

Chapter 9: Ocean, cryosphere, and sea level change

Coordinating Lead Authors:

Baylor Fox-Kemper (USA), Helene Hewitt (UK), Cunde Xiao (China)

Lead Authors:

Gudfinna Adalgeirsdottir (Iceland), Sybren Drijfhout (Netherlands), Tamsin Edwards (UK), Nicholas Golledge (New Zealand/UK), Robert Kopp (USA), Gerhard Krinner (France), Alan Mix (USA), Dirk Notz (Germany), Sophie Nowicki (USA), Intan Suci Nurhati (Indonesia), Lucas Ruiz (Argentina), Jean-Baptiste Sallée (France), Aimee Slangen (Netherlands), Yongqiang Yu (China)

Contributing Authors:

Kyle Armour (USA), Mathias Aschwanden (Switzerland), Jonathan Bamber (UK), Jason Box (Denmark), Eleanor Burke (UK), Kevin D. Burke (USA), Xavier Capet (France), John Church (Australia), Lee de Mora (UK), Chris Derksen (Canada), Catia Dominguez (Australia/Brazil), Jakob Dörr (Germany), Paul Durack (USA/Australia), Thomas Frölicher (Switzerland), Thian Gan (Canada), Gregory Garner (USA), Sebastian Gerland (Norway/Germany), Heiko Goelzer (Netherlands), Natalya Gomez (Canada), Jonathan Gregory (UK), Robert Hallberg (USA), Alexander Haumann (USA/Germany), Mark Hemer (Australia), Tim Hermans (Netherlands), Emma Hill (Singapore), Regine Hock (USA), Philippe Huybrechts (Belgium), Akm Saiful Islam (Bangladesh), Laura Jackson (UK), Nicolas Jourdain (France), Andreas Kääb (Norway), Shfaqat Abbas Khan (Denmark), Matthew Kirwan (USA), James Kossin (USA), Sophie Lewis (Australia), Shiyin Liu (China), Dan Lowry (New Zealand), Marta Marcos (Spain), Ben Marzeion (Germany), Matthew Menary (France, UK), Philip Orton (USA), Matthew Palmer (UK), Frank Pattyn (Belgium), Cécile Pellet (Switzerland), Chris Perry (UK), Mark Pickering (UK), Johannes Quaas (Germany), Roshanka Ranasinghe (Netherlands/Sri Lanka/Australia), Malcom Roberts (UK), Roelof Rietbroek (Germany), Matthew Roxy (India), Maria Santaloria Otin (France), Abhishek Savita (India, Australia), Alex Sen Gupta (Australia), Helene Seroussi (USA), Sharon Smith (Canada), Olga Solomina (Russia), Esther Stouthamer (Netherlands), William Sweet (USA), Thomas Wahl (USA), Lisan Yu (USA), Jiacan Yuan (USA), Jan Zika (Australia)

Review Editors:

Unnikrishnan Alakkat (India), Benjamin Horton (Singapore/UK), Simon Marsland (Australia)

Chapter Scientists:

Lijuan Hua (China), Tamzin Palmer (UK), Brodie Pearson (United States/UK)

Date of Draft:

01/03/2020

Note:

TSU compiled version

Table of Content

Chapter 9: Ocean, cryosphere, and sea level change	1
Executive Summary.....	5
9.1 Introduction	9
BOX 9.1: Key processes driving sea-level change.....	10
9.2 Oceans	12
9.2.1 Ocean surface	12
9.2.1.1 Sea Surface Temperature (SST)	12
9.2.1.2 Sea Surface Salinity (SSS)	14
9.2.1.3 Air-sea fluxes and winds	15
9.2.1.4 Surface Mixed Layers and Upper Ocean Stratification	17
Cross-Chapter Box 9.1: Marine Heatwaves.....	18
9.2.2 Change in Heat and Salinity	20
9.2.2.1 Ocean Heat Content and Heat Transport	20
9.2.2.2 Ocean Salt Content	24
9.2.2.3 Water Masses.....	25
9.2.3 Regional Ocean Circulation.....	28
9.2.3.1 Atlantic Meridional Overturning Circulation	28
9.2.3.2 Southern Ocean.....	32
9.2.3.3 Tropical Oceans.....	34
9.2.3.4 Western Boundary Currents and Gyres	35
9.2.3.5 Upwelling, Coastal, and Marginal Sea Systems	37
9.2.3.5.1 Eastern Boundary Upwelling Systems	37
9.2.3.5.2 Coastal Systems and Marginal Seas	38
9.2.4 Impact of Ocean Changes on Sea Level	39
9.3 Sea ice.....	40
9.3.1 Arctic Sea Ice.....	40
9.3.1.1 Arctic Sea-Ice Coverage.....	40
9.3.1.2 Arctic Sea Ice volume and thickness	44
9.3.2 Antarctic Sea Ice.....	45
9.3.2.1 Antarctic sea-ice coverage.....	45
9.3.2.2 Antarctic sea-ice thickness	48
9.4 Ice sheets	48
9.4.1 Greenland	49
9.4.1.1 Recent observations and model evaluation.....	49
9.4.1.2 Future projections	53
BOX 9.2: Insights into ice sheet evolution from paleo evidence.....	56

1	9.4.2	Antarctica	58
2	9.4.2.1	Recent observations and model evaluation.....	58
3	9.4.2.2	Future projections.....	61
4	9.5	Terrestrial cryosphere: Glaciers, permafrost and seasonal snow cover.....	67
5	9.5.1	Glaciers.....	67
6	9.5.1.1	Observed glacier extent and volume changes.....	68
7	9.5.1.2	Model evaluation.....	73
8	9.5.1.3	Projections	73
9	9.5.2	Permafrost.....	74
10	9.5.2.1	Observed changes in permafrost extent, temperature and active layer thickness.....	74
11	9.5.2.2	Model evaluation	76
12	9.5.2.3	Projections	77
13	9.5.3	Snow	78
14	9.5.3.1	Observed changes	78
15	9.5.3.2	Model evaluation	80
16	9.5.3.3	Projections	81
17	9.6	Sea Level Change and Commitment	82
18	9.6.1	Global and regional sea-level change in the instrumental era	82
19	9.6.1.1	Sea-level change since 1900.....	82
20	9.6.1.2	Sea-level change in the satellite era.....	83
21	9.6.1.3	Sea-level change acceleration.....	84
22	9.6.1.4	Regional sea-level change	85
23	9.6.1.5	Attribution and emergence of regional sea-level change.....	86
24	Cross-Chapter Box 9.2: The Global Energy Inventory and Sea Level Budget (1971-2018).....		87
25	9.6.2	Geological context of global and regional sea-level change	88
26	9.6.3	Future sea-level changes.....	90
27	9.6.3.1	Global mean sea level projections in the literature and their evaluation	90
28	9.6.3.2	Contributions to projected sea-level change.....	92
29	9.6.3.2.1	Contributors to global mean sea-level change projections	92
30	9.6.3.2.2	Contributors to regional relative sea-level change projections.....	93
31	9.6.3.3	Updated sea-level projections for the next century based on scenarios.....	95
32	9.6.3.4	Updated sea-level projections for the next century based on temperature levels.....	97
33	9.6.3.5	Multi-century and multi-millennial sea-level commitments	98
34	BOX 9.3: High and low end sea level rise projections.....		101
35	9.6.4	Extreme sea levels: Tides, surges and waves	104
36	9.6.4.1	Past changes.....	104
37	9.6.4.2	Future changes.....	106

1	9.7 Robustness and Limitations of the Assessment.....	108
2	Frequently Asked Question	111
3	FAQ 9.1: Will human-induced changes in the oceans and frozen environments be reversible?.....	111
4	FAQ 9.2: How much will sea level rise in the next few decades?.....	113
5	FAQ 9.3: Will the Gulf Stream shut down, and what happens if it does?.....	115
6	Reference.....	117
7	Appendix Chapter 9.A	181
8	Figures	185
9		
10		
11		

Executive Summary

This chapter assesses past and possible future changes in the ocean, cryosphere and sea-level at the process-level using paleo-reconstructions, modern observations and model projections, including the implications of future warming levels and long-term commitments. Unless otherwise noted, the following summary findings confirm or strengthen related findings from the IPCC Fifth Assessment Report (AR5), the Special Report on Global Warming of 1.5°C (SR1.5) and the Special Report on Ocean and Cryosphere in a Changing Climate (SROCC).

Ocean heat and salinity

The ocean has warmed at all levels and continue to do so (*very high confidence*), contributing to future sea-level rise even under low emissions scenarios. Since the 1970s, most ocean warming has been in the Southern Ocean, North Atlantic and within the upper 2000 m resulting in a *likely* global mean thermosteric sea-level rise of 0.04 ± 0.01 m over the period 1971-2015. Over years to a decade, regional heat patterns are dominated by internal and circulation variability that does not affect global heat content while at longer time-scales the pattern is dominated by additional heat gained at the surface changing water-masses. These longer time-scale processes commit the ocean to *virtually certain* warming until 2040 regardless of emissions. After 2040, similar spatial patterns of present warming will continue to warm at a rate dependent on present and future emissions (*high confidence*). Between 1996-2014 and 2100, the global mean thermosteric sea-level rise associated with ocean heat uptake is *likely* to be 0.15 (0.13-0.17) m for SSP1-2.6, 0.19 (0.17-0.22) m for SSP2-4.5, 0.24 (0.23-0.26) m for SSP3-7.0, and 0.29 (0.28-0.31) m for SSP5-8.5. {9.2.2.1; 9.2.4; 9.6.1; 9.6.3.2}

Surface-intensified temperature and salinity changes have increased the stability of the upper ocean stratification. It is *virtually certain* that near-surface ocean salinity spatial patterns are more pronounced since 1950 consistent with an enhanced hydrological cycle, and it is certain that global sea-surface temperature has increased since the beginning of the 20th century at a *very likely* rate of $0.053 \pm 0.002^\circ\text{C}$ per decade which accelerated to $0.087 \pm 0.007^\circ\text{C}$ per decade since 1979. New analysis shows that surface ocean stratification has increased from 1970 to 2018 at the rate of 5-20% per decade, more than ten times higher than reported by SROCC (*medium confidence*). Observed near-surface ocean changes and associated regional patterns are predicted to persist at a rate dependent on emissions (*high confidence*). {9.2.1.1; 9.2.1.2; 9.2.1.4}

Ocean circulation

Robust observed changes in the Atlantic Meridional Overturning Circulation (AMOC) are not presently linked to anthropogenic causes, but the AMOC will *likely* decline in response to human-induced climate change for all scenarios. The weakening of the AMOC between 2007 and 2011 is within the range of natural variability inferred from reanalysis, reconstructions and model simulations of the historical period (*high confidence*). Projected AMOC decline by 2100 is only weakly dependent on emission scenario ranging from 40% in SSP1-1.9 to 50% in SSP5-8.5 (*medium confidence*). An abrupt collapse of the AMOC before 2100 is *unlikely* with deep uncertainty beyond 2100 associated with meltwater fluxes from Greenland and Antarctica. Detecting long-term changes in other large-scale circulation systems remains challenging due to observation limitations, but the Southern Ocean overturning circulation, eastern boundary upwelling systems and Indonesian Throughflow are predicted to respond to changes in the atmospheric circulation. (*high confidence*) {9.2.3.1; 9.2.3.2, 9.2.3.4, 9.2.3.5}

Sea ice

Arctic sea ice retreated and seasonal area declined consistently since 1979. The Arctic Ocean will *likely* become practically ice-free in September for the first time before 2050 for all emission scenarios. No significant trend in overall Antarctic sea ice is detected from 1979 to 2018. Arctic summer sea-ice area during the last decade was the lowest since at least 1850, and humans have *very likely* caused at least half of Arctic summer sea ice loss since 1979. Arctic sea-ice area in all months is linearly related to global mean

atmospheric temperature and cumulative anthropogenic CO₂ emissions, which implies no tipping point for the loss of Arctic summer sea ice. The Arctic Ocean is *likely* to become sea-ice free throughout September before cumulative future anthropogenic CO₂ emissions reach 1000 GtCO₂. (*high confidence*) Slight Antarctic regional increases or decreases in ice area result from regional wind forcing (*medium confidence*), and projections require accurate modelling of regional variations (*high confidence*). Limited model resolution and poorly understood regional processes infer *low confidence* in regional projections of Antarctic sea-ice area. {9.3.1.1; 9.3.2.1; 3.4.1}

Ice sheets

It is *virtually certain* that the Greenland ice sheet has lost mass since the 1990s and will continue to lose mass. Greenland was *likely* close to mass balance in the 1990s and there is *high confidence* that annual mass losses have been consistently negative since the early 2000s. Over the period 1992-2018, Greenland *likely* contributed 0.0106 ± 0.0009 m to sea level rise with surface melt slightly exceeding losses through ice flow. Greenland ice mass loss under all emission scenarios will be mostly due to increased surface melt (*high confidence*), and it is *likely* that the Greenland ice sheet will contribute 0.07 (0.03 to 0.12) m and 0.13 (0.09-0.19) m to GMSL by 2100 under SSP1-2.6 and SSP5-8.5 respectively. {9.4.1.1; 9.4.1.2; 9.6.3.2}

It is *very likely* that the Antarctic Ice Sheet has lost mass since at least the early 1990s and *likely* that it will lose mass by the end of the century under all emissions scenarios. The grounded Antarctic Ice Sheet has *likely* contributed 0.0069 ± 0.0014 m to sea level rise over 1992-2018 and loss has accelerated over the last decades (*medium confidence*) dominated by ice discharge over the West Antarctic Ice Sheet and the Antarctic Peninsula. Ice shelf basal melting dominates current dynamical losses and will remain the dominant driver of West Antarctic mass losses (*high confidence*). Antarctic snowfall will *likely* increase by about 4-8% per °C of regional temperature change, partially compensating for dynamic losses. It is *likely* that the Antarctic ice sheet will contribute 0.12 (0.00-0.26) m to GMSL by 2100 with little scenario dependence, but there is deep uncertainty regarding the Antarctic contribution under the high emissions scenarios. {9.4.2.1; 9.4.2.2, 9.6.3.2; Box 9.3}

Glaciers

Current glacier retreat is global and unprecedented since 1850 and under future warming glaciers will lose mass (*very high confidence*). Glaciers *likely* contributed 0.025 ± 0.018 m to sea level rise between 1971 and 2016. Global glacier mass loss since the last decades of the 20th century cannot be explained without human induced warming (*high confidence*). Globally, glaciers are not in equilibrium with current climate and will diminish even if climate stabilizes (*very high confidence*). The magnitude and timing of future mass loss remains uncertain due to scenario uncertainty, low resolution global climate model (GCM) forcing, model oversimplification, and limited observations for calibration (*medium confidence*). Under continued warming glaciers will be lost globally and regions with small glaciers, such as the European Alps and low latitude mountains, will lose most or all glaciers by 2100 (*high confidence*). It is *likely* that glaciers will contribute 0.07-0.11 m and 0.11-0.18 m to GMSL by 2100 under SSP1-2.6 and SSP5-8.5, respectively [PLACEHOLDER TO UPDATE TO EMULATED GLACIERMIP PHASE 2 RESULTS FOR PROJECTIONS]. {9.5.1.1; 9.5.1.3; 9.6.3.2}

Permafrost

Arctic and mountain permafrost has been lost over recent decades and globally permafrost has warmed by 0.3°C in the last decade (*medium confidence*). Further global warming will lead to near-surface permafrost volume loss (*high confidence*). Permafrost warming is *very likely* a pan-Arctic phenomenon. Active layer thickness increase is a pan-Arctic phenomenon, subject to interannual variations. Global permafrost volume in the top 3 m will decrease by about 25 ± 5 % per °C if global air temperature remains below 4°C above preindustrial levels. (*medium confidence*) {9.5.2.1; 9.5.2.3}.

Snow cover

Northern hemisphere spring snow cover has been decreasing since at least 1978 (*very high confidence*), and the snow onset season has been delayed (*medium confidence*). Further decrease of Northern Hemisphere seasonal snow cover is *virtually certain*. There is *medium confidence* that hemispheric spring snow-cover loss trend extends back to the 1920s. Northern hemisphere spring snow mass has been decreasing since at least 1981 and mountain snow cover has decreased since 1950 globally. Northern hemisphere spring snow area changes are proportional to northern hemisphere extratropical land surface air temperature changes at $(-1.9 \pm 0.9) \times 10^6 \text{ km}^2 \text{ per } ^\circ\text{C}$ and projected snow cover sensitivity to GSAT on multi-decadal time scales of about $-2 \times 10^6 \text{ km}^2 \text{ per } ^\circ\text{C}$ from November to May. (*high confidence*) {9.5.3.1; 9.5.3.3}

Sea Level

Between 1900 and 2018 global mean sea level (GMSL) has risen by 0.19 m (*likely* 0.15-0.22 m), mostly from ocean thermal expansion and glacier melt. For 1971-2018, all major contributions to energy and sea level change are known and consistent (*high confidence*). Since 1993, the *very likely* observed GMSL change of 3.16 (2.79-3.53) mm yr⁻¹ is consistent with the sum of the independently determined contributions to sea-level change. GMSL increased faster since 1900 than any century over the last three millennia, and GMSL change has accelerated since the late 1960s, at an average rate of 0.084 (0.059-0.090) mm yr⁻² over 1993-2015. It is *very likely* that anthropogenic activities are responsible for more than half the observed GMSL change since the 1970s. (*high confidence*). {9.6.1; Cross-Chapter Box 9.2; 3.5.3.2 Box 7.2}

GMSL is *likely* to increase by 0.33-0.64 m under SSP1-2.6, 0.40-0.71 m under SSP2-4.5, 0.51-0.81 m under SSP3-7, and 0.60-0.90 m under SSP5-8.5 between 1995-2014 and 2100. GMSL is *extremely unlikely* to increase by less than 0.2 m or more than 0.9 m under SSP1-2.6 and SSP 2-4.5. It is *virtually certain* that GMSL will continue to rise beyond 2100, with GMSL rise due to thermal expansion, glacier melt, and Greenland ice sheet melt highly dependent on emissions scenario after 2050. By 2300, GMSL rise under SSP1-2.6 will *likely* be 0.3-2.9 m and *extremely likely* be below 4.7 m while GMSL rise under high emissions scenarios exhibits *deep uncertainty*. {9.6.3.1, 9.6.3.3; 9.6.3.5; Box 9.3}

Regional sea-level rise rate increased fastest in the Indo-Pacific, Northwest Pacific and subtropical North Atlantic over 1993-2015 and the projected change is largest in [PLACEHOLDER FOR REGIONAL PROJECTIONS]. Large uncertainties remain in closing the sea-level budget on regional and local scales and attribution of regional sea-level change to anthropogenic forcing also remains difficult. The anthropogenic signal in regional sea-level change will have emerged over 50% of the ocean area by 2040. (*medium confidence*) {9.6.1.4; 9.6.1.5; 9.6.3.3}

Marine and coastal extremes

Marine extremes have become more frequent and this trend will continue into the future. Marine heatwaves have become more frequent (*high confidence*) and persistent (*medium confidence*) since 1982 and are *very likely* to increase in frequency, duration, spatial extent and intensity in the coming decades. Changes in extreme still water levels over the 20th century were primarily driven by regional relative sea level change increasing median high-tide flooding frequency by 165%. By 2050, 1% average annual probability extreme water levels are projected to occur 11 and 14 times more frequently under SSP1-2.6 and SSP5-8.5 respectively, becoming an annual event at about 20% of the tide gauge stations. (*high confidence*) {Cross-Chapter Box 9.1; 9.6.4.1; 9.6.4.2}

Ocean, cryosphere and sea-level change in terms of the level of global warming

If future warming levels remain between 1.5 and 2°C for decades or longer, the Arctic Ocean will *likely* remain partly sea-ice covered all year round in most years; all small low latitude glaciers will *very likely* melt; however, 10-60% of glacier volume will remain in the polar regions; Northern hemisphere spring snow cover extent decrease will *likely* remain below $3 \cdot 10^6 \text{ km}^2$; permafrost volume decrease in the top 3 meters

1 will remain limited to less than 40%; the Greenland and the West Antarctic ice sheet will decline (*medium*
2 *confidence*). In pathways leading to about 2°C of warming, GMSL rise is *likely* to be 0.33-0.65 m between
3 1996-2014 and 2100. The committed sea level after two millennia will be 1 to 3 m GMSL rise per °C peak
4 warming (*low confidence*). {9.3.1.1; 9.4.1; 9.4.2; 9.5.1, 9.5.2; 9.5.3; 9.6.3.4; 9.6.3.5}

6 **For future warming levels between 2 and 3°C**, there is *medium confidence* that the Greenland Ice Sheet
7 will pass a threshold where long-term mass loss becomes irreversible over centennial timescales. There is
8 *high confidence* that Antarctic ice shelf basal melting will increase, but *low confidence* in the projected melt
9 rates. {9.4.1.2; 9.4.2.2}

11 **If future warming levels remain between 3 to 5 °C for decades or longer**, the Arctic Ocean will *very*
12 *likely* become sea-ice free throughout several summer months in most years and the Antarctic Ice Sheet will
13 continue to lose mass due to ocean-forced dynamic thinning and acceleration (*high confidence*). The
14 Greenland ice sheet will disappear after multiple millennia; most of the glaciers in the world will *very likely*
15 melt, only the largest glaciers will persist at 10-20% of their present size; Northern Hemisphere winter snow
16 cover will *likely* be less than 40·10⁶ km² and spring snow cover extent decrease will *likely* exceed 6·10⁶ km²;
17 permafrost volume decrease in the top 3 meters will exceed 60% compared to the present (*medium*
18 *confidence*). Furthermore, for pathways leading to about 4°C of warming, GMSL rise is *likely* to be 0.54-
19 0.85 m between 1996-2014 and 2100. The committed sea level after two millennia will be about 16 m
20 GMSL rise for 4°C peak warming (*low confidence*). {9.3.1.1; 9.4.1; 9.5.1; 9.5.2; 9.5.3; 9.6.3.4; 9.6.3.5}

9.1 Introduction

This chapter provides a holistic assessment of the physical processes underlying global and regional changes in the ocean, cryosphere and sea level, as well as improved understanding of human attribution and future projections since AR5 and SROCC. The ocean and cryosphere (defined as the frozen components of the Earth system such as sea ice, ice sheets, ice shelves, glaciers, permafrost and snow) exchange heat and freshwater with the atmosphere and each other (Figure 9.1a). In a warming climate, the combined effects of thermal expansion of seawater and melting of the terrestrial cryosphere, augmented by changes in the storage of water on land, result in global mean sea level rise (Box 9.1).

Observational advances (instrumental and reconstructions)

Advances in instrumental observations have improved understanding of the state and key processes in the ocean and cryosphere assessed in this chapter. Ocean observations have increased in both spatial and temporal coverage from the satellites measuring surface ocean properties (e.g., temperature, salinity, colour), biogeochemical (e.g., pH, dissolved oxygen) sensors in ships, mooring buoys and Argo floats mapping ocean changes across depths, and ocean circulation monitoring programs such as those measuring the Atlantic Meridional Ocean Circulation (AMOC). Similarly, progress has been made in observing sea ice and ice sheets via satellite data and improved retrieval algorithms. Sea level observations continue to improve through tide gauge, altimeter and gravimeter data.

Paleoclimate reconstructions extend the coverage of instrumental records, and together with paleoclimate modelling, play an important role in understanding changes in past ocean (e.g., sea-surface temperature, salinity, ocean heat content, thermohaline circulation), cryosphere (e.g., Arctic sea ice area, Antarctic ice shelves), and sea level (Annex II). Paleoclimate reconstructions from the Little Ice Age (LIA), the Last Millennium, mid-Holocene (MH), Last Glacial Maximum (LGM), Last Interglacial (LIG), Mid-Pliocene Warm Period (MPWP) and Early Eocene Climate Optimum (EECO) are particularly relevant to this chapter (Cross-Chapter Box 2.1). These episodes inform long-term responses of polar ice sheets to amplified warming (Dowsett et al., 2012; Rugenstein et al., 2016; Stap et al., 2017, 2018; Graversen and Langen, 2019; Smith et al., 2019), and their potential equilibrium contributions to GMSL (Box 9.2). It remains challenging, however, to quantify past rates of change with sufficient accuracy to inform projections of the near-term (decadal to centennial) future ice sheet and sea-level change. Models and geological reconstructions nonetheless consistently indicate that ice sheet equilibrium responses to warming are substantially greater (in terms of mass loss) than transient (decadal to centennial) changes, suggesting that the ice sheets are currently in disequilibrium with the warming climate. Reaching a new equilibrium state, including the slower responding elements of the ocean and cryosphere, will take thousands of years (Clark et al., 2016).

Modelling advances

In addition to the CMIP6 DECK (Diagnostic, Evaluation and Characterization of Klima) experiments utilising shared socioeconomic pathways (SSPs) (Eyring et al., 2016; O'Neill et al., 2017), a number of Model Intercomparison Projects (MIPs) inform this chapter including ScenarioMIP, HighResMIP, OMIP, FAFMIP, SIMIP, PMIP4, ISMIP6, and GlacierMIP (see details in Annex III). The fourth Paleoclimate Model Intercomparison Project (PMIP4) provides the relationships between the climate-ice sheet system and sea level during the LIG, LGM, and MPWP in this chapter. Paleoclimate modelling outputs from the LIG epoch are used in ISMIP6 ice sheet experiments to quantify past sea level changes. A particular advance relevant to this chapter is the increase in ocean resolution in the HighResMIP and OMIP2 experiments (section 9.2); the majority (but not all) of the DECK experiments have ocean resolutions of approximately 1 degree which does not resolve the Rossby radius (the length scale of the most energetic form of ocean internal variability, mesoscale eddies, and also the typical scale of boundary currents) and mesoscale eddies are parameterized (Hallberg, 2013; Hewitt et al., 2017), while HighResMIP and OMIP2 include models with ocean resolutions of approximately 1/4 degree (where mesoscale eddies are present) and 1/10 degree (where simulations are rich with mesoscale eddies) (Griffies et al., 2015; Chassignet et al., submitted).

[START FIGURE 9.1 HERE]

Figure 9.1: Components of ocean, cryosphere and sea level assessed in this chapter. (a) Schematic of processes. (b) Geographic distribution of ocean and cryosphere components (numbers indicate glacierized regions).

[END FIGURE 9.1 HERE]

Chapter outline

The outline of this chapter (Figure 9.2) explores changes in ocean, cryosphere and sea level (Sections 9.2 – 9.6). With a focus on understanding the processes driving ocean, cryosphere and sea level change on timescales from decades to millennia, this chapter sits between the discussion of large-scale observed trends, human attribution and future projections in Chapters 2- 4, and regional perspectives contained in Chapters 10-12 and the Atlas. Changes in ocean and cryosphere processes modify heat, hydrological and carbon cycles relevant to process-based Chapters 5, 7 and 8. The boxes explore themes across chapter 9 and with other chapters. These include: marine heatwaves (Cross-chapter box 9.1), paleo insights to ice sheet processes (Box 9.2), closing energy and freshwater budgets (Cross-chapter box 9.2). Given the deep uncertainty which exists around high-end sea level projections (i.e., experts and stakeholders do not know or cannot agree on: (1) appropriate conceptual models that describe relationships among key driving forces in a system; (2) the probability distributions used to represent uncertainty about key variables and parameters; and/or, (3) how to weigh and value desirable alternative outcomes (adapted from (Lempert et al., 2003; Abram et al., 2019; Marchau et al., 2019)), a storyline approach is adopted (Box 9.3) to assess climate and ice sheet processes linked to high and low-end sea level projections linked to ice sheet processes. Other themes running through the chapter are the commitment to ongoing change in the ocean, cryosphere and sea level due to the long response times of many parts of the system and the changes expected in the system at different warming levels. Section 9.7 highlights the robustness and limitations to the assessment in this chapter.

[INSERT FIGURE 9.2 HERE]

Figure 9.2: Visual guide to chapter 9 with links to relevant assessments in other chapters and common themes across the chapter.

[END FIGURE 9.2 HERE]

[START BOX 9.1 HERE]

BOX 9.1: Key processes driving sea-level change

Sea-level change (SLC) arises from many processes – in the ocean, on land, in the atmosphere, in the cryosphere and in the solid earth – acting on a range of spatial and temporal scales (Figure 9.1). **Relative sea-level (RSL) change** is the change in local mean sea level relative to the land, as measured by instruments that are fixed to the Earth’s surface, such as tide gauges. This reference frame is most relevant when considering flood hazards. In contrast, **geocentric sea-level change** is the change in the height of the water column with respect to the Earth’s centre of mass and is relevant when observing sea-level change with instruments from space. In this Box, a brief summary is provided, adopting standardized terminology (Gregory et al., 2019). The box starts with processes that act on a global scale, before defining processes acting on regional and local scales.

Global processes

Global mean sea-level (GMSL) change is the increase or decrease in the volume of the ocean divided by the

ocean surface area. It is driven by changes in ocean density through temperature changes (*global mean thermosteric sea-level change*, Section 9.2) and changes in the ocean mass as a result of changes in the cryosphere or the terrestrial water storage (*barystatic sea-level change*, Sections 9.4, 9.5, Chapter 8).

Thermosteric sea-level rise (or *thermal expansion*, Section 9.2) occurs as a result of ocean warming, which reduces the ocean density and increases the volume per unit of mass. On a local scale, ocean water density can also change due to salinity changes, called *halosteric sea-level change*: higher salinity leads to higher density and thus sea-level fall. While thermosteric sea-level rise is an important process on a global mean scale, halosteric change is only important on local scales. **Steric sea-level change** is the combination of *thermosteric sea-level change* and *halosteric sea-level change*.

The Greenland and Antarctic ice sheets (GrIS and AIS; Section 9.4) are the largest reservoirs of frozen freshwater available on Earth, and therefore potentially the largest contributors to sea-level rise. The mass balance of an ice sheet is determined by changes in ice dynamical processes and changes in the surface mass balance (SMB), which is the balance between the accumulated snow and the melted or sublimated snow and ice. Adjacent to the AIS, ice shelves (which do not directly contribute to SLC) melt from below or disintegrate from surface meltwater penetrating crevasses and reduce buttressing of the grounded outlet glaciers (which do contribute to SLC) flowing onto the ice shelves. The GrIS is expected to mainly contribute to SLC due to SMB changes, with elevation and albedo feedbacks that amplify surface melting at the margins, while the dynamic contribution is expected to decrease due to landward retreat of marine-terminating glaciers.

Glaciers (Section 9.5) contribute to sea-level change by changes in their mass, primarily as a result of surface mass balance changes. As the atmospheric temperature increases or the radiative balance on the glacier surface changes due to (for example) changes in cloud cover or darkening of the snow surface, increased melting occurs. The melted snow or ice does not all flow directly into the ocean, but in some places the meltwater refreezes in the snow and firn (although the refreezing capacity of firn deteriorates with increasing melt) or is stored in proglacial lakes for extended periods. Glacier mass loss also occurs when ice flows to lower elevations or into the ocean (tidewater glaciers).

Terrestrial water storage (TWS, Chapter 8, Section 8.2.3) includes all surface water, soil moisture, groundwater storage and snow, but excludes water stored in glaciers and ice sheets. Changes in TWS can be caused either by direct human intervention in the water cycle or by climate variations. The main anthropogenic processes that affect TWS are storage of water in reservoirs by building dams in rivers and groundwater extraction from groundwater reservoirs for consumption and irrigation. Examples of TWS changes caused by climate variations, some of which can also be indirectly affected by anthropogenic influences, include drainage of endorheic lakes and wetlands, deforestation, and changes in soil moisture, permafrost and the snowpack.

Regional and local processes

Ocean dynamic sea-level change (Section 9.2) refers to the change in mean sea level relative to the geoid, and is associated with the circulation and density-driven changes in the ocean. By convention, dynamic sea-level change is inverse-barometer corrected (i.e., the effect of the hydrostatic depression of the sea surface by atmospheric pressure changes is removed). Changes in ocean currents occur due to variations in heating and cooling, variability in winds and changes in seasonal to annual averaged air temperature and humidity. On timescales of 10-30 years, dynamic sea level variability is dominated by decadal variations and annular modes (section 9.2.4).

Changes in Earth Gravity, Earth Rotation and viscoelastic solid-Earth Deformation (GRD) result from the redistribution of mass between terrestrial ice and water reservoirs and the ocean. Terrestrial mass loss leads to elastic solid Earth uplift and a local relative sea-level fall (within ~2000 km of the shrinking terrestrial reservoir). Farther away (more than ~7000 km from the source of mass loss), RSL rises more than GMSL, due, to first order, to gravitational effects. Earth deformation associated with adding water to the oceans and a shift of the Earth's rotation axis towards the terrestrial mass source impart second-order effects that

increase spatial variability of the pattern globally. **Glacial Isostatic Adjustment (GIA)** is the ongoing GRD response of the solid Earth and the global gravity field to past changes in the distribution of ice and water on Earth's surface. On a timescale of decades to tens of millennia following mass redistribution, Earth's mantle flows viscously as it evolves toward isostatic equilibrium, causing solid Earth movement, which can result in local sea-level variations.

Vertical land motion (VLM) is the change in height of the land surface or the sea floor and can have several causes (Chapter 12) in addition to elastic deformation associated with contemporary GRD and viscoelastic deformation associated with GIA. Subsidence can, for instance, occur through compaction of alluvial sediments in deltaic regions, removal of fluids such as gas, oil, and water, or drainage of peatlands. Tectonic deformation of the Earth's crust can occur as a result of earthquakes and volcanic eruptions.

Extreme sea levels (ESLs) arise from a combination of surges, tides, and waves, superimposed on the local sea-surface height (Section 9.6.3). The combination of tide and surge is referred to as *still water level* and the combination of tide, surge, and wave effects is referred to as *total water level*. The wave contribution (wave run-up, the maximum height waves reach at the beach) can be characterized as the combination of the wave setup (time-mean sea-level elevation due to wave energy dissipation) and the swash (vertical displacement induced by individual waves). RSL change affects ESL directly by shifting the base water levels and indirectly by modulating propagation of tides, waves, and/or surges due to increased water depth. In addition, ESLs may change due to changes in the frequency, tracks, or strength of weather systems, or due to anthropogenically induced changes such as the modification of coastlines.

[END BOX 9.1 HERE]

9.2 Oceans

9.2.1 Ocean surface

9.2.1.1 Sea Surface Temperature (SST)

It is certain that global sea-surface temperature has increased since the beginning of the 20th century (*very high confidence*) (Hartmann et al., 2013). The global mean SST has increased at a *very likely* rate of $0.053 \pm 0.002^\circ\text{C}$ (90% confidence interval) per decade from 1880 to 2018 (estimated from the global interpolated SST dataset HadISST and updating AR5 estimate of $0.042 \pm 0.007^\circ\text{C}$ per decade from 1880 to 2012 using the same dataset). Global mean and regional SST observations over the past decades are more detailed and robust due to the advent of wide satellite coverage measuring surface temperature especially since 1979. The estimated global mean SST warming is $0.087 \pm 0.007^\circ\text{C}$ per decade (90% confidence interval) from 1979 to 2018 (updating AR5 estimate of $0.072 \pm 0.024^\circ\text{C}$ per decade from 1979 to 2012), faster than the mean warming rate over the past century. Associated with the observed global SST trend, regional extreme temperature events, known as marine heatwaves have become more frequent (*high confidence*) and longer in duration (*medium confidence*; Cross-Chapter Box 9.1). [PLACEHOLDER FOR GLOBAL MEAN ESTIMATES CONSISTENT WITH CHAPTER 2]

While a positive SST trend is evident globally, some regions have experienced quicker or slower warming than the global average (Figure 9.3). For instance, the Western Boundary Current of the subtropical gyres have experienced enhanced warming compared to the global mean over the past century (*medium confidence*) (Wu et al., 2012; Yang et al., 2016a), though there is still *low confidence* on the underlying dynamical reasons due to observational challenges and low agreement in the current position and intensity changes (Section 9.2.3.4). Since 1950 the tropical ocean has been warming faster than other regions, with the tropical Indian ocean warming even faster than the rest of the tropical ocean basins (*high confidence*) (Figure 9.3), due to atmosphere ocean feedback (*medium confidence*), though there is only *low confidence* on the exact mechanism driving enhanced Indian ocean warming (Rao et al., 2012; Roxy et al., 2014; Hu and Fedorov, 2019). In contrast, the eastern Pacific Ocean, North Atlantic Ocean, and Southern Ocean, have warmed

slowly or cooled due to regional ocean dynamics (Figure 9.3). In the Pacific Ocean, there is *medium confidence* that the eastern equatorial Pacific warming has been slower due to mean upwelling renewing surface water with deeper water-masses unaffected by atmospheric warming (Clement et al., 1996; Cane et al., 1997; Luo et al., 2017). Similarly, there is *medium confidence* in the slower than global average warming in the subpolar Southern Ocean since 1950s, despite sparse observations, because this pattern is consistent with the upwelling around Antarctica renewing surface water with deeper water-masses unaffected by atmospheric warming (Frölicher et al., 2015; Marshall et al., 2015b; Armour et al., 2016). Since the 1970s, the temperature around the subpolar Southern Ocean has decreased, which contrasts with marked warming directly northward of it (Fan et al., 2014; Armour and Bitz, 2015; Jones et al., 2016b) (*medium confidence*), but the cause and attribution of this cooling to anthropogenic change or internal variability remains unclear (Bintanja et al., 2013; Fan et al., 2014; Marshall et al., 2014; Ferreira et al., 2015; Purich et al., 2016a, 2018; Kostov et al., 2017, 2018; Zhang et al., 2019a; Haumann et al., submitted). In other regions, the impact of internal variability of the ocean associated with coupled ocean-atmospheric modes of variability (Cross-Chapter Box 3.1), or variability of circulation regimes is clearer. For instance, in the North Atlantic SST variability is related to the AMOC and this relationship is discussed in Section 9.2.3.1. Decadal-scale modulation of the long-term tropical ocean warming can affect global SST change, such as over the period 2001-2014 where an anomaly in Pacific trade winds and an associated increased ocean transport from the Pacific to the Indian Ocean (*high confidence*) have led to a cooling trend in the Pacific and basin-wide warming trends in the Indian Ocean (*medium confidence*) (Collins et al., 2019).

In AR5, a marginal improvement in CMIP5 climate model SST biases were noted compared to CMIP3 models in AR4, with a reduction in the magnitude of biases. In several regions large SST biases are symptomatic of errors in the representation of important processes, such as dynamics in the equatorial Pacific and Atlantic, Southern Ocean, Gulf stream (Flato et al., 2013). In these regions, shortcomings in process representation can make it difficult for models to reproduce observed historical trends giving reduced confidence in regional projections in those sectors. Common regional biases in SST or historical SST trends arise from multiple causes not always directly linked to the sole representation of the ocean (*high confidence*), including errors in the representation of long term historical trends in equatorial winds (McGregor et al., 2014; Kociuba and Power, 2015), misrepresentation of the forced equatorial ocean response (Karnauskas et al., 2012; Kohyama et al., 2017; Coats and Karnauskas, 2018), thermocline depth errors, errors in atmospheric model cloud-related short-wave (Hyder et al., 2018), biases in circulation variability (Wang et al., 2014a). In CMIP6, the mid-latitude biases in the northern hemisphere are improved in the multi-model mean and the inter-model standard deviation of the zonal mean SST error is significantly decreased south of 50°N compared to CMIP5, though biases in equatorial regions remains essentially unchanged (Section 3.5.1.1) (Figure 9.3). Despite some persisting regional biases, CMIP6 coupled climate models generally reproduce observed SST trends over the past century (Figure 9.4), and there is *high confidence* in their ability to inform on future large-scale SST changes.

[START FIGURE 9.3 HERE]

Figure 9.3: Sea Surface Temperature (SST) and its changes with time. (Top) Time series of global mean SST anomaly relative to 1970-1980 climatology. Shown are observations (HadISST), and multi-model means from the CMIP historical and SSP experiments, and from the HighResMIP experiment. (Bottom panels) Maps of observed SST (top left; 1995-2014 climatology HadISST), bias of CMIP (middle left) and HighResMIP (bottom left) over this period. Also shown are 1950-2014 historical SST changes from observations, CMIP and HighResMIP (middle column), and 2015-2079 projected changes under SSP5-8.5 for the CMIP and HighResMIP ensemble (right column).

[END FIGURE 9.3 HERE]

The global mean SST will increase during the 21st Century at a rate that depends on the future emission scenario (*very high confidence*). SROCC assessed a *very likely* range of global mean SST increased from 1986-2005 to 2081-2100 of 0.73°C under RCP2.6 (0.20°C to 1.27°C), and 2.58°C under RCP8.5 (1.64°C to 3.51°C) based on a suite of CMIP5 models. Based on CMIP6 models, estimates are revised for the period

1995-2014 to 2081-2100 to an increase of 0.9°C under SSP1-2.6 (0.5°C to 1.3°C); 1.6°C under SSP2-4.5 (1.0°C to 2.1°C); 2.4°C under SSP3-7.0 (1.6°C to 3.1°C); and 3.1°C under SSP4-8.5 (2.2°C to 4.0°C) (Figure 9.4). Coupled models project that Southern Ocean SSTs will increase under ongoing greenhouse gas forcing as upwelled waters are warmed over centuries to millennia (Li et al., 2013b; Armour et al., 2016), eventually warming by more than the tropics on centennial timescales (*high confidence*), but there is *low confidence* that this warming will emerge this century (Section 4.5.2.1, Section 7.4.4.2). Similarly, there is *medium confidence* that the observed pattern of SST change in the tropical Pacific ocean will eventually transition to a pattern similar to El Niño SST pattern on century timescales due to a weakening of the Walker circulation (Vecchi et al., 2008), weak evaporative cooling in the western equatorial Pacific (Hartmann and Michelsen, 1993; Knutson and Manabe, 1995; Xie et al., 2010; Luo et al., 2015), zonal gradients in low cloud feedbacks (Meehl and Washington, 1996), and increasing temperature of upwelled waters owing to enhanced warming in extratropical subduction regions (Barreiro and Philander, 2008; Fedorov et al., 2015) (Section 7.4.4.3). Warming of the North Atlantic subpolar gyre is projected to continue being slower than surrounding regions as AMOC further declines under greenhouse gas forcing (Section 9.2.3.1), concurrent with rapid warming in the Gulf Stream (Cheng et al., 2013) (Figure 9.4).

SST changes in the geologic past illustrate the potential for large changes as the system approaches equilibrium. For example, during the mid-Pliocene Warm Period (MPWP), when atmospheric CO₂ was near late 20th century values (Chapter 2), GMST were 1.9-3.6°C warmer than pre-industrial (Dowsett et al., 2012, 2013, 2016), substantially warmer than today (Chapter 2), but within the range of multi-millennial future projections that approach equilibrium for the lower integrated emissions scenarios of 1280GtC or 2560GtC (Clark et al., 2016). In the tropics, MPWP warming was modest but variable (-2 to +4) (Dowsett et al., 2013) but significant in the western warm pool, refuting hypothesized air-sea thermostat mechanisms (O'Brien et al., 2014), debatably (Brierley et al., 2015). Warming was much greater and seasonality lower at high northern latitudes (4-12°C north of 40°N), especially in the North Atlantic (Dowsett et al., 2013). This finding supports the concept of polar amplification, but implies that over thousands of years, regions are capable of much more warming than projected for the 21st century. Inferences that reduced MPWP thermal gradients were supported by enhanced overturning circulation (Burls et al., 2017) contrast with possible reduction of AMOC recently and in near-term projections, opening the possibility that slowdown of AMOC may be a near-term transient response that will ultimately reverse (Jansen et al., 2018). Although some models do not reproduce North Atlantic warming (Hunter et al., 2019), meridional thermal gradients can be changed by varying vertical mixing (Wiebe and Weaver, 1999; Krasting et al., 2018; Lohmann, 2019) or by changing key flows such as net freshwater transport through Bering Strait (Otto-Bliesner et al., 2017; Song et al., 2018), or as a response to Arctic sea ice loss (Howell et al., 2014). MPWP warming was less extreme in the Southern Ocean (1 to 4°C), buffered by upwelling of cooler subsurface waters as today (Dowsett et al., 2013) (*medium confidence*).

9.2.1.2 Sea Surface Salinity (SSS)

There is *very high confidence* that climatological average ocean salinity strongly resembles the surface evaporation minus precipitation (E-P) flux with evidence extending back to the early 20th century (Wüst, 1936), reflecting the long-term balance between ocean advection and mixing, ocean-cryosphere-land interactions, and E-P fluxes that maintain local salinity gradients (Durack et al., 2013; Zika et al., 2015). Since AR5, new analysis and consistency between a growing number of observation-based studies provide stronger evidence for increased contrast between high- and low- sea surface salinity regions since the 1950s. In response, we revise the AR5 reported confidence in such change from *very likely* to *extremely likely* (Section 2.3.3.2). At basin-scales, this report confirms the conclusion of AR5 and SROCC that it is *very likely* the Atlantic has become saltier and the Pacific and Southern oceans have freshened over the period of historical observational coverage (Section 2.3.3.2).

Detecting and quantifying trends at local and regional scales however remains challenging due to limitation of data coverage and the sensitivity to the analysis technique chosen, so we have only *low to medium confidence* when focusing on changes in specific regions. At a regional scale, the Western Pacific Warm (and fresh) Pool and North Pacific have freshened from the 1950s, while the subtropical Atlantic salinity

maxima have increased (Cravatte et al., 2009; Durack and Wijffels, 2010; Skliris et al., 2014) (Section 2.3.3.2). As assessed in SROCC with *medium confidence*, the surface subpolar Southern Ocean has freshened at a rate of $0.0011 \pm 0.0004 \text{ yr}^{-1}$ in the upper 100 m since the 1960s, and freshened even more over the Antarctic continental shelves (except along the Western Antarctic Peninsula), where freshening of up to 0.01 yr^{-1} is observed (Meredith et al., 2019). Surface changes of SSS propagate in the interior ocean along circulation pathways (Section 9.2.3.2). While direct estimates of E-P change remain at *low confidence* (Sections 2.3.1.3.5, 8.3.1.1.1, 9.2.2.3), observational studies coupled with modelling studies suggest E-P changes drive many observed changes to near-surface salinity, on top of additional changes specific to polar regions through a combination of changes in sea-ice regime, freshwater fluxes from rivers, as well as ice-sheet melt (Meredith et al., 2019).

The direct link between near-surface fresh salinity anomalies and terrestrial rainfall is a considerable advancement since AR5 (Rhein et al., 2013), and even though only currently established at a seasonal time-scale, clearly demonstrates mechanistic links (Li et al., 2016b; Liu et al., 2018a). In addition, recent SSS observations advances (e.g., SMOS, Aquarius and SMAP satellites; SPURS-1/2 dedicated programs) highlight the regional importance of new mechanisms linking E-P and SSS: ocean surface fresh lenses due to precipitation (Boutin et al., 2013, 2014; Drushka et al., 2016); diurnal SSS cycle in the tropics (Drushka et al., 2014a); and barrier layers affecting air-sea dynamics and climate modes (Drushka et al., 2014b). In the 21st century, CMIP5 models reproduce the observed signal in the historical simulations, and then intensify in the future under increased radiative forcing (*high confidence*), so the same processes that are driving the broadening of salinity regimes (salty gets saltier, fresh gets fresher) are set to continue in the future at an increasing rate in response to continuing anthropogenic forcing (Collins et al., 2013; Silvy et al., submitted).

9.2.1.3 Air-sea fluxes and winds

As discussed in Sections 2.3.3.1, Box 7.2, and Cross-Chapter Box 9.2, estimates of the global mean (per whole earth area, not ocean area) net air-sea heat flux inferred from observed ocean warming during the Argo era (2005-2017) range from 0.54 ± 0.02 to $0.68 \pm 0.60 \text{ W m}^{-2}$ and are close to the ensemble mean of CMIP5 models ($0.68 \pm 0.02 \text{ W m}^{-2}$) (Domingues et al., 2008; Levitus et al., 2012; Cheng et al., 2017, 2019). CMIP6 models show a similar range ($0.XX \pm 0.XX \text{ W m}^{-2}$). [PLACEHOLDER UNTIL MORE OHC PROJECTIONS ARE AVAILABLE]. There is *high confidence* that this energy imbalance is due to anthropogenic forcing as it results from increasing radiative forcing (Section 7.2, 7.3). Estimates of observed ocean warming over an earlier time window (1971-2010) are smaller (Section 2.3.3.1) so fluxes may be increasing.

Direct flux measurement trends have not emerged yet as spatial (Figure 9.4), annual (Yu, 2019), and decadal (Zanna et al., 2019) variability overwhelm detection. Air-sea flux data as of AR5 were too noisy to detect even the mean fluxes directly, which remains the case. However, since AR5 updated and new surface flux products, atmospheric reanalyses, ocean reanalyses, and satellite retrieval algorithms have become available. Comprehensive comparisons (Bentamy et al., 2017; Valdivieso et al., 2017; Yu et al., 2017) use these tools to improve surface flux uncertainty estimates, while noting uncertainties in atmospheric reanalysis-based fluxes, including global energy imbalances that exceed the observed ocean warming.

Decadal trends in satellite observed global fluxes are not detectable, but regional changes are robust in some regions in both satellite observations and projections (Figure 9.4). Recent satellite-based surface flux products with improved retrieval algorithms and new satellites, e.g., J-OFURO3 (Tomita et al., 2019) and OAFux-HR (Yu, 2019), provide a complete suite of turbulent fluxes including heat, moisture, and momentum. When combined with satellite-based surface radiation from CERES EBAF (Kato et al., 2018) and precipitation from GPCP (Adler et al., 2003), full ocean-surface forcing functions are available for the satellite era (since July 1987, Figure 9.4). These products agree with sparse buoy and ship observations (Bentamy et al., 2017). While patterns agree between models and satellites in net fluxes (Figure 9.4), the trend magnitudes are substantially weaker in models, which may reflect interannual variability in the observations or too low model sensitivity at annual or interannual timescales or in trends (Bates et al., 2012). Where the patterns agree in models and observations, warming in the North Atlantic and Southern Ocean, is

consistent with the largest changes observed in the water-masses (Sections 9.2.1.1, 9.2.2.1, 9.2.3.2). The E-P flux trend patterns in the equatorial Pacific enhance the present mean pattern of wetting and drying, while elsewhere patterns are less clear, except for a general trend toward a saltier Atlantic sector and a fresher Indian sector. Thus, satellite flux observations (see also Section 8.3.1) only partly agree with the wet gets wetter, dry gets drier simplification of freshwater forcing for regional change, which clearly shows in ocean observations (Section 9.2.1.2), but we still have *low confidence* in trends from E-P satellite flux observations. Projections (Figure 9.4, upper right) show that while some regions are expected to follow this pattern (XX) other regions do not (YY). [PLACEHOLDER FOR WHEN PROJECTION FRESHWATER FLUXES ARE AVAILABLE]

[START FIGURE 9.4 HERE]

Figure 9.4: Global maps of (left) observed mean fluxes (left column), and the observed (net heat flux 2001-2014 and P-E and winds 1995-2014) and projected (2015-2100) trends of these fluxes (middle and right columns respectively). The fluxes shown are (top) freshwater flux, (middle row) net heat flux, and (bottom) wind stress magnitude, with positive numbers indicating ocean freshening, warming, accelerating, respectively. Objective interpolation from CERES EBAF v4 (Kato et al., 2018), OAFlux-HR (Yu, 2019), and GPCP (Adler et al., 2003) of fluxes and flux trends.

[END FIGURE 9.4 HERE]

There is *low confidence* in long-term wind trends in most regions, but a few locations have *high confidence* trends over the scatterometer era (1992-present, Figure 9.4) of global synoptic measurements (Desbiolles et al., 2017; Yu, 2019). In the Southern Ocean we have *high confidence* that the zonal wind stress increased from the early 1980s to the 1990s, although since 1995 only the Indian Ocean sector (Figure 9.4) has exhibited *high confidence* trends (Jones et al., 2011). From 1992 to 2015, westerly winds in the North Pacific and North Atlantic weaken and projections also weaken (Figure 9.4). In the tropical Pacific Ocean a weakening trend in easterly winds and Walker circulation in the 20th century has been inferred based on observed sea level pressure data (Vecchi et al., 2006; Vecchi and Soden, 2007), yet since the 1980s the opposite change in observed winds has occurred at odds with the expected slowdown of the Walker circulation under greenhouse gas forcing and the CMIP6 projections (Figure 9.4). With *medium confidence*, the observed strengthening primarily results from a combination of sulphate aerosol forcing (Takahashi and Watanabe, 2016; Hua et al., 2018) and multi-decadal tropical Atlantic SST trends (Kucharski et al., 2011, 2014; McGregor et al., 2014; Li et al., 2015c; Chafik et al., 2016; Kucharski et al., 2016; Kajtar et al., 2017; Sun et al., 2017a), modulated by equatorial Indo-Pacific SST changes (Luo et al., 2012; Zhang and Karnauskas, 2017). Finally, middle and high latitude observed zonal wind and sea level pressure (SLP) are trending toward the positive phase of the annular modes (Northern Annular Mode (NAM) and Southern Annular Mode (SAM)), which are characterized by poleward shifts and strengthening of the westerly winds. This shift is also reproduced by most CMIP5 models (Yang et al., 2016a), although winds over the oceans in CMIP6 models show primarily only a SAM change, and weaker than the observed trend.

Air-sea flux biases result from common causes in most models. Important currents (e.g., Gulf Stream, Kuroshio) are often found in erroneous locations in models affecting SST and flux signatures (Bates et al., 2012), but their locations are improved in high-resolution ocean models (Chassignet and Marshall, 2008; Chassignet et al., 2017), and so with *very high confidence* high-resolution coupled models reduce these biases (Delworth et al., 2012; Sakamoto et al., 2012; Small et al., 2014b; Haarsma et al., 2016). Oceanic variability outside of the tropics is traditionally thought to stem either from internal variability or in response to atmospheric weather (Hasselmann, 1976), but these high-resolution coupled models show that the atmosphere responds to ocean eddy variability on small scales (Bishop et al., 2017), exhibiting coupled modes with an impact on clouds, precipitation, and boundary layers in the atmosphere and ocean (Chelton and Xie, 2010; Frenger et al., 2013). Even coarse resolution models, such as the ocean and sea-ice components used in CMIP6, show significant sensitivity to modest changes in flux forcing (Tsujino et al., submitted). These effects impact coupled variability (*high confidence*), but also may make the mean state and sensitivity of high-resolution coupled models systematically different from coarse resolution climate

models (*low confidence*). Finally, there is still considerable disagreement between different parameterizations of air-sea fluxes used in models and strong scatter in direct observations (Renault et al., 2016; Brodeau et al., 2017).

9.2.1.4 Surface Mixed Layers and Upper Ocean Stratification

Changes in surface temperature and salinity affect the upper-ocean stratification of the ocean, which alter the depth penetration of surface mixing, and has many important consequences on global heat and carbon cycles, as well as global ocean biogeochemistry and biology (Bindoff et al., 2019). The SROCC assesses that the upper 200 m of the ocean between 60°S and 60°N is *very likely* to have been stratifying since 1970 with a total stratification increase of 2.18%–2.42% of the mean stratification, from 1970 to 2017 (i.e. 0.46–0.51% per decade) (Bindoff et al., 2019). Using individual profiles rather than an observation-based interpolated product provides similar though slightly higher estimate of 3.3–6.1 % of the mean stratification, from 1960 to 2017 (i.e. 0.58–1.07% per decade) (Yamaguchi and Suga, 2019; Sallée et al., submitted). Stratification estimated in the upper 200 m however cuts across many different mixing regimes (within or below the upper ocean mixed-layer depending on location and season), so here we revise SROCC by providing a more physically consistent stratification change of the seasonal pycnocline (at the base of the surface mixed-layer). The stratification of the upper ocean pycnocline has *very likely* increased from 1970 to 2018, and there is *high confidence* that the summer pycnocline has increased by 5–20% per decade from 1970 to 2018, between 60°S and 60° (Sallée et al., submitted). There is *very high confidence* that increased stratification results from global ocean warming and high-latitude freshening being both surface intensified (Helm et al., 2011; De Lavergne et al., 2014; Rabe et al., 2014; Haine et al., 2015; Roemmich et al., 2015; Wijffels et al., 2016; Desbruyères et al., 2017; Sallée et al., submitted). There is *high confidence* that the stratification of the upper ocean winter pycnocline has increased, but shorter time-series in winter gives *low confidence* in the rates at global scale (Yamaguchi and Suga, 2019; Sallée et al., submitted). Paleo reconstructions and modeling show that the depth distribution of primary stratification may change through buoyancy fluxes in the Southern Ocean; warming reduces deep relative to shallow stratification (Ferrari et al., 2014; Jansen, 2017).

[START FIGURE 9.5 HERE]

Figure 9.5: Observed climatological mean mixed layer depth from IFREMER/LOS Mixed Layer Depth (updated from de Boyer Montégut et al., 2004) (first column), bias between observation-based estimate and 1995–2014 CMIP6 climatological mean mixed layer depth (second column), projected MLD change from 1995–2014 to 2081–2100 under SSP1-2.6 (third column) and SSP5-8.5 (fourth column) scenarios, for the DJF/JJA in N. Hemisphere/S. Hemisphere (upper) and JJA/DJF in N. Hemisphere/S. Hemisphere (lower). The mixed layer depth is the depth where the potential density is 0.03 kg m⁻³ denser than at 10m. Units: m.

[END FIGURE 9.5 HERE]

Changes in stratification have a climate-scale impact through modulating the surface mixed layer depth, which is set by a balance between dynamical mixing (e.g. wind, tides, wave, convective mixing) acting against the background stratification and restratification processes (e.g. solar and dynamical restratification). Change in mixed layer depth depends therefore on a delicate balance between an overall increasing stratification that act toward shoaling the mixed-layer, and large-scale changes in surface fluxes increasing or lowering mechanical and convective mixing (Section 9.2.1.4). Despite the large stratification increase observed at a global scale, the summer mixed-layer depth has deepened consistently over the globe at rate of 5–12% per decade from 1970 to 2018 between 60°S and 60°N (*medium confidence*) (Sallée et al., submitted). Such deepening in the summer mixed-layer depth has important consequences for surface ocean biological activity, and is consistent with observed global scale intensification of the surface winds (Section 9.2.1.3) (Young and Ribal, 2019; Sallée et al., submitted), possibly intensifying (*very low confidence*) turbulent surface layer mixing associated with wind, waves, and sub-mesoscales (Belcher et al., 2012; Buckingham et al., 2019; Sallée et al., submitted). Given the shorter observational record in winter than in

summer, there is *low confidence* in quantifying winter mixed-layer trends and disentangling long-term climate trend from decadal variability (Sallée et al., submitted).

Consistent with the current observed trend, climate models project that upper ocean stratification will continue to increase in the 21st century under increased radiative forcing (*high confidence*), primarily due to increased surface temperature and high-latitude surface freshening (Bindoff et al., 2019). The exact rates of stratification change as a function of depth is however still unclear due to uncertainty in process representation (Griffies et al., 2015; Fox-Kemper et al., 2019). While the global scale estimate of intensification is robust, regional biases in individual model (negative or positive) are sensitive to parameterizations (Griffies et al., 2015; Li et al., 2016c). Similarly, the mixed layer is filled with and sensitive to a range of small-scale processes (*very high confidence*), which are not resolved in climate models (Sullivan and McWilliams, 2010; Belcher et al., 2012; D'Asaro, 2014; Buckingham et al., 2019). The challenges parameterizing small-scale surface processes, along with biases in representing stratification, surface fluxes, and individual current position, lead to persisting mixed layer biases in most recent climate models, particularly in southern and northern hemisphere subpolar region (Figure 9.5) (Sallée et al., 2013a; Li and Lee, 2017). Despite these limitations, the current generation of climate model represent all the main features of historical mixed layer structure, giving *high confidence* in their first order water-mass formation mechanisms, and large-scale circulation features (Figure 9.5). In particular, improvements have been made in the parameterisation of mixed layer processes over the recent decade (Craig and Banner, 1994; Huang and Qiao, 2010; Fox-Kemper et al., 2011a; Belcher et al., 2012; Jochum et al., 2013; Li et al., 2016c) leading to improved mixed layer properties—such as depth, seasonality, and reduced pycnocline temperature and CFC biases—due to an improved treatment of mixing and restratification processes (Fox-Kemper et al., 2011b; Li et al., 2016c; Qiao et al., 2016; Reichl and Hallberg, 2018). CMIP6 climate model projections and theoretical understanding suggest that the upper-ocean stratification will continue to increase in the coming century (*very high confidence*), but challenges with mixed-layer with surface layer parameterisation lead to only *low confidence* that the global ocean mixed-layer will switch from current deepening to shallowing as stratification increase overcome increase in surface turbulence (Figure 9.5). CMIP6 climate model predict indeed a shallowing of the mixed-layer both in summer and winter by the end of the century under increased radiative forcing (Figure 9.5), with the exception of the subpolar ocean showing deepening of the mixed-layer as a result of sea-ice retreat (Lique et al., 2018) (Figure 9.5). The regions of largest shallowing are associated with the deepest climatological mixed-layer both in winter and summer, particularly affecting the North Atlantic and the Southern Ocean basins (Figure 9.5).

[START CROSS-CHAPTER BOX 9.1 HERE]

Cross-Chapter Box 9.1: Marine Heatwaves

Contributors: Mathias Aschwanden (Switzerland), Baylor Fox-Kemper (USA), Thomas Frölicher (Switzerland), Helene Hewitt (UK), Akm Saiful Islam (Bangladesh), Sophie Lewis (Australia), Intan Suci Nurhati (Indonesia), Jean-Baptiste Sallée (France)

Marine heatwaves (MHWs) are discrete prolonged anomalously warm water events (IPCC, 2018, 2019a) that can cause severe impacts on marine ecosystems that cascade to human societies (Collins et al., 2019). Unlike synoptic atmospheric heatwaves (Chapter 11), MHWs can occur over a vast spatial extent up to thousands of kilometres and can persist weeks to months (Bond et al., 2015; Perkins-Kirkpatrick et al., 2018). A range of processes may lead to extreme ocean temperature events, including anomalous air-sea heat fluxes and horizontal and vertical temperature advection due to currents, mixing processes, eddies or planetary waves. Interactions between local and remote processes may also play an important role (Holbrook et al., 2019).

Over the past two decades, MHWs have occurred in all ocean basins. Prominent examples include MHWs in the Mediterranean Sea (Garrahou et al., 2009), subtropical to equatorial Indo-Pacific (Herring et al., 2018), northeast Pacific (often called the ‘blob’; (Bond et al., 2015)) and northwest Atlantic (Mills et al., 2014; Frölicher and Laufkötter, 2018; Smale et al., 2019; Laufkötter et al., submitted). These events had severe and

potentially persisting impacts on marine ecosystems ranging from mass mortality of benthic communities including coral bleaching, shifts in species composition and geographical distribution, loss of biodiversity and increases in toxic algal blooms, to decline in aquaculture and fisheries catch (Cavole et al., 2016; Collins et al., 2019; Smale et al., 2019)(Chapter 12). For example, the Western Australia 2011 MHW wiped out hundreds of km's of kelp forests off Western Australia (Wernberg et al., 2013). These impacts are still evident today (Wernberg et al., 2016).

MHWs have become more frequent (*high confidence*) with more intense warming (*medium confidence*), longer duration (*medium confidence*) and larger spatial extent associated with the long-term ocean warming trend (*high confidence*) (Cross-Chapter Box 9.1, Figure 1a) (Frölicher and Laufkötter, 2018; Smale et al., 2019; Laufkötter et al., submitted). On the basis of satellite-derived SST data, the number of annual MHW days have doubled over the period 1982-2016 (Frölicher et al., 2018; Oliver et al., 2018) with increased intensity (0.04°C per decade) and spatial extent (+19% per decade) (Frölicher et al., 2018). Extending the satellite data using reanalysis SSTs shows a 54% increase in annual MHW days between the 1987-2016 period compared to 1925-1954 (Oliver et al., 2018) with the majority of this change driven by increase in mean SST (Oliver, 2019). Climate modes such as ENSO (Technical Annex, Section AVI.4) combined with Pacific Decadal Variability (PDV: Technical Annex, Section AVI.7) and Atlantic Multi-decadal Variability (AMV: Technical Annex, Section AVI.8) explain 18% of MHW frequency, 56% of duration and 36% of total days (Oliver et al., 2018). In the Pacific region, the interannual variation of MHWs is largely modulated by ENSO with increased frequency and intensity during moderate (1986/87/88, 1991/92, 2004/05, 2009/10) and strong El Niño events (1982/83, 1997/98, 2015/16). There is an increased incidence of high-severity (categories III and IV) during the strong ENSO events (Hobday et al., 2018). During the 2016 ENSO, high-severity MHWs affected about 70% of the ocean surface with only 5% of the ocean not affected by any class of MHWs (Hobday et al., 2016). The blob in the northeast Pacific during winter 2014/15 has been linked with the North Pacific Gyre Oscillation (NPGO) in 2014 and the PDV in 2015 (Di Lorenzo and Mantua, 2016).

Human induced global warming has already altered the probability of MHW occurrence. 84-90% of all MHWs that occurred between 2006 and 2015 are caused by human-induced global warming (Frölicher et al., 2018; IPCC, 2019b). The occurrence probability of the length and intensity of seven prominent MHWs, including the Northeast Pacific 2013-2015 MHW and the Northwest Atlantic 2012 MHW, have increased more than 10-fold due to anthropogenic global warming (Laufkötter et al., submitted).

[START CROSS-CHAPTER BOX 9.1, FIGURE 1 HERE]

Cross-Chapter Box 9.1, Figure 1: Regional probability ratio of marine heatwaves, the proportion by which the number of marine heatwave days per year has increased. A marine heatwave is defined as a deviation beyond the daily 99th percentile (11-day window) in the deseasonalized SST. (a) The marine heatwave probability ratio of satellite observations during 2007-2016 versus the 1982-2016 period. (b,c) CMIP6 simulated multi-model mean probability ratio in 2081-2100 versus preindustrial for the (b) SSP1-2.6 and (c) SSP5-8.5 scenarios. The hatched areas indicate permanent marine heatwaves in 2081-2100 (>360 marine heatwave days per year). These models are included: MIROC6, NESM3, CanESM5, EC-Earth.

[END CROSS-CHAPTER BOX 9.1, FIGURE 1 HERE]

Under further global warming, MHWs are *very likely* to increase in frequency, duration, spatial extent and intensity in all ocean basins, but with distinct spatial magnitudes (Cross-Chapter Box 9.1, Figure 1b-c) (Frölicher et al., 2018). Climate models are able to capture the broad features of MHWs although the relatively coarse ocean resolution of standard CMIP6 models is expected to bias the projections towards weaker and longer MHWs (*medium confidence*) (Frölicher et al., 2018; Pilo et al., 2019). Irrespective of the future greenhouse gas emission scenario, the largest changes are projected to occur in the tropical ocean and the Arctic Ocean (*high confidence*), because small seasonal and year-to-year SST variations lead to larger

changes in the probability of extreme temperatures). More moderate increases are projected for mid-latitudes, and only small increases are projected for the Southern Ocean (medium confidence). A near-permanent MHW state in many parts of the ocean is projected by the late 21st century with accelerated increases in the intensity and annual MHW (Oliver, 2019). Permanent MHWs (more than 360 days per year) are projected to occur in parts of the tropical ocean, the Arctic Ocean and around 45°S under the SSP5-8.5 scenario (hatched areas in Cross-Chapter Box 9.1, Figure 1c). Under the SSP1-2.6 scenario, the occurrence of such permanent MHWs can be largely avoided. Regional coupled models indicate that the intensity and duration of MHWs are also projected to increase in the semi-enclosed basin of the Mediterranean Sea (Darmaraki et al., 2019). MHWs that in pre-industrial climate occurred on average once every hundred to thousands of years are projected to become decadal to centennial events under 1.5°C warming and annual to decadal events under 3°C warming (Laufkotter et al., submitted).

These projected changes represent threats to the health and sustainability of marine ecosystems globally (Smale et al., 2019). While the physical understanding of MHWs and their ecological consequences is rapidly emerging, the science community is challenged by the ability to predict and monitor MHWs given the increasing pace at which they are occurring worldwide. Skilful MHW predictions may be improved by higher resolution models (Pilo et al., 2019) which will facilitate the utility of proactive decision-making.

[END CROSS-CHAPTER BOX 9.1 HERE]

9.2.2 Change in Heat and Salinity

9.2.2.1 Ocean Heat Content and Heat Transport

It is *virtually certain* that global ocean heat content has increased from 1971 to 2018 (*very high confidence*), and *likely* that it has increased from the 1870s to 1971 (*medium confidence*) (Figure 9.6). There is *high confidence* that this ocean heat content increase represents over 90% of the total excess heat accumulated in the climate system between 1971 and 2018 (Section 2.3.3.1; Section 7.2.2.2; Cross-Chapter Box 9.2). Confidence in this assessment is strengthened since AR5 and SROCC by longer observational time-series, further bias correction and understanding of uncertainty, consistent closure of both global sea level and energy budgets for the period 1971-present, and increasing lines of evidence from novel estimates of ocean warming largely independent of subsurface observations (Section 2.3.3.1; Section 7.2.2.2; Cross chapter Box 9.2). The ocean observing system has evolved over time, with varying depth coverage and uncertainty of sensors, meaning that confidence in decomposing global ocean temperature or heat content change into contribution by depth ranges depends on the time period and the depth range (Section 2.3.3.1). There is *high confidence* that the observed ocean heat content increase from 1970s to 2018 is distributed with 60% in the 0-700m layer, 30% in the 700-2000m layer and 10% in the 2000-6000m layer (Section 2.3.3.1; Table 2.7). Consistent with SROCC, the Southern Oceans is the major anthropogenic excess heat reservoirs, accounting for about 49% (39-61%) of the total excess heat energy stored in the upper 2000 m of the oceans from 2005 to 2018 (period where this basin is best observed), while only representing 30% of the total ocean surface (*high confidence*) (computed from NCEI dataset) (Figure 9.6). The Southern Hemisphere Mode Waters (Section 9.2.2.3) have stored large amount of heat having large impact on the global ocean energy budget (*high confidence*), by efficiently subducting surface warming at depth (Desbruyères et al., 2017; Gao et al., 2017; Häkkinen et al., 2016; Kolod et al., 2019). More generally, global ocean temperature change reflects water mass formation processes and ocean circulation, which provides confidence in assessment in sparse observation area (Desbruyères et al., 2017) (Figure 9.6).

The spatial pattern of ocean warming in the 0-700 m layer for the period 1971-present is characterized by widespread warming across all ocean basins (Figure 9.6) with particularly pronounced warming in the Atlantic sector. The spatial pattern of warming in the 0-700m layer is consistent with the observed SST trend over the overlapping period 1979-2017 (Section 9.2.1.1), with the exception of the high latitude Southern Ocean. Southern Ocean cooling is limited to the high latitude surface layer with little change in the underlying layer (Figures 9.3, 9.6). Heat taken up in the high latitude Southern Ocean is stored in the ACC and directly north of it in the 0-700m layer and deeper, so that the high latitude surface cooling contrasts

with the large warming subducted north of the ACC (Gille, 2008; Frölicher et al., 2015; Swart et al., 2018) (*medium confidence*).

Since about 2005, quasi-global coverage of the 0-2000m layer is provided by the Argo array of autonomous profiling floats (Riser et al., 2016), allowing a robust assessment of global and regional temperature (and salinity) change in the upper 2000m (Roemmich et al., 2015; Wijffels et al., 2016; Cheng et al., 2019). Consistent with AR5 and SROCC, the fastest warming in the 0-2000m region since 2005 is in the Southern Ocean (Roemmich et al., 2015; Trenberth et al., 2016), and in tropical and subtropical basins (England et al., 2014; Roemmich et al., 2015; Häkkinen et al., 2016; Desbruyères et al., 2017) (Figure 9.6, *high confidence*). The period of global Argo coverage is however relatively short so that detecting trends is constrained by the background natural climate variability. Recent cooling is observed by Argo in the subpolar North Atlantic, South-Western Indian Ocean and within most of the northern subtropics (10° N and 30° N) and eastern basins of the Pacific Ocean.

[START FIGURE 9.6 HERE]

Figure 9.6: Ocean Heat Content and its changes with time. (a) Timeseries of global ocean heat content anomaly relative to 2005-2014 climatology in the upper 2000m of the ocean. Shown are observations (Zanna et al., 2019), and multi-model means from the CMIP6 historical and SSP experiments. (b-g) Maps of Ocean Heat Content across different time periods, in different layers, and from different data sets/experiments. Maps show the observed trends of OHC for (b) 0-700m for the period 1971-2014, and (e) 0-2000m for the period 2005-2017, and the corresponding multi-model bias in these trends for CMIP6 ensembles versus observations (Zanna et al., 2019). [PLACEHOLDER UNTIL THE SIMULATION DATA ARE AVAILABLE]. Also shown are the projected change in 0-700m OHC for (d) SSP1-2.6 and (g) SSP5-8.5 in the CMIP6 ensembles, for the period 2091-2100 versus 2005-2014.

[END FIGURE 9.6 HERE]

Direct observations of ocean temperature change below 2000 m are limited to ship-based, high-quality deep ocean temperature measurements, with improved sampling from 1990 to the present due to the World Ocean Circulation Experiment (WOCE) and GO-SHIP program. There is *low confidence* in the spatial patterns of change at deeper levels for the period 1971-present due to the limited historical observations (Abraham et al., 2013; Palmer et al., 2017). Ocean data-assimilation estimates can help in data-poor regions such as the deep ocean. However, changes in the deep ocean are inconsistent across an ensemble of distinct ocean data-assimilation reanalyses, giving *low confidence* in reanalysis as a tool to estimate past changes in the deep ocean (Palmer et al., 2017). Temperature trends have been estimated for spatially coarse ocean sub-basins (Purkey and Johnson, 2010). Deep (> 2000 m) ocean temperature before the 2000s are largely consistent with post-2000 trends (Purkey and Johnson, 2010, 2013; Desbruyères et al., 2017; Purkey et al., 2019), with vertically coherent signals for the whole water column below 2000m, while uncertainty is larger in the 2000-4000 m layer than below (*medium confidence*). The ocean heat content trend below 2000 m from 1993-2018 is $1.15 \pm 0.57 \times 10^{21} \text{ J yr}^{-1}$ (Section 2.3.3.1; Table 2.7). The abyssal layer (4000-6000 m) shows the largest rate of warming of the deep layer, dominated by trends in the Southern Oceans consistent with ocean bottom water masses propagating north from a warmer Southern Ocean (Section 9.2.3.2), and the abyssal layer overall contributes to one third of the total ocean heat uptake below 2000 m (Desbruyères et al., 2016). Trends in deep waters (below 1800 m) in specific regions such as the North Indian and Pacific oceans have been attributed by inverse modelling to the continuing response to the transition from the Medieval Warm Period (950 to 1250 CE) to the Little Ice Age (1300 to 1850 CE) (*low confidence*) (Gebbie and Huybers, 2019), consistent with 19th century observations (Roemmich et al., 2012; Palmer, 2017).

The regional pattern in ocean heat content can be decomposed into a component coming from extra heat entering the ocean as if it were a passive tracer ('excess' heat), and a component related to ocean transport of background temperature field as if it were unperturbed ('redistributed' heat) (Gregory et al., 2016; Couldrey et al., submitted). An analysis of hydrographic change observed from 2006 to 2016, making an assumption of minimum water mass transformation, indicates that the observed OHC pattern is predominantly due to

redistribution of background heat (Zika et al., submitted), consistent with regional change of temperature over such relatively short time-scale being dominated by mode of variability and/or decadal change in ocean current (*high confidence*). For example, global upper ocean temperature tracks ENSO, partly balanced by subsurface anomalies of the opposite sign around 200m (Dean and John, 2011; Wijffels et al., 2016); the subpolar North Atlantic region since 2006 has cooled substantially, especially in the upper few hundred meters, as a result from a combination of reduced heat advection by the AMOC (redistribution; Section 9.2.3.1), and regional anomalously cold atmospheric conditions (excess) (Smeed et al., 2014; Robson et al., 2016; Josey et al., 2018; Bryden et al., 2019; Zika et al., submitted) (*medium confidence*). On top of this large signal from redistribution, there is also a significant signal of excess heat with a more uniform distribution, showing mid-latitude maxima consistent with subduction of water masses (Section 9.2.2.3). In contrast to the observed changes during the last decades, which is dominated by redistribution, at centennial timescales, full-depth warming can be essentially reconstructed by sea surface temperature anomalies are carried into the ocean interior by time-invariant circulation processes (Zanna et al., 2019).

[START FIGURE 9.7 HERE]

Figure 9.7: Meridional transects of potential temperature in the ocean and its changes with time in the upper 2000m of the Global, Pacific, Atlantic, and Indian Oceans. Shown are observed temperature (ARGO climatology 2005-2014), bias of the CMIP ensemble over this period, and future changes under SSP1-2.6 and SSP5-8.5.

[END FIGURE 9.7 HERE]

Concomitant development in both the ocean observing system and the climate modelling leads to an increased evidence since AR5 on the consistency between modelled and observed OHC, especially when accounting for forcing from important volcanoes eruptions, despite individual models showing larger spread in OHC than uncertainty in observations (Section 3.5.1.2). In general, climate models tend to underestimate the change in upper ocean OHC, whereas changes below 700m are overestimated before 1990, which suggest models are too diffuse (Section 3.5.1.2, Figure 9.7). In addition, historical model runs simulate different phasing of natural variability than observations (Garry et al., submitted). Growing evidence since AR5 confirm however that it is *virtually certain* that anthropogenic forcing causes the increase in OHC in the upper and intermediate ocean layers (0-2000 m), and *likely (medium confidence)* for the deeper ocean (Section 3.5.1.2). At decadal timescale the marked long-term increase in global ocean heat content shows cooling phases attributed to major volcanic eruptions (Agung in 1963, El Chichón in 1982 and Pinatubo in 1991). While, explosive volcanic eruption disturbs the earth radiative budget for only a few years, the ocean integrates the radiative cooling in its subsurface, typically in the upper 500 m (Bilbao et al., 2015; Gupta and Marshall, 2018), and deliver it back to the surface over decadal periods, extending the global ocean heat content response and associated impact of thermocline height well beyond the eruption time scale (Church et al., 2005; Fasullo et al., 2016; Fasullo and Nerem, 2018).

The ocean will continue to warm over the 21st century (*very high confidence*) (Figures 9.6, 9.7), and it is *likely* that this warming will continue at least to 2300 even for low emission scenario because of ocean inertia, but with a rate dependent on the scenario (Nauels et al., 2017a; Palmer et al., 2018). Dependence on scenarios begins only after about 2040 (Figure 9.6). SROCC assesses that the 0–2000 m layer of the ocean is projected to warm by a further 2150±540 ZJ for the RCP8.5 scenario and 900±345 ZJ from 2017 to 2100. Here we revise those estimates from CMIP6 projection estimating that heat content in the 0–2000 m layer will increase by [PLACEHOLDER COMPUTED FROM ONLY ONE MODEL; ADDITIONAL CMIP6 MODELS WILL BE USED TO UPDATE AND PROVIDE CONFIDENCE INTERVAL] 1410, 1700, 1960, or 2300 ZJ from 1995-2014 to 2081-2100 under scenarios SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively. Distribution of warming within water masses and associated thermocline sea-level are discussed below (respectively in Section 9.2.2.3 and 9.2.4).

Climate model projections show that the largest part of global ocean heat uptake occurs in the Southern Ocean, due to mid-latitude convergence and downward advection of excess heat by the residual circulation

(Banks and Gregory, 2006; Kuhlbrodt and Gregory, 2012; Marshall et al., 2015b; Armour et al., 2016). More generally, excess heat tends to follow known water mass pathways from surface formation regions to depth, primarily in the North Atlantic and Southern Ocean (Dias et al., submitted; Sallée et al., 2013; Garry et al., 2018; Couldrey et al., 2020) (Figure 9.8). In addition, strengthening of the Southern Ocean westerlies (Section 9.2.1.3) causes a northward and downward redistribution of the background heat content (Dias et al., submitted). The projected weakening of the AMOC (Section 9.2.3.1) causes redistribution of background heat from northern to low latitudes in the Atlantic (Winton et al., 2013; Marshall et al., 2015b; Gregory et al., 2016) (Figure 9.8), and the wide model spread in the magnitude of this effect (Menary and Wood, 2018) is the largest cause of model spread in the projected pattern of ocean heat content change (Couldrey et al., submitted). Redistribution of background heat content makes zero contribution to global ocean heat uptake, which therefore equals the global integral of excess heat.

[START FIGURE 9.8 HERE]

Figure 9.8: (a) Total ocean heat uptake after 100 years of simulation in an idealized simulation where atmospheric CO₂ increase by 1% per year as simulated by five Climate models (CanESM2, CanESM5, GFDL-ESM2M, 687 HadCM3, and MPI-ESM-LR), and associated decomposition of this total heat uptake into (b) passive uptake of added heat (“excess heat”; see text), and (c) pure redistribution of unperturbed heat (“redistributed heat”; see text) (Couldrey et al., submitted). (d) Relationship between northward heat transport and Atlantic Meridional Overturning Circulation in HighResMIP models (1950-2050) and observations during the RAPID period (2004-2018).

[END FIGURE 9.8 HERE]

Deep convection at subpolar latitudes effectively carries the excess heat signal downward, but due to decreasing convective activity, with decreasing efficiency (Heuzé et al., 2015). Model discrepancies in trends are greater below 700 m, presumably due to the differing representation of the overturning, large-scale circulations, and eddy transport among models. In the 700-2000m depth range, CMIP5 models project the largest warming in the North Atlantic Deep Water and Antarctic Intermediate Water (Figure 9.7). Below 2000 m, the North Atlantic consistently cools, and Antarctic Bottom Waters warm (Sallée et al., 2013b; Heuzé et al., 2015; Garry et al., submitted) (Figure 9.8, Section 4.5.2.1). This deep warming of the ocean is slow to manifest with multi-century or longer response times, so the ocean heat content and global mean thermosteric sea level will continue to rise for centuries, even in the event of cessation of greenhouse gas emissions (Ehlert and Zickfeld, 2018).

