise would remain well below +1 m over the coming centuries. The fate of the world’s largest ice sheet remains very much in our hands.

Author Contributions:

CRS developed the idea for the paper and all authors provided input on its initial contents and structure. CRS drafted Section 1. GJGP and SSRJ drafted Section 2, with contributions from MJB and TvdF. RSJ drafted Section 3 with contributions from MJB. CRS and BWJM drafted Section 4, with contributions from MHE and AF. JTML and BM drafted Section 5 with input from MAK. CR and TLE drafted Section 6, with contributions from MHE. CRS drafted Section 7 with input from TLE. All authors provided comments and edits on all sections of the paper. GJGP produced Fig. 1, with input from CRS. PLW produced Fig. 2 with input from CRS, MAK and RSJ. RSJ produced Fig. 3, with input from BWJM and CRS. AF, MHE and BWJM produced Fig. 4. JTML and BM produced Fig. 5. TLE carried out the analysis and produced Fig. 6 with input from CR.

Competing Interests:

The authors declare no competing interests.

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Figure Captions:

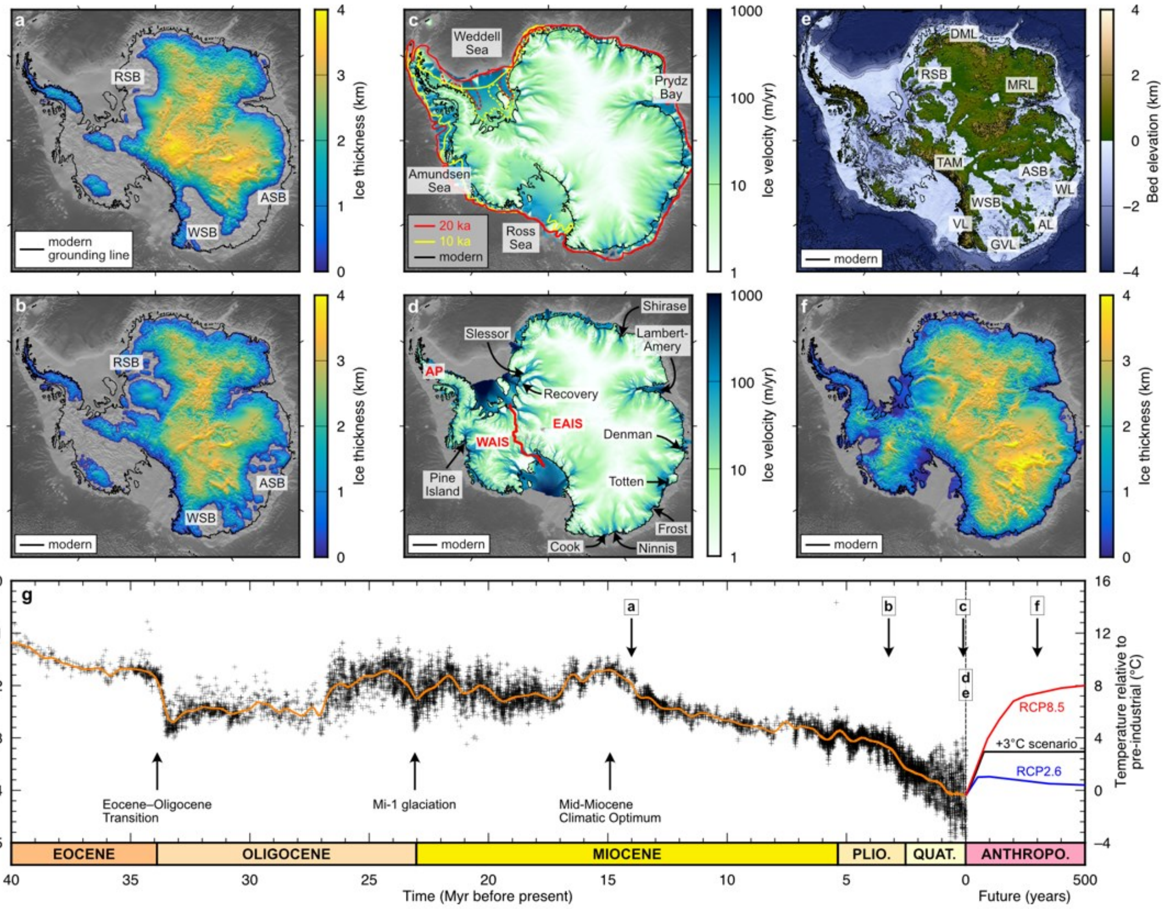

Figure 1: Grounding line extent and characteristics of the East Antarctic Ice Sheet (EAIS) at selected times in the past, present and future. (a) Modelled ice thickness during the Mid-Miocene[57] and reconstructed Mid-Miocene palaeotopography[177] in greyscale, showing deglaciation of West Antarctica and East Antarctica's three major subglacial basins: the Recovery (RSB), Wilkes (WSB) and