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Perspective

Twenty-first century sea-level rise could exceed IPCC projections for strong-warming futures

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SUMMARY

While twentieth century sea-level rise was dominated by thermal expansion of ocean water, mass loss from glaciers and ice sheets is now a larger annual contributor. There is uncertainty on how ice sheets will respond to further warming, however, reducing confidence in twenty-first century sea-level projections. In 2019, to address the uncertainty, the Intergovernmental Panel on Climate Change (IPCC) reported that sea-level rise from the 1950s levels would likely be within 0.61–1.10 m if warming exceeds 4°C by 2100. The IPCC acknowledged greater sea-level increases were possible through mechanisms not fully incorporated in models used in the assessment. In this perspective, we discuss challenges faced in projecting sea-level change and discuss why the IPCC's sea-level range for 2100 under strong warming is focused at the low end of possible outcomes. We argue outcomes above this range are far more probable than below it and discuss how decision makers may benefit from reframing IPCC's terminology to avoid unintentionally masking worst-case scenarios.

Since around 1850, the concentration of atmospheric CO_2 has risen from ~280 to over 415 parts per million (ppm), resulting in a global mean temperature rise of ~0.9°C–1.2°C.[1–4] Even if human-caused emissions are reduced to net zero by 2050, global temperatures may rise to more than 1.5°C above pre-1850 levels.[2] Global CO_2 emissions are still on the rise,[5] however, albeit with a slight coronavirus disease 2019 (COVID-19) dip,[6] and analyses of current policies suggest that greenhouse gas emissions will continue on an upward trajectory over the coming decades.[7,8] This keeps strong-warming futures, which exceed 4°C by the end of the century and continued warming thereafter, well within the realm of the possible.[9]

Sea-level rise is one of the most critical issues the world faces under global warming. According to the IPCC Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC),[10] around 680 million people (10% of the world's population) live in low-lying coastal regions that are susceptible to flooding. As sea levels rise, so too does the risk of flooding, affecting communities immediately through storm surges,[11] and in the longer term as a consequence of seawater infiltration of fresh groundwater reserves, degradation of farmland and accelerated coastal erosion, among other impacts.[12] While gradual sea-level increases, such as the ~20 cm that has occurred since 1850,[13] can be dealt with to a degree by adaptations such as flood-barriers, dykes, and natural water-system management strategies, coastal people may be unprepared for much higher rates of sea-level rise.

Will sea-level rise in the coming century be restricted to less than about 1 m, as indicated by the "likely" (IPCC terminology for >66% probability) range in the SROCC,[10] or could it be larger? Judging from records of the last deglaciation, ice sheets can respond to global warming through rapid mass loss, at rates and by amounts much higher than observed to date, leading to sea-level rise of several meters per century.[14] It is true that the ice sheets then differed from today's in many ways, but process understanding and other indicators show that rapid rise is possible under the higher levels of warming that might occur this century. In this perspective, we discuss the potential for large (>1 m) sea-level rise this century from the ice-sheet response to strong warming, assess limitations to sea-level projection, and offer ways in which the scientific problem may be better understood and translated.

PAST CLIMATE AND SEA-LEVEL CHANGE

Global sea level has risen by ~20 cm over the last 170 years or so.[10] The rate of change has been increasing through time, however, and in the early twenty-first century it is ~3.2 mm/year[10] and growing at a rate of ~0.8 mm/year per decade.[15] Globally averaged sea-level rise is probably the most important metric going forward, but it is important to note that sea-level rise is not globally uniform. For example, over short timescales, changes in winds and currents can shift water away from some coasts and against others. Over longer timescales, the

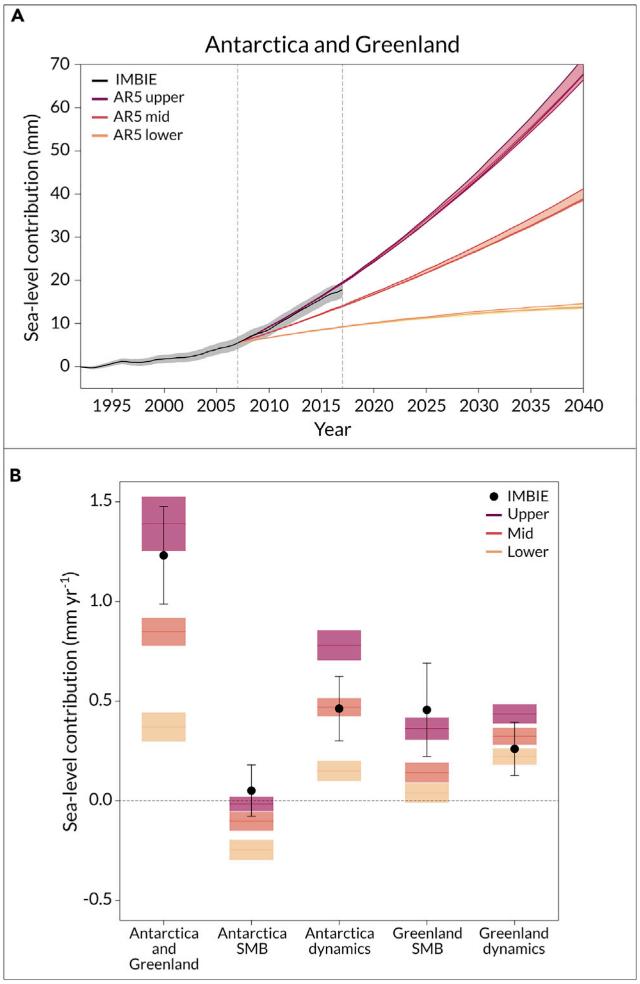

Figure 1. Analysis of ice-sheet mass balance and IPCC sea-level projections

(A) Measured ice loss from Greenland and Antarctica plotted against IPCC Fifth Assessment Report predictions. AR5 upper range relates to the business-as-usual RCP8.5 scenario, whereas the AR5 lower range corresponds to the RCP2.6 scenario of strong action on carbon dioxide emissions.[32]

(B) Components of observed and predicted, as in (A), annual sea-level contributions from Greenland and Antarctica between 2007 and 2017, broken into components of ice dynamics and surface mass balance.[32]

gravitational attraction of the gigantic polar ice sheets pulls ocean water toward them, forcing the sea level to be higher in polar regions than if the ice sheets were absent. Loss of ice from these ice sheets spreads this extra water across the ocean, and even affects the Earth's rotation a little, which also influences the pattern of sea level. Changes in the elevation of coastal land affect local sea level, and may arise from plate tectonic processes, ongoing isostatic response to past changes in loads such as removal of the weight of the ice-age ice sheets, and other processes, including groundwater/fossil-fuel extraction and compaction of coastal sediments.[16]

There are three main factors that affect global sea levels on timescales of decades to centuries:[17,18] (1) the net loss of mass from glaciers and ice sheets to the oceans; (2) the expansion of ocean water as it warms; and (3) changes in water storage on land outside of the glaciers and ice sheets, including the balance between removal of water from groundwater aquifers versus increases in impoundment of water behind dams on rivers. The oceans hold around 97% of all water on Earth. Of the remaining 3%, around two-thirds are held as ice within glaciers and ice sheets (the other third being in lakes, soils, rivers, and the atmosphere). Glaciers thus hold nearly 70% of the planet's freshwater.[19] By far the greatest ice volume is in Antarctica, where two ice sheets (the East and West Antarctic ice sheets) contain more than 25 million km^3 of ice, which, if melted, would raise global sea level by over 57 m.[20,21] The next largest ice sheet is in Greenland, containing enough water to increase sea level by about 7 m.[22–24] The remaining ice caps and glaciers in the world, if they too melted, would raise sea level by only ~0.32 m.[25] Hence, the greatest uncertainty to future sea-level rise relates to how the polar ice sheets will react to global warming.

The long response times of ice sheets to steady climate forcing, their vast expanse, and their remote and extreme conditions mean it has traditionally been difficult to determine whether they are in positive or negative balance. Field-based assessments of mass balance, while useful glaciologically, have substantial uncertainties on an ice-sheet scale.[26] More appropriate is the use of airborne and satellite measurements calibrated against field observations and extending them consistently to the whole-ice-sheet scale. Satellite measurements of ice-sheet surface elevation,[27] when repeated over a few years, allow changes to be determined with ~10-cm accuracy. Another satellite technique combines measurements of ice volume flux along the ice-sheet periphery with reconstructions of surface mass balance in the interior regions, and yields information on the components of mass change (i.e., ice flow and rates of snowfall and meltwater runoff).[28,29] Finally, satellite-based gravity measurements offer a comprehensive, precise, and largely independent calculation of ice mass changes.[30]

Such measurements, over 20 years for gravity, 28 years for altimetry, and 42 years for the mass budget, have revealed that all major sectors of Greenland's ice sheet are losing mass.[31] Furthermore, important parts of the Antarctic ice sheet, and especially West Antarctica's Pine Island and Thwaites glaciers, where the ice rests on a bed well below sea level that deepens upstream, are thinning and losing mass rapidly (Figure 1). Satellite altimetry reveals the rate of West Antarctic mass loss has increased 6-fold since the early 1990s.[27] In contrast, while the world's glaciers also contribute significantly to sea-level rise, the rate of acceleration is smaller than for the ice sheets.[33] The ice-sheet contribution already exceeds the world's glaciers, and in combination they now exceed sea-level rise from thermal expansion of the ocean. This situation seems likely to continue in the coming decades. As an illustration of the pace of regional ice retreat under current climate forcing, frontal positions of mountain glaciers ending on land typically retreat at rates of kilometers per century,[34] while some of Greenland's largest outlet glaciers (e.g., Jakobshavn Isbrae and Zachariae Isstrom) have retreated at 0.5 to 0.6 km/year.[35–37] However, glaciers in the Amundsen Sea sector of West Antarctica[38] are retreating at 1 to 2 km/year, orders of magnitude faster than typical mountain glaciers.

Further back in time, extensive evidence shows that warming has repeatedly driven large, rapid sea-level rise from ice loss. At

Box 1. Ice-sheet instability and rapid mass loss

Changes in ice-ocean interactions affecting ice-sheet flow have the greatest potential for causing large, rapid sea-level rise.[48,49] With both cold air and ocean,[50] ice flowing into the ocean generally forms a floating but attached ice shelf, calving icebergs from its terminus. Sufficient warming of air or water causes ice-shelf loss, sometimes catastrophically,[51,52] shifting calving to a non-floating (grounded) ice cliff. Essentially all ice shelves are restrained by friction with their sides or local seafloor highs, in turn restraining the flow of non-floating ice.[53] Ice-shelf reduction or loss speeds flow of that non-floating ice into the ocean, raising sea level. We focus here on ice-shelf loss and retreat from a topographic bottleneck, two well-known and often-coupled "tipping-point" behaviors that can raise sea level rapidly under strong warming.