Improved process understanding of heat transport since AR5 increases the degree of evidence, thereby providing *high confidence* in several key aspects of oceanic heat transport changes under global warming. In particular, climate models project several robust changes in meridional ocean heat transport under global warming (Chapter 7; Figure 7.5). Anomalous northward ocean heat transport in the Southern Ocean arises primarily from northward-flowing surface waters carrying anomalous heat from regions of reduced heat loss to the atmosphere around 60°S to where it is subducted to greater depth and stored in the deeper ocean further north as shown in Figure 9.7 by the sudden increase in OHC to depths of 1 – 2 km (Frölicher et al., 2015; Marshall et al., 2015b; Armour et al., 2016; Morrison et al., 2016; Liu et al., 2018b; He et al., 2019). Because eddies partly counteract these wind-driven changes in surface flow we have *low confidence* in these changes (Section 9.2.3.2). Decreased northward oceanic heat transport in the northern hemisphere mid-latitudes is mainly associated with a reduction in the strength of the AMOC (Figure 9.8; Section 9.2.3.1; Chapter 4, Section 4.3.2; Cheng, Chiang and Zhang, 2013) (*high confidence*), which occurs as greenhouse gas forcing reduces surface heat loss and increases freshwater input at high latitudes, lowering the density of surface waters in sinking regions (Rugenstein et al., 2013; Winton et al., 2013; Marshall et al., 2015b; Oldenburg et al., 2018). As a result, there is decreasing OHC in the subpolar North Atlantic (Figure 9.6). Associated with gradually northward shifting sites of deep-water formation and a retreating sea-ice edge, most models project increase in oceanic heat transport into the Arctic Ocean *medium confidence*. In many models, north of 60-70°N the overturning increases instead of decreases with global warming, and this is more pronounced in models with a high-resolution ocean component (Roberts et al., submitted) (*medium*

confidence). Also, slowly changing North Atlantic currents transport anomalous heat that would otherwise accumulate in the North Atlantic subpolar gyre, due to reduced heat loss to the atmosphere, further north into the Nordic Seas (Jungclauss et al., 2014; Koenigk and Brodeau, 2014; Marshall et al., 2015b; Nummelin et al., 2017; Oldenburg et al., 2018), leading to a dipole pattern in changing OHC in the North Atlantic with decrease between 40-60°N and increase between 60-80°N (Figures 9.6, 9.7). Observations and ocean modelling suggests that northward ocean heat transport into the Arctic already has increased in recent decades owing to changes in the temperature and volume transport of northward-flowing Atlantic water (*medium confidence*), but that these trends may reflect natural variability and cannot be ascribed yet to anthropogenic forcing (Muilwijk et al., 2018; Wang et al., 2019b).

Past changes in total OHC and oceanic heat transports illustrate scales of change and timescales of adjustment relevant to expected future changes. Paleo-observations of a thermal bipolar seesaw, in which abrupt cooling of the North Atlantic Surface waters is associated with gradual warming around Antarctica, are accounted for conceptually via reduction in net northward oceanic heat transport in the Atlantic during reduction or collapse of AMOC, compensated by a build-up of heat in the Southern Ocean (Stocker and Johnsen, 2003). Process modelling reveals a response to forced AMOC shutdown with surface ocean cooling in the northern subtropics and subpolar Atlantic, but an increase in OHC of $\sim 1.5 \times 10^{24}$ J throughout the global ocean interior (more than half in the subsurface Indo-Pacific) rather than in the Southern Ocean as previously assumed (*low confidence*). Transmission of subsurface ocean heat to Antarctica is driven in the model by increased eddy heat flux across the ACC (Pedro et al., 2018). Timescales of circulation adjustment to forced perturbation are a few hundred years, whereas timescales of thermal adjustment (consistent with data) are longer, > 500 yr (not run to equilibrium). On the glacial-interglacial scale, ice core rare gas ratios (Bereiter et al., 2018; Baggenstos et al., 2019; Shackleton et al., 2019) document a rise in global OHC relative to the last glacial maximum of $\sim 2 \times 10^{25}$ J (*medium confidence*) tracking Antarctic surface air temperature changes (Cuffey et al., 2016). These measured glacial-interglacial changes in OHC are about ten times greater than the modelled OHC changes associated with collapse of AMOC. The timing of observed OHC change on this scale suggests tight linkage, via the Southern Ocean, of mean ocean temperature and Antarctic air temperature. Consistent with paleo simulations, 10,000-year projected responses to CO₂ emissions (ensemble of four intermediate complexity models) show OHC peaking >1000 years after GMST highs, with long term OHC commitments of 0.3, 1.3, 2.0, 2.9, and 3.7×10^{25} J for future emissions of 0, 1280, 2560, and 3840 Gt respectively, with OHC peaking near 4000 AD in all projections, followed by gradual declines of about 25% of this excess OHC by 12,000 AD (calculated from data in (Clark et al., 2016)), indicating that increased ocean heat content in response to anthropogenic CO₂ emissions is essentially irreversible on human timescales (Figure 9.9).

[START FIGURE 9.9 HERE]

Figure 9.9: Measured changes in past mean ocean heat content OHC (20 kyr BCE to present) estimated from ice-core rare gases (Baggenstos et al., 2019; Shackleton et al., 2019), changes in global mean sea-surface temperature GMSST (Shakun et al., 2012; Marcott et al., 2013), and historical OHC changes from HadCM3 (Gregory et al., 2006) and observations (Durack et al., 2018; Johnson et al., 2018). Future long-term projected changes in OHC and GMSST (present to 12 kyr CE) in response to four greenhouse gas emissions scenarios over multiple millennia (Clark et al., 2016).

[END FIGURE 9.9 HERE]

9.2.2.2 Ocean Salt Content

Progress in relating ocean salinity and complex ocean-atmosphere-land freshwater interactions has improved confidence in observed salinity trends, and in the attribution to anthropogenic forcing. Observational progress has considerably expanded salinity measurement towards global coverage for the 0 – 2000m ocean since 1999 (Argo program (Riser et al., 2016)), and process-understanding (SPURS-1/2; (Lindstrom et al., 2015; SPURS-2 Planning Group 2015)). Near real-time sea surface salinity retrievals from satellites (SMOS, Aquarius and SMAP) allow atmosphere-land-ocean water cycling to be better resolved (Asher et al., 2014;

Boutin et al., 2014; Grodsky et al., 2014; Supply et al., 2018).

It is *very likely* that subsurface salinity has changed across the globe since the mid-20th century with an underlying pattern where regions of high salinity where evaporation dominates over precipitation have become more saline, while regions of low salinity where precipitation dominates over evaporation have become fresher (Section 2.3.3.2). Subsurface salinity changes are consistent with near-surface changes propagated in the interior by the large-scale circulation (Section 2.3.3.2, Section 9.2.1.2). Overall, the salinity pattern amplification - fresh regions becoming fresher, and salty regions saltier - is supported by most observational studies that have considered salinity change since the modern instrumental record began in the 1950s (Section 2.3.3.2) (Durack et al., 2013; Durack, 2015). Defining the global salinity contrast as the difference between the mean salinity for all regions where climatological salinity is higher than the global median, and the mean salinity for all regions where climatological salinity is lower than the global median, the salinity pattern amplification from 1960 to 2017 is estimated using three different observation-based products to 0.021–0.05 g.kg⁻¹ per century (Cheng et al., submitted). Since AR5, additional observational analyses have further confirmed the long-term, large-scale and regional patterns of salinity change since the 1950s (Skiriris et al., 2014; Durack, 2015; Cheng et al., submitted). These report a multidecadal freshening in the North Pacific subpolar gyre; Pacific and Indian intermediate waters (Du et al., 2015; Melzer and Subrahmanyam, 2017), Pacific Ocean (Oka et al., 2017), Greenland Seas (Friedman et al., 2017), and Southern Ocean (Azaneu et al., 2013; Bintanja et al., 2013; Purkey and Johnson, 2013; De Laverne et al., 2014; Haumann et al., 2016; Morrow and Kestenare, 2017; Purkey et al., 2019) (Section 2.3.3.2). In addition, coherent spatial changes across the Atlantic, with subpolar freshening and low-latitude salinification, were reported for a much longer period than previously (1896 to 2013) (Friedman et al., 2017). Numerical model analyses give *high confidence* that near-surface and subsurface salinity pattern changes are linked to changing E-P (Cheng et al., submitted; Durack, Wijffels and Boyer, 2013; Lago et al., 2016; Zika et al., 2018) and changes in Antarctic sea-ice (Purich et al., 2018; Haumann et al., submitted). Observed changes are replicated in free-running CMIP5 historical simulations with greater spatial variability than found across the observed estimates and correspondingly smaller amplitudes in the anthropogenically forced multi-model mean (Durack, 2015; Silvy et al., 2019; Cheng et al., submitted). These coincident model and observational analysis support the strengthened AR5 (Bindoff et al., 2013) attribution statements documenting water cycle amplification over the observed period. There is *medium confidence* in the absolute rate of salinity changes through time, however agreement exists between independent estimates when error margins are considered (Durack and Wijffels, 2010; Skiriris et al., 2014, 2016; Zika et al., 2018; Cheng et al., submitted). Considerable biases in the global zonal mean subsurface salinity field are present in CMIP5 (*high confidence*) especially the 0-200m salinity field (Flato et al., 2013) affecting the simulation of mixed layer depths (Section 9.2.1.4), subduction, and circulation pathways (Durack et al., 2012; Lago et al., 2016).

At the global scale, considering new observed and modelled studies since AR5, it is *extremely likely* that near-surface and subsurface salinity changes are a consequence of anthropogenic forcing over the 20th and early 21st centuries (see Section 3.5.2 for comprehensive assessment), involving a variety of mechanisms. A single idealized modelling analysis attributes half of the subsurface salinity change to E-P forcing, one-third from ocean warming affecting stratification and circulation, and one-sixth from land ice melting and runoff (*low confidence*) (Zika et al., 2018). The observed inter-basin halosteric sea level change made of a contrasting freshening the Pacific basin and salinification of the Atlantic basin is consistent only with CMIP5 models forced with anthropogenic greenhouse gas forcing (Durack et al., 2014). Attribution of salinity changes in specific water masses is discussed below (Section 9.2.2.3).

Future projections reproduce the fresh gets fresher, salty gets saltier paradigm on large scales (*high confidence*), including complex regional changes resulting in stronger freshening of the Pacific, Indian and Southern Oceans, with a saltier subtropical Atlantic (*medium confidence*).

9.2.2.3 Water Masses

Water masses are “formed” at the ocean surface in contact with the overlying atmosphere (and ice in the polar oceans), before being transferred (“subducted”) to the deeper ocean, away from the surface turbulent

layer. As water masses subduct, they “ventilate” the subsurface ocean, transferring the temperature, salinity, carbon and oxygen characteristics acquired at the ocean surface to the subsurface. By acting as an integrator over time of surface flux changes, water masses provide higher signal-to-noise ratios for detecting and monitoring climate change than surface fluxes (Bindoff and McDougall, 2000; Durack and Wijffels, 2010).

Upper ocean water masses ventilate the main thermocline of the ocean at mid- to low-latitudes and have circulation timescales away from the surface of the order of years to decades. Enhanced stratification and increased heating of the subtropical gyres leads to stronger gyres in the upper thermocline and weaker gyres in the lower thermocline (Wang et al., 2015a). SubTropical Mode Water (STMW) has deepened worldwide (*high confidence*) due to efficient warming resulting from subduction of this water mass, with greatest deepening in the Southern Ocean (Häkkinen et al., 2016; Desbruyères et al., 2017). There is *high confidence* that subsurface warming is larger than at the surface in the Northern Hemisphere STMW (Sugimoto et al., 2017) because they are formed in western boundary current (WBC) extensions where surface warming is larger than the global average (Section 9.2.2.1). Variability in STMW have large imprint on ocean heat content, with greatest contribution in the period 2006-2015 from the North Atlantic with a reduced heat content due to a thinning of the layer despite its warming and from the South East Pacific with a marked increase of heat content associated to a thickening and warming of the layer (Kolodziejczyk et al., 2019). Long-term repeated observation show a freshening of STMW in the North Pacific (Durack and Wijffels, 2010; Oka et al., 2017), predicted to further intensify in the future, with a signal predicted to eventually be larger than typical natural variability in 2020-2040 (at about 1.5-2.0 °C of GSAT) (Silvy et al., submitted).

Southern Ocean Sub Antarctic Mode Water (SAMW) has been observed (*high confidence*) to deepen, warm and freshen in the past decades (Bindoff and McDougall, 2000; Durack and Wijffels, 2010; Häkkinen et al., 2016; Desbruyères et al., 2017; Gao et al., 2018). The observed increase of SAMW heat content from 2005 to 2015 ($3.9 \pm 0.3 \text{ W m}^{-2}$) is primarily attributable to its thickening ($3.6 \pm 0.3 \text{ m yr}^{-1}$) (Gao et al., 2018; Kolodziejczyk et al., 2019), with a trend consistent with the 50-year time scale trend (Häkkinen et al., 2016). Over this recent period the overall volume change is associated with a thickening of the lighter part of the water-mass and thinning of the denser part, translating into an overall lightening of SAMW (Portela et al., 2019). Climate models, consistently across CMIP3 and CMIP5, generally simulate SAMW that penetrates less deeply and at a lighter density class than observed, but consistently predict that SAMW will continue to warm and freshen in the 21st century (*high confidence*), the extent depending on the emissions pathway (*likely* warming of 0.2°C – 1.3°C, and *likely* freshening 0.01 – 0.1 g kg⁻¹ in 2100 under RCP4.5, approximately double that under RCP8.5; (Sallée et al., 2013b; Silvy et al., submitted)). The freshening of SAMW appears has one of the strongest salinity signal of the world oceans that climate models predict to have already emerge from typical natural variability at the end of the 20th or early 21st century (Silvy et al., submitted).

North Atlantic SubPolar Mode Water (SPMW) featured a slight overall cooling and freshening from 1997 to 2014, but was mostly characterized (*medium confidence*) by marked decadal variability (Holliday et al., 2008, 2015; Reverdin, 2010; Gladyshev et al., 2017).

Intermediate water masses ventilate the ocean beneath the main thermocline and have circulation timescales away from the surface of order of decades to centuries. Antarctic Intermediate Waters (AAIW) has warmed and freshened (*high confidence*) at an estimated rate of 0.05 – 0.15°C, and 0.002 – 0.013 g kg⁻¹ per decade from 1970 to 2009 (Schmidtko and Johnson, 2012; Swart et al., 2018), with continued freshening at a comparable $0.006 \pm 0.003 \text{ g kg}^{-1}$ per decade in the Atlantic sector of the Southern Ocean from 2006 to 2015 (Giglio and Johnson, 2017). The warming and freshening of AAIW are inconsistent with internal variability or natural forcing and attributable (*low confidence*) to human-induced greenhouse gas increases primarily, with a secondary role for stratospheric ozone depletion (Swart et al., 2018). Climate models tend to be biased towards too light AAIW but in CMIP5 biases are getting weaker and models represent the low-salinity tongue and northward transport of AAIW (Sallée et al., 2013b; Russell et al., 2018). Under increased radiative forcing (RCP8.5), AAIW is predicted to warm by 1 – 2°C and to freshen by 0.5 to 1.5 g kg⁻¹ at the end of the century (Sallée et al., 2013b). In summary, intermediate waters in the Southern Ocean have freshened, and these observed recent trends are predicted to increase in the 21st century under increased radiative forcing (Silvy et al., submitted).

1 Deep water masses ventilate the ocean mid-depth and abyss, have very limited formation regions, and have
2 circulation timescale away from the surface of order of centuries to millennia.

3
4 The North Atlantic supplies the deep ocean with Upper and Lower North Atlantic Deep Water (NADW).
5 Upper NADW is formed by deep convection in the Labrador and Irminger Seas, while Lower NADW is
6 formed by deep overflows entraining water as they flow over sills between Scotland and Greenland.
7 Temperature and salinity variations in Upper NADW are dominated by decadal variability, and over the fifty
8 years of measurement variability is greater than any long-term trend *high confidence* (Rhein et al., 2011;
9 Mauritzen et al., 2012). Lower NADW displays a density-compensated cooling and freshening trend since
10 1955, associated with similar density-compensated changes in deep overflows (Cunningham and Alderson,
11 2007; Mauritzen et al., 2012) *medium confidence*. In CMIP5 projections NADW formation decreases,
12 associated with a decline of the AMOC in all models (Cheng et al., 2013; Reintges et al., 2017) (Section
13 9.2.3.1). CMIP5 models have large biases in NADW characteristics (Heuzé, 2017). Despite the large range
14 in model bias in NADW water mass characteristics, the projected decrease in NADW formation is consistent
15 across models (Heuzé, 2017). It is concluded that it is *likely* that NADW will decrease in volume in the
16 future, associated with a decrease in density in the northern North Atlantic and Arctic basins, where
17 freshening of the upper 1400m dominates warming in terms of density decrease.

18
19 Circumpolar Deep Water (CDW) is a thick, old ocean water-mass, marked by low oxygen, high nutrient, and
20 high carbon content, which forms as admixture of NADW and Antarctic Bottom Waters (AABW) (Talley,
21 2013). The mean transport of CDW is directed southward, across the Southern Ocean, and upwells south of
22 the ACC in the subpolar Southern Ocean, where it represents a relatively warm water-mass compared to
23 other ambient waters. The age of CDW (DeVries and Primeau, 2011) means that they have not yet been
24 affected by current climate change in their surface formation regions but are affected locally in the Southern
25 Ocean as they upwell close to the surface ocean in the Southern Ocean. SROCC assessed with *low*
26 *confidence* that CDW mean southward and upward transport toward the Southern Ocean surface has
27 increased from 1990-2000, and reduced in the following decade (Meredith et al., 2019) (Section 9.2.3.2).
28 There is *low confidence* that change in such transport is due to change in the intensity of westerly winds
29 (Waugh et al., 2019). A number of observation-based and theoretical evidence suggest that as surface
30 stratification has increased (*medium confidence*) in the subpolar Southern Ocean from 1970-2018 (Section
31 9.2.1.4), there is less volume of CDW being mixed to the surface, so its associated heat content is less
32 efficiently lost to the cold atmosphere (*low confidence*), causing CDW warming directly below the surface
33 layer in the subpolar Southern Ocean (*medium confidence*) (Schmidtke et al., 2014; Armour et al., 2016;
34 Lecomte et al., 2017; Bronselaer et al., submitted). SROCC assessed that subsurface warming directly below
35 the surface layer extends close to Antarctica, and intrudes onto the continental shelf in different regions, but
36 most widely in the Amundsen-Bellingshausen Sea where temperature has increased of 0.1-0.3°C per decade
37 from 1983-2012, plausibly due to wind forcing, and with significant decadal variability (Meredith et al.,
38 2019). New evidence since SROCC reinforced our confidence in the importance of the role of winds in
39 driving CDW heat delivery to the Amundsen-Bellingshausen Sea (Dotto et al., 2019) and suggest (*low*
40 *confidence*) that on top of a strong internal decadal wind variability, increased greenhouse gas forcing has
41 caused a slight mean change of the local winds from 1920-2018, facilitating CDW heat intrusion on the
42 Amundsen-Bellingshausen continental shelf and ice-shelf melt (Holland et al., 2019). In the future, there is
43 *high confidence* that a change of wind pattern (Spence et al., 2014), and reduction in sea-ice
44 production (Timmermann and Hellmer, 2013) are processes that can facilitate access of CDW in the sub-
45 iceshelf cavities, but there is *low confidence* on the quantification and importance of each of these processes.
46 In addition, while large regions of the Antarctic shelves are currently isolated from warm CDW, some
47 studies have predicted a possible shift from cold to warm sub-ice shelf cavities causing a sudden flush of
48 warm water underneath ice shelves, but there is *low confidence* both in the driving processes and the
49 threshold at which such a shift could happen (Hellmer et al., 2012, 2017; Silvano et al., 2018).

50
51 AABWs have *likely* warmed, freshened, and reduced in volume over the globe since the 1990s (Purkey and
52 Johnson, 2012, 2013; Desbruyères et al., 2016, 2017), and climate models predict (*low confidence*) their
53 production will further decrease and their temperature will increase in the 21st century under increased
54 radiative forcing (Sallée et al., 2013b; Heuzé et al., 2015), although the modelled formation of these water
55 masses is crude in representing continental slope and shelf processes as well as under-ice processes (Stewart

and Thompson, 2015; Manucharyan and Thompson, 2017; Horvat and Tziperman, 2018; Thompson et al., 2018), and over-representing open-ocean convection events (De Lavergne et al., 2014; Heuzé et al., 2015). Changes in the characteristics of abyssal waters can affect the ocean's large-scale circulation (including aspects of the ACC and the strength of the AMOC (Patara and Böning, 2014; Stewart and Hogg, 2017), ocean heat content (Desbruyères et al., 2016), and sea-level rise (Purkey and Johnson, 2013). The observed warming of the ocean abyss (deeper than 4000 m) is detectable in all Southern Ocean basin (Atlantic, Indian, Pacific) with a global mean trend of $0.53 \pm 0.11 \text{ } 10^{-3} \text{ } ^\circ\text{C yr}^{-1}$ from 1990 to 2010 in line with previous estimates (Purkey and Johnson, 2010; Desbruyères et al., 2016), suggesting little change in the warming rate. Regionally, observed trends in the Australian Antarctic Basin suggest a striking acceleration of the freshening between 2007 and 2016 ($0.008 \pm 0.001 \text{ g/kg decade}^{-1}$) compared to 1994 and 2007 ($0.002 \pm 0.001 \text{ g/kg decade}^{-1}$) (Menezes et al., 2017), but this abrupt freshening could be due to a specific calving event in the source region of the Australian Antarctic AABW (Kusahara et al., 2011). The scarcity of observations of AABW and the known biases in modelling of AABW formation mechanisms and convection to depth (Heuzé et al., 2013, 2015; De Lavergne et al., 2014), gives *low confidence* in attributing causes of observed trends. Nonetheless, CMIP5 models suggest (*low confidence*) that AABW will undergo a warming of $0.19 \pm 0.07^\circ\text{C}$, a salinification of $0.02 \pm 0.01 \text{ g kg}^{-1}$, and a lightening of $0.011 \pm 0.006 \text{ kg m}^{-3}$ by 2100 (Heuzé et al., 2015).

9.2.3 Regional Ocean Circulation

9.2.3.1 Atlantic Meridional Overturning Circulation

The Atlantic Meridional Overturning Circulation (AMOC) is characterised by warm, salty waters flowing poleward in the upper 1000 m, while losing buoyancy by cooling, sinking at subpolar latitudes, with a return flow of North Atlantic Deep Water between 1000 and 3000 m depth (Buckley and Marshall, 2016). There is *very high confidence* that the AMOC is linked to convective overturning in the Labrador and Irminger Seas and in the Iceland Basin and Nordic Seas (Buckley and Marshall, 2016), although the sinking branch itself is more localised in the boundary current flowing cyclonically along the perimeter of the Nordic Seas and North Atlantic subpolar gyre (Spall and Pickart, 2001; Katsman et al., 2018). In addition to convection, other key processes controlling the AMOC are the overflows between the Nordic Seas and the North Atlantic. Surface exchange over North Atlantic subpolar latitudes of heat and freshwater with the atmosphere affects convection and overflows with different time-lags, and at even larger time-lags changes in northward transport of heat and salinity by the AMOC change convection and overflows as well (Buckley and Marshall, 2016).

[START FIGURE 9.10 HERE]

Figure 9.10: (Top left) AMOC magnitude in PMIP experiments. (Top right) Timeseries of AMOC from CMIP5 and CMIP6 based on (Menary et al., submitted). (Bottom left) % change in AMOC strength per year at different resolutions over 1950-2050 period with colours for model families (Roberts et al., submitted). (Bottom right) A compilation (Jackson and Wood, 2018) of percentage changes in the simulated AMOC after applying an additional freshwater flux in the subpolar North Atlantic at the surface for a limited time (de Vries and Weber, 2005; Stouffer et al., 2006; Yin and Stouffer, 2007; Jackson, 2013; Liu and Liu, 2013; Jackson and Wood, 2018; Haskins et al., 2019). Symbols indicate whether the AMOC recovers within 200 years (circles), is starting to recover (upwards arrow) or does not recover within 200 years (downwards arrow). Symbol size indicates rate of freshwater input. (See Supplementary material for further details).

[END FIGURE 9.10 HERE]

Since 2004, the strength of the AMOC has been measured at 26.5°N by the RAPID Array (Smeed et al., 2018) (Section 2.3.3.4). The short length of the record and the surprisingly large variability compared to

CMIP models (Roberts et al., 2014) (Figure 9.10) gives *low confidence* in a meaningful long-term trend from the array even though Smeed et al. (2018) argue that between 2007 and 2011 the AMOC shifted to a state of reduced overturning. The RAPID array estimates an average AMOC from April 2004 to April 2008 of 18.8 Sv, while for the period after 2008 it is 16.1 Sv. The amount of decrease has been challenged by (Trenberth et al., 2019), arguing that in the beginning of the period part of the array was damaged by hurricanes. Their northward ocean heat transport estimate (dominated by the AMOC) shows a weaker 2004-2016 trend than the RAPID estimates, but at present it is impossible to say which estimate is better. Ocean reanalyses suggest that the AMOC reduction after 2008 followed a period of increase between the mid-90s and mid 2000s implying that AMOC changes from RAPID are due to internal variability (Jackson et al., 2016). In contrast, North Atlantic subpolar gyre temperatures suggest AMOC strengthening in the early nineties, after a period of relative constancy, followed by a decrease since the mid-nineties and subsequent cooling since 2005 (Robson et al., 2016). Longer timeseries based on AMOC temperature-fingerprints indicate a reduction of approximately 3 Sv since the mid-20th century (Dima and Lohmann, 2010; Rahmstorf et al., 2015; Caesar et al., 2018), and paleo-proxies suggest that the twentieth-century AMOC is weaker than in the previous 1500 years (Thornalley et al., 2018). The various AMOC reconstructions contain contradicting messages with one possible interpretation that various AMOC-temperature and salinity footprints are timescale-dependent, or, affected by other processes such as the NAO. In addition, AR5 showed that CMIP5 models display a weak negative trend in AMOC over the historical period (1850 – 2005; Figure 9.10) between 0 and -2 Sv per century. On top of this trend, variability on an approximate 60 year timescale in the multi-model mean was identified, attributed to variations in aerosol forcing over the historical period (Cheng et al., 2013). CMIP6 models show no, or even a slight positive trend over the historical period with a larger AMOC increase between 1850 and 1985-1995. Both CMIP5 and CMIP6, however, agree on a downward trend of approximately 1 Sv between 1995 and present (Fig. 9.10 top right panel; Menary et al. submitted). Based on these results we have *low confidence* in the reconstructed AMOC weakening over the last 150 years, *low confidence* that the presently observed AMOC reduction is anthropogenically forced and *low confidence* that the AMOC has been declining since the mid-twentieth century.

Biases in simulations of the present day AMOC and associated deep convection in the subpolar gyre and Nordic Seas remain large in CMIP5 and CMIP6 models. The simulated AMOC mean varies with a factor of 2 between CMIP5 models (Cheng et al., 2013), and many climate models (*medium confidence*) exhibit ocean convection that is too deep, over too large an area, too often and too far south (Figure 9.5), associated with biases in sea-ice extent, overflows, and freshwater forcing (Deshayes et al., 2014; Wang, Legg and Hallberg, 2015). Nevertheless, an improvement with respect to CMIP3 can be observed, in particular with generally more localised deep convection instead of deep convection over the whole subpolar gyre and convection to smaller depths (less than 2000m instead of convection to the ocean floor) (Heuzé, 2017). High-resolution coupled models show significant less bias over their lower-resolution counterparts (Roberts et al., submitted). Models and paleo-reconstructions show a close relation between AMOC variability and variability in deep convection, especially in the Labrador Sea (Eden and Willebrand, 2001; Jungclauss et al., 2005; Kuhlbrodt et al., 2007; Biastoch et al., 2008; Danabasoglu, 2008; Brodeau and Koenigk, 2016; Caesar et al., 2018; Thornalley et al., 2018) (*high confidence*). Recent observations, where the impact of Labrador Sea convection on the AMOC appears small (Lozier et al., 2019) somewhat challenge this relationship. An eddy-permitting ocean resolution coupled model showed similar weak overturning across the southern section of the Labrador Sea as was found in (Lozier et al., 2019) (Jackson et al., submitted) but this result needs to be confirmed with more and especially higher-resolution ocean-component coupled models, as eddy-permitting models still do not resolve Labrador Sea eddies, which were found instrumental in water-mass transformation and densification of the boundary current in an idealised model of a subpolar marginal sea (Brüggemann and Katsman, 2019). Based on this we have *low confidence* that CMIP models are able to correctly simulate internal variations in AMOC and deep convection, but we have *medium confidence* they correctly simulate long-term anthropogenic trends forced by large-scale changes in forcing.

Other sources of the AMOC consist of deep overflows from the Nordic Seas. There is no observational evidence for changes in these overflows as discussed in AR5 and SROCC (Hansen et al., 2016; Jochumsen et al., 2017) The representation of these deep overflows in many models is still poor and can have large impact on the simulated AMOC (Wang et al., 2015c) (*high confidence*). Some models indeed show the large benefit of implementing improved oceanic overflow representations (Legg et al., 2009).

The upper-layer northward flow in the Atlantic is mainly sourced by Agulhas leakage of warm and salty water from the Indian into the Atlantic Ocean via large eddies (Agulhas Rings) (Beal et al., 2011). Numerical model experiments suggest an intensification of the Agulhas leakage since the 1960s, which has contributed to the warming in the upper 300m of the tropical Atlantic Ocean (Lübbecke et al., 2015). There is, however, *low confidence* in a direct link between changes in Agulhas leakage and AMOC strength (Weijer and van Sebille, 2014), but *medium confidence* in the link between Agulhas leakage and the cross-hemispheric Atlantic Meridional Variability (AMV) associated with the AMOC (Bjastoch et al., 2015). Contrary to model projections, a multidecadal proxy suggests that the Agulhas Current has been broadening instead of strengthening since the early 1990s due to an increase in eddy activity (Beal and Elipot, 2016). Most CMIP models cannot directly simulate the Agulhas leakage because of their coarse resolution, so there is *low confidence* in their projections of leakage change. However, high-resolution models potentially give better projections of changes in Agulhas leakage. Based on all considerations mentioned above we have *low confidence* that CMIP models are able to correctly simulate internal variations in AMOC and deep convection, but we have *medium confidence* that they correctly simulate long-term anthropogenic trends forced by large-scale changes in forcing, at least in a qualitative sense.

The strength of the AMOC is *very likely* to decrease by 2100 (Figure 9.10), and larger decreases in volume transport are expected by 2100-2300 in high emissions scenarios. A preliminary analysis of available CMIP6 projections reveals that AMOC decline by 2100 is stronger than in previous cycles (*high confidence*) and only weakly dependent on emission scenario ranging from 40% in SPS1-1.9 to 50% in SPS5-8.5 (*medium confidence*) (Weijer et al., submitted). There is a slight tendency (*low confidence*) for higher-resolution models to show stronger declines (Figure 9.10 bottom left panel; Roberts et al. submitted). The AMOC in the historical simulations is characterized by an increase between 1960 and 1980 and a second increase in the early nineties, followed by a decrease after the mid-nineties that seems to continue uninterrupted in the SSP simulations. The AMOC evolution starts to differ between various SSP scenarios only after 2060. The simulated AMOC between 1900 and 2000 bears some resemblance with the observed shift in the nineties (Robson et al., 2012), but it should be emphasized that the a multi-model mean only contains the forced changes in AMOC, while the observations also contain internal variations. Therefore, there is *low confidence* that the simulated upward trend between 1960 and 1995 and the downward trend after 1995 by CMIP6 has actually occurred in the real world. The long-term decrease of the AMOC in the SSP simulations can be attributed to increased warming and freshening of high latitude North Atlantic surface waters under global warming, increasing their stability and decreasing deep convection and NADW formation (*high confidence*). Due to internal variability, there is *medium confidence* that in the coming century there still will be decades when increases occur.

A particular process influencing AMOC decline is the seesaw between NADW and AABW formation (Seidov, 2001; Kuhlbrodt et al., 2007; Martin et al., 2013; Jüling et al., 2018). In CMIP5 and CMIP6 AMOC decline is generally associated with an increased intrusion of modified AABW (*low confidence*), which is colder and fresher than NADW, to enter the North Atlantic from the tropical Atlantic, already visible before 2100, followed by (*low confidence*) a concomitant overall decreased northward export of AABW from the Southern Ocean after 2100 (Heuzé et al., 2015), which could help to enable AMOC recovery through the associated warming of North Atlantic deep waters if AABW decreases in volume.

An abrupt transition or collapse of the AMOC during the 21st century was assessed in AR5 as *unlikely*, since the vast majority of CMIP5 (and up till now all CMIP6) models do not project a transition; only two CMIP5 models displayed an almost complete AMOC collapse in the 21st century, FIO-ESM and FGOALS-s2, (Drijfhout et al., 2015; Sgubin et al., 2017). Here, we do not deviate from the previous A5 assessment, although there is *high confidence* that CMIP models are too stable due to tuning towards stability and model biases (Valdes, 2011; Mecking et al., 2017; Weijer et al., 2019). In addition, meltwater release from the Greenland Ice Sheet (GrIS) is not accounted for in the vast majority of scenario runs while satellite data indicate accelerated mass loss from the GrIS (Section 9.4.1.1). Since the mid-1990s there have been increases of 50% in freshwater contributions to the subpolar North Atlantic and a 50% decrease in Labrador Sea Water thickness (Yang et al., 2016b). Thus, there is *medium confidence* that GrIS melting and the volume of Labrador Sea Water are linked. There is *medium confidence* that GrIS melting has affected Labrador Sea convection since 2010 (Böning et al., 2016), but *low confidence* that this melting has already

affected the evolution of the AMOC over the 20th century, as claimed by e.g. (Rahmstorf et al., 2015; Yang et al., 2016b). Estimates suggest that GrIS melting can result in 5-15% more AMOC weakening under RCP8.5 (Lenaerts et al., 2015; Bakker et al., 2016; Golledge et al., 2019), which would imply stronger decreases in AMOC than simulated by the CMIP models neglected this effect. The sign of net freshwater transport by the AMOC across the southern boundary of the Atlantic has also been proposed as an indicator for AMOC stability (Huisman et al., 2010) but there is still debate on its usefulness (Gent, 2018). CMIP3 and CMIP5 ensembles were biased to an AMOC displaying northward freshwater transport instead of the observed southward transport (Drijfhout et al., 2011; Mecking et al., 2017), which, according to this metric, would point to a too stable AMOC. Overall, we have *high confidence* that existing model biases limit the accuracy of models displaying the correct sensitivity of the AMOC response to global warming (Swingedouw et al., 2013; Liu et al., 2017; Sgubin et al., 2017). However, there is on the other hand not sufficient support that without these biases the likelihood of an abrupt AMOC collapse before year 2100 is significantly increased.

Although an abrupt collapse of the AMOC before 2100 is *unlikely*, there is still deep uncertainty on AMOC evolution beyond 2100. After 2100 meltwater fluxes from Greenland and Antarctica affect the evolution of the AMOC. With an emulator to create a wide range of AMOC projections using the input of 8 climate models and incorporating Greenland meltwater fluxes, AMOC decline stabilizes in the RCP4.5 scenario but continues to decline an RCP8.5 scenario with 74% [+4%,100%] by 2290–2300, with a 44% likelihood of an AMOC collapse (Bakker et al., 2016). However, Antarctic melting was not accounted for and this may counteract an AMOC decline (van den Berk et al., 2019). It should be noted, however, that the relation between freshwater supply in the northern North Atlantic and AMOC-response is weak (Figure 9.10 bottom right panel). This can be understood from the perspective of a bi-stable AMOC, in which the freshwater needed to drive the AMOC from on to off-state, crucially depends on the background climate, that is, how far away the simulated AMOC is from its tipping point (Weijer et al., 2019).

Large cooling changes in the subpolar North Atlantic were simulated by 9 out of 40 CMIP5 models in their various RCP simulations (Sgubin et al., 2017). These changes can occur within a decade and can suddenly decrease the 21st century surface air temperature by several degrees over the North Atlantic and up to 2 degrees over Western Europe. Using the representation of stratification as an emergent constraint, it is *as likely as not* that rapid changes in subpolar convection and associated cooling will occur in the 21st century (Sgubin et al., 2017), because models with a better representation of stratification in the Labrador Sea and wider subpolar gyre more often simulate an abrupt collapse of Labrador Sea convection than models with a poor representation of the stratification (*low confidence*).

A major impact of the AMOC on climate is through its link with the AMV, a large-scale dipole in SST over the entire Atlantic (Knight, 2005) with a timescale of 60-80 years (Technical Annex, Section AVI.8). A similar mode appears in various models, with similarly long timescales ranging from 40-60 years to a century or more (Msadek and Frankignoul, 2009; Menary et al., 2012). This mode, which strongly projects on global mean surface temperature, however is underestimated in models, and as a result, low-frequency variations in global mean temperature are underestimated as well (*high confidence*) (Zhang et al., 2019b). Many climate models produce variability in the strength of the AMOC on shorter timescales and with positive and negative anomalies confined to the (subpolar) North Atlantic (Zhang and Wang, 2013; Ba et al., 2014; Xu et al., 2019). The mechanisms leading to different modes can be classified and appear strongly affected by model biases in the subpolar gyre (Menary et al., 2015). As a result, there is *low confidence* in the ability of models to simulate a realistic AMV and other AMOC variability, although some models perform (much) better than others in this respect. There is *medium confidence* that the low-frequency AMV mode in models is more strongly driven by stochastic atmospheric or external forcing (aerosols) (Clement et al., 2015), than in nature where the AMV appears dominated by the AMOC (Zhang et al., 2019b). Through underestimating the low-frequency AMOC-AMV link, climate models also underestimate internal variability in subpolar SST and NAO on multidecadal timescales (Kim et al., 2018). Therefore, we have *low confidence* that CMIP5/6 models correctly capture low-frequency natural variability over the North Atlantic.

The AMOC played an important role in past global scale climate change as demonstrated by the large climate transitions between cold stadial and warm interstadial phases, or “Dansgaard-Oeschger (D-O)

events” (Dansgaard et al., 1993), unrelated to orbital forcing. A strong link between AMOC variability and D-O events has been found in modelling studies (Lynch-Stieglitz, 2017), but the direct observational evidence for these links is still scarce, apart from the Younger Dryas and the last Heinrich event HS1 for which we have *high confidence* that switches between different modes of the AMOC occurred and were essential in explaining the concurrent change in temperature. Over longer time periods, AMOC reconstructions and modelling give *high confidence* that the AMOC can switch on and off; evidence includes the onset of a basin wide AMOC circa 33 million years ago switching from a Pacific Meridional Overturning Circulation (PMOC) (Hutchinson et al., 2019), a temporary return to a PMOC in the mid-Pliocene (Burls et al., 2017). Present-day boundary conditions inhibit a PMOC (Ferreira et al., 2018) and the AMOC seems to have prevailed over the last 2-3 million years. There is also evidence over the last 3,000 years of decadal to multi-centennial climate variations which could be linked to changes in AMOC deep water formation sites using high-resolution paleoceanographic data (Moffa-Sánchez and Hall, 2017). Despite these changes in deep-water formation the AMOC appears to have been remarkably stable after the Younger Dryas (Hoffmann et al., 2018; Lippold et al., 2019).

There is *low confidence* in the exact cause for mode switches in the AMOC between stadials and interstadials (Lynch-Stieglitz, 2017; Weijer et al., 2019). It appears that increasing temperatures and CO₂ concentration triggered abrupt AMOC-decreases by freshwater release from northern hemisphere icecaps on shorter timescales, but on longer timescales a decrease in sea-ice cover allowed for deeper convection in marginal seas like the Labrador and Nordic Seas than at the southern boundary of glacial sea-ice in the North Atlantic. Therefore the response of the AMOC to increasing CO₂ in colder climates appears opposite in decadal-centennial timescales to the centennial-millennial response (Weijer et al., 2019) *medium confidence*. In the warmer present-day climate, there is no indication for such an opposite response, although some CMIP5 models show AMOC recovery after 2100 for the RCP8.5 scenario (Collins et al., 2013) giving *low confidence* in the longer response of the AMOC to anthropogenic forcing.

9.2.3.2 Southern Ocean

The circulation of the Southern Ocean is composed of a strong eastward flowing current, the Antarctic Circumpolar Current (ACC), which transports approximately $173.3 \pm 10.7 \text{ } 10^6 \text{ m}^3 \text{ s}^{-1}$ (Donohue et al., 2016) of water eastward connecting the world’s three main ocean basins, and a comparatively weaker vertical overturning circulation, estimated to range from 10 to 20 $10^6 \text{ m}^3 \text{ s}^{-1}$ (Naveira Garabato et al., 2014; Cessi, 2019), which links the deep and shallow ocean. This three-dimensional circulation system has a strong influence on the global climate and its change (*high confidence*), through its central role (i) on global ocean heat content change: the region South of 30°S accounts for 49% (39-61%) of the total excess heat energy stored in the upper 2000 m of the oceans from 2005 to 2018 (*high confidence*) (Section 9.2.2.1); (ii) on global ocean anthropogenic carbon uptake: the region South of 30°S stores 40% (*high confidence*) (34-47%) of global ocean anthropogenic carbon inventory in 2016 (Lauvset et al., 2016); and (iii) on ocean-ice shelves interactions: its circulation is intimately associated with the melt of the Antarctic Ice Shelves by providing heat content available for basal melt (*high confidence*), which impacts the stability of the Antarctic Ice Sheet (Section 9.4.2) and in return the melt of the ice-shelves can feedback onto the large-scale circulation through widespread increase of the upper-ocean stratification (*high confidence*) (Fogwill et al., 2015; Bronselaer et al., 2018; Golledge et al., 2019; Lago and England, 2019; Schloesser et al., 2019a), or through facilitating heat transport toward the ice-shelf base (*low confidence*) (Hellmer et al., 2012; Donat-Magnin et al., 2017; Fyke et al., 2018; Silvano et al., 2018; Moorman et al., submitted).

The SROCC assessed observed Southern Ocean circulation change over the past few decades and the assessment is summarized here (Meredith et al., 2019). Despite observed increasing Southern Hemisphere westerly winds in the recent decades, there is no evidence that ACC transport has changed during the same period, and there is *medium confidence* that ACC transport is only weakly sensitive to changes in winds. New evidence since AR5 show that it is *unlikely* that the mean meridional position of the ACC has moved meridionally in recent decades. Theory, high-resolution ocean modelling, and observed increased of eddy activity from 1990 to 2010 suggest that the weak sensitivity of the ACC to winds is a consequence of eddy-saturation (*medium confidence*), where additional energy input from stronger winds goes into

1 increasing eddy activity rather than the intensity of ACC. The vertical upper ocean overturning rate has
2 experienced significant inter-decadal variability in response to wind forcing since the 1990s, and because of
3 this variability and the indirect nature of observational estimates, there is *low confidence* in the assessments
4 of a long-term increase in upper ocean overturning. The bottom ocean overturning rate is also difficult to
5 monitor but observed decreased volume of Antarctic Bottom Water in the global ocean (*medium confidence*)
6 is consistent with a slowdown of its circulation, which has potential implications on the rate of the AMOC.
7 Since SROCC there has been very limited additional observation-based evidence of Southern Ocean
8 circulation change. We here confirm SROCC assessment, with the one addition on the assessment of change
9 in the lower cell circulation and associated Antarctic Bottom Water volume, where new evidence suggest
10 that past observed change in volume transport in the Atlantic sector might be associated with decadal
11 variability (*low confidence*) (Abrahamsen et al., 2019). In addition to this assessment of observed circulation
12 change, Southern Ocean water-masses changes are described in Section 9.2.2.3, with a very large imprint on
13 global ocean heat and salt content change (Section 9.2.2.1).

14
15 Modelling the climate of the Southern Ocean is challenging for a number of reasons, including the
16 importance (*high confidence*) of parameterized mesoscale eddy processes in forming the horizontal and
17 vertical circulation, as well as equilibrating its future response under climate warming scenarios (Downes
18 and Hogg, 2013; Munday et al., 2013; Bishop et al., 2016; Gent, 2016; Poulsen et al., 2018), the absence of
19 ice-sheet coupling in the CMIP6 model suite implying a lack of freshwater feedback on the circulation from
20 ice sheet melting (Bronse laer et al., 2018; Golledge et al., 2019; Lago and England, 2019; Schloesser et al.,
21 2019a), and the poor representation of dense overflows from the Antarctic shelves feeding the lower
22 overturning cell (Section 9.2.3.5.2) (Snow et al., 2015; Dufour et al., 2017; Akhoudas et al., submitted).
23 Despite the above-mentioned challenges, the CMIP6 ensemble does represent the main Southern Ocean
24 circulation characteristics; the spread in CMIP6 ACC transport simulations is lower but consistent with
25 observation-based values considering spread and uncertainty envelopes (113–167 Sv across an ensemble of
26 23 CMIP6 models considered) (*medium confidence*) and the inter-model spread in ACC transport has greatly
27 reduced from previous generation of climate models from CMIP3 to CMIP6 (*high confidence*) (Meijers et
28 al., 2012; Beadling et al., 2019, submitted); and the structure (but not the magnitude) of the two-cell zonally-
29 averaged overturning is captured by most models (Sallée et al., 2013b; Russell et al., 2018; Beadling et al.,
30 2019) (*medium confidence*).

31
32 These progresses, along with a number of recent important progresses in process understanding, provides us
33 with a cluster of lines of evidences allowing us to reach *medium to high confidence* into qualitative
34 assessment of future changes for many aspects of the Southern Ocean circulation, as detailed in the
35 following paragraphs. However, the representation of Southern Ocean dynamics in climate model still suffer
36 from the number of yet unresolved challenges listed above so most quantitative predictions of future change
37 in circulation is ascribed *low confidence*.

38
39 Both observations and climate models studies provide *high confidence* that the ACC position and transport
40 will remain insensitive to predicted southward shift and intensification of winds (Section 9.2.1.3) (Meijers et
41 al., 2012; Chapman et al., submitted). There is *medium confidence* that the predicted increase in westerly
42 winds will lead to increased eddying activity as suggested by theory and observation (Meredith et al., 2019).
43 CMIP5 models show a wide variety of ACC responses to either medium or high radiative forcing scenario
44 (RCP4.5, RCP8.5), with future transport change ranging from -26 Sv to +17 Sv (*low confidence*), an inter
45 model variability mostly explained by inter model variability in future change of subpolar gyre area, with
46 models in which subpolar gyres extend northward associated with decreased ACC transport, and inversely
47 (Meijers et al., 2012). In contrast, CMIP5 models show consistently a southward shift of the northern
48 boundary of the ACC, caused by a southward shift of the core of the Southern Hemisphere subtropical gyres
49 by -1.88 ± 2.13 degree of latitude under RCP4.5 scenario and -3.20 ± 2.41 degree of latitude under RCP8.5
50 scenario (*low confidence*) (Sen Gupta et al., 2009a; Wang et al., 2011; Meijers et al., 2012).

51
52 Future intensified westerly winds will drive a stronger wind-driven upper overturning cell (*high confidence*)
53 partially compensated by an intensification of eddy-induced transport (*medium confidence*) (Downes and
54 Hogg, 2013; Morrison and Hogg, 2013; Munday et al., 2013; Sallée et al., 2013b; Doddridge et al., 2019).
55 There is therefore *medium confidence* in future increase of the upper cell overturning, but because eddies are

poorly represented in climate models and because of overall lack of confidence in Southern Ocean surface buoyancy fluxes and background density field in current climate models (Downes et al., 2018), we have only *very low confidence* in the quantification derived from CMIP5 models of an increase at the end of the 21st century of 2 ± 3 Sv under RCP4.5 and 3 ± 4 Sv under RCP8.5, of the transport of deep waters feeding the upper cell (Sallée et al., 2013b).

Under increased radiative forcing, there is *high confidence* that changes associated to air-sea exchange and ice melt, will reduce the density of the bottom water, up to a point where deep convection will reduce, and shelf water will be too light to sink to the bottom ocean, thereby slowing the lower cell overturning circulation (Sallée et al., 2013b; De Lavergne et al., 2014; Fogwill et al., 2015; Heuzé et al., 2015; Bronselaer et al., 2018; Golledge et al., 2019; Lago and England, 2019). Climate models of the CMIP5 generation consistently predict an overall reduction of 2 ± 2 Sv at the end of the century under RCP8.5, with large disparities across basins (greater and more systematic reduction across models in the Pacific sectors, compared to the Atlantic and Indian sector of the Southern Ocean) (Sallée et al., 2013b; Heuzé et al., 2015), but because of the important challenges presented above in the representation of bottom waters and their circulation in CMIP models, we have only *very low confidence* in such quantification of change. In addition, there is some indications (*low confidence*) that climate model at higher-resolution exhibit more severe slowing of the lower cell circulation than projected by coarse-resolution climate models (Newsom et al., 2016). Coarse-resolution climate model also overlook potential changes in deep interior mixing (*high confidence*), which participate with surface forcing in shaping the lower cell circulation (Garabato et al., 2007; Newsom et al., 2016; Downes et al., 2018). While predicted changes of the lower cell intensity from surface forcing are expected to overwhelm any changes of the lower cell through interior mixing (*medium confidence*) (Newsom et al., 2016), intensification of interior mixing (*low confidence*) (Sheen et al., 2014; Naveira Garabato et al., 2016) would increase the mixing of bottom water with lighter layer (Garabato et al., 2007; Akhondas et al., submitted), further reducing the amount of well-ventilated water reaching the world ocean basins to the north (*low confidence*). Open ocean polynya, which are very rarely observed, but extremely common in climate models (De Lavergne et al., 2014; Heuzé et al., 2015) are associated with high surface fluxes and water-mass transformation as well as unrealistic formation of bottom water which further limits our confidence in climate model projection of the lower cell. Limited observations of polynyas means that there is still today a weak understanding of their climate impact (Campbell et al., 2019; Zhang et al., 2019a). There is *low confidence* that a multidecadal-to-centennial mode of variability, linked to the opening of open ocean polynya, could explain observed Southern Ocean trends of the past decades (Zhang et al., 2019a), and *low confidence* that the presence of an open ocean polynya can have affect rate of melt of the Antarctic ice shelf (Naughten et al., 2019).

Projecting changes in the temperature of Antarctic shelves is a challenging task because potentially important controlling mechanisms are small and transient, which makes them difficult to observe at large-scale and are not resolved in most CMIP5/6 ocean models. Small-scale mechanisms include the heterogeneity of observed basal melt with warm water driving narrow basal channels that cut underneath the ice (Drews, 2015; Alley et al., 2016; Milillo et al., 2019), eddies and tides (Stewart et al., 2018; Jourdain et al., 2019; Hausmann et al., submitted) which can drive CDW onto the shelves or dynamically increase melting, and feedback mechanisms between ocean, atmosphere and cryosphere that can weaken or amplify initial perturbations (Donat-Magnin et al., 2017; Spence et al., 2017; Turner et al., 2017a; Silvano et al., 2018; Webber et al., 2019). Future projections of the Antarctic shelf relies on alternative tools like regional ocean models forced with future climate conditions simulated by CMIP5/6 models. Those projections and the associated melt rate are assessed in Section 9.4.2.2.

9.2.3.3 Tropical Oceans

Tropical ocean surface temperatures show warming trends from the 1950s, except in parts of the eastern tropical Pacific, which superimpose on strong internal interannual and decadal variability (Section 2.4) (*very high confidence*) (Figure 9.3). These changes are concurrent with a weakened atmospheric circulation (*high confidence*) (Section 8.2.1.3). The long-term warming of the tropical oceans is largely attributed to anthropogenic forcing and has contributed to an expansion of the tropics and the Indo-Pacific warm pool

(Allen et al., 2014; Weller et al., 2016; Staten et al., 2018; Roxy et al., 2019), recent occurrences of extreme ENSO events and marine heat waves (Santoso et al., 2017; Collins et al., 2019), and changes in oceanic circulation and heat transport (Forget and Ferreira, 2019), particularly since the 1980s.

Recent studies highlight a larger role of the tropical Indian (Izumo et al., 2010) and Atlantic Oceans (Ham et al., 2013; Cai et al., 2019) than the previous understanding of a dominant tropical Pacific influences in the pan-tropical ocean interactions (Terray et al., 2016; Cai et al., 2019; Hu and Fedorov, 2019). Tropical oceans are inter-connected via the zonal atmospheric Walker Circulation and the Indonesian Throughflow (ITF) connecting the Pacific and Indian Oceans – their responses to the changing climate involve prominent coupled ocean-atmosphere interactions (Cai et al., 2019) (*very high confidence*). Strengthened ITF transport during 1998–2013 (Nieves et al., 2015) (*high confidence*) have acted to cool the Pacific and warm the Indian Ocean (Collins et al., 2019). On a longer time-scale, there is *high confidence* that over the last two-decades, total water transport from the Pacific to the Indian Ocean via the Indonesian Throughflow has increased (Collins et al., 2019), although most models project a decline on the centennial timescale (Figure 9.11) as winds change (Sen Gupta et al., 2016, submitted).

Formally separating multi-decadal variability and anthropogenic warming in tropical oceans is challenging, particularly due to gaps in high quality long-term ocean observations (Stammer et al., 2019) and due to biases in climate models in accurately simulating the mean state and variability of the tropical oceans. Particularly, the heat budget for the tropical Indo-Pacific and the ITF is not well constrained, leading to uncertainties in the heat redistribution from the tropical oceans (Hermes et al., 2019). In the Atlantic, Pacific and Indian Oceans, CMIP5 models have biases in their tropical mean state (Chang et al., 2007; Richter and Xie, 2008; Kim et al., 2014; Li and Xie, 2014) with erroneous mean thermocline depth and slope along the equator, weak or erroneous structure of the equatorial current system, an excessive equatorial cold tongue in the Pacific (Li et al., 2015a, 2016a), and the “double ITCZ” precipitation pattern related to SST biases (Xiang et al., 2017).

Despite the biases, most CMIP5 models project increased frequency of extreme El Niño and La Niña events (*medium confidence*) and much enhanced impacts on global climate (Cai et al., 2014, 2015; Yeh et al., 2018), as well as increased frequency of positive Indian Ocean Dipole (IOD) events (*low confidence*) (Cai et al., 2013; Zheng et al., 2013) under RCP4.5 and RCP8.5 scenarios. The projected changes in these climate modes along with the rise in background temperatures may have substantial implications for marine heat wave characteristics in the future (Holbrook et al., 2019) (Cross-chapter Box 9.1). One of the clearest changes of ocean current transport simulated by climate model is a weakening of the Indonesian Throughflow projected in CMIP5 simulations under RCP4.5 and RCP8.5 scenarios (Hu et al., 2015a; Sen Gupta et al., 2016), as well as in CMIP6 simulations under SSP585 scenario (Sen Gupta et al., submitted), although observations in the past two decades have been variable (*high confidence*, Figure 9.11).

9.2.3.4 Western Boundary Currents and Gyres

The horizontal oceanic circulation in the upper ocean is characterized by basin scale gyres with intensified western boundary currents (Figure 9.11), referred to as subtropical (occurring in roughly 20°N to 40°N and 20°S to 40°S), subpolar (~40°N to ~70°N), and polar gyres (Arctic Ocean). These gyres are closely associated with overturning cells such as subtropical overturning cells and Atlantic and Southern Ocean meridional overturning cells. Both gyres and overturning cells carry heat, freshwater and carbon as mode and intermediate waters (Section 9.2.3), and these transports and water masses are sensitive to decadal variability and secular changes in the atmospheric circulation.

The gyres and associated western boundary currents (WBCs) depend on the wind stress curl which causes convergence in the upper ocean. Systematic changes in wind stress curl over both hemispheres (Figure 9.4), in particular, the poleward shift of the zero-curl line the North Pacific, the North Atlantic and the South Atlantic Oceans, may strengthen or shift the subtropical gyres poleward (Figure 9.11) which would accelerate warming poleward of the subtropical WBCs as suggested by some earlier studies (Cai, 2006; Roemmich et al., 2007; Chen and Wu, 2012; Wu et al., 2012; Zhai et al., 2014; Hu et al., 2015a; Yang et al.,

2016a). However, recent observations do not support such enhancement or poleward shift of WBCs during the past 20 years including the Kuroshio (Wang et al., 2016d; Wang and Wu, 2018), Gulf Stream (McCarthy et al., 2018; Dong et al., 2019), Agulhas Current (Beal and Elipot, 2016; Elipot and Beal, 2018), and East Australian Current (Sloyan and O’Kane, 2015). Multidecadal variability of strength and position of WBCs has been widely reported (McCarthy et al., 2018) and observed changes combine anthropogenic trends and decadal-interdecadal variability (Feng et al., 2010; Zhang and Church, 2012). There is *medium confidence* that intensifying winds increase the mesoscale eddy field of boundary currents, rather than their mean flow, in particular the Kuroshio, East Australian and Agulhas Current (Cetina-Heredia et al., 2014; Yan and Sun, 2015; Beal and Elipot, 2016). This effect is not expected in low-resolution models and has not yet been evaluated in high-resolution models. In terms of the gyres, observed changes in their circulation over the most recent decades are *likely* dominated by interannual and decadal modes of variability globally (Chen and Wu, 2012; Qiu and Chen, 2012; Melzer and Subrahmanyam, 2017; McCarthy et al., 2018). The North Atlantic subpolar gyre is strongly modulated by variability associated with the NAO and AMV (Robson et al., 2016). Subpolar gyre systems can change abruptly due to a positive feedback between convective mixing and salinity transport within the gyre (Born et al., 2013, 2016). In the Arctic, a strong increase of the Beaufort gyre is observed (Armitage et al., 2017), which might be partly due to increased wind stress, reduced ice thickness, changes morphology of ice pack, but at present there is *low confidence* in attributing causality to these changes for any of the proposed mechanisms. In the Southern Ocean, the subpolar gyres respond (*low confidence*) to Southern Hemisphere atmospheric modes of variability at interannual time-scale (Armitage et al., 2018; Dotto et al., 2018).

All climate models reproduce western boundary currents (WBCs) and gyres, but high resolution models show much better representation of these currents than low resolution models (Small et al., 2014b; Griffies et al., 2015; Chassignet et al., 2017; Hewitt et al., 2017; Roberts et al., 2018) (*very high confidence*). The typical resolution of CMIP5 models is around 100 km, which leads to weak WBCs and incorrect locations of subtropical and subpolar gyres (Figures 9.3 and 9.11). Recent studies indicate that increased resolution not only admits mesoscale eddies, but also improves simulation of the strength and position of WBCs such as the Kuroshio Current, Gulf Stream, and East Australian Current (Sasaki et al., 2004; Chassignet and Marshall, 2008; Delworth et al., 2012; Yu et al., 2012; Small et al., 2014a; Haarsma et al., 2016; Chassignet et al., 2017). Improvements to the boundary current location are closely related to model improvements in recirculation regions (Jayne et al., 2009), mean path and variability and existence of multiple stable paths (Qiu et al., 2005; Delman et al., 2015), air-sea fluxes (Small et al., 2014b), and related coastal weather patterns (Kaspi and Schneider, 2011). Some stand-alone high resolution oceanic models tend to overestimate mesoscale eddies and underestimate the strength of WBCs, which is partly attributed to a lack of mesoscale air-sea interaction (Ma et al., 2016; Renault et al., 2016), and partly attributed to incorrect representation of sub-mesoscale motions in these models (Zhang et al., 2016c) (*medium confidence*). High-resolution coupled ocean-atmosphere models have substantially weaker eddies than high-resolution ocean-only models (Renault et al., 2016; Oerder et al., 2018). As eddies potentially play a role in determining the strength of the gyre circulations and their low-frequency variability (Fox-Kemper and Pedlosky, 2004; Berloff et al., 2007), it is expected that high-resolution models will differ in the decadal variability and the sensitivity to changes in wind stress of their gyres from coarser-resolution models (*low confidence*), although some aspects of the fundamental balances setting gyre strength do not differ with resolution (Hughes and De Cuevas, 2001; Yeager, 2015). [PLACEHOLDER FOR CMIP6 RESULTS]

[START FIGURE 9.11 HERE]

Figure 9.11: (a) Mean barotropic streamfunction 1995-2014 and projected barotropic streamfunction change 2081-2100 versus 1995-2014 under (b) SSP1-2.6 and (c) SSP5-8.5. Units are Sv. (d) Mean surface (0-100 m) speed and projected surface speed change 2081-2100 versus 1995-2014 under (f) SSP1-2.6 and (g) SSP5-8.5. The units are m s^{-1} . (d,h) Median and *likely* range of 1995-2014 and 2081-2100 transport of 6 currents with the largest transport change or largest fractional change (Sen Gupta et al., submitted). (d) Deep currents: Agulhas Leakage (ACLeak), Gulf Stream (GS), Gulf Stream Extension (GSx), Tasman Leakage (TASL), East Australia Current Extension (EACx), and Indonesian Throughflow (ITF). (h) Shallow currents: as for deep but with New Guinea Current (NGC), and without GSx.

[END FIGURE 9.11 HERE]

Subtropical gyres generally strengthen under RCP4.5 and RCP8.5 scenarios and all expand southward in the Southern Hemisphere (Sen Gupta et al., 2009b, submitted; Hu et al., 2015a), which are also observed in CMIP6 simulations under SSP5-8.5 scenario (Figure 9.11). Numerical simulations indicate that ozone depletion and anthropogenic CO₂ emission contribute to the poleward intensification of the Southern Hemisphere basin scale gyre circulation (Cai et al., 2005; Cai and Cowan, 2007; Thompson et al., 2011). Driven by an expansion of the Hadley cell and poleward shift and enhancement of extratropical wind stress, global WBCs estimated from simple ocean data assimilation (SODA) tend to intensify and stronger warming tends to occur in WBC regions during the past century (Wu et al., 2012). In the North Atlantic, however, a weakened AMOC tends to reduce the Gulf Stream (Yang et al., 2016). Since the resolution of CMIP5 models and reanalysis systems do not represent mesoscale eddies and western boundary currents well, there is *low confidence* in these conclusions, in particular in the projected response of gyres to changes in winds when the energy pathway from mean to eddy kinetic energy is not (well) accounted for in low-resolution ocean components. Thus, these projections should be supported by direct observations (which at present is not the case) and higher resolution models. [PLACEHOLDER FOR CMIP6 RESULTS]

9.2.3.5 Upwelling, Coastal, and Marginal Sea Systems

9.2.3.5.1 Eastern Boundary Upwelling Systems

Eastern boundary upwelling systems (EBUS) exist where trade winds draw cold and generally less-ventilated low-pH/low-oxygen waters upward. Coastal upwelling plays a key role in supplying the food chain with nutrients, hence the richness and productivity of EBUS (Chapter 5). Coastal upwelling depends on the strength and spatio-temporal patterns of the wind component that is parallel to the shoreline. Several other physical processes modulate the intensity and variability of upwelling: the wind stress curl drives offshore vertical transport through Ekman suction and pumping; an along-shore pressure gradient that induces across-shore flow sometimes counteracting coastal upwelling (Hickey and Pola, 1983; Marchesiello and Estrade, 2010; Jacox et al., 2018); interaction of flow with bathymetry to modulate upwelling (Janowitz and Pietrafesa, 1982; Graham and Largier, 1997; Ndoye et al., 2017). The background oceanic stratification (Di Lorenzo et al., 2005) (Section 9.2.3.3), air-sea buoyancy fluxes (Section 9.2.1.4), and meso-/submeso-scale turbulence modulates the ocean response to a given wind forcing (Gruber et al., 2011; Renault et al., 2016), which can feed back onto the wind regime (Boé et al., 2011; Alves et al., 2018). These processes involve fine spatial scales which can only be properly represented with sufficient horizontal resolution in the ocean and the atmosphere (Small et al., 2015; Roberts et al., 2019).

Wind trends for the past decades have been investigated in numerous studies. Frequent inconsistencies between their findings have been reported and related to differences in wind products, season of interest, and length of the considered time series (Varela et al., 2015). In addition, these trends might only reflect natural climate variability since times of emergence for the anthropogenic upwelling is predicted to be beyond 2040-2050 (*low confidence*) in the example of the California Current system (Brady et al., 2017). However, while no consistent pattern of contemporary changes of EBUS circulation emerge from observation-based studies, numerical and theoretical work suggest that EBUS winds will change with a dipole spatial pattern within each EBUS of reduction (weaker and/or shorter) at low latitude, and enhancement (stronger and/or longer) at high latitude (Figure 9.12). Despite differences among EBUS (Wang et al., 2015) there is *high confidence* in this pattern of change for reasons detailed in this paragraph. Investigations of future physical changes have mainly focused on changes in upwelling winds. Specifically, EBUS winds are the dynamical consequence of pressure gradient forces between an offshore subtropical high and a heat low located on land. Upwelling winds have been hypothesized to intensify under climate change as a consequence of the projected increase in air temperature contrast between ocean and land (Bakun, 1990). This hypothesis is being increasingly challenged. Changes in sea level pressure and wind fields in EBUS appear to be primarily tied to those affecting subtropical highs (Garcia-Reyes et al., 2013). Poleward expansion of the Hadley cell (Staten et al., 2018) and the related poleward migration of subtropical highs (He et al., 2017; Cherchi et al., 2018), produce robust patterns of changes of reduced upwelling at low latitude and enhanced upwelling at

high latitude (Echevin et al., 2012; Belmadani et al., 2014; Bettencourt et al., 2015; Rykaczewski et al., 2015; Sousa et al., 2017; Sylla et al., 2019). These patterns are most apparent in summer. Synoptic variability of upwelling winds, which is increasingly recognised as important in the functioning of upwelling ecosystems (García-Reyes et al., 2014), may also be affected by climate change (Aguirre et al., 2019).

Predicted future annual cumulative upwelling wind changes at most locations and seasons remain within $\pm 10\text{--}20\%$ from present-day values, even in the context of high-end emission scenarios ($4\times\text{CO}_2$ or RCP8.5) (*medium confidence*, Figure 9.12). The magnitudes of upwelling changes due to wind stress curl and alongshore pressure gradients have only been considered in a few studies but do not seem to invalidate conclusions drawn from alongshore wind changes alone (Oerder et al., 2015b; Sylla et al., 2019). Direct estimation of oceanic upward transport provides a potentially more meaningful estimator of upwelling that integrates the effect of all relevant processes (Oyarzún and Brierley, 2018; Sylla et al., 2019). However, scarce availability of vertical velocity fields from climate models, plus lack of confidence in them in coarse-resolution climate models, limit this approach. Climate change induced modifications in upper ocean stratification (Section 9.2.1.4) are projected to induce an increased confinement of upwelling vertical velocities near the ocean surface (Oerder et al., 2015b; Oyarzún and Brierley, 2018) (*high confidence*).

[START FIGURE 9.12 HERE]

Figure 9.12: [PLACEHOLDER FOR REGENERATING USING REANALYSIS and CMIP6 MODELS INCLUDING HIGH-RES] Upwelling-favourable winds stress (τ_{upw}) in the four main upwelling systems (Rykaczewski et al., 2015). (a) 1985–2014 climatological mean intensity of τ_{upw} from atmospheric reanalysis; (b) 1985–2014 historical mean τ_{upw} as simulated by CMIP6 models. Change τ_{upw} from 1985–2014 to 2081–2100 predicted under (c) SSP126 scenario and (d) SSP585 scenario.

[END FIGURE 9.12 HERE]

9.2.3.5.2 Coastal Systems and Marginal Seas

Beyond the world's coastlines lie shallow estuaries, continental shelves, and deeper fjords and slopes, where depths increase rapidly from the shelves to the deep ocean floor. It is more difficult to transport fluid across the shelf-break or slope than along (Brink, 2016), and estuaries and shelves have complex circulations and mixing. These coastal processes result in only indirect connections between the inner shelves and coastlines and offshore conditions. The early chapters of this report focus on large-scale metrics of climate, while the later chapters focus on regional change. Coastal processes act as a link in both directions, from changing rivers and estuaries, melt and runoff to deep water, and how changes in deep water regions affect regional and coastal conditions.

Shelf-deep ocean exchanges involve eddying, tidal, or turbulent motions and small-scale topography such as submarine canyons, which require high resolution observations to detect and high resolution models to simulate (Greenberg et al., 2007; Capet et al., 2008; Allen and Durrieu de Madron, 2009; Colas et al., 2012; Trotta et al., 2017). Example coastal processes that introduce uncertainty into large-scale projections are exchange of CDW across the Antarctic shelf-break, which affects AABW formation and Antarctic ice shelf-ocean interaction (Stewart and Thompson, 2013, 2015), river and estuarine plumes and their responses to water level and hydrology change (Banas et al., 2009; Sun et al., 2017b), fjord dynamics linked to glacial outflows (Straneo and Cenedese, 2015), and changing formation of water masses in marginal seas (Kim et al., 2001; Greene and Pershing, 2007; Giorgi and Lionello, 2008; Renner et al., 2009). Examples of climate change downscaling to the local level allow process detail (Foreman et al., 2013; Mathis and Pohlmann, 2014; Meier, 2015; Tinker et al., 2016). Due to the complexity of shelf-deep ocean exchanges and their small scale, linking coastal ocean changes to global ocean changes with high resolution modelling (Holt et al., 2017, 2018), or nesting and local mesh refinement (Fringer et al., 2006; Zhang and Baptista, 2008; Mason et al., 2010; Dietrich et al., 2012; Hellmer et al., 2012; Ringler et al., 2013; Wang et al., 2014b; Zängl et al., 2015; Zhang et al., 2016b). Some processes require these high-resolution regions to be coupled to the larger

scales, atmosphere, cryosphere, or hydrosphere (Seo et al., 2007, 2008; Somot et al., 2008; Oerder et al., 2015a; Renault et al., 2016; Zhang et al., 2016a), many require tides to be added or forced (Janeković and Powell, 2012; Timko et al., 2013; Pickering et al., 2017), while others can be downscaled in ocean or cryosphere without coupling (Holt et al., 2016; Quante and Colijn, 2016; Mathis et al., 2018). Coarse climate models and even the HighResMIP models at near 10 km ocean resolution do not represent some coastal phenomena such as cross-shelf exchanges and sub-mesoscale eddies which require 1km or finer resolution. Thus, there is *low confidence* in projecting centennial scale coastal climate change where regional downscaling has not been implemented or assessed. There is *high confidence* in the ability of regional coupled models to improve coastal climate change process understanding and provide regional information (Section 12.4), but many sites globally await such projections to provide an assessment.

9.2.4 Impact of Ocean Changes on Sea Level

Local changes in temperature, salinity and circulation affect the sea level height. Local changes in column-integrated ocean heat and freshwater leave their imprint on local sea level through their effects on seawater density (steric change, Box 9.1). Heat and freshwater changes can arise through changes in air-sea fluxes (Section 9.2.1.3), through variations in ocean circulation (Section 9.2.3) and indirectly via the associated transports of heat and freshwater, and mixing processes (Figure 9.13, Box 9.1). Changes in the barotropic circulation also affect local and regional sea level (Figure 9.11).

Changes in globally averaged ocean heat content cause ocean thermal expansion, which contributes to global mean sea-level (GMSL) rise (Box 9.1). The observed increased ocean heat content for 1971-2015 of 367 ± 51 ZJ has led to a global mean thermosteric sea-level rise of 0.04 ± 0.01 m on a total GMSL of 0.09 ± 0.03 m (Section 2.3.3.1, Table 2.7; Table 9.4; Cross-Chapter Box 9.2). The future projection of global ocean heat content in 2100 (Section 9.2.2.1) translates into global mean thermosteric sea-level rise of 0.15 (0.13-0.17) m for SSP1-2.6, 0.19 (0.17-0.22) m for SSP2-4.5, 0.24 (0.23-0.26) m for SSP3-7.0, and 0.29 (0.28-0.31) m for SSP5-8.5 by 2100 (Section 9.6.3.2.1; Figure 9.29, Table 9.7). Both quantities show substantial spread across the model ensemble, even though the multi-model mean average removes a large part of the individual model errors. Changes in ocean heat content of deeper water masses shows considerable spread, associated with model biases in deep water formation, but their effect on SSH is smaller.

Characterization of the spatial patterns of relative sea-level (RSL) variability has been possible since the early 1990s from satellite altimeter measurements, with the largest increase in RSL in tropical and subtropical basins (Figure 9.14). Satellite measurements (<http://www.aviso.altimetry.fr>) show that regional ocean dynamic sea-level trends can strongly deviate from global mean sea-level trend by up to 100% on annual to decadal timescales. On local and regional scales, ocean dynamic sea-level changes are dominated by natural fluctuations in currents, ocean heat uptake, precipitation and evaporation. Eddy-resolving ocean models show much stronger variability, similar to observations (Figure 9.14), than coarse-resolution models (Llovel et al., 2018; Chassignet et al., submitted). Attributing the causes of observed ocean dynamic changes on a regional scale remains challenging, due to the strong imprint of internal modes of variability in the observed RSL change (Hamlington et al., 2014; Han et al., 2017). There is *medium confidence* that regionally, anthropogenically forced changes may already be emerging (Bilbao et al., 2015; Fasullo and Nerem, 2018) (Section 3.5, Section 9.6.1.5).

Better observations (from Argo and satellite altimeters), increased computing resources, improved model resolutions and improved process understanding have led to a better model representation of the rate and distribution of thermosteric and ocean dynamic sea-level change. A variety of processes and thus model parameterizations impact ocean dynamic and thermosteric sea-level change (Griffies and Greatbatch, 2012) and higher resolution models distribute heat differently and have considerably higher dynamic variability (Griffies et al., 2014). Together with scenario uncertainty, these are the primary sources of uncertainty in thermosteric and ocean dynamic sea-level change. As a result, projections of ocean dynamic and thermosteric sea-level change by coupled climate models show a substantial spread. The model disagreement appears especially large in high latitudes and the North Atlantic (Yin, 2012; Slangen et al., 2014a). In addition, a robust seesaw arises in the North Atlantic with a sea-level increase north of 40N and a smaller

sea-level increase between 10°N and 40°N, reflecting the footprint of a decreasing AMOC. The small-scale ocean dynamic sea-level change, especially near western boundary currents, as observed by satellites on timescales from seasonal to interannual is only captured by high-resolution ocean models (Figure 9.12) (Chassignet et al., submitted).

Future projections of RSL change on annual to decadal timescales diverge due to internal modes of variability dominating ocean dynamic RSL change: the arbitrary phase of the fluctuations results in a large model spread (Pardaens et al., 2011; Gregory et al., 2016). At basin scales, and on decadal-centennial timescales, the projected ocean dynamic RSL change is more robust, because patterns associated with internal fluctuations tend to average out. There is *high confidence*, that various climate models would be able to correctly simulate the anthropogenic pattern of ocean dynamic sea-level rise in a large (at least 30 member) single-model ensemble (Fasullo and Nerem, 2018). It is more difficult to extract the anthropogenic signal from the CMIP5/6 multi-model ensembles, as these do not separate internal variability and model uncertainty well. However, there is *medium confidence* that using a special technique a common pattern from CMIP5/6 can be extracted that matches with the observed trends (Marcos and Amores, 2014). At the end of the 21st century, thermosteric changes will be largest in the North Atlantic and Southern Ocean basins (*medium confidence*); halosteric changes will display a basin-scale pattern with decreased sea-level in the Atlantic and increased in the Pacific (*medium confidence*); and ocean dynamic sea-level rise will be most pronounced in the subtropical gyre (associated with spin up of the gyres (*low confidence*); Section 9.2.3.4), and along the western North Atlantic (associated with a decreasing AMOC (*high confidence*) (Chen et al., 2019) (Figure 9.13).

[START FIGURE 9.13 HERE]

Figure 9.13: CMIP6 multi-model mean projected change in ocean dynamic, thermosteric, and halosteric sea level between 1984-2015 and 2081-2100 using the a method that does not require a reference level (Landerer et al., 2007). The unit is cm.

[END FIGURE 9.13 HERE]

[START FIGURE 9.14 HERE]

Figure 9.14: Standard deviation of SSH from (left) Aviso observations, and standard deviation of total (thermo/halo/dyn) SSH from (top middle) historical CMIP, (bottom middle) historical HighResMIP, (top right) SSP585 projections of CMIP, and (bottom right) SSP585 projections of HighResMIP.
[PLACEHOLDER FOR UPDATE WHEN MORE MODELS ARE UPLOADED]

[END FIGURE 9.14 HERE]

9.3 Sea ice

9.3.1 Arctic Sea Ice

The observed decrease of Arctic sea ice is a key indicator of large-scale climate change. This section expands and explains the sea-ice related summary findings from chapters 2 to 4 by providing a more detailed assessment including regional aspects, seasonal aspects and process understanding. The findings confirm or strengthen the Arctic sea-ice related findings of AR5, SROCC and SR1.5, unless otherwise noted.

9.3.1.1 Arctic Sea-Ice Coverage

Sea-ice coverage of the Arctic Ocean has decreased in all months since at least 1979 (Figure 9.15) (Serreze and Stroeve, 2015; Stroeve and Notz, 2015, 2018). This finding is based on measurements by passive

microwave satellite sensors that provide near-continuous measurements of pan-Arctic sea-ice coverage from 1979 onwards. Irreducible uncertainties in the conversion of thermal microwave brightness temperature to sea-ice concentration and algorithm design choices cause uncertainty in observed Arctic sea-ice coverage and sea-ice loss (e.g., Comiso et al., 2017b; Niederrenk and Notz, 2018; Kern et al., 2019; Meier and Stewart, 2019). However, these uncertainties are far smaller than the observed sea-ice loss. The absolute and the relative ice loss are highest in late summer and early autumn (*high confidence*). Averaged over the decade 2009–2018, monthly-average Arctic sea-ice area in August, September and October has been around 2 million km² (or about 25%) smaller than that during 1979–1988 (*high confidence*) (Section 2.3.2.1 explains why sea-ice area rather than sea-ice extent is emphasized in this report).

[START FIGURE 9.15 HERE]

Figure 9.15: Left: Absolute anomaly of monthly-mean Arctic sea-ice area during the period 1979 to 2018 relative to the average monthly-mean Arctic sea-ice area during the period 1979 to 2008. Based on the average sea-ice area derived from OSISAF/CCI (OSI-450 for 1979–2015, OSI-430b for 2016–2018)(Lavergne et al., 2019), NASA Team (version 1, 1979–2018) (Cavalieri et al., 1996) and Bootstrap (version 3, 1979–2018) (Comiso, 2017) sea-ice concentration estimates. Right: Sea-ice coverage in the Arctic as given by the average of the three most widely used satellite-based estimates for March and September, which usually are the months of maximum and minimum sea-ice coverage, respectively. First column: Mean sea-ice coverage during the decade 1979–1988. Second column: Mean sea-ice coverage during the decade 2009–2018. Third column: Absolute change in sea-ice concentration between these two decades, with stippling indicating significant differences. Fourth column: number of available CMIP6 models that simulate a mean sea-ice concentration above 15 % for the decade 2045–2055. The observational records are based on the average sea-ice coverage of OSISAF/CCI (OSI-450 for 1979–2015, OSI-430b for 2016–2018)(Lavergne et al., 2019), NASA Team (version 1, 1979–2018)(Cavalieri et al., 1996) and Bootstrap (version 3, 1979–2018)(Comiso, 2017).

[END FIGURE 9.15 HERE]

Loss of Arctic sea ice varies by region and season; it is most pronounced during winter in the Barents Sea and the Sea of Okhotsk (Figure 9.15) (Chen et al., 2016; Onarheim et al., 2018; Peng and Meier, 2018). This winter sea-ice loss explains the decrease in area covered intermittently each year by sea ice (Bliss et al., 2019). Summer ice losses occur primarily at the summer sea-ice region margins, in particular in the East Siberian, Chukchi, Kara and Beaufort Seas (Frey et al., 2015; Chen et al., 2016; Onarheim et al., 2018; Peng and Meier, 2018). This loss of summer sea-ice area is larger than the loss of winter sea-ice area, implying that the seasonal ice zone that is only covered by sea ice in winter has expanded regionally (Bliss et al., 2019) and pan-Arctic wide (Steele and Ermold, 2015) (*high confidence*). Consistently, the average sea-ice age and fraction of multi-year ice have decreased (*high confidence*) (Tschudi et al., 2016; Stroeve and Notz, 2018). Landfast-ice coverage has also decreased, particularly around Svalbard (Muckenhuber et al., 2016; Pavlova et al., 2019). Expanding Arctic sea-ice cover in the Bering Sea during winter was observed until 2017 (Frey et al., 2015; Onarheim et al., 2018; Peng and Meier, 2018), but a marked reduction in sea-ice coverage has occurred since then (Stabeno and Bell, 2019).

Arctic sea ice retreat includes an earlier onset of surface melt in spring and a later freeze up in fall, increasing the open-water season duration in the seasonal sea-ice zone (Stroeve and Notz, 2018). However, only *low confidence* in quantification of regional trends of melt and freeze onset is assessed because different observational products do not concur (Bliss et al., 2017).

Reconstructions of Arctic sea-ice coverage put the satellite period changes into centennial context. Direct observational data coverage warrants *high confidence* that the low Arctic sea-ice area of summer 2012 is unprecedented since 1850 (Titchner and Rayner, 2014; Walsh et al., 2017). Direct wintertime observational data coverage before 1953 is too sparse to reliably assess Arctic sea-ice area. Since 1953, the years 2015 to 2018 had the four lowest values of March Arctic sea-ice area, which usually is the month of maximum ice coverage (Figure 2.18) (*high confidence*). Reconstructions of Arctic sea-ice area before 1850 remain sparse,

but as in AR5 and SROCC, there remains *medium confidence* that the current pan-Arctic sea-ice loss is unique during the past 1450 years (Kinnard et al., 2011; De Vernal et al., 2013b).

New paleo-proxy techniques indicate regional sea-ice changes over epochs and millennia. Biomarker IP25 (Belt et al., 2007) together with other sedimentary biomarkers (Belt, 2018) provide local temporal information on seasonal sea-ice coverage, permanent sea-ice coverage and ice-free waters with occasional ambiguous contrasting results (Belt, 2019). These records and other paleo proxies including bromine (Spolaor et al., 2016) and dinocyst assemblages (e.g., De Vernal et al., 2013b) give *high confidence* that in some regions sea ice fluctuations have exceeded internal variability at least since the LIG (130ka), indicating that sea-ice responds to climate forcing.

The inferred changes over millennia can be related to Northern-hemisphere temperature evolution. Consistent, Arctic-wide changes give *high confidence* in relative changes including extensive sea-ice coverage during the Younger Dryas (~12 ka) in the Bering (Méheust et al., 2018), Kara (Hörner et al., 2018), Laptev (Hörner et al., 2016) and Barents (Belt et al., 2015) Seas; little sea ice during the early Holocene when Northern hemisphere summer insolation was higher than today (8000 to 9000 years before present) in the North Icelandic Shelf area (Cabedo-Sanz et al., 2016; Xiao et al., 2017), Sea of Okhotsk (Lo et al., 2018), Canadian Arctic (Spolaor et al., 2016), Barents (Berben et al., 2017), Bering (Méheust et al., 2018), and Chukchi (Stein et al., 2017b) Seas, and north of Greenland (Funder et al., 2011); increasing sea-ice cover throughout much of the middle and late Holocene, often peaking during the Little Ice Age, around Svalbard (Knies et al., 2017), in the North Icelandic Shelf area (Cabedo-Sanz et al., 2016), Western Greenland (Kolling et al., 2018), Barents (Belt et al., 2015; Berben et al., 2017), Chukchi (De Vernal et al., 2013a; Stein et al., 2017b) and Laptev (Hörner et al., 2016) Seas. Isolated evidence provides *low confidence* that the Arctic Ocean was covered by sea ice during the last interglacial (120 ka) (Stein et al., 2017a), that the Canadian Arctic had less sea ice during the early Holocene than today (Spolaor et al., 2016), and that present spring sea-ice coverage west of Svalbard is at its lowest during the past two millennia (Cabedo-Sanz and Belt, 2016).