Aurora (ASB). **(b) Modelled ice thickness during a warm mid-Pliocene** interglacial with hydrofracturing and ice-cliff calving physics enabled[46] and reconstructed mid-Pliocene palaeotopography[177] in greyscale. **(c) Modelled Last Glacial Maximum (20 ka) ice surface velocities** from a Parallel Ice Sheet Model ensemble best-fit reference simulation[178] and RAISED consortium grounding lines at 20 ka (red) and 10 ka (yellow) inferred from empirical data[179] (dashed lines depict a RAISED scenario in the Weddell Sea that is now considered less likely[94]). **(d) Present-day ice surface velocities** derived from interferometric SAR phase mapping[180], with selected outlet glaciers labelled together with EAIS, West Antarctic Ice Sheet (WAIS) and Antarctic Peninsula (AP). Note that we use the standard definition of the EAIS as Antarctic drainage basin numbers 2-17 (e.g. refs [1, 24]). **(e) Present-day Antarctic bed topography and Southern Ocean bathymetry** from BedMachine[10] (AL = Adélie Land; DML = Dronning Maud Land; GVL = George V Land; MRL = Mac. Robertson Land; TAM = Transantarctic Mountains; WL = Wilkes Land; VL = Victoria Land). **(f) Modelled ice thickness at 2300 under a 3°C warming** scenario[46]. **(g) Global benthic oxygen isotope curve through the Cenozoic**[181] with a 1 Myr-smoothed trend line (orange). The projected temperature of the end-member RCP2.6 (blue) and RCP8.5 (red) future emissions scenarios are displayed to the year 2500. The ice configurations shown in panels a–f are labelled along the timescale.