Faster flow from thinning of ice shelves caused by warming of ocean or air can lead to fracture along ice-shelf sides and then shelf loss, as occurred at Jakobshavn Glacier in Greenland.[52] Meltwater ponded in ice-shelf surface crevasses can also accelerate iceberg calving and remove an ice shelf, as happened over a few weeks to the Larsen B ice shelf along the northern Antarctic Peninsula.[54] Ice shelves appear to be easier to lose than to regrow; despite environmental fluctuations and transient regrowth,[55] persistent ice-shelf regrowth has not yet been observed. Flow acceleration of non-floating ice has followed ice-shelf thinning and loss,[56] together with grounding-zone retreat as nearby non-floating ice thinned to flotation.

For both ice shelves and grounded calving cliffs, the grounding zone can stabilize over long periods at a bottleneck—a seafloor high and/or fjord narrowing—that partially restrains ice flow. Stability is reinforced by several processes, including sediment accumulation. If further retreat is triggered, loss of restraint from the bottleneck favors faster calving, faster ice flow and thinning, typically causing rapid grounding-zone retreat to the next bottleneck.[57,58]

These retreat processes are well known. John Muir, for example, made important observations of the ~100-km retreat from a bottleneck in Glacier Bay, Alaska, with calving as fast as 11 km/year[59] and ice thinning as much as 1,500 meters between 1750 and 1929.[60,61] Similar or faster calving occurred with flow acceleration and thinning after Jakobshavn Glacier lost its ice shelf, until the next bottleneck was reached. Various terms have been applied to variants of these processes: retreat such as that at Glacier Bay is described as part of the tidewater glacier cycle, grounding-zone retreat of a polar ice sheet may be called the marine ice-sheet instability (MISI), and retreat by calving after ice-shelf loss may be called the marine ice-cliff instability (MICI) (Box 1 Figure).

The recently observed retreats in Greenland and Antarctica have been locally spectacular but are only illustrative of the larger changes that may occur under sustained warming. Glaciers sped up 5-fold to 8-fold following collapse of Larsen B. Jakobshavn Glacier nearly tripled its speed after ice-shelf loss. Additional large changes occurred following ice-shelf loss in North Greenland, Ellesmere Island, and along the Antarctic Peninsula. Ice-shelf thinning and flow acceleration in the Amundsen Sea Embayment of West Antarctica, especially Thwaites and Pine Island Glaciers, could lead to retreat into deep interior basins with as much as 3 m of sea-level rise before stabilization on the next bottleneck. Some large East Antarctic glaciers have begun changing, and those draining the Wilkes, Aurora and other basins have even greater potential to raise sea level.[62–65] Similarly, in Greenland, marine-based outlets underlain by deep channels extending inland (e.g., Petermann, Humboldt, Zachariae, Nioghalvfjerdsfjorden, Jakobshavn)[37,52] house enough ice to raise sea level by 2.3 m.

One model that simulated ice-shelf loss and retreat from bottlenecks, including calving from grounded ice cliffs,[66] found that, for cases in which strong anthropogenic warming triggered major West Antarctic retreat, the marine basins largely deglaciated over the following century or so. This model restricted ice-cliff calving to rates that have been exceeded elsewhere; much of the contribution to rapid sea-level rise came from the great thickness of the West Antarctic ice and the huge width of the calving front that developed during retreat. A West Antarctic calving-cliff retreat would produce much higher and wider cliffs with much larger stresses than any that have been observed, so calving might be faster, or indeed much faster, than previously measured.[67,68]

Ice-shelf loss and calving-cliff retreat are well recognized at the fjord scale but often are not included well in ice-sheet models. This is partly because of inherent difficulties in simulating fracture processes. (Small differences in conditions may cause a ceramic coffee cup dropped on a hard floor to bounce unharmed or break into fragments.) Furthermore, with no recently observed catastrophic retreats on the scale of Thwaites Glacier or other major Antarctic basins, models cannot be well calibrated against observational or historical data.

The transition from non-floating to floating ice, commonly called the grounding line, is really a complex grounding zone with important but poorly known processes. In addition to calving of grounded ice blocks, retreat and faster flow are promoted by preferential melting undercutting the base of marine-ending ice cliffs,[69–73] a process not yet included in many ice-sheet models.[74,75]

More broadly, the issue of melting at grounding zones of ice shelves and grounded cliffs remains an area of active research. Ice-sheet models are sensitive to basal melt rates imposed at grounding zones.[76,77] Large observed fluctuations in grounding-zone position during tidal cycles promote ocean-water flushing and melting over many kilometers,[78] and inclusion of grounding-zone melting in models is required to match some recent observations.[79,80] Inclusion of such processes in models,[81,82] and their interactions with cliff undercutting and calving, could amplify projected ice-sheet response to warming.[77] The grounding-zone transition remains the least well explored and most challenging part of the ice-ocean system, and yet is central to the future evolution of these glaciers.

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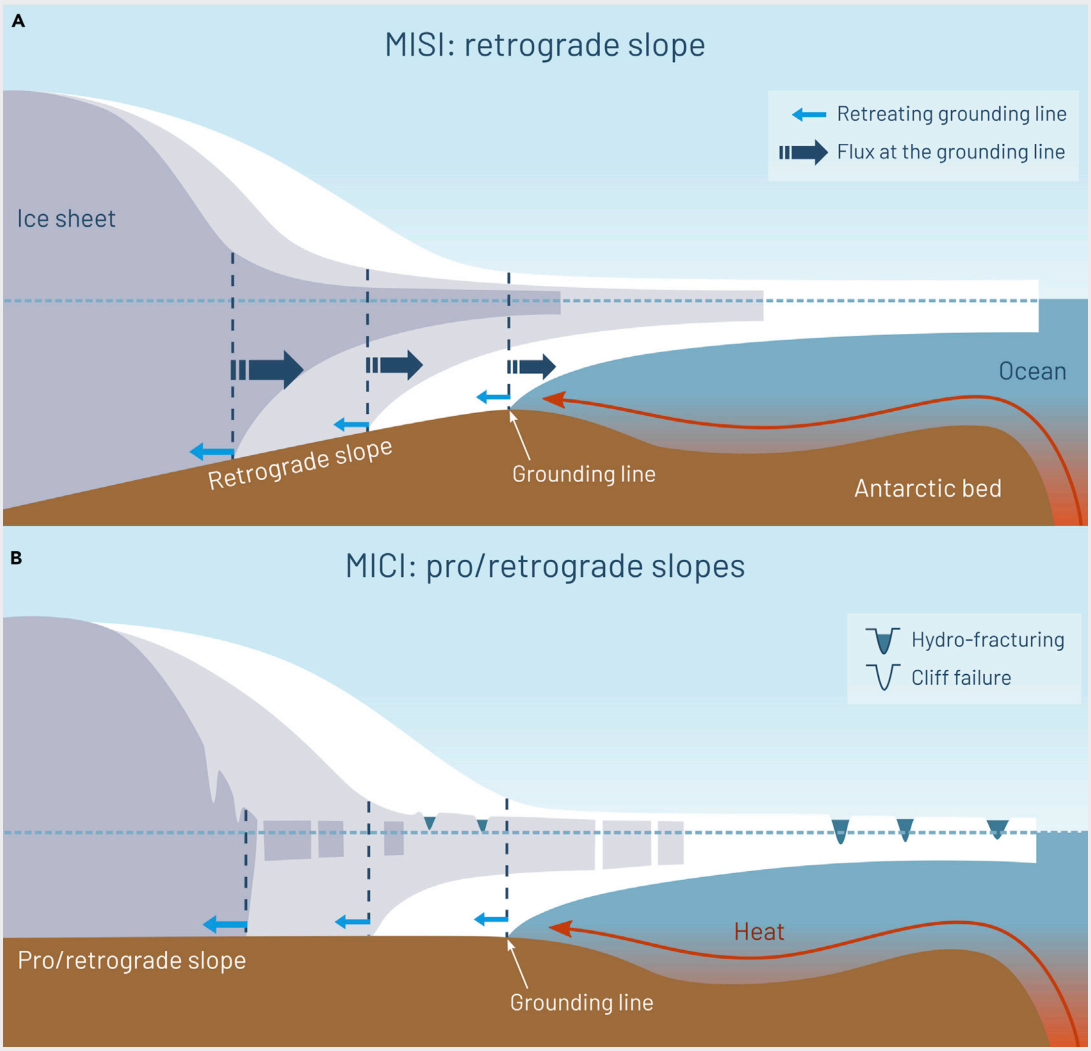
Box 1. *Continued*



Box 1 Figure. Schematic of key processes leading to rapid ice loss at marine glacier and ice-sheet margins.
(A) Marine ice-sheet instability.
(B) Marine ice-cliff instability.[83]

the Last Glacial Maximum (LGM), about 20,000 years ago, ice sheets captured so much water from the oceans that global sea level was up to 130 m lower than now.[39] Deglaciation, globalized by atmospheric CO_2 increase from 180 to 280 ppm, primarily because of CO_2 transfer from the deep ocean to the atmosphere, saw global average temperature rise by ~6°C (or 3.4°C per doubling of CO_2)[40] and sea levels increase, on average, by 1.3 m per century for about 10,000 years. While useful for a general understanding of the transition, these deglacial averages mask considerable variability at global and regional scales, such as meltwater pulses driven by the interactions between the melting ice sheets and the ocean and atmosphere that acted to modulate climate and that led to rates of sea-level rise several times higher than the average.[14,41–43] Ice sheets, and their interactions with the ocean, are critical to rapid climate change in the last deglaciation[44,45] and will likely be so this coming century.[46] They influenced climate by releasing large quantities of water, via direct melting or by iceberg calving, into the oceans, so affecting ocean circulation.[47] Ocean circulation is in turn important to glacier growth and climate change for three reasons. First, oceans are capable of transporting heat between latitudes and hemispheres. Second, ocean thermal conditions are one of several factors affecting the growth of sea ice, which is an important feedback on surface albedo. Third, ocean temperatures are a control on the rate of ice-shelf melting, grounded ice undercutting, and possibly iceberg calving (see Box 1). Hence, through ice-ocean-atmospheric interactions, the gentle rise in global temperatures through the last deglaciation was punctuated regionally by episodes of rapid (on the order of decades or shorter) and extreme (in some areas >±5°C) climate change.[84]

The lessons from the paleo record inform us that it is possible, when pushed by greenhouse gases, for the climate to change rapidly and for ice sheets to drive several meters of global sea-level rise over a century timescale.[85]

ICE-SHEET MEASUREMENT AND MODELING

Measurements of recent sea-level change cannot be used to predict the future beyond simple extrapolations that very rapidly become inadequate under enhanced global warming. The scientific community therefore often relies on computer models, including ice-sheet models, to quantify future environmental change. Ice-sheet modeling incorporates glacier mechanics into numerical schemes that allow ice flow to be calculated at a chosen scale.