The satellite record of Arctic sea-ice area from 1979 onwards is likewise strongly linearly correlated with global mean temperature (Figure 9.16a,e) (e.g., Gregory et al., 2002; Rosenblum and Eisenman, 2017) (*high confidence*). The correlation holds across all months with R^2 ranging from 0.61 to 0.81 (Niederrenk and Notz, 2018). However, in contrast to paleo-records, sea-ice fluctuations during the satellite period are only weakly correlated with Northern Hemisphere insolation (Notz and Marotzke, 2012); modern Northern Hemisphere sea-ice area is more strongly correlated with atmospheric CO₂ concentration (Johannessen, 2008; Notz and Marotzke, 2012) and cumulative anthropogenic CO₂ emissions (Figure 9.16b,f) (Zickfeld et al., 2012; Herrington and Zickfeld, 2014; Notz and Stroeve, 2016) (*high confidence*). R^2 values of the correlation between sea-ice area and cumulative CO₂ emissions range across all months from 0.76 to 0.92 (Stroeve and Notz, 2018). Because of conceptual, physical understanding of similarly high correlations between cumulative CO₂ emissions and sea-ice area in all CMIP5 and CMIP6 model simulations, and formal attribution studies (Section 3.4.1) it is *very likely* that more than half of the observed Arctic sea-ice loss in summer is anthropogenic.

In addition to climate change forcing, internal variability substantially affects Arctic sea ice, evidenced from both the paleo record (Cabedo-Sanz et al., 2016; Chan et al., 2017; Hörner et al., 2017; Kolling et al., 2018) and satellites after 1979 (e.g., Castruccio et al., 2018; Jahn et al., 2016; Kirchmeier-Young et al., 2017; Notz, 2015; Swart et al., 2015) (*high confidence*). Internal variability is estimated to have contributed 30 to 50% of the observed Arctic summer sea-ice loss since 1979 (Kay et al., 2011; Stroeve et al., 2012; Ding et al., 2017, 2019). However, this estimate from models might be biased toward internal over forced variability because the CMIP5 simulated September sea ice sensitivity to forcing is lower than observed (Notz and Stroeve, 2016; Rosenblum and Eisenman, 2017). In CMIP6, the ensemble mean well captures the observed sensitivity of sea-ice loss to CO₂ emissions, but most models still fail to capture the sensitivity of sea-ice area to CO₂ emissions and temperature correctly simultaneously (Figure 9.16h; (SIMIP Community, submitted, submitted)). Furthermore, climate models exhibit a wide range of internal variability for Arctic sea-ice area (e.g., Olonscheck and Notz, 2017), with CMIP6 estimates of internal variability for pre-industrial September sea-ice area ranging from $\sigma=0.4$ to more than $\sigma=0.8$ million km² (SIMIP Community,

submitted, submitted)(Section 3.4.1).

The understanding of physical drivers of regional change in Arctic sea ice has improved since AR5 (SROCC). Consensus across existing studies implies *high confidence* that variations in air temperature (Frey et al., 2015; Lynch et al., 2016; Mudryk et al., 2018), moisture transport (Park et al., 2015; Mortin et al., 2016; Woods and Caballero, 2016; Lee et al., 2017b), wind patterns (Frey et al., 2015; Smedsrud et al., 2017; Moore and McNeil, 2018; Graham et al., 2019), cloud coverage (Cox et al., 2016; Letterly et al., 2016; Mortin et al., 2016), downwelling longwave radiation (Park et al., 2015; Kapsch et al., 2016; Mortin et al., 2016; Ding et al., 2017), and oceanic heat flux (Nakanowatari et al., 2014; Cornford et al., 2015; Hutchings et al., 2015; Timmermans, 2015; Polyakov et al., 2017; Wang et al., 2018) all contribute to the regional evolution of sea-ice coverage. However, sparse observations, disagreement among CMIP5 models (Burgard and Notz, 2017) and limited understanding of the variation of each factor with global mean temperature implies *low confidence* in the relative contribution of each physical cause to sea-ice projections.

There is *high confidence* that Arctic summer sea ice area will remain highly correlated with global mean temperature until the ice has vanished (Li et al., 2013a; Rosenblum and Eisenman, 2017; Niederdrenk and Notz, 2018) (Figure 9.16a,e). Observations of sea-ice evolution and global mean temperature, observationally constrained model estimates with CMIP5 models, and CMIP6 simulations all show that a reduction of September-mean sea-ice area to below 1 million km², effectively a sea-ice-free Arctic Ocean, is more probable for a global mean warming of 2°C compared to global mean warming of 1.5°C (Kjellsson et al., 2015; Screen and Williamson, 2017; Jahn, 2018; Niederdrenk and Notz, 2018; Ridley and Blockley, 2018; Sigmond et al., 2018; SIMIP Community, submitted, submitted) (*high confidence*). However, because of CMIP5 and CMIP6 models' generally too low sensitivity of sea-ice loss to global warming, there is only *low confidence* regarding the specific warming level at which the Arctic Ocean becomes sea-ice free for the first time (SIMIP Community, submitted, submitted).

[START FIGURE 9.16 HERE]

Figure 9.16: Monthly mean March (a-d) and September (e-h) sea-ice area as a function of global surface air temperature (GSAT) anomaly (a,e); cumulative anthropogenic CO₂ emissions (b,f); year (c,g) in CMIP6 model simulations (shading, ensemble mean as bold line) and in observations (black dots). Panels d and h show the sensitivity of sea-ice loss to anthropogenic CO₂ emissions as a function of the modelled sensitivity of GSAT to anthropogenic CO₂ emissions (after SIMIP Community, submitted, submitted).

[END FIGURE 9.16 HERE]

Based on CMIP6 simulations, the Arctic Ocean will *likely* become practically sea-ice free throughout September for the first time for future CO₂ emissions of less than 1000 Gt and before the year 2050 in all emission scenarios considered here (SIMIP Community, submitted, submitted). This is consistent with an observation-based projection of a near-ice free Arctic Ocean in September for additional anthropogenic CO₂ emissions of 800± 330 GtCO₂ beyond the year 2018 (Notz and Stroeve, 2018; Stroeve and Notz, 2018). This estimate may, however, be conservative due to neglecting possible future reduction in atmospheric aerosol load (Gagné et al., 2015a; Wang et al., 2018) and is subject to the same constraints as the carbon budget analysis for global mean temperature (see section 5.5 for details).

CMIP5 model projections and reanalyses show a large spread of simulated regional Arctic sea-ice concentration (Laliberté et al., 2016; Chevallier et al., 2017), and they have difficulty simulating realistic landfast sea ice (Laliberté et al., 2018). This implies *low confidence* in the models capturing the regional processes governing Arctic sea-ice evolution and *low confidence* in future regional sea-ice evolution estimates, including projected low sea-ice concentration, for example for shipping routes across the Northern Sea Route and Northwest Passage (Stephenson and Smith, 2015; Wang and Overland, 2015; Laliberté et al., 2016; Melia et al., 2016).

CMIP5 models also have issues with capturing the seasonal cycle of observed changes in Arctic sea-ice drift speed, which impacts their simulation of regional sea-ice concentration patterns. Direct measurements of Arctic sea-ice from drift buoys and satellites show with *high confidence* that drift speed of Arctic sea ice has increased over the satellite period in all seasons (Rampal et al., 2009; Spreen et al., 2011; Kwok et al., 2013; Laxon et al., 2013; Olason and Notz, 2014; Docquier et al., 2017). CMIP5 models generally capture the observed acceleration of Arctic drift speed throughout winter. In summer, however, CMIP5 models show a slowdown of Arctic sea-ice drift rather than the observed acceleration (Tandon et al., 2018). The drift acceleration is primarily caused by the decrease in sea-ice concentration and sea-ice thickness both in the observational record (Rampal et al., 2009; Spreen et al., 2011; Olason and Notz, 2014; Docquier et al., 2017) and in CMIP5 models (Tandon et al., 2018) (*high confidence*). Changes in wind speed are less important for the observed large-scale changes (Spreen et al., 2011; Vihma et al., 2012; Olason and Notz, 2014; Docquier et al., 2017; Tandon et al., 2018).

There is *high confidence* that no “tipping point” or critical threshold exists in global mean temperature beyond which the loss of summer sea ice becomes self-accelerating and irreversible. Recovery of summer ice cover through stabilizing feedbacks from increased heat fluxes through thin ice and thin snow and from increased emission of longwave radiation from open water dominate positive feedbacks such as the ice-albedo feedback (Tietsche et al., 2011; e.g., Eisenman, 2012; Wagner and Eisenman, 2015) (see section 7.4.2 for details on the individual feedbacks). Relatedly, observed and modelled Arctic summer sea ice and global mean temperature are linked with little temporal delay (Armour et al., 2011; Ridley et al., 2012; Li et al., 2013a; Jahn, 2018). In contrast, winter sea-ice loss versus warming in some, but not all, CMIP5 projections eventually accelerates for very high warming levels (Winton, 2006; Armour et al., 2011; Ridley et al., 2012; Li et al., 2013a; Drijfhout et al., 2015; Bathiany et al., 2016) (*medium confidence* due to the limited number of models).

9.3.1.2 Arctic Sea Ice volume and thickness

Arctic sea-ice volume has decreased in all months over the period 1979-2018. It is clear that sea-ice area has decreased substantially over this period (section 9.3.1.1), and *extremely likely* that the remaining sea-ice has become thinner (Lindsay and Schweiger, 2015; Kwok, 2018; Schweiger et al., 2019), leading to a clear signal of Arctic sea-ice volume loss. There is, however, only *low confidence* in quantitative estimates of the sea-ice volume loss over this period because of a lack of reliable, long-term, pan-Arctic observations and substantial spread in available reanalyses (Chevallier et al., 2017). Current best estimates from reanalysis give an average loss of annual mean sea-ice volume of between 2800 and 3700 km³ per decade over the period 1979 to 2010 (Schweiger et al., 2019).

For the more recent past, direct ice-thickness estimates from satellites exist. Based on these direct observations, there is *medium confidence* that Arctic sea-ice volume has decreased since 2003 (Kwok and Cunningham, 2015; Kwok, 2018). There is, however, only *low confidence* in the amount of decrease over this period and over the CryoSat-2 period from 2011 onwards because of snow-induced uncertainties in the retrieval algorithms, the shortness of the record, and the small identified trend (Kwok and Cunningham, 2015; Ricker et al., 2015; Tilling et al., 2015; Lee et al., 2016; Wang et al., 2016c; Nicolaus et al., 2017; Bunzel et al., 2018; Petty et al., 2018). (See Atlas.5.10.1.3 for a detailed description of the role of snow for satellite retrievals and sea-ice thermodynamics).

Observations of regional changes in sea-ice thickness vary in quality. Analysis of submarine data shows that sea ice has become thinner in the central Arctic Ocean over the past 60 years (*high confidence*). Over this period, agreement between satellites, submarines, and aircraft measurements leads to *medium confidence* that central Arctic average thinning is around 2 m or about two thirds of the initial ice thickness (Lindsay and Schweiger, 2015; Kwok, 2018). On smaller regions, data is too sparse to allow for quantitative estimates of long-term trends (King et al., 2017; Rösel et al., 2018), but a clear thinning signal over 10 to 20 years has been found for sea ice in Fram Strait (Hansen et al., 2013; Renner et al., 2014), north of Canada (Haas et al., 2017) and for landfast ice in Kongsfjorden/Svalbard (Pavlova et al., 2019). CMIP5 models and reanalyses fail to capture the observed distribution (Stroeve et al., 2014; Shu et al., 2015) and evolution (Chevallier et

al., 2017) of Arctic sea-ice thickness, which leads to *low confidence* in estimates of thickness from reanalyses and CMIP5 models, and these models' projections of sea-ice volume.

9.3.2 Antarctic Sea Ice

The multi-annual evolution of Antarctic sea ice is regionally more diverse than that of Arctic sea ice. This section expands and explains the Antarctic sea-ice related summary findings from chapters 2 and 3 by providing a more detailed assessment including regional aspects, seasonal aspects and process understanding. The findings differ from previous assessments in particular because of the observed, substantial decline in Antarctic sea-ice area in all months since 2016.

9.3.2.1 Antarctic sea-ice coverage

There is no significant trend in annual mean Antarctic sea-ice area over the period of reliable satellite retrievals from 1979 to 2018 (Maksym, 2019), which includes the record-high satellite-retrieved sea-ice coverage in 2015 and the substantial decline in sea-ice area afterward (*high confidence*) (Turner et al., 2017b; Schlosser et al., 2018) (Figure 9.17, Figure 2.18). Additionally, in contrast to AR5, recent research suggests only *low confidence* into the significance of the positive trend in mean Antarctic sea-ice area over the period 1979 and 2015 (Figure 2.18, Section 2.3.2.1.2) (Parkinson and Cavalieri, 2012; Simmonds, 2015; Comiso et al., 2017a), because the year-to-year fluctuations are large in all months (Figure 9.17) and the processes and reliability of year-to-year correlation of Antarctic sea-ice area is not well understood (Yuan et al., 2017).

The large-scale variability of Antarctic sea-ice coverage derives from regional details of ice formation, transport and decay across the Southern Ocean. Most of the ice is formed in polynyas near the continental margin (Haumann et al., 2016), and the maximum northward extent is largely governed by a balance between the northward advection of the ice and the lateral and subsurface melting from the warmer ocean (Martinson, 2012; Su, 2017; Maksym, 2019). Northward advection, mostly in the Ross and Weddell Sea, occurs under prevailing southerly winds and northward ocean currents (Haumann et al., 2016; Holland and Kimura, 2016). Fluctuations in ocean currents and surface winds can cause substantial regional and temporal fluctuations of Antarctic sea-ice area (Figure 9.17) (Stammerjohn and Maksym, 2016) although bathymetry and coastal topography limit the fluctuations (Nghiem et al., 2016). The evolution of mean Antarctic sea-ice area is the result of opposing regional trends, with slightly decreasing sea-ice cover during the period 1979 to 2018 in the Amundsen Sea and the Bellingshausen Sea, particularly during summer, and slightly increasing sea-ice cover in the eastern parts of the Weddell Sea and the Ross Sea (*medium confidence*; see Figure 9.17) (Stammerjohn et al., 2012; Holland, 2014; Yuan et al., 2017; SROCC). With the exception of the Ross Sea, these trends are not significant considering the large variability of the time series (Yuan et al., 2017).

[START FIGURE 9.17 HERE]

Figure 9.17: Left: Absolute anomaly of monthly-mean Antarctic sea-ice area during the period 1979 to 2018 relative to the average monthly-mean Antarctic sea-ice area during the period 1979 to 2008. Based on the average sea-ice area derived from OSISAF/CCI (OSI-450 for 1979-2015, OSI-430b for 2016-2018)(Lavergne et al., 2019), NASA Team (version 1, 1979-2018)(Cavalieri et al., 1996) and Bootstrap (version 3, 1979-2018)(Comiso, 2017) sea-ice concentration estimates. Right: Sea-ice coverage in the Antarctic as given by the average of the three most widely used satellite-based estimates for September and February, which usually are the months of maximum and minimum sea-ice coverage, respectively. First column: Mean sea-ice coverage during the decade 1979-1988. Second column: Mean sea-ice coverage during the decade 2009-2018. Third column: Absolute change in sea-ice concentration between these two decades, with stippling indicating significant differences. Fourth column: number of available CMIP6 models that simulate a mean sea-ice concentration above 15 % for the decade 2045-2055. The observational records are based on the average sea-ice coverage of OSISAF/CCI (OSI-450 for 1979-2015, OSI-430b for 2016-2018)(Lavergne et al., 2019), NASA Team (version 1, 1979-2018)(Cavalieri et al., 1996) and Bootstrap

(version 3, 1979-2018)(Comiso, 2017).

[END FIGURE 9.17 HERE]

Regional variations of Antarctic sea-ice area throughout the satellite record from 1979 onwards is potentially driven by many causes. Quantifying their different contribution is difficult among others because of temporally and spatially remote linkages due to sea-ice drift (Pope et al., 2017), and the tight coupling of Antarctic sea-ice coverage to drift patterns linked to regional surface winds and ocean currents. Strong correlations exist between sea-ice drift variability and meridional and zonal winds (Holland and Kwok, 2012; Kwok et al., 2017), and between sea-ice drift and cyclone activity (Schemm, 2018). In particular, a strengthening of the Amundsen Sea low (Raphael et al., 2016; Turner et al., 2016) can be linked to increasing sea-ice coverage in the Amundsen Sea (Holland and Kwok, 2012; Haumann et al., 2014, 2016; Kwok et al., 2016; Holland et al., 2017b, 2018) (*high confidence*). Multi-decadal variations in the tropical Pacific strengthen the Amundsen Sea low (Kwok et al., 2016; Meehl et al., 2016; Purich et al., 2016b) and influence other climate modes affecting Southern Ocean conditions, such as cyclone activity, which impact regional sea-ice coverage (Simpkins et al., 2012; Cerrone et al., 2017; Schemm, 2018) (*high confidence*).

There is *low confidence* that regional sea-ice trends in the austral fall are linked to changes in westerly winds and the Southern Annual Mode (SAM) in summer and spring (Holland, 2014; Doddridge and Marshall, 2017; Holland et al., 2017a), because of differing regional patterns (Simpkins et al., 2013). Increased near-surface ocean stratification contributes to the increase in sea-ice coverage (Haid et al., 2017; Lecomte et al., 2017; Su, 2017; Purich et al., 2018; Zhang et al., 2019a) (*medium confidence*). The stratification changes result partly from surface freshening (De Lavergne et al., 2014) (*high confidence*) from increased northward sea-ice advection (Haumann et al., 2016, submitted, submitted) and melting of the Antarctic ice sheet (Bintanja et al., 2013; Pauling et al., 2017; Merino et al., 2018; Haumann et al., submitted, submitted; Mackie et al., submitted, submitted) (*medium confidence*) amplified by local ice-ocean feedbacks (Goosse and Zunz, 2014; Lecomte et al., 2017; Goosse et al., 2018). In contrast, in the Amundsen Sea, very strong ice-shelf melting can cause sea-ice melt by entraining warm Circumpolar Deep Water (*medium confidence*) (Jourdain et al., 2017; Merino et al., 2018). Disagreement between models and unresolved processes (Section 9.2.3.2) lead to *low confidence* on the spatial distribution and relative contributions of the meltwater fluxes and ice-ocean feedbacks. Modifications of sea ice along the ice edge have been suggested to occur due to waves that propagate into the sea-ice region and break the ice (Kohout et al., 2014; Bennetts et al., 2017; Roach et al., 2018b), but there is *low confidence* in the magnitude of this process. There is *low confidence* that the increase in ice coverage results from a long-term Southern Ocean surface cooling trend (Fan et al., 2014; Comiso et al., 2017a; De Santis et al., 2017) because of potential intricate feedbacks between sea-ice change and surface cooling (Haumann et al., submitted).

There is *low confidence* in which physical drivers caused the decline in Antarctic sea ice since 2016. Change in atmospheric circulation (Latif et al., 2017; Stuecker et al., 2017; Turner et al., 2017b; Schlosser et al., 2018; Meehl et al., 2019; Wang et al., 2019a) and in subsurface ocean heat flux (Stuecker et al., 2017; Meehl et al., 2019) contribute. Since 2015, the large-scale trends in sea-ice drift speed have weakened or even reversed (Kwok et al., 2017) (*medium confidence*).

Before satellites, observational scarcity results in *low confidence* in large-scale decadal fluctuations in sea-ice coverage caused by large-scale temperature and wind forcing. Sparse direct observations suggest a decrease in sea-ice coverage from the 1950s to the 1970s, followed by an increasing sea-ice coverage since then (Fan et al., 2014; Armour and Bitz, 2015). On even longer time scales, paleo-proxy data indicate that on multi-centennial time scales sea-ice coverage of the Antarctic Ocean follows large-scale temperature trends, and that during the LGM Antarctic sea ice extended to about the polar front latitude in most regions during winter, whereas the extent during summer is less well understood (Gersonde et al., 2005; Benz et al., 2016; Xiao et al., 2016). Very limited proxy coverage leads to *low confidence* paleo-reconstructions of Antarctic sea ice.

CMIP5 models simulate more internal variability than in the observational record (Zunz et al., 2013), and

1 this remains true for CMIP6 models (Roach et al., submitted, submitted). Many individual CMIP5 models
2 misrepresent the seasonal cycle in sea-ice coverage (Mahlstein et al., 2013; Turner et al., 2013; Holmes et
3 al., 2019), but the multi-model mean seasonal cycle agrees well with observations (Shu et al., 2015). Again,
4 both findings remain valid for CMIP6 models (Roach et al., submitted, submitted). CMIP5 models also do
5 not realistically simulate the evolution of Antarctic sea-ice volume (Shu et al., 2015) and consistently
6 overestimate the amount of low concentration sea ice and underestimate the amount of high concentration
7 sea ice (Roach et al., 2018a). CMIP6 models, in contrast, simulate a more realistic distribution of regional
8 sea-ice coverage than CMIP5 models (Roach et al., submitted, submitted). Most CMIP5 models poorly
9 represent Antarctic sea-ice drift (Uotila et al., 2014; Lecomte et al., 2016; Schroeter et al., 2017, 2018;
10 Holmes et al., 2019) affecting simulated historical trends, with models that simulate a strong sea-ice motion
11 showing more variability in sea-ice coverage than models with weaker sea-ice motion (Schroeter et al.,
12 2018). There is hence *low confidence* regarding CMIP5 models' capability to realistically simulate the long-
13 term evolution of Antarctic sea ice, even though discrepancy between observed and modelled trends is often
14 within the range of simulated internal variability (Mahlstein et al., 2013; Turner et al., 2013; Zunz et al.,
15 2013; Olonscheck and Notz, 2017) and that a state-of-the-art representation of Southern Ocean processes, at
16 least in principle, allows for a realistic representation (Holmes et al., 2019). A lack of process understanding
17 of the still substantial spread in CMIP6 model simulations limits the robustness also of their future
18 projections to *low confidence* (Roach et al., submitted, submitted).

19
20 The discrepancy between the CMIP5 modelled and the observed evolution and regional patterns of Antarctic
21 sea ice has been related to a number of causes: a mis-representation of the detailed time-evolution of the sea-
22 ice response to a changing SAM (Ferreira et al., 2015; Holland et al., 2017a), an underestimation of changes
23 in the zonal asymmetries of atmospheric circulation (Haumann et al., 2014; Lecomte et al., 2016; Schroeter
24 et al., 2017) and an underestimation of seasonal changes in westerly winds related to stratospheric ozone
25 depletion and greenhouse gas increase (Jones et al., 2016b; Purich et al., 2016a) associated with increased
26 northward Ekman transport of the surface waters and upwelling (Gent et al., 2006; Bitz and Polvani, 2012).
27 Also the low resolution of CMIP5 models and their lack of ability in representing eddies might cause
28 shortcomings in the Southern Ocean, as temperature and other biases in the Southern Ocean have been found
29 to be reduced in a high-resolution model setup (Sidorenko et al., 2019). There is *low confidence* regarding
30 the relative importance of these possible shortcomings.

31
32 Most of the positive Antarctic trend over the period 1979 to 2015 has been attributed to internal variability
33 based on an analysis of the observational record (Meier et al., 2013; Gallaher et al., 2014; Gagné et al.,
34 2015b), reconstructions from early observations (Fan et al., 2014; Edinburgh and Day, 2016) and proxy data
35 (Hobbs et al., 2016), and model simulations (Turner et al., 2013; Zunz et al., 2013). However, CMIP5
36 models fail to simulate the observed trends, and the Southern Ocean response to external forcing takes longer
37 than the length of the observational record (Goosse and Renssen, 2001; Armour et al., 2016). Proxy data
38 from ice cores consistently indicate that the regional increase in the Ross Sea and the decrease in the
39 Bellingshausen Sea are part of longer centennial trends and exceed internal variability on multi-decadal time-
40 scales (Sinclair et al., 2014; Porter et al., 2016; Thomas and Abram, 2016) (*low confidence*). These
41 centennial trends are consistent with simulations from CMIP5 models (Hobbs et al., 2016; Jones et al.,
42 2016b; Kimura et al., 2017). Nonetheless, without accurate simulations of observed changes possible
43 contribution of anthropogenic forcing to the regional changes remains unclear (Hosking et al., 2013; Turner
44 et al., 2013; Haumann et al., 2014) and there is *low confidence* in the attribution to internal variability
45 (Section 3.4.1)

46
47 Because of our lack of detailed understanding of the governing processes, there is as in AR5 only *low*
48 *confidence* of the CMIP5 model finding of a future long-term decline of Antarctic sea ice in response to
49 ongoing global warming, and of a mitigation of the sea-ice loss by stratospheric ozone recovery during the
50 first half of the century (Smith et al., 2012) and by an increased freshwater input from melting of the
51 Antarctic ice sheet (Bronselaer et al., 2018).

9.3.2.2 Antarctic sea-ice thickness

There are only sparse observational records of Antarctic sea-ice thickness. Data from ICESat-1 laser altimetry (Kurtz and Markus, 2012), from Operation IceBridge (Kwok and Kacimi, 2018), and long-term ship-board observations collected in the ASPeCt data set (Worby et al., 2008) suggest that sea ice thicker than 1 m prevails in regions of multi-year ice along the eastern coast of the Antarctic Peninsula in the Weddell Sea, in the high-latitude embayment of the Weddell Sea, and along the coast of the Amundsen Sea, with remaining regions dominated by thinner first year sea ice (*high confidence*). Regional patterns in ice thickness are affected by large snow deposition and resulting snow-ice formation (Massom et al., 2001; Maksym and Markus, 2008), and deformation, ridging, and rafting that regionally cause formation of very thick sea ice (Williams et al., 2015).

There is *low confidence* in the long-term evolution of Antarctic sea-ice thickness. Both ASPeCt and ICESat-1 measurements are biased low in regions with thick ice (Kern and Spreen, 2015) consistent with results from reanalyses (Massonnet et al., 2013; Haumann et al., 2016) and observations with autonomous vehicles under sea ice (Williams et al., 2015). Estimates of sea-ice thickness from Cryosat-2 do not substantially reduce uncertainty, primarily because of the unknown snow thickness and radar scattering above the snow-ice interface (Kern and Spreen, 2015; Kern et al., 2016; Bunzel et al., 2018; Kwok and Kacimi, 2018). Reanalyses suggest increasing sea-ice thickness and volume overall between 1980 and 2010 (Holland et al., 2014; Zhang, 2014; Massonnet et al., 2015), with thickening in the high-latitude embayment of the Weddell Sea and in the open ocean Ross Sea, and thinning in the Bellingshausen Sea and along the coast lines of the Antarctic Peninsula. This is in contrast to CMIP5 model simulations that simulate a decrease in Antarctic sea-ice volume over the historical period (Shu et al., 2015; Schroeter et al., 2018), so there is *low confidence* in CMIP5 simulated future Antarctic sea-ice thickness.

9.4 Ice sheets

Observations and modelling indicate that both the Greenland and Antarctic ice sheets are currently losing mass at accelerating rates (Figure 9.18), due to atmospheric, oceanic, and dynamic processes, and are contributing to sea level rise (Bamber et al., 2018a; The IMBIE Team, 2018; Holland et al., 2019; Mouginot et al., 2019; Rignot et al., 2019; The IMBIE Team, 2019). Based on multiple modelling studies of past and future climate scenarios there is *high confidence* that freshwater input to the oceans from ice sheet melting also triggers local (Figure 9.10) and global climate responses (Morrison et al., 2015; Ivanovic et al., 2017; Bronselaer et al., 2018; Park and Latif, 2018; Golledge et al., 2019) but the magnitude of responses may depend on processes incorporated in these models (Pauling et al., 2017; Merino et al., 2018; Schloesser et al., 2019b).

[START FIGURE 9.18 HERE]

Figure 9.18: (a) Timeseries of mass changes in Greenland for each of the major drainage basins shown in the inset figure (Colgan et al., 2019; Mouginot et al., 2019) for the periods 1972 – 2018 and 1992-2018. (b) Timeseries of mass changes of three portions of Antarctica (Bamber et al., 2018a; The IMBIE Team, 2018) for the period 1992 – 2018.

[END FIGURE 9.18 HERE]

Ice sheet models (including both standalone ice sheet models and ice sheet models coupled within climate models) represent many of the environmental and ice-dynamic processes known to influence ice sheet behaviour (Goelzer et al., 2018; Pattyn, 2017; Pattyn et al., 2017; Turner et al., 2017) but limitations or simplifications in terms of physics, numerical schemes, or spatial resolution, still persist (Seroussi et al., 2019). For this reason, assessments in this report of the future contribution to sea level from the Greenland and Antarctic ice sheets (Sections 9.4.1.2, 9.4.2.2) are based primarily on the results of process-based model

intercomparison exercises (Nowicki et al., 2016; Goelzer et al., 2018a; Seroussi et al., 2019; Goelzer et al., submitted; Levermann et al., submitted, a; Nowicki et al., submitted, a, submitted, b; Seroussi et al., submitted) but are further contextualised with results from Structured Expert Judgement exercises (Bamber and Aspinall, 2013; Bamber et al., 2019).

9.4.1 Greenland

9.4.1.1 Recent observations and model evaluation

The AR5 and SROCC stated with *very high confidence* that the Greenland ice sheet (GrIS) lost mass and contributed to sea level rise over the past 25 years (Figure 9.18a). In AR5 the mass loss during 2002–2011 ($215 [157 \text{ to } 275] \text{ Gt yr}^{-1}$) was shown to have increased 6-fold compared to the preceding decade. In Chapter 2 (Section 2.3.2.4) an account of the total Greenland mass loss is given in a global context, while in this section we present regional time series and assess the different processes causing this increase in mass loss. Since AR5, spatial and temporal estimates of GrIS mass change have been improved through continued airborne and satellite radar and laser altimetry, satellite-based gravimetry, use of historical aerial photographs (1931–1984), mass discharge calculations, modelling and combination of methods, as well as improvements in data processing, model corrections (e.g. elastic land uplift from GNET GNSS observations), and the synthesis of data from different measurement methods (Shepherd et al., 2012; Box and Colgan, 2013; Velicogna et al., 2014; Khan et al., 2015; Kjeldsen et al., 2015; McMillan et al., 2016; van den Broeke et al., 2017; Forsberg et al., 2017; Bamber et al., 2018b; Sandberg Sørensen et al., 2018; King et al., 2018; The IMBIE Team, 2019; Groh et al., 2019; Mankoff et al., 2019; Mouginot et al., 2019). There is *very high confidence* in the mass loss record after 1992, with a reconciled estimate based on 26 satellite estimates complemented with 10 models of surface mass balance and 6 models of glacial isostatic adjustment (The IMBIE Team, 2019). The ice sheet was close to a state of balance in the 1990s, then increased its annual mass loss, with *high confidence* that the year to year variability in total mass loss is large. The ice sheet experienced high melt rates in 2010 and exceptionally high melt rates in 2012 when 97% of the total surface area had signs of melt (Hanna et al., 2012, 2014; Nghiem et al., 2012; Tedesco et al., 2013; Bamber et al., 2018b). According to the reconciled estimate from satellite data, the peak mass losses were in [PLACEHOLDER FOR VALUES: 2011 at $335 \pm 62 \text{ Gt}$ and 2012 at $324 \pm 70 \text{ Gt}$ (The IMBIE Team, 2019)]. The record summer melt of 2012 led to a mass loss of $627 \pm 89 \text{ Gt}$ during June–August (Tedesco et al., 2013), but relatively high accumulation rates and refreezing of much of the melt on the ice sheet prevented this from becoming a record mass loss year (van Angelen et al., 2014). The 2012 event resulted in melt-generated ice layers in the firn that have made the use of radar altimetry for height and mass change estimates more complicated (Nilsson et al., 2015). Mass losses between 2013 and 2018 were lower than the preceding few years because of cooler regional atmospheric and ocean temperatures (van den Broeke et al., 2017; Bevis et al., 2019; Khazendar et al., 2019a; Mouginot et al., 2019; The IMBIE Team, 2019). During summer 2019 the ice sheet experienced another exceptional melt season due to persistence of anticyclonic conditions as well as an enhanced melt-albedo feedback, bringing the amount of melt to pre-2013 values (Tedesco and Fettweis, 2019). As presented in Chapter 2, the Greenland Ice Sheet lost $3800 \pm 339 \text{ Gt}$ of ice between 1992 and 2018, causing sea level to rise by $10.6 \pm 0.9 \text{ mm}$ (The IMBIE Team, 2019). Greenland ice sheet was *likely* close to mass balance ($-18 \pm 28 \text{ Gt yr}^{-1}$) between 1992 and 1997, but a peak in mass loss occurred between 2007 and 2012 ($-270 \pm 28 \text{ Gt yr}^{-1}$). The mass loss rates are shown in Figures 2.23, 9.18 and 9.19 and Cross-Chapter Box 9.2, Figure 1. These estimates agree within uncertainties with other post-AR5 reviews (Bamber et al., 2018b; Cazenave et al., 2018; Mouginot et al., 2019) and show that the largest mass losses are along the west coast and in southeast Greenland. This is additionally supported by independent GNSS observations from the Greenland GPS network which show elastic bedrock uplift of tens of centimeters between 2007–2019 as a result of ongoing ice mass loss (Bevis et al., 2019).

Only a few estimates of mass change during the 20th century are presently available. A few studies use aerial stereo photogrammetric imagery to map trimlines, as well as lateral and end moraines marking the maximum extent of the Greenland ice sheet during the Little Ice Age, and calculate frontal retreat and ice mass losses from 1900 to 2012 (Khan et al., 2014; Schenk et al., 2014; Kjeldsen et al., 2015). Average mass change during the period 1900–1983 is found to be $-75 \pm 29 \text{ Gt yr}^{-1}$, with an increase in mass losses after 2000 (Box

and Colgan, 2013; Khan et al., 2015; Kjeldsen et al., 2015). During the 20th century many outlet glaciers in Greenland underwent asynchronous retreat after the LIA (Lea et al., 2014b), loss of floating ice tongues resulting in increased calving rates in the early 20th century (Andresen et al., 2017), and small changes from 1912 until after 2000, with historically unprecedented retreats in the 21st century (Lüthi et al., 2016). Land-terminating glaciers showed more rapid retreat in the 1930s than the 2000s, whereas marine-terminating glaciers retreated more rapidly in the 2000s (Bjørk et al., 2012).

The increase in mass losses at the end of the 1990s is caused by a number of spatially and temporally variable factors. The mass losses can be partitioned between surface mass balance and glacier dynamics; the latter results in calving of icebergs and submarine melting where the ice is in contact with the ocean. For the period 1972-2018, the GrIS lost $66 \pm 8\%$ of its mass due to ice dynamics and $34 \pm 8\%$ by surface mass balance (Mouginot et al., 2019). This ratio changes to 48% glacier dynamics and 52% surface melt in the period 1992-2018 (The IMBIE Team, 2019), and a more detailed study shows that glacier dynamics accounted for 58% of mass lost in the period 2000-2005, decreasing to 36% in the period 2005-2009 and 32% between 2009 and 2012 (Enderlin et al., 2014). The surface mass balance is observed to vary on a shorter time scale than the glacier dynamics (McMillan et al., 2016). A study with near-complete spatial coverage (using 276 flux gates) reports approximately steady solid ice discharge from 1986 to 2000, then a sharp increase during 2000-2005, followed by approximately steady ice discharge of $500 \pm 50 \text{ Gt yr}^{-1}$ through 2017, with the highest discharge in southeast Greenland and large regional and glacier variability (Mankoff et al., 2019). A similar increase in discharge in the early 2000s and then leveling off is observed in other ice sheet wide records of discharge (Khan et al., 2015; King et al., 2018; Mouginot et al., 2019).

Surface melt depends on the energy balance at the surface (Bennartz et al., 2013; Van Tricht et al., 2016; Hofer et al., 2017). A number of positive feedback mechanisms will amplify the melting (Fyke et al., 2018). There is *very high confidence* that increased melt at the ice sheet margins decreases the surface elevation, which increases the local air temperature and further increases the melt. This is called the SMB-elevation feedback (van den Broeke et al., 2017). There is *very high confidence* that surface darkening (Tedesco et al., 2016; Tedstone et al., 2017), and the positive albedo feedback (Box et al., 2012; Fyke et al., 2018) that results from the melting of bright winter snow cover that reveals darker bare ice beneath, lower the albedo and lead to further melting (Ryan et al., 2019). Darkening due to hampered percolation and meltwater saturation in lower accumulation areas (Charalampidis et al., 2015), and from dark particulate matter from natural and anthropogenic sources that are deposited on the surface and decrease the albedo (Stibal et al., 2017; Ryan et al., 2018; Cook et al., 2019b), have enhanced the recent increasing trend in mass loss. Cloud cover has various effects on melt; decreasing cloud cover can increase melt (Hofer et al., 2017), low level liquid clouds increase melt (Bennartz et al., 2013), and clouds can reduce meltwater refreezing, thereby accelerating bare-ice exposure and enhance meltwater runoff (Van Tricht et al., 2016). Interannual variation in total column water was positively correlated with the number of melt days, cumulative melt extent, and modelled melt amount over Greenland in the period 2000-2014, indicating that the cloud longwave forcing dominates over the shortwave forcing with a net warming effect on the surface (Välisuo et al., 2018). The meltwater does not all run off, because meltwater can be retained and refreeze in the snowpack and firn (de la Peña et al., 2015; Charalampidis et al., 2016; Machguth et al., 2016a; Sommers et al., 2017; Heilig et al., 2018; MacFerrin et al., 2019; Vandecrux et al., 2019) and thereby buffer the contribution to sea level rise. Recently discovered liquid perennial firn aquifers (Koenig et al., 2014; Miège et al., 2016) store large amounts of meltwater. The estimated aquifer area is about 70,000 km² and the depth to the top of the water table is 5–50 m (Forster et al., 2014). This liquid aquifer in the firn is potentially important for the ice sheets mass and energy budget.

[START FIGURE 9.19 HERE]

Figure 9.19: Estimated sea level equivalent volume relative to present day (top left panel, detail in supplementary material) and spatial extent (bottom left) of the Greenland ice sheet during the mid-Pliocene Warm Period, the Last Interglaciation, and the Last Glacial Maximum (see Annex II for details). Paleo extents in bottom panels are interpreted from individual reconstructions (Lecavalier et al., 2014; Koenig et al., 2015; Goelzer et al., 2017). Grey shading shows extent of grounded ice. (top right) Cumulative mass loss (sea level equivalent) from the Greenland ice sheet since 1840 (updated from (Box and Colgan, 2013;

Kjeldsen et al., 2015) reconstruction), with satellite observations shown from 1972 (Mouginot et al., 2019) and 1993 (Bamber et al., 2018b; The IMBIE Team, 2019), and projections from ISMIP6 by 2100 under RCP8.5/SSP5-8.5 and RCP2.6/SSP1-2.6 scenarios (thin lines from (Goelzer et al., submitted; Nowicki et al., submitted, a) and likely range of the ISMIP6 emulation (shades and bold line (Edwards et al., submitted)). (bottom right) Maps of mean elevation changes 2010-2017 derived from CryoSat 2 radar altimetry (Bamber et al., 2018b) and ISMIP6 model mean (2061-2100) projected changes (Goelzer et al., submitted) .

[END FIGURE 9.19 HERE]

The outlet glaciers that are in contact with the ocean, the tidewater glaciers, are responsible for the mass loss through discharge, which increased by 18% in the period 1972-2018 (Mouginot et al., 2019). Although only 8% of the length of the ice sheet periphery has ice grounded below sea level, this small fraction controls 88% of the ice discharge from Greenland (Morlighem et al., 2014). The outlet glaciers have shown temporally variable and asynchronous behaviour (Khan et al., 2015; Moon et al., 2015; Bunce et al., 2018; Vijay et al., 2019). There is *high confidence* that submarine melt and iceberg calving contributed to the acceleration of mass loss from Greenland, particularly in the period 2000-2010 when accelerated retreat was observed in the southeast and northwest. The fastest glacier on Earth, Jakobshavn Isbræ, which drains ~6% of the GrIS, has retreated and thinned several times since 1900 (Csatho et al., 2008). Jakobshavn retreated and thinned between 2003 and 2016, and thickened and readvanced between 2016-2019, which was linked to cooling of ocean water in Disko Bay (Khazendar et al., 2019b). Despite thickening near the margin it is still thinning farther upstream, leading to a net basin-wide mass loss (Mouginot et al., 2019). Glacier speed has increased about 10% for Petermann, Humboldt and Nioghalvfjærdssjøfjorden in the north (Khan et al., 2014). The velocity of Zachariæ Isstrøm in northeast Greenland, containing 0.5 m sea level equivalent, increased after its ice shelf began to break up in 2002-2003: from late 2012 the velocity tripled, and overall the melting of its ice shelf and thinning of its grounded portion have doubled (Khan et al., 2014; Mouginot et al., 2015). North East Greenland Ice stream (NEGIS) initiated speed-up and dynamic mass loss between 2003 and 2006, following almost three decades of dormancy (Khan et al., 2014). Changes in front position, and their resulting persistent changes to the balance of forces at the glacier terminus, appear to dominate multi-year variability in regional and total discharge; however, seasonal variations are substantially controlled by the timing of meltwater input (King et al., 2018). There is *high confidence* that warmer ocean waters reaching the tidewater glaciers trigger retreat and calving processes (Xu et al., 2012; Straneo and Heimbach, 2013; Gladish et al., 2015; Wood et al., 2018). Surface melting can increase subglacial runoff and enhance ocean melt rates and increase plume-driven ice face melting (Xu et al., 2013; Slater et al., 2016), even though ocean warming may have a stronger effect on future melt rates than increased runoff (Morlighem et al., 2019). Migration of subtropical water masses around the coast of Greenland (Holland et al., 2008) and its occupation of coastal fjords (Straneo et al., 2010; Christoffersen et al., 2011) will affect retreat and calving rates (Benn et al., 2017), but there are gaps in understanding of the processes by which warmer Atlantic waters are driven toward the ice sheet (Straneo et al., 2013; Xu et al., 2013; Bendtsen et al., 2015). Recent studies show the spatial pattern of accelerating mass changes reflects the geography of NAO-driven shifts in atmospheric forcing and the ice sheet's sensitivity to that forcing. The decadal acceleration in mass loss in southwest Greenland arose due to the combination of sustained global warming and positive fluctuations in temperature and insolation driven by the NAO (van Angelen et al., 2014; Bevis et al., 2019).

Primary observational improvements for the Greenland Ice Sheet since AR5 include mapping of firn aquifers (Miège et al., 2016), understanding of the storage of meltwater in firn (Machguth et al., 2016a; MacFerrin et al., 2019), processing of historical datasets to derive ice discharge (King et al., 2018; Mankoff et al., 2019; Mouginot et al., 2019), the digital elevation model from the beginning of the 1980s made from aerial images (Schenk et al., 2014; Kjeldsen et al., 2015; Korsgaard et al., 2016) and the increasing availability of satellite observations (CryoSat-2, Landsat-8, Sentinel 1-3, RADARSAT-2) that are continuously acquiring data (McMillan et al., 2016; Mouginot et al., 2017; Gourmelen et al., 2018; Sandberg Sørensen et al., 2018). A new Greenland bed topography dataset has been generated by inferring ice thickness from mass conservation at a high level of spatial detail (150 m resolution) and precision (Morlighem et al., 2017). The higher resolution bedrock data reveal increased vulnerability of marine-based glaciers to Atlantic water, due to the presence of submarine ice-covered valleys that extend deep below sea level and far inland (Morlighem et al.,

2014). There are, however, still areas within the ice sheet that have poor coverage of ice thickness and bedrock observations.

Model evaluation

Different types of models are used to reconstruct or project the evolution of the Greenland Ice sheet; surface mass balance models for the atmosphere-ice interaction, ice dynamic models, and submarine melt and calving models for the ice-ocean interaction. These have not yet been coupled together to include all feedbacks. Large scale global models are discussed in Chapter 3 (Section 3.4.3.2), and the different process models are discussed here. Since AR5, regional climate models (RCMs) that are routinely used to simulate SMB of the GrIS have improved (van den Broeke et al., 2017). Increased availability and quality of remotely sensed (Koenig et al., 2016; Overly et al., 2016) and in-situ observations (Machguth et al., 2016b; Fausto et al., 2018; Vandecrux et al., 2019) of the ice sheet, required to evaluate and validate SMB models, have enabled improvements in their simulations. Reconstructions of SMB evolution extend further back in time with models now covering all or a large part of the 20th century (Hanna et al., 2011; Box and Colgan, 2013; Kjeldsen et al., 2015; Noël et al., 2016; Fettweis et al., 2017; Wilton et al., 2017) with a spatial resolution of up to 1 km (Noël et al., 2016; Wilton et al., 2017). The SMB reconstructions are, however, diverse and provide only *medium confidence* in whether recent (post-1990) surface mass losses are higher than in earlier periods of high mass loss (van den Broeke et al., 2017). Uncertainties in SMB projections remain relating to processes that are currently not included or poorly represented in RCMs, but that are expected to have an impact on future SMB changes and could give rise to important feedback mechanisms (van den Broeke et al., 2017). These include the presentation of clouds (Van Tricht et al., 2016; Hofer et al., 2019), blocking index over Greenland (Van Den Broeke et al., 2016; Delhasse et al., 2018; Hanna et al., 2018), turbulent heat exchange over rough ice surfaces (Smith et al., 2016), representation of blowing snow processes (Lenaerts et al., 2012), the biological darkening of the ice sheet surface (Musilova et al., 2016; Stibal et al., 2017; Ryan et al., 2018) and the retention and lateral and heterogeneous vertical transport of meltwater in the ice (Machguth et al., 2016a; Steger et al., 2017; MacFerrin et al., 2019). Computing SMB directly within global models is key to coupled ice-sheet-climate simulations, and considerable development in Earth system models (ESMs) has occurred since AR5. ESMs are starting to incorporate multi-layer snow models and full energy balance models (Punge et al., 2012; Cullather et al., 2014; van Kampenhout et al., 2017, submitted; Alexander et al., 2019) as well as the use of elevation classes to compensate for their coarser resolution (Lipscomb et al., 2013; Sellevold et al., 2019), resulting in SMB simulations that compare well with RCMs and observations (Alexander et al., 2019; van Kampenhout et al., submitted). Overall there is *medium confidence* in SMB projections for the GrIS.

Ice flow models used to project dynamic changes of the GrIS have improved since AR5 (Section 3.4.3.2), particularly in terms of model resolution (Aschwanden et al., 2016), data assimilation for initialisation of projections (Larour et al., 2014, 2016; Perego et al., 2014; Goldberg et al., 2015; Lee et al., 2015; Schlegel et al., 2015; Mosbeux et al., 2016), the use of increasingly large ensembles to study model sensitivity and parameter uncertainty (Saito et al., 2016; Aschwanden et al., 2019), and simulations that combine a long-term paleoclimate spinup with incremental grid refinement and nudging to derive a fully thermally and dynamically-evolved ice sheet that has a geometry almost identical to present day (Golledge et al., 2019). This improvement is closely linked to the growing number and quality of satellite and airborne observations of the ice sheet (Howat et al., 2014; MacGregor et al., 2015; Watkins et al., 2015) and new techniques used to generate internally consistent input data sets (Morlighem et al., 2014, 2016b). In particular, the improved bedrock topography map of Greenland (Morlighem et al., 2017) has provided greater detail for models to capture the complex flow pattern of outlet glaciers (Aschwanden et al., 2016). There is, however, still insufficiently accurate observations of subglacial topography in southeast Greenland, which results in poor agreement between modelled and observed surface velocity in this area (Aschwanden et al., 2016). Models that assimilate contemporary data for their initialisation poorly represent the long-term effects of ice thermodynamics and bedrock response to ice loading changes, while models that can include long-term temperature evolution have lower resolution and do not resolve details at the ice sheet margin (Fürst et al., 2015; Saito et al., 2016), which can lead to large mismatches with present day observed geometry and velocity of the ice sheet (Goelzer et al., 2018a). There is *medium confidence* in ice flow models projecting the future of the GrIS. Nonetheless, as for SROCC, uncertainties in future SMB are assessed to be greater

than uncertainties in ice dynamics arising from bathymetric boundary conditions (*medium confidence*).

The physical process of calving (Benn et al., 2017) and ice-ocean interactions remain poorly understood, so parameterizations are suggested for marine-terminating glaciers for calving (Mercenier et al., 2018) and submarine and plume-driven melt (Xu et al., 2012; Fried et al., 2015; Nienow et al., 2017; Cowton et al., 2018; Slater et al., 2019a, 2019b). Owing to the difficulties of modelling the large number of marine-terminating glaciers, the majority of recent modelling work on Greenland is focused on individual or a limited number of glaciers (Beckmann et al., 2018; Bondzio et al., 2016, 2017; Krug et al., 2014; Morlighem et al., 2016a; Muresan et al., 2016), or a specific region (Morlighem et al., 2019). A study of 12 marine-terminating outlet glaciers with flowline models (Beckmann et al., 2018) concludes that the AR5 upscaling of contributions from 4 of the largest glaciers (Nick et al., 2013) probably overestimated the total glacier contribution from the ice sheet, due to differences in response between large and small glaciers. Morlighem et al. (2019) show similar results for two of the glaciers studied by Beckmann et al. (2018), and emphasise the importance of bedrock features as deep valleys can drive rapid retreat (Morlighem et al., 2017). In summary, we have *medium confidence* in quantitative, scenario-dependent projections for the GrIS outlet glacier dynamic contribution to sea level.

Improved SMB simulations (van den Broeke et al., 2017) and coupled RCM-ice sheet (e.g., Le clec'h et al., 2019) and ESM-ice sheet (e.g., Muntjewerf et al., submitted, b)) models have the potential to reduce uncertainty in projections of near-term GrIS change and the reversibility of GrIS mass losses. However, fully coupled models that include all important feedback mechanisms (Fyke et al., 2018) are computationally expensive (Ridley et al., 2010) and still require extensive development (Fyke et al., 2014; Vizcaino, 2014; Vizcaino et al., 2015; Helsen et al., 2017).

9.4.1.2 Future projections

The SROCC assessed the projected GrIS contribution to GMSL as unchanged from AR5 (e.g. *likely* ranges 4-10 cm under RCP2.6; 7–21 cm under RCP8.5 from 1986-2005 to 2081-2100) (Church et al., 2013; SROCC), based on the good agreement with more recent modelling results with multiple climate and ice sheet models (Edwards et al., 2014; Fürst et al., 2015; Calov et al., 2018; Rückamp et al., 2018; Aschwanden et al., 2019; Golledge et al., 2019), and taking into account a recent expert elicitation study (Bamber et al., 2019). Figure 9.19 and Table 9.1 show the projected GrIS contribution to GMSL by 2100 under selected scenarios. New projections since SROCC come primarily as part of the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6, with protocols described in (Nowicki et al., 2016)). The contribution from 14 ice sheet modelling groups (Figure 9.19) forced with one global climate model for RCP2.6 and six for RCP8.5 (Goelzer et al., submitted), one climate model for SSP1-2.6 and four for SSP5-8.5 (Nowicki et al., submitted, a), result in sea level contributions by 2100 of 0.031 ± 0.016 m, 0.089 ± 0.051 m, 0.057 ± 0.045 m and 0.153 ± 0.038 m respectively (Table 9.1). These projections are relative to 2015 and do not contain the additional mass change due to climatic changes prior to 2015 (called committed mass change), due to the uncertainty in initializing ice sheet models (Goelzer et al., 2018a) and in climatic forcing required to reproduce observed changes over the GrIS. The CMIP5 climate models selected by ISMIP6 represent a range of possible future climates (Barthel et al., 2019), while the CMIP6 climate models were chosen based on their availability (Nowicki et al., submitted, b). Statistical emulation of the ISMIP6 ensemble from (Edwards et al., submitted) allows representation of a larger range of uncertainties in climate models and scenarios, yielding median [5, 95 percentiles] sea level contributions at 2100 of: 0.02 [-0.06, 0.13] m for SSP1-1.9, 0.03 [-0.05, 0.13] m for SSP1-2.6, 0.05 [-0.03, 0.15] m for SSP2-4.5; 0.08 [0, 0.18] m for SSP3-7.0; and 0.09 [0.01, 0.20] for SSP5-8.5. The coupled atmosphere-ice sheet ESM CESM2-CISM2 projects 0.104 m at year 2100 relative to 2015 under SSP5-8.5 (Muntjewerf et al., submitted, a) and a similar contribution under the idealized 1% increase in CO₂ scenario (Muntjewerf et al., submitted, b). [PLACEHOLDER FOR ADDITIONAL ASSESSMENT OF THE GREENLAND CONTRIBUTION AT 2100 WILL BE INCLUDED ONCE NEW STUDIES ARE AVAILABLE THAT EXPLORE CLIMATE AND ICE SHEET MODEL UNCERTAINTIES MORE COMPREHENSIVELY].

[START TABLE 9.1 HERE]

Table 9.1: Projected sea level contributions in meters from the Greenland ice sheet by 2100 for selected CMIP5 RCP and CMIP6 SSP scenarios. SMB is surface mass balance, DYN the dynamical contributions. *denotes median values and likely ranges for projection reported in IPCC SROCC (Oppenheimer et al., 2019) Table 4.4 for 2081-2100 relative to 1986-2005, ** denotes contribution in 2100 relative to 2000 reported in SROCC Table 4.2, unless noted. *** denotes contribution in 2100 relative to 2015 and excludes committed ice mass change

Study	RCP2.6	RCP4.5	RCP8.5	Notes
*SMB only IPCC AR5	0.03 (0.01 to 0.07)	0.04 (0.02 to 0.09)	0.07 (0.03 to 0.17)	Median (<i>likely</i> range)
*DYN only IPCC AR5	0.04 (0.01 to 0.06)	0.04 (0.01 to 0.06)	0.05 (0.02 to 0.07)	Median (<i>likely</i> range)
**IPCC AR5 (Church et al., 2013a)	0.04 to 0.10	0.04 to 0.13	0.07 to 0.21	<i>Likely</i> range
**(Fürst et al., 2015)	0.014 to 0.078	0.028 to 0.093	0.051 to 0.016	Range from 10 GCMs
**(Vizcaino et al., 2015)	0.037	---	0.068	
**(Aschwanden et al., 2019)	0.05 to 0.019	0.08 to 0.23	0.14 to 0.33	16-84% range relative to 2008
**(Calov et al., 2018)	---	0.019 to 0.056	0.046 to 0.13	Range from 3 GCMs
**(Golledge et al., 2019)	---	0.109	0.112	
**IPCC SROCC (Oppenheimer et al., 2019) Table 4.2	0.063	0.078	0.119	Process based median of post AR5 studies
*** (Goelzer et al., submitted)	0.031 ± 0.016	---	0.089 ± 0.051	ISMIP6 multi-model ensemble from 1 ESM for RCP2.6 and 6 ESMs for RCP8.5
<i>Any other study</i>				
PLACEHOLDER				
Study	SSP1-26	SSP2-45	SSP5-85	
*** (Nowicki et al., submitted, b)	0.057 ± 0.045	---	0.153 ± 0.038	ISMIP6 multi-model ensemble from 1 ESM for SSP1-26 and 4 ESMs for SSP5-85
*** (Edwards et al., submitted)	0.03 (-0.05 to 0.13)	0.05 (-0.03 to 0.15)	0.09 (0.01 to 0.20)	Emulator of ISMIP6, median (5-95% range)
(Muntjewerf et al., submitted, a)	---	---	0.104	Coupled CESM2-CISM2 relative to 2015
<i>Any other study</i>				
PLACEHOLDER				

[END TABLE 9.1 HERE]

The projected mass loss of GrIS is predominantly due to increased melt resulting in decreasing SMB (*high confidence*), concurrent with rising temperatures (Fettweis et al., 2013; Vizcaino et al., 2015; Le clec'h et al., 2019; Muntjewerf et al., submitted, a, submitted, b). There is *high confidence* that mass changes due to SMB and outlet glacier dynamics are not independent (Goelzer et al., 2013; Fürst et al., 2015; Rückamp et al., submitted). There is *high confidence* that the mass loss through ice discharge will reduce in future as marginal thinning and retreat become self-limiting (Fürst et al., 2015; Aschwanden et al., 2019; Golledge et al., 2019; Goelzer et al., submitted; Muntjewerf et al., submitted, a), because an increase in discharge leads

in most areas to a retreat of the glacier margin onto land above sea level, isolating the ice sheet from marine influence. Representation of ice-ocean interactions does however remain challenging in continental scale ice sheet models, due to the limited understanding and computational cost. In ISMIP6, uncertainty in ice-ocean interaction was explored by the use of “open” and “standard” experiments, with the latter taking the form of three distinct timelines of tidewater glacier retreat per climate model to investigate ice sheet sensitivity to ocean variability (Slater et al., 2019a, 2019b; Nowicki et al., submitted, b). In the open approach, models were free to use their own parameterization of ice-ocean interactions, including calving. The open approach models generally show smaller mass loss than the standard approach (Goelzer et al., submitted). At the end of the 21st century under RCP8.5, ISMIP6 ice sheet model uncertainty explains 40% of the ensemble spread, while uncertainty due to climate model and ocean forcing account for 40% and 20%, respectively (Goelzer et al., submitted). The emulated projections use double the ISMIP6 range of ocean forcings, so this uncertainty is a larger contribution: 37% of the 5-95% interval under SSP5-8.5 (Edwards et al., submitted). (Aschwanden et al., 2019) ascribe 25-51% of the uncertainty in 21st century projections (depending on the RCP) to uncertainty in the parameterization of basal sliding, and this reduces to 4-34% in the 22nd century and 1-29% in 2300. (Fürst et al., 2015) estimate the partitioning between basal/surface mass balance and dynamic discharge will change from approximately 50/50 in 2010 to 86-97%/3-14% by 2100. (Golledge et al., 2019), and the coupled ESM-ice sheet simulations of (Muntjewerf et al., submitted, a) also find that mass loss in the latter half of the 21st century is primarily due to declining net surface mass balance. In agreement with SROCC, we have *high confidence* that the future evolution of the GrIS will be controlled more by changes in its SMB, and that the relative importance of SMB will increase in time over that of ocean-forced dynamical changes.

The SMB-elevation feedback (Section 9.4.1.1) increases projected sea level contributions when included through coupling or parameterisation in standalone ice sheet models (Edwards et al., 2014). The feedback is parameterized in many standalone ice sheet modelling studies (Edwards et al., 2014; Fürst et al., 2015; Calov et al., 2018; Rückamp et al., 2018), but may increasingly diverge from coupled model results over centennial timescales (Vizcaino et al., 2015; Le clec’h et al., 2019). Estimates of the proportional increase depend on the climate model from which it is derived (i.e. scenario and timescale), and has been estimated to be around 5% for scenario RCP8.5 at 2100 (Fettweis et al., 2013; Edwards et al., 2014; Vizcaino et al., 2015; Le clec’h et al., 2019; Goelzer et al., submitted), with *medium confidence* in projections of this contribution from standalone ice sheet models beyond 2100.

Basal lubrication of glaciers by surface meltwater drainage is no longer expected to accelerate future Greenland mass losses (*high confidence*), due both to a more nuanced understanding of the influence of basal meltwater (Nienow et al., 2017; Flowers, 2018), as well as modelling studies which have shown that this lubrication effect is of secondary importance (Aschwanden et al., 2019; Fürst et al., 2015; Golledge et al., 2019; Shannon et al., 2013). Basal lubrication is expected to add less than 5% to the sea level contribution (up to 0.9 cm by 2200) in a multi-model ensemble (Shannon et al., 2013) and with negligible effect (< 1%) in more recent individual model studies (Fürst et al., 2015; Aschwanden et al., 2019; Golledge et al., 2019). Given the small expected contribution to GrIS mass loss, lubrication effects were not included in the ISMIP6 projections.

Studies of the sea level rise commitment from Greenland (Levermann et al., 2013; Clark et al., 2016; Mengel et al., 2016) have largely relied upon one model (Robinson et al., 2012) that found highly non-linear response to temperature. The threshold for an irreversible decline of the GrIS in that study was found to occur between 1.1–2.3 °C of regional summer warming above pre-industrial levels (95% interval) with a best estimate of 1.8 °C (Pattyn et al., 2018). The corresponding threshold in global mean temperature for a decline of the GrIS can be calculated based on climate model-estimated scaling coefficients (Frieler et al., 2017) to be 0.8–3.2 °C above pre-industrial levels (95% interval) (*medium confidence*), with a best estimate of 1.6 °C (Robinson et al., 2012). Using the model of Robinson et al. (2012), Clark et al. (2016) found a 10,000-year commitment of about 0.4 m GMSL equivalent/°C up to 2.7°C, above which the commitment increased to about 3 m/°C, leading to a complete loss of the GrIS within 10,000 years by about 5°C of warming. The rate at which the commitment is realized depends strongly upon the amount of warming, with complete loss taking about 8000 years at 5.5°C peak warming and about 3000 years at 8.6°C. These estimates rely upon a single, relatively low-complexity model, but they are consistent at an order of

magnitude level with predictions from other studies that show a logarithmic relationship between a given atmospheric temperature anomaly and the time taken to lose a certain proportion of ice volume (Ridley et al., 2010; Huybrechts et al., 2011; Golledge, in press). Furthermore, these studies are consistent with estimates from an analytical approach that predicts GrIS decay time as a function of atmospheric lapse rate, melt sensitivity, and the initial equilibrium line altitude (Levermann and Winkelmann, 2016). The most rapid estimate of full loss is after 1000 years under RCP8.5 (7.28 m at 84th percentile) (Aschwanden et al., 2019). Orders of magnitude agreement between these studies gives *medium confidence* in plausible timeframes for long-term mass loss. [PLACEHOLDER FOR FURTHER ASSESSMENT WITH NEW LITERATURE].

[START BOX 9.2 HERE]

BOX 9.2: Insights into ice sheet evolution from paleo evidence

Environmental controls

Atmospheric forcing

In ice sheet model simulations the GrIS tends to equilibrate with MPWP forcing after 5,000-10,000 years or longer, giving *high confidence* that the GrIS took at least one to two thousand years to reach a minimum extent and volume during this period (Yan et al., 2014; Koenig et al., 2015). Simulations constrained by paleo-environmental proxy data indicate that Pliocene ice loss was probably driven by changes in atmospheric temperatures arising from elevated CO₂, rather than warming of the oceans (Hill et al., 2010). Some simulations of the AIS during the MPWP show that East Antarctic accumulation changes resulting from the greater moisture holding capacity of the atmosphere may have been sufficient to completely or partially offset sea level rise associated with dynamic losses from West Antarctica (Yan et al., 2016). Other ice sheet model simulations of the MPWP AIS suggest that dynamic losses outweighed surface mass balance changes, producing a net positive contribution to sea level (DeConto and Pollard, 2016; Golledge et al., 2017b).

During the Pliocene-Pleistocene transition (3-2.5 Ma), the GrIS is thought to have been sensitive to declining levels of atmospheric CO₂ (*medium confidence*) with no perennial ice sheet while CO₂ concentrations were above ~320 ppm and large volume changes occurring when CO₂ was in the range ~280-320 ppm (Tan et al., 2018). This study suggested a more important role for CO₂ than insolation in determining the eventual onset of the ice sheet at ~2.5 Ma (Bierman et al., 2016), and potentially supports a threshold for long-term GrIS loss below current CO₂ levels (~400 ppm) (Section 9.4.1.2). During Marine Isotope Stage 11 (MIS 11: ~395-424 ka) the GrIS may have sustained greater losses than during the Last Interglaciation (LIG, MIS 5e: ~116-129 ka) despite a smaller insolation forcing, due to the longer duration of MIS 11 (Robinson et al., 2017). This suggests (*medium confidence*) that GrIS ice loss is more or less proportional to the magnitude of time-integrated atmospheric warming.

The increasing use of coupled climate-ice models that use a full surface energy balance forcing rather than positive-degree-day schemes and which capture albedo-melt feedbacks has led to increased incompatibility between LIG simulations and proxy-based inferences. This may be a consequence of increased ice sheet model sensitivity to summer insolation and warming (e.g. Goelzer et al., 2016; Robinson and Goelzer, 2014; Van De Berg et al., 2011; Yau et al., 2016) or could indicate that proxy-based temperature reconstructions are over-estimated. Because of this discrepancy we have only *low confidence* in the utility of GrIS LIG studies to constrain future projections of the ice sheet. In Antarctica, the LIG climate remained too cold for widespread surface melting (Deconto and Pollard, 2016; Goelzer et al., 2016a). We therefore have *low confidence* that large-scale Antarctic ice shelf collapse necessary to trigger marine ice cliff instability occurred during the LIG.

Warming during the LIG occurred under lower-than-present atmospheric CO₂ levels as a consequence of orbitally-induced higher summer insolation (Dutton et al., 2015b; Goelzer et al., 2016a). It is not therefore a direct analogue for future warming, in which elevated CO₂ will result in year-round atmospheric warming. However, the LIG may be a useful analogue for assessing regional responses to warmer climates, particularly in the Arctic and Greenland.

Oceanic forcing

The growth of marine-based sectors of the Antarctic ice sheet most probably first occurred during the Early Oligocene (c. 32.8 Ma), becoming more established after c. 24 Ma when atmospheric CO₂ fell below 600 ppm (Levy et al., 2019). Based on evidence of ice-proximal glaciomarine sedimentation as well as changes in palynological and geochemical indicators there is *high confidence* that dynamic marine-based ice sheets existed at times through the Miocene (Sangiorgi et al., 2018; Levy et al., 2019) and into the Pliocene (5 – 3 Ma; (Patterson et al., 2014)). At this time oceanic controls on ice sheet behaviour became increasingly important with proxy-based sea-level reconstructions suggesting repeated growth and collapse of marine-based Antarctic ice sheets at CO₂ levels 300 – 500 ppm (Grant et al., 2019). Marine-terminating margins of the Greenland Ice Sheet only developed after 2.7 Ma as a consequence of changes in the Western Boundary Undercurrent (Blake-Mizen et al., 2019).

Marine Isotope Stage 31 (MIS31, 1.08 to 1.06 Ma) was a warmer-than-present interglacial that occurred during a period of peak Southern Hemisphere (SH) summer insolation. Based on modelled and proxy evidence there is *medium confidence* that MIS31 was associated with warmer ocean temperatures and reduced sea ice, as well as substantial retreat of marine-based Antarctic margins (DeConto et al., 2012; Beltran et al., 2020). Proxy-based sea-surface temperatures 5°C above present values are consistent with a scenario in which increased SH insolation intensifies polar westerly winds and encourages more active upwelling of Circumpolar Deep Water (Beltran et al., 2020).

Ocean temperatures during the LIG are effectively unconstrained by paleo-environmental proxy records at latitudes above 57°S (Capron et al., 2014). Some ice sheet models attempting to capture Antarctic ice sheet retreat have therefore imposed additional Southern Ocean subsurface warming of at least 2°C relative to preindustrial levels (DeConto and Pollard, 2016; Sutter et al., 2016). Simulations in which ocean temperature anomalies are used directly from ocean-atmosphere climate models typically yield lower mass loss estimates (Goelzer et al., 2018b; Clark et al., in press). It has also been proposed that LIG Antarctic ice loss was primarily driven by sea-level and ice viscosity changes, rather than ocean heat (Goelzer et al., 2016a). Disagreement between studies means that whilst we have *medium confidence* that ocean forcing dominated over atmospheric forcing at this time, we have *low confidence* in the magnitude of Antarctic mass loss during the LIG that can be attributed to ocean forcing.

Retreat of marine-based Antarctic margins following ice-sheet expansion at the Last Glacial Maximum (LGM) has been shown in one modelling to have been amplified by meltwater-induced ice-ocean feedbacks (Golledge et al., 2014). Enhanced Southern Ocean stratification triggered by northern hemisphere meltwater and reduced meridional overturning trapped warm water at depth on the continental shelf, initiating margin retreat. Antarctic ice sheet melt further strengthened the stratification and promoted further grounding-line retreat (Golledge et al., 2014). Other studies have also interpreted sub-surface ocean warming as a key driver of LGM ice sheet retreat in Antarctica (Weber et al., 2014) and in the Eurasian Ice Sheet complex (Brendryen et al., submitted).

Sea ice forcing

Biological proxies from an Antarctic marine sediment core show that expansion of the marine-based ice sheet at around 14–15 Ma (mid-Miocene) coincided with the establishment of perennial sea ice (Levy et al., 2019), suggesting sea ice presence may reduce losses of grounded ice. Similar links between sea-ice extent and marine-based ice sheet stability have also been interpreted from changes in iceberg-rafted debris concentrations in Pliocene and early Pleistocene sediments (Patterson et al., 2014). Based on consistency between these reconstructions and recent observations and modelling studies (Miles et al., 2013, 2016, 2017; Robel, 2017; Greene et al., 2018), there is *high confidence* that sea ice plays a role in controlling the behaviour of floating and grounded ice.

Dynamic controls

Ice streaming

Ice streams are responsible for the majority of ice sheet discharge (Rignot et al., 2011, 2014) but their role in ice sheet evolution is unclear. Theoretical analyses shows that large ice sheets should become more sensitive

to environmental forcing as they grow larger, because ice streams become more mature (Robel and Tziperman, 2016). However, a paleoglaciological reconstruction of the Laurentide Ice Sheet has been used to suggest that ice stream activity at the LGM scaled with ice sheet volume (Stokes et al., 2016), with decreased streaming during deglaciation. During deglaciation of the last British-Irish Ice Sheet one of its major ice streams underwent accelerated retreat when grounding lines retreated onto more resistant bedrock and encountered topography that favoured a Marine Ice Sheet Instability (MISI)-style response (Bradwell et al., 2019), consistent with findings from Antarctica (Jones et al., 2015b). Problems relating geological evidence to glaciological behaviour, coupled with an apparently high dependence of ice stream behaviour on local basal conditions mean that there is only *low confidence* in linking ice stream activity with overall stability of an ice sheet.

Instability mechanisms

The slow advance and rapid retreat of marine-based parts of the West and to a lesser extent East Antarctic ice sheets (WAIS, EAIS) during glacial-interglacial cycles suggests the existence of positive feedbacks. The MISI and marine ice cliff instability (MICI) are two mechanisms of self-sustaining ice loss that have been proposed to explain the apparent collapse of marine-based sectors of the AIS in the past. However, the use of paleo-environmental proxy records to evaluate whether MISI or MICI mechanisms were responsible for past AIS retreats is complicated because reconstructions of the magnitude, location and rates of ice loss are typically only poorly constrained. Models of long-term AIS evolution are necessarily low in spatial resolution, and ice dynamic processes are simplified, due to computational expense. The sub-kilometre resolution required to resolve grounding line migration and ice cliff geometry has so far only been used for simulations of the whole ice sheet spanning several hundred and up to 1000 years in length (Cornford et al., 2016; Martin et al., 2019). There is very limited paleogeological evidence to directly inform the MICI hypothesis. Iceberg keel marks have been interpreted as records of repeated collapse of tall coastal cliffs in the Amundsen Sea Embayment during the last deglaciation (Wise et al., 2017), but individual cliff collapses do not necessarily imply a sustained, rapid instability. MICI is currently incorporated in only one ice sheet model, and it improves agreement of this model with reconstructions by substantially increasing the sensitivity to warming during the Pliocene and Miocene (DeConto and Pollard, 2016; Gasson et al., 2016a; Pollard et al., 2015). However, it is not necessarily required (Edwards et al., 2019b), and better agreement might also be achieved with models that do not include MICI, or impose greater ocean warming (which is currently not well constrained), or incorporate meltwater-ocean feedbacks that increase AIS ice loss (Golledge et al., 2019). Furthermore, (Raymo et al., 2018) consider the uncertainties of geochemical sea level proxies and dynamic topography to be so large they do not allow ‘a robust estimate’ of the AIS contribution to Pliocene sea level, thus preventing evaluation of whether higher or lower model estimates are more plausible. There is therefore *low confidence* in the existence of a MICI-style instability occurring in past warm climates.

Solid earth feedbacks

The rate at which ice sheet retreat occurs is reduced by the rate of isostatic rebound as well as by the stabilizing influence of reduced ice-ocean-gravitation (Box 9.1). There is high confidence that such processes affected Antarctic ice sheet behaviour during the geological past (Gomez et al., 2018; McKay et al., 2016). There is *medium confidence*, based on observations and modelling, that ongoing viscous uplift of the bed following a halt in ice sheet retreat can stabilise or initiate readvance of marine margins (Kingslake et al., 2018; Pollard et al., 2017; Larour et al., 2019).

[END BOX 9.2 HERE]

9.4.2 Antarctica

9.4.2.1 Recent observations and model evaluation

For the period 1992 to 2016-2017, the average grounded Antarctic Ice Sheet (AIS) mass loss was *likely* about $100 \pm 50 \text{ Gt yr}^{-1}$, amounting to about $2500 \pm 500 \text{ Gt}$ over this period (see Figure 9.18) (Bamber et al., 2018b; The IMBIE Team, 2018), equivalent to $6.9 \pm 1.4 \text{ mm}$ sea level rise. These estimates of AIS mass

loss, which combine observational estimates of the evolution of the grounded AIS mass balance based on satellite altimetry, gravity anomalies (GRACE), and the input-output method (e.g., Rignot et al., 2019), are, respectively, a review of post-AR5 studies up to 2016 (Bamber et al., 2018b) and a multi-team intercomparison including new results up to 2017 (The IMBIE Team, 2018), and agree within the estimated uncertainties. Methodological limitations, linked in particular to estimates of firn compaction rates and corrections for glacial isostatic adjustment (Rignot et al., 2019) and snowfall variability (Palerme et al., 2017a), imply that there is only *medium confidence* that the East Antarctic Ice Sheet (EAIS) has been close to balance during this period (Bamber et al., 2018b; The IMBIE Team, 2018). There is *high confidence* that mass loss of the AIS has been dominated by high ice discharge rates over the West Antarctic Ice Sheet (WAIS) and the Antarctic Peninsula (Mouginot et al., 2014; Sutterley et al., 2014; Wouters et al., 2015), leading to substantial observed surface height changes (McMillan et al., 2014; Bamber et al., 2018b; Schröder et al., 2019; Shepherd et al., 2019), particularly in coastal regions, as can be seen in Figure 9.20. In summary, the published estimates using different techniques are overall consistent and agree on the geographic patterns of recent Antarctic mass change. Based on these studies, there is *very high confidence* that the grounded AIS has lost mass since the early 1990s. A longer time series based on the input/output method (Rignot et al., 2019) suggests that the AIS was close to balance between 1979 and 1992.

Recent studies and syntheses (Bamber et al., 2018b; The IMBIE Team, 2018; Rignot et al., 2019) consistently suggest an acceleration of Antarctic mass loss over the last decades (*medium confidence*). Mass loss of the WAIS and Antarctic Peninsula has *likely* accelerated over this period and particularly since about 2000, essentially due to increased ice discharge (Harig and Simons, 2015; Paolo et al., 2015; Forsberg et al., 2017; Bamber et al., 2018b; Gardner et al., 2018; The IMBIE Team, 2018; Rignot et al., 2019) (Figure 9.18). Two recent syntheses for the period after 1992 (Bamber et al., 2018b; The IMBIE Team, 2018) agree on an increase of the mass imbalance of the combined WAIS and Antarctic Peninsula from $60 \pm 30 \text{ Gt yr}^{-1}$ in the early to mid 1990s to about $180 \pm 40 \text{ Gt yr}^{-1}$ 20 years later. Multiple studies show that this mass loss increase is *very likely* caused by glacier flow acceleration (dynamic thinning) (e.g., Mouginot et al., 2014; Sutterley et al., 2014; Wouters et al., 2015), most extensively in the Amundsen Sea Embayment (ASE) of West Antarctica. Because of consistent multiple evidence, there is *high confidence* in mass loss of the Totten Glacier of the EAIS (Miles et al., 2013; Li et al., 2016d; Mohajerani et al., 2018; Rignot et al., 2019; Schröder et al., 2019; Shepherd et al., 2019) since about 2000, linked to changes in coastal ice dynamics (Li et al., 2016d).

There is robust evidence and therefore *high confidence* that mass loss of the AIS is primarily due to reductions in the thickness and extent of floating ice shelves (Pritchard et al., 2012; Paolo et al., 2015), while there is no trend in ice-sheet-wide surface mass balance since 1979 (The IMBIE Team, 2018). Ice-shelf thinning and retreat reduce buttressing of the glaciers that feed them. This can lead to glacier speed-up and thinning (e.g., De Angelis and Skvarca, 2003), unless the ice-shelf mass loss concerns its “passive” fringe, defined as the part of the shelf that can be removed without major effects on the ice shelf dynamics (Fürst et al., 2016; Reese et al., 2018). This passive fringe is very limited or absent for the ice shelves in the Amundsen and Bellingshausen sectors (Fürst et al., 2016).

It is *very likely* that volume loss from Antarctic ice shelves has been accelerating overall since the early 1990s (e.g., Paolo et al., 2015). Antarctic ice shelves primarily lose mass through basal melting and iceberg calving (e.g., Depoorter et al., 2013; Rignot et al., 2013; Liu et al., 2015), and although surface melting can be important in some cases (Lenaerts et al., 2017), basal melt *likely* dominates Antarctic ice shelf mass losses (Liu et al., 2015). There is evidence (*medium confidence*) that ice-shelf thinning through basal melting can trigger increased iceberg calving (Liu et al., 2015), with calving rates often as high as basal melting in thinning ice shelves (Depoorter et al., 2013; Rignot et al., 2013; Liu et al., 2015). Consistent and robust evidence across multiple studies (e.g., Jacobs et al., 2011; Schmidtko et al., 2014; Cook et al., 2016; Khazendar et al., 2016; Rintoul et al., 2016; Turner et al., 2017a; Jenkins et al., 2018), supported by paleoclimatic evidence (Hillenbrand et al., 2017), suggests that the acceleration of ice-shelf mass losses has *very likely* been driven primarily by an increase in ocean thermal forcing associated with upwelling Circumpolar Deep Water (CDW) (Sections 9.2.2.3, 9.2.3.2) that reaches the grounding lines of glaciers (*high confidence*). However, in situ measurements within the ice-ocean boundary layer are extremely sparse, and modelling ice-shelf basal melt remains challenging because of insufficient process understanding and the

paucity of in-situ observations (e.g., Turner et al., 2017a).

[START FIGURE 9.20 HERE]

Figure 9.20: Estimated sea level equivalent volume relative to present day (top left panel, detail in supplementary material) and spatial extent (bottom left) of the Antarctic ice sheet during the mid-Pliocene Warm Period, the Last Interglaciation, and the Last Glacial Maximum (see Annex II for details) compared to present-day extent (Fretwell et al., 2013). Paleo extents are interpreted from individual reconstructions (Anderson et al., 2002; Bentley et al., 2014; De Boer et al., 2015; Goelzer et al., 2016a). Grey shading shows extent of grounded ice. (top right) Cumulative mass loss (sea level equivalent) from the Antarctic ice sheet since 2015, with satellite observations shown from 1993 (Bamber et al., 2018a; The IMBIE Team, 2018; WCRP Global Sea Level Budget Group, 2018), and ISMIP6 projected changes by 2100 under RCP8.5/SSP5-8.5 and RCP2.6/SSP1-2.6 scenarios (thin lines from (Seroussi et al., 2019) and (Nowicki et al., submitted, a)) and likely range of the ISMIP6 emulation (shaded line, (Edwards et al., submitted)). (bottom right) Maps of mean elevation changes 1978-2017 derived from restored analog radar records (Schroeder et al., 2019) and projected changes by 2100 under selected RCP 8.5 scenarios (Seroussi et al., 2019).

[END FIGURE 9.20 HERE]

Although a mechanistic link between sea ice presence and ice sheet stability has not yet been proven, it is *likely* that wind-driven reductions in sea ice formation rate also lead to reduced vertical mixing and increased ocean stratification, allowing warming of the subsurface due to the penetration of warm CDW onto the continental shelf (Miles et al., 2016). There is *high confidence* from consistent observations that changes in sea ice thickness and extent modulate ice shelf flow and glacier terminus position (Miles et al., 2013, 2016, 2017; Greene et al., 2018; Massom et al., 2018), which may in turn affect the dynamics of upstream grounded ice (*medium confidence*).

A substantial increase in tributary glacier flow had been observed after the major Larsen A and B ice-shelf collapse events in 1995 and 2002, respectively (e.g., De Angelis and Skvarca, 2003). Mass losses of glaciers in this region are currently continuing (*high confidence*) (Rott et al., 2018; Seehaus et al., 2018; The IMBIE Team, 2018), with glaciers affected by ice shelf disintegration showing the largest retreat since the mid-1980s (Seehaus et al., 2018). However, although flow speed of these glaciers is still above that observed before the ice shelf disintegration about two decades ago (Seehaus et al., 2018) (*medium confidence*), a general trend towards calving flux decrease, a possible sign of stabilization, was reported after 2011 (Rott et al., 2018), albeit with substantial variability between the individual glaciers.

Changes in mass or volume of the ice sheet can also be due to changes of the horizontal extent of the grounded ice. Remotely-sensed reconstructions of glacier grounding line positions reveal that over recent decades there has been widespread recession of grounding lines in several locations around the continent (Li et al., 2015b; Christie et al., 2016; Konrad et al., 2018). There is *high confidence* that grounding line retreat is pervasive along the Bellingshausen (Christie et al., 2016) and Amundsen Sea (Rignot et al., 2014) sectors of West Antarctica since about 1990. There is *medium confidence* that grounding line retreat has been occurring on the East Antarctic Totten Glacier since 1996 (Li et al., 2015b). Furthermore, grounding line retreat in excess of 1.5 km between 2003 and 2015 has been observed in parts of Marie Byrd Land (Christie et al., 2018). There is *limited evidence (low confidence)* that grounding line retreat occurred along parts of the English Coast in the Southern Antarctic Peninsula since about 1973 (Holt et al., 2013; Wouters et al., 2015).

Since 2010, grounding line retreat rates estimated to exceed the mean rate of retreat following the LGM (typically about 25 m yr⁻¹) have been mapped across 22%, 3%, and 10% of West Antarctica, East Antarctica, and Antarctic Peninsula margins, respectively (Konrad et al., 2018), but there is *low confidence* in this estimate because of the high temporal variability, short observational record and large uncertainties concerning retreat rates since the LGM. It is *about as likely as not (medium confidence)* that collapse of the

WAIS due to grounding line retreat in the Amundsen Sea Embayment (ASE), as suggested by some modelling studies (Favier et al., 2014; Joughin et al., 2014; Feldmann and Levermann, 2015), is presently close to being underway. For example, grounding line retreat of Pine Island Glacier, interpreted as indicative of a probably engaged Marine Ice Sheet Instability (MISI; Box 9.3) (Favier et al., 2014), seems to have stabilized recently (Konrad et al., 2018).