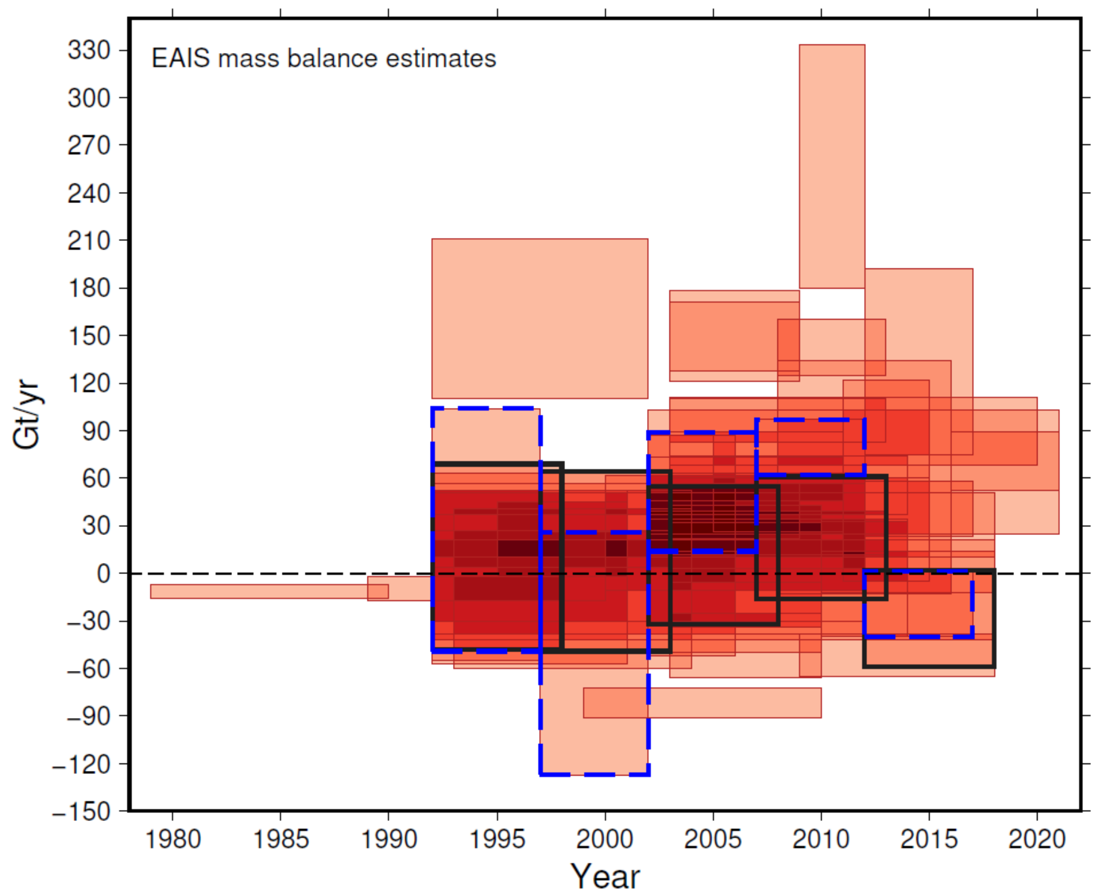

Figure 2: Published estimates of the net mass balance of the East Antarctic Ice Sheet. Each box represents a single estimate of net mass balance with overlapping estimates indicated by darker shading. The horizontal extent of each box represents the survey period. Most studies provide annual data plotted from 1st January to 31st December for any given year. The vertical extent of each box represents the stated uncertainties. Survey areas may vary slightly between different studies, but only those that partition the net mass balance of the EAIS or a large part thereof are included (see Source Data file for numeric values and references). Two recent attempts to reconcile data from multiple methods are highlighted in black[1] and dashed blue[2] lines.