The earliest ice-sheet models were relatively basic, involving grids with cells tens of kilometers wide.[86] While such coarse assessments worked reasonably well near the ice-sheet center, where the flow of ice is simple to calculate and speeds are very low, they broke down toward the coast where flow funnels into fast-flowing ice streams. These regions require enhanced physics and higher resolutions. Modern ice-sheet models have improved substantially in their representation of ice streams,[87] using adaptive grids to provide increased resolution where needed.[88] However, despite these impressive advances, models still face major difficulties in accurately deriving the modern flow of ice as measured by satellites.[28,29] These difficulties arise from shortcomings in some combination of the models themselves, knowledge of the physical processes, and awareness of environmental conditions such as the nature of the glacier bed and thermal forcing from the ocean. Consequently, fully process-based ice-sheet models that adequately capture critical dynamics occurring in ice sheets are not (yet) available for projections of century-scale sea-level change.

Ice-sheet models typically are tuned to observations in some way; for example, by adjusting basal conditions (i.e., the friction between the ice and its bed) until an optimization between ice-flow speeds and satellite observations is reached. The advantage of this approach is that the modern ice-sheet flow can be quantified well; the models represent a reasonable snapshot of ice flow today. Large disadvantages may arise from this approach, however. For example, it is possible to reasonably accurately fit the modern ice sheet assuming either a viscous bed or a plastic one,[89] but this significant rheological difference, and therefore the choice of model, can control whether future warming produces ice-sheet collapse or not.[76] In some places, data suggest that the bed actually alternates between viscous and plastic, and simple modeling shows that, while such a bed usually behaves in an intermediate manner, it may exhibit behavior more extreme than either of these situations.[90] Clearly, improved physical understanding of the geotechnical conditions at the ice-sheet bed is a requirement for confident projections of future ice-sheet behavior.

Another level of modeling uncertainty relates to the interaction between the ice sheet and the ocean. The Antarctic ice sheet is changing mostly because of ocean-driven melting beneath its ice shelves, including in the complex grounding zones.[91] Because of this, physical processes acting within ice-shelf cavities are essential to being able to model ice-sheet change. Despite progress in coupling ice sheet and ocean models,[83,92] coupled ice-ocean models presently fail to capture much of the detail revealed by observations.[93] Further developments are thus required to improve ice-ocean processes in models while, at the same time, allowing them to be run efficiently (enough to permit the exploration of ensembles of runs over time scales of ice-sheet response) and appropriately (covering a very uncertain parameter space).

At least three major problems have limited progress in modeling ice-ocean processes. First, the bathymetry along the coastline of Antarctica and Greenland is not known completely, and, where absent, has been replaced by a mathematical interpolation that smooths out deeper trenches,[20,22] which tends to artificially stabilize the modeled ice sheets. Today's ice sheets are generally surrounded by exceptionally cold ocean waters, with warmer waters that often stay offshore of the edge of the continental shelf.[94] Around Greenland, this water is 300–400 m below the surface and is associated with Atlantic Intermediate Water. Around Antarctica, warm water occurs 400–700 m below the surface, associated with Circumpolar Deep Water. These warm waters are stable at depth because they are more dense than the colder, fresher surface waters. Changes in prevailing winds and currents may draw up these warm waters over the continental shelf edge, where they can then drain downward along deep troughs carved by formerly more extensive ice streams to reach the modern ice edge.[95–97] Many of these deep troughs have not been mapped satisfactorily, especially in East Antarctica.[62] As these natural conduits that allow warm waters to reach the ice sheet are missing in models, the impact of ocean-driven forcing on ice sheets is likely to be systematically underestimated.

Second, global ocean models often operate at a coarse spatial resolution (100 km) that does not resolve eddies ($\sim$1 km) and therefore cannot fully reproduce the transfer of ocean heat from the outer ocean onto the continental shelf, into the fjords and ice-shelf cavities.[98] As a result, the source of thermal forcing from the ocean onto the ice sheets is not fully resolved in most models, despite recent progress in incorporating the action of mesoscale eddies and tidal pumping.[99] To represent that ocean heat transfer adequately, it is essential to reproduce the westerly wind regime around Antarctica and its temporal trend, which remains a challenge for most climate models in Antarctica.[100]

Third, ocean observations are limited along the coastline of Antarctica and Greenland. Argo floats do not operate in these shallow regions infested with sea ice, and use of other techniques remains logistically challenging. As a result of these limitations, thermal ocean forcing of the ice sheets, which has long been known to be essential for accurate understanding of their response,[101] is neither well known nor likely to be represented properly in models.[102,103] In the absence of an observation network of ocean conditions adjacent to the ice sheets, modelers must deal with significant and unsatisfactory uncertainties in applying ocean thermal forcing at marine ice-sheet margins.[104]

The floating ice shelves surrounding grounded ice sheets are critical to ice-sheet dynamics. Receiving warmth from both the ocean and atmosphere under climate change, ice shelves are subject to mass loss from their upper and lower boundaries. Ice-shelf disintegration in the northern Antarctic Peninsula has occurred suddenly.[105] Following breakup, grounded glaciers began to flow more quickly through removal of the back-force that was provide by the ice shelf.[54,106] Under strong-warming scenarios, ice shelves bordering the grounded ice sheets in East and West Antarctica may also be vulnerable,[107] which could in turn lead to ice-sheet instabilities and substantial rapid loss of mass to the ocean (see Box 1).

An additional area of concern lies in the subglacial boundary conditions used as input to models. Much progress has been made to improve the description of ice thickness and bed elevation of Greenland glaciers, with corrections to earlier datasets up to many hundreds of meters, which can be the difference between a glacier terminating in warm, deep water or a glacier terminating on land.[22] Similarly, significant advances have been made in Antarctica, especially in West Antarctica, via dedicated airborne and other geophysical surveys in the last decade tied to targeted modeling.[20] However, the shapes of ice-shelf cavities have not been mapped in detail around Antarctica, although progress has been made in key sectors.[108–113] Moreover, the grounding zone (the elusive region where ice meets seawater sloshing back and forth with ocean tides) is challenging to observe with *in situ* devices as it is overlain by hundreds of meters of ice that is often heavily crevassed. Remarkably clever studies have partially overcome these difficulties in restricted places,[114,115] with more work planned,[116] but large data gaps remain. A lack of precise measurements in sub-ice-shelf cavities, and across grounding zones, obstructs our fundamental ability to model ice-ocean processes.

In summary, using ice-sheet models, coupled with ocean models, to derive probabilistic accounts of future sea-level rise remains problematic, especially under situations where the environment is changing rapidly. We can exemplify the problem in two ways. First, no ice-sheet model has been able to prognostically replicate the 6-fold increase in ice-sheet mass imbalance observed over the last 30 years.[27,117] Second, a recent assessment of how a range of ice-sheet models reacted to a scenario where all Antarctic ice shelves are removed (implausible but interesting from a modeling perspective) revealed substantial variation in how much mass was lost in 500 years, ranging from 1 to 12 m in sea-level-equivalent terms depending on the model.[118]

IPCC EMISSIONS SCENARIOS AND SEA LEVEL

Ice-sheet models have had many successes and their projections provide useful guidance when forced by slowly evolving climate and ocean systems, if thresholds leading to more rapid changes are not crossed. While it is possible that West Antarctica is already committed to future large and rapid changes under marine ice-sheet instability,[78,119,120] other modeling indicates that the threshold may not yet have been crossed.[66,76] Since the likelihood of irreversible rapid marine ice loss obviously increases with more warming, the size and rate of future warming are thus of great importance.

IPCC scenarios are analytical tools that can help explore the parameter space of high- and low-emission futures. For example, the climate research community often falls back to using a core set of standardized pathways that reach varying levels of climate forcing by 2100. These include the representative concentration pathways (RCPs) published about a decade ago[121] and more recent scenarios based on a common set of socioeconomic assumptions known as shared socioeconomic pathways (SSPs).[122,123] The lowest of the RCPs (RCP2.6) is most compatible with the Paris climate agreement, as it assumes atmospheric CO_2 concentration ~400 ppm by 2100 and projects global temperature increases likely to be held below 2°C relative to preindustrial levels.[1] The more recent and more stringent SSP1-1.9 scenario gives an even better chance of meeting the Paris Agreement's long-term temperature goal[124] by keeping peak warming well below 2°C and leading to less than 1.5°C of warming by 2100.[123] In contrast, at the high end, the RCP8.5 and SSP5-8.5 scenarios involve no effective response to greenhouse gas emissions and a recarbonization of economic growth, leading to ~1,000 ppm by 2100 and a high probability of global warming in excess of 4°C by 2100. Emissions trends and assessments of current policies indicate that global emissions scenarios are currently neither tracking the high, nor in line with the low, end of this range.[7,125] However, this does not mean that the high warming projected under RCP8.5 cannot occur given current trends[9] or that future policy decisions cannot reduce emissions significantly further, especially considering recent studies that suggest cumulative emissions between 2005 and 2020 are more in line with the upper range of scenarios.[126]

The SROCC[10] states that "the global mean sea level (GMSL) rise under RCP2.6 is projected to be … 0.43 m (0.29–0.59 m, likely range) in 2100 with respect to 1986–2005". For warming as projected under RCP8.5, the SROCC considers the sea-level rise "to be 0.84 m (0.61–1.10 m, likely range) in 2100". It considers the uncertainty in sea level by 2100 to be largely "determined by the ice sheets, especially in Antarctica". Thus, sea-level rise in excess of 1 m by 2100 is not discounted by the IPCC.

Existing ice-sheet models are more likely to provide reliable projections if global warming is kept below 2°C, but a world in which warming exceeds 4°C presents a much more challenging situation. It is quite possible that this extreme situation will lead to reactions and feedbacks in the atmosphere-ocean-ice systems that cannot be adequately modeled at present in terms of process interactions and their consequences. Ice-sheet mass-balance changes observed in the last few decades have tended to occur sooner and more rapidly than midrange values of prior projections (Figure 1),[127,128] at least in part because of processes that were not well resolved in earlier models. For example, ocean thermal forcing around ice sheets did not increase primarily as a result of the slow warming of the world's ocean waters, but rather as a result of changes in the wind regime,[46,95] which acted rapidly to push warm waters more into previously cool regions (note that the wind shifts occurred at least in part because of anthropogenic forcing).[46,129] This served to more than offset the expected growth of the Antarctic ice sheet from increasing precipitation in a warming world, which has been evident only in limited areas such as the Antarctic Peninsula and Dronning Maud Land in East Antarctica. In other words, the physical mechanisms that could mitigate the impact of climate warming, as expected in climate models, have so far proved inadequate to do so.