Although progress has been made recently in quantifying current Antarctic precipitation rates using CloudSat (Palermo et al., 2014), there is an uncertainty of at least 10-15% (*medium confidence*) in current observation- and model-based values of the surface mass balance (SMB) of the grounded parts of the AIS of around 2100 Gt yr⁻¹ (Favier et al., 2013; Van Wessem et al., 2018; Agosta et al., 2019). Precipitation variability dominates the interannual to decadal AIS mass balance variability (Rignot et al., 2019), although compensation effects due to the spatial heterogeneity of this precipitation variability dampen its impact on the ice-sheet wide SMB (Fyke et al., 2017). Contrary to older studies (Monaghan et al., 2006), there is some recent evidence (*low confidence*) from modelling, observations and reanalyses that increased Antarctic precipitation might have mitigated dynamical ice mass losses by about 50 Gt yr⁻¹ since about 1990 (Lenaerts et al., 2018; Medley and Thomas, 2019) and at lower rates over the entire 20th century (Medley and Thomas, 2019). On sub-continental scales, it is not yet clear whether accumulation increase on multi-decadal timescales in West Antarctica (Thomas et al., 2015) is due to warming (Bromwich et al., 2013) or variations in the intensity of the Amundsen-Bellingshausen Sea Low (Thomas et al., 2015). However, since about the year 2000, an absence of warming in the region (Nicolas and Bromwich, 2014) has coincided with an absence of accumulation trends (Seo et al., 2015).

Ice sheet modelling of Antarctica is less mature than numerical modelling of the global climate system, due in part to the relatively short length of the satellite records that first highlighted rapid changes. Since AR5, simulations are now more commonly evaluated or formally calibrated with modern observations (i.e. separate to initialisation) and/or paleodata, either through perturbed parameter ensembles or model intercomparison projects. This has improved understanding and confidence in ice sheet model projections especially with respect to the future contribution of Antarctica to global sea level rise. However, modelling of the ice-ocean interface remains challenging. Three approaches exist: standalone ice sheet models driven by (i) basal melting from ocean-ice-shelf models or (ii) thermal forcing from ocean models, via a temperature-melt parameterisation, or (iii) coupled ocean-ice-sheet models. All have known limitations due to, among other things, poor knowledge of the bathymetry and ocean state beneath ice shelves (Donat-Magnin et al., 2017), parameterised modelling of grounding line evolution (i.e. cavity geometry changes) in most ice sheet models, and uncertainties about possible marine ice sheet instabilities. Although some models now account for ice-shelf basal topography and sub-ice-shelf buoyant flow (Asay-Davis et al., 2017), fine-scale complexities such as seawater intrusion along preferential channels at tidal frequencies, which causes local high melt rates (Milillo et al., 2019), are not represented in coupled ice sheet-ocean models.

There are some indications of limited improvement in the representation of Antarctic atmospheric fields such as precipitation and temperature from CMIP5 to CMIP6 (Gorte et al., 2019; Roussel et al., submitted). An analysis of CMIP5 and CMIP6 long-term (1850-2000) mean SMB, its long-term trends and interannual variability suggests, on average, better agreement of CMIP6 with ice-core-based reconstructions, although this might be due to reduced sample size (Gorte et al., 2019). However, concerning continental-mean and regional-scale snowfall rates, a comparison with 2007-2010 CloudSat estimates suggests no substantial improvement from CMIP5 to CMIP6, neither in atmosphere-only nor in coupled model setups, in spite of reduced surface air temperature errors (Roussel et al., submitted). [PLACEHOLDER FOR FURTHER QUANTITATIVE ASSESSMENT OF CMIP6 RESULTS, IN PARTICULAR ISMIP6]

9.4.2.2 Future projections

Based on recent observations and numerical simulations the AIS will *likely (medium confidence)* lose mass by the end of the century. The SROCC reassessed the contribution of Antarctica to GMSL for this century based on post-AR5 process-based studies, which are able to reproduce marine ice sheet instability (MISI) (Levermann et al., 2014; Golledge et al., 2015, 2019; Ritz et al., 2015; Bulthuis et al., 2019). The latter had

been estimated to not exceed several tenths of a meter in AR5 (Church et al., 2013a), but had been excluded in the AR5 projection for Antarctica. The SROCC assessed this contribution by ice dynamics to be *likely* 0.16 m (0.02 to 0.37 m 5-95% range) in 2100 under RCP8.5, resulting in projected Antarctica median contribution of 0.04 m (0.01 to 0.1 m *likely* range) and 0.12 m (0.03 to 0.28 m *likely* range) in 2100 under RCP 2.6 and RCP8.5, respectively. This range is smaller than a structured expert judgement study (Bamber et al., 2019), which projected median (5-95% range) contributions from Antarctica by 2100 of 0.09 (0.08 to 0.53) m and 0.21 (-0.11 to 1.32) m for scenarios approximately equivalent to RCP2.6 and RCP8.5, respectively. We note that Bamber et al. (2019) provides higher and more uncertain values than the previous exercise (Bamber and Aspinall, 2013). Figure 9.20 and Table 9.2 show the projected AIS contribution to GMSL by 2100 under selected scenarios. New projections since SROCC come primarily from two efforts: the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6; (Nowicki et al., submitted, a, submitted, b; Seroussi et al., submitted)) and the Linear Antarctic Response Model Intercomparison Project, Phase 2 (LARMIP-2, (Levermann et al., submitted, a)). The contribution for 15 ice sheet modelling groups (Figure 9.20) forced by two global climate models for RCP2.6 and six for RCP8.5 (Seroussi et al., submitted), one for SSP1-2.6 and four for SSP5-8.5 (Nowicki et al., submitted, a), result in a sea level contribution by 2100 that ranges from -0.01 to 0.16 m under RCP2.6 and -0.08 to 0.3 m under RCP8.5, and similar mass losses are projected with the equivalent SSPs (Table 9.2; (Nowicki et al., submitted, a; Seroussi et al., submitted)). As for the GrIS, the CMIP5 models used to force the ice sheet models were selected to sample a range of future climate (Barthel et al., 2019), while the CMIP6 models were selected because of their availability. The statistical emulation of (Edwards et al., submitted) extends the ISMIP6 projections to a larger range of future climate, projecting median (5 to 95 percentiles) sea level of 0.07 (-0.12 to 0.33) m for SSP1-2.6 and SSP2-4.5, and 0.07 (-0.12 to 0.32) m for SSP5-8.5, suggesting a limited scenario dependence of projected mass losses of Antarctica. As shown in (Seroussi et al., submitted), this is in part due to the complex interplay between mass gain from increased snowfall and mass loss due to ice shelf basal melting (and consequent speed up of glacier flow). (Edwards et al., submitted) attributes 2/3 of the uncertainty in projections to model structural uncertainty in climate models and ice sheet models (ISMs), and 1/3 to ocean driven melt. The ISMIP6 based projections are relative to 2015 and the ice mass change due to climatic changes prior to 2015 has been removed, as these historical trends may not always represent physical changes but rather limitations in the procedures used to initialize ice sheet models (Seroussi et al., submitted). LARMIP-2 derives linear response function from the ISMIP6 models (without removing historical trends in the ice sheet simulation) and explores the uncertainty of the AIS that arises from large uncertainty in ocean forcing (Levermann et al., submitted, a). Unlike ISMIP6, LARMIP-2 does not account for ice sheet response to changes in surface mass balance and results in a higher range of ice shelf basal melt rates, yielding sea level projections that are much larger than ISMIP6. [PLACEHOLDER FOR THE ASSESSMENT OF THE ANTARCTIC CONTRIBUTION AT 2100 UNTIL NEW STUDIES ARE AVAILABLE FROM ISMIP6 THAT EXPLORE CLIMATE AND ICE SHEET MODEL UNCERTAINTIES MORE COMPREHENSIVELY].

[START TABLE 9.2 HERE]

Table 9.2: Projected sea level contributions in meters from the Antarctic ice sheet by 2100 relative to 2000 for selected CMIP5 RCP and CMIP6 SSP scenarios, unless noted otherwise. *denotes contribution reported in SROCC Table 4.3. ** denotes contribution in 2100 relative to 2015 and excludes committed ice mass change

Study	RCP2.6	RCP4.5	RCP8.5	Notes
IPCC AR5 (Church et al., 2013a)Table 13.5	0.05(-0.03 to 0.14)	0.05(-0.04 to 0.13)	0.04(-0.06 to 0.12)	Median (<i>likely</i> range) contributions in 2081-2100 relative to 1986-2005
*(Levermann et al., 2014)	0.07	0.09	0.11	Median, ocean forcing only
*(Gollledge et al., 2015)	0.02	0.05	0.18	Median

*(Golledge et al., 2019)	---	0.04	0.11	Average from two basal melt parameterisation
*(Bulthuis et al., 2019)	0.03	0.05	0.11	Average from multiple sensitivity experiments
IPCC SROCC (Oppenheimer et al., 2019) Table 4.4	0.04(0.01 to 0.11)	0.06(0.01 to 0.15)	0.12(0.03 to 0.28)	Median (<i>likely</i> range)
*(Deconto and Pollard, 2016)	0.14	0.41	0.79	With MICI, not used in SROCC
(Edwards et al., 2019a)	0.187(-0.026 to 0.48)		0.788(0.202 to 1.566)	Median (5 to 95%), with MICI, not used in SROCC
(Edwards et al., 2019a)	-0.01(-0.087 to 0.13)		0.206(0.089 to 0.388)	Median (5 to 95%), no MICI, not used in SROCC
** (Seroussi et al., submitted)	-0.01 to 0.16	---	-0.08 to 0.3	ISMIP6 multi-model ensemble from 2 ESMs for RCP2.6 and 6 ESMs for RCP8.5
(Levermann et al., submitted, a)	0.14(0.05 to 0.4)	0.15(0.05 to 0.47)	0.18(0.06 to 0.61)	Median (5 to 95%), ocean forcing only
<i>Placeholder for any other study</i>				
Study	SSP1-2.6	SSP2-4.5	SSP5-8.5	
** (Nowicki et al., submitted, a)	-0.04 to 0.11	---	-0.09 to 0.15	ISMIP6 multi-model ensemble from 1 ESM for SSP1-2.6 and 4 ESMs for SSP5-8.5
** (Edwards et al., submitted)	0.07(-0.12 to 0.33)	0.07(-0.12 to 0.33)	0.07(-0.12 to 0.32)	Emulator of ISMIP6, median (5 to 95%)
<i>PLACEHOLDER for any other study</i>				

[END TABLE 9.2 HERE]

Recent studies (e.g., Lenaerts et al. 2016) confirm the AR5 conclusion that increased Antarctic precipitation, due to increased moisture availability in a warmer climate, will *very likely* result in a negative contribution to future sea level rise. The physical reasoning is supported by paleo-evidence (Box 9.2), large scale global climate simulations (Genthon et al., 2009b; Lenaerts et al., 2016; Palerme et al., 2017b) as well as regional atmospheric simulations (Ligtenberg et al., 2013). The total snowfall increase is *likely* between 4% and 8% per °C of spatially-averaged temperature increase over Antarctica (Frieler et al., 2015; Previdi and Polvani, 2016). CMIP6 models simulate a precipitation sensitivity of about 4.5-6% per °C of continental warming (Bracegirdle et al., submitted). CMIP5 models that are in better agreement with observations of present precipitation rates tend to predict larger future increases (Palerme et al., 2017b), although long-term continental scale changes are controlled by thermodynamic constraints on the availability of heat and moisture, at least for substantial warming (Krinner et al., 2014; Frieler et al., 2015; Medley and Thomas, 2019). Regional patterns depend on atmospheric circulation. The snowfall is *very likely* to be enhanced near the coast, simulated consistently by many models (e.g., Genthon et al., 2009a), linked to much higher current precipitation rates in these regions. It is also *very likely* that surface melting will increase with rising air temperatures, i.e. contributing to sea level, but there is *low confidence* in the exact magnitude and location

(Trusel et al., 2015). Surface melting is predicted to remain confined to low-lying areas at the margins during the 21st century even under scenarios of high greenhouse gas concentrations, except over the Antarctic Peninsula (e.g., Trusel et al. 2015). Overall, ice loss due directly to surface melting is projected to be lower than the gain from additional snowfall, so the future net SMB contribution up to 2100 is negative (*high confidence*). As no radical improvement in the quality of present Antarctic surface climate in CMIP6 compared to CMIP5 has been reported, substantial changes to SMB projections since AR5 (e.g. *likely* 0.01-0.07 m by 2100 under RCP8.5) are not expected as an outcome of CMIP6. CMIP6 projections suggest some scenario dependency of the weak Antarctic climate change amplification: projected Antarctic surface warming by 2100 is somewhat stronger in stabilization and aggressive mitigation scenarios than in higher emission scenarios, due to stronger global-relative Southern Ocean warming and relatively stronger effects of ozone recovery (Bracegirdle et al., submitted). [PLACEHOLDER: THE ASSESSMENT WILL BE COMPLETED WHEN MORE CMIP6 RESULTS ARE AVAILABLE.]

Sub-ice-shelf melting is the most investigated perturbation to the AIS, because it drives current dynamical losses (Section 9.4.2.1). Ice sheet models (e.g. Arthern and Williams, 2017; Nias et al., 2016) predict Antarctica will continue to contribute to GMSL this century even with no additional changes in climate (*medium confidence*). However, projecting future changes in ice shelf basal melting remains a key uncertainty in AIS projections (see Figure 9.1; Box 9.3; Section 9.7). The main process for increased ice shelf basal melting is enhanced intrusion of warm CDW onto the Antarctic shelves (Khazendar et al., 2013; Schmidtko et al., 2014; Mohajerani et al., 2018; Naughten et al., 2018), and warming surface waters intruding under cavities (Naughten et al., 2018). However, there is still little consensus regarding future change in CDW (Section 9.2.2.3), and more generally *low confidence* in future change in the temperature of Antarctic ice shelf cavities (Section 9.2.3.2). Two models that resolve ice shelf cavities have been used to make basal melting projections for SRES and RCP scenarios over the next one or two centuries, forced by atmospheric models (Hellmer et al., 2012; Timmermann and Hellmer, 2013; Timmermann and Goeller, 2017; Naughten et al., 2018), though other models exist so greater diversity for RCP projections is expected in the future (Dinniman et al., 2016; Jourdain et al., 2017; Mathiot et al., 2017; Merino et al., 2018; Naughten et al., 2018; Favier et al., 2019). New generation coupled ocean--ice-sheet--ice-shelf models (Asay-Davis et al., 2016; De Rydt and Gudmundsson, 2016; Favier et al., 2019; Goldberg et al., 2012, Jourdain et al., 2018; Seroussi et al., 2017) will enable feedbacks between the ice sheet and oceanic and atmospheric forcing to be represented. The majority of Antarctic ice sheet projections use ice sheet models that are not coupled to ocean models, instead using a variety of methods to prescribe sub-shelf melt: (i) Prescribed melt with ad-hoc perturbation (uniform or depth-dependent, e.g. (Seroussi et al., 2019)); (ii) Prescribed onset date for grounding line retreat (Ritz et al., 2015); (iii) Prescribed melt rate from ocean projections (with static ice shelf cavities), themselves forced by a CMIP simulation (Cornford et al., 2015); (iv) Parameterized melt rates, i.e. melt rates are calculated at the simulated ice draft depth, based on ocean temperatures (and possibly salinity) from a climate simulation (Deconto and Pollard, 2016; Golledge et al., 2019; Nowicki et al., submitted, a; Seroussi, submitted). The limitations of the above methods to prescribe basal melt rates are the following: method (i) does not enable a connection between emission scenarios and melt rates, and depth-dependent formulations do not compare well to ice/ocean coupled models ((De Rydt and Gudmundsson, 2016; Seroussi et al., 2017), (ii) is empirical and hides poorly known processes behind an onset date, (iii) does not account for feedbacks between the ocean circulation and temperature and the geometry of the ice shelf base, (e.g (Donat-Magnin et al., 2017; Timmermann and Goeller, 2017) and needs to be extrapolated where grounding lines retreat, (iv) includes a variety of parameterizations (e.g. (Lazeroms et al., 2018; Reese et al., 2018; Pelle et al., 2019)) that respond differently to ocean warming (Favier et al., 2019) and that are usually calibrated to match present-day melt rates (using various methodologies as done for ISMIP6 (Jourdain et al., submitted)). The parameterizations and model simulations produce various melting patterns (Figure 9.21), which can make a large difference to ice sheet response (e.g. (Seroussi et al., 2017, submitted; Goldberg et al., 2018)). ISMIP6 use two calibrations of their parameterization (Figure 9.21: Mean Ant and PIGL), to reproduce mean basal melt observations for Antarctica or maximum melt values in Pine Island Glacier (Jourdain et al., submitted), and the latter results in larger melt rates close to the grounding line. Doubling of basal melt leads to more than ten times greater ice sheet mass loss (Seroussi et al., submitted), highlighting the non-linear ice sheet response to submarine melt. There is *high confidence* that Antarctic ice shelf basal melting will increase under the RCP4.5 and RCP8.5 scenarios, but *low confidence* in the projected melt rates because of present-day model uncertainty, limited exploration of possible climate

forcing and uncertainty in sub-shelf melt rate parameterization.

[START FIGURE 9.21 HERE]

Figure 9.21: Ice shelf basal melt rates for present-day (upper panels) and changes from present-day to the end of the 21st century under the RCP8.5 scenario (lower panels). Present-day melt rates were estimated through: the input-output method constrained by satellite data and atmosphere/snow simulations (Rignot et al., 2013) and representative of 2003–2008 (upper left); the non-local-PIGL parameterization constrained by observation-based ocean properties (Jourdain et al., submitted) and representative of 1995–2014 (upper centre); the Finite Element Sea-ice/ice-shelf Ocean Model (FESOM) simulation over 2006–2015, forced by atmospheric conditions from a CMIP5 multi-model mean (MMM) under the RCP8.5 scenario (Naughten et al., 2018) (upper right). Future anomalies are calculated as 2081–2100 minus present-day using the ISMIP6 non-local-MeanAnt and non-local-PIGL parameterizations ((Jourdain et al., submitted) lower left and centre respectively) based on projections from the NorESM1-M CMIP5 model, and the FESOM-MMM projection (lower right). Note the symmetric-log colour bar (linear around zero, logarithmic for stronger negative and positive values). Inset highlights the Amundsen Sea Region.

[END FIGURE 9.21 HERE]

Uncertainties of the ice sheet models themselves are generally becoming better quantified, through systematic model intercomparison projects and perturbed parameter ensembles (Bulthuis et al., 2019; Cornford et al., 2015; DeConto et al., 2019; DeConto and Pollard, 2016; Durand and Pattyn, 2015; Edwards et al., 2019; Levermann et al., 2014; Nias et al., 2016; Nowicki et al., 2016; Ritz et al., 2015; Ruckert et al., 2017; Seroussi et al., 2019), particularly for ice-ocean interactions (Pattyn and Durand, 2013; Seroussi et al., 2014; Asay-Davis et al., 2016; Arthern and Williams, 2017). As well as the coupled ocean/ice-sheet modelling described above, studies increasingly use ice sheet models coupled to regional ocean or atmosphere models or global climate models to capture local feedbacks including those of meltwater on climate (Goelzer et al., 2012; DeConto and Pollard, 2016; Golledge et al., 2019). Sub-kilometre resolution is used in some models to resolve grounding line processes (e.g. Cornford et al., 2015; Nias et al., 2016). Overall confidence in continental-scale and regional ice sheet projections for Antarctica has therefore greatly increased since AR5, but confidence is generally lower in ice sheet models than in climate models as they have been developed and evaluated for a shorter time (both in terms of satellite records and, related to this, the history of the field). Key processes of dynamic ice loss are still heavily parameterised in the vast majority of models, including submarine basal melting (see earlier discussion) and grounding line evolution (Pollard and DeConto, 2012; Pattyn et al., 2017). Treatment of sliding has a strong influence on the sensitivity of the ice sheet response to melting (Ritz et al., 2015; Tsai et al., 2015; Nias et al., 2016; Brondex et al., 2017; Pattyn, 2017; Bulthuis et al., 2019; Sun et al., submitted). How to best initialize ice sheet models for projections (e.g. Nowicki and Seroussi, 2018; Seroussi et al., 2019), and how to capture historical ice sheet trends (e.g. (Goldberg et al., 2015; Reese et al., submitted)), remains open questions. This means that despite process and resolution improvements since AR5, large uncertainty remains, particularly with respect to the upper tail of the distribution of possible sea level contributions (discussed further in Box 9.3).

The largest uncertainty in future Antarctic mass losses are the possibility of rapid and/or irreversible ice losses through instability of marine parts of the ice sheet, proposed via MISI and Marine Ice Cliff Instability (MICI) (Box 9.3), and whether this will lead to a collapse of the WAIS. The theoretical possibility of MISI in regions with downsloping bedrock below sea level, triggered by sub-ice-shelf melting or ice-shelf collapse, has long been suggested based on physical understanding and observational data (Weertman, 1974; Mercer, 1978). As discussed in Section 9.4.2.1, it is *as likely as not* that MISI is underway in the Amundsen Sea Embayment, and under future warming there is *medium confidence* that this could continue and potentially initiate in other marine-based parts of Antarctica for decades to centuries, with projections suggesting that parts of the Weddell Sea and Ross Sea sectors could be vulnerable (Levermann et al., 2014, submitted, b; Ritz et al., 2015; Golledge et al., 2017a). Topographic conditions that would allow for MISI are also present in East Antarctica, for example under the large marine-based Wilkes Basin (Fretwell et al., 2012). The maximum rate and time scale of sustained ice discharge under MISI are not well-constrained but

depend on bed conditions and topography (e.g. Ritz et al., 2015). Numerical simulations are divided on whether the ice discharge would be self-sustaining in the Amundsen Sea Embayment even if the original melt forcing was reduced or removed (Favier et al., 2014; Seroussi et al., 2014; Feldmann and Levermann, 2015; Arthern and Williams, 2017). Rates of sea level contribution exceeding 50 cm per century have been simulated in studies that use unrealistically extreme basal melt rates and/or collapse of all ice shelves (Cornford et al., 2016; Pattyn, 2017; Schlegel et al., 2018). [PLACEHOLDER FOR FURTHER ANALYSIS OF ISMIP6 RESULTS].

The MICI hypothesis of rapid, unmitigated calving triggered by ice shelf collapse (Box 9.3) has been included in one ice sheet model (Pollard et al., 2015), partly motivated by improving agreement of that model with reconstructed past sea level contributions. Initial projections exceeded 1 m by 2100 under RCP8.5 when calibrated with mid-Pliocene and Last Interglacial (LIG) Antarctic sea level contributions (Deconto and Pollard, 2016). The two main reasons for the high projections at 2100 are sensitivity to the lower bound used for the mid-Pliocene calibration (10 m) and high estimates of future ice shelf surface melting. A probabilistic reanalysis of DeConto and Pollard (2016) by (Edwards et al., 2019b) estimates a lower contribution (median 0.79 m by 2100), though still with large uncertainties (5-95% range 0.2 to 1.57 m, Table 9.2). The large uncertainties in paleoclimate data (Box 9.2) result in uncertainties in future projections: the higher DeConto and Pollard (2016) projections and Edwards et al. (2019) are based on the same model ensemble but differ in their choice of paleoclimate calibration (10 m versus 5 m for the mid-Pliocene minimum). We note that the SROCC does not incorporate the original MICI projections nor studies that build on this work (DeConto and Pollard, 2016; Edwards et al., 2019) in their revised assessment for Antarctica, due predominantly to the high ice shelf melting. [PLACEHOLDER FOR AN UPDATE]. There is therefore *low confidence* in simulating mechanisms that have the potential to cause widespread, sustained and very rapid ice loss from Antarctica this century through MICI.

Projections of ice shelf collapse due to melting at the ice sheet surface are uncertain both in terms of the atmospheric warming forcing and the response of the ice shelf surface. Surface air temperature projections in coastal regions are very sensitive to oceanic forcing (i.e. SST and sea-ice concentration in the vicinity of the continent) for which the spatial patterns are highly variable in multi-model ensembles (Krinner et al., 2014; Bracegirdle et al., 2015). The influence of atmospheric temperature on surface meltwater and firn depletion is highly uncertain, with predictions of melting by 2100 and subsequent ice shelf disintegration under RCP8.5 varying widely (Trusel et al., 2015; Deconto and Pollard, 2016; Bell et al., 2018; DeConto et al., submitted). Simulations with the regional climate model RACMO coupled with a multilayer snow model (Trusel et al., 2015) agree better with current melt rates (Bell et al., 2018) and predict less future melting than DeConto and Pollard (2016). Atmospheric feedbacks in which ice sheet discharge leads to surface ocean and atmosphere cooling may also reduce predictions of future surface melting (Golledge et al., 2019). In ISMIP6, mass loss due to hydrofracturing yields an additional mass loss of 8 mm compared to simulations without hydrofracture collapse (Seroussi et al., submitted). There is however *very low confidence* in projections of the possible timing and location of future ice-shelf disintegration due to surface meltwater.

Projections at 2200 include a GMSL contribution from Antarctica of 0.04-0.72 m (5-95%) under SRES A1B (Ritz et al., 2015), 0.25-0.29 m and 0.88-1.52 m (ranges) under RCP4.5 and RCP8.5 respectively (Golledge et al., 2015), and 0.0-0.3m and 0.08-0.8 m (5-95% range) under RCP4.5 and RCP8.5 respectively (Bulthuis et al. 2019). Projections with the MICI hypothesis by Deconto and Pollard (2016) reanalysed by (Edwards et al., 2019b) are far higher, 4.0-6.9 m (5-95% range) with MICI and 3.7-4.2m even without MICI (5-95%) for RCP8.5. Overall, these results mean that it is *very likely* that the sea level contribution from Antarctica will continue in the 22nd century, but there is *low confidence* in projecting the magnitude of the GMSL contribution.

Earlier studies found low or negative sea level contributions from the Antarctic ice sheet on millennial timescales, even under very high emission scenarios (e.g. Vizcaino et al., 2008; Goelzer et al., 2012). This is due to the dominant role of SMB and a lack of realistic ocean coupling, with negligible dynamic ice losses, in contrast to current observations and process understanding. More recent ice sheet models with improved representations of ocean-ice coupling used to estimate multi-millennial sea level commitment clearly show stronger responses, mostly around 1.6 m/°C (Levermann et al., 2013; Golledge et al., 2015; Winkelmann et

al., 2015; Clark et al., 2016; Deconto and Pollard, 2016). Golledge et al., (2015) find around 2.5-3 m and 5-9 m of sea level commitment for a warming of 2 °C and 8 °C after 5000 years, with most of the ice loss occurring in the first 1000 years of the simulations. Deconto and Pollard (2016), incorporating MICI, found a 5000-year commitment of 4.5 m and 15.4 m for RCP 4.5 and RCP 8.5, peaking at ~15 m and ~20 m after around 1000 years, respectively. Clark et al. (2016) found a 10,000-year commitment of about 0.7 m by 1°C warming, 2.4 m by 1.4°C of warming, 16 m by 2.7°C of warming, and about 4 m/°C warming thereafter without inclusion of MICI, with much of the response within 3000 years of peak emissions and at least 80% of the commitment within 5000 years under almost all their scenarios. This suggests that the long-term commitment projections may be less susceptible to the uncertainties about MICI and is governed more by the regional limits (or otherwise) of ice losses. Numerical modeling suggests Antarctica will become practically ice-free on multi-millennial time scales if all available fossil fuels are burned (Winkelmann et al., 2015). However, most multi-millennial Antarctic studies use the same two ice sheet models (Levermann et al., 2013; Golledge et al., 2015; Winkelmann et al., 2015; Clark et al., 2016; Deconto and Pollard, 2016). Furthermore, computational expense prohibits exploration of modelling uncertainties on these multi-millennial timescales. Due to uncertainties in long-term future climate trajectories, as well as uncertainties in ice sheet processes that may operate under much warmer than present conditions, we have *very low confidence* in the magnitude of multi-millennial Antarctic commitment to sea level.

It is *likely* that the Amundsen Sea Sector will remain Antarctica's largest contributor to global sea level rise throughout this century (*high confidence*), based on the consistency of satellite observations (Section 9.4.2.1) and ice sheet model projections. The volume of the WAIS that is susceptible to the MISI is equivalent to 3.3 m SLE (Bamber et al., 2009). There may be a threshold for collapse of the marine sector of the WAIS that is susceptible to the MISI related to emissions, level of warming or basal melting (DeConto and Pollard, 2016; Feldmann and Levermann, 2015; Golledge et al., 2015; Nowicki et al., 2013; Ritz et al., 2015), consistent with physical understanding that MISI would be self-sustaining (Weertman, 1974; Mercer, 1978). However there is *low confidence* in estimating such a threshold. As discussed above, it is not known to which degree buttressing effects within the ice sheet can stabilise ice discharge after destabilisation (Gudmundsson et al., 2012), whether MISI would continue if basal melting decreased, and whether it might be stabilised by solid earth influences on millennial-timescales (Gomez et al., 2015b; Larour et al. 2019). RCP8.5 simulations, driven by ISMIP6 ocean forcing only, trigger a long-term retreat of the WAIS, with mass loss accelerating late in the 21st century and rising steadily over the next several centuries without levelling off (Lipscomb et al., submitted). Projections for RCP2.6 suggest that this emissions scenario might avoid full collapse of the WAIS: at 2500, projections by Golledge et al., (2015) reach only 0.21-0.29 m, and even projections with MICI (Deconto and Pollard, 2016)(Deconto and Pollard, 2016)(Deconto and Pollard, 2016)(Deconto and Pollard, 2016; Gasson et al., 2016b) reach only 0.30-1.88 m (5-95%) (Edwards et al., 2019b), again suggesting the importance of regional constraints on sustained ice loss under a given emissions scenario [PLACEHOLDER FOR FURTHER ASSESSMENT INCLUDING ISMIP6 RESULTS]. Alternatively, a destabilisation of the WAIS might already have been triggered (see earlier discussion). If WAIS were lost it would take several millennia to regrow the ice sheet, which implies an irreversible contribution to GMSL on those timescales.

9.5 Terrestrial cryosphere: Glaciers, permafrost and seasonal snow cover

9.5.1 Glaciers

Glacier mass (Harrison, 2013; Roe et al., 2017; Marzeion et al., 2018) and thermal regime (e.g., Gilbert and Vincent, 2013) are sensitive to climate variability and change. Glaciers are therefore important indicators and symbols of climate change (Zemp et al., 2015, 2019; Wouters et al., 2019). In this section (Table 9.3), we consider ice masses outside of or peripheral to the large ice sheets on Greenland and Antarctica that are discussed in Section 9.4. Progress in observing, understanding and modelling the evolution of glaciers in the past, present and future since AR5 and SROCC is assessed. Global and large-scale changes of glaciers are summarized in Chapter 2 (Section 2.3.2.3). Glacier contribution to sea level rise assessed in this section is put into global context in Section 9.6. The impact of glacier changes on the hydrology of different regions is assessed in Chapter 8 and attribution of glacier changes in Chapter 3 (Section 3.4.3.1).

Glaciers are important seasonal to long-term water reservoirs that regulate the river runoff on a daily, seasonal and inter-annual time scales (Bliss et al., 2014). SROCC concluded with *very high confidence* that glacier decline has affected and will continue to change the amounts and seasonality of river runoff in many glacier-fed river basins (Huss and Hock, 2018). Glaciers are large freshwater reservoirs and can therefore alleviate effects of drought (Pritchard, 2019; Tiel et al., 2019). Glaciers can directly and indirectly cause natural hazards during their retreat (Veh et al., 2019b, 2019a), as assessed in Chapter 12.

AR5 stated with *high confidence* and SROCC with *very high confidence* that the world's glaciers are presently in disequilibrium due to the warming of recent decades. The observed retreat of glaciers is not the full response to already realized warming (Christian et al., 2018), and they are committed to losing considerable mass in the future, even without further change in temperature (Jouvet et al., 2011; Mernild et al., 2013; Zekollari and Huybrechts, 2015; Zemp et al., 2015; Marzeion et al., 2018). As glaciers respond to changing climate, they act as integrators of the climate forcing (Roe et al., 2017) filtering out the inter-annual variability (Beedle et al., 2009). Although the point surface mass balance is a direct and immediate function of accumulation and ablation, the glacier geometry adjusts over a longer time to changed atmospheric conditions. This adjustment time is variable from one glacier to another. It depends strongly on the glacier geometry (thickness/vertical extent and slope/inclination) as well as the mass turnover and flow speed (Jóhannesson et al., 1989; Harrison et al., 2001; Lüthi, 2009). The range of adjustment time is from a few years, for smaller and steeper glaciers (Beedle et al., 2009; Lüthi and Bauder, 2010; Rabatel et al., 2013), to decades and up to hundreds of years for larger and low-pitched glaciers (e.g., Lüthi et al., 2010). As the glaciers retreat and become smaller, their sensitivity to climate change intensifies due to feedback mechanisms (Harrison, 2013). There is therefore *very high confidence* that climatic disequilibrium of glaciers will persist as warming continues.

9.5.1.1 Observed glacier extent and volume changes

The time series of measured mass changes of monitored glaciers in the world is presented in Chapter 2 (Section 2.3.2.3, Figure 2.21). Here we present regional patterns of mass changes. The data presented here come from the Global Terrestrial Network for Glaciers (GTN-G), in which the world is divided into 19 first-order GTN-G regions, also known as Randolph Glacier Inventory (RGI) regions, as shown in figure 9.1 and listed in Table 9.3. This assessment uses version 6.0 of the RGI (RGI Consortium, 2017b). The RGI has been gradually improving since AR5 due to a high volume of satellite data and the inclusion of detailed regional glacier inventories; the number of glaciers in the inventory has increased by 28%, often due to subdivision of larger ice caps and glaciers, but the total area covered by inventoried glaciers has decreased by 2.8% (compared to RGI 2.0 used in AR5). The total area, covered by the more than 215,000 glaciers with surface area in excess of 0.01 km² counted in RGI 6.0 (Table 9.3), is $(706 \pm 30) \cdot 10^3$ km² (*high confidence*). There is *medium confidence* in the total number of glaciers because it is subject to uncertainties in counting very small glaciers, automated methods to separate ice masses, and the area threshold above which glaciers are included in the inventory (Bahr and Radić, 2012; Parkes and Marzeion, 2018). Based on the worldwide glacier thickness measurement database (GlaThiDa 3.0; (GlaThiDa Consortium, 2019)), a new estimate for global glacier thicknesses, obtained using five numerical models, has been produced (Farinotti et al., 2019). Combined with the RGI 6.0 glacier area estimates, the total volume of glaciers in the world is 324 ± 84 mm potential sea level equivalent (SLE). This is about 21% lower global glacier SLE estimate than that given in AR5, because it takes into account that 15% of the total glacier volume is situated below sea level and will therefore not contribute to sea level rise if melted (Farinotti et al., 2019).

Figure 9.22 presents the regional glacier volume change since the beginning of the last century from different sources (Marzeion et al., 2015a, submitted; Bamber et al., 2018b; Dussaillant et al., 2019; Groh et al., 2019; Wouters et al., 2019; Zemp et al., 2019). [PLACEHOLDER FOR ADDITIONAL REGIONAL ESTIMATES] Analysis of these data reveals a large variability between regions, but a global retreat from the Little Ice Age maximum position is observed (Figure 9.22), indicating moderate decadal ice losses of a few tenths of m w.e. a⁻¹ (water equivalent per year) in the second half of 19th and the beginning of 20th century, followed by increased ice losses around 0.40 m w.e. a⁻¹ in the 1940s and 1950s, and then a period of moderate to little ice loss between mid-1960s and mid-1980s. The last decade of the 20th century and the first

decade of the 21st century exhibit the most negative mass balances since the beginning of the observational record (Zemp et al., 2015, 2019). The global average mass loss of glaciers in the period 2006-2016 is *very likely* -0.38 ± 0.16 m w.e. a⁻¹ (-274 ± 113 Gt a⁻¹) (Table 9.3). In the period 1971-2016 it is *likely* that glaciers contributed 25 ± 18 mm to sea-level rise. [PLACEHOLDER FOR ESTIMATE BASED ON NEWER PUBLICATIONS COMING]. Sources of uncertainty of the past glacier contribution to sea-level rise are missing (uncharted, because too small) glaciers and already disappeared glaciers. Because of an inherent lack of information, there is *low confidence* in the estimated contribution of these glaciers to sea-level rise of 17 to 48 mm to global sea level between 1901 and 2015 (Parkes and Marzeion, 2018). Multiple and consistent studies provide *very high confidence* that glacier retreat is an unprecedented global phenomenon since the beginning of the observational record in the second half of the 19th century (Roe et al., 2017; Zemp et al., 2019).

Regional anomalies do exist, most notably the “Karakoram anomaly” (Gardelle et al., 2012; Kääb et al., 2012, 2015; Sakai and Fujita, 2017; de Kok et al., 2018; Berthier and Brun, 2019; Bonekamp et al., 2019; Farinotti et al., 2020) centred on the Kunlun range at about 80°E/35°N on the north-western fringe of the Tibetan Plateau (Brun et al., 2017). Positive mass balance anomaly in Karakoram, dating back to 1970s (Bolch et al., 2017; Brun et al., 2017; Zhou et al., 2017), has been linked to atmospheric circulation variability leading to cooler regional conditions in summer (Forsythe et al., 2017). However, this anomaly is spatially very limited, negative mass balances being predominant in the remaining High Mountain Asia regions (Brun et al., 2017) and at further distance, for example in Iran (Karimi et al., 2015). In Western North America outside of Alaska, a fourfold increase in mass loss observed between 2000–2009 (2.9 ± 3.1 Gt yr⁻¹) and 2009–2018 (12.3 ± 4.6 Gt yr⁻¹) has been attributed to a shift in regional meteorological conditions driven by the location and strength of upper level zonal wind (Menounos et al., 2019). Conversely, the impact of volcanoes on glacier retreat in South America and Iceland is not significant (Rivera et al., 2012; Björnsson et al., 2013; Reinthaler et al., 2019). Although the Southern Patagonia Icefield has had mass loss since at least 2000 of 0.96 ± 0.29 m w.e. a⁻¹ (Dussaillant et al., 2019), the biggest glacier of the Southern Patagonia Icefield, Pio XI, is at present advancing, thickening and pushing moraines (Wilson et al., 2016) with a positive mass balance for the last 20 years of 0.41 ± 0.07 m w.e. a⁻¹ (Braun et al., 2019; Dussaillant et al., 2019). [PLACEHOLDER FOR FURTHER REGIONAL ANOMALIES, CANADIAN ARCTIC, GREENLAND AND ICELAND REGIONS HAVE HAD NEAR BALANCE IN 2017, 2018, BUT BACK TO NEGATIVE IN 2019, NZ SITUATION, OTHERS]

[START TABLE 9.3 HERE]

Table 9.3: Number of glaciers, area, glacier volume (potential sea level equivalence), recent glacier mass change and future contribution of glaciers to sea level rise, globally and differentiated for the 19 regions used throughout the chapter. Number of glaciers and extent (glacierized area) are derived from the RGI 6.0 (RGI Consortium, 2017b). Glacier volume is derived from Farinotti et al. (2019), which is based on RGI 6.0. Recent (2006-2016) mass change is based on SROCC Chapter 2 Table 2A.1, except those marked by *1, which are derived from (Zemp et al., 2019) and *2 which are derived from (Dussaillant et al., 2019). Total projected sea level rise contribution between 2015-2100 is derived from GlacierMIP Phase 2 results (Marzeion et al., submitted). See Figure 9.22 for time series of changes in each region.

Region	OI	Num ber of Glaci ers	Glacieriz ed area (km ²)	Glacier volume		Glacier mass balance change 2006-2016		Total projected glacier change between 2016-2100 (mm sle)			
				(mm sle)	Uncert ainty	(mm sle)	Uncert ainty	RCP 2.6	Unc ertai nty	RCP 8.5	Uncerta inty
World	0	2155 47	705738.8	324.3	84.1	-7.58	3.09	81.3	16.4	162.0	24.0

Alaska	1	2710 8	86725.1	43.3	11.2	-1.70	0.40	13.2	6.6	23.6	9.1
Western Canada and US	2	1885 5	14524.2	2.6	0.7	-0.20	0.40	1.5	0.5	2.1	0.4
Arctic Canada North	3	4556	105110.6	64.8	16.8	-1.10	0.20	11.5	6.2	25.5	12.0
Arctic Canada South	4	7415	40888.2	20.5	5.3	-0.90	0.30	5.6	3.5	10.8	4.4
Greenland Periphery	5	1930 6	89717.1	33.6	8.7	-1.30	0.40	8.4	6.0	17.2	6.6
Iceland	6	568	11059.7	9.1	2.4	-0.20	0.10	2.5	1.7	4.2	2.1
Svalbard	7	1615	33958.9	17.3	4.5	-0.2	0.10	8.3	4.5	14.4	5.1
Scandinavi a	8	3417	2949.1	0.7	0.2	-0.06	0.03	0.3	0.1	0.5	0.1
Russian Arctic	9	1069	51591.6	32	8.3	-0.40	0.08	10.2	5.0	20.8	7.5
North Asia	10	5151	2410.1	0.3	0.1	-0.03	0.03	0.2	0.1	0.3	0.1
Central Europe	11	3927	2092.1	0.3	0.1	-0.06	0.01	0.2	0.1	0.3	0.0
Caucasus and Middle East	12	1888	1307.0	0.2	0	-0.03	0.03	0.1	0.0	0.1	0.0
Central Asia	13	5442 9	49303.4	7.9	2	- 0.19 ^{*1}	0.17 ^{*1}	3.8	1.6	6.9	2.4
South Asia West	14	2798 8	33568.3	6.9	1.8	- 0.03 ^{*1}	0.11 ^{*1}	2.9	1.2	5.3	1.6
South Asia East	15	1311 9	14734.2	2.1	0.5	- 0.14 ^{*1}	0.06 ^{*1}	1.0	0.6	1.8	0.7
Low Latitudes	16	2939	2341.0	0.2	0.1	- 0.03 ^{*2}	0.03 ^{*2}	0.1	0.1	0.2	0.1
Southern Andes	17	1590 8	29429.1	12.8	3.3	- 0.70 ^{*2}	0.10 ^{*2}	3.3	2.1	6.3	2.2
New Zealand	18	3537	1161.8	0.2	0	-0.03	0.03	0.1	0.0	0.1	0.0

Antarctic and Subantarctic	19	2752	132867.2	69.4	18	-0.30	3.00	8.1	9.0	21.5	14.5
----------------------------------	----	------	----------	------	----	-------	------	-----	-----	------	------

[END TABLE 9.3 HERE]

A range of natural and anthropogenic factors and processes modulate surface climate-induced glacier melt and retreat, many of them resulting in widespread glacier shrinkage. Collapse of mountain glaciers (Gilbert et al., 2015; Kääb et al., 2018) is an extreme example of processes enhancing glacier response to climatic forcing. Increasingly thick debris cover on retreating glaciers, particularly in the lower ablation zone, can slow down glacier melt, and reduce their sensitivity to climate (Pratap et al., 2015; Brun et al., 2016). Recently, supraglacial debris cover extent has been assessed globally (Scherler et al., 2018), although only less than 5% of glacier area is covered by debris, it is highly heterogeneous, with low-latitude high mountain areas (e.g. South-Central Asia, Alaska, Central Southern Andes, Caucasus and Western North America) exhibiting the highest proportion of debris-covered glacier surface. Deposition of light-absorbing particles on snow, in particular black carbon or dust, as well as development of snow algae, can lead to enhanced surface melt (Ginot et al., 2014; Ménégoz et al., 2014; Gabbi et al., 2015; Lutz et al., 2016; Wittmann et al., 2017; Magalhães et al., 2019). With warming, the refreezing capacity of firn gradually decreases, which results in larger increase in runoff than in meltwater production, as less meltwater is refrozen, as observed on the peripheral glaciers in Greenland (Noël et al., 2017). Other processes, such as solar radiation absorption in superficial ponds on debris covered glaciers and ice-cliff wasting have been shown to lead to substantially enhanced glacier melt rates on debris-covered glaciers (e.g., Brun et al., 2016). Tidewater and lake terminating glaciers lose mass through calving, the rate of which can be substantially amplified by warming ocean/lake temperatures (e.g. (Luckman et al., 2015; Minowa et al., 2017)) and increased undermining of calving fronts driven by enhanced submarine meltwater runoff (O’Leary and Christoffersen, 2013; Lea et al., 2014a; Fried et al., 2018). However, at least for the Canadian Arctic marine-terminating glaciers, surface air temperature change seems to be the main driver of recent retreat (Cook et al., 2019a). Improved understanding and quantification of the effect that these processes have on glaciers retreat lead to *high confidence* that the main driver of observed global glacier retreat is changing climate.

The current global retreat of glaciers appears unusual during the post-glacial period: it occurs worldwide with very few regional exceptions, at a very high rate, during an interval of orbital forcing favourable for glacier growth, rather than degradation, and despite of various regional trends in precipitation regime (Solomina et al., 2015, 2016, and references therein). High resolution reconstructions of glaciers variations in the Mont Blanc and Grindelwald regions in the European Alps show that now the glaciers are smaller than ever since the mid-16th century (Nussbaumer and Zumbühl, 2012). In Norway, archaeological artefacts were continuously covered with ice since 9th century CE (Nesje et al., 2012), implying that glaciers in the last millennium were larger than in early 21st century. Exposure of plants emerging from beneath glaciers confirm that glaciers are retreating from areas that have been covered with glaciers for at least 1300 years in Svalbard (Miller et al., 2017) and 40,000 years in Canada (Pendleton et al., 2019). Koch et al. (2014) provided numerous examples of dramatic loss of snow and ice in mountains that occurred in the last decades in the Western Canada, as well as in Europe and in the tropics. They demonstrated that in many regions the glaciers and snow patches are smaller now than at any time in the past several thousand years. Although the millennial glacier length variations records are incomplete and discontinuous (the advancing glaciers are constantly overriding previous geomorphic and sedimentological evidences), and glacier fluctuations depend on multiple factors (temperature, precipitation, topography, internal glacial dynamic etc.), the observed almost uniform global glacier retreat implies a common climatic signal driving this process. The studies above and statistical confirmation that current retreat can only happen with anthropogenic climate change (Roe et al., 2016) support the previous IPCC conclusions (AR4 and AR5) with *high confidence* that the character of the global glacier retreat in the 20th through early 21st centuries is exceptional within the current interglacial.

Glacier retreat during the last deglaciation (from 19 to 11.5 ka) in mid-to-low-latitudes is linked to increasing greenhouse gas concentration, based on numerous cosmogenic-nuclide production-rate calibrations and transient climate model simulations (Shakun et al., 2015), although at the regional level other forcings, such as insolation, ice sheets and ocean circulation, modulated the glacier responses (the glacial-interglacial temperature timeseries (Cross-Chapter Box 2.1, Figure 1), observations of orbital and greenhouse gas forcings (Section 2.2) and glacial-interglacial and Holocene changes in the carbon cycle (Section 5.1.3) are discussed in other chapters). The variability of glacier extents on millennial time scales in the Holocene shows a certain correlation with orbital forcing: in the Northern Hemisphere the size of the glaciers was generally increasing over the Holocene (Briner et al., 2016; Kaufman et al., 2016; Lecavalier et al., 2017; Zhang et al., 2017; Axford et al., 2019; Geirsdóttir et al., 2019; Larsen et al., 2019) while the glaciers in Patagonia, New Zealand and in the tropics were generally decreasing (Solomina et al., 2015, and references therein). In Central Asia (Murari et al., 2014; Owen and Dortch, 2014; Batbaatar et al., 2018; Saha et al., 2018), in southern South America (Kaplan et al., 2016), and in some other regions glacier fluctuations were more strongly controlled by regional atmospheric circulation patterns. Solar and volcanic activity supported by positive feedbacks in the climate system were forcing glacier variability at the decadal and multidecadal frequencies, though no single mechanism of glacier fluctuations over the Holocene on a global scale is identified (Solomina et al., 2015, 2016; Brönnimann et al., 2019). In summary, there is *high confidence* that greenhouse gas concentration and orbital forcing are the main drivers of glacier extent on millennial time scale.

Using the Open Global Glacier Model and a selection of six climate model outputs for the period 850–2004 CE (Parkes and Goosse, 2019) conclude that temperature explains more variance in glacier lengths than precipitation, however in some regions snowfall increase may counterbalance warming. The combined effect of precipitation and temperature may induce significant interregional variability, during the last deglaciation, during the Holocene and for future projections. In certain regions, such as the Tropical Andes (e.g., Martin et al., 2018), glaciers stayed close to their maximum local position until the end of the Heinrich 1 event, about 15 ka, because the precipitation increase between 17 and 15 ka compensated for the synchronous post-LGM warming. Batbaatar et al. (2018) identified a local early Holocene glacier maximum at 7–8 ka due to the increase in precipitation in arid Mongolia (Gichginii Range) where the glaciers are located at high altitudes with low precipitation and sub-freezing sublimation regimes. Modelling experiments show that the retreat of Greenland Ice Sheet in the Kangerlussuaq region (West Greenland) would be more rapid during the Holocene thermal maximum without elevated precipitation (Downs et al., 2019). Increased precipitation is largely responsible for the last phase of the Little Ice Age advance in the mid-19th century of Alpine glaciers (Brönnimann et al., 2019). There is *high confidence* that combined effect of precipitation and temperature induces interregional variability in glacier retreat.

[START FIGURE 9.22 HERE]

Figure 9.22: Modelled, observed and project volume change of glaciers in the 19 GTN-G regions from 1902 to 2100 (see Figure 9.1 for location) relative to the volume in year 2016. Regions are ordered from largest mass loss in 21st century in top left and the smallest mass loss in bottom right. Light brown lines and shaded areas indicate the regional glacier change through the 20th century and its respective uncertainty given by Marzeion et al. (2015b). Black lines and grey shaded areas indicate the regional glacier mass change between 1961 to 2016 and its respective uncertainty obtained by Zemp et al. (2019). Green lines and shaded areas (Region 16 and 19 [Placeholder for more regional studies]) indicate regional estimates where available (Dussailant et al., 2019). Two scenarios, RCP2.6 and RCP8.5 from GlacierMIP Phase 2 (Marzeion et al., submitted), are shown. See Table 9.3 for projected mass change in each region. For a better comparison between regions, the maximum relative volume was set to 200%, although for some regions the volume changes between 1902 and 2016 exceeded that value. For the low latitude, New Zealand and South Asia (East) glaciers the changes were larger than 1000%, 350%, and 250%, respectively.

[END FIGURE 9.22 HERE]

9.5.1.2 Model evaluation

Global scale modelling is challenging because of a scarcity of accurate glacier thickness data for model initialisation and calibration. With improved inventories (RGI Consortium, 2017a; GlaThiDa Consortium, 2019) and modelling approaches (Farinotti et al., 2019), the situation is improving, although large uncertainties remain. Due to the coordinated efforts of the global glacier model intercomparison project (GlacierMIP) that harmonizes both input and forcing data (Hock et al., 2019; Marzeion et al., submitted), rapid progress is being made in projecting the future of glaciers in the world. In Phase 1 (Hock et al., 2019), 6 models contributed with 214 simulations, whereas in Phase 2 (Marzeion et al., submitted), 11 models with 279 simulations contributed to the intercomparison. SROCC stated with *high confidence* that glaciers will lose mass until the end of the century, but expressed *medium confidence* in the magnitude and timing, due to relatively simple models in the intercomparison study (Hock et al., 2019). However, compared to simpler, empirical parameterizations, full energy-balance models are not necessarily the most appropriate choice for simulating future glacier response to climate change even at the local scale (Réveillet et al., 2017, 2018) because of parameter and forcing uncertainties. The global surface mass balance (SMB) models applied in GlacierMIP are of varying complexity, including mass balance sensitivity approaches (Slangen et al., 2012), temperature-index methods (Marzeion et al., 2012; Hirabayashi et al., 2013; Radić et al., 2014; Huss and Hock, 2015; Maussion et al., 2018; Zekollari et al., 2019), simplified energy balance calculations (Giesen and Oerlemans, 2013), and inclusion of glacierized surfaces at multiple heights within a single grid cell in the Joint UK Land Environment Simulator (JULES (Shannon et al., 2019)). All models account for glacier retreat and advance (Hock et al., 2019), but only two models that include glacier dynamics (Maussion et al., 2019; Zekollari et al., 2019) contributed to GlacierMIP Phase 2 (Marzeion et al., submitted).

Secondary processes such as debris-cover thickening, albedo changes due to dust and snow algae development, as discussed in Section 9.5.1.1, are not represented in the global glacier models, resulting in both underestimated and overestimated sensitivity to warming, and induce complexity and uncertainty in attribution studies. Furthermore, the low resolution and high spatial variability at subgrid scale of the precipitation amount requires downscaling to the spatial scale of a glacier, and model uncertainties persist due to the SMB and ice dynamic models used, and due to uncertainties in the thickness of the modelled glaciers. Nevertheless, the reported progress in global scale modelling efforts allows *medium confidence* in the capacity of current-generation glacier models to simulate reasonably the magnitude and timing of mass loss. Although formal attribution of glacier changes is currently not possible, the assessment in Chapter 3 is that anthropogenic forcings *very likely* contributed to the observed retreat of glaciers.

9.5.1.3 Projections

The projected mass loss for the 19 GTN-G regions is shown in Figure 9.22, which displays results from GlacierMIP Phase 2 (Marzeion et al., submitted) and regional projections where available. Regional projections have been made for western Canada (Clarke et al., 2015), high mountain Asia (Kraaijenbrink et al., 2017) and European Alps (Zekollari et al., 2019) and a number of individual glacier projections have been developed (Ziemen et al., 2016; Gilbert et al., 2017; Zekollari et al., 2017; Schmidt et al., 2019). The projected mass loss under both RCP2.6 and RCP8.5 for global glacier mass until 2100 is substantial in all regions and all models agree that a few regions (Low latitudes, Central Europe, Caucasus and Middle East, Western Canada and US, North Asia and Scandinavia) will lose most if not all glacier mass by the end of the century under RCP8.5. There is, however, a large spread in the projections and they all suffer from the uncertainties discussed above. The greatest source of uncertainty in glacier contribution to sea level rise until the middle of the century is disagreement between glacier models and the emission pathway is the dominant uncertainty for how much glacier mass will remain at the end of the 21st century (Marzeion et al., submitted). The projected global glacier mass loss by 2100 [PLACEHOLDER FOR UPDATES WHEN PAPER IS ACCEPTED: is 17% (75 ± 64 mm SLE) and 38% (165 ± 98 mm SLE) for RCP2.6 and RCP8.5, respectively]. Within error bars, this agrees with the former estimation of GlacierMIP Phase 1 (Hock et al., 2019), although it indicates a slightly smaller mass loss due to inclusion of two new models with lower overall climate sensitivity than previous model ensemble (Marzeion et al., submitted). Figure 9.22 and Table 9.3 show time series and mass change, respectively, in each region. [PLACEHOLDER: SECTION 9.6.3.2

CURRENTLY USES THE RESULTS FROM GLACIERMIP PHASE 1 TO UPDATE THE PARAMETRIC REPRESENTATION OF GLACIER MASS LOSS AS A POWER FUNCTION OF CUMULATIVE TEMPERATURE THAT WAS EMPLOYED IN AR5 CALCULATE THE GLACIER CONTRIBUTION FOR THE SEA LEVEL PROJECTIONS PRESENTED IN 9.6.3.3. IN THE FINAL DRAFT, ASSESSMENT OF CONTRIBUTION OF GLACIERS TO SEA LEVEL WILL INCLUDE THE EMULATED GLACIERMIP PHASE 2 RESULTS (Edwards et al., submitted)].

The largest contribution to sea level rise comes from the Arctic and Antarctic regions (67% of the total). The regions with large glacier mass have the smallest relative loss, but large contribution to sea level rise. The regions with glaciers that adjust quickly to climate change and small total glacier volume are the regions that will lose nearly all (>85%) glacier mass by 2100 in the RCP 8.5 scenario. The mass loss rates vary between regions and there are distinctively different patterns between the RCPs. The rates range from 0.5 mm SLE yr⁻¹ (RCP4.5) to 3.0 mm SLE yr⁻¹ (RCP8.5) for the period 2015 – 2100 with maximum occurring at the end of the 21st century. The peak in mass loss rate followed by reduction is due to both decreasing glacier volume and stabilizing mass balance (under RCP2.6) (Marzeion et al., submitted).

As expressed in Section 9.5.1.2, there is *medium confidence* in the projected rates of mass loss. However, there is *high confidence* in some important consistent and already currently observed features of the projections, most notably that (1) the projected glacier mass loss will be largely of global extent, (2) the largest glacierized regions will provide a largest contribution to future SLR, (3) regions with a dominant population of small glaciers will lose most if not all of their glacier mass by the end of the century, although the SLR contribution from these regions is negligible, (4) after about 2050, loss rates will depend on the emission scenario. Two models that have run longer simulations (Levermann et al., 2013) show that under the RCP2.6 scenario, 10-60% of the glacier volume will remain in polar regions, although all low latitude and small glaciers will melt and for the RCP8.5 scenario only the largest glaciers will survive at sizes that are less than 10-20% of their present size (Marzeion et al., 2012).

As discussed in more detail in Chapter 8, the projected changes in glacial meltwater runoff vary greatly. It is assessed in SROCC that average winter runoff is expected to increase in mountain regions (*high confidence*) and spring peak maxima will occur earlier (*very high confidence*) due to increase in air temperature and phase change of precipitation. Although observed and projected trends in annual runoff vary substantially among regions and can even be opposite in sign, there is *high confidence* that average annual runoff from glaciers will have reached a peak, with declining runoff thereafter, at the latest by the end of the 21st century in most regions.

9.5.2 Permafrost

This section assesses current understanding of the physical aspects of permafrost as an element of the climate system, focusing on recent developments since the publication of AR5 and SROCC. Observed large-scale permafrost changes are summarized in Chapter 2 (Section 2.3.2.5) and the permafrost carbon feedback is assessed in Chapter 5.

9.5.2.1 Observed changes in permafrost extent, temperature and active layer thickness

According to a recent observationally based model analysis (Obu et al., 2019) which is in good agreement with previous estimates (Zhang et al., 1999; Gruber, 2012), permafrost (perennially frozen ground) underlies about 22% of the Northern Hemisphere land and covers more than 50% areal fraction of unglaciated land north of 60°N. It is also found in many high-altitude areas of mountain ranges in both hemispheres, representing about 27-29% of the global permafrost area (*medium confidence*, see SROCC) and unglaciated areas in Antarctica. The global permafrost extent is about $22 \pm 3 \times 10^6$ km² (Gruber, 2012) (*medium confidence*). The Earth's perennial ground ice volume is estimated to be equivalent to 0.027 to 0.088 m of global sea level rise (Zhang et al., 2000) (*low confidence* because of limited evidence).

There are consistent, but geographically unevenly distributed reports of observed loss of permafrost in areas of discontinuous and sporadic permafrost since about 1980 (Camill, 2005; James et al., 2013; Jones et al., 2016a; e.g., Borge et al., 2017; Chasmer and Hopkinson, 2017; Gibson et al., 2018). Since observations are geographically uneven, there is only *medium confidence* that permafrost loss is a pan-Arctic phenomenon, although there has been a widespread Arctic warming in recent decades. In spite of increasing evidence of landscape changes from site studies and remote sensing, quantifying permafrost extent change in the Arctic remains challenging (e.g., Jorgenson and Grosse, 2016). Paleoclimate studies (Treat and Jones, 2018) suggest a strong sensitivity of permafrost extent to climatic variations over the Holocene in the Northern Hemisphere (*medium confidence*) and report widespread northern peatland thaw since about 250 years ago. Analysis of Siberian speleothem records over the last 500,000 years (Vaks et al., 2013) suggests that climatic conditions only about 1.5°C warmer than the preindustrial could, at equilibrium, lead to extensive permafrost loss in Central Siberia, with permafrost-free conditions north of 60°N (*low confidence* because of limited evidence).

Average permafrost borehole temperatures have been increasing across the Arctic permafrost region, with few exceptions, since the 1980s, reaching record high temperatures at many long-term monitoring sites (*very high confidence*, see SROCC), and given the substantial number of boreholes and large geographical coverage (Romanovsky et al., 2018), there is *high confidence* that this is a pan-Arctic phenomenon. Between 2007 and 2016, the global permafrost temperature has increased at $0.29 \pm 0.12^{\circ}\text{C}$ near the depth of zero annual amplitude (*medium confidence*) (Biskaborn et al., 2019). Stronger temperature trends dominate in the continuous permafrost zone ($0.39 \pm 0.15^{\circ}\text{C}$) than in the discontinuous zone ($0.20 \pm 0.10^{\circ}\text{C}$) (*medium confidence* because of substantial inter-site variability). Except in Interior Alaska, Arctic permafrost temperatures have continued to increase, particularly at cold sites (below -4°C or so), broadly in agreement with strong positive Arctic air temperature trends and partially modulated by snow thickness changes. Warmer sites, with permafrost temperatures closer to 0°C , show weaker positive trends because of the large amount of energy required for melting the ice (Figure 9.23), and reduced effect of the Arctic amplification in more southerly locations (Romanovsky et al., 2017). Mountain permafrost temperature trends are rather heterogeneous, reflecting variations in local conditions, including topography, surface type and snow cover (Noetzli et al., 2019). The observed mountain permafrost warming rates tend to be weakest at warm sites close to 0°C , as in the Arctic, and again because of ice-water phase changes (Mollaret et al., 2019; PERMOS, 2019). Biskaborn et al. (2019) estimate an average permafrost warming trend in mountain zones of $0.19 \pm 0.05^{\circ}\text{C}$ between 2007 and 2016 (*medium confidence*). In summary, because of relative sparse geographic coverage in mountain areas and ice-free areas of the Antarctic, there is *medium confidence* that permafrost warming is a global phenomenon (Biskaborn et al., 2019).

[START FIGURE 9.23 HERE]

Figure 9.23: Observed recent rate of terrestrial Arctic permafrost temperature change ($^{\circ}\text{C}/\text{decade}$) (updated from (Romanovsky et al., 2018)) as a function of local average permafrost temperature ($^{\circ}\text{C}$). Coastal regions are identified by the name of the adjacent sea (Baffin Bay/Davis Strait, Barents and Beaufort Seas).

[END FIGURE 9.23 HERE]

Closer to the surface, the annual maximum thaw depth defines the active layer thickness (ALT), which responds more strongly to short-term climate variations and can therefore exhibit substantial interannual variability. Syntheses of Northern Hemisphere ALT change from 1990 to 2015 (Luo et al., 2016) and of pan-Arctic ALT change extending to 2017 (Romanovsky et al., 2018) report widespread, but non-uniform, ALT increases across the Northern Hemisphere permafrost zone. However, less than half of the observed ALT trends are significant at the 5% level. A clear global picture has not yet emerged because of (1) uneven distribution of observing sites, (2) substantial variability among the existing sites, strongly influenced by local conditions (soil constituents and moisture, snow cover, vegetation) in addition to surface temperature as primary driver, and (3) thaw settlement in ice rich terrain (Streletskiy et al., 2017). As assessed in SROCC, it can only be stated with *medium confidence* that ALT increase is a pan-Arctic phenomenon. Records of

Antarctic ALT changes are too limited for an assessment of long-term trends (Hrbáček et al., 2018). Documented increased ALT in European Alps, Scandinavia, Tibetan Plateau, and Andes during the past decade (Christiansen et al., 2010; Wu et al., 2015; PERMOS, 2019; Trombotto-Liaudat and Bottegal, 2019) indicates widespread mountain permafrost degradation (*medium confidence*). Although most pieces of evidence are restricted to the European Alps, there is *high agreement* that acceleration (Scapozza et al., 2014; Kaufmann, 2016; Kellerer-Pirklbauer and Kaufmann, 2018; PERMOS, 2019) and destabilization, including collapse, of rock glaciers are indicators of mountain permafrost degradation (*medium confidence*) related to warming temperatures and increase of water content (Deline et al., 2015; Cicoira et al., 2019; Marcer et al., 2019).

As summarized in SROCC, consistent post-AR5 evidence from observations (e.g., Portnov et al., 2013; Ruppel et al., 2016) and modelling (Angelopoulos et al., 2019) suggests that the extent of subsea permafrost formed before submersion on Arctic continental shelves is much reduced compared to older studies that had estimated that the entire formerly exposed Arctic shelf area was underlain by permafrost (*high confidence*). This is supported by observations (Shakhova et al., 2017) that show rapid thaw of recently submerged permafrost on the East Siberian Shelf (see also Section 12.4.8.5).

There is a lack of formal studies attributing observed permafrost changes (thaw depth, thermal state) or associated landscape changes to human activity. However, the observed Arctic warming has been attributed to human influence (e.g., Najafi et al., 2015) and an obvious physical link exists between ground temperatures (and thus permafrost) and surface air temperatures. Therefore, there is *medium confidence* (because of physical consistency and convergent evidence, but an absence of formal attribution studies) that human influence is overall the dominant cause of the observed pan-Arctic permafrost changes.

9.5.2.2 Model evaluation

Coupled models contributing to CMIP5 showed large inter-model variability of permafrost extent due to deficiencies in reproducing surface characteristics and processes (Koven et al., 2013), particularly thermal properties of the ground and snow, leading to several new developments since AR5. Soil freezing and its thermal and hydrological effects are now included in a large number of land-surface modules that are part of the CMIP6 ensemble (Chadburn et al., 2015; Hagemann et al., 2016; Cuntz and Haverd, 2018; Guimberteau et al., 2018), sometimes allowing for the effects of excess ice (Lee et al., 2014). Improved representation of snow insulation in models has led to more realistic simulated permafrost extents (e.g., Paquin and Sushama, 2014); in a post-CMIP5 ensemble of land-surface models driven by observed meteorological conditions (McGuire et al., 2016), inter-model spread was substantially reduced when the ensemble was restricted to models that correctly represented the effect of snow insulation on the underlying soil (Wang et al., 2016b). More detailed descriptions of high-latitude vegetation characteristics, vegetation dynamics, and snow-vegetation interactions were included in several models (e.g., Lawrence and Swenson 2011; Chadburn et al. 2015b; Porada et al. 2016; Druehl et al. 2017).

Despite these improvements, several types of abrupt physical disturbances, in particular fire and thermokarst formation (thaw of ice-rich permafrost leading to soil subsidence and erosion, further enhancing thaw), are usually not represented in coupled climate models. As discussed in detail and assessed in SROCC, there is *high confidence* that permafrost degradation through fire (Jones et al., 2015a; Gibson et al., 2018) is currently occurring faster than historical observations suggest, and *medium confidence* that thermokarst formation, to which about 20% of the northern permafrost region is vulnerable (Olefeldt et al., 2016), has the potential to accelerate large-scale permafrost decay in response to climate change. Another process that can accelerate permafrost thaw is strong thermal insulation by organic material which creates a positive feedback loop, usually not represented in the current (CMIP6) generation of climate models, whereby permafrost thaw leads to mineralization of the organic material, reducing the thermal insulation of frozen ground from the warm surface in summer (Zhu et al., 2019). In summary, because current-generation climate models tend to neglect several abrupt physical disturbances impacting frozen ground, it is possible that permafrost decay rates simulated in these models in response to climate change are currently underestimated (*low confidence*).

In terms of simulated permafrost extent, output from available CMIP6 models shows no substantial improvement with respect to CMIP5 (see Figure 9.24a). Compared to observations and reanalysis-based estimates of present permafrost extent (Zhang et al., 1999; Gruber, 2012; Obu et al., 2019), the extent of the region where perennially frozen soil is simulated within the top 15 m below the surface in the Northern Hemisphere for the 1979-1998 (corresponding approximately to the average period of the observational estimates) is characterized by a very large scatter in the coupled CMIP5 and CMIP6 historical simulations, possibly slightly reduced in CMIP6 with respect to CMIP5. Outliers with very low simulated permafrost extent are models that have only a very shallow soil column and, at the same time, do not take into account soil water phase changes. This leads to an overestimate of seasonal thaw depth, exceeding the total thickness of the models' soil column (Burke et al., submitted). Excessive simulated permafrost extent can in several cases be traced to insufficient thermal insulation by the winter snow cover (Burke et al., submitted). Figure 9.24a also shows that the corresponding AMIP land-atmosphere simulations, and the LS3MIP (van den Hurk et al., 2016) land-only simulations with prescribed reanalysis-based meteorological forcing, do not provide an improved simulation of the current permafrost extent, although, by construction, they can be expected to exhibit lower land surface climate biases. This further points to deficiencies in the land modules as the main reason for biases, consistent with conclusions drawn from the analysis of CMIP5 output (Koven et al., 2013).

[START FIGURE 9.24 HERE]

Figure 9.24: a) Diagnosed NH permafrost extent (area with perennially frozen ground at 15 m depth, or at the deepest model soil level if this is above 15 m) for 1979-1998, for available CMIP5 and CMIP6 models, from the first ensemble member of the historical coupled run, and for CMIP6 Amip (atmosphere+land surface, ocean fixed) and LS3MIP land-hist (land only, atmosphere prescribed) runs. Observed and reanalysis-based current permafrost extents are indicated as black symbols (triangle: Obu et al. (2018); star: Zhang et al. (1999); circle: central value and associated range from Gruber (2012)). b) Simulated global permafrost volume change between the surface and 3 m depth as a function of the simulated GSAT change, from the first ensemble members of a selection of scenarios, for available CMIP6 models.

[END FIGURE 9.24 HERE]

9.5.2.3 Projections

Constraining CMIP5 output with diagnosed relationships between the observed present-day spatial distribution of permafrost and air temperature allowed to infer an equilibrium sensitivity of permafrost loss to stabilized global mean warming of about $4.0 \times 10^6 \text{ km}^2 \text{ }^\circ\text{C}^{-1}$ (Chadburn et al., 2017). This equilibrium permafrost sensitivity is about 20% higher than the transient centennial-scale near-surface permafrost sensitivity (diagnosed from seasonal thaw down to 3 m depth) suggested by direct analysis of CMIP5 output (Slater and Lawrence, 2013).

In spite of the large spread of simulated permafrost extent in the top 15 m, the available CMIP6 models consistently suggest a strong temperature sensitivity of the volume of perennially frozen soil in the top 3 m below the surface (see Figure 9.24b). The sensitivity of the modelled permafrost volume in the top 3 m to global surface air temperature (GSAT) changes (with respect to the 1995-2014 average and up to $+3^\circ\text{C}$ change, that is, about up to $+4^\circ\text{C}$ with respect to preindustrial levels) is about $(25 \pm 5)\%/^\circ\text{C}$ (Burke et al., submitted), but there can be only *medium confidence* in this range because of the model deficiencies discussed above. The temperature sensitivity of permafrost volume is slightly stronger in the strong-mitigation SSP1-2.6 scenario than in the less mitigated ones because subsurface temperature lag increases with higher atmospheric warming rates, particularly when soil ice melting induces additional delays.

In summary, based on high agreement across CMIP6 and older model projections (see AR5), fundamental process understanding, and paleoclimate evidence, it appears *virtually certain* that near-surface permafrost extent will shrink as global climate warms.

New permafrost will generally form after disturbance or transient warming, particularly in the continuous permafrost zones (Briggs et al., 2014a). Irreversibility of permafrost loss on large spatial scales on the centennial time scale is therefore not to be expected (*high confidence*). Potential abruptness and irreversibility of permafrost carbon loss is assessed in Chapter 5.

9.5.3 Snow

9.5.3.1 Observed changes

Northern Hemisphere mean snow cover extent in January and February, the usual months of maximum extent, corresponds to about 45% of the Northern Hemisphere land surface (in excess of 45 million km² over the 1967-2014 period (Estilow et al., 2015)). In contrast, maximum seasonal snow cover in South America, the dominant ice-free land mass in Southern Hemisphere in terms of seasonal snow cover extent, remains below 1 million km² (Foster et al., 2009). Terrestrial snow cover is most frequently characterized via three main variables: how much area is covered by snow (snow cover extent - SCE), how long snow continuously remains on the land surface (snow cover duration (SCD) or snow-on and snow-off dates) and how much water is stored in solid or liquid form by the snowpack (snow water equivalent - SWE). In addition, snow depth, linked to SWE, is sometimes reported in observational studies. Observed large-scale snow cover changes, their attribution to human activity and their effects on the hydrological cycle are also discussed in Chapters 2, 3 and 8 of this report.

Satellite-derived estimates of SCE compiled within the NOAA snow chart Climate Data Record (NOAA CDR) extend back to 1967, providing one of the longest environmental data records from spaceborne measurements (Estilow et al., 2015). Continental trends from these coarse resolution estimates (~200 km) show declining snow cover during the spring period, consistent with surface warming (Hernández-Henríquez et al., 2015; Mudryk et al., 2017). Negative monthly trends in SCE are also captured by other datasets (Figure 9.25a and b) over the 1981-2010 period used for inter-comparison (Mudryk et al., submitted). On the hemispheric scale, SCE reductions during that period in the combined observational record are strongest during the transitional seasons (early winter and spring). Studies consistently report low uncertainty on the magnitude of May SCE trends of -3 to -5% per decade (*high confidence*) and higher uncertainties on the magnitude of June SCE trends because of a large inter-dataset spread (-5% to -16% per decade, *medium confidence*) (Hori et al., 2017; Mudryk et al., 2017). The loss of spring SCE is also reflected in earlier spring snow melt, derived from surface observations (Bulygina et al., 2011; Brown et al., 2017), satellite data (Wang et al., 2013; Estilow et al., 2015), and model-based analyses (Liston and Hiemstra, 2011). There is considerable inter-dataset variability, but all trends of spring snow-off dates from all datasets are negative (Brown et al., 2017). There is therefore *very high confidence* that the northern hemisphere spring SCE has been decreasing since 1978, and based on the NOAA record only, there is *medium confidence* that this trend extends back to 1967.

There is consistent evidence and thus *high confidence* that changes in SCE can be directly related to extratropical temperature increases (Brutel-Vuilmet et al., 2013; Thackeray et al., 2016; Mudryk et al., 2017). Consistent between multiple observational products and historical climate model simulations, the observed northern hemisphere spring SCE sensitivity to northern hemisphere land (>30°N) warming (Mudryk et al., 2017) is approximately $-1.9 \times 10^6 \text{ km}^2/\text{°C}$ (95% confidence range of $\pm 0.9 \times 10^6 \text{ km}^2/\text{°C}$).

Reconstructions with simple snow models have extended the satellite record as far back as 1915 (Brown and Robinson, 2011), putting the satellite era in historical context. There is *medium confidence (limited evidence)* in this extension that shows that overall, Northern Hemisphere spring snow cover extent has been declining since the 1920s, particularly in Eurasia, with accelerating decline since about 1970. Analysis of paleoclimate records (Pederson et al., 2011; Belmecheri et al., 2016) suggests that recent snowpack reductions in western North America are exceptional on a millennial timescale (*medium confidence*).

The coarse spatial resolution of the NOAA-CDR means it is inappropriate for mountain regions: anomalies from the NOAA CDR are not always consistent with the snowline elevation anomalies derived from 500 m

resolution MODIS imagery (Bormann et al., 2018). While moderate resolution optical sensors with wide swath coverage such as MODIS can provide appropriate measurements for determining SCE in mountainous regions (Riggs et al., 2017), these are short records for climate applications (~20 years). Longer time series are available from older-generation sensors such as AVHRR, for example over the 1985–2011 period for the European Alps (Hüsler et al., 2014). SROCC reported *high confidence* in a general decline (with regional variations) of mountain SCD since the middle of the 20th century (Hock et al., 2019) and *high confidence* in Arctic spring SCE trends, broadly consistent with hemispheric trends reported above (Meredith et al., 2019).

Less attention has been paid to changes in northern hemisphere snow cover during the onset period in the fall, a challenging period to retrieve snow information from satellite imagery due to persistent cloud and decreased solar illumination at higher latitudes. Positive recent trends in October and November SCE in the NOAA-CDR (Hernández-Henríquez et al., 2015) are not replicated in other surface, satellite, and model datasets (Brown and Derksen, 2013; Mudryk et al., 2014). Historical surface observations, model derived analyses, and independent satellite datasets consistently report trends toward later autumn snow-on dates since about 1980 (–0.6 to –1.4 days per decade (updated from Derksen et al., 2017)). SCE trend sensitivity to surface temperature forcing in the NOAA-CDR is anomalous compared to other datasets during October and November (Mudryk et al., 2017), so there is *medium confidence* that the hemispheric SCE trend for the 1981–2016 period was negative also during these two months.

Gridded products from passive microwave remote sensing and land surface models driven by reanalysis identify negative trends in maximum pre-melt SWE over the 1981 – 2016 period over both Eurasia and North America (Mudryk et al., 2017; Mortimer et al., submitted); Figure 9.25c and d). Considerable regional and inter-dataset variability in the spatial patterns of change limit the confidence in the amplitude of these trends (Liston and Hiemstra, 2011; Brown et al., 2017); specifically SWE estimates with passive microwave sensors are problematic in mountainous and/or forested terrain and deep snow (Dozier et al., 2016), with possible future progress based on synthetic aperture radar measurements (Lievens et al., 2019). However, consistency across datasets allows *high confidence* in the existence of a general decline of hemispheric maximum pre-melt SWE over that period. There is also consistent evidence of negative trends in pre-melt maximum snow depth from climate station measurements across Eurasia (earlier positive trends over the 1966–2010 period described in Bulygina et al. (2011) were replaced by predominantly negative trends when the time series was updated in Brown et al. (2017)).

[START FIGURE 9.25 HERE]

Figure 9.25: Observed monthly northern hemisphere snow cover trends (a) and anomalies (b), and snow mass trends (c) and anomalies (d) from the observation-based ensemble discussed in the text (Mudryk et al., submitted). Trends and anomalies are calculated over the 1981–2017 period.

[END FIGURE 9.25 HERE]

Determining snow water equivalent (SWE) trends in mountain regions is challenging because coarse resolution (25 to 50 km) gridded SWE products are inadequate in areas of mountainous terrain (Snauffer et al., 2016). Surface observations are typically too sparse to capture the complex spatial distribution of SWE, including gradients related to elevation (Dozier et al., 2016). Furthermore, there are no internationally coordinated assessments of mountain snow cover; nearly all trend studies focus on a single region or mountain range. There is significant interannual and decadal variability in reported SWE trends because they are impacted by both temperature and precipitation (amount and phase) changes along elevation gradients (Sospedra-Alfonso et al., 2015). There is general consistency in greater reductions in SWE at lower elevations (Marty et al., 2017; Beniston et al., 2018), impacted by shifts from solid to liquid precipitation (Kapnick and Hall, 2012; Marty et al., 2017). SROCC reports *high confidence* in a general reduction of SWE in mountain areas since the middle of the 20th century (Hock et al., 2019) and *medium confidence* in Arctic SWE trends, globally consistent with the hemispheric trends reported here (Meredith et al., 2019).

Estimates of the radiative effect of the Northern Hemisphere cryosphere (Flanner et al., 2011) were recently followed up by a more detailed account of the effect of seasonal snow (Singh et al., 2015). Ongoing and projected decrease of seasonal snow cover in the Northern Hemisphere will *very likely* lead to a reduction of this radiative forcing (Perket et al., 2014) (Section 7.4), and consequently on the snow albedo feedback strength. The impact of aerosol deposition on snow albedo and associated climate forcing is discussed in Section 6.3.

9.5.3.2 Model evaluation

Progress in the observation, description and understanding of snow microstructure (Gallet et al., 2009; Calonne et al., 2017) and its links to physical (thermal and radiative) properties (Domine et al., 2012; Calonne et al., 2014) has prompted efforts to represent physical properties as a function of the evolving snow microstructure in models (Carmagnola et al., 2014; Calonne et al., 2015). However, even state-of-the-art snow models intended for meteorological and climate applications still struggle to correctly represent the time evolution of the snow thermal properties, particularly of cold and dry tundra snow (Domine et al., 2016). Regardless of these shortcomings, snow modules of climate models continue to be improved. Recent progress includes the inclusion of multiple energy balances within the canopy and between subgrid-tiles with different snow heights (Aas et al., 2017; Boone et al., 2017) and inclusion of advanced specific snow models in coupled climate models (Vionnet et al., 2012), opening the prospect of future progress in quantifying snow-related feedbacks in a changing climate. Recently developed multi-physics snow models (Essery, 2015; Lafaysse et al., 2017), which are able to emulate the behaviour of a large number of models in a broad range of climates, allow model shortcomings and key parameter uncertainties, for example concerning snow masking by vegetation or snow thermal conductivity, to be identified.

At the time of AR5, inaccuracies in the simulated snow insulation effect as represented by several CMIP5 models were identified (Koven et al., 2013). A concise metric of snow insulation, building on an observed relation between effective seasonal mean snow depth and the dampening of winter season temperature decrease within the soil, was developed for efficient diagnosis of these inadequacies (Slater et al., 2017), providing guidance for future model improvement.

The is *high confidence* that large inter-model variations in the snow-cover sensitivity to temperature can largely be explained by inaccuracies in the simulated snow albedo feedback (SAF) (Qu and Hall, 2014); a multi-model sub-ensemble of CMIP5 models that simulate a correct magnitude of this feedback presents a 40% reduced spread in the projected 21st century Northern Hemisphere land warming trend (Thackeray et al., 2016). Errors of the simulated feedback strength were linked to 1) systematic positive albedo biases over the boreal forest belt, mostly due to an unrealistic treatment of vegetation masking (Thackeray and Fletcher, 2015), 2) inaccurate prescribed tree cover fraction and inappropriate parameterization of leaf area index in some models (Lorant et al., 2014; Wang et al., 2016a) and 3) low spatial resolution leading to inaccuracies in the strength of the simulated SAF in mountainous regions (Letcher and Minder, 2015). Although the representation of SAF improved in many CMIP5 models over CMIP3, some models deteriorated (Thackeray et al., 2018).

Analysis of the available CMIP6 historical simulations for the 1981-2014 shows that on average, CMIP6 models simulate well the observed snow extent (Thackeray et al., 2016; Mudryk et al., submitted), except for outliers and a median low bias during the winter months (Figure 9.26a) (Mudryk et al., submitted). This is an improvement over CMIP5 (Mudryk et al., submitted), in which many snow-related biases were linked to inadequacies of the vegetation masking of snow cover over the boreal forests (Thackeray et al., 2015). A comparison between CMIP5 and CMIP6 results (Mudryk et al., submitted) shows that there is no notable progress in the quality of the representation of the observed 1981-2014 monthly snow cover trends. Because of (1) physical consistency of the observed spring snowpack and surface temperature changes both in observations and models, (2) the strong observed hemispheric and regional spring SCE and SWE trends, and (3) the general attribution of hemispheric temperature changes to human influence, the observed spring reduction of Northern Hemisphere snow cover in spring is assessed in Section 3.4.2 as *very likely* associated with anthropogenic warming of near-surface air, and thus as also caused by human activities.

[START FIGURE 9.26 HERE]

Figure 9.26: a) Simulated CMIP6 and observed (Mudryk et al., submitted) snow cover extent (in millions of km²) for 1981-2014. Boxes and whiskers with outliers represent monthly mean values for the individual CMIP6 models averaged over 1981-2014, with the red bar indicating median of the CMIP6 multi-model ensemble for that period. The observed interannual distribution over the period is represented in green, with the yellow bar indicating the median. b) Spring (March to May) NH snow cover extent against GSAT (relative to the 1995-2014 average) for the CMIP6 Tier 1 scenarios (SSP1-2.6, SSP2-4.5, SSP3-7.0 and SSP5-.85), with linear regressions. Each data point is the mean for one CMIP6 simulation (first ensemble member for each available model) in the corresponding temperature bin.