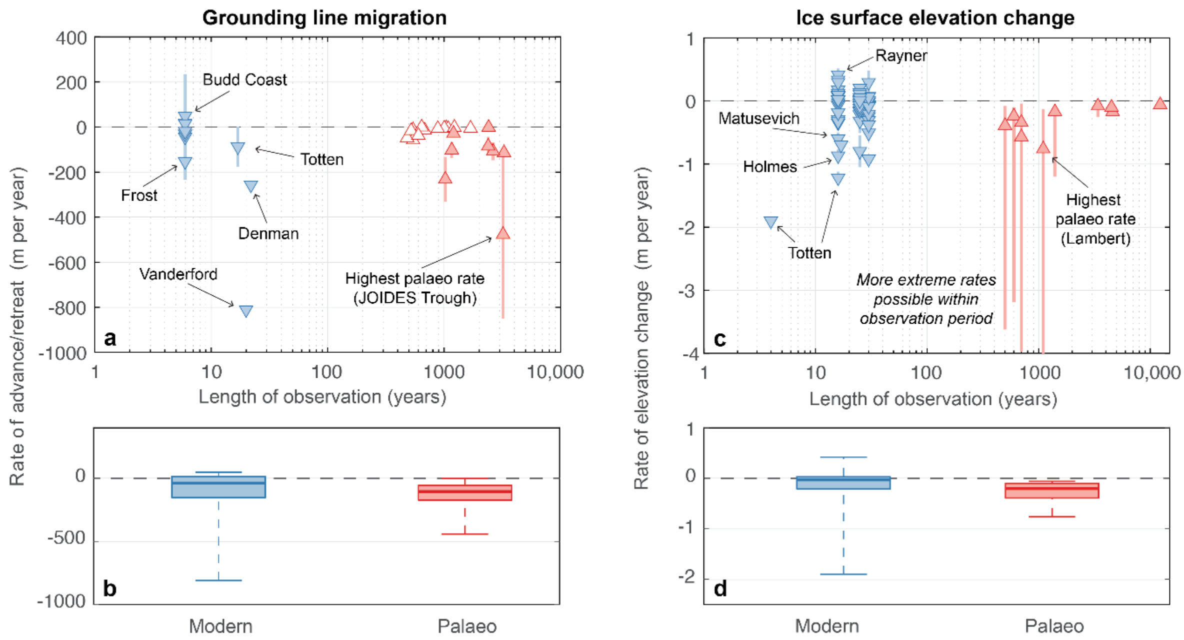

Figure 3: Comparison between published estimates of modern and palaeo (last deglaciation) rates of grounding line migration and ice surface elevation change. (a) Rates of grounding line advance (positive) and retreat (negative) for modern (blue) and palaeo (red) estimates plotted against length of observations. For modern estimates, the triangle marker denotes the mean and the vertical line extends to the maximum possible advance or retreat value quoted in the study. For palaeo estimates, the triangle represents the mean and the vertical line represents the minimum-maximum range, where available. White triangles are minimum palaeo estimates based on the grounding line reaching its present-day position zero years ago. **(b) Box and whisker plots for the range of modern and palaeo mean estimates of grounding line migration.** The median and interquartile range is represented by the horizontal line and the box extent, respectively, while the range is shown by the vertical dashed line. **(c) Rates of ice sheet thickening (positive) and thinning (negative) for modern (blue) and palaeo (red) estimates** plotted against the length of observations. Modern rates from selected East Antarctic outlet glaciers are mean rates extracted from a 20 km x 20 km box immediately up-ice of the grounding line from three recently published altimetry studies[5-7]. Triangle markers and vertical lines represent the mean and published uncertainty range for the modern estimates, and the median and 95% confidence range for the palaeo estimates. **(d) Box and whisker plots for the range of modern (mean) and palaeo (median) estimates of ice surface elevation change** (as in 'c'). See Source Data for numeric values, uncertainties and references. Note that all palaeo-estimates are time-averaged rates for the period of observation and actual rates could have been lower/higher within the period.