Comparison of IPCC projections with those made in other ways suggests they are consistent with, or lower than, contemporaneous expert insights, but not higher. For example, a recent expert survey[130] found, among those who responded, a 45%

likelihood that sea-level rise by 2100 under RCP8.5 would exceed the upper end of the likely range given in the IPCC Fifth Assessment Report, which is defined as having a 17% chance of exceedance. A 2014 expert elicitation by an overlapping team of authors[131] similarly found a 42% likelihood of exceedance. A comprehensive survey of sea-level-rise projections made between 1983 and 2018[132] found many who agreed with IPCC projections but more that anticipated greater rise than in the IPCC, and few that projected less rise than in the IPCC. (Note that the Fourth Assessment Report of the IPCC projected sea-level rise ''excluding future rapid dynamical changes in ice flow,'' and thus was not a complete projection and cannot be compared directly, although it often is.) Comparisons of projections from earlier IPCC reports with subsequent sea-level rise found that sea level has been tracking near the upper limits and well above central estimates.[127,133] Furthermore, a more recent study[128] noted that Antarctic ice loss was running ahead of central values in earlier IPCC projections, although within the considerable uncertainties, while another has shown that ice-sheet mass losses from both Greenland and Antarctica are in line with the upper range of IPCC projections (Figure 1).[32] These considerations suggest that the probability of sea-level rise greater than the IPCC range may be higher than generally expected, especially under higher emissions scenarios, and that there is little probability of sea-level rise less than the IPCC range.

FUTURE RESEARCH DIRECTIONS TO IMPROVE SEA-LEVEL PROJECTIONS

Here, we identify some broad areas for future research to narrow uncertainties on this important issue. These overlap in considerable degree and could be integrated in research planning going forward.

Ice-sheet modeling has played, and must continue to play, a crucial role in understanding how ice sheets operate and the processes that are most important to ice-sheet change. We have gained substantial knowledge of ice-sheet sensitivities to warming from models.[134–136] Considerable improvements in ice-sheet models have been made in recent years, moving the community away from complete reliance on shallow-ice or shallow-shelf approximation models by allowing appropriate use of higher-order and full-Stokes models over the entire ice sheet, constrained by satellite observations of surface elevation, surface ice speed, and reconstructions of surface climate conditions. The suite of models (ranging from simple to complex) still provides the only means by which ice-flow processes at a continental scale can be integrated and projected. Progress has been spectacular. However, for ice-sheet models to be capable of making accurate projections about future sea level under high-warming scenarios, further major advances are needed. Important work is already ongoing addressing many issues (e.g., the International Thwaites Glacier Collaboration[137] and other efforts) but much additional sustained work is warranted, intensifying research in other parts of Antarctica, and engaging a larger community of scientists, especially young researchers. Proposals for this research properly would be supplied by the broad polar-scientific community in consultation with policymakers; here, we offer some ideas based on our discussions and experiences, highlighting six important areas that must be advanced greatly within 10 years (Figure 2).

The first is to develop a more precise and complete map of the subglacial topography and seafloor bathymetry around the Antarctic continent, at the scale useful to ice-sheet and ocean modeling, requiring targeted airborne (radar, gravity), ship-based (multibeam echo sounding), and ground-based (seismic and other techniques) measurements, and a suite of geophysical drones and autonomous underwater vehicles (AUVs). By having better measurements of the topography and bathymetry, more-effective coupling between ice-sheet and ocean models will become possible.

Second, there is an urgent need to collect oceanographic data around the ice-sheet margins. Ice-prone shallow-operating Argo-type floats, and other types of AUV and sampling strategies, now exist but need to be deployed more broadly.[138] Sea mammals have been used with tremendous success in both hemispheres[139] to supplement Argo-type programs, but provide data only in their foraging grounds. It is also important to explore sub-ice-shelf cavities with robotic technologies to provide long-term measurements at and close to the ice-sheet grounding zone.[140,141] While limited in scope, targeted field campaigns, including on-ice geophysics and drilling, are essential for major advances in scientific understanding of the complex interactions taking place at glacier grounding zones, several hundred meters below the water line.

A third area is to use data-assimilation techniques more systematically to incorporate new, expanded geophysical information regarding basal conditions and ice-sheet structures into ice-sheet models. Doing so would improve the ability of ice-sheet models to incorporate physical processes, as opposed to using blanket parameterizations, in their depictions of ice flow. The actual bed of a glacier, as known from deglaciated surfaces, does not match gridded bed products[20–22] in aspects other than gross (>500 m scale) morphology.[142] These differences are important, however, and spatially variable flow laws, as well as incorporation of appropriate topography, bed rheology, and basal lubrication, are needed in next-generation models.[90,143]

Fourth, it is essential to improve the coupling between ice-sheet, ocean, and atmosphere models, and this requires a community-wide effort and vastly enhanced computing resources to resolve critical detail at high spatial resolution. These developments are needed to replicate the observational record and increase the level of confidence in projections. While major progress has been made along these lines,[77,103] significant efforts are still required to bring these coupled models to a level of performance that can be vetted by the observational record.

Fifth, fracture plays a central role in determining whether (1) ice shelves will persist or break off, and whether calving cliffs will retreat from stabilizing basal sills or remain pinned; (2) calving will be limited to the fastest rates so far observed or go much higher; and (3) surface meltwater will wedge open crevasses to drive faster calving or drain away without doing so. This is a challenging area because, as noted in Box 1, even small errors in knowledge of forcing or materials properties can lead to very large errors in predicted behavior. Practicing engineers committed to avoiding catastrophic failure design for several-fold safety margins for very good reasons, and the need for such engineering safety margins is of concern in making projections for ice-sheet processes that involve fracture. Extensive

Figure 2. Six ways in which ice-sheet modeling, and thus sea-level predictions, may be improved under further research

observations in the field and the laboratory, combined with expanded modeling, appear necessary.

Sixth, we have never directly observed an ice-sheet collapse, to guide construction and calibration of better models. Engineers avidly study past failures for clues to avoiding future ones. For ice sheets, those past failures are recorded in the paleoclimatic record (and to a much smaller extent in the instrumental record). Ice-sheet models can be tested against those past changes, but the uncertainties in past configurations remain large enough that they translate into major uncertainties in projections.[66,144,145] Improved understanding of past changes is thus required to ascertain, for example, whether marine ice-cliff instability has occurred in the past[146] and, if so, the conditions that led to it occurring. We note, though, that future forcing may move outside of the envelope of paleoclimatic forcing, perhaps bringing novel behaviors, and requiring physics-based models calibrated against laboratory data.

While these and related efforts should improve the ability of models to provide process-based projections of glaciological change, they may not quickly give the desired high level of confidence in those projections, especially under the highest warming scenarios. More-prominent cautionary notes accompanying the assessments would provide valuable guidance to users, based on known and recognized limitations of physical models, and backed up by expert judgments,[23,24,147] which project systematically higher rates of sea level change than are currently projected by many models.[148]

We also recognize the value of an early-warning system for observable steps that could lead to a >1-m sea-level rise by 2100.[85] Coupled ice-sheet/ocean modeling could play an important role in understanding how significant sea-level rise is delivered and could guide construction of such a system.[149] An international effort of observations, and an appreciation of what to look out for, especially under the most extreme of scenarios, would certainly add confidence in sea-level projections.

In many ways, we already have a good start on such an early-warning system comprising satellites, airborne platforms, robotic devices, field investigators, and expert knowledge. While this network is growing and getting stronger, it has major weaknesses at ice-sheet boundaries that require urgent action. We need to develop a capability to collect continuous observations of ocean temperature and salinity along ice-sheet boundaries, with an array of robotic devices in key parts of Antarctica and Greenland that are most vulnerable and capable of causing rapid sea-level rise in the future. Monitoring meltwater ponding in ice-shelf crevasses, retreating grounding zones, and accelerating iceberg calving will also help provide early warnings of the onset of rapid sea-level rise. The required technology for achieving these goals is largely available.

WAYS FORWARD FOR THE IPCC

The IPCC has the difficult task of gathering and evaluating all available evidence, and presenting a consolidated assessment of the state of scientific knowledge on climate change. This is particularly challenging for sea-level rise because, as we discussed, model projections generally do not capture the abrupt and highly non-linear dynamics that are being observed today. These challenges are not new. Indeed, several studies cited above have shown that, over the past few decades, the IPCC's assessments of future sea-level rise have tended to be low, with no tendency to be high or alarmist. Social-scientific research has provided possible insights to this tendency.[150] IPCC assessments of twenty-first century sea-level rise generally place some or much reliance on outputs of models, but many models do not include a complete suite of ice-loss physical processes that could cause rapid sea-level rise, including catastrophic calving of ice shelves and removal of grounded ice blocks and rapid melt of grounded ice by the ocean at the grounding line, some of which are becoming apparent in present-day observations. By omitting some rapid mass-loss processes in ice-sheet models, the upper end of the range of projected sea-level rise becomes inevitably restricted.

Careful reading of the underlying IPCC chapters over multiple reports shows that many of the issues highlighted here have been considered. For example, in discussing the large ice sheets "… new evidence … has again raised the possibility of larger dynamical changes in the future than are projected by state-of-the-art continental models … because these models do not incorporate all the processes responsible for the rapid marginal thinning currently taking place …".[151] Except for the Fourth Assessment Report,[152] however, such concerns have not prevented the IPCC from assigning upper limits on the likely range for future sea-level rise (the Fourth Assessment assigned this upper limit "excluding future rapid dynamical changes in ice flow", as noted above). Those likely ranges have been emphasized in the summaries for policymakers that receive greatest public attention. We suggest that the IPCC could greatly aid stakeholders by more strongly and prominently bringing forward (1) appreciation that ice-sheet and ocean dynamics can act to deliver higher, but not lower, sea level outcomes; (2) acknowledgment that such processes are omitted within many ice-sheet models relied on by the IPCC, as there is still uncertainty on how they operate and interact on sub-century time scales; and (3) awareness that observations of rapid mass loss in both Greenland and Antarctica over the last ~40 years provide a wealth of evidence that these physical processes are already active today in some places and could drive much larger, more rapid sea-level rise under further warming. The example of the underlying text in the Fourth Assessment provides possible guidance on dealing with the deep uncertainties affecting the upper bound on the likely range.

The discussion above shows that the uncertainties in scientific projections of sea-level rise are skewed; the rise may be a little less than central projections, or a little more, with some chance of rise being much more but no similarly large chance of the rise being much less. Even under strong warming, the IPCC's likely projections for 2100 involve the ice sheets providing less than 1% of their potential sea-level rise to the ocean, leaving a substantial possible contribution. Although the potential of sea-level rise exceeding the upper likely range is discussed in the IPCC summaries for policymakers, the asymmetry of the uncertainties is not strongly reported. More-explicit statements addressing this issue could usefully inform policymakers and the public on the possible costs of warming and on the value of additional research addressing the long tail of the probability distribution of large, rapid sea-level rise.