[END FIGURE 9.26 HERE]

9.5.3.3 Projections

Based on one study (Brown et al., 2017) applying a method developed by de Elía et al. (2013) to a CMIP5 sub-ensemble, there is *medium confidence* that the projected decrease of annual maximum snow water equivalent (SWE_{max}) will exceed natural variability later than the decrease of snow cover duration (SCD) over most of the Northern Hemisphere. The same study reports that over large parts of Eastern and Western North America and Europe, forced SCD changes are projected to exceed natural variability in the 2020s both in spring and autumn, while the signals tend to emerge later in the Arctic regions and particularly late, after 2060, in Eastern Siberia under the RCP8.5 scenario. Thackeray et al. (2016) have shown that inter-model spread in projected spring SCE trends could be reduced through improved simulation of spring season warming because of the tight coupling between temperature and SCE linked to the SAF (Qu and Hall, 2014; Thackeray and Fletcher, 2015).

Reductions of snow cover (duration and/or mass) are projected using downscaling approaches in various mountainous areas, such as the Western USA (Fyfe et al., 2017), the Subtropical Central Andes (Zazulie et al., 2018), or the Himalaya and Hindu Kush Karakoram regions (Terzago et al., 2014). In Europe, recent studies (Gobiet et al., 2014; Beniston et al., 2018; Verfaillie et al., 2018) consistently suggest strong future snow cover decreases in all scenarios especially near the snow/rain transition level (typically about 1500-2000 m in the Alps), with stabilization after the mid-21st century only in RCP2.6. In higher areas, increased snowfall rates can partially compensate for melt increases, with total compensation (no net decrease) remaining a possibility in the highest parts of the Scandinavian mountains (Beniston et al., 2018). Although in mountain ranges, details of climate model projections depend on methodological choices (e.g., Mendoza et al., 2015), the large overall consistency across a number of studies covering a large diversity of projection methods and regions make it *very likely* that seasonal snow in mountain areas will undergo very strong reductions in maximum SWE especially under low-mitigation scenarios (RCP4.5 and above) and at lower altitudes. Substantially above the current altitude of the snow/rain transition, there is *low confidence* that compensation of increased snowmelt by increased snowfall rates will occur in some cases. This is linked to the fact that below a monthly mean temperature of about $-5 \pm 1^\circ\text{C}$, mountain snow cover is mainly sensitive to solid precipitation rates in the preceding months, while it is mainly sensitive to temperature variations in warmer conditions (Sospedra-Alfonso and Merryfield, 2017). Other factors such as air humidity also impact snow mass balance and modulate the response (Harpold and Brooks, 2018).

Across all emission scenarios and with negligible scenario dependence (Figure 9.26b), CMIP6 models consistently (all models and all months) simulate Northern Hemisphere snow cover decrease in response to future global surface air temperature change (Mudryk et al., submitted). The sensitivity is particularly strong during the transitional season months, with values exceeding $-2.0 \cdot 10^6 \text{ km}^2/^\circ\text{C}$ (Mudryk et al., submitted). The simulated snow cover decrease is close to a linear function of global temperature change for all months (Figure 9.26b), except when snow cover vanishes, which occurs at about $+2^\circ\text{C}$ of GSAT change above the 1995-2014 level (that is, about $+3^\circ\text{C}$ above the preindustrial level) for the months of July and August, and at about $+3^\circ\text{C}$ above the 1995-2014 level for June and September.

In summary, consistent projections from all generations of global climate models, elementary process

understanding and strong covariance between snow cover and temperature on several time scales make it *virtually certain* that future Northern Hemisphere snow cover extent and duration will continue to decrease as global climate continues to warm, and process understanding strongly suggests (*high confidence*) that this also applies to Southern Hemisphere seasonal snow cover.

Seasonal snow cover, by definition, has a clear annual cycle with usually complete disappearance in summer and re-formation in autumn or winter. Therefore, there is *very high confidence* that the current and projected changes to seasonal snow cover are reversible (Verfaillie et al., 2018). In the case of global or regional cooling, abrupt large-scale snow-cover changes with a transition from seasonal to persistent snow cover due to a strong snow albedo feedback are a typical feature of glacial inception (e.g., Baum and Crowley, 2003; Calov et al., 2005), and these can be irreversible on centennial or longer timescales because of this feedback. In summary, based on physical understanding and absence of occurrence of such events in climate model projections, abrupt future changes of seasonal snow cover on large scales in the absence of concomitant abrupt atmospheric change as a driver appear *very unlikely* in the context of current and projected warming.

9.6 Sea Level Change and Commitment

9.6.1 Global and regional sea-level change in the instrumental era

9.6.1.1 Sea-level change since 1900

Since AR5, new statistical methodologies for reconstructing global mean sea level (GMSL) from tide-gauge data over the 20th century have been developed and applied, including one GMSL curve (Dangendorf et al., submitted) in addition to the reconstructions presented in SROCC. Based on the tide-gauge record and given uncertainties from different estimates, the long-term GMSL trend estimate was *very likely* 1.25 [0.98-1.52] mm yr⁻¹ over 1900-1990 and total sea-level rise was 0.15-0.22 m between 1900 and 2018 (*high confidence*) (Table 9.4). Since SROCC, a new reconstruction of historical ocean heat content back to 1870, combining observed sea surface temperature observations with ocean transport information derived from an observation-driven model (Zanna et al., 2019), gives a first observation-based estimate of the thermosteric contribution of 0.38 [0.32-0.44] mm yr⁻¹ (*medium confidence*). In combination with the individual observed contributions of glaciers, Greenland Ice Sheet (GrIS) and Terrestrial Water Storage (TWS) in 1900-1990, the sum of the observations (1.10 [0.96-1.24] mm yr⁻¹) agrees with the total observed GMSL (1.25 [0.98-1.52] mm yr⁻¹) within their respective uncertainties, closing the observational GMSL budget for the period 1900-1990 (*medium confidence*, Table 9.4).

Model simulations have been used to complement the observations in sea-level budget studies (Gregory et al., 2013; Meyssignac et al., 2017; Slangen et al., 2017b) and for individual contributions (e.g., (Marzeion et al., 2015b) for glaciers, (Noël et al., 2015) for Greenland SMB and (Doell et al., 2014; Wada, 2016) for TWS changes). The thermosteric model estimate includes 22 CMIP5 models (compared to 12 in SROCC), corrected for the absence of volcanic forcing in the pre-industrial control run (Slangen et al., 2014b, 2015; Meyssignac et al., 2017; Slangen et al., 2017b), but the additional models do not lead to a change in the mean estimated trend. [PLACEHOLDER FOR ADDITION OF CMIP6 THERMOSTERIC HISTORICAL IN FGD]. The glacier estimate is based on 16 model runs forced by CMIP5 data (compared to 12 in SROCC), which leads to minor changes of the order of 0.01 mm yr⁻¹ in the mean and spread (Marzeion et al., 2015b) [PLACEHOLDER FOR NEW GRIS/AIS MODEL SIMULATIONS IN FGD]. The average sum of simulated contributions is within the 5-95% range of the reconstructed GMSL for 1900-1990, but a large residual remains (*medium confidence*, Table 9.4). Proposed hypotheses for the residual include i) a bias in the early 20th century for Greenland SMB and glacier mass loss as a result of internal climate variability or climate model deficiencies (Slangen et al., 2016, 2017b) or ii) the glacier estimate being too low because small glaciers might not be represented in the inventories or glaciers might have disappeared since 1900 (Parkes and Marzeion, 2018). Both hypotheses are assigned *low confidence* as a result of limited evidence.

9.6.1.2 Sea-level change in the satellite era

Based on satellite data (Cazenave et al., 2018), SROCC concluded that GMSL *very likely* rose 3.16 [2.79-3.53] mm yr⁻¹ from 1993-2015 (*high confidence*) and 3.58 [3.10-4.06] mm yr⁻¹ from 2006-2015 in the GRACE/Argo era (*high confidence*, Table 9.4). Individual estimates of GMSL from tide gauge and satellite altimetry records agree within their respective uncertainties, including the faster rate of rise since 1993 compared to the rate since 1900. The time series of all contributions have been extended to 2018 where possible [PLACEHOLDER TO UPDATE FOR FGD]: the glacier contribution is now based on (Zemp et al., 2019), AIS based on (The IMBIE Team, 2018) and GrIS is based on (The IMBIE Team, 2019). Adding data points to these relatively short time series (going from 23 years to 26 years, and from 9 to 13 years, respectively) decreases the potential impact of internal variability on the trend estimates. For 1993-2018, the additional years lead to only minor variations in the trend estimates compared to SROCC, strengthening the *high confidence* in these observational estimates (Table 9.4). For 2006-2018, the impact of the additional years is larger but still within the *likely* range. Thermal expansion and glacier changes continued to be major contributors to GMSL change in both of these periods (*high confidence*). As a result of improved data quality and coverage due to the availability of satellite data and Argo data, the total observed GMSL change is consistent with the sum of the contributions (*high confidence*, Table 9.4).

Using RCP8.5 model output, the simulated estimates of GMSL have also been extended to 2018. The sum of the simulated contributions is [PLACEHOLDER: 2.49 + ice sheets + TWS] mm yr⁻¹ compared to a total observed GMSL change of 3.16 [2.79-3.53] mm yr⁻¹ for 1993-2018. For 2006-2018, the sum of the simulated contributions is [PLACEHOLDER: 2.74 + ice sheets + TWS] mm yr⁻¹ compared to a total observed change of 3.58 [3.10-4.06] mm yr⁻¹. This closes the simulated vs. observation budget for both satellite periods (PLACEHOLDER: *high confidence*). The comparison of observations and simulations for individual contributions also agree within the *very likely* range. There is *high confidence* that the simulations are a good representation of the observed processes in sea-level change in the satellite altimetry era and the Argo/GRACE era. [PLACEHOLDER FOR NEW LITERATURE ON GMSL SIMULATIONS VS OBSERVATIONS]

[START TABLE 9.4 HERE]

Table 9.4: Contributions to global mean sea level change from observations and from climate models, over four different periods (selected for observational availability), representing the linear trends (mm yr⁻¹). Uncertainties represent the 5-95% range (1σ *1.645).

OBSERVED	1900-1990 {9.6.1.1}	1971-2018 {Cross-Chapter Box 9.2}	1993-2018 {9.6.1.2}	2006-2018 {9.6.1.2}
Thermal expansion (full depth – see Table 2.7 for depth intervals & details)	0.38 [0.30-0.47] (Zanna et al., 2019)	0.88 [0.75-1.01] (Domingues et al., 2008; Purkey and Johnson, 2010; Levitus et al., 2012)	1.08 [1.02-1.14] (Domingues et al., 2008; Purkey and Johnson, 2010; Levitus et al., 2012)	1.44 [1.39-1.49] (Table 2.7)
Glaciers (Excl. peripheral glaciers)	0.50 [0.44-0.56] (Marzeion et al., 2015a)	0.48 [0.22-0.74]* (Zemp et al., 2019)	0.67 [0.40-0.94]* (Zemp et al., 2019)	0.60 [0.35-0.85] (Table 9.4)
Peripheral glaciers Greenland	0.12 [0.06-0.18] (Marzeion et al., 2015a)	0.07 (0.04-0.10)* (Zemp et al., 2019)	0.11 (0.07-0.15)* (Zemp et al., 2019)	0.13 (0.03-0.23)* (Table 9.4)
Peripheral glaciers Antarctica	--	-0.01 (-0.32-0.30)* (Zemp et al., 2019)	0.02 (-0.29-0.33)* (Zemp et al., 2019)	0.03 (0.00-0.06)* (Table 9.4)
Greenland ice sheet	0.24 [0.20-0.28]	0.29 [0.25-0.33]	0.44 [0.38-0.50]	0.67 [0.59-0.75]

	(Kjeldsen et al., 2015)	(Mouginot et al., 2019)	(The IMBIE Team, 2019)	(The IMBIE Team, 2019)
Antarctic ice sheet	--	--	0.30 [0.19-0.41]* (The IMBIE Team, 2018)	0.49 [0.32-0.66]* (The IMBIE Team, 2018)
Terrestrial water storage	-0.14 [-0.20-0.09]^ (Wada et al., 2016)	-0.05 [-0.19-0.08]^ (Wada et al., 2016)	0.15 [-0.00-0.30]^ (Wada et al., 2016)	0.21 [0.04-0.38]^ (Wada et al., 2016)
Sum of observed contributions	1.10 [0.96-1.24]	1.66 [1.21-2.11]	2.77 [2.31-3.23]	3.57 [3.20-3.94]
Observed GMSL change	1.25 [0.98-1.52]^# (Dangendorf et al., submitted)	2.17 [1.76-2.59]^# (Dangendorf et al., submitted)	3.16 [2.79-3.53]^# (Oppenheimer et al., 2019)	3.58 [3.10-4.06]^# (Oppenheimer et al., 2019)
MODELLED	1900-1990 {9.6.1.1}	1971-2018 {Cross-Chapter Box 9.2}	1993-2018 {9.6.1.2}	2006-2018 {9.6.1.2}
Thermal expansion ¹	0.32 [0.02-0.62]	1.02 [0.49-1.55]	1.50 [0.88-2.12]	1.65 [0.94-2.36]
Glaciers ²	0.52 [0.35-0.68]	0.74 [0.43-1.07]	0.99 [0.45-1.53]	1.08 [0.44-1.72]
Greenland	[placeholder for ice sheet modelling results]	[placeholder for ice sheet modelling results]	[placeholder for ice sheet modelling results]	[placeholder for ice sheet modelling results]
Antarctica	[placeholder for ice sheet modelling results]	[placeholder for ice sheet modelling results]	[placeholder for ice sheet modelling results]	[placeholder for ice sheet modelling results]
TWS	[placeholder for TWS modelling results]	[placeholder for TWS modelling results]	[placeholder for TWS modelling results]	[placeholder for TWS modelling results]
Sum of modelled contributions	[placeholder]	[placeholder]	[placeholder]	[placeholder]
Residual (Observed - Modelled)	[placeholder]	[placeholder]	[placeholder]	[placeholder]

*source time series ends in 2016; will be updated for FGD if data allows

^source time series ends in 2014; will be updated for FGD if data allows

#source time series ends in 2015; will be updated for FGD if data allows

¹CMIP5-based thermal expansion includes ACCESS1_0^v, ACCESS1_3^v, CCSM4, CNRM_CM5, CSIRO_Mk3_6_0, CanESM2^v, GFDL_CM3^v, GFDL_ESM2G^v, GFDL_ESM2M, GISS_E2_R, HadGEM2_ES, IPSL_CM5A_LR, IPSL_CM5A_MR, MIROC5, MIROC_ESM, MIROC_ESM_CHEM, MPI_ESM_LR^v, MPI_ESM_MR^v, MRI_CGCM3^v, NorESM1_M, NorESM1_ME, inmcm4^v, where ^v indicates correction for no volcanic in pre-industrial control (Slangen et al., 2014b, 2015; Meyssignac et al., 2017; Slangen et al., 2017b)

²CMIP5-based glacier contribution includes ACCESS1-3, bcc-csm1-1, CanESM2, CCSM4, CNRM-CM5, CSIRO-Mk3-6-0, GFDL-CM3, GISS-E2-R, HadGEM2-ES, inmcm4, IPSL-CM5A-LR, MIROC5, MIROC-ESM, MPI-ESM-LR, MRI-CGCM3, NorESM1-M (Slangen et al., 2017b), based on (Marzeion et al., 2012), upd. with RGI4.0.

[END TABLE 9.4 HERE]

9.6.1.3 Sea-level change acceleration

The combination of tide gauge observations and geological reconstructions indicate an acceleration of GMSL and RSL rise in many locations at a centennial timescale. A review of tide gauge and geological data from seven regions (all in the North Atlantic or Australasia) visually identified inflections in sea-level trends between 1895-1935, with the RSL rates increasing at different sites by 0.8 to 2.5 mm yr⁻¹ across the inflection (Gehrels and Woodworth, 2013). A statistical meta-analysis of globally distributed geological and tide gauge data (Kopp et al., 2016) found that, in all 20 examined regions with geological records stretching

back at least 2000 years, it was *virtually certain* that the rate of RSL rise in the 20th century was faster than the average over 0-1700 CE. In four of the twenty regions, all in the North Atlantic (Connecticut, New Jersey, North Carolina, and Iceland), the 19th century rate was also *likely* faster than the 0-1700 CE average. The associated GMSL reconstruction (Kopp et al., 2016; Kemp et al., 2018) found no evidence of GMSL rise over 1800-1840 and an average rate of rise over 1850-1890 of 0.4 ± 0.4 mm yr⁻¹ (90% CI), and found it *extremely likely* that all 40-year intervals following 1850-1890 exhibited positive rates of GMSL rise. The same meta-analysis also found that the average rate of change over 1900-2000 CE (which it identified as 1.4 ± 0.1 mm yr⁻¹) was faster than that average rate over the 19th century (0.3 ± 0.3 mm yr⁻¹), the 18th century (-0.1 ± 0.3 mm yr⁻¹), and indeed over any century since at least the one beginning in 1000 BCE (Section 9.6.2). Consistent with AR5's conclusion, these lines of evidence provide *high confidence* that an acceleration in GMSL occurred over the late 19th-20th centuries and led to a rate of rise over the last century that was faster than that over any preceding century in the last three millennia.

Tide gauge records support a continued long-term GMSL acceleration over the 20th century. As assessed by SROCC (Oppenheimer et al., 2019) and Chapter 2, estimates of the acceleration of sea-level rise over the twentieth century are generally consistent with those of AR5, with four of five available reconstructions showing robust accelerations from 1902-2010 (*high confidence*).

A new tide-gauge analysis (Dangendorf et al., submitted) decomposes the 20th century acceleration into two episodes -- one in the 1920s and 1930s, and one beginning in the 1960s and continuing to today -- separated by a deceleration over the 1940s and 1950s. The estimated acceleration from 1968-2015 is *likely* 0.06 ± 0.01 mm yr⁻² and is responsible for the net acceleration seen since 1900. The GMSL acceleration that began in the 1960s has led to rates of GMSL rise since the early 1990s that are substantially faster than the 20th century average (Hay et al., 2015; Dangendorf et al., 2017, submitted).

Drift-corrected satellite data (Chen et al., 2017; Dieng et al., 2017; Nerem et al., 2017) confirm the acceleration, with the satellite-estimated rate of GMSL rise since the early 1990s more than doubling relative to the 1900-1990 average (Table 9.4), leading to a *very likely* acceleration of 0.084 ± 0.025 mm yr⁻² over the satellite era (Nerem et al., 2017; Cazenave et al., 2018). This acceleration is primarily due to acceleration in GrIS and AIS mass loss and the impacts of increased GHG concentrations on all contributing processes (*high confidence*); (Velicogna et al., 2014; Harig and Simons, 2015; Slangen et al., 2016; Dieng et al., 2017; Cazenave et al., 2018; Fasullo and Nerem, 2018; Oppenheimer et al., 2019). The acceleration uncertainty (0.025 mm yr⁻²) includes effects of forced natural variability, such as the acceleration after the eruption of Mount Pinatubo (0.01 mm yr⁻²), and effects of internal variability, such as ENSO/PDO modes (0.01 mm yr⁻²) (Nerem et al., 2018). Compared to 1993-2014, the acceleration tripled over 2005-2015, to 0.27 ± 0.17 mm yr⁻². This increase is primarily tied to interannual-to-decadal variability in land water storage (0.11 ± 0.02 mm yr⁻²) and thermosteric sea-level rise (0.12 ± 0.06 mm yr⁻²) (Yi et al., 2017); it would thus be premature to project a continued increase in this elevated acceleration.

9.6.1.4 Regional sea-level change

Both tide gauge and satellite-altimetry records provide measurements of regional sea-level changes, though tide-gauges measure RSL and altimetry measures geocentric sea level. SROCC concluded that sea-level trend patterns since the 1970s in the Pacific and Indian Ocean were driven primarily by surface winds, associated with ENSO, PDO, NPGO, and IOD oscillations and that, in the Atlantic Ocean, these were primarily driven by NAO and AMOC variations (Oppenheimer et al., 2019). Over 1993-2015, analysis of tide-gauge and satellite data indicate that regional-mean relative sea-level (after removal of the GIA signal) has risen fastest in the Indo-Pacific region (3.3 ± 0.3 mm yr⁻¹), the Northwest Pacific (3.0 ± 0.3 mm yr⁻¹), and the subtropical North Atlantic (3.0 ± 0.4 mm yr⁻¹), slower in the south Atlantic (2.5 ± 0.3 mm yr⁻¹) and subpolar North Atlantic (2.2 ± 0.3 mm yr⁻¹), and slowest in the eastern Pacific (1.3 ± 0.7 mm yr⁻¹), Arctic Ocean (1.1 ± 0.8 mm yr⁻¹), and Southern Ocean (1.7 ± 0.4 mm yr⁻¹) (Dangendorf et al., submitted). These patterns are consistent with the suppression of sea-level rise in the eastern Pacific over much of this time period by a negative phase of the PDO (Hamlington et al., 2014), and potentially also by the GRD effects of land ice mass losses at high latitudes. A *likely* sea-level rise acceleration of 0.06 ± 0.02 mm yr⁻² over 1968-

2015 is apparent in all these regions except the eastern Pacific and the polar oceans (Dangendorf et al., submitted), where accelerations are within 1 standard error of zero (*medium confidence*).

There is limited literature on regional sea-level budget closure. Some observation-based studies focus on specific areas, such as the Mediterranean (García et al., 2006), the South China Sea (Feng et al., 2012), the North Atlantic (Kleinherenbrink et al., 2016) or the Northwestern European continental shelf sea (Frederikse et al., 2016). Regional budget studies using tide gauge data and observation-based estimates of the contributions find that local agreement is not yet possible, but that the observational sea-level budget can be closed for some regions on a basin scale (Slangen et al., 2014c; Frederikse et al., 2018). A budget analysis for the GRACE era confirms that the budget closes in some but not all coastal regions: substantial parts of the sea-level change signal cannot be explained by steric or mass variations (Rietbroek et al., 2016), and this part of the signal was attributed to interannual variability. This is in line with other work comparing model estimates to 20th century tide gauge observations, where the majority of local variability is determined by the variability in steric component. In this study, vertical land motion (VLM) is found to be a second major cause explaining local variability (Meyssignac et al., 2017). [PLACEHOLDER FOR NEW LITERATURE ON REGIONAL MODEL PROJECTIONS VS OBSERVATIONS] Literature shows that the regional budget, using either observations or models, can so far only be closed on basin scales (*medium confidence*), with large uncertainties remaining.

9.6.1.5 Attribution and emergence of regional sea-level change

Based on detection and attribution studies (See Cross Chapter Box 1.4 for a definition) of total GMSL (Slangen et al., 2016) and of the individual contributions to GMSL (Marcos and Amores, 2014; Marzeion et al., 2014; Slangen et al., 2014b), Chapter 3), it is *very likely* that anthropogenic forcing is the dominant cause of GMSL rise since 1970 (Chapter 3, SROCC 4.2.2.5). For regional sea-level change, formal detection and attribution of 20th century changes to external forcing is more challenging than for GMSL, because the regional variability is larger than the GMSL variability, and therefore the signal-to-noise ratio decreases (Richter and Marzeion, 2014; Monselesan et al., 2015; Palanisamy et al., 2015).

Formal detection and attribution of regional sea-level change for all ocean basins and including all sea-level contributions is not available in the literature. However, several detection-only studies have been published, focusing on specific ocean areas or individual sea-level processes. A region of interest is for instance the East-West dipole in the Tropical Pacific, where the observed changes have proven difficult to detect and explain using climate model data (Meyssignac et al., 2012; Bilbao et al., 2015; Palanisamy et al., 2015). However, after removing the variability associated with the PDO, (Hamlington et al., 2014) did find a residual trend pattern in the Tropical Pacific that could be linked to warming of the Tropical Indian Ocean. At a set of 59 tide gauge locations, changes at two-thirds of the stations were detected to be outside the range of internal variability (Becker et al., 2014). Subsequently, the minimum anthropogenic contribution was found to range between 20% near Hawaii to 80% in the North Sea and the Baltic Sea (Dangendorf et al., 2017). Several studies have focused on the ocean dynamic and thermosteric sea-level change contributions. (Marcos and Amores, 2014) find that the anthropogenically forced thermosteric change in the upper 700 m in different ocean basins accounts for part (North Atlantic, 65%) or all (tropical Pacific, Southern Ocean) of the observed change. This is in line with in other studies, where observed spatial patterns in the satellite era driven by external forcing could be isolated using model simulations (Fasullo and Nerem, 2018), and the simulated response to external forcings was found to result in detectable changes outside the range of simulated internal variability (Slangen et al., 2015). However, there is evidence that the range of simulated internal variability is sensitive to the model resolution and increases for finer resolutions, which would increase the detection threshold (Sérazin et al., 2016). In summary, detection of forced regional changes for specific ocean areas and contributions is possible (*medium confidence*), but formal attribution of regional sea-level change over longer periods (20th century), for all ocean basins and for all contributions is still lacking.

The time of emergence of an anthropogenically forced sea-level signal shows regional differences: 2-10 years of altimeter data are sufficient to capture the forced signals in the tropical Pacific, while 10–14 years

are required in the eastern Indian Ocean, and 20 years or more in the north Pacific and Atlantic oceans, except in the subpolar gyre (Llovel et al., 2018). Longer records are needed to isolate the forced part of sea level trends in regions where the internal variability is large. Time of emergence studies using sea-level projections from climate models find similar results: the anthropogenic signal in regional sea-level change is expected to emerge first in regions showing relatively small internal variability in the models, such as the tropical Atlantic Ocean and the tropical Indian Ocean (Jordà, 2014; Lyu et al., 2014; Richter and Marzeion, 2014; Bilbao et al., 2015). In regions where the climate model internal variability is large and projected changes are small, such as the Southern Ocean, the signal is not found to emerge before 2100. Adding the projected sea-level change from land ice mass loss and groundwater extraction amplifies the forced signal and leads to earlier times of emergence in most parts of the ocean, except close to sources of mass loss (Lyu et al., 2014; Richter et al., 2017). Based on the available literature on time of emergence in sea-level observations and model projections, there is *medium confidence* that the anthropogenic forced signal in regional sea-level change will emerge in 50% of the ocean area before 2040.

[START CROSS-CHAPTER BOX 9.2 HERE]

Cross-Chapter Box 9.2: The Global Energy Inventory and Sea Level Budget (1971-2018)

Contributors: Catia M. Domingues (Australia/Brazil), Matthew D. Palmer (UK), Johannes Quaas (Germany), Aimée B.A. Slangen (Netherlands)

[note for SOD: reported numbers are for 1971-2015; for FGD this is intended to report 1971-2018]

Increased atmospheric greenhouse gas emissions since the 19th century have led to a net positive radiative forcing of the Earth system (Sections 7.3, 2.2.3) and a corresponding accumulation of energy in the global climate system. Quantification of this energy change is essential to our understanding of the rate of global climate change and for estimates of climate sensitivity (Section 7.5).

It is *virtually certain* that the Earth system gained substantial energy over the period 1971-2015, with a best estimate of 367 ± 84 ZJ or 0.52 ± 0.12 W m⁻² expressed relative to the surface area of the Earth (Cross-Chapter Box 9.2, Figure 1a; Chapter 5; Chapter 7). Ocean warming dominated the energy inventory change (*high confidence*), accounting for 92% of the observed energy change for the period 1971-2015, of which the upper ocean warming (0-700 m) accounted for 60%. Much smaller amounts went into melting of ice (~2%), heating of the land surface (~5%) and atmosphere (<1%). The dominant role of the ocean in the energy budget on annual and longer timescales (Palmer and McNeall, 2014; Johnson et al., 2016) makes ocean heat content change one of the most robust quantities for tracking the rate of global warming and less susceptible to unforced decadal variability than global surface temperature (Palmer and McNeall, 2014; Wijffels et al., 2016).

Changes in heat storage in the climate system have direct implications for the global mean sea level (GMSL) budget (Section 9.6.1), as increased ocean heat storage leads to warming of the oceans and thermal expansion (Section 9.2.4) and heating and melting of land-based ice adds water to the oceans (Sections 9.4, 9.5.1). The sum of the observed contributions indicates that GMSL *very likely* increased by 0.079 ± 0.028 m over the period 1971-2015 (Cross-Chapter Box 9.2, Figure 1b, Table 9.5). The main contributions to the GMSL change over this period are from ocean thermal expansion (52%), glacier mass loss (28% excluding peripheral glaciers), and ice sheet mass loss (22%). Changes in land water storage are a small negative contribution over this period due to reservoir storage (~-3%).

The comparison of the sum of the observed contributions to an independent reconstruction of total GMSL change ((Dangendorf et al., submitted), Cross-Chapter Box 9.2, Figure 1c) demonstrates closure of the global sea level budget for the period 1971-2018 within the estimated uncertainty (90% confidence interval), strengthening the increased confidence of SROCC (Oppenheimer et al., 2019) in the observed sea-level change and understanding of the contributions. Similarly, the observed energy inventory changes are consistent with revised estimates of effective radiative forcing and Earth's radiative response (Chapter 7,

Box 7.2). Therefore, using a common observational basis for the ocean and cryosphere changes, a consistent closure of the global energy and sea level budgets for the period 1971-2015 can be demonstrated (*high confidence*), increasing confidence in our ability to quantify these fundamental aspects of climate change relative to AR5.

[START CROSS-CHAPTER BOX 9.2, FIGURE 1 HERE]

Cross-Chapter 9.2, Figure 1: a) Observed changes in the global energy inventory components for 1971-2015; b) Observed changes in global mean sea level components for 1971-2015; c) Sea level budget for 1971-2015, comparing the sum of components in b) to the total reconstructed sea-level change (Dangendorf et al., submitted). a) and b) show individual components in shaded colours as indicated in the figure legends. Shading in c) indicates the 5-95 percentile range. Full details of the datasets and methods used available in the Annex I. Corresponding sea-level trends are in Table 9.4.

[END Cross-Chapter BOX 9.2, FIGURE 1 HERE]

[END CROSS-CHAPTER BOX 9.2 HERE]

9.6.2 Geological context of global and regional sea-level change

Past warm periods differ from future warm periods in forcing, including the spatial and temporal distribution of temperature. Nonetheless, they provide indications of the response of sea level to warmer conditions, as well as key targets for testing and calibrating the climate and ice-sheet models used to project future conditions. This section focuses on assessing the drivers of past GMSL changes and areas where there are climatically coherent areas of RSL variability; past changes in GMSL as a global metric are assessed in (Chapter 2).

During the mid-Pliocene Warm Period (MPWP, 3.3-3.0 Ma), GMST was *very likely* $3 \pm 1^\circ\text{C}$ higher than pre-Industrial, CO_2 concentrations were 350-400 ppm, and GMSL was between 5 and 25 m higher than today (*medium confidence*) (Cross-Chapter Box 2.4). This large reconstructed GMSL range means that the MPWP provides only a relatively weak constraint for individual ice-sheet simulations. Due in part to a broad spread in simulations of summer and mean annual air temperatures over Greenland (Yan et al., 2014; Dolan et al., 2015), simulations of MPWP GrIS volume change are effectively unconstrained, ranging from growth from the present volume by 0.6 m sea-level equivalent (SLE) to complete ice sheet loss (Hill et al., 2010; Yan et al., 2014; Contoux et al., 2015; Dolan et al., 2015; Koenig et al., 2015). Geological evidence of Pliocene GrIS extent is sparse and indirect (e.g., Blake-Mizzen et al., 2019), so it is challenging to distinguish between these different model simulations. Both budgetary accounting and model simulations suggest AIS contributed at least about half the total increase in MPWP GMSL relative to today (Gasson et al., 2016b; Yan et al., 2016; Golledge et al., 2017b), implying contributions from both WAIS and EAIS, but constraints are similarly poor. Although there is *high confidence* that GMSL was higher than present during the MPWP, due to contributions from GrIS and AIS, quantification is not possible due to climate and ice sheet modelling uncertainties and sparse data.

During the Last Interglacial (LIG, ~129-116 ka), GMST was *very likely* $1\text{-}2^\circ\text{C}$ above pre-Industrial temperatures (Chapter 2). It is *virtually certain* GMSL was higher than today, *likely* by 5–9 m (Chapter 2). No more than 0.3 ± 0.1 m of this GMSL rise could have come from glaciers (Section 9.5.1), and no more than 0.4 ± 0.3 m from global mean thermal expansion (McKay et al., 2011), so all but 0.7 ± 0.3 m of the GMSL rise must be sourced from the polar ice sheets (Dutton et al., 2015a). As noted by SROCC (Oppenheimer et al., 2019), GrIS modelling studies yield peak GrIS contributions to GMSL between 0.3 m (5% estimates of (Stone et al., 2013)) and 6.2 m (97.5% estimate of (Yau et al., 2016)) (Helsen et al., 2013; Quiquet et al., 2013; Stone et al., 2013; Dutton et al., 2015a; Goelzer et al., 2016a; Yau et al., 2016). This spread arises because models match either evidence for continuous ice presence and small elevation change at the North Greenland Eemian Ice Drilling (NEEM) site, leading to lower estimates of GrIS contributions (Dahl-Jensen et al., 2013; Helsen et al., 2013; Goelzer et al., 2016a; Landais et al., 2016) or evidence for

high temperatures reconstructed from the NEEM, GRIP or GISP2 ice cores (Robinson et al., 2011; e.g., Goelzer et al., 2016a; Yau et al., 2016). While model-based and geological constraints on LIG ice-sheet extent and volume are broad, they are consistent in indicating that the initial LIG sea-level rise was driven by reduced AIS volume, while later in the LIG, both AIS and GrIS contributed to high sea levels (Rohling et al., 2019; Clark et al., in press), with a peak GrIS contribution at 123-122 ka, consistent with the asymmetry in the timing of peak polar temperatures (Capron et al., 2017) (*medium confidence*). Asynchronous minima in AIS and GrIS volume may explain evidence for multiple sea-level peaks during the LIG, but there is *low confidence* in whether multiple LIG GMSL peaks exist (Chapter 2). Based on one climate/ice-sheet modelling study (Clark et al., in press), the oceanic forcing associated with a weakened AMOC during the deglaciation was primarily responsible for excess ice-sheet loss during the LIG relative to the Holocene with a maximum ice-sheet contribution of ~4 m (although this amount of melting is too small to simultaneously fit all geological constraints) (*low confidence*).

Comparison of geological proxies and GIA models indicates that at the LGM (~21-19 ka) GMSL was 125-134 m below present (Lambeck et al., 2014; Yokoyama et al., 2018) (Chapter 2). Of this total, the central 67% probability contributions are estimated at 76 ± 7 m in the North American Laurentide ice sheet, 18 ± 5 m in the Eurasian ice sheet, 10 ± 2 m in Antarctica, 4 ± 1 m in Greenland, 5.5 ± 0.5 m in glaciers and ice caps, and 2.4 ± 0.3 m due to an increase in ocean density (Simms et al., 2019). Of the residual, up to about 1.4 m may be ascribed to groundwater, leaving a shortfall of 16 ± 10 m yet to be allocated among land ice reservoirs or lakes (Simms et al., 2019).

During Meltwater pulse 1A (MWP-1A) (14.65-14.31 ka), GMSL *very likely* (*medium confidence*) rose between 8 and 15 m (Liu et al., 2016), an average rate of 24-44 mm yr⁻¹ (Chapter 2). The spatial patterns of RSL change over this interval are inadequately observed to constrain the relative contributions of the North American and Antarctic ice sheets (Liu et al., 2016). Alternative North American ice sheet modelling studies permit a 3-6 m contribution (Gregoire et al., 2016) or a 9-13 m contribution over 500 yr (6-9 m over the duration of MWP-1A) (Tarasov et al., 2012). A pulse of ice-rafted debris (Weber et al., 2014) provides near-field evidence for an Antarctic contribution, consistent with the AIS modelling study of Golledge et al. (Golledge et al., 2014), while a recent modelling study suggests a 3-7 m contribution from the Eurasian Ice Sheet (Brendryen et al., submitted), a possibility not considered in the meta-analysis of Liu et al. (2016). Thus, MWP-1A appears to have been driven by a combination of melt in North America (*high confidence*), Eurasia (*low confidence*), and Antarctica (*low confidence*), but the budget is not closed.

Post-glacial climatic warming during the early Holocene, following an orbitally forced insolation maximum at 10 ka (Blaschek and Renssen, 2013), led to increased summer warmth of 3 ± 1 °C above pre-Industrial levels in Arctic Canada and Greenland that was earliest north of ~66°N and later in the south (Briner et al., 2016). Several lines of evidence (lake sediments, radiocarbon-dated moraines, and cosmogenic surface exposure dates) indicate that this spatially-diachronous Holocene Thermal Maximum (HTM) warming triggered retreat of GrIS margins, leading to reduced GrIS extent compared to present-day, especially in south and west Greenland (Carlson et al., 2014; Larsen et al., 2014, 2015; Young and Briner, 2015). It is therefore *likely* that total GrIS volume was also reduced. Numerical modelling of the HTM GrIS is broadly consistent with geological reconstructions but uncertainties exist regarding when the ice sheet reached its minimum HTM extent and volume, associated with climate and margin reconstructions from proxy data, as well as spatial differences in the timing of margin retreats, but it is *likely* (*medium confidence*) that diachronous minima were reached in many areas during the period 8-3 ka BP (Larsen et al., 2015; Young and Briner, 2015; Briner et al., 2016). There is *low confidence* in the 0.16 m SLE GrIS contribution estimated with an ice-sheet model informed by lake records, due to known model inadequacies that tend to underestimate the response (Larsen et al., 2015). Despite spatial differences in timing, geochronological and numerical modelling studies indicate that it is *likely* (*medium confidence*) that the period of smaller-than-present ice extent in all sectors of GrIS persisted for at least 2000 to 3000 years (Larsen et al., 2015; Young and Briner, 2015; Briner et al., 2016; Nielsen et al., 2018). However, due to some combination of masking by other factors (e.g., ongoing melt from the Laurentide ice sheet until ~6700 BP and limited resolution of available records), evidence for subsequent GrIS advance is not apparent in sea-level proxy records.

9.6.3 Future sea-level changes

9.6.3.1 Global mean sea level projections in the literature and their evaluation

Since AR5, numerous integrated projections of GMSL rise and several global sets of integrated RSL rise projections have been published (Palmer et al.; Slangen et al., 2014a; Kopp et al., 2014, 2017; Grinsted et al., 2015; Schleussner et al., 2016; Jackson and Jevrejeva, 2016; Mengel et al., 2016; Bakker et al., 2017; Bittermann et al., 2017; Nauels et al., 2017a, 2017b; Wong et al., 2017; Le Bars et al., 2017; Rasmussen et al., 2018; Jackson et al., 2018; Le Cozannet et al., 2019) (see (Garner et al., 2018) for a database). Most of these projections fall into one of three categories (Horton et al., 2018). (1) Central-range projections combine estimated scenario-conditional probability distributions for the different factors driving global and regional sea-level change, to estimate a central range (e.g., a likely range) of projections under different scenarios. These projections are most directly comparable to those presented in AR5. (2) Probabilistic projections are distinguished from central-range projections in that they explicitly consider outcomes at more than three probability levels, including low-probability outcomes. They were developed in part as an effort to unify central-range projections and physically plausible high-impact scenarios in a single coherent framework. (3) Semi-empirical projections are based upon statistically identified relationships between past GMSL changes and climate variables. Semi-empirical models (SEMs) were in general systematically higher than central-range projections in AR5, and *low confidence* was assigned to SEMs because of the lack of a physical explanation for this discrepancy. Since the AR5, the use of a two-millennium GMSL reconstruction for calibration (Kopp et al., 2016) and the calibration of individual contributions (Mengel et al., 2016) has generated SEM results that are consistent with the AR5 projections. Consistent with AR5, however, there remains *low confidence* in the ability of semi-empirical models to project GMSL rise under future scenarios that are considerably warmer than during the calibration period.

Partial cumulative distribution functions (CDFs) of GMSL projections from the literature for different post-AR5 studies in 2050 and 2100 under different RCPs and temperature stabilization scenarios are shown in Figure 9.27. For each study, the values along the x-axis represent different GMSLs (standardized to account for minor differences in time periods) and the y-axis represents the probability that each study ascribes to that level not being exceeded, conditional upon the scenario. The dispersion among the different CDFs represents deep uncertainty, which arises as a result of low agreement regarding appropriate conceptual models describing ice sheet behaviour and low agreement regarding probability distributions used to represent key uncertainties. One way of representing this limited agreement is with ‘probability boxes’ or ‘p-boxes’ (Kriegler and Held, 2005; Le Cozannet et al., 2017), which represent upper and lower bounds to plausible CDFs. A p-box in principle should encompass any plausible CDF that a reasonable expert might generate; and in practice, the empirical p-boxes here encompass all the CDFs represented in the assessed studies (Figure 9.27, shaded area). To the extent that the available literature projections represent the plausible range of probability estimates, it is, by construction, *extremely likely* that GMSL rise will be above the lowest 5th percentile in the p-box and *extremely likely* that it will be below the highest 95th percentile in the p-box (Table 9.5). (This description of the p-box is used to clearly distinguish this broad, literature-based set of projections from the new integrated projections presented in Section 9.6.3.3 where a *likely/very likely* range is employed). Figure 9.27 distinguishes the area demarcated by studies that neither include MICI ((Deconto and Pollard, 2016); Box 9.3) nor rely strongly on structured expert judgement (SEJ) ((Bamber and Aspinall, 2013; Bamber et al., 2019); 9.4) (shaded in dark grey, *medium confidence*), from the area demarcated by studies including MICI ((Deconto and Pollard, 2016), used in (Kopp et al., 2017; Wong et al., 2017)) or relying strongly on SEJ ((Bamber and Aspinall, 2013), used in (Jevrejeva et al., 2014; Grinsted et al., 2015), and one set of projections in (Jackson and Jevrejeva, 2016), as well as (Bamber et al., 2019)) (shaded in light grey, *low confidence*).

There is a *high agreement* among published GMSL projections for 2050, reflected in the relatively narrow width of the p-box (Table 9.5 and Figure 9.27, left panels). Similarly, there is *high agreement* on a low level of emissions scenario sensitivity, reflected in the high degree of overlap among the p-boxes for different scenarios. Conversely, for 2100, there is *low agreement* (Table 9.5 and Figure 9.27, right panels), particularly for higher emissions scenarios, as well as a higher degree of sensitivity to the choice of emissions scenario. It is *extremely likely* that 21st century GMSL rise under different emissions scenarios will

be above the lowest 5th percentile of projections (*high confidence*). There is *high confidence* that GMSL rise will *extremely likely* be below the highest of the 95th percentile projections including MICI and SEJ, and *medium confidence* that GMSL rise will *extremely likely* be below the highest 95th percentile projections excluding studies relying on MICI or SEJ.

[START TABLE 9.5 HERE]

Table 9.5: Overview of end member global mean sea-level projections from the assessed post-AR5 literature, for RCP scenarios and temperature stabilization scenarios, relative to a 1995-2014 baseline. Updated sea-level projections, based on CMIP6 data, will be provided in Section 9.6.3.2 to 9.6.3.4, Table 9.6 and 9.7 and Figures 9.29 to 9.31.

		<i>Extremely likely above</i> (lowest 5th percentile)	<i>Extremely likely below</i> (highest 95th percentile)		
			Excluding MICI and SEJ	Including MICI	Including SEJ
RCP 8.5	2050	0.12 m (1)	0.36 m (2)	0.46 m (3)	0.59 m (4)
RCP 4.5	2050	0.11 m (1)	0.33 m (2)	0.41 m (3)	-
RCP 2.6	2050	0.10 m (1)	0.31 m (2)	0.40 m (3)	-
RCP 8.5	2100	0.34 m (5)	1.57 m (9)	2.41 m (3)	2.36 m (4)
RCP 4.5	2100	0.19 m (5)	0.94 m (9)	1.56 m (3)	-
RCP 2.6	2100	0.14 m (6)	0.81 m (2)	0.97 m (3)	-
2.0°C	2100	0.21 m (7)	0.94 m (8)	-	1.24 m (4)
1.5°C	2100	0.17 m (7)	0.80 m (8)	-	-

Study providing the most extreme projection at specific percentile: (1) (Mengel et al., 2016). (2) (Kopp et al., 2014). (3) (Kopp et al., 2017). (4) (Bamber et al., 2019). (5) (Jackson and Jevrejeva, 2016). (6) (Le Cozannet et al., 2019). (7) (Jackson et al., 2018). (8) (Rasmussen et al., 2018). (9) (Bakker et al., 2017). Reference baseline adjusted to 1996-2014 assuming an average rate of 3.1 mm yr⁻¹ from 1995 to 2005. For studies reporting projections for 2081-2100 rather than for 2100, values are linearly extrapolated to 2100.

[END TABLE 9.5 HERE]

[START FIGURE 9.27 HERE]

Figure 9.27: Partial cumulative distribution functions (CDFs) from literature global mean sea level (GMSL) projections (m) for 2050 (left) and 2100 (right), for RCP8.5 (upper), RCP4.5 (middle) and RCP2.6 (lower). For each study, the values along the x-axis represent different GMSLs (standardized to account for minor differences in time periods) and the y-axis represents the probability that each study ascribes to that level not being exceeded, conditional upon the scenario. Dark brown shading indicates the *medium confidence* p-box (excluding studies incorporating MICI or relying on structured expert judgement [SEJ]) and light brown shading indicates the *low confidence* p-box (including MICI and studies relying on SEJ). Because of limited availability of reported outcomes, the p-box is not shaded below the 5th or above the 95th percentile.

[END FIGURE 9.27 HERE]

9.6.3.2 Contributions to projected sea-level change

This section describes the choices made for the contributions to the updated sea-level projections (Section 9.6.3.3) based on assessments earlier in this chapter. There is no single model that can directly compute all of the contributions to sea-level change (Box 9.1) which means that all of the components have to be computed separately and then combined. Here, a probabilistic framework is used following the procedure in Figure 9.28 to produce projections of GMSL change, RSL change, and ESL change. The methodology is similar to that presented in IPCC AR5 Chapter 13 (Church et al., 2013a), adapted to allow for fully probabilistic projections. For all global-mean components except thermal expansion, projections are forced by GSAT projections from the FAIR simple climate model (Smith et al., 2018), calibrated to be consistent with the assessment of ECS and TCR in chapter 7 (see discussion in (Forster et al., 2019)).

9.6.3.2.1 Contributors to global mean sea-level change projections

The global mean thermosteric sea-level contribution is based on output from CMIP6 models (Section 9.2.4). Following the approach of (Kopp et al., 2014), a multivariate t-distribution is fitted to ocean dynamic and global mean thermosteric sea-level terms from CMIP6, after drift-correcting the global mean thermosteric terms using pre-industrial control runs. Consistent with the judgement that the 5th-95th percentiles of the GCM distribution represent a *likely* range (Oppenheimer et al., 2019), the standard deviation of the fitted distribution is multiplied by 1.7, so that ± 1 standard deviation of the scaled fitted distribution corresponds to a central 66% credible range. This yields a *likely* global-mean thermosteric contribution of 0.15 (0.13-0.17) m for SSP1-2.6, 0.19 (0.17-0.22) m for SSP2-4.5, 0.24 (0.23-0.26) m for SSP3-7.0, and 0.29 (0.28-0.31) m for SSP5-8.5 by 2100 (Figure 9.29, Table 9.6). The number of CMIP6 models reporting global mean thermosteric sea-level rise for control, historical, and future simulations is currently extremely limited, with only 2 models reporting this variable for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, and only 1 model reporting this variable for SSP1-1.9. These projections are thus expected to change as more CMIP6 model results become available. Projections will also be developed based on scaling GSAT and deep ocean temperature from the two-layer climate model presented in Chapter 7. Accounting for the shift in baseline from 1986-2005 to 1996-2014 between AR5 and AR6, AR5 projected 0.14 (0.10-0.19) m for RCP2.6, 0.19 (0.14-0.24) m for RCP4.5, and 0.31 (0.24-0.38) m for RCP8.5. The differences in median global mean thermal expansion is thus currently minimal between CMIP5 and CMIP6, and differences in spread are small and hard to interpret in light of the more restricted CMIP6 model availability.

The projected glacier contribution is based on the GlacierMIP Phase 1 results (Section 9.5.1) (Hock et al., 2019), which are driven by CMIP5 data. The GlacierMIP results are used to update the parametric representation of glacier mass loss as a power function of cumulative temperature that was employed in AR5 (Church et al., 2013a), which is then driven with FAIR projections of global mean temperature. [PLACEHOLDER TO UPDATE IN THE FINAL DRAFT TO INCLUDE THE EMULATED GLACIERMIP PHASE 2 RESULTS (Edwards et al., submitted)]. This yields a *likely* global-mean glacier contribution of 0.09 (0.07-0.11) m for SSP1-2.6, 0.11 (0.08-0.14) m for SSP2-4.5, 0.13 (0.10-0.16) m for SSP3-7.0, and 0.14 (0.11-0.18) m for SSP5-8.5 by 2100 (Figure 9.29, Table 9.6). By comparison, accounting for the shift in baseline from 1986-2005 to 1996-2014 between AR5 and AR6, AR5 projected 0.10 (0.04-0.16) m for RCP2.6, 0.12 (0.06-0.19) m for RCP4.5, and 0.17 (0.09-0.25) m for RCP8.5. As with global mean thermal expansion, differences between the AR5 and AR6 projections are minimal.

The terrestrial water storage (TWS) projections are based on statistical relationships between global population and dam impoundment and between global population and groundwater extraction (Rahmstorf et al., 2012; Kopp et al., 2014). Whereas the rate of increase of dam impoundment declines with population growth, the rate of groundwater extraction is projected to increase due to higher water demand and increased evaporation loss (e.g., (Wada et al., 2012)), with *medium confidence* that ~80% ends up in the ocean (Wada et al., 2016). Projections of natural contributions of TWS have not been included in published sea-level projections, as they are not well quantified or are considered negligible (see for example permafrost, Section 9.5.2). The combined cumulative contribution of groundwater extraction (multiplied by 0.8 following (Wada et al., 2016)) and reservoir impoundment as a function of global population yields a *likely* global-mean TWS contribution of 0.04 (0.02-0.05) m for SSPs 1, 2, 4 and 5, and 0.04 (0.03-0.06) m for SSP 3 (Figure 9.29,

Table 9.6). By comparison, AR5 projects 0.05 (-0.01-0.11) m for all scenarios.

Several different alternative options were explored for ice sheet projections (Sections 9.4.1.2, 9.4.2.2), including (1) using ISMIP6 results directly (mapping SSP1-2.6, SSP2-4.5 and SSP5-8.5 to RCP2.6, RCP4.5, and RCP8.5) (Nowicki et al., 2016), (2) using emulated ISMIP6 results for the SSPs (Edwards, submitted), and (3) using results from LARMIP2 (Levermann et al., submitted, b) for the Antarctic contribution. ISMIP6 results are analysed relative to a control run based on constant 2015 conditions, so historical trends are added to the projections (West Antarctic Ice Sheet: 0.28 ± 0.03 mm yr⁻¹; Antarctic Peninsula: 0.06 ± 0.01 mm yr⁻¹; East Antarctic Ice Sheet: -0.02 ± 0.05 mm yr⁻¹; Greenland Ice Sheet: 0.46 ± 0.04 mm yr⁻¹) assuming that current trends reflect forced change rather than internal variability, an assumption that may require further evaluation. LARMIP2 results were considered as an option but these include only the dynamic Antarctic contribution for basal ice shelf melt, and therefore may overestimate the Antarctic sea-level contribution due to their exclusion of surface mass-balance. [PLACEHOLDER FOR EXTENSIVE ASSESSMENT OF ANTARCTIC DYNAMIC CONTRIBUTIONS FOR ISMIP6 AND LARMIP2 IN 9.4 AND 9.6]. The following are also considered for context: (1) an approach similar to that employed by Kopp et al. (2014), which yields probabilistic projections with *likely* ranges consistent with SROCC, and (2) the SEJ projections of Bamber et al. (2019), for context in interpreting SSP5-8.5 projections (described also in Section 9.4). Comparison of the ice sheet options for different scenarios is in Tables 9.1, 9.2, 9.A.1. The primary sea-level projection results presented here, in Figures 9.29 to 9.31 and Tables 9.6 and 9.7, are based upon the emulated ISMIP6 results, driven by the FAIR temperature projections.

For GrIS, the emulated ISMIP6 results yield *likely* projections of 0.07 (0.03-0.12) m, 0.09 (0.05-0.14) m, 0.12 (0.08-0.16) m, and 0.13 (0.09-0.19) m for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5 by 2100 (Figure 9.29, Table 9.6). This compares with AR5 projections of 0.08 (0.04-0.12) m, 0.09 (0.05-0.16) m, and 0.15 (0.09-0.28) m by 2100 for RCP2.6, RCP4.5, and RCP8.5, respectively. The ISMIP6 projections for Greenland are similar to AR5 for lower emissions scenarios, and more narrow than those of AR5 for high emissions. By comparison, the Bamber et al. (2019) SEJ indicates a *likely* contribution of 0.23 m (0.10-0.60) m under a scenario similar to SSP5-8.5, higher and with larger uncertainty than both the primary projections and AR5.

For AIS, the emulated ISMIP6 results yield *likely* AIS projections of 0.12 (0.01-0.25) m, 0.11 (0.00-0.24) m, 0.12 (0.00-0.24) m, and 0.11 (0.01-0.26) m by 2100 for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively (Figure 9.29, Table 9.6). Notably, the ISMIP6 ensemble exhibits no significant scenario dependence in Antarctic projections. This contrasts with the LARMIP2 *likely* projections of 0.12 (0.06-0.25) m, 0.14 (0.07-0.28) m, 0.15 (0.08-0.31) m, and 0.17 (0.08-0.36) m by 2100 for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively. It also compares with SROCC projections of 0.04 (0.01-0.11) m, 0.06 (0.01-0.15) m, and 0.12 (0.03-0.28) m by 2100 for RCP2.6, RCP4.5, and RCP8.5, respectively. Median projections with the emulated ISMIP6 ensemble are similar for SSP5-8.5 and RCP8.5, but the AR5 projections exhibit scenario sensitivity. Notably, the uncertainty ranges for the LARMIP2 projections are substantially broader than for the ISMIP6 ensemble. By comparison, the Bamber et al. (2019) SEJ indicates a *likely* contribution of 0.21 m (0.02-0.57) m, similar to the 5th-95th percentile range from LARMIP2 and substantially higher than the primary projection.

9.6.3.2.2 Contributors to regional relative sea-level change projections

Like the global mean thermosteric sea-level contribution, the ocean dynamic sea-level contribution to relative sea-level change (Figure 9.29) is based on output from CMIP6 models (Section 9.2.4). Following the approach of (Kopp et al., 2014), a multivariate t-distribution is fitted to ocean dynamic sea-level terms from CMIP6. Correlation with global mean thermal expansion is not currently taken into account, due to limited model availability. Consistent with the judgement that the 5th-95th percentiles of the GCM distribution represent a *likely* range, the standard deviation of the fitted distribution is multiplied by 1.7, so that ± 1 standard deviation of the scaled fitted distribution corresponds to a central 66% credible range. The number of CMIP6 models reporting ocean dynamic sea-level rise is currently quite limited: 8 models for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, and 5 for SSP1-1.9.

For each of the projected mass change contributions to sea-level change (glaciers, ice sheets and TWS), GRD effects (Box 9.1) lead to distinct variations in the RSL change pattern, which are similar across a range of benchmarked GRD solvers (Martinec et al., 2018; Palmer et al., submitted). RSL rise associated with GRD is *very likely* to be largest in the Pacific, due to the combined effects of projected GrIS, AIS and glacier mass loss (*high confidence*) (e.g., Kopp et al., 2014; Slangen et al., 2014a; Larour et al., 2017; Mitrovica et al., 2018). The GRD effect will be greater in Australia associated with mass loss from the Antarctic Peninsula as opposed to Thwaites Glacier, and the opposite signs in parts of northeastern North America and northwestern Europe of the GRD effects from northern and southern Greenland (*high confidence*) (Larour et al., 2017; Mitrovica et al., 2018). The RSL patterns are computed using a gravitationally self-consistent GRD solver using the amounts, locations and timing of the projected mass changes of glaciers, ice sheets and TWS (Figure 9.29). The GRD solver used (following (Slangen et al., 2014a)) solves the sea-level equation with a pseudo-spectral approach (Farrell and Clark, 1976; Mitrovica et al., 2001), and considers changes in the Earth's gravitational field, in the solid-earth deformation and in the Earth's rotation vector. The Earth model used is based on PREM (Dziewonski and Anderson, 1981), and is elastic, compressible and radially stratified.

GIA (ongoing GRD as a result of mass loss of ice sheets after the last glacial maximum) leads to VLM and RSL changes (Box 9.1). GIA uncertainty is partly caused by the uncertainty in the rheological structure of the solid Earth which drives the longer-term viscous Earth deformation. Another reason for GIA uncertainty is the treatment and derivation of the global ice history (e.g., (Whitehouse, 2018)). Although GIA modelling efforts have been substantial and evolving, there are only few global model solutions that could be used in global sea-level projections: the ICE5G models from (Peltier et al., 2015) and the models from (Lambeck et al., 2014). VLM other than GIA or GRD, for instance due to subsidence or tectonics, can be an important local factor to RSL change, but complete global projections are not available because of the small spatial scale, the sensitivity of subsidence to local human activities, and the stochasticity of tectonics. Integrated RSL projections to date have therefore modelled VLM either by incorporating only GIA from models (e.g., (Church et al., 2013a)) or by using a constant VLM rate derived from historical tide gauge trends (e.g., (Kopp et al., 2014)). Here, the VLM fields are derived from tide gauges and assumed to be constant over the projected period and independent of the scenario followed (Figure 9.29). The tide gauge trends are extrapolated to global coverage following a spatiotemporal statistical approach (Kopp et al., 2014), using the ICE5G-VM2-90 model as a statistical prior.

[START FIGURE 9.28 HERE]

Figure 9.28: Flowchart illustrating the procedure for computing the global mean sea level change (red), regional relative sea level change (purple) and extreme sea level change (purple) projections framework. White blocks with black outline represent external forcing, blue blocks represent model options, orange block represents tide gauge data, yellow blocks represent sea-level contributors, white blocks with purple outline the corrections for vertical land movement.

[END FIGURE 9.28 HERE]

[START FIGURE 9.29 HERE]

Figure 9.29: Global mean and regional relative sea-level projections (m) by contribution for the SSP5-8.5 and SSP1-2.6 scenarios. (a) and (b) global mean contributions to sea-level change as a function of time, relative to 1995-2014, (b)-(k) regional projections of the sea-level contributions in 2100 relative to 1995-2014 for SSP5-8.5 (c, e, g) and SSP1-2.6 (d, f, h). (c)-(d) Ocean contribution (includes ocean dynamics, steric change, inverse barometer), (d)-(f) ice sheets and glaciers, (g)-(h) Terrestrial Water Storage, (i) Vertical Land Movement (common to both SSPs).

[END FIGURE 9.29 HERE]

9.6.3.3 Updated sea-level projections for the next century based on scenarios

The updated sea-level projections presented in this section are based on the contributions and the methodology presented in Section 9.6.3.2. In contrast to IPCC AR5 and SROCC, the sea-level projections are based on CMIP6 scenarios. While CMIP5 data were available for four RCP scenarios (RCP2.6, 4.5, 6.0 and 8.5), CMIP6 data are becoming available for 5 cross-WG agreed scenarios: SSP1-1.9, SSP1-2.6, SSP2-4.5, SSP3-7.0 and SSP5-8.5 (Chapter 1). The scenario-based projections presented in this section are up to 2100 [PLACEHOLDER FOR UP TO 2150], while projections based on warming levels are presented in Section 9.6.3.4 and projections beyond 2100 are discussed in Section 9.6.3.5. Due to current limited availability of CMIP6 model runs, results for SSP1-1.9 are not presented.

Up to 2050, consistent with AR5 and SROCC, GMSL projections exhibit little scenario dependence (Figure 9.30, Table 9.6) (*high confidence*), with *likely* sea-level rise between 1996-2014 and 2050 of 0.15-0.23 m under SSP1-2.6 and 0.16-0.26 m under SSP5-8.5. Beyond 2050, the scenarios increasingly diverge. Between 1996-2014 and 2100, conditional upon the modelling choices described in Section 9.6.3.2, GMSL is *likely* to rise by 0.47 m (0.33-0.64) m, 0.55 (0.40-0.71) m, 0.65 (0.51-0.81) m, and 0.73 (0.60-0.90) m under SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively. Conditional upon these modelling choices, it is *very likely* to rise by 0.21-0.84 m, 0.28-0.89 m, 0.41-1.00 m, and 0.50-1.07 m under SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively (Table 9.6).

As noted in Section 9.6.3.2, differences in ice sheet projections, particularly Antarctic ice sheet projections, are the primary driver of differences between the AR5 and SROCC projections for RCP2.6, RCP4.5 and RCP8.5, and the projections here for SSP1-2.6, SSP2-4.5 and SSP5-8.5. Employing SROCC-consistent ice sheets together with the updated thermal expansion, glacier, and terrestrial water storage projections yields *likely* GMSL projections for these three scenarios of 0.41 (0.33-0.53) m, 0.51 (0.41-0.68) m, and 0.80 (0.64-1.09) m, respectively (Table 9.A.2). These projections with SROCC-consistent ice sheets are close to indistinguishable from the SROCC GMSL projections, which, adjusted for the change in baseline, are 0.40 (0.26-0.56) m, 0.52 (0.36-0.69) m, and 0.81 (0.58-1.07) m, respectively.

The *likely* projections under the standard assumptions for SSP1-2.6 and SSP2-4.5 are quite similar to SROCC projections for RCP2.6 and RCP4.5, but are lower and narrower than for SSP5-8.5/RCP8.5. The greater disagreement for higher emissions scenarios highlights the presence of deep uncertainty in GMSL projections for SSP5-8.5, consistent with the assessment based on prior literature GMSL projections (Section 9.6.3.1). Compared to literature projections, the very likely ranges for SSP1-2.6 and SSP2-4.5 are consistent with the range of the lowest 5th and highest 95th percentile for RCP2.6 (0.21-0.84 m vs. 0.14-0.81 m) and RCP4.5 (0.28-0.89 m vs. 0.19-0.94 m) excluding studies incorporating MICI, whereas the very likely range for SSP5-8.5 is substantially narrower than the comparable literature range for RCP8.5 (0.50-1.07 m vs. 0.34 m-1.57 m). The divergence of the projections for SSP5-8.5 from comparable projections employing SEJ for ice sheets (Table 9.A.2; likely range of 0.71-1.63 m) again highlights the deep uncertainty under the highest emissions scenarios. This deep uncertainty motivates the examination of physical pathways that could, under a high-emissions scenario, lead to high-end GMSL outcomes above substantially above the very likely range of the primary projections (Box 9.3).

[PLACEHOLDER FOR DISCUSSION OF UNCERTAINTIES IN TIME RATHER THAN IN LEVEL E.G. WHEN WILL 1 M GMSL CHANGE OCCUR UNDER DIFFERENT SCENARIOS]

[START FIGURE 9.30 HERE]

Figure 9.30: Global mean sea-level change for 3 SSP-RCP scenarios (a) and regional relative sea-level (b-g) change for 3 scenarios between 1995-2014 and 2100 (m): (b, e) SSP5-8.5, (c, f) SSP2-4.5, and (d, g) SSP1-2.6; projection median (b-d) and 17th-83rd percentile range (e-g). [PLACEHOLDER TO UPDATE TO 5 SCENARIOS FOR FGD WHEN DATA IS AVAILABLE]

[END FIGURE 9.30 HERE]

[START TABLE 9.6 HERE]

Table 9.6: Global mean sea-level projections for 5 SSP-RCP scenarios, for total change and individual contributions, median values, (*likely*) and [*very likely*] ranges of the process-based model ensemble, for 1995-2014 to 2100 in meters. Average rates for total sea-level change in mm yr⁻¹.

	SSP1-1.9	SSP1-2.6	SSP2-4.5	SSP3-7.0	SSP5-8.5
Thermosteric		0.15 (0.13--0.17) [0.07--0.22]	0.19 (0.17--0.22) [0.11--0.28]	0.24 (0.23--0.26) [0.19--0.30]	0.29 (0.28--0.31) [0.25--0.34]
GIC		0.09 (0.07--0.11) [0.05--0.14]	0.11 (0.08--0.14) [0.06--0.16]	0.13 (0.10--0.16) [0.07--0.19]	0.14 (0.11--0.18) [0.08--0.22]
 AIS		0.12 (0.01--0.25) [-0.07--0.39]	0.11 (-0.00--0.24) [-0.08--0.36]	0.12 (-0.00-- 0.24) [-0.07-- 0.36]	0.11 (0.01--0.26) [-0.05--0.37]
 - EAIS		-0.04 (-0.10--0.02) [-0.15--0.08]	-0.05 (-0.11--0.01) [-0.15--0.07]	-0.05 (-0.10-- 0.01) [-0.15-- 0.05]	-0.05 (-0.11-- 0.01) [-0.16-- 0.05]
 - WAIS		0.12 (0.04--0.23) [-0.02--0.35]	0.12 (0.03--0.23) [-0.04--0.34]	0.13 (0.04--0.23) [-0.02--0.35]	0.12 (0.04--0.24) [-0.02--0.37]
 - AP		0.03 (0.01--0.05) [0.00--0.06]	0.03 (0.02--0.05) [0.00--0.07]	0.04 (0.02--0.05) [0.01--0.07]	0.04 (0.02--0.05) [0.01--0.07]
GrIS		0.07 (0.03--0.12) [- 0.00--0.19]	0.09 (0.05--0.14) [0.01--0.21]	0.12 (0.08--0.16) [0.05--0.22]	0.13 (0.09--0.19) [0.06--0.26]
TWS	0.04 (0.02-- 0.05) [0.01-- 0.06]	0.04 (0.02--0.05) [0.01--0.06]	0.04 (0.03--0.06) [0.01--0.07]	0.04 (0.02--0.05) [0.01--0.06]	0.04 (0.02--0.05) [0.01--0.06]
Total short term (2040)		0.14 (0.10--0.17) [0.08--0.20]	0.14 (0.10--0.17) [0.08--0.19]	0.14 (0.11--0.17) [0.08--0.20]	0.14 (0.11--0.18) [0.08--0.22]
Total medium term (2050)		0.19 (0.15--0.23) [0.12--0.27]	0.19 (0.15--0.24) [0.12--0.28]	0.19 (0.15--0.24) [0.12--0.27]	0.21 (0.16--0.26) [0.13--0.30]
Total long term (2090)		0.39 (0.28--0.52) [0.19--0.65]	0.45 (0.35--0.58) [0.26--0.71]	0.51 (0.41--0.64) [0.33--0.75]	0.59 (0.48--0.71) [0.39--0.83]
Total (2100)		0.47 (0.33--0.64) [0.21--0.84]	0.55 (0.40--0.71) [0.28--0.89]	0.65 (0.51--0.81) [0.41--1.00]	0.73 (0.60--0.90) [0.50--1.07]
Total (2150)					
Total (2300)					
Total rate medium term (2040-2060)		5.3 (4.2--6.6) [3.4- 7.8]	5.5 (4.3--6.8) [3.5- 8.3]	6.0 (4.8--7.2) [4.2--8.0]	6.6 (5.6--7.9) [4.9--9.1]
Total rate long term (2080- 2100)		6.6 (4.0--9.7) [1.6- 14.7]	7.8 (5.0--11.2) [2.3--15.5]	11.5 (8.8--14.2) [6.8--19.2]	13.1 (11.0--17.0) [9.4--20.3]

* The ISMIP6 emulator used to project ice sheet contributions does not account for temporal correlations, and so sample-wise uncertainties on rates would be overestimated. For the SOD, we report the trends in the 17th-83rd percentile and 5th-95th percentile estimates as the *likely* and *very likely* ranges. This approach effectively assumes perfect temporal correlation over 20-year periods, and may therefore underestimate uncertainties.

[END TABLE 9.6 HERE]

9.6.3.4 Updated sea-level projections for the next century based on temperature levels

This section presents the sea-level projections based on warming levels, defined as the 20-yr running mean of the GSAT increase of the projections with respect to the period 1850-1900 crossing a certain warming level at 2100. The contributions and general methodology for the projections are the same as in the scenario-based projections (Section 9.6.3.2). The warming levels included are [PLACEHOLDER:1.5], 2.0, 3.0, 4.0 and [PLACEHOLDER:5.0] °C GSAT increase, using all the available SSP-RCP model results, sorted into warming levels. [Due to current limited CMIP6 model availability, 1.5 and 5.0 degree scenarios are not presented.] A certain warming level may therefore include a mixture of one or more SSP-RCP scenarios. It is important to keep in mind that different pathways in GSAT can be followed to reach a certain temperature level, which affects the temporal evolution of the different contributors to sea-level change. For instance, there will be different ice sheet and glacier responses to a fast increase to a peak warming of 2 degrees in 2050 followed by a plateau or a decrease, compared to a gradual increase to the same level of warming in 2100. The sea-level projections presented might include different pathways to the same warming level in 2100, which is reflected in the uncertainty ranges, and should therefore be interpreted as an illustration of sea-level scenarios under a certain warming level. The sea-level projections presented here do not go beyond 2100, as the different pathways in temperature evolution will become increasingly relevant for longer time periods.

Between 1996-2014 and 2100, conditional upon the modelling choices described in Section 9.6.3.2, GMSL is likely to rise by 0.48 (0.33-0.65) m, 0.56 (0.41-0.71) m, and 0.69 (0.54-0.85) m along climate trajectories leading respectively to 2.0°C, 3.0°C, and 4.0°C increase in GSAT between 1850-1900 and 2081-2100 (Figure 9.31, Table 9.7). Conditional upon these modelling choices, GMSL is *very likely* to rise by 0.21-0.83 m, 0.31-0.90 m, and 0.44-1.00 m under 2°C, 3°C and 4°C warming, respectively. The projections for 2°C and 3°C are closely aligned with the projections for respectively SSP1-2.6 and SSP2-4.5, while those for 4°C are intermediate between those for SSP3-7.0 and SSP5-8.5. Consistent with the discussion in Section 9.6.3.3, there is deep uncertainty in the projections for temperature levels above 3°C, and alternative approaches to projecting ice sheet changes may yield substantially different projections in 4°C and 5°C futures. For example, the 5°C scenario described in Bamber et al. (2019) suggests a *likely* rise of 1.0 (0.7-1.6) m.

[START FIGURE 9.31 HERE]

Figure 9.31: Global mean sea-level change for 3 GSAT temperature levels (a) and regional relative sea-level (b-g). The change for 3 temperature levels between 1850-1900 and 2081-2100: (b, e) 4 °C, (c, f) 3 °C, and (d, g) 2 °C. 1996-2014 relative to 2100 projection median (b-d) and 17th-83rd percentile range (e-g). [PLACEHOLDER PANEL e AWAITING DATA UPLOAD]

[END FIGURE 9.31 HERE]

[START TABLE 9.7 HERE]

Table 9.7: Global mean sea-level projections for exceedance of 5 GSAT temperature levels, by sorting GSAT change in 2081-2100 w.r.t. 1850-1900, for total change, median values and (*likely*) [*very likely*] range of the process-based model ensemble, for 1995-2014 to 2100 in meters and the average rate in 2081-2100 in mm yr⁻¹.

	1.5 °C	2 °C	3 °C	4 °C	5 °C
Total change long-term (2100)		0.48 (0.33--0.65) [0.21--0.83]	0.56 (0.41--0.71) [0.31--0.90]	0.69 (0.54--0.85) [0.44--1.00]	
Rate 2081-2100 (mm/yr)		6.5 (3.9--9.9) [1.4--13.6]	8.2 (5.6--11.3) [3.4--15.6]	12.3 (9.7--15.5) [7.8--18.4]	

* The ISMIP6 emulator used to project ice sheet contributions does not account for temporal correlations, and so uncertainties on rates are overestimated.

[END TABLE 9.7 HERE]

9.6.3.5 Multi-century and multi-millennial sea-level commitments

Only a few integrated sea-level rise projection studies have examined GMSL rise projections along the extended RCPs beyond 2100 (Palmer et al.; Kopp et al., 2014, 2017; Nauels et al., 2017a, 2019; Bamber et al., 2019) (Table 9.8). There is substantial deep uncertainty on these timescales, associated with the deep uncertainty in the ice sheet response. For 2300, considering the 17th-83rd percentile projections of the two studies excluding MICI (Kopp et al., 2014; Nauels et al., 2017a, 2019; Palmer et al.) suggests a *likely* GMSL rise of 0.3-2.9 m under RCP2.6, and 1.7-6.8 m under RCP8.5. The sole study incorporating MICI (Kopp et al., 2017), which employs an Antarctic ice sheet model that is very sensitive to atmospheric forcing (Deconto and Pollard, 2016), projects 17th-83rd percentile rises of 0.8-2.3 m under RCP2.6 and 9.8-14.0 m under RCP8.5, highlighting the limited significance of MICI under low-emissions scenario but potentially great significance under high-emissions on multi-century timescales. Combining RCP8.5 for non-ice sheet components with GMST stabilization at 5°C for ice sheets, a recent SEJ study (Bamber et al., 2019) projects 17th-83rd percentile rises of 2.6-6.5 m, with a 95th percentile of 11.8 m, suggesting concerns about ice-sheet instability on these timescales. Based on these integrated literature projections, GMSL rise through 2300 under RCP2.6 will *likely* be between 0.3 and 2.9 m, the range defined by the lowest 17th and highest 83rd percentile projection, and *extremely likely* be below 4.7 m, the highest 95th percentile projection (*medium confidence*). There is *low confidence* in central projections of GMSL rise under higher emissions scenarios, and *medium confidence* that GMSL rise through 2300 will *extremely likely* be less than 7.0 m under RCP4.5 and 15.5 m under RCP8.5. To a first approximation, these conclusions about RCP2.6 and RCP8.5 can be extended to the three-century extensions of SSP1-2.6 and SSP5-8.5.

[START TABLE 9.8 HERE]

Table 9.8: Global mean sea-level rise projections for 2000-2300 from literature (m), for different RCP scenarios.

Study	Scenario	5 th percentile	17 th percentile	83 rd percentile	95 th percentile
Kopp et al. 2014	RCP8.5	1.0 m	1.8 m	5.2 m	7.4 m
	RCP4.5	0.0 m	0.7 m	3.5 m	5.3 m
	RCP2.6	-0.2 m	0.3 m	2.9 m	4.7 m
Nauels et al. 2017a	RCP8.5	-	3.4 m	6.8 m	-
	RCP4.5	-	1.8 m	2.3 m	-
	RCP2.6	-	0.8 m	1.4 m	-
Kopp et al. 2017 (MICI)	RCP8.5	9.1 m	9.8 m	14.1 m	15.5 m
	RCP4.5	2.1 m	2.8 m	6.0 m	7.0 m
	RCP2.6	0.5 m	0.8 m	2.3 m	3.0 m
Bamber et al. 2019 (SEJ)	RCP8.5 / 5°C	1.8 m	2.6 m	6.5 m	11.8 m
	2°C	0.5 m	1.2 m	3.6 m	5.3 m
Palmer et al. (submitted)*	RCP8.5	-	1.7 m	4.5 m	-
	RCP4.5	-	0.9 m	2.6 m	-

	RCP2.6	-	0.6 m	-2.2 m	-
Nauels et al. (2019)	Historical emissions thru 1991	-	0.5 m	0.7 m	-
	Historical emissions thru 2016	-	0.7 m	1.1 m	-
	Pledged emissions thru 2030	-	0.8 m	1.4 m	-

* Palmer et al. (submitted) 5th-95th percentile of simulated projections are constructed to be analogous to AR5/SROCC *likely* ranges and so are presented here as 17th-83rd percentile projections.

[END TABLE 9.8 HERE]

Only one study since AR5 (Clark et al., 2016) has produced projections for the integrated GMSL commitment on multimillennial timescales (Figure 9.32), which are indexed here based on the peak GMST reached in the period following cessation of emissions. At all levels of warming considered, Antarctica is projected to be the single largest contributor to the sea-level commitment, with a share ranging from about 40% (1.1°C peak warming) to 90% (2.7°C peak warming), then declining somewhat as Greenland passes a critical transition and its overall contribution to GMSL rise increases. The projections show a 2000-year commitment of ~1.1 m associated with 1.0°C of peak warming, increasing to ~3 m/°C between 1.0 and 2.0°C of peak warming, and to ~6 m/°C at higher levels of peak warming. Sea-level rise continues to increase after 2000 years, leading to a 10,000-year commitment of 1.9 m associated with 1.0°C of peak warming, 5 m associated with 1.5°C of peak warming, and 12 m associated with 2.0°C of peak warming, with an average slope of about 8 m/°C between 2.0°C and 6.0°C of peak warming. These estimated commitments are substantially higher than that assessed by AR5, which concluded with *low confidence* that the multi-millennial commitment was about 1 to 3 m/°C (Church et al., 2013a).

Paleo-sea level reconstructions (Chapter 2, Section 9.6.2) can provide a limited degree of additional information about the multimillennial sea-level commitment. Heuristically, comparing GMST and GMSL during past multimillennial warm periods can provide an indicative metric for the equilibrium response. However, of past warm periods, only during the LIG is GMSL known with enough confidence to provide a useable indicator, and differences in orbital and AMOC forcing prevent using it as a direct analogue. During the LIG, GMSL was *likely* 5–9 m higher than today and GMST was *likely* 1-2°C warmer than pre-industrial, with greater warming over the poles (Chapter 2). In addition, the forcing provided to marine-based ice sheets by a reduced deglacial AMOC may have been stronger than during the last deglaciation (Clark et al., in press). Given these ambiguities, there is *medium confidence* that the sea-level commitment for 1-2°C is below LIG GMSL. An additional paleo constraint comes from the early Eocene Climatic Optimum, where ice-free conditions and a warmer ocean (which would lead to a *likely* GMSL rise of 71-74 m if they occurred today, based on current ice volume and projected global mean thermal expansion; (MORLIGHAM et al.; Morlighem et al., 2017; Fischer et al., 2018; Farinotti et al., 2019)) were associated with a *very likely* 12-16°C GMST anomaly (Chapter 2), suggesting an average sea-level commitment for warming from present conditions to ~14°C of ~5 m/°C (*low confidence*).

On the basis of these studies, single-ice sheet studies finding multimillennial nonlinear responses from both GrIS and AIS (Sections 9.4.1.2, 9.4.2.2), and the underlying physics, it is concluded that the sea-level commitment is nonlinear in peak warming (*medium confidence*). The multimillennial sea-level commitment is consistent with the AR5 assessment of 1 to 3 m/°C (*low confidence*) up to global-mean peak warming of 2°C on 2000-year timescales and 1.5°C on 10,000-year timescales (Figure 9.32). Above 2°C, the 2000-year commitment is ~6 m/°C (*low confidence*), while the 10,000 year commitment is more nonlinear in peak warming, and may in some temperature ranges (e.g., between 1.5°C and 2.5°C) exceed 13 m/°C (*low confidence*). Although thermosteric sea level may start to decline slowly about 2000 years after the end of

emissions, the slower responses from GrIS and AIS mean that GMSL will under most scenarios continue to rise throughout the next 10,000 years (*medium confidence*).

The approximately linear relationship between cumulative carbon emissions and peak GMST increase (Chapter 5), combined with the relationship between sea-level commitment and peak GMST increase, allows the sea-level commitment to be linked to carbon budgets (Clark et al., 2018). Given a *likely* TCRE of 1.0-2.2 K EgC⁻¹ (Section 5.5.1), the relationships above yield a sea-level commitment of 1.2 to 5.2 m EgC⁻¹ up to 2°C (~900-2000 GtC) on 2000-year timescales and to 1.5°C (~700-1500 GtC) on 10,000-year timescales. At higher levels of emissions, the 2,000-year commitment is ~6 to 13 m EgC⁻¹ on 2000-year timescales, and in some temperature ranges may exceed 13 to 29 m EgC⁻¹ on 10,000-year timescales (*low confidence*). One study (Clark et al., 2016) relates that the human carbon footprint of about 470 Gt C by 2000 to have already committed a GMSL rise of ~1.7 m (range of 1.2 to 2.2 m), while if another 470 Gt C is released, it will result in a further rise of ~9 m, illustrating the strong nonlinearity to forcing of the model projections. However, the particular relationship between emissions and GMSL commitment in this study relies on the climate sensitivity of the particular model used in that study and is therefore assigned *low confidence*.

Short-lived climate forcers (Chapter 6) are associated with a sea-level commitment that lasts substantially longer than their atmospheric forcing, although not as long lived as the sea-level commitment associated with CO₂ emissions. For example, (Zickfeld et al., 2017) find that about 70% of the thermosteric sea-level rise associated with methane forcing would persist 100 years after the elimination of methane emissions and 40% would persist for over 500 years.

A small number of modelling studies have examined the reversibility of the multimillennial sea-level commitment under carbon dioxide removal, albedo modification or local ice-shelf engineering. Rapid reversion to pre-industrial CO₂ concentrations has been found to be ineffective at fostering regrowth of the AIS but did stop ice sheet mass loss within two centuries, which reduces the sea-level commitment (Deconto and Pollard, 2016). Altering sub-ice shelf bathymetry (Wolovick and Moore, 2018) or triggering an ice-shelf advance through massive snow deposition (~2 mm yr⁻¹ SLE) (Feldmann et al., 2019) might interrupt MISI (Box 9.3) and thus reduce sea-level commitment. A geoengineered reversion to pre-industrial GrIS temperatures effectively stopped mass loss in Greenland but led to minimal regrowth (Applegate and Keller, 2015). A reversion to pre-industrial CO₂ after a quadrupling of CO₂ concentrations was found to stop thermosteric sea-level rise in about 80 years, with a subsequent thermosteric sea-level fall that was substantially slower than the preceding rise, under most parameterizations taking over a millennium to reverse (Ehlert and Zickfeld, 2018). Thus, the few studies conducted to date suggest that geoengineering may be effective at reducing the yet-to-be-realized sea-level commitment but is ineffective at reversing GMSL rise (*low confidence*).

[START FIGURE 9.32 HERE]

Figure 9.32: Global mean sea-level commitment as a function of global mean surface temperature from models and paleo data. Grey dots and ranges indicate constraints from the Common Era, the LIG, the MPWP, and the early Eocene Climatic Optimum. Blue, green and purple dots and fits indicate respectively the 10,000-year sea-level commitments, 2,000-year sea-level commitments and 300-year sea-level commitments from (Clark et al., 2016). Boxes indicate the scenario ranges from CMIP5.