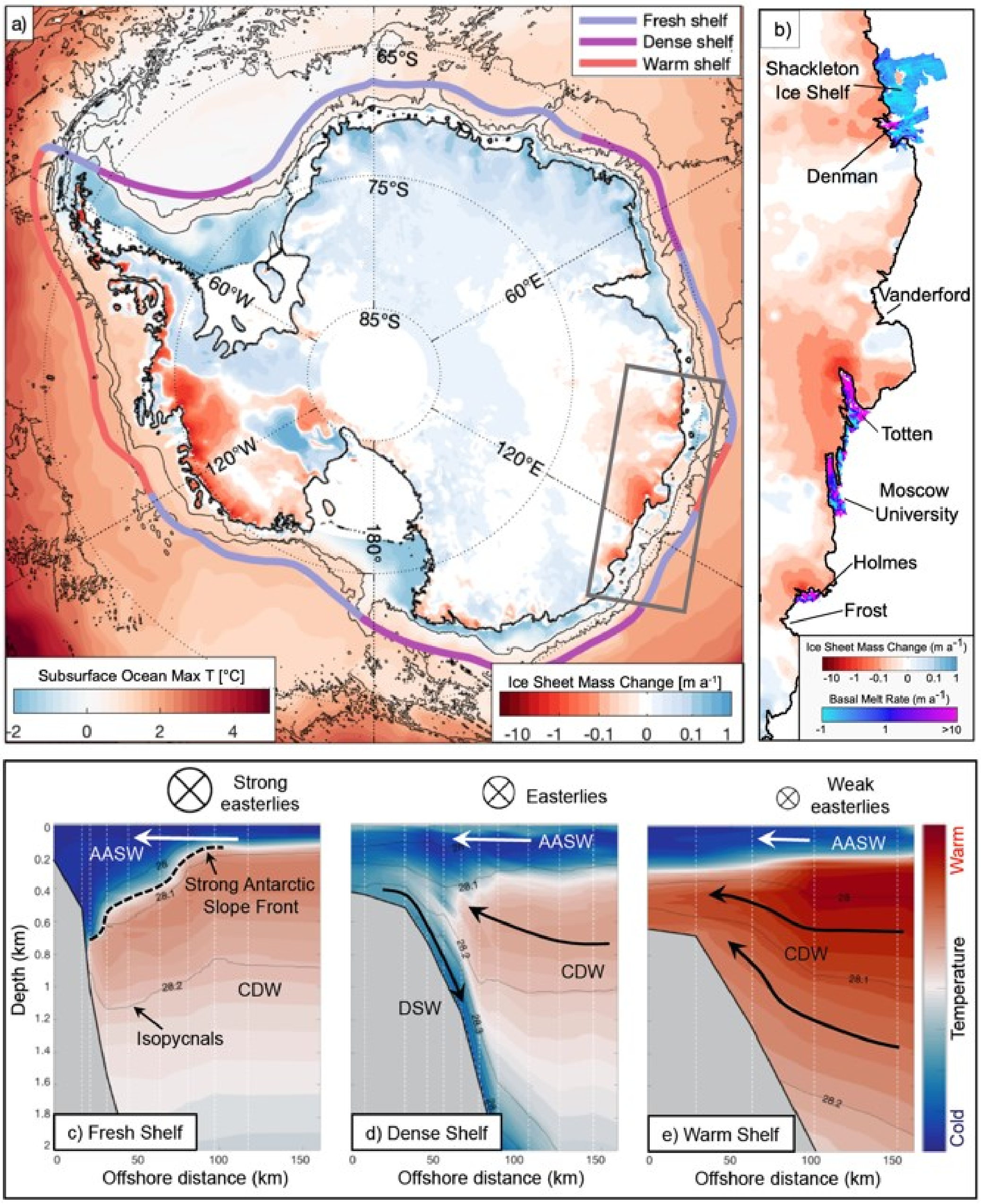

Figure 4: Modern oceanic conditions and characteristic shelf/slope regimes around East Antarctica in relation to recent ice sheet mass changes. (a) Oceanic colours show the 2005-2010 mean subsurface ocean potential temperature maximum from the Southern Ocean State Estimate[182]. Black lines indicate isobaths from ETOPO2v2 (ref. [183]), contoured every 2000 m from the 1000-m isobath; thick black line is the Antarctic continental coast. The thick coloured line parallel to the coast differentiates the three main oceanic shelf regimes (fresh shelf, dense shelf, warm shelf: from ref. [110]).Continental colours represent data from a recent altimetry study[7] of ice sheet elevation change (2003-2019), corrected for firn air content to reflect mass change. **(b) Firn Air Content-corrected ice elevation change in Wilkes Land**, as in (a) with location shown, together with time-averaged ice shelf basal melt rates (2010-2018) from ref. [116]. Note the correspondence between mass loss and high basal melt rates. **(c, d, e) Schematic latitude-depth transects indicating typical winds, subsurface ocean circulation, temperature and density structure in a (c) fresh shelf, (d) dense shelf and (e) warm shelf regime** (modified from ref. [110]). Colours represent temperature and black contours isopycnals of neutral density, with the bold black dashed line in (c) indicating the sharp density gradient across the Antarctic Slope Front. Cross-slope circulation is shown schematically with black and white arrows, and wind direction and strength by arrow tails going into the page. Water masses shown include Antarctic Surface Water (AASW), Circumpolar Deep Water (CDW), and Dense Shelf Water (DSW, also referred to High Salinity Shelf Water in some sectors).