Recent scholarship shows that some policymakers, and members of the public, misinterpret the IPCC's communication of uncertainty and risk.[153–155] The IPCC has established a calibrated language for developing expert judgments and for communicating the degree of certainty in findings of their assessment process.[156] This language relies on two metrics: confidence and likelihood statements. Confidence statements express the confidence in the validity of a finding and are based on an expert judgment of the type, amount, quality, and consistency of available evidence, and the degree of agreement. For example, the SROCC concluded with very high confidence that "acceleration of ice flow and retreat in Antarctica, which has the potential to lead to sea-level rise of several metres within a few centuries, is observed in the Amundsen Sea Embayment of West Antarctica and in Wilkes Land, East Antarctica". Likelihood statements are intended to express quantified uncertainty and use calibrated terms to reflect a specific likelihood. For example, *likely* for an outcome that has 66%–100% probability of being correct, and *very likely* if it has 90%–100% probability. When the IPCC thus reports that future sea-level rise under 4°C warming is *likely* in the 0.6 to 1.1 m range, it means that there could be a 1 in 3 chance that reality falls outside the reported range. These technical definitions are commonly used alongside alternative definitions in non-technical settings, which cause confusion in user communities.[157] Some research also suggests that these definitions may be applied by the IPCC in ways that are too conservative, tending to underestimate scientific confidence on some issues and the scale of potential climate disruption.[158] In our considerable but anecdotal experience, many coastal planners, policymakers, and members of the general public fail to fully appreciate the meaning of the likely range in the assessed sea-level-rise projections. Instead, some treat the high end of the likely range as a worst-case scenario or else the upper end for practical designs. Such a situation is far from ideal because the upper limit of the IPCC likely range was never intended as a worst-case scenario.

Sea-level rise will be one of the most challenging issues faced by society in the coming decades unless we decarbonize fully by mid-century. An objective appreciation and more-effective dissemination of what sea-level rise is possible under strong warming, as opposed to what is deemed likely or is currently accounted for by numerical models, would better inform decision makers, who must increase decarbonization ambition to avoid the most severe of outcomes.

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REFERENCES

1. IPCC (2013). Climate change (2013). The physical science basis. In Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change, T.F. Stocker, et al., eds. (Cambridge University Press), p. 1535.
2. IPCC. (2018). Summary for policymakers. In Global Warming of 1.5°C. An IPCC Special Report on the Impacts of Global Warming of 1.5°C above Pre-industrial Levels and Related Global Greenhouse Gas Emission Pathways, in the Context of Strengthening the Global Response to the Threat of Climate Change, Sustainable Development, and Efforts to Eradicate Poverty, V. Masson-Delmotte, et al., eds. (World Meteorological Organization), p. 32.
3. Schurer, A.P., Mann, M.E., Hawkins, E., Tett, S.F.B., and Hegerl, G.C. (2017). Importance of the pre-industrial baseline for likelihood of exceeding Paris goals. Nat. Clim. Chang. *7*, 563–567.
4. Millar, R.J., Fuglestvedt, J.S., Friedlingstein, P., Rogelj, J., Grubb, M.J., Matthews, H.D., Skeie, R.B., Forster, P.M., Frame, D.J., and Allen, M.R. (2017). Emission budgets and pathways consistent with limiting warming to 1.5 °C. Nat. Geosci. *10*, 741–747.

5. Friedlingstein, P., Jones, M.W., O'Sullivan, M., Andrew, R.M., Hauck, J., Peters, G.P., Peters, W., Pongratz, J., Sitch, S., Le Quéré, C., et al. (2019). Global carbon budget 2019. Earth Syst. Sci. Data *11*, 1783–1838.

6. Le Quéré, C., Jackson, R.B., Jones, M.W., Smith, A.J.P., Abernethy, S., Andrew, R.M., De-Gol, A.J., Willis, D.R., Shan, Y., Canadell, J.G., et al. (2020). Temporary reduction in daily global CO_2 emissions during the COVID-19 forced confinement. Nat. Clim. Chang. *10*, 647–653.

7. Rogelj, J., den Elzen, M., Höhne, N., Fransen, T., Fekete, H., Winkler, H., Schaeffer, R., Sha, F., Riahi, K., and Meinshausen, M. (2016). Paris Agreement climate proposals need a boost to keep warming well below 2 °C. Nature *534*, 631–639.

8. Höhne, N., den Elzen, M., Rogelj, J., Metz, B., Fransen, T., Kuramochi, T., Olhoff, A., Alcamo, J., Winkler, H., Fu, S., et al. (2020). Emissions: world has four times the work or one-third of the time. Nature *579*, 25–28.

9. Fawcett, A.A., Iyer, G.C., Clarke, L.E., Edmonds, J.A., Hultman, N.E., McJeon, H.C., Rogelj, J., Schuler, R., Alsalam, J., Asrar, G.R., et al. (2015). Can Paris pledges avert severe climate change? Science *350*, 1168–1169.

10. IPCC. (2019). Summary for policymakers. In IPCC Special Report on the Ocean and Cryosphere in a Changing Climate, H.-O. Pörtner, D.C. Roberts, V. Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, K. Mintenbeck, A. Alegria, M. Nicolai, and A. Okem, et al., eds., p. 36. https://www.ipcc.ch/srocc/chapter/summary-for-policymakers/.

11. Leuven, J.R.F.W., Pierik, H.J., van der Vegt, M., Bouma, T.J., and Kleinhans, M.G. (2019). Sea-level-rise-induced threats depend on the size of tide-influenced estuaries worldwide. Nat. Clim. Chang *9*, 986–992.

12. Kirezci, E., Young, I.R., Ranasinghe, R., Muis, S., Nicholls, R.J., Lincke, D., and Hinkle, J. (2020). Projections of global-scale extreme sea levels and resulting episodic coastal flooding over the 21st Century. Sci. Rep. *10*, 11629.

13. Jevrejeva, S., Grinsted, A., and Moore, J.C. (2009). Anthropogenic forcing dominates sea level rise since 1850. Geophys. Res. Lett. *36*, L20706.

14. Liu, J., Milne, G., Kopp, R.E., Clark, P.U., and Shennan, I. (2016). Sea-level constraints on the amplitude and source distribution of Meltwater Pulse 1A. Nat. Geosci. *9*, 130–134.

15. Nerem, R.S., Beckley, B.D., Fasullo, J.T., Hamlington, B.D., Masters, D., and Mitchum, G.T. (2018). Climate-change–driven accelerated sea-level rise detected in the altimeter era. Proc. Natl. Acad. Sci. U S A *115*, 2022–2025.

16. Mitrovica, J.X., Gomez, N., Morrow, E., Hay, C., Latychev, K., and Tamisiea, M.E. (2011). On the robustness of predictions of sea level fingerprints. Geophys. J. Int. *187*, 729–742.

17. Frederikse, T., Landerer, F., Caron, L., Adhikari, S., Parkes, D., Humphrey, V.W., Dangendorf, S., Hogarth, P., Zanna, L., Cheng, L., and Wu, Y.-H. (2020). The causes of sea-level rise since 1900. Nature *584*, 393–397.

18. Horton, B.P., Kopp, R.E., Garner, A.J., Hay, C.C., Khan, N.S., Roy, K., and Shaw, T.A. (2018). Mapping sea-level change in time, space, and probability. Annu. Rev. Environ. Resour. *43*, 481–521.

19. Siegert, M.J. (2006). Role of glaciers and ice sheets in climate and the global water cycle. In Encyclopaedia of Hydrological Sciences, *Vol. 4*, M.G. Anderson, ed. (J. Wiley and Sons), p. 164, part 14.

20. Morlighem, M., Rignot, E., Binder, T., Blankenship, D.D., Drews, R., Eagles, G., Eisen, O., Ferraccioli, F., Forsberg, R., Fretwell, P., et al. (2020). Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the Antarctic ice sheet. Nat. Geosci. *13*, 132–137.

21. Fretwell, P., Pritchard, H.D., Vaughan, D.G., Bamber, J.L., Barrand, N.E., Bell, R., Bianchi, C., Bingham, R.G., Blankenship, D.D., Casassa, G., et al. (2013). Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. Cryosphere *7*, 375–393.

22. Morlighem, M., Williams, C.N., Rignot, E., An, L., Arndt, J.E., Bamber, J.L., Catania, G., Chauché, N., Dowdeswell, J.A., Dorschel, B., et al. (2017). BedMachine v3: complete bed topography and ocean bathymetry mapping of Greenland from multibeam echo sounding combined with mass conservation. Geophys. Res. Lett. *44*, 11051–11061.

23. Bamber, J., and Aspinall, W. (2013). An expert judgement assessment of future sea level rise from the ice sheets. Nat. Clim. Chang. *3*, 424–427.

24. Bamber, J.L., Oppenheimer, M., Kopp, R.E., Aspinall, W.P., and Cooke, R.M. (2019). Ice sheet contributions to future sea-level rise from structured expert judgment. Proc. Natl. Acad. Sci. U S A *116*, 11195–11200.

25. Farinotti, D., Huss, M., Fürst, J.J., Landmann, J., Machguth, H., Maussion, F., and Pandit, A. (2019). A consensus estimate for the ice thickness distribution of all glaciers on Earth. Nat. Geosci. *12*, 168–173.

26. Jacobs, S.S., Hellmer, H.H., Doake, C.S.M., Jenkins, A., and Frolich, R.M. (1992). Melting of ice shelves and the mass balance of Antarctica. J. Glaciol. *38*, 375–387.

27. Shepherd, A., Ivins, E., Rignot, E., Smith, B., van den Broeke, M., Velicogna, I., Whitehouse, P., Briggs, K., Joughin, I., Krinner, G., et al. (2018). Mass balance of the Antarctic ice sheet from 1992 to 2017. Nature *558*, 219–222.

28. Mouginot, J., Rignot, R., Bjørk, A.A., van den Broeke, M., Millan, R., Morlighem, M., Noël, B., Scheuchl, B., and Wood, M. (2019). Forty-six years of Greenland Ice Sheet mass balance from 1972 to 2018. Proc. Natl. Acad. Sci. U S A *116*, 9239–9244.

29. Rignot, E., Mouginot, J., Scheuchl, B., van den Broeke, M., van Wessem, M.J., and Morlighem, M. (2019). Four decades of Antarctic ice sheet mass balance from 1979–2017. Proc. Natl. Acad. Sci. U S A *116*, 1095–1103.

30. Velicogna, I., Mohajerani, Y., Landerer, A.G., Mouginot, F., Noel, J., Rignot, B., Sutterley, E., van den Broeke, T.M., van Wessem, M., and Wiese, D. (2020). Continuity of Ice Sheet Mass Loss in Greenland and Antarctica from the GRACE and GRACE Follow-on Missions. Geophys. Res. Lett. *47*, e2020GL087291.

31. Sasgen, I., Wouters, B., Gardner, A.S., King, M.D., Tedesco, M., Landerer, F.W., Dahle, C., Save, H., and Fettweis, X. (2020). Return to rapid ice loss in Greenland and record loss in 2019 detected by the GRACE-FO satellites. Commun. Earth Environ. *1*, 8.

32. Slater, T., Hogg, A.E., and Mottram, R. (2020). Ice-sheet losses track high-end sea-level rise projections. Nat. Clim. Chang. *10*, 879–881.

33. Ciracì, E., Velicogna, I., and Swenson, S. (2020). Continuity of the mass loss of the world's glaciers and ice caps from the GRACE and GRACE follow-on missions. Geophys. Res. Lett. *47*, e2019GL086926.