[END FIGURE 9.32 HERE]

[START BOX 9.3 HERE]

BOX 9.3: High and low end sea level rise projections

Projections of global mean sea-level change are affected by two substantial areas of uncertainty. First is that which surrounds future climate trajectory, specifically the rate and magnitude of changes, and the temporal shape of these changes (Chapter 3). Second is the response of individual components of the physical system, which depends specifically on the processes that are included in the models used for projections, and the way that these processes interact in a manner that can allow natural feedbacks to propagate. A matrix of possible futures therefore exists. The *deep uncertainty* in sea-level projections for the late-21st century and beyond, as well as the upward shift of the central range of literature sea level rise projections since AR5 (Garner et al., 2018)(Section 9.6.3.1) highlights the importance of assessing high-end, physically plausible sea-level projections, even if they are currently viewed as low probability. A deliberate effort to explore high-impact, physically plausible sea-level scenarios of low but deeply uncertain probability, and the early indicators that would be associated with the occurrence of such scenarios, thus provides a complementary approach that can be valuable for risk assessment (Miller et al., 2013; Kopp et al., 2017; Hinkel et al., 2019).

Here we assess the mechanisms by which contrasting sea level projections may arise. In one possible pathway, minor environmental changes from curtailed emissions combine with self-stabilising (negative) process feedbacks and the system responds linearly to external drivers. Such a scenario would be characterised not just by relatively modest changes to Global Mean Sea Level (GMSL), but also by comparatively low uncertainties associated with these changes. An alternative pathway exists, however, in which high emissions and exacerbated environmental responses combine with self-sustaining (positive) earth system feedbacks and a tendency for an integrated sea-level response that is above-linear with respect to the initial forcing. In contrast with the ‘low-end’ case, this ‘high-end’ pathway is characterised both by higher GMSL and also considerably greater uncertainty.

A pathway for low-end sea-level rise

Substantial global mitigation endeavours could lead to an emissions trajectory that follows emissions scenarios consistent with the more aggressive Paris Agreement targets, leading to GSAT stabilizing around 1.5°C-2°C above pre-industrial levels by 2100, and physical uncertainties associated with the different contributions to GMSL rise could all break toward the low end of their *likely* ranges. By 2100, oceanic heat uptake would lead to modest warming of global oceans and about 0.13 m GMSL increase from thermosteric expansion. Mountain glaciers might contribute 0.07 m to GMSLR. Groundwater withdraw could contribute another 0.03 m. In Greenland, ice-sheet surface melt and ocean-driven retreat of marine-terminating glaciers might contribute only 0.03 m to GMSL, due to stabilizing feedbacks such as refreezing of surface meltwater within the snowpack and meltwater-induced cooling of the proximal sea surface (Bakker et al., 2017; Caesar et al., 2018), which could reduce local air temperature increases (Golledge et al., 2019)). Similarly, meltwater entering the Southern Ocean from Antarctica may reduce warming of the proximal ocean by up to 0.3°C under RCP 2.6 (Lago and England, 2019), enhancing sea ice expansion and mollifying atmospheric warming (Bronse laer et al., 2018; Golledge et al., 2019). Where ice margins are already retreating, rapid bedrock rebound (observed up to 41 mm yr⁻¹; Barletta et al., 2018)) as well as gravitational lowering of relative sea level may slow the rate of recession or allow future stabilization (Gomez et al., 2015; Larour et al., 2019). The result, combined with a modest increase in snowfall in East Antarctica, would be little net GMSL contribution from Antarctica. The resulting total would bring GMSL rise to less than 0.3 m over the course of the century, at an average rate below the average rate of GMSL rise since the early 1990s – but achieving such an outcome would require both stringent greenhouse gas mitigation and physical luck.

A pathway for high-end sea-level rise

Probabilistic projections based incorporating ice-sheet structured expert judgment (Bamber et al., 2019) suggest that GMSL rise of 2.0 ± 0.1 m over the course of this century can *very likely* only be achieved with a Greenland contribution *very likely* in excess of 25 cm (and *likely* in excess of 46 cm) and an Antarctic contribution in excess of 18 cm (and *likely* in excess of 42 cm). Achieving such large ice-sheet contributions would require a high-emissions trajectory (e.g., SSP5-8.5), leading to substantial atmospheric and ocean

warming. Warmer atmospheric and oceanic conditions would have to precipitate a cascade of feedbacks that produce substantial contributions to GMSL from the Greenland and Antarctic ice sheets, exceeding those simulated by the ISMIP6 ensemble (Edwards, submitted; Seroussi, submitted). By comparison, in a 4°C warmer world, the primary projections in this chapter, based on the ISMIP6 ensemble, ascribe a ~5% probability to GMSL rise in excess of 1.0 m, and if all individual components contributed at the upper end of their *very likely* ranges, would lead to a GMSL rise of 1.2 m. Alternative approaches considered lead to 95th percentile projections under SSP5-8.5 of 1.3 m (using LARMIP projections of Antarctic contributions), 1.5 m (using SROCC-consistent ice sheet projections), or 2.3 m (using structured expert judgement for ice sheet projections).

In Greenland, the ‘elevation–mass balance feedback’ (Robinson et al., 2012; Pattyn et al., 2018) means that surface melting lowers the ice surface and exposes it to warmer air temperatures, which in turn promotes greater melt. Concurrence between model-based studies suggests *high confidence* that such a mechanism exists. Modelling experiments indicate that inclusion of this process can increase contributions to GMSL by 8–11% by 2100 and 24–31% by 2300, depending on emissions scenario (Vizcaino et al., 2015). Surface lowering of ice sheet margins leads to a greater change in ice volume than interior lowering due to changes in surface albedo from snowline fluctuations, steepening of the ice surface gradient from increased runoff, and consequent enhanced flow (Vizcaino et al., 2010; Vizcaino et al., 2015). In marginal areas surface mass balance decreases 10–100 mm yr^{−1} for every metre of surface lowering. Due to the logarithmic relationship between atmospheric warming and the time taken for a given amount of ice loss, initial warming leads to a rapid decrease in the timescale of mass loss such that a rise from 1 to 2°C above the melting threshold reduces the timescale for 10% volume loss from 2040 (1320–4120) years to 1140 (760–2210) years, whereas a rise from 4 to 5°C above the melting threshold reduces the timescale for 10% volume loss from 610 (410–1150) years to 500 (330–930) years (Levermann and Winkelmann, 2016).

Based on consistency between multiple studies there is *medium confidence* that that over decadal to multi-decadal timeframes, processes such as subglacial melt at marine margins (Wilson et al., 2016; Slater et al., 2018) enhanced calving and dynamic thinning (Hill et al., 2018), as well as surface darkening (Cook et al., 2019a; Ryan et al., 2019), reduced cloud cover (Hofer et al., 2017) and reduced snowpack permeability (Machguth et al., 2016b; MacFerrin et al., 2019) may exert a substantial control on the Greenland Ice Sheet response to atmospheric warming. Surface darkening due to increased growth of pigmented algae on the ice sheet surface increased summer melt by 6–9 % during 2016 (Cook et al., 2019b), whereas darkening from snowline-driven albedo changes are associated with 5 times higher surface melting than that arising from hydrological or biological processes alone (Ryan et al., 2019). Reductions in cloud cover above the Greenland Ice Sheet may lead to 27 ± 13 Gt additional summer melt for every percent change in cloud area (Hofer et al., 2017). Century-end projections of surface meltwater that runs off the ice sheet under simulated RCP 4.5 and 8.5 conditions may be underestimated by a factor of 2 in models that do not include the presence and growth of low-permeability ice slabs (MacFerrin et al., 2019) but it is considered *unlikely* that meltwater-enhanced basal sliding will play a substantial role in increasing the dynamic component of ice sheet discharge, except perhaps over diurnal to seasonal timescales (Schoof, 2010; Flowers, 2018).

AR5 concluded that only the collapse of marine-based sectors of the Antarctic ice-sheet could cause GMSL rise substantially above the 0.98 m *likely*-range upper bound for RCP 8.5, and expressed *medium confidence* that such collapse would not exceed several tenths of a meter during the 21st century. By revising the Antarctic contribution, the SROCC (Oppenheimer et al., 2019) raised the *likely*-range upper bound for RCP 8.5 to 1.1 m, indicating increased likelihood not just in the possible occurrence of such collapse under RCP 8.5, but also that the rate of collapse would be sufficiently high to produce substantial sea-level impacts within the current century. New literature, not assessed in the SROCC, now provides a more complete basis for assessing the likelihood of a high-magnitude, decadal-scale, contribution from the Antarctic ice sheet.

The Antarctic Ice Sheet is most substantially influenced by oceanic (rather than atmospheric) forcing. Furthermore, its response is controlled to a large extent by the topographic configuration of underlying bedrock, such that a ‘marine ice sheet instability’ (MISI) exists in which thinning and flotation of ice at the grounded-floating transition results in landward migration of the grounding line (Weertman, 1974; Thomas and Bentley, 1978; Schoof, 2007; Pattyn et al., 2012). Where the bedrock deepens toward the interior of the

ice sheet, the grounding line encounters ever thicker ice. Thicker ice results in increased ice flux, which further thins (and eventually floats) the ice sheet margin, resulting in further retreat into deeper water (and thicker ice) and so on (Figure 9.1). The rate at which MISI proceeds is controlled by the thermal forcing provided by the adjacent ocean, with one model indicating a four times increase in sea-level contribution in response to ten times increase in ocean-driven sub-ice shelf melt rate (Arthern and Williams, 2017). In Thwaites Glacier, the importance of ocean thermal forcing may increase as the grounding-line proceeds further toward the interior of the ice sheet (Seroussi et al., 2017), although other model experiments indicate a contribution to sea-level that is more linearly related to the mean ocean temperature anomaly (Alevropoulos-Borrill et al., 2019). Despite some model dependency, agreement between models suggests at least *medium confidence* that, in the context of a high-end sea-level pathway, the response of Thwaites Glacier is of particular importance, perhaps responsible for a three times increase in future SLR from West Antarctica compared to a pathway in which its retreat is prevented (Cornford et al., 2015). MISI may halt in areas where underlying bedrock begins to rise, or if the lateral drag provided by topographically-pinned ice shelves exerts sufficient buttressing to stop further grounding-line retreat (Pegler, 2018). There is *medium confidence* that Pine Island Glacier may already have retreated in part due to ocean-forced MISI (Rignot et al., 2014; Christianson et al., 2016), and its recent slowdown in retreat rate has been attributed to colder ocean conditions 2012-2013 compared to 2000-2011 (Milillo et al., 2017; Konrad et al., 2018). Based on simulations using three different ice-sheet models, (Favier et al., 2014) conclude that even after stabilisation, Pine Island Glacier will remain in a state of mass imbalance three to six times higher than prior to perturbation.

There are two additional mechanisms that might lead to high-end sea-level contributions from Antarctica. First, a positive ice-ocean feedback is thought to exist in which ice sheet meltwater can stratify the proximal oceanic water column and allow deeper and warmer water to penetrate further into ice-shelf cavities (Menviel et al., 2010; Bronselaer et al., 2018), where it could increase basal melting by 9-34% (95% uncertainty range) under a simulated RCP 8.5 scenario (Bronselaer et al., 2018). In one model this mechanism nearly doubles the 21st century Antarctic sea-level contribution, increasing it from 24 to 46 mm (RCP 4.5) and from 78 to 140 mm (RCP 8.5) (Golledge et al., 2019). There is *medium confidence* that meltwater discharge to the Southern Ocean could also be associated with a time-averaged steric contribution to local sea-level of 0-1 mm yr⁻¹ in basins where warmer water is trapped (Golledge et al., 2019; Lago and England, 2019).

A second mechanism has been proposed in which the collapse of ice shelves exposes vertical cliffs of grounded ice that collapse in a self-sustaining manner (Deconto and Pollard, 2016). Using their ‘Marine Ice-Cliff Instability’ (MICI) parameterization, DeConto and Pollard (2016) suggested that the Antarctic Ice Sheet could contribute ‘more than a metre’ to GMSL by 2100 and more than 13 m by 2500. Independent analysis has subsequently shown that, compared to a ‘no-MICI’ simulation, the inclusion of MICI can increase the mean 2100 CE SLE contribution from Antarctica by 20, 42, or 61 cm under RCPs 2.6, 4.5, 8.5 respectively, or by as much as 35, 82, or 118 cm under the same scenarios if the upper bound of the 90% confidence interval is taken (Edwards et al., 2019). The MICI mechanism requires an initial trigger of ice-shelf hydrofracture through surface meltwater ponding and crevasse wedging, followed by the establishment of subaerial ice cliffs in excess of 90 m high. Other studies suggest there is *low confidence* in whether or not this mechanism would take place either this century or in the future as a) it is unclear whether sufficient surface meltwater could be generated and maintained on Antarctic ice shelf surfaces under projected 21st century conditions (Trusel et al., 2015), particularly in light of meltwater-induced high-latitude sea-surface and atmospheric cooling (Bronselaer et al., 2018; Golledge et al., 2019) (Bell et al., 2018) and b) the localized nature of ice shelf hydrofracture tends to place a ‘speed limit’ on ice shelf disintegration (Robel and Banwell, 2019), restricting the rate at which a grounded ice cliff might be formed. In a model that explicitly calculates the complete stress field of vertical ice cliffs (rather than parameterizing a collapse rate) the 90 m high threshold required for the proposed MICI mechanism could only be achieved in areas where the entire ice shelf collapses in less than an hour (Clerc et al., 2019), which is considered unrealistic given the observational evidence for much slower ice shelf disintegration (e.g. years, (Cook and Vaughan, 2010)), even in cases considered to be unusually fast (e.g. weeks, (Robel and Banwell, 2019)).

[END BOX 9.3 HERE]

9.6.4 Extreme sea levels: Tides, surges and waves

9.6.4.1 Past changes

Long-term trends in extreme still water levels (combining RSL, tide and surge; see Box 9.1) as observed by tide gauges primarily follow trends in RSL (Menéndez and Woodworth, 2010), consistent with AR5 (Church et al., 2013a) (Figure 9.33). The robustness of estimates of long-term trends in extreme still water levels increases with record length. The continued development of the GESLA tide gauge database (Woodworth et al., 2016) in conjunction with data archaeology (Talke and Jay, 2013) extends the tide gauge records to earlier periods. Modelled ESL data sets at global or regional scales, such as the Global Tide and Surge Reanalysis (GTSR) (Muis et al., 2016), currently cover only a few decades and include important uncertainties, due to the model limitations and inhomogeneities of atmospheric hindcast data, and are therefore not yet long or accurate enough to be useful in assessing long-term trends in extreme still water levels.

Observations show that RSL rise is the primary driver of changes in extreme still water levels at most locations and has led to a median 165% increase in high-tide flooding (Figure 9.33) (*high confidence*) (Menéndez and Woodworth, 2010; Mawdsley and Haigh, 2016; Marcos and Woodworth, 2017). However, some locations exhibit substantial differences between long-term RSL trends and extreme still water levels (*high confidence*). For example, at New York City, the 10% annual probability ESL has increased by 0.72 m since the mid-19th century, compared to 0.44 m of RSL rise over the same period (Talke et al., 2014). Tide gauges in the German Bight exhibit trends over 1951–2008 in high percentiles of still water levels (after removing RSL rise) of 0.9–3.7 mm yr⁻¹ for the 95th percentile and 2.4–4.8 mm yr⁻¹ for the 99th percentile (Dangendorf et al., 2013). These larger trends in extreme still water levels may be a result of changes (either positive or negative) in the surge component or in the tidal component, and can include non-linear interactions between tide, surge, and RSL (Arns et al., 2015; Schindelegger et al., 2018). In addition to any long-term trends, extreme still water levels also exhibit significant and regionally coherent multi-decadal variability in the order of up to several centimetres (Marcos et al., 2015; Wahl and Chambers, 2015).

Secular, non-astronomically driven changes in tidal amplitude and tidal phase have been observed in paleo-oceanographic proxies and tide-gauge records (Woodworth, 2010; Müller et al., 2011; Mawdsley et al., 2015; e.g., Devlin et al., 2017; Schindelegger et al., 2018), and in numerical modelling re-analyses (Egbert et al., 2004; Uehara et al., 2006; Arbic et al., 2008; Wilmes and Green, 2014). Tide gauge data show that principle tidal components have varied in amplitude in the order of 2% to 10% per century (Jay, 2009; Ray, 2009). Tidal changes may be influenced by multiple interacting processes on different temporal and spatial scales (Talke and Jay, 2020; Idier et al., submitted), including strong drivers of a localised anthropogenic nature (e.g. port development, dredging, flood defences, land reclamation). However, attributing direct causality and proportionality (with only modest correlations and instrumental error) of the respective drivers remains challenging (Haigh et al., 2019). RSL rise causes water depth based alterations to the resonant characteristics of the basin, changes the bottom friction and increases in the wave speed (Pickering et al., 2012) and remains one of the primary hypotheses for observed tidal changes. This is confirmed with global tidal models forced by observed RSL change, which explain 30–40% of the magnitude of measured M2 changes (Müller et al., 2011) and 80% of the sign of the tidal amplitude trends (Schindelegger et al., 2018). Combined, the observations and models indicate that RSL rise and direct anthropogenic factors are the primary drivers of tidal changes observed at tide gauge stations (*medium confidence*).

The evidence for changes in storm surges since 1900 and their influence on ESL varies for different locations. Along the US east coast, there has been an increase in an aggregate measure of the frequency of ESL events due to tropical cyclone changes since 1923 that can be statistically linked to changes in global average temperature (Grinsted et al., 2013), but this change in the overall distribution of events may not clearly emerge from the noise until around 2030 (Lee et al., 2017a). Observations from 1929 to 2013 indicate ESL increases in the southeastern US (RSL rise and tidal changes removed), though this effect is most regionally coherent in the winter, outside of hurricane season (Wahl and Chambers, 2015). The 99th percentile of surge-driven ESLs at 220 tide gauges (RSL rise removed) exhibits negative trends over the last four decades in parts of northern Australia, the US West Coast, and the European Atlantic coast, and positive

trends in the Gulf of Mexico (Mawdsley and Haigh, 2016). The Baltic Sea coastline exhibits significant trends over 1916-2005 in annual maxima only at the northernmost locations (Ribeiro et al., 2014). At century and longer timescales, geological proxies, such as overwash deposits in coastal lagoons or sinkholes, can be used to reconstruct past changes in storm activity (e.g., Brandon et al., 2013; Lin et al., 2014) and put recent events into historical perspective (e.g., (Brandon et al., 2015)). However, in the absence of rigorous statistical analyses integrating geological proxies (which most commonly reflect simply the occurrence of an event) and instrumental records of ESLs, there is *low confidence* in the current ability to directly compare these two data sets. At the local and regional scale, and for tropical cyclones, which are underrepresented in the observational records, hydrodynamic numerical model simulations forced by observed and synthetic storm data sets can be used to estimate changes in the storm surge activity at centennial to millennial time scales (e.g., (Reed et al., 2015)). As a result of the large regional variations and limited observational data, there is *low confidence* in observations of trends in the surge contribution to increasing ESL.

[START FIGURE 9.33 HERE]

Figure 9.33: Historical occurrences of minor extreme sea levels, defined as the 99th percentile of daily observed water levels over 1995-2014. (a) Percent change in occurrences over 1995-2014 relative to those over 1960-1980. (c-e) Annual mean sea level (blue) and annual occurrences of extreme sea levels over the 1995-2014 99th percentile daily maximum (yellow) at six selected tide gauge locations. [PLACEHOLDER FIGURE FROM SROCC WILL BE REDRAWN]

[END FIGURE 9.33 HERE]

In addition to extreme still water levels, ESL at the coast is also affected by surface wind-waves via wave setup. The most common observation platform for surface waves over the past ~30 years are in-situ ocean buoys. A number of studies indicate that changes in buoy have resulted in inhomogeneous datasets generally unsuitable for long term trend analysis (e.g., Gemmrich et al., 2011). An alternative approach is to use wave reanalysis and hindcast products to investigate effects of historical climate driven changes in wind-waves and their contribution to total water level at global scale (Melet et al., 2018; Reguero et al., 2019), but the drawback is the inhomogeneity among reanalysis and hindcast wave products associated with the temporal changes in assimilated data. Therefore, the use of these products is limited to understanding the relationships between seasonal, inter-annual to inter-decadal variability of climate indices and the global wind-wave climate (Marshall et al., 2015a, 2018; Kumar et al., 2016; Stopa et al., 2016). The satellite altimeter-derived wave record now spans over 33 years (1985 to present), and shows small increases in mean wave height over 1985-2018, with stronger increases in extreme wave heights (90th percentile), and strongest increases in the Southern Ocean (Young and Ribal, 2019). However, due to observational limitations, there is *low confidence* in trends in surface wind-wave conditions.

Detection and attribution of changes in ESL can be viewed as two problems: detection and attribution of changes associated with RSL change (Section 9.6.1.5), and detection and attribution of changes associated with other climatic factors. By approximating local RSL change as the sum of GMSL change attributable to human-caused emissions, other GMSL change and local RSL change, (Strauss et al., 2016) concluded that about two-thirds of high tide flooding between 1950 and 2014 would not have occurred in the absence of attributed GMSL rise. By contrast, detection and attribution of the portion of changes in ESL driven by climatic factors other than RSL rise remains a challenge, and only few studies have attempted to quantify the role of climate change in ESL events (e.g.(Mori et al., 2014)(Takayabu et al., 2015)). Nevertheless, the close relationship between local ESL and long term RSL change, combined with the attributability of GMSL change (Section 9.6.1.5), implies that observed changes in ESL can *very likely* be attributed to human-caused climate change (*medium confidence*).

9.6.4.2 Future changes

Currently, there are two distinct methods used to project future ESL changes. The “static” or “MSL offset” approach assumes that there are no changes in the distributions of the tidal, storm surge, and wave components and that mean RSL rise is the only relevant oceanographic driver for future changes in the coastal flood hazard. Under this assumption, statistical extreme value models can be used to parametrize observed or modelled ESLs in the past (accounting for past mean RSL rise) and shift the modelled distribution based on RSL projections. The results can be used to quantify future changes in the frequency (often expressed as “frequency amplification factors”) of ESL events (Buchanan et al., 2017; Vitousek et al., 2017; Wahl et al., 2017; Rasmussen et al., 2018; Oppenheimer et al., 2019). Results from these assessments are sensitive to the type of ESL probability distribution assumed (Buchanan et al., 2016; Wahl et al., 2017), whether still water levels or total water levels are analysed (Vitousek et al., 2017), and the magnitude and uncertainty of the projected RSL change (Slangen et al., 2017a; e.g., Wahl et al., 2017). However, all of the aforementioned studies conclude that even small to moderate changes in mean RSL can lead to hundred- to thousand-fold increases in the frequencies with which certain critical levels are exceeded; i.e. what is currently a 1-in-100 year ESL height (1% annual probability or 0.01 expected annual events) will be expected once or even multiple times per year in the future in many locations. Frequency amplification factors tend to be largest in tropical regions, due in part to higher RSL rise projections but primarily to the relative rarity of high extreme water levels in areas with little historical exposure to tropical or extratropical cyclones (*high confidence*) (Figure 9.34).

Like the RSL projections, frequency amplification factors increase with increasing emissions, and the differences among scenarios increase over time. Using the updated RSL projections (Section 9.6.3.3), the median increase in the amplification factor at 634 tide gauge stations is 11 times for SSP1-2.6, 11 times for SSP2-4.5 and 14 times for SSP5-8.5 by 2050 (*high confidence*) (Figure 9.34). This means that a historical 1% annual probability ESL will have an increased to an 11-14% annual probability. At 18% (SSP1-2.6), 18% (SSP2-4.5) and 19% (SSP5-8.5) of the 634 stations, the 1% historical annual probability has become an annual event by 2050. This confirms the conclusion of SROCC that, as a result of RSL rise, ESL that are currently rare will become increasingly common, with some regions experiencing historical 1% annual probability ESLs on an annual basis by 2050 (Oppenheimer et al., 2019). By 2100, the median amplification factor has increased 79 times for SSP1-2.6, 112 times for SSP2-4.5 and 251 times for SSP5-8.5, with respectively 46%, 54% and 71% of the stations at least yearly experiencing a currently 1% annual probability event (*high confidence*) (Figure 9.34).

The “dynamic approach” uses hydrodynamic models to simulate future ESLs (or their components) by considering atmospheric forcing from GCMs. This is computationally expensive and, while regional and local assessments have been performed (e.g., (Woth et al., 2006; Marcos et al., 2011)), it has only recently been successfully implemented at large spatial or global scales (Vousdoukas et al., 2017, 2018). This approach, further presented in Chapter 12, combines tide, surge, and wave models with probabilistic RSL projections to assess future changes in ESLs (expressed as total water levels including wave setup). Consistent with studies using the static approach, the global dynamic analysis (Vousdoukas et al., 2018) found that the current 1% average annual probability height will be exceeded, on average, every 2-50 years by 2050 for most regions in higher latitudes in both hemispheres, and expected more often (up to multiple times a year) in the tropics, under both RCP4.5 and 8.5, mainly because of RSL change. For 2100, changes are most pronounced along the majority of coastlines in the southern hemisphere, where present-day 1% average annual probability levels will be exceeded multiple times each year. Currently, individual models are used to simulate different components in this type of large-scale studies and hence not all non-linear interactions are fully captured. Overall, there is *high confidence* that RSL change will be the main driver in changes in future ESL return periods in most areas.

Comprehensive high-resolution ($\geq 1/8$ degree) numerical modelling studies provide evidence for spatially coherent increases and decreases in tidal amplitudes in shelf seas as a result of RSL rise ((Haigh et al., 2019), and references therein). Regional studies for Northwest Europe, Patagonia, Gulf of Maine, Australasia and Antarctica, and also for large coastal cities, have shown changes in M2 tidal amplitudes of up to around 10-20% per meter of RSL rise (Pickering et al., 2017). Studies of the European Shelf in scenarios assuming

coastal defences at the present day coastline show tidal changes with a medium to high level of agreement (Pickering et al., 2012; Pelling et al., 2013; Idier et al., 2017; Howard et al., 2019). Global tidal models have shown agreement on the magnitude of response and prolific global distribution of tidal changes in shelf seas (Pickering et al., 2017; Wilmes et al., 2017; Schindelegger et al., 2018). These studies also show that deep ocean-driven tidal changes are of the order of a few centimetres, whilst shelf seas globally in both semi-diurnal and diurnal dominated tidal regions can respond with tidal changes of tens of centimetres. There is *high confidence* that GMSL rise will be the primary driver of global tidal amplitude increases and decreases over the next 100-200 years, changing the baseline tide that ESLs are imposed upon. At local and regional scales, anthropogenic factors such as major land reclamation efforts (e.g. East China Sea) or differing national coastal management strategies (maintaining the present coastline position or managed retreat) are *very likely* to locally modulate the influence of GMSL rise on the tidal amplitude (*medium confidence*).

There is evidence that changes in tropical cyclone climatology will contribute to the frequency and magnitude of future ESL surge events, although estimates of the magnitude of this contribution range widely (Lin et al., 2012; McInnes et al., 2014; Little et al., 2015; Garner et al., 2017; Mori et al., 2019). In the Gulf of Mexico, changes in ESL due to tropical cyclone activity may be as important as SLR in enhancing future flood hazards (Marsooli et al., 2019). For the Korean Peninsula, the maximum change in the 100-year return height associated with typhoon-induced storm surges is 10% under 4 °C warming (Yang et al., 2018). The effects of projected changes in tropical cyclone intensity may be enhanced or offset in different locations by the effects of changes in tracks (Garner et al., 2017) (Chapter 11). Thus, while there is *high confidence* that ESLs driven by tropical cyclones will increase as a result of RSL rise, there is *low confidence* in projected changes in ESL driven by changes in tropical cyclone climatology.

Changes in ocean surface wave conditions can occur in response to changes in the frequency, intensity and position of forcing winds and storms, and result in changes in coastal morphology (Chapter 12) (Storlazzi et al., 2018). Changes in significant wave height can be divided into changes in mean wave height and changes in extreme wave height (for instance as a result of tropical cyclones). An assessment of 91 publicly available wave climate projection studies (Morim et al., 2018) concluded that projection uncertainty surrounding wave climate projections has been poorly sampled and that the current knowledge on projected changes in wave climate is limited to mean wave height. Qualitatively, studies agree on a future increase in the mean significant wave height across the Southern Ocean, tropical eastern Pacific and Baltic Sea, a projected decrease in the mean significant wave height over the North Atlantic and Mediterranean Sea, a projected increase in extreme wave heights in the Southern Ocean, and a projected decrease in extreme wave heights in the North Atlantic (*medium confidence*) (Morim et al., 2018).

A quantitative assessment of a community ensemble of global wind-wave projections (COWCLiP) (Morim et al., 2019) found robust projected changes of ~5-10% (positive or negative, depending on the region) in annual mean significant wave height, mean wave period, and/or mean wave directions along ~52% of the world's coastline that exceed internal climate variability only under RCP8.5 by 2100 (*medium confidence*). Ocean surface waves display considerable spatial variability at the coastal scale, which is poorly resolved by the climate models, as they are unable to resolve local scale geography (e.g., bathymetry, coastal orientation) and the forcing winds in extratropical or tropical storms. These higher-resolution features are required to accurately determine the contribution of storm waves to ESLs in the coastal zone and individual studies have sought to improve resolution to address these issues (e.g., Timmermans et al., 2017), but quantitative agreement on projected changes is not currently possible.

Correlations between changes in sea level-forced drivers of ESLs (mean sea-level changes and tidal changes) and changes in atmospherically forced drivers (ocean surface waves and surges) have only been considered in a few studies, although high surge and high waves co-occur along a majority of the world's coastlines (Marcos et al., 2019). Along the US east coast, ocean dynamic RSL change and change in power dissipation index (a proxy for North Atlantic tropical cyclone activity) are correlated across CMIP5 GCMs (Little et al., 2015), resulting in an increase in ESLs relative to analyses assuming independence of these changes. In the Irish Sea, dynamically coupled wave-tide modelling indicates that the tidal effect on wave propagation leads high water wave heights to be up to 20% higher than in an uncoupled analysis (Lewis et al., 2019). In the German Bight, RSL rise relaxes the breaking criterion of nearshore waves (assuming no geomorphological

response), allowing larger waves to propagate closer to the shore and leading to increased wave runup (Arns et al., 2017). Thus, depending on local geography, bathymetry, geomorphology and climatology, projections of ESL that do not consider correlations between and among sea level-forced and climate-forced drivers can differ strongly from coupled projections (*medium confidence*).

Combinations of storm surges and rain or river streamflow are often called “compound” surge-rain or surge-discharge events. Statistical studies have shown that high rain or streamflow often co-occurs with storm surge (Wahl and Chambers, 2015; e.g., Moftakhari et al., 2017; Ward et al., 2018; Couasnon et al., 2019). Dynamical modelling studies showed that the co-occurrence of flood-drivers raises ESLs at some locations in estuaries (Zhong et al., 2013; van den Hurk et al., 2015; Chen and Liu, 2016; Orton et al., 2018), particularly when hydrologic catchments are steep and cause high rainfall near the coast (Svensson and Jones, 2004). However, at other locations co-occurrence was found not to be important because the timing of streamflow is not coincident with the coastal peak storm surge (Orton et al., 2012; Klerk et al., 2015). Climate change will lead to increased compound surge-rain flooding by increasing RSLs and by increased storm surge, rain, or correlations between these flood-drivers (e.g., Moftakhari et al., 2017; Orton et al., 2018). RSL rise can impact tidal river reaches, extending its effects hundreds of kilometres inland, yet hydrodynamic modelling studies show that the assumption that RSL rise increases flood elevations equivalently in all locations can overestimate the increase in flooding at upriver locations (Zhong et al., 2013; Chen and Liu, 2016; Orton et al., 2018). However, the number of studies on compound events is still limited and therefore there is *low confidence* in understanding to which extent compound surge-rain events will change in response to RSL and climate change.

[START FIGURE 9.34 HERE]

Figure 9.34: Projected frequency amplification factors for the 1% average annual probability coastal extreme water level in 2050 (left) and 2100 (right), based on a statistical analysis applied to historical extreme water level return period curves following SROCC, using the regional sea-level projections of this chapter (Section 9.6.3.3).

[END FIGURE 9.34 HERE]

9.7 Robustness and Limitations of the Assessment

The process-based assessment of observed and projected change in the ocean, cryosphere and sea level here reveals knowledge and gaps in reconstructions, observations, models and process understanding. Those gaps that can be linked to an outcome that would revise the overall assessments are notable, as are those that belong to deep uncertainty and cannot have a probability assigned. Revisiting the updated assessments since AR5 and SROCC helps to gauge the robustness of understanding and quantitative assessments.

The largest uncertainties in future sea level and cryospheric change are related to the GrIS and AIS. Detection and attribution of changes in ice sheet extents is lacking due to both short observational records and inadequate models, for which initialization remains problematic (Sections 9.4.1.1; 9.4.2.1). While the ISMIP6 and InitMIP protocols provide simulations to estimate the uncertainties due to initial condition errors and emulators are employed to expand the scope of scenarios explored, remaining deep process uncertainties and a wide range of projections under different assumptions continue to reduce confidence levels. Future projections of GrIS and AIS changes remain highly uncertain, especially under strong warming, largely due to weaknesses in process understanding (Sections 9.4.1.2, 9.4.2.2). Lack of understanding of surface processes over Greenland, solid-Earth feedbacks, feedbacks between freshwater melt and regional climate, and the potential range of timescales for the marine ice shelf instability widen the uncertainty range of projected ice mass loss for a given scenario by a particular horizon. Potential additional ice loss mechanisms over Antarctica (hydrofracturing, cliff calving and surface meltwater) and warm water intrusions into ice shelf cavities inducing basal melting tend toward more rapid projected losses than those estimated here if these processes become significant. The importance of these processes in the past could provide physically

plausible levels of ice loss, but uncertainties in present reconstructions do not make these constraints very proscriptive. Critically, the uncertainty in ice sheet projections is the leading uncertainty in projections of future sea level for the second half of this century and beyond (Section 9.6.3).

Glacier inventory and projection uncertainty has been a significant source of past sea level budget uncertainty and remains a dominant uncertainty until mid-century. Emission scenario becomes the largest source of glacier change uncertainty by 2100 (Marzeion et al., submitted) just as the relative importance of glacier model uncertainty decreases over time (Section 9.5.1). Uncertainty in the natural variability of glacier response is small on the global scale but has substantial effect on small regional scales. The accuracy and representation of RGI 6.0 varies considerably along different glacierized regions depending on the quality and amount of available data (RGI Consortium, 2017a; Farinotti et al., 2019) but it is not possible to assess the uncertainty on a global scale (Marzeion et al., submitted). Particularly over the Southern Andes RGI 6.0 and other regional inventories disagree in glacier count and division (Barcaza et al., 2017; Zalazar et al., submitted), which influences the ice thickness distribution and projection uncertainty although more so for local less massive glaciers than larger glaciers or global inventories (Millan et al., 2019b; Zorzut et al., submitted). Improved depiction of glaciers through glacier inventories, mass balance observations, ice thickness observation and ice surface velocity in regions with limited or discontinuous data will improve projections (RGI Consortium, 2017b; WGMS, 2017; GlaThiDa Consortium, 2019; Millan et al., 2019a)

The Southern Ocean is difficult to access and this has limited discovery. Change in the Southern Ocean and adjacent shelves (Section 9.2.3.2) is intimately linked to the future of the AIS (Section 9.4.2.2), and the Southern Ocean is a complex region with both oceanic and atmospheric drivers affecting heat and carbon uptake and sea ice. The Southern Ocean has been a significant reservoir taking up heat and carbon, and projections predict this will continue (Section 9.2.2.1). However, recent wind stress changes have made it clear that the response of ocean eddies (which are poorly resolved or parameterized in climate models) and other ocean mixing processes to changing Southern Ocean forcing is a major reason why models disagree in this region (Section 9.2.3.2). The new high-resolution climate models show that SST, overturning, ocean heat content change and sea ice cover in this region is considerably improved in most models when compared to the coarser models that parameterize eddies (Newsom et al., 2016; Chassignet et al., submitted). Pan-Antarctic sea ice change is also lacking clear attribution to anthropogenic forcing and future projections remain uncertain, as regional patterns show sea ice gain while others show loss, and these regional changes are influenced by small-scale processes such as eddies, waves, upwelling regions and katabatic winds (Section 9.3.2). If the AIS begins to melt rapidly due any of the potential uncertainties mentioned above, the freshwater input is expected to have strong effects on both regional and global climate (Section 9.2.3).

Modelling uncertainties, particularly process representation and resolution, have been highlighted throughout this chapter. By including and excluding processes such as the MICI (Box 9.3; Section 9.4.2.2), its effects have been shown to accelerate ice mass loss and increase the *extremely likely* upper global sea level rise to 2.41 m instead of 1.57 m for RCP8.5 by 2100 (Section 9.6.3.1, Table 9.5). However, the probability of this occurrence is so low as to be inestimable, so the main projections in this chapter neglect these effects while noting that deep uncertainty persists, especially under strong emissions. These higher projection ranges can be kept in mind if and when MICI or other processes potentially contributing to high-end sea level are observed to be occurring (Box 9.3). The potential for abrupt changes in the AMOC (Section 9.2.3.1) is also demonstrated to be sensitive to model representation of overflows and convection in ocean models: higher resolution models tend to have AMOC systematically stronger, more variable, and less stable than coarser models do, although there are fewer high resolution models over fewer scenarios so it may be that these high-resolution models are not representative and some models do not adhere to this rule (Chassignet et al., submitted; Jackson et al., submitted). High resolution coupled simulations are beginning to demonstrate the interdependency of ocean processes and air-sea fluxes (Section 9.2.2) which is also part of the differences in projected changes in the ocean between low and high-resolution models. CMIP6 models have advanced representation of the ocean through parameterizations of unresolved processes (Fox-Kemper et al., 2011b, 2019; Li et al., 2016c; Qiao et al., 2016; Mathiot et al., 2017; Sun et al., 2017b; Reichl and Hallberg, 2018), and improved numerical schemes (Hu et al., 2015b) which are summarized for different models in Annex III. However, resolution in ocean models is a major limitation (Hallberg, 2013; Hewitt et al., 2017) with most CMIP6 models having resolutions that lack the ability to directly represent mesoscale, coastal and regional

1 shallow-water processes such as estuaries (Sun et al., 2017b; Chassignet et al., submitted; Jackson et al.,
2 submitted), the ocean beneath Antarctic ice shelves (Heuzé et al., 2013; Mathiot et al., 2017), the shelf seas
3 regions (Huthnance et al., 2009), Greenland fjords (Joughin et al., 2012), eastern boundary upwelling and
4 shelf seas (Sections 9.2.3.5). Coarse representation of these processes affects both regional impacts of
5 climate change and global consequences through water mass formation and ocean-ice interactions. For
6 permafrost (Section 9.5.2.2), the representation of thermokarst in land models is critical, and this process
7 could significantly accelerate future loss.

8
9 The observations of the ocean and cryosphere have been lacking through both short historical timeseries and
10 sparse observations (glaciers, permafrost, snow, lake and river ice (Section 9.5), deep ocean below 2000m
11 (Section 9.2.2)). In the case of snow, inconsistencies in different datasets leads to uncertainties in trends
12 (Section 9.5.3.2). The sea level budget for the 20th century has not been closed due a lack of observations
13 from many of the ocean and cryosphere components, but more recent periods show consistency among
14 observational budgets (Section 9.6.1, Cross-chapter Box 9.2). Paleo reconstructions of ice-sheet extent and
15 sea level, which have the potential to calibrate the sensitivity of ice sheets to global and regional climate
16 changes, remain sparse and noisy for past warm periods (Box 9.2; Section 9.6.2).

17
18 The cryospheric elements that are quick to respond to warming (sea ice area (Figure 9.16); permafrost
19 (Figure 9.24) and snow (Figure 9.26)) are found to track warming levels closely across many observations
20 and projections independent of scenario. This is a major new source of robustness since AR5 and SROCC,
21 which allows for rapid assessment of change in these elements based on warming level or cumulative CO₂
22 emissions.

23
24 Both AR5 and SROCC used the CMIP5 ensemble for assessing projections of global mean sea level rise and
25 regional sea level rise. These projections have recently been shown to be quantitatively accurate and precise
26 (Wang et al., submitted). The identification of the major constituents of sea level rise were clear even in
27 earlier assessment reports, but only in this report are the observations of constituents and total energy and sea
28 level rise shown to be quantitatively consistent. The CMIP6 family of models builds upon the experience of
29 the CMIP5 models, and the projections of ISMIP6, GlacierMIP, InitMIP, and LARMIP2 strengthen
30 understanding and together with the emulators of these simulations (Edwards et al., submitted) and statistical
31 approaches (Kopp et al., 2014, 2017) help quantify the probable projections and their uncertainties.

Frequently Asked Question

FAQ 9.1: Will human-induced changes in the oceans and frozen environments be reversible?

Some of the processes we are currently witnessing, such as sea level rise, the release of carbon stores into the atmosphere from thawing permafrost and other natural sources, as well as changes in ocean chemistry, are effectively irreversible on human timescales. Evidence from the distant past shows that some parts of the planet might take hundreds to thousands of years to fully adjust to changes in climate. This means that some of the consequences of human-induced climate change will continue for a very long time, even if atmospheric greenhouse gas levels and temperatures are stabilised or reduced in future.

Humans are changing the climate, and although many of the consequences may take hundreds or thousands of years to be fully realised, a large part of the responses in the oceans and in frozen environments (the cryosphere) will occur over decades to centuries. An important question therefore is how long it will take for these changes to eventually reverse, once levels of greenhouse gases in the atmosphere are stabilized or reduced by humans and natural processes. Records from the past can help us answer this question.

For at least the last 800,000 years, the Earth has followed cycles of gradual cooling followed by rapid warming caused by natural processes. During each cycle, sea level slowly lowered as large ice sheets grew, and then rose much more quickly as ice sheets melted. Ice sheets retreat more quickly than they grow because growth relies on the steady accumulation of falling snow that eventually compacts into ice, whereas retreat can be affected by instability processes in which rapid loss can occur once a particular threshold is reached (Box 9.3). Under certain circumstances, several threshold or instability processes may combine, forming a self-reinforcing cycle, or positive feedback. For example, during a period of relatively rapid global warming 55 million years ago (the ‘Paleocene–Eocene Thermal Maximum’), scientists think that warming was accelerated by the release of carbon dioxide or methane into the atmosphere from thawing of perennially frozen ground (permafrost) in Arctic and mountain regions (FAQ 5.2) or by the release of methane from frozen crystals (hydrates) in the deep ocean. If either were triggered in the future, global warming would increase rapidly, further melting permafrost areas, glaciers, and ice sheets, resulting in additional warming.

This kind of heightened response of the Earth system explains why relatively small increases in global average temperature have led to very large changes in the past, such as the collapse of the West Antarctic ice sheet and the loss of much of the Greenland ice sheet around three million years ago (the mid-Pliocene Warm Period) when global average temperature was only around 2°C above present. At that time, carbon dioxide concentrations in the atmosphere were about the same as they are today, but collapse of the ice sheets meant that global sea level was 10 to 30 metres higher than now, changing coastlines around the world. Analysis of geological records from many different warm periods of the past show that this relationship is remarkably consistent.

The implications of these insights from the past are that future changes, for example in terms of thinning or retreat of the large ice sheets, sea level rise, or thawing of permafrost, may lead to environmental changes that take thousands of years to return to their present state, even if global temperatures decrease within this or the next century. Warming of the oceans as well as ocean acidification (in which carbon dioxide dissolves into seawater) would also take centuries to reverse because the heat and CO₂ are transported down to the deep oceans, and take hundreds of years to circulate back to the atmosphere.

Changes in some elements of the cryosphere and oceans are more readily reversed, however. Mountain glaciers are much smaller than ice sheets, so they would regrow faster, taking centuries rather than millennia. Seasonal sea ice and snow cover in the northern hemisphere, once lost, would also probably return within a few years if air temperatures decreased to pre-industrial levels. Thawing of permafrost is expected to be reversible in most places, but the timescales to lock up the carbon again in permafrost and hydrates are very uncertain.

1 **[START FIGURE FAQ9.1, FIGURE 1 HERE]**

2
3 **FAQ 9.1, Figure 1:** Placeholder for figure with elements similar to above. Source: Nick Golledge

4
5 **[END FIGURE FAQ9.1, FIGURE 1 HERE]**

FAQ 9.2: How much will sea level rise in the next few decades?

Scientists estimate that, as of 2015, global average sea level was already about 20 cm higher than in 1900 and 5 cm higher than in 2000, and that sea level will rise by a further 7–25 cm by 2050. Two major reasons for this are the thermal expansion of water from increasing ocean temperatures and the melting of glaciers and ice sheets. Local sea level change can be higher or lower than the global mean, with the lowest rates of rise in formerly glaciated areas and the highest rates of rise in low-lying river delta regions.

Across the globe, sea level is rising. The rate of global mean sea level change was 3 millimetres per year from 1993 to 2017, which is about double the average rate observed during the 20th century. Scientists estimate the main causes of sea level rise for the earlier part of the last century were natural factors, such as the delayed response of mountain glaciers to northern hemisphere warming in the 19th century after the Little Ice Age. However, since at least 1970, human activities have been the dominant cause of global average sea level rise, and this will continue to be the case for centuries into the future.

Sea level rises when the oceans expand, either through warming of ocean waters, through the addition of water or ice previously stored on land, or through vertical movement of the land. The single largest source of current sea level rise is the expansion of ocean water due to global warming (39% of the rise observed since 1993), followed by melting of ice from mountain glaciers and ice caps around the world (24%). Melting of the two large ice sheets have played an increasing role since the 1990s, with average contributions of 16% from Greenland and 11% from Antarctica. Another source is changes in land water storage (5%), particularly due to extraction of water from groundwater reservoirs for consumption and irrigation.

The amount by which global mean sea level will rise over the next three decades will not depend much on whether or not greenhouse gas emissions are reduced. This is partly because the emissions are similar under all realistic scenarios in the near term and partly because oceans, glaciers and ice sheets mostly respond on slow timescales (decades, centuries and longer), so they will still be responding to warming that has occurred since the pre-industrial era. Scientists project there is a 90% or greater chance that global average sea level rise will be between 0.2 m and 1.1 m between the average level over 1995–2014 and the level in 2100 under all scenarios (SSP1–2.6 to SSP5–8.5). This means that the projected annual average rates of sea level rise range from the current rate of about 3 millimetres per year up to about four times faster than that.

Sea level change has large regional variations, so in many places local sea level change will be higher or lower than the global mean. Changes in ocean circulation and wind, which are the primary drivers of year-to-year local sea level variability, can lead to long-term local sea level change. In regions where large ice sheets covered the land during the last ice age, such as Scandinavia and North America, the land is still slowly rising up now that the downward pressure exerted by the extra weight of the ice sheets is gone. This local recovery is compensating global sea level rise in these regions and can even lead to a local fall in sea level. In regions just beyond where the former ice sheets reached, where the Earth bulged upwards, the land is now falling, and as a result local sea level rise is faster than the global rate. In many cities within low-lying delta regions, the land is rapidly subsiding (falling) because of human activities such as groundwater or fossil fuel extraction. In some cases, this subsidence reaches tens of millimetres per year, greatly amplifying local sea level rise. A further reason for local changes, which may sound counterintuitive, is that when an ice sheet melts, it has less gravitational pull on the ocean water. This causes sea level to fall close to the (now less massive) ice sheet while causing sea level to rise farther away because the melted ice sheet adds more water to the oceans overall. Melt from an ice sheet therefore affects sea level most in the opposite hemisphere from where the melting occurs.

Sea level rise will increase the frequency and severity of extreme sea level events at coasts. Scientists predict that in some regions, extreme sea level events that are currently expected once in 100 years will be experienced many times more frequently by 2050: they will occur every 2 to 50 years for most regions in high northern and southern latitudes, and they may occur multiple times per year in the tropics.

Beyond 2050, the amount by which sea level will rise is more uncertain. Some predictions show sea level rise continuing at a similar rate, while others show substantial acceleration. Present and future greenhouse

gas emissions play a big role in determining how much rise there will be beyond 2050. Under SSP1-2.6, there is a 95 % chance that sea level rise will remain below 4.7 m by 2300. However, if the climate warms substantially due to sustained future emissions above this level, the Antarctic ice sheet may begin to lose ice much more quickly than today, leading to much higher sea levels.

[START FIGURE FAQ9.2, FIGURE 1 HERE]

FAQ 9.2, Figure 1: Placeholder for figure with elements similar to some of the above. Source: Sophie Berger & Nick Golledge & Aimée Slangen

[END FAQ9.2, FIGURE 1 HERE]

FAQ 9.3: Will the Gulf Stream shut down, and what happens if it does?

The Gulf Stream is part of a large ocean current in the North Atlantic called the Atlantic Meridional Overturning Circulation (AMOC). Scientists estimate that this current is slowing down and will slow more in the future. Most climate models project that the AMOC could slow or shut down in the 22nd century under most future emissions scenarios. As the Gulf Stream is a warm current, it affects the weather around the tropics where it begins and the North Atlantic where it ends. If it slows further or shuts down, North America will see higher sea levels and Europe will see changes in weather and reduced warming relative to other regions. Similar processes may also affect the Southern Hemisphere.

The Gulf Stream is the biggest current in the North Atlantic Ocean and one of the largest currents in the world in terms of water carried. Each second, it transports about 30 billion kilograms of water northward past any location along its path along the east coast of North America. It is a warm current, with temperatures typically 5° to 15°C warmer than surrounding waters. As it moves northward, it carries a large amount of thermal energy (i.e., energy stored in warmer water) from its southern origins, and as it travels it releases this thermal energy to the atmosphere and surrounding water. Thus, the Gulf Stream warms poleward latitudes and cools equatorward and southward latitudes.

The Gulf Stream is one of many currents in the world that flow along the western side of ocean basins. The Kuroshio is a similar current in the North Pacific Ocean that travels northward along the east coast of Asia. However, the Gulf Stream is special in that it serves a second role: supplying the source for cold, deep waters that circulate back towards the equator. Cold water is denser than warm water, and in the Labrador and Irminger Seas at the northern end of the Gulf Stream path, cold wintertime conditions tend to make surface waters cold enough to sink to a depth of about 1500 metres. This cold water returns southward, spilling over features on the ocean floor and mixing with other deep waters in the Atlantic to form North Atlantic Deep Water. This overturning flow, with the Gulf Stream in the upper kilometre flowing northward and the colder Deep Water flowing southward, is called the Atlantic Meridional Overturning Circulation. There is no comparable deep overturning circulation in the North Pacific. So, the Gulf Stream completes both the horizontal circulation of the North Atlantic surface waters, driven by the winds, and the vertical overturning circulation, resupplying the water that sinks and becomes North Atlantic Deep Water.

The Kuroshio and the part of the Gulf Stream that transports warm surface waters result primarily from wind patterns and are not expected to change much. These currents will continue to transport thermal energy poleward from the equator much as they do now, although there will be some changes as the atmosphere and surrounding waters continue to warm under climate change.

However, the Atlantic Meridional Overturning Circulation is slowing and is expected to slow even more by the end of this century. This is mainly due to freshening (reduction in salt content) of the ocean waters due to meltwater from the Greenland ice sheet and Arctic sea ice and increased precipitation in the North Atlantic. This is making waters in the Labrador and Irminger Seas warmer and fresher and thus less likely to sink in this region. Since 2005, an array of moorings across the Atlantic at 26.5°N has been monitoring the Gulf Stream. The slowdown is not yet apparent in the data from this new monitoring system, but there are other signs that the overturning is slowing—for example, the region where the Gulf Stream's surface waters sink is not warming up as fast as its surroundings with global warming. Climate models show that this pattern of relatively cooler waters is consistent with a slowing of the overturning. Models also agree that under all emissions scenarios, this slowdown is expected to intensify by 2100 and beyond, with some models projecting a shutdown in the 22nd century. Paleoclimate evidence shows that the overturning has weakened in the past, especially during the ends of ice ages as ice sheets were melting and temperatures were rising.

What happens as the overturning slows down? The atmosphere adjusts somewhat to compensate for the changes in the ocean overturning by transporting some of the missing heat transport. But the reduced oceanic poleward heat transport means the North Atlantic will warm more slowly than other regions, which in turn means that parts of Europe will warm more slowly than other regions. As a result of the additional heat accumulated in the Gulf Stream area and the slowing of this current, sea level along the coast of North America will rise more than it would without these changes. Models indicate that weather patterns around

1 the Atlantic will be affected, with reduced precipitation in the mid-latitudes, changes to where strong
2 precipitation occurs in the tropics and Europe, and stronger storms in the North Atlantic Storm Track. We
3 have not been monitoring the changes in the AMOC long enough to be sure all of these model projections
4 are accurate, and present variability of the AMOC do not yet indicate a long-term shutdown.
5

6 The North Atlantic is not the only site of meridional overturning. Around Antarctica, the densest water in the
7 world is formed by freezing conditions that cool the water and form ice, adding salt to the remaining liquid
8 water. Recent studies have shown that melting of the Antarctic Ice Sheet can reduce the salinity of this water,
9 slowing the rate at which it sinks and thereby slowing a different, southern branch of the Meridional
10 Overturning. This slowdown also results in a change in surface conditions near the sinking regions which in
11 turn affects the weather locally and potentially across the Southern Ocean and the southern ends of South
12 America, Australia, and Africa. As in the North Atlantic, as this shutdown has not begun in earnest these
13 effects have not been observed, only projected by models. However, these changing weather patterns are not
14 as well understood as those that would result from a shutdown of the overturning portion of the Gulf Stream.
15

16
17 **[START FIGURE FAQ9.3, FIGURE 1 HERE]**
18

19 **FAQ 9.3, Figure 1:** [Placeholder for Gulf Stream Figure taken from RAPID/MOCHA website].
20

21 **[END FIGURE FAQ9.2, FIGURE 1 HERE]**
22

Reference

- Aas, K. S., Gislås, K., Westermann, S., and Berntsen, T. K. (2017). A Tiling Approach to Represent Subgrid Snow Variability in Coupled Land Surface–Atmosphere Models. *J. Hydrometeorol.* 18, 49–63. doi:10.1175/JHM-D-16-0026.1.
- Abraham, J. P., Baringer, M., Bindoff, N. L., Boyer, T., Cheng, L. J., Church, J. A., et al. (2013). A review of global ocean temperature observations: Implications for ocean heat content estimates and climate change. *Rev. Geophys.* doi:10.1002/rog.20022.
- Abrahamsen, E. P., Meijers, A. J. S., Polzin, K. L., Naveira Garabato, A. C., King, B. A., Firing, Y. L., et al. (2019). Stabilization of dense Antarctic water supply to the Atlantic Ocean overturning circulation. *Nat. Clim. Chang.* 9, 742–746. doi:10.1038/s41558-019-0561-2.
- Abram, N., Gatsutto, J.-P., and Prakash, A. et al. (2019). “Chapter 1: Framing and Context of the Report,” in *IPCC SR Ocean and Cryosphere*.
- Adler, R. F., Huffman, G. J., Chang, A., Ferraro, R., Xie, P.-P., Janowiak, J., et al. (2003). The Version-2 Global Precipitation Climatology Project (GPCP) Monthly Precipitation Analysis (1979–Present). *J. Hydrometeorol.* 4, 1147–1167. doi:10.1175/1525-7541(2003)004<1147:TVGPCP>2.0.CO;2.
- Agosta, C., Amory, C., Kittel, C., Orsi, A., Favier, V., Gallée, H., et al. (2019). Estimation of the Antarctic surface mass balance using the regional climate model MAR (1979–2015) and identification of dominant processes. *Cryosph.* 13, 281–296. doi:10.5194/tc-13-281-2019.
- Aguirre, C., Rojas, M., Garreaud, R. D., and Rahn, D. A. (2019). Role of synoptic activity on projected changes in upwelling-favourable winds at the ocean’s eastern boundaries. *npj Clim. Atmos. Sci.* 2, 44. doi:10.1038/s41612-019-0101-9.
- Akhoudas, C. H., Sallée, J.-B., Reverdin, G., Meredith, M. P., Naveira-Garabato, A., Haumann, A., et al. (submitted). Ventilation of the abyss in the Atlantic sector of the Southern Ocean. *Nat. Commun.* (submitted).
- Alevropoulos-Borrill, A., Nias, I. J., Payne, A. J., Golledge, N. R., and Bingham, R. J. (2019). Ocean forced evolution of the Amundsen Sea catchment, West Antarctica, by 2100. *Cryosph. Discuss.*, doi.org/10.5194/tc-2019-202.
- Alexander, P. M., LeGrande, A. N., Fischer, E., Tedesco, M., Fettweis, X., Kelley, M., et al. (2019). Simulated Greenland Surface Mass Balance in the GISS ModelE2 GCM: Role of the Ice Sheet Surface. *J. Geophys. Res. Earth Surf.* 124, 750–765. doi:10.1029/2018JF004772.
- Allen, R. J., Norris, J. R., and Kovilakam, M. (2014). Influence of anthropogenic aerosols and the Pacific Decadal Oscillation on tropical belt width. *Nat. Geosci.* 7, 270–274. doi:10.1038/ngeo2091.
- Allen, S. E., and Durrieu de Madron, X. (2009). A review of the role of submarine canyons in deep-ocean exchange with the shelf. *Ocean Sci.* 5, 607–620. doi:10.5194/os-5-607-2009.
- Alley, K. E., Scambos, T. A., Siegfried, M. R., and Fricker, H. A. (2016). Impacts of warm water on Antarctic ice shelf stability through basal channel formation. *Nat. Geosci.* doi:10.1038/ngeo2675.
- Alves, J. M. R., Peliz, A., Caldeira, R. M. A., and Miranda, P. M. A. (2018). Atmosphere-ocean feedbacks in a coastal upwelling system. *Ocean Model.* 123, 55–65.
- Anderson, J. B., Shipp, S. S., Lowe, A. L., Wellner, J. S., and Mosola, A. B. (2002). The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history: A review. *Quat. Sci. Rev.* 21, 49–70. doi:10.1016/S0277-3791(01)00083-X.
- Andresen, C. S., Kokfelt, U., Sicre, M.-A., Knudsen, M. F., Dyke, L. M., Klein, V., et al. (2017). Exceptional 20th century glaciological regime of a major SE Greenland outlet glacier. *Sci. Rep.* 7, 13626. doi:10.1038/s41598-017-13246-x.
- Angelopoulos, M., Westermann, S., Overduin, P. P., Faguet, A., Olenchenko, V., Grosse, G., et al. (2019). Heat and Salt Flow in Subsea Permafrost Modeled with CryoGRID2. *J. Geophys. Res. F Earth Surf.* 124, 920–937. doi:10.1029/2018JF004823.
- Applegate, P. J., and Keller, K. (2015). How effective is albedo modification (solar radiation management geoengineering) in preventing sea-level rise from the Greenland Ice Sheet? *Environ. Res. Lett.* 10, 84018. doi:10.1088/1748-9326/10/8/084018.
- Arbic, B. K., Mitrovica, J. X., MacAyeal, D. R., and Milne, G. A. (2008). On the factors behind large Labrador Sea tides during the last glacial cycle and the potential implications for Heinrich events. *Paleoceanography* 23, n/a–n/a. doi:10.1029/2007PA001573.
- Argus, D. F., and Peltier, W. R. (2010). Constraining models of postglacial rebound using space geodesy: a detailed assessment of model ICE-5G (VM2) and its relatives. *Geophys. J. Int.* 181, 697–723.
- Argus, D. F., Peltier, W. R., Drummond, R., and Moore, A. W. (2014). The Antarctica component of postglacial rebound model ICE-6G_C (VM5a) based on GPS positioning, exposure age dating of ice thicknesses, and relative sea level histories. *Geophys. J. Int.* 198, 537–563.
- Armitage, T. W. K., Bacon, S., Ridout, A. L., Petty, A. A., Wolbach, S., and Tsamados, M. (2017). Arctic Ocean surface geostrophic circulation 2003–2014. *Cryosph.* 11, 1767–1780. doi:10.5194/tc-11-1767-2017.
- Armitage, T. W. K., Kwok, R., Thompson, A. F., and Cunningham, G. (2018). Dynamic Topography and Sea Level

- 1 Anomalies of the Southern Ocean: Variability and Teleconnections. *J. Geophys. Res. Ocean.* 123, 613–630.
2 doi:10.1002/2017JC013534.
- 3 Armour, K. C., and Bitz, C. M. (2015). Observed and Projected Trends in Antarctic Sea ice. *US Clivar.*
4 doi:10.1002/rog.20022.
- 5 Armour, K. C., Eisenman, I., Blanchard-Whigglesworth, E., McCusker, K. E., and Bitz, C. M. (2011). The reversibility
6 of sea ice loss in a state-of-the-art climate model. *Geophys. Res. Lett.* 38. doi:10.1029/2011GL048739.
- 7 Armour, K. C., Marshall, J., Scott, J. R., Donohoe, A., and Newsom, E. R. (2016). Southern Ocean warming delayed by
8 circumpolar upwelling and equatorward transport. *Nat. Geosci.* 9, 549. doi:10.1038/ngeo2731.
- 9 Arns, A., Dangendorf, S., Jensen, J., Talke, S., Bender, J., Pattiaratchi, C., et al. (2017). Sea-level rise induced
10 amplification of coastal protection design heights. *Sci. Rep.* 7, 40171. doi:10.1038/srep40171.
- 11 Arns, A., Wahl, T., Haigh, I. D., and Jensen, J. (2015). Determining return water levels at ungauged coastal sites: a case
12 study for northern Germany. *Ocean Dyn.* doi:10.1007/s10236-015-0814-1.
- 13 Arthern, R. J., and Williams, C. R. (2017). The sensitivity of West Antarctica to the submarine melting feedback.
14 *Geophys. Res. Lett.* 44, 2352–2359. doi:10.1002/2017GL072514.
- 15 Asay-Davis, X. S., Cornford, S. L., Durand, G., Galton-Fenzi, B. K., Gladstone, R. M., Gudmundsson, G. H., et al.
16 (2016). Experimental design for three interrelated marine ice sheet and ocean model intercomparison projects:
17 MISMIP v. 3 (MISMIP +), ISOMIP v. 2 (ISOMIP +) and MISOMIP v. 1 (MISOMIP1). *Geosci. Model Dev.* 9,
18 2471–2497. doi:10.5194/gmd-9-2471-2016.
- 19 Asay-Davis, X. S., Jourdain, N. C., and Nakayama, Y. (2017). Developments in Simulating and Parameterizing
20 Interactions Between the Southern Ocean and the Antarctic Ice Sheet. *Curr. Clim. Chang. Reports* 3, 316–329.
21 doi:10.1007/s40641-017-0071-0.
- 22 Aschwanden, A., Fahnestock, M. A., and Truffer, M. (2016). Complex Greenland outlet glacier flow captured. *Nat.*
23 *Commun.* 7, 10524. doi:10.1038/ncomms10524.
- 24 Aschwanden, A., Fahnestock, M. A., Truffer, M., Brinkerhoff, D. J., Hock, R., Khroulev, C., et al. (2019). Contribution
25 of the Greenland Ice Sheet to sea level over the next millennium. *Sci. Adv.* 5. doi:10.1126/sciadv.aav9396.
- 26 Asher, W. E., Jessup, A. T., Branch, R., and Clark, D. (2014). Observations of rain-induced near-surface salinity
27 anomalies. *J. Geophys. Res. Ocean.* 119, 5483–5500. doi:10.1002/2014JC009954.
- 28 Axford, Y., Lasher, G. E., Kelly, M. A., Osterberg, E. C., Landis, J., Schellinger, G. C., et al. (2019). Holocene
29 temperature history of northwest Greenland – With new ice cap constraints and chironomid assemblages from
30 Deltasø. *Quat. Sci. Rev.* 215, 160–172. doi:https://doi.org/10.1016/j.quascirev.2019.05.011.
- 31 Azaneu, M., Kerr, R., Mata, M. M., and Garcia, C. A. E. (2013). Trends in the deep Southern Ocean (1958–2010):
32 Implications for Antarctic Bottom Water properties and volume export. *J. Geophys. Res. Ocean.* 118, 4213–4227.
33 doi:10.1002/jgre.20303.
- 34 Ba, J., Keenlyside, N. S., Latif, M., Park, W., Ding, H., Lohmann, K., et al. (2014). A multi-model comparison of
35 Atlantic multidecadal variability. *Clim. Dyn.* 43, 2333–2348. doi:10.1007/s00382-014-2056-1.
- 36 Baggenstos, D., Schmitt, J., Shackleton, S. A., Birner, B., Severinghaus, J. P., Kellerhals, T., et al. (2019). Earth's
37 radiative imbalance from the Last Glacial Maximum to the present. 116, 14881–14886.
38 doi:10.1073/pnas.1905447116.
- 39 Bahr, D. B., and Radić, V. (2012). Significant contribution to total mass from very small glaciers. *Cryosph.* 6, 763–770.
40 doi:10.5194/tc-6-763-2012.
- 41 Bakker, A. M. R., Wong, T. E., Ruckert, K. L., and Keller, K. (2017). Sea-level projections representing the deeply
42 uncertain contribution of the West Antarctic ice sheet. *Sci. Rep.* 7, 3880. doi:10.1038/s41598-017-04134-5.
- 43 Bakker, P., Schmittner, A., Lenaerts, J. T. M., Abe-Ouchi, A., Bi, D., van den Broeke, M. R., et al. (2016). Fate of the
44 Atlantic Meridional Overturning Circulation: Strong decline under continued warming and Greenland melting.
45 *Geophys. Res. Lett.* 43, 12,252–12,260. doi:10.1002/2016GL070457.
- 46 Bakun, A. (1990). Global climate change and intensification of coastal ocean upwelling. *Science* 247, 198–201.
- 47 Bamber, J. L., and Aspinall, W. P. (2013). An expert judgement assessment of future sea level rise from the ice sheets.
48 *Nat. Clim. Chang.* 3, 424–427. doi:10.1038/nclimate1778.
- 49 Bamber, J. L., Oppenheimer, M., Kopp, R. E., Aspinall, W. P., and Cooke, R. M. (2019). Ice sheet contributions to
50 future sea-level rise from structured expert judgment. *Proc. Natl. Acad. Sci.* 116, 11195–11200.
51 doi:10.1073/pnas.1817205116.
- 52 Bamber, J. L., Riva, R. E. M., Vermeersen, B. L. A., and Le Brocq, A. M. (2009). Reassessment of the Potential Sea-
53 Level Rise from a Collapse of the West Antarctic Ice Sheet. *Science* 324, 901–903.
- 54 Bamber, J. L., Westaway, R. M., Marzeion, B., and Wouters, B. (2018a). A new synthesis of annual land ice mass
55 trends 1992 to 2016. doi:10.1594/PANGAEA.890030.
- 56 Bamber, J. L., Westaway, R. M., Marzeion, B., and Wouters, B. (2018b). The land ice contribution to sea level during
57 the satellite era. *Environ. Res. Lett.* 13, 063008. doi:10.1088/1748-9326/aac2f0.
- 58 Banas, N. S. S., MacCready, P., and Hickey, B. M. M. (2009). The Columbia River plume as cross-shelf exporter and
59 along-coast barrier. *Cont. Shelf Res.* 29, 292–301. doi:10.1016/J.CSR.2008.03.011.
- 60 Banks, H. T., and Gregory, J. M. (2006). Mechanisms of ocean heat uptake in a coupled climate model and the
61 implications for tracer based predictions of ocean heat uptake. *Geophys. Res. Lett.* 33.

- doi:10.1029/2005GL025352.
- Barcaza, G., Nussbaumer, S. U., Tapia, G., Valdés, J., García, J.-L., Videla, Y., et al. (2017). Glacier inventory and recent glacier variations in the Andes of Chile, South America. *Ann. Glaciol.* 58, 166–180. doi:10.1017/aog.2017.28.
- Barletta, V. R., Bevis, M., Smith, B. E., Wilson, T., Brown, A., Bordoni, A., et al. (2018). Observed rapid bedrock uplift in Amundsen Sea Embayment promotes ice-sheet stability. *Science* 360, 1335–1339.
- Barreiro, M., and Philander, S. G. (2008). Response of the tropical Pacific to changes in extratropical clouds. *Clim. Dyn.* doi:10.1007/s00382-007-0363-5.
- Barthel, A., Agosta, C., Little, C. M., Hatterman, T., Jourdain, N. C., Goelzer, H., et al. (2019). CMIP5 model selection for ISMIP6 ice sheet model forcing: Greenland and Antarctica. *Cryosph. Discuss.* 2019, 1–34. doi:10.5194/tc-2019-191.
- Batbaatar, J., Gillespie, A., Fink, D., Matmon, A., and Fujioka, T. (2018). *Asynchronous glaciations in arid continental climate*. doi:10.1016/j.quascirev.2017.12.001.
- Bates, S. C., Fox-Kemper, B., Jayne, S. R., Large, W. G., Stevenson, S., and Yeager, S. G. (2012). Mean biases, variability, and trends in air-sea fluxes and sea surface temperature in the CCSM4. *J. Clim.* 25. doi:10.1175/JCLI-D-11-00442.1.
- Bathiany, S., Notz, D., Mauritsen, T., Raedel, G., and Brovkin, V. (2016). On the potential for abrupt Arctic winter sea-ice loss. *J. Clim.* doi:10.1175/JCLI-D-15-0466.1.
- Baum, S. K., and Crowley, T. J. (2003). The snow/ice instability as a mechanism for rapid climate change: A Neoproterozoic Snowball Earth model example. *Geophys. Res. Lett.* 30, 2030. doi:10.1029/2003GL017333.
- Beadling, R. L., Russell, J. L., Stouffer, R. J., Goodman, P. J., and Mazloff, M. (2019). Assessing the quality of Southern Ocean circulation in CMIP5 AOGCM and Earth System Model simulations. *J. Clim.* 32, 5915–5940. doi:10.1175/JCLI-D-19-0263.1.
- Beadling, R. L., Russell, J. L., Stouffer, R. J., Mazloff, M., Talley, L. D., Goodman, P. J., et al. (submitted). Representation of Southern Ocean properties across Coupled Model Intercomparison Project generations: CMIP3 to CMIP6. *J. Clim.* (submitted).
- Beal, L. M., De Ruijter, W. P. M., Biastoch, A., Zahn, R., 136, S. W. G., Cronin, M., et al. (2011). On the role of the Agulhas system in ocean circulation and climate. *Nature* 472, 429–436. doi:10.1038/nature09983.
- Beal, L. M., and Elipot, S. (2016). Broadening not strengthening of the Agulhas Current since the early 1990s. *Nature* 540, 570–573. doi:10.1038/nature19853.
- Becker, M., Karpytchev, M., and Lennartz-Sassinek, S. (2014). Long-term sea level trends: Natural or anthropogenic? *Geophys. Res. Lett.* 41, 5571–5580. doi:10.1002/2014GL061027.
- Beckmann, J., Perrette, M., Beyer, S., Calov, R., Willeit, M., and Ganopolski, A. (2018). Modeling the response of Greenland outlet glaciers to global warming using a coupled flowline-plume model. *Cryosph. Discuss.* 2018, 1–32. doi:10.5194/tc-2018-89.
- Beedle, M. J., Menounos, B., Luckman, B. H., and Wheate, R. (2009). Annual push moraines as climate proxy. *Geophys. Res. Lett.* 36, L20501. doi:10.1029/2009GL039533.
- Belcher, S. E., Grant, A. L. M., Hanley, K. E., Fox-Kemper, B., Van Roekel, L., Sullivan, P. P., et al. (2012). A global perspective on Langmuir turbulence in the ocean surface boundary layer. *Geophys. Res. Lett.* 39. doi:10.1029/2012GL052932.
- Bell, R. E., Banwell, A. F., Trusel, L. D., and Kingslake, J. (2018). Antarctic surface hydrology and impacts on ice-sheet mass balance. *Nat. Clim. Chang.* 8, 1044–1052. doi:10.1038/s41558-018-0326-3.
- Belmadani, A., Echevin, V., Codron, F., Takahashi, K., and Junquas, C. (2014). What dynamics drive future wind scenarios for coastal upwelling off Peru and Chile? *Clim. Dyn.* 43, 1893–1914.
- Belmecheri, S., Babst, F., Wahl, E. R., Stahle, D. W., and Trouet, V. (2016). Multi-century evaluation of Sierra Nevada snowpack. *Nat. Clim. Chang.* 6, 2–3. doi:10.1038/nclimate2809.
- Belt, S. T. (2018). Source-specific biomarkers as proxies for Arctic and Antarctic sea ice. *Org. Geochem.* 125, 277–298. doi:10.1016/j.orggeochem.2018.10.002.
- Belt, S. T. (2019). What do IP25 and related biomarkers really reveal about sea ice change? *Quat. Sci. Rev.* 204, 216–219. doi:10.1016/j.quascirev.2018.11.025.
- Belt, S. T., Cabedo-Sanz, P., Smik, L., Navarro-Rodriguez, A., Berben, S. M. P., Knies, J., et al. (2015). Identification of paleo Arctic winter sea ice limits and the marginal ice zone: Optimised biomarker-based reconstructions of late Quaternary Arctic sea ice. *Earth Planet. Sci. Lett.* 431, 127–139. doi:10.1016/j.epsl.2015.09.020.
- Belt, S. T., Massé, G., Rowland, S. J., Poulin, M., Michel, C., and LeBlanc, B. (2007). A novel chemical fossil of palaeo sea ice: IP25. *Org. Geochem.* 38, 16–27. doi:10.1016/j.orggeochem.2006.09.013.
- Beltran, C., Golledge, N. R., Christian, O., Kowalewski, D. E., Marie-Alexandrine, S., Hageman, K. J., et al. (2020). Southern Ocean temperature records and ice-sheet models demonstrate rapid Antarctic ice sheet retreat under low atmospheric CO₂ during Marine Isotope Stage 31. *Quat. Sci. Rev.* 228. doi:10.1016/j.quascirev.2019.106069.
- Bendtsen, J., Mortensen, J., Lennert, K., and Rysgaard, S. (2015). Heat sources for glacial ice melt in a west Greenland tidewater outlet glacier fjord: The role of subglacial freshwater discharge. *Geophys. Res. Lett.* doi:10.1002/2015GL063846.

- Beniston, M., Farinotti, D., Stoffel, M., Andreassen, L. M., Coppola, E., Eckert, N., et al. (2018). The European mountain cryosphere: a review of its current state, trends, and future challenges. *Cryosph.* 12, 759–794. doi:10.5194/tc-12-759-2018.
- Benn, D. I., Cowton, T., Todd, J., and Luckman, A. (2017). Glacier Calving in Greenland. *Curr. Clim. Chang. Reports* 3, 282–290. doi:10.1007/s40641-017-0070-1.
- Bennartz, R., Shupe, M. D., Turner, D. D., Walden, V. P., Steffen, K., Cox, C. J., et al. (2013). July 2012 Greenland melt extent enhanced by low-level liquid clouds. *Nature* 496, 83. Available at: <https://doi.org/10.1038/nature12002>.
- Bennetts, L. G., O’Farrell, S., and Uotila, P. (2017). Brief communication: Impacts of ocean-wave-induced breakup of Antarctic sea ice via thermodynamics in a stand-alone version of the CICE sea-ice model. *Cryosph.* 11, 1035–1040. doi:10.5194/tc-11-1035-2017.
- Bentamy, A., Piollé, J. F., Grouazel, A., Danielson, R., Gulev, S., Paul, F., et al. (2017). Review and assessment of latent and sensible heat flux accuracy over the global oceans. *Remote Sens. Environ.* 201, 196–218. doi:10.1016/J.RSE.2017.08.016.
- Bentley, M. J., Ocofaigh, C., Anderson, J. B., Conway, H., Davies, B., Graham, A. G. C., et al. (2014). A community-based geological reconstruction of Antarctic Ice Sheet deglaciation since the Last Glacial Maximum. *Quat. Sci. Rev.* 100, 1–9. doi:10.1016/j.quascirev.2014.06.025.
- Benz, V., Esper, O., Gersonde, R., Lamy, F., and Tiedemann, R. (2016). Last Glacial Maximum sea surface temperature and sea-ice extent in the Pacific sector of the Southern Ocean. *Quat. Sci. Rev.* 146, 216–237. doi:10.1016/j.quascirev.2016.06.006.
- Berben, S. M. P., Husum, K., Navarro-Rodriguez, A., Belt, S. T., and Aagaard-Sørensen, S. (2017). Semi-quantitative reconstruction of early to late Holocene spring and summer sea ice conditions in the northern Barents Sea. *J. Quat. Sci.* 32, 587–603. doi:10.1002/jqs.2953.
- Bereiter, B., Shackleton, S., Baggenstos, D., Kawamura, K., and Severinghaus, J. (2018). Mean global ocean temperatures during the last glacial transition. *Nature* 553, 39–44. doi:10.1038/nature25152.
- Berloff, P., Hogg, A. M. C., Dewar, W., Berloff, P., Hogg, A. M. C., and Dewar, W. (2007). The Turbulent Oscillator: A Mechanism of Low-Frequency Variability of the Wind-Driven Ocean Gyres. *J. Phys. Oceanogr.* 37, 2363–2386. doi:10.1175/JPO3118.1.
- Berthier, E., and Brun, F. (2019). Karakoram geodetic glacier mass balances between 2008 and 2016: persistence of the anomaly and influence of a large rock avalanche on Siachen Glacier. *J. Glaciol.* 65, 494–507. doi:10.1017/jog.2019.32.
- Bettencourt, J. H., López, C., Hernández-García, E., Montes, I., Sudre, J., Dewitte, B., et al. (2015). Boundaries of the Peruvian oxygen minimum zone shaped by coherent mesoscale dynamics. *Nat. Geosci.* 8, 937.
- Bevis, M., Harig, C., Khan, S. A., Brown, A., Simons, F. J., Willis, M., et al. (2019). Accelerating changes in ice mass within Greenland, and the ice sheet’s sensitivity to atmospheric forcing. *Proc. Natl. Acad. Sci. U. S. A.* 116, 1934–1939. doi:10.1073/pnas.1806562116.
- Bjastoch, A., Böning, C. W., Getzlaff, J., Molines, J.-M., and Madec, G. (2008). Causes of Interannual–Decadal Variability in the Meridional Overturning Circulation of the Midlatitude North Atlantic Ocean. *J. Clim.* 21, 6599–6615. doi:10.1175/2008JCLI2404.1.
- Bjastoch, A., Durgadoo, J. V., Morrison, A. K., van Seville, E., Weijer, W., and Griffies, S. M. (2015). Atlantic multi-decadal oscillation covaries with Agulhas leakage. *Nat. Commun.* 6, 10082. doi:10.1038/ncomms10082.
- Bierman, P. R., Shakun, J. D., Corbett, L. B., Zimmerman, S. R., and Rood, D. H. (2016). A persistent and dynamic East Greenland Ice Sheet over the past 7.5 million years. *Nature* 540, 256–260. doi:10.1038/nature20147.
- Bilbao, R. A. F., Gregory, J. M., and Bouttes, N. (2015). Analysis of the regional pattern of sea level change due to ocean dynamics and density change for 1993–2009 in observations and CMIP5 AOGCM’s. *Clim. Dyn.* doi:10.1007/s00382-015-2499-z.
- Bindoff, N. L., Cheung, W. W. L., Kairo, G. J., Aristegui, J., Guinder, V. A., Hallberg, R., et al. (2019). “Changing Ocean, Marine Ecosystems, and Dependent Communities,” in *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*, eds. H.-O. Pörtner, D. C. Roberts, V. Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, et al.
- Bindoff, N. L., and McDougall, T. J. (2000). Decadal Changes along an Indian Ocean Section at 32°S and Their Interpretation. *J. Phys. Oceanogr.* 30, 1207–1222. doi:10.1175/1520-0485(2000)030<1207:DCAAIO>2.0.CO;2.
- Bindoff, N. L., Stott, P. A., AchutaRao, K. M., Allen, M. R., Gillett, N., Gutzler, D., et al. (2013). Detection and Attribution of Climate Change: from Global to Regional. in *{Climate Change 2013}: The Physical Science Basis. {Contribution of Working Group I} to the {Fifth Assessment Report of the IPCC}*, eds. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, et al. (Cambridge University Press), 867–952. Available at: <http://bit.ly/2wNISEX>.
- Bintanja, R., Van Oldenborgh, G. J., Drijfhout, S. S., Wouters, B., and Katsman, C. A. (2013). Important role for ocean warming and increased ice-shelf melt in Antarctic sea-ice expansion. *Nat. Geosci.* 6, 376–379. doi:10.1038/ngeo1767.
- Bishop, S. P., Gent, P. R., Bryan, F. O., Thompson, A. F., Long, M. C., and Abernathey, R. (2016). Southern Ocean

- overturning compensation in an eddy-resolving climate simulation. *J. Phys. Oceanogr.* 46, 1575–1592.
- Bishop, S. P., Small, R. J., Bryan, F. O., Tomas, R. A., Bishop, S. P., Small, R. J., et al. (2017). Scale Dependence of Midlatitude Air–Sea Interaction. *J. Clim.* 30, 8207–8221. doi:10.1175/JCLI-D-17-0159.1.
- Biskaborn, B. K., Smith, S. L., Noetzi, J., Matthes, H., Vieira, G., Streletskiy, D. A., et al. (2019). Permafrost is warming at a global scale. *Nat. Commun.* 10, 264. doi:10.1038/s41467-018-08240-4.
- Bittermann, K., Rahmstorf, S., Kopp, R. E., and Kemp, A. C. (2017). Global mean sea-level rise in a world agreed upon in Paris. *Environ. Res. Lett.* 12. doi:10.1088/1748-9326/aa9def.
- Bitz, C. M., and Polvani, L. M. (2012). Antarctic climate response to stratospheric ozone depletion in a fine resolution ocean climate model. *Geophys. Res. Lett.* 39. doi:10.1029/2012GL053393.
- Björk, A. A., Kjær, K. H., Korsgaard, N. J., Khan, S. A., Kjeldsen, K. K., Andresen, C. S., et al. (2012). An aerial view of 80 years of climate-related glacier fluctuations in southeast Greenland. *Nat. Geosci.* 5, 427–432. doi:10.1038/ngeo1481.
- Björnsson, H., Pálsson, F., Gudmundsson, S., Magnússon, E., Adalgeirsdóttir, G., Jóhannesson, T., et al. (2013). Contribution of Icelandic ice caps to sea level rise: Trends and variability since the Little Ice Age. *Geophys. Res. Lett.* 40, 1546–1550. doi:10.1002/grl.50278.
- Blake-Mizen, K., Hatfield, R. G., Stoner, J. S., Carlson, A. E., Xuan, C., Walczak, M., et al. (2019). Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene intensification of Northern Hemisphere glaciation. *Quat. Sci. Rev.* 209, 40–51. doi:10.1016/j.quascirev.2019.01.015.
- Blaschek, M., and Renssen, H. (2013). The Holocene thermal maximum in the Nordic Seas: the impact of Greenland Ice Sheet melt and other forcings in a coupled atmosphere-sea-ice-ocean model. *Clim. Past* 9, 1629–1643.
- Bliss, A. C., Miller, J. A., and Meier, W. N. (2017). Comparison of passive microwave-derived early melt onset records on Arctic sea ice. *Remote Sens.* doi:10.3390/rs9030199.
- Bliss, A. C., Steele, M., Peng, G., Meier, W. N., and Dickinson, S. (2019). Regional variability of Arctic sea ice seasonal change climate indicators from a passive microwave climate data record. *Environ. Res. Lett.* 14. doi:10.1088/1748-9326/aafb84.
- Bliss, A., Hock, R., and Radić, V. (2014). Global response of glacier runoff to twenty-first century climate change. *J. Geophys. Res. Earth Surf.* 119, 717–730. doi:10.1002/2013JF002931.
- Boé, J., Hall, A., Colas, F., McWilliams, J. C., Qu, X., Kurian, J., et al. (2011). What shapes mesoscale wind anomalies in coastal upwelling zones? *Clim. Dyn.* 36, 2037–2049.
- Bolch, T., Pieczonka, T., Mukherjee, K., and Shea, J. (2017). Brief communication: Glaciers in the Hunza catchment (Karakoram) have been nearly in balance since the 1970s. *Cryosph.* 11, 531–539. doi:10.5194/tc-11-531-2017.
- Bond, N. A., Cronin, M. F., Freeland, H., and Mantua, N. (2015). Causes and impacts of the 2014 warm anomaly in the NE Pacific. *Geophys. Res. Lett.* 42, 3414–3420. doi:10.1002/2015GL063306. Received.
- Bondzio, J. H., Morlighem, M., Seroussi, H., Kleiner, T., Rückamp, M., Mouginot, J., et al. (2017). The mechanisms behind Jakobshavn Isbræ’s acceleration and mass loss: A 3-D thermomechanical model study. *Geophys. Res. Lett.* 44, 6252–6260. doi:10.1002/2017GL073309.
- Bondzio, J. H., Seroussi, H., Morlighem, M., Kleiner, T., Rückamp, M., Humbert, A., et al. (2016). Modelling calving front dynamics using a level-set method: application to Jakobshavn Isbræ, West Greenland. *Cryosph.* 10, 497–510. doi:10.5194/tc-10-497-2016.
- Bonekamp, P. N. J., de Kok, R. J., Collier, E., and Immerzeel, W. W. (2019). Contrasting Meteorological Drivers of the Glacier Mass Balance Between the Karakoram and Central Himalaya. *Front. Earth Sci.* 7, 107. doi:10.3389/feart.2019.00107.
- Böning, C. W., Behrens, E., Biastoch, A., Getzlaff, K., and Bamber, J. L. (2016). Emerging impact of Greenland meltwater on deepwater formation in the North Atlantic Ocean. *Nat. Geosci.* 9, 523–527. doi:10.1038/ngeo2740.
- Boone, A., Samuelsson, P., Gollvik, S., Napoly, A., Jarlan, L., Brun, E., et al. (2017). The interactions between soil-biosphere-atmosphere land surface model with a multi-energy balance (ISBA-MEB) option in SURFEXv8-Part 1: Model description. *Geosci. Model Dev.* doi:10.5194/gmd-10-843-2017.
- Borge, A. F., Westermann, S., Solheim, I., and Etzelmüller, B. (2017). Strong degradation of palsas and peat plateaus in northern Norway during the last 60 years. *Cryosphere* 11, 1–16. doi:10.5194/tc-11-1-2017.
- Bormann, K., Brown, R., Derksen, C., and Painter, T. (2018). Estimating snow cover trends from space. *Nat. Clim. Chang.* 8, 924–928.
- Born, A., and Nisancioglu, K. H. (2012). Melting of Northern Greenland during the last interglaciation. *Cryosph.* 6, 1239–1250.
- Born, A., Stocker, T. F., Raible, C. C., and Levermann, A. (2013). Is the Atlantic subpolar gyre bistable in comprehensive coupled climate models? *Clim. Dyn.* 40, 2993–3007. doi:10.1007/s00382-012-1525-7.
- Born, A., Stocker, T. F., and Sandø, A. B. (2016). Transport of salt and freshwater in the Atlantic Subpolar Gyre. *Ocean Dyn.* 66, 1051–1064. doi:10.1007/s10236-016-0970-y.
- Boutin, J., Martin, N., Reverdin, G., Morisset, S., Yin, X., Centurioni, L., et al. (2014). Sea surface salinity under rain cells: SMOS satellite and in situ drifters observations. *J. Geophys. Res. Ocean.* 119, 5533–5545. doi:10.1002/2014JC010070.