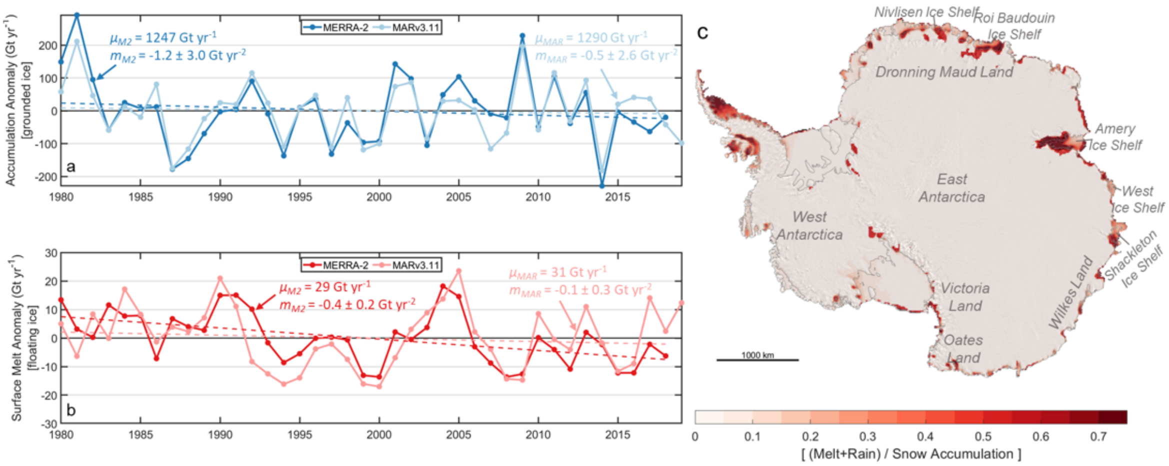

Figure 5: Recent temporal and spatial trends in Antarctic snow accumulation and surface melt. (a) Annual snow accumulation rates integrated over the entire grounded EAIS expressed as an anomaly from the 39-year mean (1980–2018; μ_{M2} = 1,247 Gt yr^{-1}; μ_{MAR} = 1,290 Gt yr^{-1}; from ref. [7] and [130], respectively). The 1980–2018 trends are displayed as dashed lines. (**b**) **As in (a) but for annual surface melt rates over floating ice** only (μ_{M2} = 29 Gt yr^{-1}; μ_{MAR} = 31 Gt yr^{-1}). **(c) The average liquid-to-solid ratio from MERRA-2[7] and MAR[130] over both the WAIS and EAIS** (grounded and floating), where values approaching zero reflect areas of a thick, porous firn column capable of storing liquid water and those approaching one reflect areas with little to no pore space. See Source Data file for numeric values.