34. Zemp, M., Huss, M., Thibert, E., Eckert, N., McNabb, R., Huber, J., Barandun, M., Machguth, H., Nussbaumer, S.U., Gärtner-roer, I., et al. (2019). Global glacier mass changes and their contributions to sea-level rise from 1961 to 2016. Nature *568*, 382–386.

35. King, M.D., Howat, I.M., Candela, S.G., Noh, M.J., Jeong, S., Noël, B.P., van den Broeke, M.R., Wouters, B., and Negrete, A. (2020). Dynamic ice loss from the Greenland Ice Sheet driven by sustained glacier retreat. Commun. Earth Environ. *1*, 1.

36. An, L., Rignot, E., Elief, S., Paden, J., Morlighem, M., Mouginot, J., Holland, D.M., and Holland, D. (2017). Bed elevation of Jakobshavn Isbrae, West Greenland, from high-resolution airborne gravity and other data. Geophys. Res. Lett. *44*, 3728–3736.

37. Mouginot, J., Rignot, E., Scheuchl, B., Fenty, I., Khazendar, A., Morlighem, M., and Paden, J. (2015). Fast retreat of Zachariæ Isstrøm, Northeast Greenland. Science *350*, 1,357–1,361.

38. Scheuchl, B., Mouginot, J., Rignot, E., Morlighem, M., and Khazendar, A. (2016). Grounding line retreat of Pope, Smith, and Kohler glaciers, west Antarctica, measured with Sentinel-1a radar interferometry data. Geophys. Res. Lett. *43*, 8572–8579.

39. Lambeck, K., Rouby, H., Purcell, A., Sun, Y., and Sambridge, M. (2014). Sea level and ice volume since the glacial maximum. Proc. Natl. Acad. Sci. U S A *111*, 15,296–15,303.

40. Tierney, J.E., Zhu, J., King, J., Malevich, S.B., Hakim, G.J., and Poulsen, C.J. (2020). Glacial cooling and climate sensitivity revisited. Nature *584*, 569–573.

41. Weaver, A.J., Saenko, O.A., Clark, P.U., and Mitrovica, J.X. (2003). Meltwater pulse 1A from Antarctica as a trigger of the Bølling-Allerød warm interval. Science *299*, 1709–1713.

42. Gregoire, L.J., Otto-Bliesner, B., Valdes, P.J., and Ivanovic, R. (2016). Abrupt Bølling warming and ice saddle collapse contributions to the Meltwater Pulse 1a rapid sea level rise. Geophys. Res. Lett. *43*, 130–139, 137.

43. Harrison, S., Smith, D.E., and Glasser, N.F. (2019). Late Quaternary meltwater pulses and sea level change. J. Quat. Sci. *34*, 1–15.

44. Mayewski, P.A., Maasch, K.A., Dixon, D., Sneed, S.B., Oglesby, R., Korotkikh, E., Potocki, M., Grigholm, B., Kreutz, K., Kurbatov, A.V., et al. (2013). West Antarctica's sensitivity to natural and human-forced climate change over the Holocene. J. Quat. Sci. *28*, 40–48.

45. Abram, N.J., Mulvaney, R., Vimeux, F., Phipps, S.J., Turner, J., and England, M.H. (2014). Evolution of the southern annular mode during the past millennium. Nat. Clim. Change *4*, 564–569.

46. Holland, P.R., Bracegirdle, T.J., Dutrieux, P., Jenkins, A., and Steig, E.J. (2019). West Antarctic ice loss influenced by internal climate variability and anthropogenic forcing. Nat. Geosci. *12*, 718–724.

47. Broecker, W. (2010). The Great Ocean Conveyor: Discovering the Trigger for Abrupt Climate Change (Princeton University Press), p. 176.

48. Joughin, I., Alley, R.B., and Holland, D.M. (2012). Ice-sheet response to oceanic forcing. Science *338*, 1172–1176.

49. Scambos, T., Bell, R.E., Alley, R.B., Anandakrishnan, S., Bromwich, D.H., Brunt, K., Christianson, K., Creyts, T., Das, S.B., DeConto, R., et al. (2017). How much, how fast?: a review and science plan for research on the instability of Antarctica's Thwaites Glacier in the 21st century. Glob. Planet. Change *153*, 16–34.

50. Ferrari, R., Jansen, M.F., Adkins, J.F., Burke, A., Stewart, A.L., and Thompson, A.F. (2014). Antarctic sea ice control on ocean circulation in present and glacial climates. Proc. Natl. Acad. Sci. U S A *111*, 8753–8758.

51. MacAyeal, D.R., Scambos, T.A., Hulbe, C.L., and Fahnestock, M.A. (2003). Catastrophic ice–shelf break-up by an ice–shelf-fragment-capsize mechanism. J. Glaciol. *49*, 22–36.

52. Joughin, I., Howat, M., Fahnestock, M., Smith, B., Krabill, W., Alley, R.B., Stern, H., and Truffer, M. (2008). Continued evolution of Jakobshavn Isbrae following its rapid speedup. J. Geophys. Res. *113*, F04006.

53. Fürst, J.J., Durand, G., Gillet-Chaulet, F., Tavard, L., Rankl, M., Braun, M., and Gagliardini, O. (2016). The safety band of Antarctic ice shelves. Nat. Clim. Chang. *6*, 479–482.

54. Scambos, T.A., Bohlander, J., Shuman, C.A., and Skvarca, P. (2004). Glacier acceleration and thinning after ice shelf collapse in the Larsen B embayment, Antarctica. Geophys. Res. Lett. *31*, L18402.

55. Joughin, I., Shean, D.E., Smith, B.E., and Floricioiu, D. (2020). A decade of variability on Jakobshavn Isbræ: ocean temperatures pace speed through influence on mélange rigidity. Cryosphere *14*, 211–227.

56. Gudmundsson, G.H., Paolo, F.S., Adusumilli, S., and Fricker, H.A. (2019). Instantaneous Antarctic ice-sheet mass loss driven by thinning ice shelves. Geophys. Res. Lett. *46*, 13903–13909.

57. Alley, R.B., Anandakrishnan, S., Dupont, T.K., Parizek, B.R., and Pollard, D. (2007). Effect of sedimentation on ice-sheet grounding-line stability. Science *315*, 1838–1841.

58. Aitken, A.R.A., Roberts, J.L., Van Ommen, T., Young, D.A., Golledge, N.R., Greenbaum, J.S., Blankenship, D.D., and Siegert, M.J. (2016). Repeated large-scale retreat and advance of Totten Glacier indicated by inland bed erosion. Nature *533*, 385–389.

59. Meier, M.F., and Post, A. (1987). Fast tidewater glaciers. J. Geophys. Res. *92*, 9051–9058.

60. Larsen, C.F., Motyka, R.J., Freymuller, J.T., Echelmeyer, K., and Ivins, E.R. (2005). Rapid viscoelastic uplift in southeast Alaska caused by post-Little Ice Age glacial retreat. E. Plan. Sci. Lett. *237*, 548–560.

61. Larsen, C.F., Motyka, R.J., Arendt, A.A., Echelmeyer, K.A., and Geissler, P.E. (2007). Glacier changes in southeast Alaska and northwest British Columbia and contribution to sea level rise. J. Geophys. Res. *112*, F01007.

62. Greenbaum, J.S., Blankenship, D.D., Young, D.A., Aitken, A.R.A., Richter, T.G., Roberts, J.L., Warner, J.C., van Ommen, T.D., and Siegert, M.J. (2015). Ocean access to a cavity beneath Totten Glacier in East Antarctica. Nat. Geosci. *8*, 294–298.

63. Rintoul, S.R., Silvano, A., Pena-Molino, B., van Wijk, E., Rosenberg, M.A., Greenbaum, J.S., and Blankenship, D.D. (2016). Ocean heat drives rapid basal melt of. Totten Ice Shelf. Sci. Adv. *2*, e1601610.

64. Silvano, A., Rintoul, S.R., and Herraiz-Borreguero, L. (2016). Ocean-ice shelf interaction in East Antarctica. Oceanography *29*, 100–113.

65. Brancato, V., Rignot, E., Milillo, P., Morlighem, M., Mouginot, J., An, L., Scheuchl, B., Jeong, S., Rizzoli, P., Bueso-Bello, J., and Prats-Iraola, P. (2020). Retreat of Denman glacier, East Antarctica. Geophys. Res. Lett. *47*, e2019GL086291.

66. DeConto, R.M., and Pollard, D.M. (2016). Contribution of Antarctica to past and future sea-level rise. Nature *531*, 591–597.

67. Parizek, B.R., Christianson, K., Alley, R.B., Voytenko, D., Vaňková, I., Dixon, T.H., Walker, R.T., and Holland, D.M. (2019). Ice-cliff failure via retrogressive slumping. Geology *47*, 449–452.

68. Bassis, J.N., and Walker, C.C. (2012). Upper and lower limits on the stability of calving glaciers from the yield strength envelope of ice. Proc. Roy. Soc. Lon. A *468*, 913–931.

69. Sutherland, D., Jackson, R., Kienholz, C., Amundson, J., Dryer, W., Duncan, D., Eidam, E., Motyka, R., and Nash, J. (2019). Direct observations of submarine melt and subsurface geometry at a tidewater glacier. Science *365*, 369–374.

70. Xu, Y., Rignot, E., Fenty, I., Menemenlis, D., and Mar Flexas, M. (2013). Subaqueous melting of Store Glacier, West Greenland from three-dimensional, high-resolution numerical modeling and ocean observations. Geophys. Res. Lett. *40*, 4648–4653.

71. Rignot, E., Yun, X., An, L., van den Broeke, M., Cai, C., Fenty, I., Li, X., Menemenlis, D., Mouginot, J., and Morlighem, M. (2016). Modeling of ocean-induced ice melt rates of five Greenland glaciers over the past two decades. Geophys. Res. Lett. *43*, 6374–6382.

72. An, L., Rignot, E., Wood, M., Willis, J., Mouginot, J., and Khan, S. (2020). The tale of two ice shelves: Zachariae Isstrøm and Nioghalvfjerdsfjorden, Northeast Greenland data set. Zenodo. https://doi.org/10.7280/D19987.

73. Truffer, M., and Motyka, R.J. (2016). Where glaciers meet water: Subaqueous melt and its relevance to glaciers in various settings. Rev. Geophys. *54*, 220–239.

74. Choi, Y., Morlighem, M., Rignot, E., Mouginot, J., and Wood, M. (2017). Modeling the response of Nioghalvfjerdsfjorden and Zachariae Isstrøm glaciers, Greenland, to ocean forcing over the next century. Geophys. Res. Lett. *44*, 11071–11079.

75. Morlighem, M., Wood, M., Seroussi, H., Choi, Y., and Rignot, E. (2019). Modeling the response of northwest Greenland to enhanced ocean thermal forcing and subglacial discharge. Cryosphere *13*, 723–734.