- Boutin, J., Martin, N., Reverdin, G., Yin, X., and Gaillard, F. (2013). Sea surface freshening inferred from SMOS and ARGO salinity: impact of rain. *Ocean Sci.* 9, 183–192. doi:10.5194/os-9-183-2013.
- Box, J. E., and Colgan, W. (2013). Greenland Ice Sheet Mass Balance Reconstruction. Part III: Marine Ice Loss and Total Mass Balance (1840–2010). *J. Clim.* 26, 6990–7002. doi:10.1175/JCLI-D-12-00546.1.
- Box, J. E., Fettweis, X., Stroeve, J. C., Tedesco, M., Hall, D. K., and Steffen, K. (2012). Greenland ice sheet albedo feedback: thermodynamics and atmospheric drivers. *Cryosph.* 6, 821–839. doi:10.5194/tc-6-821-2012.
- Bracegirdle, T. J., Stephenson, D. B., Turner, J., and Phillips, T. (2015). The importance of sea-ice area biases in 21 st century multi-model projections of Antarctic temperature and precipitation. *Geophys. Res. Lett.* 42, 10832–10839. doi:10.1002/2015GL067055.
- Bracegirdle, T., Krinner, G., Tonelli, M., Haumann, A., Naughten, K. A., Rackow, T., et al. (submitted). Twenty-first century changes in Antarctic and Southern Ocean surface climate in CMIP6. *Atmos. Sci. Lett.* (submitted).
- Bradwell, T., Small, D., Fabel, D., Smedley, R. K., Clark, C. D., Saher, M. H., et al. (2019). Ice-stream demise dynamically conditioned by trough shape and bed strength. *Sci. Adv.* 5. doi:10.1126/sciadv.aau1380.
- Brady, R. X., Alexander, M. A., Lovenduski, N. S., and Rykaczewski, R. R. (2017). Emergent anthropogenic trends in California Current upwelling. *Geophys. Res. Lett.* 44, 5044–5052.
- Brandon, C. M., Woodruff, J. D., Donnelly, J. P., and Sullivan, R. M. (2015). How Unique was Hurricane Sandy? Sedimentary Reconstructions of Extreme Flooding from New York Harbor. *Sci. Rep.* 4, 7366. doi:10.1038/srep07366.
- Brandon, C. M., Woodruff, J. D., Lane, D., and Donnelly, J. P. (2013). Tropical cyclone wind speed constraints from resultant storm surge deposition: A 2500 year reconstruction of hurricane activity from St. Marks, FL. *Geochemistry, Geophys. Geosystems* 14, 2993–3008.
- Braun, M. H., Malz, P., Sommer, C., Farias-Barahona, D., Sauter, T., Casassa, G., et al. (2019). Constraining glacier elevation and mass changes in South America. *Nat. Clim. Chang.* 9, 130–136. doi:10.1038/s41558-018-0375-7.
- Brendryen, J., Hafliðason, H., Yokoyama, Y., Haaga, K. A., Hannisdal, B., and Jo Brendryen, Hafliði Hafliðason, Yusuke Yokoyama, K. A. H. and B. H. (submitted). Collapse of Eurasian ice sheets 14,600 years ago was a major source of global Meltwater Pulse 1a. *Nat. Geosci.* (submitted).
- Brierley, C., Burls, N., Ravelo, C., and Fedorov, A. (2015). Pliocene warmth and gradients. *Nat. Geosci.* 8, 419–420. doi:10.1038/ngeo2444.
- Briggs, M. A., Walvoord, M. A., McKenzie, J. M., Voss, C. I., Day-Lewis, F. D., and Lane, J. W. (2014a). New permafrost is forming around shrinking Arctic lakes, but will it last? *Geophys. Res. Lett.* doi:10.1002/2014GL059251.
- Briggs, R. D., Pollard, D., and Tarasov, L. (2014b). A data-constrained large ensemble analysis of Antarctic evolution since the Eemian. *Quat. Sci. Rev.* 103, 91–115.
- Briner, J. P., McKay, N. P., Axford, Y., Bennike, O., Bradley, R. S., de Vernal, A., et al. (2016). Holocene climate change in Arctic Canada and Greenland. *Quat. Sci. Rev.* 147, 340–364.
- Brink, K. H. (2016). Cross-shelf exchange. *Ann. Rev. Mar. Sci.* 8, 59–78.
- Brodeau, L., Barnier, B., Gulev, S. K., and Woods, C. (2017). Climatologically Significant Effects of Some Approximations in the Bulk Parameterizations of Turbulent Air–Sea Fluxes. *J. Phys. Oceanogr.* 47, 5–28. doi:10.1175/JPO-D-16-0169.1.
- Brodeau, L., and Koenigk, T. (2016). Extinction of the northern oceanic deep convection in an ensemble of climate model simulations of the 20th and 21st centuries. *Clim. Dyn.* 46, 2863–2882. doi:10.1007/s00382-015-2736-5.
- Bromwich, D. H., Nicolas, J. P., Monaghan, A. J., Lazzara, M. A., Keller, L. M., Weidner, G. A., et al. (2013). Central West Antarctica among the most rapidly warming regions on Earth. *Nat. Geosci.* 6, 139–145. doi:10.1038/ngeo1671.
- Brondex, J., Gagliardini, O., Gillet-Chaulet, F., and Durand, G. (2017). Sensitivity of grounding line dynamics to the choice of the friction law. *J. Glaciol.* 63, 854–866. doi:10.1017/jog.2017.51.
- Brönnimann, S., Franke, J., Nussbaumer, S. U., Zumbühl, H. J., Steiner, D., Trachsel, M., et al. (2019). Last phase of the Little Ice Age forced by volcanic eruptions. *Nat. Geosci.* 12, 650–656. doi:10.1038/s41561-019-0402-y.
- Bronselaer, B., Russell, J. L., Winton, M., Williams, N. L., Key, R. M., Dunne, J. P., et al. (submitted). Impact of wind and meltwater on recent observed Southern Ocean physical and chemical patterns of change. (submitted).
- Bronselaer, B., Winton, M., Griffies, S. M., Hurlin, W. J., Rodgers, K. B., Sergienko, O. V., et al. (2018). Change in future climate due to Antarctic meltwater. *Nature* 564, 53–58. doi:10.1038/s41586-018-0712-z.
- Brown, R. D., and Derksen, C. (2013). Is Eurasian October snow cover extent increasing? *Environ. Res. Lett.* 8, 024006. doi:10.1088/1748-9326/8/2/024006.
- Brown, R. D., and Robinson, D. A. (2011). Northern Hemisphere spring snow cover variability and change over 1922–2010 including an assessment of uncertainty. *Cryosph.* 5, 219–229. doi:10.5194/tc-5-219-2011.
- Brown, R., Vikhamar-Schuler, D., Bulygina, O., Derksen, C., Loujus, K., Mudryk, L., et al. (2017). “Arctic terrestrial snow cover,” in *Snow, Water, Ice and Permafrost in the Arctic (SWIPA) 2017* (Oslo, Norway: Arctic Monitoring and Assessment Programme (AMAP)), 25–64.
- Brüggemann, N., and Katsman, C. A. (2019). Dynamics of Downwelling in an Eddying Marginal Sea: Contrasting the Eulerian and the Isopycnal Perspective. *J. Phys. Oceanogr.* 49, 3017–3035. doi:10.1175/JPO-D-19-0090.1.

- 1 Brun, F., Berthier, E., Wagnon, P., Kääb, A., and Treichler, D. (2017). A spatially resolved estimate of High Mountain
2 Asia glacier mass balances from 2000 to 2016. *Nat. Geosci.* 10, 668–673. doi:10.1038/ngeo2999.
- 3 Brun, F., Buri, P., Miles, E. S., Wagnon, P., Steiner, J., Berthier, E., et al. (2016). Quantifying volume loss from ice
4 cliffs on debris-covered glaciers using high-resolution terrestrial and aerial photogrammetry. *J. Glaciol.* 62, 684–
5 695. doi:10.1017/jog.2016.54.
- 6 Brutel-Vuilmet, C., Ménégoz, M., and Krinner, G. (2013). An analysis of present and future seasonal Northern
7 Hemisphere land snow cover simulated by CMIP5 coupled climate models. *Cryosph.* 7, 67–80. doi:10.5194/tc-7-
8 67-2013.
- 9 Bryden, H. L., Johns, W. E., King, B. A., McCarthy, G., McDonagh, E. L., Moat, B. I., et al. (2019). Reduction in
10 ocean heat transport at 26°N since 2008 cools the eastern subpolar gyre of the North Atlantic Ocean. *J. Clim.* 0,
11 null. doi:10.1175/JCLI-D-19-0323.1.
- 12 Buchanan, M. K., Kopp, R. E., Oppenheimer, M., and Tebaldi, C. (2016). Allowances for evolving coastal flood risk
13 under uncertain local sea-level rise. *Clim. Change* 137. doi:10.1007/s10584-016-1664-7.
- 14 Buchanan, M. K., Oppenheimer, M., and Kopp, R. E. (2017). Amplification of flood frequencies with local sea level
15 rise and emerging flood regimes. *Environ. Res. Lett.* 12. doi:10.1088/1748-9326/aa6cb3.
- 16 Buckingham, C. E., Lucas, N., Belcher, S., Rippeth, T., Grant, A., Le Sommer, J., et al. (2019). The contribution of
17 surface and submesoscale processes to turbulence in the open ocean surface boundary layer. *J. Adv. Model. Earth*
18 *Syst.* n/a. doi:10.1029/2019MS001801.
- 19 Buckley, M. W., and Marshall, J. (2016). Observations, inferences, and mechanisms of the Atlantic Meridional
20 Overturning Circulation: A review. *Rev. Geophys.* 54, 5–63. doi:10.1002/2015RG000493.
- 21 Bulthuis, K., Arnst, M., Sun, S., and Pattyn, F. (2019). Uncertainty quantification of the multi-centennial response of
22 the antarctic ice sheet to climate change. *Cryosphere* 13, 1349–1380. doi:10.5194/tc-13-1349-2019.
- 23 Bulygina, O. N., Groisman, P. Y., Razuvaev, V. N., and Korshunova, N. N. (2011). Changes in snow cover
24 characteristics over Northern Eurasia since 1966. *Environ. Res. Lett.* 6, 0452204. doi:10.1088/1748-
25 9326/6/4/045204.
- 26 Bunce, C., Carr, J. R., Nienow, P. W., Ross, N., and Killick, R. (2018). Ice front change of marine-terminating outlet
27 glaciers in northwest and southeast Greenland during the 21st century. *J. Glaciol.* 64, 523–535.
28 doi:10.1017/jog.2018.44.
- 29 Bunzel, F., Notz, D., and Pedersen, L. T. (2018). Retrievals of Arctic Sea-Ice Volume and Its Trend Significantly
30 Affected by Interannual Snow Variability. *Geophys. Res. Lett.* 45, 11,751–11,759. doi:10.1029/2018GL078867.
- 31 Burgard, C., and Notz, D. (2017). Drivers of Arctic Ocean warming in CMIP5 models. *Geophys. Res. Lett.* 44, 4263–
32 4271. doi:10.1002/2016GL072342.
- 33 Burke, E., Zhang, Y., and Krinner, G. (submitted). Evaluating permafrost physics in the CMIP6 models and their
34 sensitivity to climate change. *Cryosph. Discuss.* (submitted).
- 35 Burls, N. J., Fedorov, A. V., Sigman, D. M., Jaccard, S. L., Tiedemann, R., and Haug, G. H. (2017). Active Pacific
36 meridional overturning circulation (PMOC) during the warm Pliocene. *Sci. Adv.* 3, e1700156.
37 doi:10.1126/sciadv.1700156.
- 38 Cabedo-Sanz, P., and Belt, S. T. (2016). Seasonal sea ice variability in eastern Fram Strait over the last 2000 years.
39 *Arktos* 2, 22. doi:10.1007/s41063-016-0023-2.
- 40 Cabedo-Sanz, P., Belt, S. T., Jennings, A. E., Andrews, J. T., and Geirsdóttir, Á. (2016). Variability in drift ice export
41 from the Arctic Ocean to the North Icelandic Shelf over the last 8000 years: A multi-proxy evaluation. *Quat. Sci.*
42 *Rev.* 146, 99–115. doi:10.1016/j.quascirev.2016.06.012.
- 43 Caesar, L., Rahmstorf, S., Robinson, A., Feulner, G., and Saba, V. (2018). Observed fingerprint of a weakening
44 Atlantic Ocean overturning circulation. *Nature*. doi:10.1038/s41586-018-0006-5.
- 45 Cai, W. (2006). Antarctic ozone depletion causes an intensification of the Southern Ocean super-gyre circulation.
46 *Geophys. Res. Lett.* 33, L03712.
- 47 Cai, W., Borlace, S., Lengaigne, M., Van Rensch, P., Collins, M., Vecchi, G., et al. (2014). Increasing frequency of
48 extreme El Niño events due to greenhouse warming. *Nat. Clim. Chang.* 4, 111–116. doi:10.1038/nclimate2100.
- 49 Cai, W., and Cowan, T. (2007). Trends in Southern Hemisphere Circulation in IPCC AR4 Models over 1950–99:
50 Ozone Depletion versus Greenhouse Forcing. *J. Clim.* 20, 681–693. doi:10.1175/JCLI4028.1.
- 51 Cai, W., Shi, G., Cowan, T., Bi, D., and Ribbe, J. (2005). The response of the Southern Annular Mode, the East
52 Australian Current, and the southern mid-latitude ocean circulation to global warming. *Geophys. Res. Lett.* 32,
53 L23706. doi:10.1029/2005GL024701.
- 54 Cai, W., Wang, G., Santoso, A., McPhaden, M. J., Wu, L., Jin, F.-F., et al. (2015). Increased frequency of extreme La
55 Niña events under greenhouse warming. *Nat. Clim. Chang.* 5, 132–137. doi:10.1038/nclimate2492.
- 56 Cai, W., Wu, L., Lengaigne, M., Li, T., McGregor, S., Kug, J.-S., et al. (2019). Pantropical climate interactions. *Science*
57 363, eaav4236. doi:10.1126/SCIENCE.AAV4236.
- 58 Cai, W., Zheng, X.-T., Weller, E., Collins, M., Cowan, T., Lengaigne, M., et al. (2013). Projected response of the
59 Indian Ocean Dipole to greenhouse warming. *Nat. Geosci.* 6, 999–1007. doi:10.1038/ngeo2009.
- 60 Calonne, N., Flin, F., Geindreau, C., Lesaffre, B., and Rolland Du Roscoat, S. (2014). Study of a temperature gradient
61 metamorphism of snow from 3-D images: Time evolution of microstructures, physical properties and their

- associated anisotropy. *Cryosph.* 8, 2255–2274. doi:10.5194/tc-8-2255-2014.
- Calonne, N., Geindreau, C., and Flin, F. (2015). Macroscopic modeling of heat and water vapor transfer with phase change in dry snow based on an upscaling method: Influence of air convection. *J. Geophys. Res. F Earth Surf.* 120, 2476–2497. doi:10.1002/2015JF003605.
- Calonne, N., Montagnat, M., Matzl, M., and Schneebeli, M. (2017). The layered evolution of fabric and microstructure of snow at Point Barnola, Central East Antarctica. *Earth Planet. Sci. Lett.* 460, 293–301. doi:10.1016/j.epsl.2016.11.041.
- Calov R, Robinson A, Perrette M, G. A. (2015). Simulating the Greenland ice sheet under present-day and palaeo constraints including a new discharge parameterization. *Cryosph.* 9, 179–196.
- Calov, R., Beyer, S., Greve, R., Beckmann, J., Willeit, M., Kleiner, T., et al. (2018). Simulation of the future sea level contribution of Greenland with a new glacial system model. *Cryosph.* 12, 3097–3121. doi:10.5194/tc-12-3097-2018.
- Calov, R., Ganopolski, A., Claussen, M., Petoukhov, V., and Greve, R. (2005). Transient simulation of the last glacial inception. Part I: Glacial inception as a bifurcation in the climate system. *Clim. Dyn.* 24, 545–561. doi:10.1007/s00382-005-0007-6.
- Camill, P. (2005). Permafrost thaw accelerates in boreal peatlands during late-20th century climate warming. *Clim. Change* 68, 135–152. doi:10.1007/s10584-005-4785-y.
- Campbell, E. C., Wilson, E. A., Moore, G. W. K., Riser, S. C., Brayton, C. E., Mazloff, M. R., et al. (2019). Antarctic offshore polynyas linked to Southern Hemisphere climate anomalies. *Nature*. doi:10.1038/s41586-019-1294-0.
- Cane, M. A., Clement, A. C., Kaplan, A., Kushnir, Y., Pozdnyakov, D., Seager, R., et al. (1997). Twentieth-century sea surface temperature trends. *Science* doi:10.1126/science.275.5302.957.
- Capet, X., Campos, E. J., and Paiva, A. M. (2008). Submesoscale activity over the Argentinian shelf. *Geophys. Res. Lett.* 35, L15605. doi:10.1029/2008GL034736.
- Capron, E., Govin, A., Stone, E. J., Masson-Delmotte, V., Mulitza, S., Otto-Bliesner, B., et al. (2014). Temporal and spatial structure of multi-millennial temperature changes at high latitudes during the Last interglacial. *Quat. Sci. Rev.* 103, 116–133. doi:10.1016/j.quascirev.2014.08.018.
- Carlson, A. E., Winsor, K., Ullman, D. J., Brook, E. J., Rood, D. H., Axford, Y., et al. (2014). Earliest Holocene south Greenland ice sheet retreat within its late Holocene extent. *Geophys. Res. Lett.* 41, 5514–5521.
- Carmagnola, C. M., Morin, S., Lafaysse, M., Domine, F., Lesaffre, B., Lejeune, Y., et al. (2014). Implementation and evaluation of prognostic representations of the optical diameter of snow in the SURFEX/ISBA-Crocus detailed snowpack model. *Cryosph.* 8, 417–437. doi:10.5194/tc-8-417-2014.
- Cavalieri, D. J., Parkinson, C. L., Gloersen, P., and Zwally, H. J. (1996). Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave Data, Version 1. *Boulder, Color. USA. NASA Natl. Snow Ice Data Cent. Distrib. Act. Arch. Cent.* doi:10.5067/8GQ8LZQVL0VL.
- Cavole, L. M., Demko, A. M., Diner, R. E., Giddings, A., Koester, I., Pagniello, C. M. L. S., et al. (2016). Biological impacts of the 2013–2015 warm-water anomaly in the northeast Pacific: Winners, Losers, and the Future. *Oceanography* 29, 273–285. doi:10.5670/oceanog.2016.32.
- Cazenave, A., Meyssignac, B., Ablain, M., Balmaseda, M., Bamber, J., Barletta, V., et al. (2018). Global sea-level budget 1993-present. *Earth Syst. Sci. Data*. doi:10.5194/essd-10-1551-2018.
- Cerrone, D., Fusco, G., Simmonds, I., Aulicino, G., and Budillon, G. (2017). Dominant covarying climate signals in the southern ocean and antarctic sea ice influence during the last three decades. *J. Clim.* 30, 3055–3072. doi:10.1175/JCLI-D-16-0439.1.
- Cessi, P. (2019). The Global Overturning Circulation. *Ann. Rev. Mar. Sci.* 11, 249–270. doi:10.1146/annurev-marine-010318-095241.
- Cetina-Heredia, P., Roughan, M., van Sebille, E., and Coleman, M. A. (2014). Long-term trends in the East Australian Current separation latitude and eddy driven transport. *J. Geophys. Res. Ocean.* 119, 4351–4366. doi:10.1002/2014JC010071.
- Chadburn, S., Burke, E., Essery, R., Boike, J., Langer, M., Heikenfeld, M., et al. (2015). An improved representation of physical permafrost dynamics in the JULES land-surface model. *Geosci. Model Dev.* 8, 1493–1508. doi:10.5194/gmd-8-1493-2015.
- Chadburn, S. E., Burke, E. J., Cox, P. M., Friedlingstein, P., Hugelius, G., and Westermann, S. (2017). An observation-based constraint on permafrost loss as a function of global warming. *Nat. Clim. Chang.* 7, 340–344. doi:10.1038/nclimate3262.
- Chafik, L., Häkkinen, S., England, M. H., Carton, J. A., Nigam, S., Ruiz-Barradas, A., et al. (2016). Global linkages originating from decadal oceanic variability in the subpolar North Atlantic. *Geophys. Res. Lett.* doi:10.1002/2016GL071134.
- Chan, P., Halfar, J., Adey, W., Hetzinger, S., Zack, T., Moore, G. W. K., et al. (2017). Multicentennial record of Labrador Sea primary productivity and sea-ice variability archived in coralline algal barium. *Nat. Commun.* 8, 15543. doi:10.1038/ncomms15543.
- Chang, C. Y., Carton, J. A., Grodsky, S. A., and Nigam, S. (2007). Seasonal climate of the tropical Atlantic sector in the NCAR community climate system model 3: Error structure and probable causes of errors. *J. Clim.*

- doi:10.1175/JCLI4047.1.
- Chapman, C., Lea, M.-A., Meyer, A., Sallvé, J.-B., and Hindell, M. (submitted). Defining Southern Ocean fronts and their influence on biological and physical processes in a changing climate. *Nat. Clim. Chang.* (submitted).
- Charalampidis, C., van As, D., Box, J. E., van den Broeke, M. R., Colgan, W. T., Doyle, S. H., et al. (2015). Changing surface–atmosphere energy exchange and refreezing capacity of the lower accumulation area, West Greenland. *Cryosph.* 9, 2163–2181. doi:10.5194/tc-9-2163-2015.
- Charalampidis, C., Van As, D., Colgan, W. T., Fausto, R. S., Macferrin, M., and Machguth, H. (2016). Thermal tracing of retained meltwater in the lower accumulation area of the Southwestern Greenland ice sheet. *Ann. Glaciol.* 57, 1–10. doi:DOI: 10.1017/aog.2016.2.
- Chasmer, L., and Hopkinson, C. (2017). Threshold loss of discontinuous permafrost and landscape evolution. *Glob. Chang. Biol.* 23, 2672–2686. doi:10.1111/gcb.13537.
- Chassignet, E. P., and Marshall, D. P. (2008). “Gulf Stream separation in numerical ocean models,” in (American Geophysical Union (AGU)), 39–61. doi:10.1029/177GM05.
- Chassignet, E. P., Xu, X., Chassignet, E. P., and Xu, X. (2017). Impact of Horizontal Resolution (1/12° to 1/50°) on Gulf Stream Separation, Penetration, and Variability. *J. Phys. Oceanogr.* 47, 1999–2021. doi:10.1175/JPO-D-17-0031.1.
- Chassignet, E. P., Yeager, S. G., Fox-Kemper, B., Bozec, A., Castruccio, F. S., Danabasoglu, G., et al. (submitted). Impact of horizontal resolution on global ocean-sea-ice model simulations based on the experimental protocols of the Ocean Model Intercomparison Project phase 2 (OMIP-2). *Geosci. Model Dev. Discuss.* (submitted).
- Chelton, D. B., and Xie, S.-P. (2010). Coupled Ocean-Atmosphere Interaction at Oceanic Mesoscales. *Oceanography* 23, 52–69. doi:10.2307/24860862.
- Chen, C., Liu, W., and Wang, G. (2019). Understanding the Uncertainty in the 21st Century Dynamic Sea Level Projections: The Role of the AMOC. *Geophys. Res. Lett.* 46, 210–217. doi:10.1029/2018GL080676.
- Chen, H. W., Alley, R. B., and Zhang, F. (2016). Interannual Arctic sea ice variability and associated winter weather patterns: A regional perspective for 1979–2014. *J. Geophys. Res.* 121, 14433–14455. doi:10.1002/2016JD024769.
- Chen, W.-B., and Liu, W.-C. (2016). Assessment of storm surge inundation and potential hazard maps for the southern coast of Taiwan. *Nat. Hazards* 82, 591–616.
- Chen, X., Zhang, X., Church, J. A., Watson, C. S., King, M. A., Monselesan, D., et al. (2017). The increasing rate of global mean sea-level rise during 1993–2014. *Nat. Clim. Chang.* 7, 492–495. doi:10.1038/nclimate3325.
- Chen, Z., and Wu, L. (2012). Long-term change of the Pacific North Equatorial Current bifurcation in SODA. *J. Geophys. Res. Ocean.* 117, n/a–n/a. doi:10.1029/2011JC007814.
- Cheng, L., Abraham, J., Hausfather, Z., and Trenberth, K. E. (2019). How fast are the oceans warming? *Science* 363, 128–129. doi:10.1126/science.aav7619.
- Cheng, L., Trenberth, K. E., Fasullo, J., Boyer, T., Abraham, J., and Zhu, J. (2017). Improved estimates of ocean heat content from 1960 to 2015. *Sci. Adv.* 3, e1601545. doi:10.1126/sciadv.1601545.
- Cheng, L., Trenberth, K. E., Mann, M. E., Abraham, J. P., Gruber, N., Fasullo, J., et al. The emergence of anthropogenic climate change in ocean salinity. *Nature*.
- Cheng, W., Chiang, J. C. H., and Zhang, D. (2013). Atlantic Meridional Overturning Circulation (AMOC) in CMIP5 Models: RCP and Historical Simulations. *J. Clim.* 26, 7187–7197. doi:10.1175/JCLI-D-12-00496.1.
- Cherchi, A., Ambrizzi, T., Behera, S., Freitas, A. C. V., Morioka, Y., and Zhou, T. (2018). The response of subtropical highs to climate change. *Curr. Clim. Chang. Reports* 4, 371–382.
- Chevallier, M., Smith, G. C., Dupont, F., Lemieux, J.-F., Forget, G., Fujii, Y., et al. (2017). Intercomparison of the Arctic sea ice cover in global ocean–sea ice reanalyses from the ORA-IP project. *Clim. Dyn.* 49, 1107–1136. doi:10.1007/s00382-016-2985-y.
- Christian, J. E., Koutnik, M., and Roe, G. (2018). Committed retreat: controls on glacier disequilibrium in a warming climate. *J. Glaciol.* 64, 675–688. doi:DOI: 10.1017/jog.2018.57.
- Christiansen, H. H., Etzelmüller, B., Isaksen, K., Juliussen, H., Farbrot, H., Humlum, O., et al. (2010). The thermal state of permafrost in the nordic area during the international polar year 2007–2009. *Permafr. Periglac. Process.* 21, 156–181. doi:10.1002/ppp.687.
- Christianson, K., Bushuk, M., Dutrieux, P., Parizek, B. R., Joughin, I. R., Alley, R. B., et al. (2016). Sensitivity of Pine Island Glacier to observed ocean forcing. *Geophys. Res. Lett.* 43, 10,810–817,825. doi:10.1002/2016GL070500.
- Christie, F. D. W., Bingham, R. G., Gourmelen, N., Steig, E. J., Bisset, R. R., Pritchard, H. D., et al. (2018). Glacier change along West Antarctica’s Marie Byrd Land Sector and links to inter-decadal atmosphere-ocean variability. *Cryosph.* 12, 2461–2479. doi:10.5194/tc-12-2461-2018.
- Christie, F. D. W., Bingham, R. G., Gourmelen, N., Tett, S. F. B., and Muto, A. (2016). Four-decade record of pervasive grounding line retreat along the Bellingshausen margin of West Antarctica. *Geophys. Res. Lett.* 43, 5741–5749. doi:10.1002/2016GL068972.
- Christoffersen, P., Mugford, R. I., Heywood, K. J., Joughin, I., Dowdeswell, J. A., Syvitski, J. P. M., et al. (2011). Warming of waters in an East Greenland fjord prior to glacier retreat: Mechanisms and connection to large-scale atmospheric conditions. *Cryosphere*. doi:10.5194/tc-5-701-2011.
- Church, J. A., Clark, P. U., Cazenave, A., Gregory, J. M., Jevrejeva, S., Levermann, A., et al. (2013a). Sea Level

- Change. in *{Climate Change 2013}: The Physical Science Basis. {Contribution of Working Group I} to the {Fifth Assessment Report of the IPCC}*, eds. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, et al. (Cambridge University Press), 1137–1216. Available at: <http://bit.ly/2wNISEX>.
- Church, J. A., Monselesan, D., Gregory, J. M., and Marzeion, B. (2013b). Evaluating the ability of process based models to project sea-level change. *Environ. Res. Lett.* 8, 14051. doi:10.1088/1748-9326/8/1/014051.
- Church, J. A., White, N. J., and Arblaster, J. M. (2005). Significant decadal-scale impact of volcanic eruptions on sea level and ocean heat content. *Nature* 438, 74–77. doi:10.1038/nature04237.
- Cicoira, A., Beutel, J., Faillettaz, J., and Vieli, A. (2019). Water controls the seasonal rhythm of rock glacier flow. *Earth Planet. Sci. Lett.* 528. doi:10.1016/j.epsl.2019.115844.
- Clark, P. U., He, F., Golledge, N. R., Mitrovica, J. X., Dutton, A., Hoffman, J. S., et al. (in press). Oceanic forcing of penultimate deglacial and last interglacial sea-level rise. *Nature* (in press).
- Clark, P. U., Mix, A. C., Eby, M., Levermann, A., Rogelj, J., Nauels, A., et al. (2018). Sea-level commitment as a gauge for climate policy. *Nat. Clim. Chang.* doi:10.1038/s41558-018-0226-6.
- Clark, P. U., Shakun, J. D., Marcott, S. A., Mix, A. C., Eby, M., Kulp, S., et al. (2016). Consequences of twenty-first-century policy for multi-millennial climate and sea-level change. *Nat. Clim. Chang.* 6, 360–369. doi:10.1038/nclimate2923.
- Clarke, G. K. C., Jarosch, A. H., Anslow, F. S., Radić, V., and Menounos, B. (2015). Projected deglaciation of western Canada in the twenty-first century. *Nat. Geosci.* 8, 372. doi:10.1038/NGEO2407.
- Clement, A., Bellomo, K., Murphy, L. N., Cane, M. A., Mauritsen, T., Radel, G., et al. (2015). The Atlantic Multidecadal Oscillation without a role for ocean circulation. *Science* 350, 320–324. doi:10.1126/science.aab3980.
- Clement, A. C., Seager, R., Cane, M. A., and Zebiak, S. E. (1996). An Ocean Dynamical Thermostat. *J. Clim.* 9, 2190–2196.
- Clerc, F., Minchew, B. M., and Behn, M. D. (2019). Marine Ice Cliff Instability Mitigated by Slow Removal of Ice Shelves. *Geophys. Res. Lett.*, doi: 10.1029/2019GL084183.
- Coats, S., and Karnauskas, K. B. (2018). A role for the equatorial undercurrent in the ocean dynamical thermostat. *J. Clim.* doi:10.1175/JCLI-D-17-0513.1.
- Colas, F., McWilliams, J. C., Capet, X., and Kurian, J. (2012). Heat balance and eddies in the Peru-Chile current system. *Clim. Dyn.* 39, 509–529. doi:10.1007/s00382-011-1170-6.
- Colgan, W., Mankoff, K. D., Kjeldsen, K. K., Bjørk, A. A., Box, J. E., Simonsen, S. B., et al. (2019). Programme for monitoring of the Greenland ice sheet (PROMICE): Ice sheet mass balance (1995-2015). Available at: doi: https://doi.org/10.22008/promice/data/mass_balance/input_output/perimeter.
- Colleoni, F., Masina, S., Cherchi, A., Navarra, A., Ritz, C., Peyaud, V., et al. (2014). Modeling Northern Hemisphere ice-sheet distribution during MIS 5 and MIS 7 glacial inception. *Clim. Past* 10, 269–291. doi:10.5194/cp-10-269-2014.
- Collins, M., Knutti, R., Arblaster, J., Dufresne, J.-L., Fichet, T., Friedlingstein, P., et al. (2013). “Long-term Climate Change: Projections, Commitments and Irreversibility,” in *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, 1029–1136. doi:10.1017/CBO9781107415324.024.
- Collins, M., Sutherland, M., Bouwer, L., Cheong, S.-M., Frölicher, T., Combes, H. J. Des, et al. (2019). “Extremes, Abrupt Changes and Managing Risks,” in *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*, eds. H.-O. Pörtner, D. C. Roberts, V. Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, et al. (In Press), in press.
- Colville, E. J., Carlson, A. E., Beard, B. L., Hatfield, R. G., Stoner, J. S., Reyes, A. V., et al. (2011). Sr-Nd-Pb isotope evidence for ice-sheet presence on southern Greenland during the Last Interglacial. *Science* 333, 620–623.
- Comiso, J. C. (2017). Bootstrap Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I-SSMIS, Version 3. Boulder, Color. USA. NASA Natl. Snow Ice Data Cent. Distrib. Act. Arch. Cent. doi:10.5067/7Q8HCCWS4I0R.
- Comiso, J. C., Gersten, R. A., Stock, L. V., Turner, J., Perez, G. J. A. Y. J., Kohei, C. H. O., et al. (2017a). Positive Trend in the Antarctic Sea Ice Cover and Associated Changes in Surface Temperature. *J. Clim.* 30, 2251–2267. doi:10.1175/JCLI-D-16-0408.1.
- Comiso, J. C., Meier, W. N., and Gersten, R. (2017b). Variability and trends in the Arctic Sea ice cover: Results from different techniques. *J. Geophys. Res. Ocean.* 122, 6883–6900. doi:10.1002/2017JC012768.
- Contoux, C., Dumas, C., Ramstein, G., Jost, A., and Dolan, A. M. (2015). Modelling Greenland ice sheet inception and sustainability during the Late Pliocene. *Earth Planet. Sci. Lett.* 424, 295–305.
- Cook, A. J., Copland, L., Noël, B. P. Y., Stokes, C. R., Bentley, M. J., Sharp, M. J., et al. (2019a). Atmospheric forcing of rapid marine-terminating glacier retreat in the Canadian Arctic Archipelago. *Sci. Adv.* 5, eaau8507. doi:10.1126/sciadv.aau8507.
- Cook, A. J., Holland, P. R., Meredith, M. P., Murray, T., Luckman, A., and Vaughan, D. G. (2016). Ocean forcing of glacier retreat in the western Antarctic Peninsula. *Science* doi:10.1126/science.aae0017.
- Cook, A. J., and Vaughan, D. G. (2010). Overview of areal changes of the ice shelves on the Antarctic Peninsula over the past 50 years. *Cryosphere* 4, 77–98. doi:10.5194/tc-4-77-2010.

- Cook, J. M., Tedstone, A. J., Williamson, C., McCutcheon, J., Hodson, A. J., Dayal, A., et al. (2019b). Glacier algae accelerate melt rates on the western Greenland Ice Sheet. *Cryosph. Discuss.*, 1–31. doi:10.5194/tc-2019-58.
- Cornford, S. L., Martin, D. F., Lee, V., Payne, A. J., and Ng, E. G. (2016). Adaptive mesh refinement versus subgrid friction interpolation in simulations of Antarctic ice dynamics. *Ann. Glaciol.* 57, 1–9. doi:10.1017/aog.2016.13.
- Cornford, S. L., Martin, D. F., Payne, A. J., Ng, E. G., Le Brocq, A. M., Gladstone, R. M., et al. (2015). Century-scale simulations of the response of the West Antarctic Ice Sheet to a warming climate. *Cryosph.* 9, 1579–1600. doi:10.5194/tc-9-1579-2015.
- Couasnon, A., Eilander, D., Muis, S., Veldkamp, T. I. E., Haigh, I. D., Wahl, T., et al. (2019). Measuring compound flood potential from river discharge and storm surge extremes at the global scale and its implications for flood hazard. *Nat. Hazards Earth Syst. Sci. Discuss.* doi:10.5194/nhess-2019-205.
- Couldrey, M. P., Gregory, J., Garuba, O., Griffies, S. M., Haak, H., Hu, A., et al. (submitted). What causes the spread of model projections of ocean dynamic sea level change in response to greenhouse gas forcing? *Clim. Dyn.* (submitted).
- Cowton, T. R., Sole, A. J., Nienow, P. W., Slater, D. A., and Christoffersen, P. (2018). Linear response of east Greenland's tidewater glaciers to ocean/atmosphere warming. *Proc. Natl. Acad. Sci.* 115, 7907–7912. doi:10.1073/pnas.1801769115.
- Cox, C. J., Uttal, T., Long, C. N., Stone, R. S., Shupe, M. D., and Starkweather, S. (2016). The role of springtime Arctic clouds in determining autumn sea ice extent. *J. Clim.* 29, 6581–6596. doi:10.1175/JCLI-D-16-0136.1.
- Craig, P. D., and Banner, M. L. (1994). Modeling Wave-Enhanced Turbulence in the Ocean Surface Layer. *J. Phys. Oceanogr.* 24, 2546–2559. doi:10.1175/1520-0485(1994)024<2546:MWETIT>2.0.CO;2.
- Cravatte, S., Delcroix, T., Zhang, D., McPhaden, M., and Leloup, J. (2009). Observed freshening and warming of the western Pacific Warm Pool. *Clim. Dyn.* doi:10.1007/s00382-009-0526-7.
- Csatho, B., Schenk, T., Van Der Veen, C. J., and Krabill, W. B. (2008). Intermittent thinning of Jakobshavn Isbræ, West Greenland, since the Little Ice Age. *J. Glaciol.* 54, 131–144. doi:DOI: 10.3189/002214308784409035.
- Cuffey, K. M., Clow, G. D., Steig, E. J., Buizert, C., Fudge, T. J., Koutnik, M., et al. (2016). Deglacial temperature history of West Antarctica. *Proc. Natl. Acad. Sci.* 113, 14249 LP – 14254. doi:10.1073/pnas.1609132113.
- Cullather, R. I., Nowicki, S. M. J., Zhao, B., and Suarez, M. J. (2014). Evaluation of the Surface Representation of the Greenland Ice Sheet in a General Circulation Model. *J. Clim.* 27, 4835–4856. doi:10.1175/JCLI-D-13-00635.1.
- Cunningham, S. A., and Alderson, S. (2007). Transatlantic temperature and salinity changes at 24.5°N from 1957 to 2004. *Geophys. Res. Lett.* 34, L14606. doi:10.1029/2007GL029821.
- Cuntz, M., and Haverd, V. (2018). Physically Accurate Soil Freeze-Thaw Processes in a Global Land Surface Scheme. *J. Adv. Model. Earth Syst.* 10, 54–77. doi:10.1002/2017MS001100.
- D'Asaro, E. A. (2014). Turbulence in the Upper-Ocean Mixed Layer. *Ann. Rev. Mar. Sci.* 6, 101–115. doi:10.1146/annurev-marine-010213-135138.
- Dahl-Jensen, D., Albert, M. R., Aldahan, A., Azuma, N., Balslev-Clausen, D., Baumgartner, M., et al. (2013). Eemian interglacial reconstructed from a Greenland folded ice core. *Nature* 493, 489–494. doi:10.1038/nature11789.
- Danabasoglu, G. (2008). On Multidecadal Variability of the Atlantic Meridional Overturning Circulation in the Community Climate System Model Version 3. *J. Clim.* 21, 5524–5544. doi:10.1175/2008JCLI2019.1.
- Danabasoglu, G., Large, W. G., and Briegleb, B. P. (2010). Climate impacts of parameterized Nordic Sea overflows. *J. Geophys. Res.* 115, C11005. doi:10.1029/2010JC006243.
- Dangendorf, S., Hay, C. C., Calafat, F. M., Marcos, M., Berk, K., and Jensen, J. (submitted). Persistent acceleration in global sea-level rise since the 1970s. (submitted).
- Dangendorf, S., Marcos, M., Wöppelmann, G., Conrad, C. P., Frederikse, T., and Riva, R. (2017). Reassessment of 20th century global mean sea level rise. *Proc. Natl. Acad. Sci.* 114, 5946–5951. doi:10.1073/pnas.1616007114.
- Dangendorf, S., Mudersbach, C., Jensen, J., Anette, G., and Heinrich, H. (2013). Seasonal to decadal forcing of high water level percentiles in the German Bight throughout the last century. *Ocean Dyn.* doi:10.1007/s10236-013-0614-4.
- Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gundestrup, N. S., Hammer, C. U., et al. (1993). Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364, 218–220. doi:10.1038/364218a0.
- Darmaraki, S., Somot, S., Sevault, F., Nabat, P., Cabos Narvaez, W. D., Cavicchia, L., et al. (2019). Future evolution of Marine Heatwaves in the Mediterranean Sea. *Clim. Dyn.* 53, 1371–1392. doi:10.1007/s00382-019-04661-z.
- De Rydt, J., and Gudmundsson, G. H. (2016). Coupled ice shelf-ocean modeling and complex grounding line retreat from a seabed ridge. *J. Geophys. Res. Earth Surf.* 121, 865–880. doi:10.1002/2015JF003791.
- De Angelis, H., and Skvarca, P. (2003). Glacier surge after ice shelf collapse. *Science* 299, 1560–1562. doi:10.1126/science.1077987.
- De Boer, B., Dolan, A. M., Bernal, J., Gasson, E., Goelzer, H., Golledge, N. R., et al. (2015). Simulating the Antarctic ice sheet in the late-Pliocene warm period: PLISMIP-ANT, an ice-sheet model intercomparison project. *Cryosphere* 9. doi:10.5194/tc-9-881-2015.
- de Boer, B., Haywood, A. M., Dolan, A. M., Hunter, S. J., and Prescott, C. L. (2017). The transient response of ice volume to orbital forcing during the warm late Pliocene. *Geophys. Res. Lett.* 44, 10–486.

- de Boyer Montégut, C., Madec, G., Fischer, A. S., Lazar, A., and Iudicone, D. (2004). Mixed layer depth over the global ocean: an examination of profile data and a profile-based climatology. *J. Geophys. Res.* 109, C12003. Available at: http://www.google.fr/search?client=safari&rls=10_7_4&q=Mixed+layer+depth+over+the+global+ocean+an+examination+of+profile+data+and+a+profile+based+climatology&ie=UTF-8&oe=UTF-8&redir_esc=&ei=Wz_PUP3eAoTDhAfNuoCwAQ.
- de Elía, R., Biner, S., and Frigon, A. (2013). Interannual variability and expected regional climate change over North America. *Clim. Dyn.* 41, 1245–1267. doi:10.1007/s00382-013-1717-9.
- de Kok, R. J., Tuinenburg, O. A., Bonekamp, P. N. J., and Immerzeel, W. W. (2018). Irrigation as a Potential Driver for Anomalous Glacier Behavior in High Mountain Asia. *Geophys. Res. Lett.* 45, 2047–2054. doi:10.1002/2017GL076158.
- de la Peña, S., Howat, I. M., Nienow, P. W., van den Broeke, M. R., Mosley-Thompson, E., Price, S. F., et al. (2015). Changes in the firm structure of the western Greenland Ice Sheet caused by recent warming. *Cryosph.* 9, 1203–1211. doi:10.5194/tc-9-1203-2015.
- De Lavergne, C., Palter, J. B., Galbraith, E. D., Bernardello, R., and Marinov, I. (2014). Cessation of deep convection in the open Southern Ocean under anthropogenic climate change. *Nat. Clim. Chang.* 4, 278. doi:10.1038/nclimate2132.
- De Santis, A., Maier, E., Gomez, R., and Gonzalez, I. (2017). Antarctica, 1979–2016 sea ice extent: total *versus* regional trends, anomalies, and correlation with climatological variables. *Int. J. Remote Sens.* 38, 7566–7584. doi:10.1080/01431161.2017.1363440.
- De Vernal, A., Gersonde, R., Goosse, H., Seidenkrantz, M. S., and Wolff, E. W. (2013a). Sea ice in the paleoclimate system: The challenge of reconstructing sea ice from proxies - an introduction. *Quat. Sci. Rev.* 79, 1–8. doi:10.1016/j.quascirev.2013.08.009.
- De Vernal, A., Hillaire-Marcel, C., Rochon, A., Fréchette, B., Henry, M., Solignac, S., et al. (2013b). Dinocyst-based reconstructions of sea ice cover concentration during the Holocene in the Arctic Ocean, the northern North Atlantic Ocean and its adjacent seas. *Quat. Sci. Rev.* 79, 111–121. doi:10.1016/j.quascirev.2013.07.006.
- de Vries, P., and Weber, S. L. (2005). The Atlantic freshwater budget as a diagnostic for the existence of a stable shut down of the meridional overturning circulation. *Geophys. Res. Lett.* 32. doi:10.1029/2004GL021450.
- Dean, R., and John, G. (2011). The global ocean imprint of ENSO. *Geophys. Res. Lett.* 38. doi:10.1029/2011GL047992.
- Deconto, R. M., and Pollard, D. (2016). Contribution of Antarctica to past and future sea-level rise. *Nature* 531, 591–597. doi:10.1038/nature17145.
- DeConto, R. M., Pollard, D., Alley, R. B., Velicogna, I., and Gasson, E. (submitted). The Paris Climate Agreement and future sea-level rise from Antarctica. (submitted).
- DeConto, R. M., Pollard, D., and Kowalewski, D. E. (2012). Modeling Antarctic ice sheet and climate variations during Marine Isotope Stage 31. *Glob. Planet. Change* 88–89, 45–52.
- Delhasse, A., Fettweis, X., Kittel, C., Amory, C., and Agosta, C. (2018). Brief communication: Impact of the recent atmospheric circulation change in summer on the future surface mass balance of the Greenland Ice Sheet. *Cryosph.* 12, 3409–3418. doi:10.5194/tc-12-3409-2018.
- Deline, P., Gruber, S., Delaloye, R., Fischer, L., Geertsema, M., Giardino, M., et al. (2015). “Ice Loss and Slope Stability in High-Mountain Regions,” in *Snow and Ice-Related Hazards, Risks, and Disasters*, eds. John F. Shroder, W. Haeberli, and C. Whiteman (Academic Press), 521–561. doi:10.1016/B978-0-12-394849-6.00015-9.
- Delman, A. S., McClean, J. L., Sprintall, J., Talley, L. D., Yulaeva, E., Jayne, S. R., et al. (2015). Effects of Eddy Vorticity Forcing on the Mean State of the Kuroshio Extension. *J. Phys. Oceanogr.* 45, 1356–1375. doi:10.1175/JPO-D-13-0259.1.
- Delworth, T. L., Rosati, A., Anderson, W., Adcroft, A. J., Balaji, V., Benson, R., et al. (2012). Simulated Climate and Climate Change in the GFDL CM2.5 High-Resolution Coupled Climate Model. *J. Clim.* 25, 2755–2781. doi:10.1175/JCLI-D-11-00316.1.
- Depoorter, M. A., Bamber, J. L., Griggs, J. A., Lenaerts, J. T. M., Ligtenberg, S. R. M., Van Den Broeke, M. R., et al. (2013). Calving fluxes and basal melt rates of Antarctic ice shelves. *Nature* 502, 89–92. doi:10.1038/nature12567.
- Derksen, C., Brown, R., Mudryk, L., and Luojus, K. (2017). “Terrestrial Snow,” in *State of the Climate in 2016*, S143–S145. doi:10.1175/2017BAMSStateoftheClimate.1.
- Desbiolles, F., Bentamy, A., Blanke, B., Roy, C., Mestas-Nuñez, A. M., Grodsky, S. A., et al. (2017). Two decades [1992–2012] of surface wind analyses based on satellite scatterometer observations. *J. Mar. Syst.* 168, 38–56. doi:10.1016/J.JMARSYS.2017.01.003.
- Desbruyères, D. G., Purkey, S. G., McDonagh, E. L., Johnson, G. C., and King, B. A. (2016). Deep and abyssal ocean warming from 35 years of repeat hydrography. *Geophys. Res. Lett.* 43, 310–356. doi:10.1002/2016GL070413.
- Desbruyères, D., McDonagh, E. L., King, B. A., and Thierry, V. (2017). Global and Full-Depth Ocean Temperature Trends during the Early Twenty-First Century from Argo and Repeat Hydrography. *J. Clim.* 30, 1985–1997. doi:10.1175/JCLI-D-16-0396.1.

- Deshayes, J., Curry, R., Msadek, R., Deshayes, J., Curry, R., and Msadek, R. (2014). CMIP5 Model Intercomparison of Freshwater Budget and Circulation in the North Atlantic. *J. Clim.* 27, 3298–3317. doi:10.1175/JCLI-D-12-00700.1.
- Devlin, A. T., Jay, D. A., Zaron, E. D., Talke, S. A., Pan, J., and Lin, H. (2017). Tidal Variability Related to Sea Level Variability in the Pacific Ocean. *J. Geophys. Res. Ocean.* doi:10.1002/2017JC013165.
- DeVries, T., and Primeau, F. (2011). Dynamically-and observationally-constrained estimates of water-mass distributions and ages in the global ocean. *J. Phys. Oceanogr.* 41, 2381–2401. Available at: <http://journals.ametsoc.org/doi/pdf/10.1175/JPO-D-10-05011.1>.
- Di Lorenzo, E., and Mantua, N. (2016). Multi-year persistence of the 2014/15 North Pacific marine heatwave. *Nat. Clim. Chang.* 6, 1–7. doi:10.1038/nclimate3082.
- Di Lorenzo, E., Miller, A. J., Schneider, N., and McWilliams, J. C. (2005). The warming of the California Current System: Dynamics and ecosystem implications. *J. Phys. Oceanogr.* 35, 336–362.
- Dias, F. B., Fiedler, R., Marsland, S. J., Domingues, C. M., Clement, L., Rintoul, S. R., et al. (submitted). Ocean heat uptake driven by redistribution of heat in response to ocean circulation changes. *J. Clim.* (submitted).
- Dieng, H. B., Cazenave, A., Meyssignac, B., and Ablain, M. (2017). New estimate of the current rate of sea level rise from a sea level budget approach. *Geophys. Res. Lett.* doi:10.1002/2017GL073308.
- Dietrich, J. C., Tanaka, S., Westerink, J. J., Dawson, C. N., Luettich, R. A., Zijlema, M., et al. (2012). Performance of the Unstructured-Mesh, SWAN+ADCIRC Model in Computing Hurricane Waves and Surge. *J. Sci. Comput.* 52, 468–497. doi:10.1007/s10915-011-9555-6.
- Dima, M., and Lohmann, G. (2010). Evidence for Two Distinct Modes of Large-Scale Ocean Circulation Changes over the Last Century. *J. Clim.* 23, 5–16. doi:10.1175/2009JCLI2867.1.
- Ding, Q., Schweiger, A., L’Heureux, M., Battisti, D. S., Po-Chedley, S., Johnson, N. C., et al. (2017). Influence of high-latitude atmospheric circulation changes on summertime Arctic sea ice. *Nat. Clim. Chang.* 7, 289–295. doi:10.1038/nclimate3241.
- Ding, Q., Schweiger, A., L’Heureux, M., Steig, E. J., Battisti, D. S., Johnson, N. C., et al. (2019). Fingerprints of internal drivers of Arctic sea ice loss in observations and model simulations. *Nat. Geosci.* 12, 28. doi:10.1038/s41561-018-0256-8.
- Dinniman, M. S., Asay-Davis, X. S., Galton-Fenzi, B. K., Holland, P. R., Jenkins, A., and Timmermann, R. (2016). Modeling Ice Shelf/Ocean in Antarctica: A REVIEW. *OCEANOGRAPHY* 29, 144–153. doi:10.5670/oceanog.2016.106.
- Docquier, D., Massonnet, F., Barthélemy, A., Tandon, N. F., Lecomte, O., and Fichet, T. (2017). Relationships between Arctic sea ice drift and strength modelled by NEMO-LIM3.6. *Cryosph.* 11, 2829–2846. doi:10.5194/tc-11-2829-2017.
- Doddridge, E. W., and Marshall, J. (2017). Modulation of the Seasonal Cycle of Antarctic Sea Ice Extent Related to the Southern Annular Mode. *Geophys. Res. Lett.* 44, 9761–9768. doi:10.1002/2017GL074319.
- Doddridge, E. W., Marshall, J., Song, H., Campin, J.-M., Kelley, M., and Nazarenko, L. (2019). Eddy Compensation Dampens Southern Ocean Sea Surface Temperature Response to Westerly Wind Trends. *Geophys. Res. Lett.* 46, 4365–4377. doi:10.1029/2019GL082758.
- Doell, P., Mueller Schmied, H., Schuh, C., Portmann, F. T., and Eicker, A. (2014). Global-scale assessment of groundwater depletion and related groundwater abstractions: Combining hydrological modeling with information from well observations and GRACE satellites. *Water Resour. Res.* 50, 5698–5720.
- Dolan, A. M., Haywood, A. M., Hill, D. J., Dowsett, H. J., Hunter, S. J., Lunt, D. J., et al. (2011). Sensitivity of Pliocene ice sheets to orbital forcing. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 309, 98–110.
- Dolan, A. M., Hunter, S. J., Hill, D. J., Haywood, A. M., Koenig, S. J., Otto-Bliesner, B. L., et al. (2015). Using results from the PlioMIP ensemble to investigate the Greenland Ice Sheet during the mid-Pliocene Warm Period. *Clim. Past* 11, 403–424.
- Domine, F., Barrere, M., and Sarrazin, D. (2016). Seasonal evolution of the effective thermal conductivity of the snow and the soil in high Arctic herb tundra at Bylot Island, Canada. *Cryosph.* 10, 2573–2588. doi:10.5194/tc-10-2573-2016.
- Domine, F., Gallet, J. C., Bock, J., and Morin, S. (2012). Structure, specific surface area and thermal conductivity of the snowpack around Barrow, Alaska. *J. Geophys. Res. Atmos.* 117, D00R14. doi:10.1029/2011JD016647.
- Domingues, C. M., Church, J. A., White, N. J., Gleckler, P. J., Wijffels, S. E., Barker, P. M., et al. (2008). Improved estimates of upper-ocean warming and multi-decadal sea-level rise. *Nature* 453, 1090–1093. doi:10.1038/nature07080.
- Donat-Magnin, M., Jourdain, N. C., Spence, P., Le Sommer, J., Gallée, H., and Durand, G. (2017). Ice-Shelf Melt Response to Changing Winds and Glacier Dynamics in the Amundsen Sea Sector, Antarctica. *J. Geophys. Res. Ocean.* 122, 10206–10224. doi:10.1002/2017JC013059.
- Dong, S., Baringer, M. O., and Goni, G. J. (2019). Slow Down of the Gulf Stream during 1993–2016. *Sci. Rep.* 9, 6672. doi:10.1038/s41598-019-42820-8.
- Donohue, K. A., Tracey, K. L., Watts, D. R., Chidichimo, M. P., and Chereskin, T. K. (2016). Mean Antarctic Circumpolar Current transport measured in Drake Passage. *Geophys. Res. Lett.* 43, 11,711–760,767.

- doi:10.1002/2016GL070319.
- Dotto, T. S., Naveira Garabato, A., Bacon, S., Tsamados, M., Holland, P. R., Hooley, J., et al. (2018). Variability of the Ross Gyre, Southern Ocean: Drivers and Responses Revealed by Satellite Altimetry. *Geophys. Res. Lett.* 45, 6195–6204. doi:10.1029/2018GL078607.
- Dotto, T. S., Naveira Garabato, A. C., Bacon, S., Holland, P. R., Kimura, S., Firing, Y. L., et al. (2019). Wind-Driven Processes Controlling Oceanic Heat Delivery to the Amundsen Sea, Antarctica. *J. Phys. Oceanogr.* 49, 2829–2849. doi:10.1175/JPO-D-19-0064.1.
- Downes, S. M., and Hogg, A. M. (2013). Southern Ocean Circulation and Eddy Compensation in CMIP5 Models. *J. Clim.* 26, 7198–7220. doi:10.1175/JCLI-D-12-00504.1.
- Downes, S. M., Spence, P., and Hogg, A. M. (2018). Understanding variability of the Southern Ocean overturning circulation in CORE-II models. *Ocean Model.* 123, 98–109. doi:https://doi.org/10.1016/j.ocemod.2018.01.005.
- Downs, J., Johnson, J., Briner, J., Young, N., Lesnek, A., and Cuzzone, J. (2019). West Greenland ice sheet retreat history reveals elevated precipitation during the Holocene thermal maximum. *Cryosph. Discuss.* 2019, 1–29. doi:10.5194/tc-2019-129.
- Dowsett, H., Dolan, A., Rowley, D., Pound, M., Salzmann, U., Robinson, M., et al. (2016). The PRISM4 (mid-Piacenzian) palaeoenvironmental reconstruction. *Clim. Past Discuss.* 12, 1519–1538.
- Dowsett, H. J., Foley, K. M., Stoll, D. K., Chandler, M. A., Sohl, L. E., Bentsen, M., et al. (2013). Sea Surface Temperature of the mid-Piacenzian Ocean: A Data-Model Comparison. *Sci. Rep.* 3, 2013. doi:10.1038/srep02013.
- Dowsett, H. J., Robinson, M. M., Haywood, A. M., Hill, D. J., Dolan, A. M., Stoll, D. K., et al. (2012). Assessing confidence in Pliocene sea surface temperatures to evaluate predictive models. *Nat. Clim. Chang.* 2, 365–371. doi:10.1038/nclimate1455.
- Dozier, J., Bair, E. H., and Davis, R. E. (2016). Estimating the spatial distribution of snow water equivalent in the world's mountains. *Wiley Interdiscip. Rev. Water* 3, 461–474. doi:10.1002/wat2.1140.
- Drews, R. (2015). Evolution of ice-shelf channels in Antarctic ice shelves. *Cryosph.* 9, 1169–1181. doi:10.5194/tc-9-1169-2015.
- Drijfhout, S., Bathiany, S., Beaulieu, C., Brovkin, V., Claussen, M., Huntingford, C., et al. (2015). Catalogue of abrupt shifts in Intergovernmental Panel on Climate Change climate models. *Proc. Natl. Acad. Sci.* 112, E5777–E5786. doi:10.1073/pnas.1511451112.
- Drijfhout, S. S., Weber, S. L., and van der Waluw, E. (2011). The stability of the MOC as diagnosed from model projections for pre-industrial, present and future climates. *Clim. Dyn.* doi:10.1007/s00382-010-0930-z.
- Drushka, K., Asher, W. E., Ward, B., and Walesby, K. (2016). Understanding the formation and evolution of rain-formed fresh lenses at the ocean surface. *J. Geophys. Res. Ocean.* 121, 2673–2689. doi:10.1002/2015JC011527.
- Drushka, K., Gille, S. T., and Sprintall, J. (2014a). The diurnal salinity cycle in the tropics. *J. Geophys. Res. Ocean.* 119, 5874–5890. doi:10.1002/2014JC009924.
- Drushka, K., Sprintall, J., and Gille, S. T. (2014b). Subseasonal variations in salinity and barrier-layer thickness in the eastern equatorial Indian Ocean. *J. Geophys. Res. Ocean.* 119, 805–823. doi:10.1002/2013JC009422.
- Du, Y., Zhang, Y., Feng, M., Wang, T., Zhang, N., and Wijffels, S. (2015). Decadal trends of the upper ocean salinity in the tropical Indo-Pacific since mid-1990s. *Sci. Rep.* 5, 16050. doi:10.1038/srep16050.
- Dufour, C. O., Morrison, A. K., Griffies, S. M., Frenger, I., Zanowski, H., and Winton, M. (2017). Preconditioning of the Weddell Sea Polynya by the Ocean Mesoscale and Dense Water Overflows. *J. Clim.* 30, 7719–7737. doi:10.1175/JCLI-D-16-0586.1.
- Durack, P. J. (2015). Ocean Salinity and the Global Water Cycle. *Oceanography* 28. Available at: <https://doi.org/10.5670/oceanog.2015.03>.
- Durack, P. J., Gleckler, P. J., Purkey, S. G., Johnson, G. C., Lyman, J. M., and Boyer, T. P. (2018). Ocean Warming. *Oceanography* 31, 41–51.
- Durack, P. J., and Wijffels, S. E. (2010). Fifty-Year trends in global ocean salinities and their relationship to broad-scale warming. *J. Clim.* doi:10.1175/2010JCLI3377.1.
- Durack, P. J., Wijffels, S. E., and Boyer, T. P. (2013). “Chapter 28 - Long-term Salinity Changes and Implications for the Global Water Cycle,” in *Ocean Circulation and Climate*, eds. G. Siedler, S. M. Griffies, J. Gould, and J. A. B. T.-I. G. Church (Academic Press), 727–757. doi:https://doi.org/10.1016/B978-0-12-391851-2.00028-3.
- Durack, P. J., Wijffels, S. E., and Gleckler, P. J. (2014). Long-term sea-level change revisited: The role of salinity. *Environ. Res. Lett.* 9. doi:10.1088/1748-9326/9/11/114017.
- Durack, P. J., Wijffels, S. E., and Matear, R. J. (2012). Ocean salinities reveal strong global water cycle intensification during 1950 to 2000. *Science* doi:10.1126/science.1212222.
- Durand, G., and Pattyn, F. (2015). Reducing uncertainties in projections of Antarctic ice mass loss. *Cryosph.* 9, 2043–2055. doi:10.5194/tc-9-2043-2015.
- Dussaillant, I., Berthier, E., Brun, F., Masiokas, M., Hugonnet, R., Favier, V., et al. (2019). Two decades of glacier mass loss along the Andes. *Nat. Geosci.* 12, 802–808. doi:10.1038/s41561-019-0432-5.
- Dutton, A., Carlson, A. E., Long, A. J., Milne, G. A., Clark, P. U., DeConto, R., et al. (2015a). Sea-level rise due to polar ice-sheet mass loss during past warm periods. *Science* 349, aaa4019. doi:10.1126/science.aaa4019.
- Dutton, A., Webster, J. M., Zwartz, D., Lambeck, K., and Wohlfarth, B. (2015b). Tropical tales of polar ice: evidence

- of Last Interglacial polar ice sheet retreat recorded by fossil reefs of the granitic Seychelles islands. *Quat. Sci. Rev.* 107, 182–196.
- Dziewonski, A. M., and Anderson, D. L. (1981). Preliminary reference Earth model (PREM). *Phys. Earth Planet. Inter.* 25, 297–356. doi:10.1016/0031-9201(81)90046-7.
- Echevin, V., Goubanova, K., Belmadani, A., and Dewitte, B. (2012). Sensitivity of the Humboldt Current system to global warming: a downscaling experiment of the IPSL-CM4 model. *Clim. Dyn.* 38, 761–774.
- Eden, C., and Willebrand, J. (2001). Mechanism of Interannual to Decadal Variability of the North Atlantic Circulation. *J. Clim.* 14, 2266–2280. doi:10.1175/1520-0442(2001)014<2266:MOITDV>2.0.CO;2.
- Edinburgh, T., and Day, J. J. (2016). Estimating the extent of Antarctic summer sea ice during the Heroic Age of Antarctic Exploration. *Cryosph.* 10, 2721–2730. doi:10.5194/tc-10-2721-2016.
- Edwards, T. L., Brandon, M. A., Durand, G., Edwards, N. R., Golledge, N. R., Holden, P. B., et al. (2019a). Revisiting Antarctic ice loss due to marine ice cliff instability. *Nature* 566.
- Edwards, T. L., Brandon, M. A. M. A., Durand, G., Edwards, N. R., Golledge, N. R., Holden, P. B., et al. (2019b). Revisiting Antarctic ice loss due to marine ice-cliff instability. *Nature* 566, 58–64. doi:10.1038/s41586-019-0901-4.
- Edwards, T. L. et al. (submitted). Quantifying uncertainties in the land ice contribution to sea level rise this century. (submitted).
- Edwards, T. L., Fettweis, X., Gagliardini, O., Gillet-Chaulet, F., Goelzer, H., Gregory, J. M., et al. (2014). Effect of uncertainty in surface mass balance-elevation feedback on projections of the future sea level contribution of the Greenland ice sheet. *Cryosphere* 8. doi:10.5194/tc-8-195-2014.
- Edwards, T. L., Nowicki, S., Goelzer, H., Seroussi, H., Smith, C. J., Jourdain, N. C., et al. (submitted). Quantifying uncertainties in the land ice contribution to sea level rise this century. *Nature* (submitted).
- Egbert, G. D., Ray, R. D., and Bills, B. G. (2004). Numerical modeling of the global semidiurnal tide in the present day and in the last glacial maximum. *J. Geophys. Res. Ocean.* doi:10.1029/2003jc001973.
- Ehlert, D., and Zickfeld, K. (2018). Irreversible ocean thermal expansion under carbon dioxide removal. *Earth Syst. Dyn.* 9, 197–210. doi:10.5194/esd-9-197-2018.
- Eisenman, I. (2012). Factors controlling the bifurcation structure of sea ice retreat. *J. Geophys. Res. Atmos.* 117. doi:10.1029/2011JD016164.
- Elipot, S., and Beal, L. M. (2018). Observed Agulhas Current sensitivity to interannual and long-term trend atmospheric forcings. *J. Clim.*, JCLI-D-17-0597.1. doi:10.1175/JCLI-D-17-0597.1.
- Enderlin, E. M., Howat, I. M., Jeong, S., Noh, M.-J., van Angelen, J. H., and van den Broeke, M. R. (2014). An improved mass budget for the Greenland ice sheet. *Geophys. Res. Lett.* 41, 866–872. doi:10.1002/2013GL059010.
- England, M. H., McGregor, S., Spence, P., Meehl, G. A., Timmermann, A., Cai, W., et al. (2014). Recent intensification of wind-driven circulation in the Pacific and the ongoing warming hiatus. *Nat. Clim. Chang.* doi:10.1038/nclimate2106.
- Essery, R. (2015). A factorial snowpack model (FSM 1.0). *Geosci. Model Dev.* 8, 3867–3876. doi:10.5194/gmd-8-3867-2015.
- Estilow, T. W., Young, A. H., and Robinson, D. A. (2015). A long-term Northern Hemisphere snow cover extent data record for climate studies and monitoring. *Earth Syst. Sci. Data* 7, 137–142. doi:10.5194/essd-7-137-2015.
- Eyring, V., Bony, S., Meehl, G. A., Senior, C. A., Stevens, B., Stouffer, R. J., et al. (2016). Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. *Geosci. Model Dev.* 9, 1937–1958. doi:10.5194/gmd-9-1937-2016.
- Fan, T., Deser, C., and Schneider, D. P. (2014). Recent Antarctic sea ice trends in the context of Southern Ocean surface climate variations since 1950. *Geophys. Res. Lett.* 41, 2419–2426. doi:10.1002/2014GL059239.
- Farinotti, D., Huss, M., Fürst, J. J., Landmann, J., Machguth, H., Maussion, F., et al. (2019). A consensus estimate for the ice thickness distribution of all glaciers on Earth. *Nat. Geosci.* 2019, 1. doi:10.1038/s41561-019-0300-3.
- Farinotti, D., Immerzeel, W. W., Kok, R. J. De, Quincey, D. J., and Dehecq, A. (2020). Manifestations and mechanisms of the Karakoram glacier Anomaly. *Nat. Geosci.* 13, 8–16. doi:10.1038/s41561-019-0513-5.
- Farrell, W. E., and Clark, J. A. (1976). On Postglacial Sea Level. *Geophys. J. R. Astron. Soc.* 46, 647–667. doi:10.1111/j.1365-246X.1976.tb01252.x.
- Fasullo, J. T., and Nerem, R. S. (2018). Altimeter-era emergence of the patterns of forced sea-level rise in climate models and implications for the future. *Proc. Natl. Acad. Sci.* 115, 12944–12949. doi:10.1073/pnas.1813233115.
- Fasullo, J. T., Nerem, R. S., and Hamlington, B. (2016). Is the detection of accelerated sea level rise imminent? *Sci. Rep.* 6, 31245. doi:10.1038/srep31245.
- Fausto, R. S., Box, J. E., Vandecrux, B., van As, D., Steffen, K., MacFerrin, M. J., et al. (2018). A Snow Density Dataset for Improving Surface Boundary Conditions in Greenland Ice Sheet Firm Modeling. *Front. Earth Sci.* 6, 51. doi:10.3389/feart.2018.00051.
- Favier, L., Durand, G., Cornford, S. L., Gudmundsson, G. H., Gagliardini, O., Gillet-Chaulet, F., et al. (2014). Retreat of Pine Island Glacier controlled by marine ice-sheet instability. *Nat. Clim. Chang.* 5, 117–121. doi:10.1038/nclimate2094.
- Favier, L., Jourdain, N. C., Jenkins, A., Merino, N., Durand, G., Gagliardini, O., et al. (2019). Assessment of Sub-Shelf

- Melting Parameterisations Using the Ocean-Ice Sheet Coupled Model NEMO(v3.6)-Elmer/Ice(v8.3). *Geosci. Model Dev. Discuss.* 2019, 1–40. doi:10.5194/gmd-2019-26.
- Favier, V., Agosta, C., Parouty, S., Durand, G., Delaygue, G., Gallée, H., et al. (2013). An updated and quality controlled surface mass balance dataset for Antarctica. *Cryosph.* 7, 583–597. doi:10.5194/tc-7-583-2013.
- Fedorov, A. V., Burls, N. J., Lawrence, K. T., and Peterson, L. C. (2015). Tightly linked zonal and meridional sea surface temperature gradients over the past five million years. *Nat. Geosci.* doi:10.1038/ngeo2577.
- Feldmann, J., and Levermann, A. (2015). Collapse of the West Antarctic Ice Sheet after local destabilization of the Amundsen Basin. *Proc. Natl. Acad. Sci.* 112, 14191–14196. doi:10.1073/pnas.1512482112.
- Feldmann, J., Levermann, A., and Mengel, M. (2019). Stabilizing the West Antarctic Ice Sheet by surface mass deposition. *Sci. Adv.* 5, eaaw4132.
- Feng, M., McPhaden, M. J., and Lee, T. (2010). Decadal variability of the Pacific subtropical cells and their influence on the southeast Indian Ocean. *Geophys. Res. Lett.* 37, n/a-n/a. doi:10.1029/2010GL042796.
- Feng, W., Zhong, M., and Xu, H. Z. (2012). Sea level variations in the South China Sea inferred from satellite gravity, altimetry, and oceanographic data. *Sci. China Earth Sci.* doi:10.1007/s11430-012-4394-3.
- 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.* 111, 8753–8758. doi:10.1073/pnas.1323922111.
- Ferreira, D., Cessi, P., Coxall, H. K., de Boer, A., Dijkstra, H. A., Drijfhout, S. S., et al. (2018). Atlantic-Pacific Asymmetry in Deep Water Formation. *Annu. Rev. Earth Planet. Sci.* 46, 327–352. doi:10.1146/annurev-earth-082517-010045.
- Ferreira, D., Marshall, J., Bitz, C. M., Solomon, S., and Plumb, A. (2015). Antarctic ocean and sea ice response to ozone depletion: A two-time-scale problem. *J. Clim.* 28, 1206–1226. doi:10.1175/JCLI-D-14-00313.1.
- Fettweis, X., Box, J. E., Agosta, C., Amory, C., Kittel, C., Lang, C., et al. (2017). Reconstructions of the 1900–2015 Greenland ice sheet surface mass balance using the regional climate MAR model. *Cryosph.* 11, 1015–1033. doi:10.5194/tc-11-1015-2017.
- Fettweis, X., Franco, B., Tedesco, M., van Angelen, J. H., Lenaerts, J. T. M., van den Broeke, M. R., et al. (2013). Estimating the Greenland ice sheet surface mass balance contribution to future sea level rise using the regional atmospheric climate model MAR. *Cryosph.* 7, 469–489. doi:10.5194/tc-7-469-2013.
- Fischer, H., Meissner, K. J., Mix, A. C., Abram, N. J., Austermann, J., Brovkin, V., et al. (2018). Palaeoclimate constraints on the impact of 2 °C anthropogenic warming and beyond. *Nat. Geosci.* 11, 474–485. doi:10.1038/s41561-018-0146-0.
- Flanner, M. G., Shell, K. M., Barlage, M., Perovich, D. K., and Tschudi, M. A. (2011). Radiative forcing and albedo feedback from the Northern Hemisphere cryosphere between 1979 and 2008. *Nat. Geosci.* 4, 151–155. doi:10.1038/ngeo1062.
- Flato, G., Marotzke, J., Abiodun, B., Braconnot, P., Chou, S. C., Collins, W., et al. (2013). Evaluation of Climate Models. in *{Climate Change 2013}: The Physical Science Basis. {Contribution of Working Group I} to the {Fifth Assessment Report of the IPCC}*, eds. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, et al. (Cambridge University Press), 741–866. Available at: <http://bit.ly/2wNMqXD>.
- Flowers, G. E. (2018). Hydrology and the future of the Greenland Ice Sheet. *Nat. Commun.* 9, 2729. doi:10.1038/s41467-018-05002-0.
- Fogwill, C. J., Phipps, S. J., Turney, C. S. M., and Golledge, N. R. (2015). Sensitivity of the Southern Ocean to enhanced regional Antarctic ice sheet meltwater input. *Earth's Futur.* 3, 317–329. doi:10.1002/2015EF000306.
- Foreman, M. G. G., Callendar, W., Masson, D., Morrison, J., and Fine, I. (2013). A Model Simulation of Future Oceanic Conditions along the British Columbia Continental Shelf. Part II: Results and Analyses. doi:10.1080/07055900.2013.873014.
- Forget, G., and Ferreira, D. (2019). Global ocean heat transport dominated by heat export from the tropical Pacific. *Nat. Geosci.*, 1. doi:10.1038/s41561-019-0333-7.
- Forsberg, R., Sørensen, L., and Simonsen, S. (2017). Greenland and Antarctica Ice Sheet Mass Changes and Effects on Global Sea Level. *Surv. Geophys.* 38, 89–104. doi:10.1007/s10712-016-9398-7.
- Forster, P. M., Maycock, A. C., McKenna, C. M., and Smith, C. J. (2019). Latest climate models confirm need for urgent mitigation. *Nat. Clim. Chang.* doi:10.1038/s41558-019-0660-0.
- Forster, R., Box, J., Van den Broeke, M., Mieke, C., Burgess, E., van angelen, J., et al. (2014). Extensive liquid meltwater storage in firn within the Greenland ice sheet. *Nat. Geosci.* 7, 95–98. doi:10.1038/ngeo2043.
- Forsythe, N., Fowler, H. J., Li, X. F., Blenkinsop, S., and Pritchard, D. (2017). Karakoram temperature and glacial melt driven by regional atmospheric circulation variability. *Nat. Clim. Chang.* 7, 664–670. doi:10.1038/nclimate3361.
- Foster, J. L., Hall, D. K., Kelly, R. E. J., and Chiu, L. (2009). Seasonal snow extent and snow mass in South America using SMMR and SSM/I passive microwave data (1979–2006). *Remote Sens. Environ.* 113, 291–305. doi:10.1016/j.rse.2008.09.010.
- Fox-Kemper, B., Adcroft, A., Böning, C. W., Chassignet, E. P., Curchitser, E., Danabasoglu, G., et al. (2019). Challenges and Prospects in Ocean Circulation Models. *Front. Mar. Sci.* 6, 65. doi:10.3389/fmars.2019.00065.
- Fox-Kemper, B., Danabasoglu, G., Ferrari, R., Griffies, S. M., Hallberg, R. W., Holland, M. M., et al. (2011a).