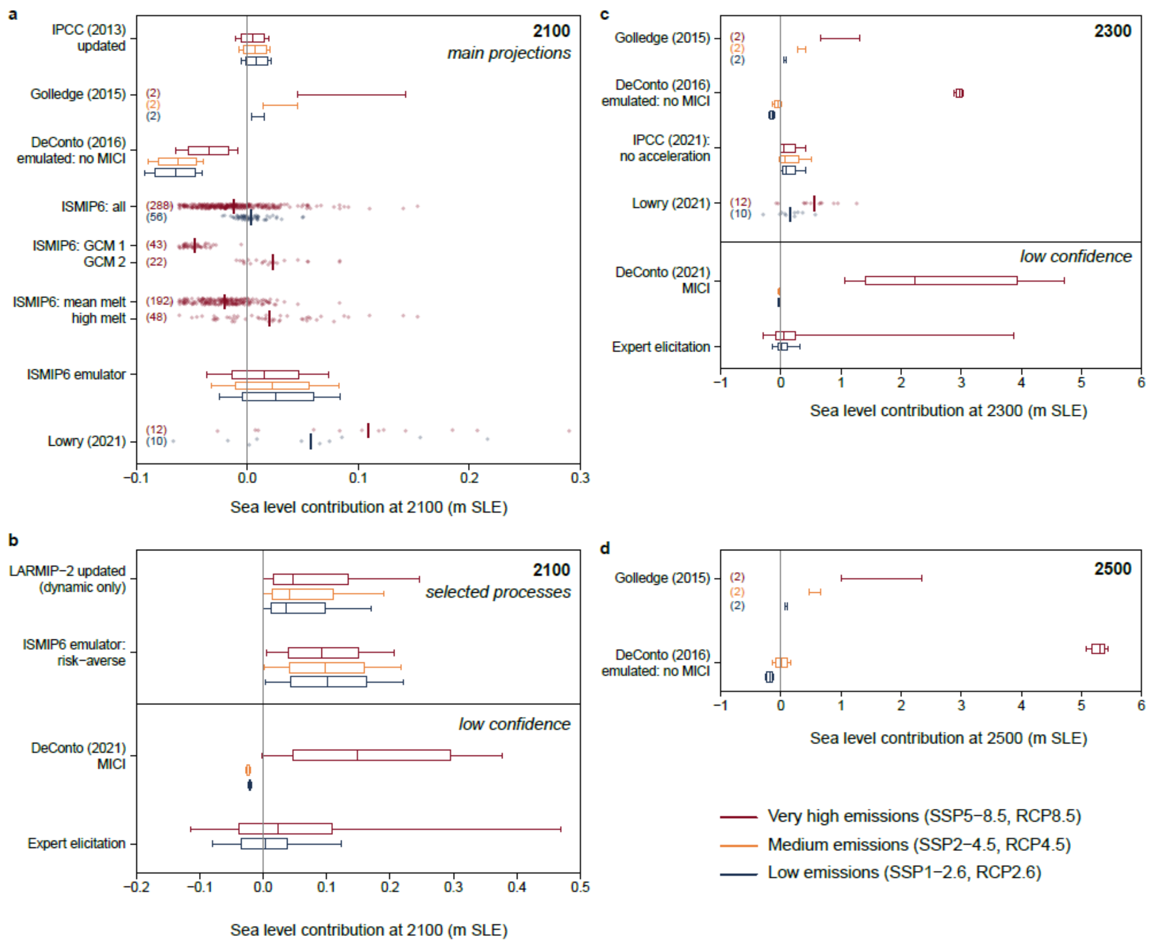

Figure 6: Projected sea level contribution from the East Antarctic Ice Sheet at 2100, 2300 and 2500 under very high, medium and low emissions scenarios. (a) Projections at 2100 from: IPCC (2013) method, re-estimated for IPCC (2021)[151]; ref. [43]; emulated estimate of ref. [44] without Marine Ice Cliff Instability (MICI) mechanism using method of ref.[86]; ISMIP6 multi-model ensemble[25,148,152]; sub-sets of ISMIP6 ensemble using climate forcing from two different Global Climate Models (CCSM4 and HadGEM2-ES), with mean sub-ice shelf melting; subsets of ISMIP6 ensemble using mean versus high sub-ice shelf melting treatment; emulated ISMIP6 projections[152] re-estimated for IPCC (2021)[151], including the addition of 0.09 mm/yr response to pre-2015 climate change; ref.[153], subtracting control simulation and adding the same pre-2015 response. **(b) Projections at 2100 for selected processes and 'low confidence' projections** from: LARMIP-2 dynamic-only contribution[156] re-estimated for IPCC (2021)[151]; emulated ISMIP6 'risk-averse' projections[152] using a high sea-level subset of models and parameter values, with +1.1 cm contribution added to approximate re-estimation for IPCC (2021)[151]; with MICI enabled[46]; expert elicitation[160]. **(c) Projections at 2300** from: ref. [43]; emulated estimate of ref. [44] without MICI, using method of ref. [86]; p-box of IPCC (2013) method[150] and dynamic-only contribution[156], extrapolated beyond 2100 with fixed rate mass loss from IPCC (2021)[151]; ref. [153], subtracting control simulation; with MICI[46]; expert elicitation[160]. **(d) Projections at 2500** from: ref. [43]; emulated estimate of ref. [44] without MICI using method of ref. [86]. Small dots show individual simulations, with short vertical lines showing ensemble means; whiskers without box show range of two simulations. Numbers of simulations are given in brackets. Central line and whiskers show median and 5-95% range; box shows 16%-84% for ref. [44] or 17-83% otherwise. All relative to 1995-2014 baseline[151] except for refs [43,44], relative to 2000; ISMIP6 ensemble, relative to 2015; and ref. [153] for 2105 and 2301, relative to 2025. All use identical climate forcing under Shared Socioeconomic Pathways (SSPs) from IPCC (2021)[151], except for refs [43,44,46], forced with regional climate model (RegCM3) under Representative Concentration Pathways (RCPs); ISMIP6 simulations, forced with various global climate models under RCPs and SSPs; IPCC (2021) no acceleration[151], which has no climate dependence beyond 2100; and expert elicitation, where warming scenarios are interpreted as SSPs following ref.[151]. See Source Data file for numeric values.