76. Parizek, B.R., Christianson, K., Anandakrishnan, S., Alley, R.B., Walker, R.T., Edwards, R.A., Wolfe, D.S., Bertini, G.T., Rinehart, S.K., Bindschadler, R.A., and Nowicki, S.M.J. (2013). Dynamic (in)stability of Thwaites Glacier, West Antarctica. J. Geophys. Res. *118*, 638–655.

77. Seroussi, H., and Morlighem, M. (2018). Representation of basal melting at the grounding line in ice flow models. Cryosphere *12*, 3085–3096.

78. Milillo, P., Rignot, E., Rizzoli, P., Scheuchl, B., Bueso-Bello, J., and Prats-Iraola, P. (2019). Heterogeneous retreat and ice melt of Thwaites Glacier, West Antarctica. Sci. Adv. *5*, eaau3433.

79. Lilien, D.A., Joughin, I., Smith, B., and Shean, D.E. (2018). Changes in flow of Crosson and Dotson ice shelves, West Antarctica, in response to elevated melt. Cryosphere *12*, 1145–1431.

80. Goldberg, D.N., Gourmelen, N., Kimura, S., Millan, R., and Snow, K. (2019). How accurately should we model ice shelf melt rates? Geophys. Res. Lett. *46*, 189–199.

81. Horgan, H.J., Alley, R.B., Christianson, K., Jacobel, R.W., Anandakrishnan, S., Muto, A., Beem, L.H., and Siegfried, M.R. (2013). Estuaries beneath ice sheets. Geology *41*, 1159–1162.

82. Jeofry, H., Ross, N., Le Brocq, A., Graham, A., Li, J., Gogineni, P., Morlighem, M., Jordan, T., and Siegert, M.J. (2018). Hard rock landforms generate 130 km ice shelf channels through water focusing in basal corrugations. Nat. Commun. *9*, 4576.

83. Pattyn, F., Favier, L., Sun, S., and Durand, G. (2017). Progress in numerical modeling of Antarctic ice-sheet dynamics. Curr. Clim. Change Rep. *3*, 174–184.

84. Alley, R.B., and Cuffey, K.M. (2001). Oxygen- and hydrogen-isotopic ratios of water in precipitation: beyond paleothermometry. In Stable Isotope Geochemistry, *43*, J.W. Valley and D. Cole, eds. (Reviews in Mineralogy and Geochemistry), pp. 527–553.

85. National Research Council (2013). Abrupt Impacts of Climate Change: Anticipating Surprises (National Academies Press).

86. Budd, W.F., and Smith, I.N. (1982). Large-scale numerical modelling of the Antarctic ice sheet. Ann. Glaciol. *3*, 42–49.

87. Nowicki, S.M.J., Payne, T., Larour, E., Seroussi, H., Goelzer, H., Lipscomb, W., Gregory, J., Abe-Ouchi, A., and Shepherd, A. (2016). Ice sheet model intercomparison project (ISMIP6) contribution to CMIP6. Geosci. Model Dev. *9*, 4521–4545.

88. Cornford, S.L., Martin, D.F., Graves, D.T., Ranken, D.F., Le Brocq, A.M., Gladstone, R.M., Payne, A.J., Ng, E.G., and Lipscomb, W.H. (2013). Adaptive mesh, finite volume modeling of marine ice sheets. J. Comput. Phys. *232*, 529–549.

89. Joughin, I., MacAyeal, D.R., and Tulaczyk, S. (2004). Basal shear stress of the Ross ice streams from control method inversions. J. Geophys. Res. *109*, B09405.

90. Koellner, S., Parizek, R.R., Alley, R.B., Muto, A., and Holschuh, N. (2019). The impact of spatially-variable basal properties on outlet glacier flow. E. Plan. Sci. Lett. *515*, 200–208.

91. Alley, R.B., Anandakrishnan, S., Christianson, K., Horgan, H.J., Muto, A., Parizek, B.R., Pollard, D., and Walker, R.T. (2017). Oceanic forcing of ice-sheet retreat: West Antarctica and more. Ann. Rev. E. Plan. Sci. *43*, 207–231.

92. Golledge, N.R., Keller, E.D., Gomez, N., Naughten, K.A., Bernales, J., Trusel, L.D., and Edwards, T.L. (2019). Global environmental consequences of twenty-first-century ice-sheet melt. Nature *566*, 65–72.

93. Seroussi, H., Nakayama, Y., Larour, E., Menemenlis, D., Morlighem, M., Rignot, E., and Khazendar, A. (2017). Continued retreat of Thwaites Glacier, West Antarctica, controlled by bed topography and ocean circulation. Geophys. Res. Lett. *44*, 6191–6199.

94. Holland, P.R., Jenkins, A., and Holland, D.M. (2010). Ice and ocean processes in the Bellingshausen Sea, Antarctica. J. Geophys. Res. *115*, C05020.

95. Spence, P., Grifes, S.M., England, M.H., Hogg, A.M.C., Saenko, O.A., and Jourdain, N.C. (2014). Rapid subsurface warming and circulation changes of Antarctic coastal waters by poleward shifting winds. Geophys. Res. Lett. *41*, 4601–4610.

96. Jordan, T.A., Porter, D., Tinto, K., Millan, R., Muto, A., Hogan, K., Larter, R.D., Graham, A.G.C., and Paden, J.D. (2020). New gravity-derived bathymetry for the Thwaites, Crosson, and Dotson ice shelves revealing two ice shelf populations. Cryosphere *14*, 2869–2882.

97. Hogan, K.A., Larter, R.D., Graham, A.G.C., Arthern, R., Kirkham, J.D., Totten Minzoni, R., Jordan, T.A., Clark, R., Fitzgerald, V., Wåhlin, A.K., et al. (2020). Revealing the former bed of Thwaites Glacier using sea-floor bathymetry: implications for warm-water routing and bed controls on ice flow and buttressing. Cryosphere *14*, 2883–2908.

98. Mazloff, M.R., Heimbach, P., and Wunsch, C. (2010). An eddy-permitting Southern Ocean state estimate. J. Phys. Oceanogr. *40*, 880–899.

99. Stewart, A.L., Klocker, A., and Menemenlis, D. (2018). Circum-Antarctic shoreward heat transport derived from an eddy- and tide-resolving simulation. Geophys. Res. Lett. *45*, 834–845.

100. Spence, P., Holmes, R.M., Hogg, A., McC, Gries, S.M., Stewart, K.D., and England, M.H. (2017). Localized rapid warming of West Antarctic subsurface waters by remote winds. Nature Clim. Chang *7*, 595–603.

101. Warner, R.C., and Budd, W.K. (1998). Modelling the long-term response of the Antarctic ice sheet to global warming. Ann. Glaciol *27*, 161–168.

102. Seroussi, H., Nowicki, S., Simon, E., Abe-Ouchi, A., Albrecht, T., Brondex, J., Cornford, S., Dumas, C., Gillet-Chaulet, F., Goelzer, H., et al. (2019). initMIP-Antarctica: an ice sheet model initialization experiment of ISMIP6. Cryosphere *13*, 1441–1471.

103. Goelzer, H., Nowicki, S., Edwards, T., Beckley, M., Abe-Ouchi, A., Aschwanden, A., Calov, R., Gagliardini, O., Gillet-Chaulet, F., Golledge, N.R., et al. (2018). Design and results of the ice sheet model initialisation experiments initMIP-Greenland: an ISMIP6 intercomparison. Cryosphere *12*, 1433–1460.

104. Newman, L., Heil, P., Trebilco, R., Katsumata, K., Constable, A., van Wijk, E., Assmann, K., Beja, J., Bricher, P., Coleman, R., et al. (2019). Delivering sustained, coordinated, and integrated observations of the Southern Ocean for global impact. Front. Mar. Sci. *6*, 433.

105. Banwell, A.F., MacAyeal, D.R., and Sergienko, O.V. (2013). Breakup of the Larsen B Ice Shelf triggered by chain reaction drainage of supraglacial lakes. Geophys. Res. Lett. *40*, 5872–5876.

106. Rignot, E., Casassa, G., Gogineni, P., Krabill, W., Rivera, A., and Thomas, R. (2004). Accelerated ice discharge from the Antarctic Peninsula following the collapse of Larsen B ice shelf. Geophys. Res. Lett. *31*, L18401.

107. Lai, C., Kingslake, J., Wearing, M.G., Chen, P.-H.C., Gentine, P., Li, H., Spergel, J.J., and van Wessem, M. (2020). Vulnerability of Antarctica's ice shelves to meltwater-driven fracture. Nature *584*, 574–578.

108. Tinto, K.J., and Bell, R.E. (2011). Progressive unpinning of Thwaites Glacier from newly identified offshore ridge: constraints from aerogravity. Geophys. Res. Lett. *38*, L20503.

109. Schodlok, M.P., Menemenlis, D., Rignot, E., and Studinger, M. (2012). Sensitivity of the ice shelf ocean system to the sub-ice shelf cavity shape measured by NASA Ice Bridge in Pine Island Glacier, West Antarctica. Ann. Glaciol *53*, 156–162.

110. Cochran, J.R., and Bell, R.E. (2012). Inversion of IceBridge gravity data for continental shelf bathymetry beneath the Larsen Ice Shelf, Antarctica. J. Glaciol *5*, 540–552.

111. Millan, R., Rignot, E., Bernier, V., Morlighem, M., and Dutrieux, P. (2017). Bathymetry of the Amundsen Sea embayment sector of West Antarctica from operation ice bridge gravity and other data. Geophys. Res. Lett. *44*, 1360–1368.

112. Tinto, K.J., Padman, L., Siddoway, C.S., Springer, S.R., Fricker, H.A., Das, I., Caratori Tontini, F., Porter, D.F., Frearson, N.P., Howard, S.L., et al. (2019). Ross Ice Shelf response to climate driven by the tectonic imprint on seafloor bathymetry. Nat. Geosci. *12*, 441–449.

113. Millan, R., St Laurent, P., Rignot, E., Morlighem, M., Mouginot, J., and Scheuchl, B. (2020). Constraining an ocean model under Getz Ice Shelf, Antarctica using a gravity-derived bathymetry. Geophys. Res. Lett. *47*, e2019GL086522.

114. Sugiyama, S., Sawagaki, T., Fukuda, T., and Aoki, S. (2014). Active water exchange and life near the grounding line of an Antarctic outlet glacier. E. Plan. Sci. Lett. *399*, 52–60.

115. Fox, D. (2020). Scientists drill through 2,400 feet of Antarctic ice for climate clues. Scientific Am. *16*, 2015.

116. Barbuzano, J. (2020). Diagnosing Thwaites. Eos *101*, https://doi.org/10.1029/2020EO139694.

117. Shepherd, A., Ivins, E., Rignot, E., Smith, B., van den Broeke, M., Velicogna, I., Whitehouse, P., Briggs, K., Joughin, I., Krinner, G., et al. (2020). Mass balance of the Greenland ice sheet from 1992 to 2018. Nature *579*, 233–239.

118. Sun, S., Pattyn, F., Simon, E., Albrecht, T., Cornford, S., Calov, R., Dumas, C., Gillet-Chaulet, F., Goelzer, H., Golledge, N.R., et al. (2020). Antarctic ice sheet response to sudden and sustained ice-shelf collapse (ABUMIP). J. Glaciol, 1–14, https://doi.org/10.1017/jog.2020.67.

119. Joughin, I., Smith, B.E., and Medley, B. (2014). Marine ice sheet collapse potentially under way for the Thwaites Glacier basin, West Antarctica. Science *344*, 735–738.

120. Rignot, E., Mouginot, J., Morlighem, M., Seroussi, H., and Scheuchl, B. (2014). Widespread, rapid grounding line retreat of pine Island, Thwaites, Smith, and Kohler glaciers, West Antarctica, from 1992 to 2011. Geophys. Res. Lett. *41*, 3502–3509.

121. van Vuuren, D.P., Edmonds, J., Kainuma, M., Riahi, K., Thomson, A., Hibbard, K., Hurtt, G., Kram, T., Krey, V., Lamarque, J.-F., et al. (2011). The representative concentration pathways: an overview. Climat. Chang. *109*, 5–31.

122. Riahi, K., van Vuuren, D.P., Kriegler, E., Edmonds, J., O'Neill, B.C., Fujimori, S., Bauer, N., Calvin, K., Dellink, R., Fricko, O., et al. (2017). The Shared Socioeconomic Pathways and their energy, land use, and greenhouse gas emissions implications: an overview. Glob. Environ. Change *42*, 153–168.

123. Rogelj, J., Popp, A., Calvin, K.V., Luderer, G., Emmerling, J., Gernaat, D., Fujimori, S., Strefler, J., Hasegawa, T., Marangoni, G., et al. (2018). Scenarios towards limiting global mean temperature increase below 1.5 °C. Nat. Clim. Chang. *8*, 325–332.

124. Schleussner, C.-F., Rogelj, J., Schaeffer, M., Lissner, T., Licker, R., Fischer, E.M., Knutti, R., Levermann, A., Frieler, K., and Hare, W. (2016). Science and policy characteristics of the Paris Agreement temperature goal. Nat. Clim. Chang. *6*, 827–835.

125. Hausfather, Z., and Peters, G.P. (2020). Emissions – the 'business as usual' story is misleading. Nature *577*, 618–620.

126. Schwalm, C.R., Glendon, S., and Duffy, P.B. (2020). RCP8.5 tracks cumulative CO_2 emissions. Proc. Natl. Acad. Sci. U S A *117*, 19656–19657.

127. Rahmstorf, S., Foster, G., and Cazenave, A. (2012). Comparing climate projections to observations up to 2011. Environ. Res. Lett. *7*, 044035.

128. Slater, T., and Shepherd, A. (2018). Antarctic ice losses tracking high. Nat. Clim. Chang. *8*, 1025–1026.

129. Thompson, D.W.J., and Solomon, S. (2002). Interpretation of recent Southern Hemisphere climate change. Science *296*, 895–899.

130. Horton, B.P., Khan, N.S., Cahill, N., Lee, J.S.H., Shaw, T.A., Garner, A.J., Kemp, A.C., Engelhart, S.E., and Rahmstorf, S. (2020). Estimating global mean sea-level rise and its uncertainties by 2100 and 2300 from an expert survey. Clim. Atmos. Sci. *3*, 18.

131. Horton, B.P., Rahmstorf, S., Engelhart, S.E., and Kemp, A.C. (2014). Expert assessment of sea-level rise by AD 2100 and AD 2300. Quat. Sci. Rev. *84*, 1–6.

132. Garner, A.J., Weiss, J.L., Parris, A., Kopp, R.E., Horton, R.M., Overpeck, J.T., and Horton, B.P. (2018). Evolution of 21st century sea level rise projections. Earth's Future *6*, 1603–1615.

133. Rahmstorf, S., Cazenave, A., Church, J.A., Hansen, J.E., Keeling, R.F., Parker, D.E., and Somerville, R.C.J. (2007). Recent climate observations compared to projections. Science *316*, 709.

134. Cornford, S.L., Martin, D.F., Payne, A.J., Ng, E.G., Le Brocq, A.M., Gladstone, R.M., Edwards, T.L., Shannon, S.R., Agosta, C., van den Broeke, M.R., et al. (2015). Century-scale simulations of the response of the West Antarctic ice sheet to a warming climate. Cryosphere *9*, 1579–1600.

135. Wright, A.P., Le Brocq, A.M., Cornford, S.L., Bingham, R.G., Corr, H.F.J., Ferroccioli, F., Jordan, T.A., Payne, A.J., Rippin, D.M., Ross, N., and Siegert, M.J. (2014). Sensitivity of the Weddell Sea sector ice streams to sub-shelf melting and surface accumulation. Cryosphere *8*, 2119–2134.

136. Ritz, C., Edwards, T., Durand, G., Payne, A.J., Payaud, V., and Hindmarsh, R.C.A. (2015). Potential sea-level rise from Antarctic ice-sheet instability constrained by observations. Nature *528*, 115–118.

137. International Thwaites Glacier Collaboration. (2015). https://thwaitesglacier.org/.

138. Wallace, L.O., van Wijk, E.M., Rintoul, S.R., and Hally, B. (2020). Bathymetry-constrained navigation of Argo floats under sea ice on the Antarctic continental shelf. Geophys. Res. Lett. *47*, e2020GL087019.

139. Padman, L., Costa, D.P., Bolmer, S.T., Goebel, M.E., Huckstadt, L.A., Jenkins, A., McDonald, B.I., and Shoosmith, D.R. (2010). Seals map bathymetry of the Antarctic continental shelf. Geophys. Res. Lett. *37*, L21601.

140. Miles, T., Lee, S.H., Wahlin, A., Ha, H.K., Kim, T.W., Assman, K., and Schofield, O. (2016). Glider observations of the Dotson ice shelf outflow. Deep Sear Res. *123*, 16–29.

141. Barker, L., Jakuba, M., Bowen, A., German, C., Maksym, T., Mayer, L., Boetius, A., Dutrieux, P., and Whitcomb, L.L. (2020). Scientific challenges and present capabilities in underwater robotic vehicle design and navigation for oceanographic exploration under-ice. Remote Sens. *12*, 2588.

142. Holschuh, N., Christianson, K., Paden, J., Alley, R.B., and Anandakrishnan, S. (2020). Linking postglacial landscapes to glacier dynamics using swath radar at Thwaites Glacier. Geology *48*, 268–272.

143. Muto, A., Anandakrishnan, S., Alley, R.B., Horgan, H.J., Parizek, B.R., Koellner, S., Christianson, K., and Holschuh, N. (2019). Relating bed character and subglacial morphology using seismic data from Thwaites Glacier, West Antarctica. E. Plan. Sci. Lett. *507*, 199–206.

144. Edwards, T.L., Brandon, M.A., Durand, G., Edwards, N.R., Golledge, N.R., Holden, P.B., Nias, I.J., Payne, A.J., Ritz, C., and Wernecke, A. (2019). Revisiting Antarctic ice loss due to marine ice-cliff instability. Nature *566*, 58–64.

145. Gasson, E.G.W., and Keisling, B.A. (2020). The Antarctic ice sheet: a paleoclimate modeling perspective. Oceanography *33*, 90–100.

146. Dowdeswell, J.A., Batchelor, C.L., Montelli, A., Ottesen, D., Christie, F.D.W., Dowdeswell, E.K., and Evans, J. (2020). Delicate seafloor landforms reveal past Antarctic grounding-line retreat of kilometers per year. Science *368*, 1020–1024.

147. Mach, J.J., Mastrandrea, M.D., Freeman, P.T., and Field, C.B. (2017). Unleashing expert judgment in assessment. Glob. Environ. Chang *44*, 1–14.

148. Oppenheimer, M., Glavovic, B.C., Hinkel, J., van de Wal, R., Magnan, A.K., Abd-Elgawad, A., Cai, R., Cifuentes-Jara, M., Deconto, R.M., Ghosh, T., et al. (2019). Chapter 4: Sea level rise and implications for low-lying islands, coasts and communities. In IPCC Special Report on the Ocean and Cryosphere in a Changing Climate, H.-O. Pörtner, D.C. Roberts, V. Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, K. Mintenbeck, A. Alegria, M. Nicolai, A. Okem, J. Petzold, B. Rama, and N.M. Weyer, eds., p. 169.

149. Rosier, S.H.R., Reese, R., Donges, J.F., De Rydt, J., Gudmundsson, G.H., and Winkelmann, R. (2020). The tipping points and early-warning indicators for Pine Island Glacier, West Antarctica. Cryosphere Discuss. https://doi.org/10.5194/tc-2020-186.

150. Oppenheimer, M., Oreskes, N., Jamieson, D., Brysse, K., O'Reilly, J., Shindell, M., and Wazeck, M. (2019). Discerning Experts: The Practices of Scientific Assessment for Environmental Policy (University of Chicago Press), p. 304.

151. Meehl, G.A., Stocker, T.F., Collins, W.D., Friedlingstein, P., Gaye, T., Gregory, J.M., Kitoh, A., Knutti, R., Murphy, J.M., Noda, A., et al. (2007). Chapter 10: Global climate projections. In Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, S. Solomon, et al., eds. (Cambridge University Press), pp. 747–846.

152. IPCC (2007). Summary for policymakers. In Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change Solomon, S.D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, and H.L. Miller, eds. (Cambridge University Press), p. 18.

153. Budescu, D., Por, H., Broomell, S., and Smithson, M. (2014). The interpretation of IPCC probabilistic statements around the world. Nat. Clim. Chang. *4*, 508–512.

154. Budescu, D.V., Por, H., and Broomell, S.B. (2012). Effective communication of uncertainty in the IPCC reports. Clim. Chang *113*, 181–200.

155. NASEM (2017). National Academies of Sciences, Engineering, and Medicine; Division of Behavioral and Social Sciences and Education; Committee on the Science of Science Communication: A Research Agenda (Communicating Science Effectively: A Research Agenda. Washington (DC): National Academies Press (US)).

156. Mastrandrea, M.D., Field, C.B., Stocker, T.F., Edenhofer, O., Ebi, K.L., Frame, D.J., Held, H., Kriegler, E., Mach, K.J., Matschoss, P.R., et al. (2010). Guidance Notes for Lead Authors of the IPCC Fifth Assessment Report on Consistent Treatment of Uncertainties (IPCC). https://www.ipcc.ch/site/assets/uploads/2017/08/AR5_Uncertainty_Guidance_Note.pdf.

157. Collins, L.C., and Nerlich, B. (2016). How certain is 'certain'? Exploring how the English-language media reported the use of calibrated language in the Intergovernmental Panel on Climate Change's Fifth Assessment Report. Public Understanding Sci. *25*, 656–673.

158. Herrando-Pérez, S., Bradshaw, C.J.A., Lewandowsky, S., and Vieites, D.R. (2019). Statistical language backs conservatism in climate-change assessments. BioScience *69*, 209–219.