- Parameterization of mixed layer eddies. III: Implementation and impact in global ocean climate simulations. *Ocean Model.* 39. doi:10.1016/j.ocemod.2010.09.002.
- Fox-Kemper, B., Danabasoglu, G., Ferrari, R., Griffies, S. M. S. M., Hallberg, R. W. R. W., Holland, M. M. M., et al. (2011b). Parameterization of mixed layer eddies. III: Implementation and impact in global ocean climate simulations. *Ocean Model.* 39, 61–78. doi:https://doi.org/10.1016/j.ocemod.2010.09.002.
- Fox-Kemper, B., and Pedlosky, J. (2004). Wind-driven barotropic gyre I: Circulation control by eddy vorticity fluxes to an enhanced removal region. *J. Mar. Res.* 62, 169–193. doi:10.1357/002224004774201681.
- Frederikse, T., Jevrejeva, S., Riva, R. E. M., Dangendorf, S., Frederikse, T., Jevrejeva, S., et al. (2018). A Consistent Sea-Level Reconstruction and Its Budget on Basin and Global Scales over 1958–2014. *J. Clim.* 31, 1267–1280. doi:10.1175/JCLI-D-17-0502.1.
- Frederikse, T., Riva, R., Slobbe, C., Broerse, T., and Verlaan, M. (2016). Estimating decadal variability in sea level from tide gauge records: An application to the North Sea. *J. Geophys. Res. Ocean.* doi:10.1002/2015JC011174.
- Frenger, I., Gruber, N., Knutti, R., and Münnich, M. (2013). Imprint of Southern Ocean eddies on winds, clouds and rainfall. *Nat. Geosci.* 6, 608–612. doi:10.1038/ngeo1863.
- Fretwell, P., Pritchard, H. D., Vaughan, D. G., Bamber, J. L., Barrand, N. E., Bell, R., et al. (2013). Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. *Cryosphere* 7, 375–393. doi:10.5194/tc-7-375-2013.
- Frey, K. E., Moore, G. W. K., Cooper, L. W., and Grebmeier, J. M. (2015). Divergent patterns of recent sea ice cover across the Bering, Chukchi, and Beaufort seas of the Pacific Arctic Region. *Prog. Oceanogr.* 136, 32–49. doi:10.1016/j.pocean.2015.05.009.
- Fried, M. J., Catania, G. A., Bartholomäus, T. C., Duncan, D., Davis, M., Stearns, L. A., et al. (2015). Distributed subglacial discharge drives significant submarine melt at a Greenland tidewater glacier. *Geophys. Res. Lett.* 42, 9328–9336. doi:10.1002/2015GL065806.
- Fried, M. J., Catania, G. A., Stearns, L. A., Sutherland, D. A., Bartholomäus, T. C., Shroyer, E., et al. (2018). Reconciling Drivers of Seasonal Terminus Advance and Retreat at 13 Central West Greenland Tidewater Glaciers. *J. Geophys. Res. Earth Surf.* 123, 1590–1607. doi:10.1029/2018JF004628.
- Friedman, A. R., Reverdin, G., Khodri, M., and Gastineau, G. (2017). A new record of Atlantic sea surface salinity from 1896 to 2013 reveals the signatures of climate variability and long-term trends. *Geophys. Res. Lett.* 44, 1866–1876. doi:10.1002/2017GL072582.
- Frieler, K., Clark, P. U., He, F., Buizert, C., Reese, R., Ligtenberg, S. R. M., et al. (2015). Consistent evidence of increasing Antarctic accumulation with warming. *Nat. Clim. Chang.* 5, 348–352. doi:10.1038/nclimate2574.
- Frieler, K., Lange, S., Piontek, F., Reyer, C. P. O., Schewe, J., Warszawski, L., et al. (2017). Assessing the impacts of 1.5 °C global warming – simulation protocol of the Inter-Sectoral Impact Model Intercomparison Project (ISIMIP2b). *Geosci. Model Dev.* 10, 4321–4345. doi:10.5194/gmd-10-4321-2017.
- Fringer, O. B., Gerritsen, M., and Street, R. L. (2006). An unstructured-grid, finite-volume, nonhydrostatic, parallel coastal ocean simulator. doi:10.1016/j.ocemod.2006.03.006.
- Frölicher, T. L., Fischer, E. M., and Gruber, N. (2018). Marine heatwaves under global warming. *Nature* 560, 360–364. doi:10.1038/s41586-018-0383-9.
- Frölicher, T. L., and Laufkötter, C. (2018). Emerging risks from marine heat waves. *Nat. Commun.* 9, 650. doi:10.1038/s41467-018-03163-6.
- Frölicher, T. L., Sarmiento, J. L., Paynter, D. J., Dunne, J. P., Krasting, J. P., and Winton, M. (2015). Dominance of the Southern Ocean in Anthropogenic Carbon and Heat Uptake in CMIP5 Models. *J. Clim.* 28, 862–886. doi:10.1175/JCLI-D-14-00117.1.
- Funder, S., Goosse, H., Jepsen, H., Kaas, E., Kjær, K. H., Korsgaard, N. J., et al. (2011). A 10,000-year record of Arctic Ocean Sea-ice variability - View from the beach. *Science* 333, 747–750. doi:10.1126/science.1202760.
- Fürst, J. J., Durand, G., Gillet-Chaulet, F., Tavard, L., Rankl, M., Braun, M., et al. (2016). The safety band of Antarctic ice shelves. *Nat. Clim. Chang.* 6, 479–482. doi:10.1038/nclimate2912.
- Fürst, J. J., Goelzer, H., and Huybrechts, P. (2015). Ice-dynamic projections of the Greenland ice sheet in response to atmospheric and oceanic warming. *Cryosph.* 9, 1039–1062.
- Fyfe, J. C., Derksen, C., Mudryk, L., Flato, G. M., Santer, B. D., Swart, N. C., et al. (2017). Large near-Term projected snowpack loss over the western United States. *Nat. Commun.* 8, 14996. doi:10.1038/ncomms14996.
- Fyke, J. G., Sacks, W. J., and Lipscomb, W. H. (2014). A technique for generating consistent ice sheet initial conditions for coupled ice sheet/climate models. *Geosci. Model Dev.* 7, 1183–1195. doi:10.5194/gmd-7-1183-2014.
- Fyke, J. G., Weaver, A. J., Pollard, D., Eby, M., Carter, L., and Mackintosh, A. (2011). A new coupled ice sheet/climate model: description and sensitivity to model physics under Eemian, Last Glacial Maximum, late Holocene and modern climate conditions. *Geosci. Model Dev.* 4, 117–136.
- Fyke, J., Lenaerts, J. T. M., and Wang, H. (2017). Basin-scale heterogeneity in Antarctic precipitation and its impact on surface mass variability. *Cryosph.* 11, 2595–2609. doi:10.5194/tc-11-2595-2017.
- Fyke, J., Sergienko, O., Löfverström, M., Price, S., and Lenaerts, J. T. M. (2018). An Overview of Interactions and Feedbacks Between Ice Sheets and the Earth System. *Rev. Geophys.* 56.
- Gabbi, J., Huss, M., Bauder, A., Cao, F., and Schwikowski, M. (2015). The impact of Saharan dust and black carbon on

- albedo and long-term mass balance of an Alpine glacier. *Cryosph.* 9, 1385–1400. doi:10.5194/tc-9-1385-2015.
- Gagné, M., Gillett, N. P., and Fyfe, J. C. (2015a). Impact of aerosol emission controls on future Arctic sea ice cover. *Geophys. Res. Lett.* 42, 8481–8488. doi:10.1002/2015GL065504.
- Gagné, M., Gillett, N. P., and Fyfe, J. C. (2015b). Observed and simulated changes in Antarctic sea ice extent over the past 50 years. *Geophys. Res. Lett.* 42, 90–95. doi:10.1002/2014GL062231.
- Gallaher, D. W., Campbell, G. G., and Meier, W. N. (2014). Anomalous Variability in Antarctic Sea Ice Extents During the 1960s With the Use of Nimbus Data. *IEEE J. Sel. Top. Appl. Earth Obs. Remote Sens.* 7, 881–887. doi:10.1109/JSTARS.2013.2264391.
- Gallet, J.-C. C., Domine, F., Zender, C. S., and Picard, G. (2009). Measurement of the specific surface area of snow using infrared reflectance in an integrating sphere at 1310 and 1550 nm. *Cryosph.* 3, 167–182. doi:10.5194/tc-3-167-2009.
- Gao, L., Rintoul, S. R., and Yu, W. (2018). Recent wind-driven change in Subantarctic Mode Water and its impact on ocean heat storage. *Nat. Clim. Chang.* 8, 58–63. doi:10.1038/s41558-017-0022-8.
- Garabato, A. C. N., Stevens, D. P., Watson, A. J., and Roether, W. (2007). Short-circuiting of the overturning circulation in the Antarctic Circumpolar Current. *Nature* 447, 194–197. doi:10.1038/nature05832.
- García-Reyes, M., Sydean, W. J., Black, B. A., Rykaczewski, R. R., Schoeman, D. S., Thompson, S. A., et al. (2013). Relative influence of oceanic and terrestrial pressure systems in driving upwelling-favorable winds. *Geophys. Res. Lett.* 40, 5311–5315.
- García-Reyes, M., Largier, J. L., and Sydean, W. J. (2014). Synoptic-scale upwelling indices and predictions of phyto-and zooplankton populations. *Prog. Oceanogr.* 120, 177–188.
- García, D., Chao, B. F., Del Río, J., Vigo, I., and García-Lafuente, J. (2006). On the steric and mass-induced contributions to the annual sea level variations in the Mediterranean Sea. *J. Geophys. Res.* 111, C09030. doi:10.1029/2005JC002956.
- Gardelle, J., Berthier, E., and Arnaud, Y. (2012). Slight mass gain of Karakoram glaciers in the early twenty-first century. *Nat. Geosci.* 5, 322–325. doi:10.1038/ngeo1450.
- Gardner, A. S., Moholdt, G., Scambos, T., Fahnestock, M., Ligtenberg, S., Van Den Broeke, M., et al. (2018). Increased West Antarctic and unchanged East Antarctic ice discharge over the last 7 years. *Cryosphere* 12, 521–547. doi:10.5194/tc-12-521-2018.
- Garner, A. J., Mann, M. E., Emanuel, K. A., Kopp, R. E., Lin, N., Alley, R. B., et al. (2017). Impact of climate change on New York City’s coastal flood hazard: Increasing flood heights from the preindustrial to 2300 CE. *Proc. Natl. Acad. Sci. U. S. A.* 114. doi:10.1073/pnas.1703568114.
- Garner, A. J., Weiss, J. L., Parris, A., Kopp, R. E., Horton, R. M., Overpeck, J. T., et al. (2018). Evolution of 21 st Century Sea-level Rise Projections. *Earth’s Futur.* 6, 1603–1615. doi:10.1029/2018EF000991.
- Garrabou, J., Coma, R., Bensoussan, N., Bally, M., Chevaldonne, P., Cigliano, M., et al. (2009). Mass mortality in Northwestern Mediterranean rocky benthic communities: effects of the 2003 heat wave. *Glob. Chang. Biol.* 15, 1090–1103. doi:10.1111/j.1365-2486.2008.01823.x.
- Garry, F., Roberts, C., Blaker, A., McDonagh, E., Frajka-Williams, E., and King, B. (submitted). Spatial variation of ocean heat content change in CMP5 models and implications for deep ocean observing. *Clim. Dyn.* (submitted).
- Gasson, E., DeConto, R., and Pollard, D. (2016a). Dynamic Antarctic ice sheet during the early to mid-Miocene. *Natl. Acad. Sci.* 113, 3459–3464. Available at: <http://www.pnas.org/content/113/13/3459.short> [Accessed September 26, 2018].
- Gasson, E., DeConto, R. R. M., Geology, D. P., 2016, U., and Pollard, D. (2016b). Modeling the oxygen isotope composition of the Antarctic ice sheet and its significance to Pliocene sea level. *Geology* 44, 827–830. doi:10.1130/G38104.1.
- Gebbie, G., and Huybers, P. (2019). The Little Ice Age and 20th Century Deep Pacific Cooling. *Science* 363, 70–74. doi:10.1126/science.aar8413.
- Gehrels, W. R., and Woodworth, P. L. (2013). When did modern rates of sea-level rise start? *Glob. Planet. Change* 100, 263–277. doi:10.1016/j.gloplacha.2012.10.020.
- Geirsdóttir, Á., Miller, G. H., Andrews, J. T., Harning, D. J., Anderson, L. S., Florian, C., et al. (2019). The onset of neoglaciation in Iceland and the 4.2\,ka event. *Clim. Past* 15, 25–40. doi:10.5194/cp-15-25-2019.
- Gemrich, J., Thomas, B., and Bouchard, R. (2011). Observational changes and trends in northeast Pacific wave records. *Geophys. Res. Lett.* doi:10.1029/2011GL049518.
- Gent, P. R. (2016). Effects of Southern Hemisphere Wind Changes on the Meridional Overturning Circulation in Ocean Models. *Ann. Rev. Mar. Sci.* 8, 79–94. doi:10.1146/annurev-marine-122414-033929.
- Gent, P. R. (2018). A commentary on the Atlantic meridional overturning circulation stability in climate models. *Ocean Model.* 122, 57–66. doi:10.1016/j.ocemod.2017.12.006.
- Gent, P. R., Holland, M. M., Lindsay, R., Woodgate, R. A., and Bitz, C. M. (2006). The Influence of Sea Ice on Ocean Heat Uptake in Response to Increasing CO₂. *J. Clim.* 19, 2437–2450. doi:10.1175/jcli3756.1.
- Genthon, C., Krinner, C., and Castebrunet, H. (2009a). Antarctic precipitation and climate-change predictions: Horizontal resolution and margin vs plateau issues. *Ann. Glaciol.* 50, 55–60. doi:10.3189/172756409787769681.
- Genthon, C., Krinner, G., and Castebrunet, H. (2009b). Antarctic precipitation and climate-change predictions:

- Horizontal resolution and margin vs plateau issues. *Ann. Glaciol.* 50, 55–60. doi:10.3189/172756409787769681.
- Gersonde, R., Crosta, X., Abelmann, A., and Armand, L. (2005). Sea-surface temperature and sea ice distribution of the Southern Ocean at the EPILOG Last Glacial Maximum - A circum-Antarctic view based on siliceous microfossil records. *Quat. Sci. Rev.* 24, 869–896. doi:10.1016/j.quascirev.2004.07.015.
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nat. Commun.* 9, 3041. doi:10.1038/s41467-018-05457-1.
- Giesen, R. H., and Oerlemans, J. (2013). Climate-model induced differences in the 21st century global and regional glacier contributions to sea-level rise. *Clim. Dyn.* 41, 3283–3300. doi:10.1007/s00382-013-1743-7.
- Giglio, D., and Johnson, G. C. (2017). Middepth decadal warming and freshening in the South Atlantic. *J. Geophys. Res. Ocean.* 122, 973–979. doi:10.1002/2016JC012246.
- Gilbert, A., Flowers, G. E., Miller, G. H., Refsnider, K. A., Young, N. E., and Radić, V. (2017). The projected demise of Barnes Ice Cap: Evidence of an unusually warm 21st century Arctic. *Geophys. Res. Lett.* 44, 2810–2816. doi:10.1002/2016GL072394.
- Gilbert, A., and Vincent, C. (2013). Atmospheric temperature changes over the 20th century at very high elevations in the European Alps from englacial temperatures. *Geophys. Res. Lett.* 40, 2102–2108. doi:10.1002/grl.50401.
- Gilbert, A., Vincent, C., Gagliardini, O., Krug, J., and Berthier, E. (2015). Assessment of thermal change in cold avalanching glaciers in relation to climate warming. *Geophys. Res. Lett.* 42, 6382–6390. doi:10.1002/2015GL064838.
- Gille, S. T. (2008). Decadal-Scale Temperature Trends in the Southern Hemisphere Ocean. *J. Clim.* 21, 4749–4765. doi:10.1175/2008JCLI2131.1.
- Ginot, P., Dumont, M., Lim, S., Patris, N., Taupin, J.-D., Wagnon, P., et al. (2014). A 10 year record of black carbon and dust from a Mera Peak ice core (Nepal): variability and potential impact on melting of Himalayan glaciers. *Cryosph.* 8, 1479–1496. doi:10.5194/tc-8-1479-2014.
- Giorgi, F., and Lionello, P. (2008). Climate change projections for the Mediterranean region. *Glob. Planet. Change* 63, 90–104. doi:10.1016/J.GLOPLACHA.2007.09.005.
- Gladish, C. V., Holland, D. M., Rosing-Asvid, A., Behrens, J. W., and Boje, J. (2015). Oceanic Boundary Conditions for Jakobshavn Glacier. Part I: Variability and Renewal of Ilulissat Icefjord Waters, 2001–14. *J. Phys. Oceanogr.* 45, 3–32. doi:10.1175/JPO-D-14-0044.1.
- Gladyshev, S. V., Gladyshev, V. S., Gulev, S. K., and Sokov, A. V. (2017). Subpolar mode water classes in the northeast Atlantic: Interannual and long-term variability. *Dokl. Earth Sci.* 476, 1203–1206. doi:10.1134/S1028334X17100166.
- GlaThiDa Consortium (2019). Glacier Thickness Database 3.0.1. doi:10.5904/wgms-glathida-2019-03.
- Gobiet, A., Kotlarski, S., Beniston, M., Heinrich, G., Rajczak, J., and Stoffel, M. (2014). 21st century climate change in the European Alps-A review. *Sci. Total Environ.* 493, 1138–1151. doi:10.1016/j.scitotenv.2013.07.050.
- Goelzer, H., Huybrechts, P., Fürst, J. J., Nick, F. M., Andersen, M. L., Edwards, T. L., et al. (2013). Sensitivity of Greenland ice sheet projections to model formulations. *J. Glaciol.* 59. doi:10.3189/2013JoG12J182.
- Goelzer, H., Huybrechts, P., Loutre, M.-F. M.-F., and Fichet, T. (2016a). Last Interglacial climate and sea-level evolution from a coupled ice sheet–climate model. *Clim. Past* 12, 2195–2213. doi:10.5194/cp-12-2195-2016.
- Goelzer, H., Huybrechts, P., Loutre, M.-F. M.-F., and Fichet, T. (2016b). Last Interglacial climate and sea-level evolution from a coupled ice sheet–climate model. *Clim. Past* 12, 2195–2213. doi:10.5194/cp-12-2195-2016.
- Goelzer, H., Huybrechts, P., Raper, S. C. B., Loutre, M.-F. F., Goosse, H., and Fichet, T. (2012). Millennial total sea-level commitments projected with the Earth system model of intermediate complexity LOVECLIM. *Environ. Res. Lett.* 7, 045401. doi:10.1088/1748-9326/7/4/045401.
- Goelzer, H., Nowicki, S., Edwards, T., Beckley, M., Abe-Ouchi, A., Aschwanden, A., et al. (2017). Design and results of the ice sheet model initialisation experiments initMIP-Greenland: an ISMIP6 intercomparison. *Cryosph.* 12, 1433–1460. doi:10.5194/tc-2017-129.
- Goelzer, H., Nowicki, S., Edwards, T., Beckley, M., Abe-Ouchi, A., Aschwanden, A., et al. (2018a). Design and results of the ice sheet model initialisation experiments initMIP-Greenland: an ISMIP6 intercomparison. *Cryosph.* 12, 1433–1460. doi:10.5194/tc-12-1433-2018.
- Goelzer, H., Nowicki, S., Edwards, T., Beckley, M., Abe-Ouchi, A., Aschwanden, A., et al. (2018b). Design and results of the ice sheet model initialisation experiments initMIP-Greenland: an ISMIP6 intercomparison. *Cryosph.* 12, 1433–1460. doi:10.5194/tc-12-1433-2018.
- Goelzer, H., Nowicki, S., Payne, A., Larour, E., Seroussi, H., William, H., et al. (submitted). The future sea-level contribution of the Greenland ice sheet: a multi- model ensemble study of ISMIP6. *Cryosph.* (submitted).
- Goldberg, D. N., Gourmelen, N., Kimura, S., Millan, R., and Snow, K. (2018). How accurately should we model ice shelf melt rates? doi:10.1029/2018GL080383.
- Goldberg, D. N., Heimbach, P., Joughin, I., and Smith, B. (2015). Committed retreat of Smith, Pope, and Kohler Glaciers over the next 30 years inferred by transient model calibration. *Cryosph.* 9, 2429–2446. doi:10.5194/tc-9-2429-2015.
- Golledge, N. R. (in press). Long-term projections of sea-level rise from ice sheets. *Wiley Interdiscip. Rev. Clim. Chang.*

- (in press).
- Golledge, N. R., Fogwill, C. J., Mackintosh, A. N., and Buckley, K. M. (2012). Dynamics of the last glacial maximum Antarctic ice-sheet and its response to ocean forcing. *Proc. Natl. Acad. Sci. U. S. A.* 109. doi:10.1073/pnas.1205385109.
- Golledge, N. R., Keller, E. D., Gomez, N., Naughten, K. A., Bernales, J., Trusel, L. D., et al. (2019). Global environmental consequences of twenty-first-century ice-sheet melt. *Nature* 566. doi:10.1038/s41586-019-0889-9.
- Golledge, N. R., Kowalewski, D. E., Naish, T. R., Levy, R. H., Fogwill, C. J., and Gasson, E. G. W. (2015). The multi-millennial Antarctic commitment to future sea-level rise. *Nature* 526. doi:10.1038/nature15706.
- Golledge, N. R., Levy, R. H., McKay, R. M., Fogwill, C. J., White, D. A., Graham, A. G. C., et al. (2013). Glaciology and geological signature of the Last Glacial Maximum Antarctic ice sheet. *Quat. Sci. Rev.* 78. doi:10.1016/j.quascirev.2013.08.011.
- Golledge, N. R., Levy, R. H., McKay, R. M., and Naish, T. R. (2017a). East Antarctic ice sheet most vulnerable to Weddell Sea warming. *Geophys. Res. Lett.* 44, 2343–2351. doi:10.1002/2016GL072422.
- Golledge, N. R., Menviel, L., Carter, L., Fogwill, C. J., England, M. H., Cortese, G., et al. (2014). Antarctic contribution to meltwater pulse 1A from reduced Southern Ocean overturning. *Nat. Commun.* 5. doi:10.1038/ncomms6107.
- Golledge, N. R., Thomas, Z. A., Levy, R. H., Gasson, E. G. W., Naish, T. R., McKay, R. M., et al. (2017b). Antarctic climate and ice-sheet configuration during the early Pliocene interglacial at 4.23 Ma. *Clim. Past* 13. doi:10.5194/cp-13-959-2017.
- Gomez, N., Pollard, D., and Holland, D. (2015). Sea-level feedback lowers projections of future Antarctic Ice-Sheet mass loss. *Nat. Commun.* 6, 8798.
- Gomez, N., Pollard, D., and Mitrovica, J. X. (2013). A 3-D coupled ice sheet – sea level model applied to Antarctica through the last 40 ky. *Earth Planet. Sci. Lett.* 384, 88–99. doi:10.1016/j.epsl.2013.09.042.
- Goosse, H., Kay, J. E., Armour, K. C., Bodas-Salcedo, A., Chepfer, H., Docquier, D., et al. (2018). Quantifying climate feedbacks in polar regions. *Nat. Commun.* 9, 1919. doi:10.1038/s41467-018-04173-0.
- Goosse, H., and Renssen, H. (2001). A two-phase response of the Southern Ocean to an increase in greenhouse gas concentrations. *Geophys. Res. Lett.* 28, 3469–3472. doi:10.1029/2001GL013525.
- Goosse, H., and Zunz, V. (2014). Decadal trends in the Antarctic sea ice extent ultimately controlled by ice-ocean feedback. *Cryosph.* 8, 453–470. doi:10.5194/tc-8-453-2014.
- Gorte, T., Lenaerts, J. T. M., and Medley, B. (2019). Scoring Antarctic surface mass balance in climate models to refine future projections. *Cryosph. Discuss.* in review. doi:10.5194/tc-2019-240.
- Gourmelen, N., Escorihuela, M. J., Shepherd, A., Foresta, L., Muir, A., Garcia-Mondéjar, A., et al. (2018). CryoSat-2 swath interferometric altimetry for mapping ice elevation and elevation change. *Adv. Sp. Res.* 62, 1226–1242. doi:https://doi.org/10.1016/j.asr.2017.11.014.
- Graham, R. M., Itkin, P., Meyer, A., Sundfjord, A., Spreen, G., Smedsrud, L. H., et al. (2019). Winter storms accelerate the demise of sea ice in the Atlantic sector of the Arctic Ocean. *Sci. Rep.* 9. doi:10.1038/s41598-019-45574-5.
- Graham, W. M., and Largier, J. L. (1997). Upwelling shadows as nearshore retention sites: the example of northern Monterey Bay. *Cont. Shelf Res.* 17, 509–532.
- Grant, G. R., Naish, T. R., Dunbar, G. B., Stocchi, P., Kominz, M. A., Kamp, P. J. J., et al. (2019). The amplitude and origin of sea-level variability during the Pliocene epoch. *Nature*. doi:10.1038/s41586-019-1619-z.
- Greenberg, D. A., Dupont, F., Lyard, F. H., Lynch, D. R., and Werner, F. E. (2007). Resolution issues in numerical models of oceanic and coastal circulation. *Cont. Shelf Res.* 27, 1317–1343. doi:10.1016/j.csr.2007.01.023.
- Greene, C. A., Young, D. A., Gwyther, D. E., Galton-Fenzi, B. K., and Blankenship, D. D. (2018). Seasonal dynamics of Totten Ice Shelf controlled by sea ice buttressing. *Cryosph.* 12, 2869–2882. doi:10.5194/tc-12-2869-2018.
- Greene, C. H., and Pershing, A. J. (2007). Oceans. Climate drives sea change. *Science* 315, 1084–5. doi:10.1126/science.1136495.
- 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, 9130–9137.
- Gregory, J. M., Bouttes, N., Griffies, S. M., Haak, H., Hurlin, W. J., Jungclaus, J., et al. (2016). The Flux-Anomaly-Forced Model Intercomparison Project (FAFMIP) contribution to CMIP6: investigation of sea-level and ocean climate change in response to CO₂ forcing. *Geosci. Model Dev.* 9, 3993–4017. doi:10.5194/gmd-9-3993-2016.
- Gregory, J. M., Griffies, S. M., Hughes, C. W., Lowe, J. A., Church, J. A., Fukimori, I., et al. (2019). Concepts and Terminology for Sea Level: Mean, Variability and Change, Both Local and Global. *Surv. Geophys.* 40, 1251–1289. doi:10.1007/s10712-019-09525-z.
- Gregory, J. M., Lowe, J. A., and Tett, S. F. B. (2006). Simulated Global-Mean Sea Level Changes over the Last Half-Millennium. *J. Clim.* 19, 4576–4591. doi:10.1175/JCLI3881.1.
- Gregory, J. M., Stott, P. A., Cresswell, D. J., Rayner, N. A., Gordon, C., and Sexton, D. M. H. (2002). Recent and future changes in Arctic sea ice simulated by the HadCM3 AOGCM. *Geophys. Res. Lett.* 29, 28-1-28-4. doi:10.1029/2001GL014575.
- Gregory, J. M., White, N. J., Church, J. A., Bierkens, M. F. P., Box, J. E., Van Den Broeke, M. R., et al. (2013).

- Twentieth-century global-mean sea level rise: Is the whole greater than the sum of the parts? *J. Clim.* 26, 4476–4499. doi:10.1175/JCLI-D-12-00319.1.
- Griffies, S. M., and Greatbatch, R. J. (2012). Physical processes that impact the evolution of global mean sea level in ocean climate models. *Ocean Model.* 51, 37–72. doi:10.1016/j.ocemod.2012.04.003.
- Griffies, S. M., Winton, M., Anderson, W. G., Benson, R., Delworth, T. L., Dufour, C. O., et al. (2015). Impacts on ocean heat from transient mesoscale eddies in a hierarchy of climate models. *J. Clim.* 28, 952–977.
- Griffies, S. M., Yin, J., Durack, P. J., Goddard, P., Bates, S. C., Behrens, E., et al. (2014). An assessment of global and regional sea level for years 1993–2007 in a suite of interannual CORE-II simulations. *Ocean Model.* 78, 35–89. doi:10.1016/J.OCEMOD.2014.03.004.
- Grinsted, A., Jevrejeva, S., Riva, R. E. M., and Dahl-Jensen, D. (2015). Sea level rise projections for Northern Europe under RCP8.5. *Clim. Res.* 64 (1), 2015. doi:10.3354/cr01309.
- Grinsted, A., Moore, J. C., and Jevrejeva, S. (2013). Projected Atlantic hurricane surge threat from rising temperatures. *Proc. Natl. Acad. Sci.* 110, 5369–5373. doi:10.1073/pnas.1209980110.
- Grodsky, S. A., Carton, J. A., and Bryan, F. O. (2014). A curious local surface salinity maximum in the northwestern tropical Atlantic. *J. Geophys. Res. Ocean.* 119, 484–495. doi:10.1002/2013JC009450.
- Groh, A., Horwath, M., Horvath, A., Meister, R., Sørensen, L. S., Barletta, V. R., et al. (2019). Evaluating GRACE Mass Change Time Series for the Antarctic and Greenland Ice Sheet—Methods and Results. *Geosciences* 9, 415. doi:10.3390/geosciences9100415.
- Gruber, N., Lachkar, Z., Frenzel, H., Marchesiello, P., Münnich, M., McWilliams, J. C., et al. (2011). Eddy-induced reduction of biological production in eastern boundary upwelling systems. *Nat. Geosci.* 4, 787. Available at: <https://doi.org/10.1038/ngeo1273>.
- Gruber, S. (2012). Derivation and analysis of a high-resolution estimate of global permafrost zonation. *Cryosph.* 6, 221–233. doi:10.5194/tc-6-221-2012.
- Gudmundsson, G., Krug, J., Durand, G., Favier, L., and Gagliardini, O. (2012). The stability of grounding lines on retrograde slopes. *Cryosphere* 6, 1497–1505.
- Guimberteau, M., Zhu, D., Maignan, F., Huang, Y., Yue, C., Dantec-N d lec, S., et al. (2018). ORCHIDEE-MICT (v8.4.1), a land surface model for the high latitudes: model description and validation. *Geosci. Model Dev.* 11, 121–163. doi:10.5194/gmd-11-121-2018.
- Gupta, M., and Marshall, J. (2018). The Climate Response to Multiple Volcanic Eruptions Mediated by Ocean Heat Uptake: Damping Processes and Accumulation Potential. *J. Clim.* 31, 8669–8687. doi:10.1175/JCLI-D-17-0703.1.
- Haarsma, R. J., Roberts, M. J., Vidale, P. L., Senior, C. A., Bellucci, A., Bao, Q., et al. (2016). High Resolution Model Intercomparison Project (HighResMIP v1.0) for CMIP6. *Geosci. Model Dev.* 9, 4185–4208. doi:10.5194/gmd-9-4185-2016.
- Haas, C., Beckers, J., King, J., Silis, A., Stroeve, J., Wilkinson, J., et al. (2017). Ice and Snow Thickness Variability and Change in the High Arctic Ocean Observed by In Situ Measurements. *Geophys. Res. Lett.* doi:10.1002/2017GL075434.
- Hagemann, S., Blome, T., Ekici, A., and Beer, C. (2016). Soil-frost-enabled soil-moisture-precipitation feedback over northern high latitudes. *Earth Syst. Dyn.* 7, 611–625. doi:10.5194/esd-7-611-2016.
- Haid, V., Iovino, D., and Masina, S. (2017). Impacts of freshwater changes on Antarctic sea ice in an eddy-permitting sea-ice-ocean model. *Cryosph.* 11, 1387–1402. doi:10.5194/tc-11-1387-2017.
- Haigh, I. D., Pickering, M. D., Green, J. A. M., Arbic, B. K., Arns, A., Dangendorf, S., et al. (2019). The Tides They Are a-Changin’: A comprehensive review of past and future non-astronomical changes in tides, their driving mechanisms and future implications. *Rev. Geophys.*, 2018RG000636. doi:10.1029/2018RG000636.
- Haine, T. W. N., Curry, B., Gerdes, R., Hansen, E., Karcher, M., Lee, C., et al. (2015). Arctic freshwater export: Status, mechanisms, and prospects. *Glob. Planet. Change.* doi:10.1016/j.gloplacha.2014.11.013.
- Häkkinen, S., Rhines, P. B., and Worthen, D. L. (2016). Warming of the Global Ocean: Spatial Structure and Water-Mass Trends. *J. Clim.* 29, 4949–4963. doi:10.1175/JCLI-D-15-0607.1.
- Hallberg, R. (2013). Using a resolution function to regulate parameterizations of oceanic mesoscale eddy effects. *Ocean Model.* 72, 92–103. doi:10.1016/J.OCEMOD.2013.08.007.
- Ham, Y.-G., Kug, J.-S., Park, J.-Y., and Jin, F.-F. (2013). Sea surface temperature in the north tropical Atlantic as a trigger for El Niño/Southern Oscillation events. *Nat. Geosci.* 6, 112–116. doi:10.1038/ngeo1686.
- Hamlington, B. D., Strassburg, M. W., Leben, R. R., Han, W., Kim, K.-Y., and Nerem, R. S. (2014). Uncovering an anthropogenic sea-level rise signal in the Pacific Ocean. *Nat. Clim Chang.* 4, 782–785. doi:10.1038/nclimate2307.
- Han, W., Meehl, G. A., Stammer, D., Hu, A., Hamlington, B., Kenigson, J., et al. (2017). Spatial Patterns of Sea Level Variability Associated with Natural Internal Climate Modes. *Surv. Geophys.* 38, 217–250. doi:10.1007/s10712-016-9386-y.
- Hanna, E., Fettweis, X., and Hall, R. J. (2018). Brief communication: Recent changes in summer Greenland blocking captured by none of the CMIP5 models. *Cryosph.* 12, 3287–3292. doi:10.5194/tc-12-3287-2018.
- Hanna, E., Fettweis, X., Mernild, S. H., Cappelen, J., Ribergaard, M. H., Shuman, C. A., et al. (2014). Atmospheric and

- oceanic climate forcing of the exceptional Greenland ice sheet surface melt in summer 2012. *Int. J. Climatol.* 34, 1022–1037. doi:10.1002/joc.3743.
- Hanna, E., Huybrechts, P., Cappelen, J., Steffen, K., Bales, R. C., Burgess, E., et al. (2011). Greenland Ice Sheet surface mass balance 1870 to 2010 based on Twentieth Century Reanalysis, and links with global climate forcing. *J. Geophys. Res. Atmos.* 116. doi:10.1029/2011JD016387.
- Hanna, E., Mernild, S. H., Cappelen, J., and Steffen, K. (2012). Recent warming in Greenland in a long-term instrumental (1881–2012) climatic context: I. Evaluation of surface air temperature records. *Environ. Res. Lett.* 7, 45404. doi:10.1088/1748-9326/7/4/045404.
- Hansen, B., Húsgarð Larsen, K. M., Hátún, H., and Østerhus, S. (2016). A stable Faroe Bank Channel overflow 1995–2015. *Ocean Sci.* 12, 1205–1220. doi:10.5194/os-12-1205-2016.
- Hansen, E., Gerland, S., Granskog, M. A., Pavlova, O., Renner, A. H. H., Haapala, J., et al. (2013). Thinning of Arctic sea ice observed in Fram Strait: 1990–2011. *J. Geophys. Res. Ocean.* doi:10.1002/jgrc.20393.
- Harig, C., and Simons, F. J. (2015). Accelerated West Antarctic ice mass loss continues to outpace East Antarctic gains. *Earth Planet. Sci. Lett.* 415, 134–141.
- Harpold, A. A., and Brooks, P. D. (2018). Humidity determines snowpack ablation under a warming climate. *Proc. Natl. Acad. Sci. U. S. A.* 115, 1215–1220. doi:10.1073/pnas.1716789115.
- Harrison, W. D. (2013). How do glaciers respond to climate? Perspectives from the simplest models. *J. Glaciol.* 59, 949–960. doi:DOI: 10.3189/2013JoG13J048.
- Harrison, W. D., Elsberg, D. H., Echelmeyer, K. A., and Krimmel, R. M. (2001). On the characterization of glacier response by a single time-scale. *J. Glaciol.* 47, 659–664. doi:DOI: 10.3189/172756501781831837.
- Hartmann, D. L., and Michelsen, M. L. (1993). Large-scale effects on regulation of tropical sea surface temperature. *J. Clim.* 6, 2049–2062. doi:10.1175/1520-0442(1993)006<2049:LSEOTR>2.0.CO;2.
- Hartmann, D. L., Tank, A. M. G. K., Rusticucci, M., Alexander, L. V., Brönnimann, S., Charabi, Y., et al. (2013). “Observations: Atmosphere and Surface,” in *Climate Change 2013: the Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, eds. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, A. J. Boschung, et al. (Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.).
- Haskins, R. K., Oliver, K. I. C., Jackson, L. C., Drijfhout, S. S., and Wood, R. A. (2019). Explaining asymmetry between weakening and recovery of the AMOC in a coupled climate model. *Clim. Dyn.* 53, 67–79. doi:10.1007/s00382-018-4570-z.
- Hasselmann, K. (1976). Stochastic climate models Part I. Theory. *Tellus* 28, 473–485. doi:10.1111/j.2153-3490.1976.tb00696.x.
- Haumann, F. A., Gruber, N., and Münnich, M. (submitted). Sea-ice induced Southern Ocean subsurface warming and surface cooling in a warming climate. *AGU Adv.* (submitted).
- Haumann, F. A., Gruber, N., Münnich, M., Frenger, I., and Kern, S. (2016). Sea-ice transport driving Southern Ocean salinity and its recent trends. *Nature* 537, 89–92. doi:10.1038/nature19101.
- Haumann, F. A., Notz, D., and Schmidt, H. (2014). Anthropogenic influence on recent circulation-driven Antarctic sea ice changes. *Geophys. Res. Lett.* 41, 8429–8437. doi:10.1002/2014GL061659.
- Hausmann, U., Sallee, J.-B., Jourdain, N. C., Mathiot, P., Rousset, C., Madec, G., et al. (submitted). The role of tides in ocean–ice–shelf interactions in the southwestern Weddell Sea. *JGR* (submitted).
- Hay, C. C., Morrow, E., Kopp, R. E., and Mitrovica, J. X. (2015). Probabilistic reanalysis of twentieth-century sea-level rise. *Nature* 517. doi:10.1038/nature14093.
- He, C., Liu, Z., and Hu, A. (2019). The transient response of atmospheric and oceanic heat transports to anthropogenic warming. *Nat. Clim. Chang.* 9, 222–226. doi:10.1038/s41558-018-0387-3.
- He, C., Wu, B., Zou, L., and Zhou, T. (2017). Responses of the summertime subtropical anticyclones to global warming. *J. Clim.* 30, 6465–6479.
- Heilig, A., Eisen, O., MacFerrin, M., Tedesco, M., and Fettweis, X. (2018). Seasonal monitoring of melt and accumulation within the deep percolation zone of the Greenland Ice Sheet and comparison with simulations of regional climate modeling. *Cryosph.* 12, 1851–1866. doi:10.5194/tc-12-1851-2018.
- Hellmer, H. H., Kauker, F., Timmermann, R., Determann, J., and Rae, J. (2012). Twenty-first-century warming of a large Antarctic ice-shelf cavity by a redirected coastal current. *Nature* 485, 225. doi:10.1038/nature11064.
- Hellmer, H. H., Kauker, F., Timmermann, R., and Hattermann, T. (2017). The fate of the Southern Weddell sea continental shelf in a warming climate. *J. Clim.* 30, 4337–4350. doi:10.1175/JCLI-D-16-0420.1.
- Helm, K. P., Bindoff, N. L., and Church, J. A. (2011). Observed decreases in oxygen content of the global ocean. *Geophys. Res. Lett.* doi:10.1029/2011GL049513.
- Helsen, M. M., Van De Berg, W. J., van de Wal, R. S. W., van den Broeke, M. R., and Oerlemans, J. (2013). Coupled regional climate-ice-sheet simulation shows limited Greenland ice loss during the Eemian. *Clim. Past* 9, 1773–1788. doi:10.5194/cp-9-1773-2013.
- Helsen, M. M., Van De Wal, R. S. W., Reerink, T. J., Bintanja, R., Madsen, M. S., Yang, S., et al. (2017). On the importance of the albedo parameterization for the mass balance of the Greenland ice sheet in EC-Earth. *Cryosphere* 11, 1949–1965. doi:10.5194/tc-11-1949-2017.

- Hermes, J. C., Masumoto, Y., Beal, L. M., Roxy, M. K., Vialard, J., Andres, M., et al. (2019). A Sustained Ocean Observing System in the Indian Ocean for Climate Related Scientific Knowledge and Societal Needs. *Front. Mar. Sci.* 6, 355. doi:10.3389/fmars.2019.00355.
- Hernández-Henríquez, M. A., Déry, S. J., and Derksen, C. (2015). Polar amplification and elevation-dependence in trends of Northern Hemisphere snow cover extent, 1971–2014. *Environ. Res. Lett.* 10, 044010. doi:10.1088/1748-9326/10/4/044010.
- Herring, S. C., Christidis, N., Hoell, A., Kossin, J. P., Schreck, C. J., and Stott, P. A. (2018). Explaining Extreme Events of 2016 from a Climate Perspective. *Bull. Am. Meteorol. Soc.* 99, S1–S157. doi:10.1175/BAMS-ExplainingExtremeEvents2016.1.
- Herrington, T., and Zickfeld, K. (2014). Path independence of climate and carbon cycle response over a broad range of cumulative carbon emissions. *Earth Syst. Dyn.* 5, 409–422. doi:10.5194/esd-5-409-2014.
- Heuzé, C. (2017). North Atlantic deep water formation and AMOC in CMIP5 models. *Ocean Sci.* 13, 609–622. doi:10.5194/os-13-609-2017.
- Heuzé, C., Heywood, K. J., Stevens, D. P., and Ridley, J. K. (2013). Southern Ocean bottom water characteristics in CMIP5 models. *Geophys. Res. Lett.* 40, 1409–1414. doi:10.1002/grl.50287.
- Heuzé, C., Heywood, K. J., Stevens, D. P., and Ridley, J. K. (2015). Changes in Global Ocean Bottom Properties and Volume Transports in CMIP5 Models under Climate Change Scenarios. *J. Clim.* 28, 2917–2944. doi:10.1175/JCLI-D-14-00381.1.
- Hewitt, H. T., Bell, M. J., Chassignet, E. P., Czaja, A., Ferreira, D., Griffies, S. M., et al. (2017). Will high-resolution global ocean models benefit coupled predictions on short-range to climate timescales? *Ocean Model.* 120. doi:10.1016/j.ocemod.2017.11.002.
- Hickey, B. M., and Pola, N. E. (1983). The seasonal alongshore pressure gradient on the west coast of the United States. *J. Geophys. Res. Ocean.* 88, 7623–7633.
- Hill, D. J., Dolan, A. M., Haywood, A. M., Hunter, S. J., and Stoll, D. K. (2010). Sensitivity of the Greenland Ice Sheet to Pliocene sea surface temperatures. *Stratigraphy* 7, 111–121.
- Hill, E. A., Hilmar Gudmundsson, G., Rachel Carr, J., and Stokes, C. R. (2018). Velocity response of Petermann Glacier, northwest Greenland, to past and future calving events. *Cryosphere* 12, 3907–3921. doi:10.5194/tc-12-3907-2018.
- Hillenbrand, C. D., Smith, J. A., Hodell, D. A., Greaves, M., Poole, C. R., Kender, S., et al. (2017). West Antarctic Ice Sheet retreat driven by Holocene warm water incursions. *Nature* 547, 43–48. doi:10.1038/nature22995.
- Hinkel, J., Church, J. A., Gregory, J. M., Lambert, E., Le Cozannet, G., Lowe, J., et al. (2019). Meeting User Needs for Sea Level Rise Information: A Decision Analysis Perspective. *Earth's Futur.* 7, 320–337. doi:10.1029/2018EF001071.
- Hirabayashi, Y., Zang, Y., Watanabe, S., Koirala, S., and Kanae, S. (2013). Projection of glacier mass changes under a high-emission climate scenario using the global glacier model HYOGA2. *Hydrol. Res. Lett.* 7, 6–11. doi:10.3178/hrl.7.6.
- Hobbs, W., Curran, M., Abram, N., and Thomas, E. R. (2016). Century-scale perspectives on observed and simulated Southern Ocean sea ice trends from proxy reconstructions. *J. Geophys. Res.* 121, 7804–7818. doi:10.1002/2016JC012111.
- Hobday, A. J., Alexander, L. V., Perkins, S. E., Smale, D. A., Straub, S. C., Oliver, E. C. J., et al. (2016). A hierarchical approach to defining marine heatwaves. *Prog. Oceanogr.* 141, 227–238. doi:10.1016/j.pocean.2015.12.014.
- Hobday, A. J., Oliver, E. C., Sen Gupta, A., Benthuyssen, J. A., Burrows, M. T., Donat, M. G., et al. (2018). Categorizing and Naming Marine Heatwaves. *Oceanography* 31. Available at: <https://doi.org/10.5670/oceanog.2018.205>.
- Hock, R., Bliss, A., Marzeion, B., Giesen, R., Hirabayashi, Y., Huss, M., et al. (2019). GlacierMIP - A model intercomparison of global-scale glacier mass-balance models and projections. *J. Glaciol.* 65, 453–467.
- Hofer, S., Tedstone, A. J., Fettweis, X., and Bamber, J. L. (2017). Decreasing cloud cover drives the recent mass loss on the Greenland Ice Sheet. *Sci. Adv.* 3. doi:10.1126/sciadv.1700584.
- Hofer, S., Tedstone, A. J., Fettweis, X., and Bamber, J. L. (2019). Cloud microphysics and circulation anomalies control differences in future Greenland melt. *Nat. Clim. Chang.* 9, 523–528. doi:10.1038/s41558-019-0507-8.
- Hoffmann, S. S., McManus, J. F., and Swank, E. (2018). Evidence for Stable Holocene Basin-Scale Overturning Circulation Despite Variable Currents Along the Deep Western Boundary of the North Atlantic Ocean. *Geophys. Res. Lett.* 45, 13,427–13,436. doi:10.1029/2018GL080187.
- Holbrook, N. J., Scannell, H. A., Sen Gupta, A., Benthuyssen, J. A., Feng, M., Oliver, E. C. J., et al. (2019). A global assessment of marine heatwaves and their drivers. *Nat. Commun.* 10, 2624. doi:10.1038/s41467-019-10206-z.
- Holland, D. M., Thomas, R. H., De Young, B., Ribergaard, M. H., and Lyberth, B. (2008). Acceleration of Jakobshavn Isbr triggered by warm subsurface ocean waters. *Nat. Geosci.* doi:10.1038/ngeo316.
- Holland, M. M., Landrum, L., Kostov, Y., and Marshall, J. (2017a). Sensitivity of Antarctic sea ice to the Southern Annular Mode in coupled climate models. *Clim. Dyn.* 49, 1813–1831. doi:10.1007/s00382-016-3424-9.
- Holland, M. M., Landrum, L., Raphael, M. N., and Kwok, R. (2018). The regional, seasonal, and lagged influence of the Amundsen Sea Low on Antarctic Sea Ice. *Geophys. Res. Lett.* doi:10.1029/2018GL080140.

- 1 Holland, M. M., Landrum, L., Raphael, M., and Stammerjohn, S. (2017b). Springtime winds drive Ross Sea ice
2 variability and change in the following autumn. *Nat. Commun.* 8. doi:10.1038/s41467-017-00820-0.
- 3 Holland, P. R. (2014). The seasonality of Antarctic sea ice trends. *Geophys. Res. Lett.* 41, 4230–4237.
4 doi:10.1002/2014GL060172.
- 5 Holland, P. R., Bracegirdle, T. J., Dutrieux, P., Jenkins, A., and Steig, E. J. (2019). West Antarctic ice loss influenced
6 by internal climate variability and anthropogenic forcing. *Nat. Geosci.* 12, 718–724. doi:10.1038/s41561-019-
7 0420-9.
- 8 Holland, P. R., Bruneau, N., Enright, C., Losch, M., Kurtz, N. T., and Kwok, R. (2014). Modeled trends in Antarctic
9 sea ice thickness. *J. Clim.* 27, 3784–3801. doi:10.1175/JCLI-D-13-00301.1.
- 10 Holland, P. R., and Kimura, N. (2016). Observed concentration budgets of Arctic and Antarctic sea ice. *J. Clim.* 29,
11 5241–5249. doi:10.1175/JCLI-D-16-0121.1.
- 12 Holland, P. R., and Kwok, R. (2012). Wind-driven trends in Antarctic sea-ice drift. *Nat. Geosci.* 5, 872–875.
13 doi:10.1038/ngeo1627.
- 14 Holliday, N. P., Cunningham, S. A., Johnson, C., Gary, S. F., Griffiths, C., Read, J. F., et al. (2015). Multidecadal
15 variability of potential temperature, salinity, and transport in the eastern subpolar North Atlantic. *J. Geophys. Res.*
16 *Ocean.* 120, 5945–5967. doi:10.1002/2015JC010762.
- 17 Holliday, N. P., Hughes, S. L., Bacon, S., Beszczynska-Möller, A., Hansen, B., Lavín, A., et al. (2008). Reversal of the
18 1960s to 1990s freshening trend in the northeast North Atlantic and Nordic Seas. *Geophys. Res. Lett.* 35.
19 doi:10.1029/2007GL032675.
- 20 Holmes, C. R., Holland, P. R., and Bracegirdle, T. J. (2019). Compensating biases and a noteworthy success in the
21 CMIP5 representation of Antarctic sea ice processes. *Geophys. Res. Lett.* 0. doi:10.1029/2018GL081796.
- 22 Holt, J., Hyder, P., Ashworth, M., Harle, J., Hewitt, H. T., Liu, H., et al. (2017). Prospects for improving the
23 representation of coastal and shelf seas in global ocean models. *Geosci. Model Dev.* 10, 499–523.
24 doi:10.5194/gmd-10-499-2017.
- 25 Holt, J., Polton, J., Huthnance, J., Wakelin, S., O’Dea, E., Harle, J., et al. (2018). Climate-Driven Change in the North
26 Atlantic and Arctic Oceans Can Greatly Reduce the Circulation of the North Sea. *Geophys. Res. Lett.* 45, 11,827-
27 11,836. doi:10.1029/2018GL078878.
- 28 Holt, J., Schrum, C., Cannaby, H., Daewel, U., Allen, I., Artioli, Y., et al. (2016). Potential impacts of climate change
29 on the primary production of regional seas: A comparative analysis of five European seas. *Prog. Oceanogr.* 140,
30 91–115. doi:10.1016/j.pocean.2015.11.004.
- 31 Holt, T. O., Glasser, N. F., Quincey, D. J., and Siegfried, M. R. (2013). Speedup and fracturing of George VI Ice Shelf,
32 Antarctic Peninsula. *Cryosph.* 7, 797–816. doi:10.5194/tc-7-797-2013.
- 33 Hori, M., Sugiura, K., Kobayashi, K., Aoki, T., Tanikawa, T., Kuchiki, K., et al. (2017). A 38-year (1978–2015)
34 Northern Hemisphere daily snow cover extent product derived using consistent objective criteria from satellite-
35 borne optical sensors. *Remote Sens. Environ.* 191, 402–418. doi:10.1016/j.rse.2017.01.023.
- 36 Hörner, T., Stein, R., and Fahl, K. (2017). Evidence for Holocene centennial variability in sea ice cover based on
37 IP25biomarker reconstruction in the southern Kara Sea (Arctic Ocean). *Geo-Marine Lett.* 37, 515–526.
38 doi:10.1007/s00367-017-0501-y.
- 39 Hörner, T., Stein, R., and Fahl, K. (2018). Paleo-sea ice distribution and polynya variability on the Kara Sea shelf
40 during the last 12 ka. *Arktos* 4, 6. doi:10.1007/s41063-018-0040-4.
- 41 Hörner, T., Stein, R., Fahl, K., and Birgel, D. (2016). Post-glacial variability of sea ice cover, river run-off and
42 biological production in the western Laptev Sea (Arctic Ocean) - A high-resolution biomarker study. *Quat. Sci.*
43 *Rev.* 143, 133–149. doi:10.1016/j.quascirev.2016.04.011.
- 44 Horton, B. P., Kopp, R. E., Garner, A. J., Hay, C. C., Khan, N. S., Roy, K., et al. (2018). Mapping Sea-Level Change in
45 Time, Space, and Probability. *Annu. Rev. Environ. Resour.* 43, 481–521. doi:10.1146/annurev-environ-102017-
46 025826.
- 47 Horvat, C., and Tziperman, E. (2018). Understanding Melting due to Ocean Eddy Heat Fluxes at the Edge of Sea-Ice
48 Floes. *Geophys. Res. Lett.* 45, 9721–9730. doi:10.1029/2018GL079363.
- 49 Hosking, J. S., Orr, A., Marshall, G. J., Turner, J., and Phillips, T. (2013). The influence of the Amundsen-
50 Bellingshausen Seas low on the climate of West Antarctica and its representation in coupled climate model
51 simulations. *J. Clim.* 26, 6633–6648. doi:10.1175/JCLI-D-12-00813.1.
- 52 Howard, T., Palmer, M. D., and Bricheno, L. M. (2019). Contributions to 21st century projections of extreme sea-level
53 change around the {UK}. *Environ. Res. Commun.* 1, 95002. doi:10.1088/2515-7620/ab42d7.
- 54 Howat, I. M., Negrete, A., and Smith, B. E. (2014). The Greenland Ice Mapping Project (GIMP) land classification and
55 surface elevation data sets. *Cryosph.* 8, 1509–1518. doi:10.5194/tc-8-1509-2014.
- 56 Howell, F. W., Haywood, A. M., Dolan, A. M., Dowsett, H. J., Francis, J. E., Hill, D. J., et al. (2014). Can uncertainties
57 in sea ice albedo reconcile patterns of data-model discord for the Pliocene and 20th/21st centuries? *Geophys. Res.*
58 *Lett.* 41, 2011–2018. doi:10.1002/2013GL058872.
- 59 Hrbáček, F., Vieira, G., Oliva, M., Balks, M., Guglielmin, M., de Pablo, M. Á., et al. (2018). Active layer monitoring in
60 Antarctica: an overview of results from 2006 to 2015. *Polar Geogr.* 0, 1–15.
61 doi:10.1080/1088937X.2017.1420105.

- Hu, D., Wu, L., Cai, W., Gupta, A. Sen, Ganachaud, A., Qiu, B., et al. (2015a). Pacific western boundary currents and their roles in climate. *Nature* 522, 299–308. doi:10.1038/nature14504.
- Hu, S., and Fedorov, A. V (2019). Indian Ocean warming can strengthen the Atlantic meridional overturning circulation. *Nat. Clim. Chang.* 9, 747–751. doi:10.1038/s41558-019-0566-x.
- Hu, Y., Huang, X., Baker, A. H., Tseng, Y.-H., Bryan, F. O., Dennis, J. M., et al. (2015b). Improving the Scalability of the Ocean Barotropic Solver in the Community Earth System Model. *f Int. Conf. High Perform. Comput. Networking, Storage Anal.* doi:10.1145/2807591.2807596.
- Hua, W., Dai, A., and Qin, M. (2018). Contributions of Internal Variability and External Forcing to the Recent Pacific Decadal Variations. *Geophys. Res. Lett.* doi:10.1029/2018GL079033.
- Huang, C. J., and Qiao, F. (2010). Wave-turbulence interaction and its induced mixing in the upper ocean. *J. Geophys. Res.* 115, C04026. doi:10.1029/2009JC005853.
- Hughes, C. W., and De Cuevas, B. A. (2001). Why Western Boundary Currents in Realistic Oceans are Inviscid: A Link between Form Stress and Bottom Pressure Torques. Available at: <http://pordlabs.ucsd.edu/wryoung/theorySeminar/pdf16/HughesDeCuevas2001a.pdf> [Accessed February 6, 2019].
- Huisman, S. E., den Toom, M., Dijkstra, H. A., and Drijfhout, S. (2010). An Indicator of the Multiple Equilibria Regime of the Atlantic Meridional Overturning Circulation. *J. Phys. Oceanogr.* doi:10.1175/2009JPO4215.1.
- Hunter, S. J., Haywood, A. M., Dolan, A. M., and Tindall, J. C. (2019). The HadCM3 contribution to PlioMIP phase 2. *Clim. Past* 15, 1691–1713. doi:10.5194/cp-15-1691-2019.
- Hüsler, F., Jonas, T., Riffler, M., Musial, J. P., and Wunderle, S. (2014). A satellite-based snow cover climatology (1985–2011) for the European Alps derived from AVHRR data. *Cryosph.* 8, 73–90. doi:10.5194/tc-8-73-2014.
- Huss, M., and Hock, R. (2015). A new model for global glacier change and sea-level rise. *Front. Earth Sci.* 3. doi:10.3389/feart.2015.00054.
- Huss, M., and Hock, R. (2018). Global-scale hydrological response to future glacier mass loss. *Nat. Clim. Chang.* 8, 135–140. doi:10.1038/s41558-017-0049-x.
- Hutchings, J., Layton, C., Winsor, P., Ruddick, B., Perovich, D., Kelley, D., et al. (2015). Toward Quantifying the Increasing Role of Oceanic Heat in Sea Ice Loss in the New Arctic. *Bull. Am. Meteorol. Soc.* 96, 2079–2105. doi:10.1175/bams-d-13-00177.1.
- Hutchinson, D. K., Coxall, H. K., O'Regan, M., Nilsson, J., Caballero, R., and de Boer, A. M. (2019). Arctic closure as a trigger for Atlantic overturning at the Eocene-Oligocene Transition. *Nat. Commun.* 10, 3797. doi:10.1038/s41467-019-11828-z.
- Huthnance, J. M., Holt, J. T., and Wakelin, S. L. (2009). Deep ocean exchange with west-European shelf seas. *Ocean Sci.* 5, 621–634. doi:10.5194/os-5-621-2009.
- Huybrechts, P., Goelzer, H., Janssens, I., Driesschaert, E., Fichet, T., Goosse, H., et al. (2011). Response of the Greenland and Antarctic Ice Sheets to Multi-Millennial Greenhouse Warming in the Earth System Model of Intermediate Complexity LOVECLIM. *Surv. Geophys.* 32, 397–416. doi:10.1007/s10712-011-9131-5.
- Hyder, P., Edwards, J. M., Allan, R. P., Hewitt, H. T., Bracegirdle, T. J., Gregory, J. M., et al. (2018). Critical Southern Ocean climate model biases traced to atmospheric model cloud errors. *Nat. Commun.* 9, 3625. doi:10.1038/s41467-018-05634-2.
- Idier, D., Bertin, X., Thompson, P. R., and Pickering, M. D. (submitted). Surveys in Geophysics Interactions Between mean sea level, Tide, Surge, Waves and Flooding: Mechanisms and Contributions to Sea Level Variations at the Coast. *Surv. Geophys.* (submitted).
- Idier, D., Paris, F., Cozannet, G. Le, Boulahya, F., and Dumas, F. (2017). Sea-level rise impacts on the tides of the European Shelf. *Cont. Shelf Res.* doi:10.1016/j.csr.2017.01.007.
- IPCC (2018). “Annex I: Glossary [Matthews, J.B.R. (ed.)],” 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*, eds. V. Masson-Delmotte, P. Zhai, H.-O. Pörtner, D. Roberts, J. Skea, P. R. Shukla, et al. (In Press).
- IPCC (2019a). “Annex I: Glossary [Weyer N.M. (ed.)],” in *IPCC, 2019: Annex I: Glossary. 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. Alegría, M. Nicolai, A. Okem, J. Petzold, B., ed. N. M. Weyer (In Press).*
- IPCC (2019b). “Summary for Policymakers,” in *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. Alegría, M. Nicolai, A. Okem, J. Petz.*
- Ivanovic, R. F., Gregoire, L. J., Wickert, A. D., Valdes, P. J., and Burke, A. (2017). Collapse of the North American ice saddle 14,500 years ago caused widespread cooling and reduced ocean overturning circulation. *Geophys. Res. Lett.* 44, 383–392. doi:10.1002/2016GL071849.
- Ivins, E. R., JAMES, T. S., Wahr, J., O Schrama, E. J., Landerer, F. W., and Simon, K. M. (2013). Antarctic contribution to sea level rise observed by GRACE with improved GIA correction. *J. Geophys. Res. Solid Earth* 118, 3126–3141.
- Izumo, T., Vialard, J., Lengaigne, M., de Boyer Montegut, C., Behera, S. K., Luo, J.-J., et al. (2010). Influence of the

- state of the Indian Ocean Dipole on the following year's El Niño. *Nat. Geosci.* 3, 168–172. doi:10.1038/ngeo760.
- Jackson, L. C. (2013). Shutdown and recovery of the AMOC in a coupled global climate model: The role of the advective feedback. *Geophys. Res. Lett.* 40, 1182–1188. doi:10.1002/grl.50289.
- Jackson, L. C., Peterson, K. A., Roberts, C. D., and Wood, R. A. (2016). Recent slowing of Atlantic overturning circulation as a recovery from earlier strengthening. *Nat. Geosci.* 9, 518. Available at: <https://doi.org/10.1038/ngeo2715>.
- Jackson, L. C., Roberts, M. J., Hewitt, H. T., Iovino, D., Koenigk, T., Meccia, V. L., et al. (submitted). Does ocean resolution affect the rate of AMOC weakening? *Clim. Dyn.* (submitted).
- Jackson, L. C., and Wood, R. A. (2018). Timescales of AMOC decline in response to fresh water forcing. *Clim. Dyn.* doi:10.1007/s00382-017-3957-6.
- Jackson, L. P., Grinsted, A., and Jevrejeva, S. (2018). 21st century sea-level rise in line with the Paris accord. *Earth's Futur.*, 2017EF000688. doi:10.1002/2017EF000688.
- Jackson, L. P., and Jevrejeva, S. (2016). A probabilistic approach to 21st century regional sea-level projections using RCP and High-end scenarios. *Glob. Planet. Change* 146, 179–189. doi:10.1016/j.gloplacha.2016.10.006.
- Jacobs, S. S., Jenkins, A., Giulivi, C. F., and Dutrieux, P. (2011). Stronger ocean circulation and increased melting under Pine Island Glacier ice shelf. *Nat. Geosci.* 4, 519–523. doi:10.1038/ngeo1188.
- Jacox, M. G., Edwards, C. A., Hazen, E. L., and Bograd, S. J. (2018). Coastal Upwelling Revisited: Ekman, Bakun, and Improved Upwelling Indices for the US West Coast. *J. Geophys. Res. Ocean.* 123, 7332–7350.
- Jahn, A. (2018). Reduced probability of ice-free summers for 1.5°C compared to 2°C warming. *Nat. Clim. Chang.* 8, 409–413. doi:10.1038/s41558-018-0127-8.
- Jahn, A., Kay, J. E., Holland, M. M., and Hall, D. M. (2016). How predictable is the timing of a summer ice-free Arctic? *Geophys. Res. Lett.* 43, 9113–9120. doi:10.1002/2016GL070067.
- James, M., Lewkowicz, A. G., Smith, S. L., and Miceli, C. M. (2013). Multi-decadal degradation and persistence of permafrost in the Alaska Highway corridor, northwest Canada. *Environ. Res. Lett.* 8, 045013. doi:10.1088/1748-9326/8/4/045013.
- Janeković, I., and Powell, B. (2012). Analysis of imposing tidal dynamics to nested numerical models. *Cont. Shelf Res.* 34, 30–40. doi:10.1016/j.csr.2011.11.017.
- Janowitz, G. S., and Pietrafesa, L. J. (1982). The effects of alongshore variation in bottom topography on a boundary current—(topographically induced upwelling). *Cont. Shelf Res.* 1, 123–141.
- Jansen, M. F. (2017). Glacial ocean circulation and stratification explained by reduced atmospheric temperature. *Proc. Natl. Acad. Sci. U. S. A.* 114, 45–50. doi:10.1073/pnas.1610438113.
- Jansen, M. F., Nadeau, L.-P., and Merlis, T. M. (2018). Transient versus Equilibrium Response of the Ocean's Overturning Circulation to Warming. *J. Clim.* 31, 5147–5163. doi:10.1175/JCLI-D-17-0797.1.
- Jay, D. A. (2009). Evolution of tidal amplitudes in the eastern Pacific Ocean. *Geophys. Res. Lett.* doi:10.1029/2008GL036185.
- Jayne, S. R., Hogg, N. G., Waterman, S. N., Rainville, L., Donohue, K. A., Randolph Watts, D., et al. (2009). The Kuroshio Extension and its recirculation gyres. *Deep Sea Res. Part I Oceanogr. Res. Pap.* 56, 2088–2099. doi:10.1016/J.DSR.2009.08.006.
- Jenkins, A., Shoosmith, D., Dutrieux, P., Jacobs, S., Kim, T. W., Lee, S. H., et al. (2018). West Antarctic Ice Sheet retreat in the Amundsen Sea driven by decadal oceanic variability. *Nat. Geosci.* 11, 733–738. doi:10.1038/s41561-018-0207-4.
- Jevrejeva, S., Grinsted, A., and Moore, J. C. (2014). Upper limit for sea level projections by 2100. *Environ. Res. Lett.* 9, 104008. doi:10.1088/1748-9326/9/10/104008.
- Jochum, M., Briegleb, B. P., Danabasoglu, G., Large, W. G., Norton, N. J., Jayne, S. R., et al. (2013). The Impact of Oceanic Near-Inertial Waves on Climate. *J. Clim.* 26, 2833–2844. doi:10.1175/JCLI-D-12-00181.1.
- Jochumsen, K., Moritz, M., Nunes, N., Quadfasel, D., Larsen, K. M. H., Hansen, B., et al. (2017). Revised transport estimates of the Denmark Strait overflow. *J. Geophys. Res. Ocean.* 122, 3434–3450. doi:10.1002/2017JC012803.
- Johannessen, O. M. (2008). Decreasing Arctic Sea Ice Mirrors Increasing CO₂ on Decadal Time Scale. *Atmos. Ocean. Sci. Lett.* 1, 51–56. doi:10.1080/16742834.2008.11446766.
- Jóhannesson, T., Raymond, C., and Waddington, E. (1989). Time-Scale for Adjustment of Glaciers to Changes in Mass Balance. *J. Glaciol.* 35, 355–369. doi:DOI: 10.3189/S002214300000928X.
- Johnson, G. C., Lyman, J. M., Boyer, T., Cheng, L., Domingues, C. M., Gilson, J., et al. (2018). Ocean heat content [in State of the Climate in 2017]. *Bull. Am. Meteorol. Soc.* 99, S72–S77.
- Johnson, G. C., Lyman, J. M., and Loeb, N. G. (2016). Improving estimates of Earth's energy imbalance. *Nat. Clim. Chang.* 6, 639–640. doi:10.1038/nclimate3043.
- Jones, B. M., Baughman, C. A., Romanovsky, V. E., Parsekian, A. D., Babcock, E. L., Stephani, E., et al. (2016a). Presence of rapidly degrading permafrost plateaus in south-central Alaska. *Cryosph.* 10, 2673–2692. doi:10.5194/tc-10-2673-2016.
- Jones, B. M., Grosse, G., Arp, C. D., Miller, E., Liu, L., Hayes, D. J., et al. (2015a). Recent Arctic tundra fire initiates widespread thermokarst development. *Sci. Rep.* doi:10.1038/srep15865.
- Jones, D. C., Ito, T., and Lovenduski, N. S. (2011). The transient response of the Southern Ocean pycnocline to

- changing atmospheric winds. doi:10.1029/2011GL048145.
- Jones, J. M., Gille, S. T., Goosse, H., Abram, N. J., Canziani, P. O., Charman, D. J., et al. (2016b). Assessing recent trends in high-latitude Southern Hemisphere surface climate. *Nat. Clim. Chang.* 6, 917. doi:10.1038/nclimate3103.
- Jones, R. S., Mackintosh, A. N., Norton, K. P., Golledge, N. R., Fogwill, C. J., Kubik, P. W., et al. (2015b). Rapid Holocene thinning of an East Antarctic outlet glacier driven by marine ice sheet instability. *Nat. Commun.* 6. doi:10.1038/ncomms9910.
- Jordà, G. (2014). Detection time for global and regional sea level trends and accelerations. *J. Geophys. Res. Ocean.* 119, 7164–7174. doi:10.1002/2014JC010005.
- Jorgenson, M. T., and Grosse, G. (2016). Remote Sensing of Landscape Change in Permafrost Regions. in *Permafrost and Periglacial Processes* doi:10.1002/ppp.1914.
- Josey, S. A., Hirschi, J. J.-M., Sinha, B., Duche, A., Grist, J. P., and Marsh, R. (2018). The Recent Atlantic Cold Anomaly: Causes, Consequences, and Related Phenomena. *Ann. Rev. Mar. Sci.* 10, 475–501. doi:10.1146/annurev-marine-121916-063102.
- Joughin, I., Alley, R. B., and Holland, D. M. (2012). Ice-Sheet Response to Oceanic Forcing. *Science* 338, 1172 LP – 1176. doi:10.1126/science.1226481.
- 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. doi:10.1126/science.1249055.
- Jourdain, N. C., Asay-Davis, X., Hattermann, T., Straneo, F., Seroussi, H., Little, C. M., et al. (submitted). A protocol for calculating basal melt rates in the ISMIP6 Antarctic ice sheet projections. *Cryosph. Discuss.* (submitted). doi:10.5194/tc-2019-277.
- Jourdain, N. C., Mathiot, P., Merino, N., Durand, G., Le Sommer, J., Spence, P., et al. (2017). Ocean circulation and sea-ice thinning induced by melting ice shelves in the Amundsen Sea. *J. Geophys. Res.* 122, 2550–2573. doi:10.1002/2016JC012509.
- Jourdain, N. C., Molines, J.-M., Le Sommer, J., Mathiot, P., Chanut, J., de Lavergne, C., et al. (2019). Simulating or prescribing the influence of tides on the Amundsen Sea ice shelves. *Ocean Model.* 133, 44–55. doi:10.1016/j.ocemod.2018.11.001.
- Jouvet, G., Huss, M., Funk, M., and Blatter, H. (2011). Modelling the retreat of Grosser Aletschgletscher, Switzerland, in a changing climate. *J. Glaciol.* 57, 1033–1045. doi:DOI: 10.3189/002214311798843359.
- Jüling, A., Viebahn, J. P., Drijfhout, S. S., and Dijkstra, H. A. (2018). Energetics of the Southern Ocean Mode. *J. Geophys. Res. Ocean.* 123, 9283–9304. doi:10.1029/2018JC014191.
- Jungclauss, J. H., Haak, H., Latif, M., and Mikolajewicz, U. (2005). Arctic–North Atlantic Interactions and Multidecadal Variability of the Meridional Overturning Circulation. *J. Clim.* 18, 4013–4031. doi:10.1175/JCLI3462.1.
- Jungclauss, J. H., Lohmann, K., and Zanchettin, D. (2014). Enhanced 20th-century heat transfer to the Arctic simulated in the context of climate variations over the last millennium. *Clim. Past* 10, 2201–2213. doi:10.5194/cp-10-2201-2014.
- Kääb, A., Berthier, E., Nuth, C., Gardelle, J., and Arnaud, Y. (2012). Contrasting patterns of early twenty-first-century glacier mass change in the Himalayas. *Nature* 488, 495–498. doi:10.1038/nature11324.
- Kääb, A., Leinss, S., Gilbert, A., Bühler, Y., Gascoin, S., Evans, S. G., et al. (2018). Massive collapse of two glaciers in western Tibet in 2016 after surge-like instability. *Nat. Geosci.* 11, 114–120. doi:10.1038/s41561-017-0039-7.
- Kääb, A., Treichler, D., Nuth, C., and Berthier, E. (2015). Brief Communication: Contending estimates of 2003–2008 glacier mass balance over the Pamir–Karakoram–Himalaya. *Cryosph.* 9, 557–564. doi:10.5194/tc-9-557-2015.
- Kajtar, J. B., Santoso, A., England, M. H., and Cai, W. (2017). Tropical climate variability: interactions across the Pacific, Indian, and Atlantic Oceans. *Clim. Dyn.* doi:10.1007/s00382-016-3199-z.
- Kaplan, M. R., Schaefer, J. M., Strelin, J. A., Denton, G. H., Anderson, R. F., Vandergoes, M. J., et al. (2016). Patagonian and southern South Atlantic view of Holocene climate. *Quat. Sci. Rev.* 141, 112–125. doi:https://doi.org/10.1016/j.quascirev.2016.03.014.
- Kapnick, S., and Hall, A. (2012). Causes of recent changes in western North American snowpack. *Clim. Dyn.* 38, 1885–1899. doi:10.1007/s00382-011-1089-y.
- Kapsch, M. L., Graversen, R. G., Tjernström, M., and Bintanja, R. (2016). The effect of downwelling longwave and shortwave radiation on Arctic summer sea ice. *J. Clim.* 29, 1143–1159. doi:10.1175/JCLI-D-15-0238.1.
- Karimi, N., Eftekhari, M., Farajzadeh, M., Namdari, S., Moridnejad, A., and Karimi, D. (2015). Use of multitemporal satellite images to find some evidence for glacier changes in the Haft-Khan glacier, Iran. *Arab. J. Geosci.* 8, 5879–5896. doi:10.1007/s12517-014-1578-5.
- Karnauskas, K. B., Johnson, G. C., and Murtugudde, R. (2012). An equatorial ocean bottleneck in global climate models. *J. Clim.* doi:10.1175/JCLI-D-11-00059.1.
- Kaspi, Y., and Schneider, T. (2011). Winter cold of eastern continental boundaries induced by warm ocean waters. *Nature* 471, 621–624. doi:10.1038/nature09924.
- Kato, S., Rose, F. G., Rutan, D. A., Thorsen, T. J., Loeb, N. G., Doelling, D. R., et al. (2018). Surface Irradiances of Edition 4.0 Clouds and the Earth’s Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) Data

- Product. *J. Clim.* 31, 4501–4527. doi:10.1175/JCLI-D-17-0523.1.
- Katsman, C. A., Drijfhout, S. S., Dijkstra, H. A., and Spall, M. A. (2018). Sinking of Dense North Atlantic Waters in a Global Ocean Model: Location and Controls. *J. Geophys. Res. Ocean.* doi:10.1029/2017JC013329.
- Kaufman, D. S., Axford, Y. L., Henderson, A. C. G., McKay, N. P., Oswald, W. W., Saenger, C., et al. (2016). Holocene climate changes in eastern Beringia (NW North America) – A systematic review of multi-proxy evidence. *Quat. Sci. Rev.* 147, 312–339. doi:https://doi.org/10.1016/j.quascirev.2015.10.021.
- Kaufmann, V. (2016). 20 years of geodetic monitoring of Dösen rock glacier (Ankogel Group, Austria): A short review. *Joannea / Geol. und Paläontologie* 12, 37–44.
- Kay, J. E., Holland, M. M., and Jahn, A. (2011). Inter-annual to multi-decadal Arctic sea ice extent trends in a warming world. *Geophys. Res. Lett.* 38. doi:10.1029/2011GL048008.
- Kellerer-Pirklbauer, A., and Kaufmann, V. (2018). Deglaciation and its impact on permafrost and rock glacier evolution: New insight from two adjacent cirques in Austria. *Sci. Total Environ.* 621, 1397–1414. doi:10.1016/j.scitotenv.2017.10.087.
- Kemp, A. C., Wright, A. J., Edwards, R. J., Barnett, R. L., Brain, M. J., Kopp, R. E., et al. (2018). Relative sea-level change in Newfoundland, Canada, during the past ~3000 years. *Quat. Sci. Rev.* 201, 89–110.
- Kern, S., Laverigne, T., Notz, D., Pedersen, L. T., Tonboe, R. T., Saldo, R., et al. (2019). Satellite Passive Microwave Sea-Ice Concentration Data Set Intercomparison: Closed Ice and Ship-Based Observations. *Cryosph.* doi:10.5194/tc-2019-120.
- Kern, S., Ozsoy-Çiçek, B., and Worby, A. (2016). Antarctic sea-ice thickness retrieval from ICESat: Inter-comparison of different approaches. *Remote Sens.* 8, 538. doi:10.3390/rs8070538.
- Kern, S., and Spreen, G. (2015). Uncertainties in Antarctic sea-ice thickness retrieval from ICESat. *Ann. Glaciol.* 56, 107–119. doi:10.3189/2015AoG69A736.
- Khan, S. A., Aschwanden, A., Bjørk, A. A., Wahr, J., Kjeldsen, K. K., and Kjær, K. H. (2015). Greenland ice sheet mass balance: a review. *Reports Prog. Phys.* 78, 46801. doi:10.1088/0034-4885/78/4/046801.
- Khan, S. A., Kjær, K. H., Bevis, M., Bamber, J. L., Wahr, J., Kjeldsen, K. K., et al. (2014). Sustained mass loss of the northeast Greenland ice sheet triggered by regional warming. *Nat. Clim. Chang.* 4, 292–299. doi:10.1038/nclimate2161.
- Khan, S. A., Sasgen, I., Bevis, M., van Dam, T., Bamber, J. L., Wahr, J., et al. (2016). Geodetic measurements reveal similarities between post--Last Glacial Maximum and present-day mass loss from the Greenland ice sheet. *Sci. Adv.* 2, e1600931.
- Khazendar, A., Fenty, I. G., Carroll, D., Gardner, A., Lee, C. M., Fukumori, I., et al. (2019a). Interruption of two decades of Jakobshavn Isbrae acceleration and thinning as regional ocean cools. *Nat. Geosci.* 12, 277–283. doi:10.1038/s41561-019-0329-3.
- Khazendar, A., Fenty, I. G., Carroll, D., Gardner, A., Lee, C. M., Fukumori, I., et al. (2019b). Interruption of two decades of Jakobshavn Isbrae acceleration and thinning as regional ocean cools. *Nat. Geosci.* 12, 277–283. doi:10.1038/s41561-019-0329-3.
- Khazendar, A., Rignot, E., Schroeder, D. M., Seroussi, H., Schodlok, M. P., Scheuchl, B., et al. (2016). Rapid submarine ice melting in the grounding zones of ice shelves in West Antarctica. *Nat. Commun.* doi:10.1038/ncomms13243.
- Khazendar, A., Schodlok, M. P., Fenty, I., Ligtenberg, S. R. M. R. M., Rignot, E., and van den Broeke, M. R. R. (2013). Observed thinning of Totten Glacier is linked to coastal polynya variability. *Nat. Commun.* 4, 2857. doi:10.1038/ncomms3857.
- Kim, K., Kim, K.-R., Min, D.-H., Volkov, Y., Yoon, J.-H., and Takematsu, M. (2001). Warming and structural changes in the east (Japan) Sea: A clue to future changes in global oceans? *Geophys. Res. Lett.* 28, 3293–3296. doi:10.1029/2001GL013078.
- Kim, S. T., Cai, W., Jin, F.-F., and Yu, J.-Y. (2014). ENSO stability in coupled climate models and its association with mean state. *Clim. Dyn.* 42, 3313–3321. doi:10.1007/s00382-013-1833-6.
- Kim, W. M., Yeager, S., Chang, P., and Danabasoglu, G. (2018). Low-Frequency North Atlantic Climate Variability in the Community Earth System Model Large Ensemble. *J. Clim.* 31, 787–813. doi:10.1175/JCLI-D-17-0193.1.
- Kimura, S., Jenkins, A., Regan, H., Holland, P. R., Assmann, K. M., Whitt, D. B., et al. (2017). Oceanographic Controls on the Variability of Ice-Shelf Basal Melting and Circulation of Glacial Meltwater in the Amundsen Sea Embayment, Antarctica. *J. Geophys. Res. Ocean.* 122, 10131–10155. doi:10.1002/2017JC012926.
- King, J., Spreen, G., Gerland, S., Haas, C., Hendricks, S., Kaleschke, L., et al. (2017). Sea-ice thickness from field measurements in the northwestern Barents Sea. *J. Geophys. Res. Ocean.* 122, 1497–1512. doi:10.1002/2016JC012199.
- King, M. D., Howat, I. M., Jeong, S., Noh, M. J., Wouters, B., Noël, B., et al. (2018). Seasonal to decadal variability in ice discharge from the Greenland Ice Sheet. *Cryosph.* 12, 3813–3825. doi:10.5194/tc-12-3813-2018.
- Kingslake, J., Scherer, R. P., Albrecht, T., Coenen, J., Powell, R. D., Reese, R., et al. (2018). Extensive retreat and re-advance of the West Antarctic Ice Sheet during the Holocene. *Nature* 558, 430–434. doi:10.1038/s41586-018-0208-x.
- Kinnard, C., Zdanowicz, C. M., Fisher, D. A., Isaksson, E., De Vernal, A., and Thompson, L. G. (2011). Reconstructed

- changes in Arctic sea ice over the past 1,450 years. *Nature* 479, 509–512. doi:10.1038/nature10581.
- Kirchmeier-Young, M. C., Zwiers, F. W., and Gillett, N. P. (2017). Attribution of extreme events in Arctic Sea ice extent. *J. Clim.* 30, 553–571. doi:10.1175/JCLI-D-16-0412.1.
- Kjeldsen, K. K., Korsgaard, N. J., Bjørk, A. A., Khan, S. A., Box, J. E., Funder, S., et al. (2015). Spatial and temporal distribution of mass loss from the Greenland Ice Sheet since AD 1900. *Nature* 528, 396. doi:10.1038/nature16183.
- Kjellsson, J., Holland, P. R., Marshall, G. J., Mathiot, P., Aksenov, Y., Coward, A. C., et al. (2015). Model sensitivity of the Weddell and Ross seas, Antarctica, to vertical mixing and freshwater forcing. *Ocean Model.* 94, 141–152. doi:https://doi.org/10.1016/j.ocemod.2015.08.003.
- Kleinherenbrink, M., Riva, R., and Sun, Y. (2016). Sub-basin-scale sea level budgets from satellite altimetry, Argo floats and satellite gravimetry: A case study in the North Atlantic Ocean. *Ocean Sci.* doi:10.5194/os-12-1179-2016.
- Klerk, W.-J., Winsemius, H. C., Van Verseveld, W. J., Bakker, A. M. R., and Diermanse, F. L. M. (2015). The co-occurrence of storm surges and extreme discharges within the Rhine--Meuse Delta. *Environ. Res. Lett.* 10, 35005.
- Knies, J., Pathirana, I., Cabedo-Sanz, P., Banica, A., Fabian, K., Rasmussen, T. L., et al. (2017). Sea-ice dynamics in an Arctic coastal polynya during the past 6500 years. *Arktos* 3, 1. doi:10.1007/s41063-016-0027-y.
- Knight, J. R. (2005). A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophys. Res. Lett.* 32, L20708. doi:10.1029/2005GL024233.
- Knutson, T. R., and Manabe, S. (1995). Time-mean response over the tropical Pacific to increased CO₂ in a coupled ocean atmosphere model. *J. Clim.* doi:10.1175/1520-0442(1995)008<2181:TMROTT>2.0.CO;2.
- Koch, J., Clague, J. J., and Osborn, G. (2014). Alpine glaciers and permanent ice and snow patches in western Canada approach their smallest sizes since the mid-Holocene, consistent with global trends. *The Holocene* 24, 1639–1648. doi:10.1177/0959683614551214.
- Kociuba, G., and Power, S. B. (2015). Inability of CMIP5 models to simulate recent strengthening of the walker circulation: Implications for projections. *J. Clim.* doi:10.1175/JCLI-D-13-00752.1.
- Koenig, L. S., Ivanoff, A., Alexander, P. M., MacGregor, J. A., Fettweis, X., Panzer, B., et al. (2016). Annual Greenland accumulation rates (2009–2012) from airborne snow radar. *Cryosph.* 10, 1739–1752. doi:10.5194/tc-10-1739-2016.
- Koenig, L. S., Miège, C., Forster, R. R., and Brucker, L. (2014). Initial in situ measurements of perennial meltwater storage in the Greenland firn aquifer. *Geophys. Res. Lett.* 41, 81–85. doi:10.1002/2013GL058083.
- Koenig, S. J., Dolan, A. M., de Boer, B., Stone, E. J., Hill, D. J., DeConto, R. M., et al. (2015). Ice sheet model dependency of the simulated Greenland Ice Sheet in the mid-Pliocene. *Clim. Past* 11, 369–381.
- Koenigk, T., and Brodeau, L. (2014). Ocean heat transport into the Arctic in the twentieth and twenty-first century in EC-Earth. *Clim. Dyn.* doi:10.1007/s00382-013-1821-x.
- Kohout, A. L., Williams, M. J. M., Dean, S. M., and Meylan, M. H. (2014). Storm-induced sea-ice breakup and the implications for ice extent. *Nature* 509, 604–607. doi:10.1038/nature13262.
- Kohyama, T., Hartmann, D. L., and Battisti, D. S. (2017). La Niña-like mean-state response to global warming and potential oceanic roles. *J. Clim.* doi:10.1175/JCLI-D-16-0441.1.
- Kolling, H. M., Stein, R., Fahl, K., Perner, K., and Moros, M. (2018). New insights into sea ice changes over the past 2.2 kyr in Disko Bugt, West Greenland. *Arktos* 4, 11. doi:10.1007/s41063-018-0045-z.
- Kolodziejczyk, N., Llovel, W., and Portela, E. (2019). Interannual Variability of Upper Ocean Water Masses as Inferred From Argo Array. *J. Geophys. Res. Ocean.* 124, 6067–6085. doi:10.1029/2018JC014866.
- Konrad, H., Shepherd, A., Gilbert, L., Hogg, A. E., McMillan, M., Muir, A., et al. (2018). Net retreat of Antarctic glacier grounding lines. *Nat. Geosci.* 11, 258–262. doi:10.1038/s41561-018-0082-z.
- Kopp, R. E., DeConto, R. M., Bader, D. A., Hay, C. C., Horton, R. M., Kulp, S., et al. (2017). Evolving Understanding of Antarctic Ice-Sheet Physics and Ambiguity in Probabilistic Sea-Level Projections. *Earth's Futur.* 5, 1217–1233. doi:10.1002/2017EF000663.
- Kopp, R. E., Horton, R. M., Little, C. M., Mitrovica, J. X., Oppenheimer, M., Rasmussen, D. J., et al. (2014). Probabilistic 21st and 22nd century sea-level projections at a global network of tide gauge sites. *Earth's Futur.* 2, 383–406. doi:10.1002/2014EF000239.
- Kopp, R. E., Kemp, A. C., Bittermann, K., Horton, B. P., Donnelly, J. P., Gehrels, W. R., et al. (2016). Temperature-driven global sea-level variability in the Common Era. *Proc. Natl. Acad. Sci.* 113, E1434–E1441. doi:10.1073/pnas.1517056113.
- Korsgaard, N. J., Nuth, C., Khan, S. A., Kjeldsen, K. K., Bjørk, A. A., Schomacker, A., et al. (2016). Digital elevation model and orthophotographs of Greenland based on aerial photographs from 1978–1987. *Sci. Data* 3, 160032. doi:10.1038/sdata.2016.32.
- Kostov, Y., Ferreira, D., Armour, K. C., and Marshall, J. (2018). Contributions of Greenhouse Gas Forcing and the Southern Annular Mode to Historical Southern Ocean Surface Temperature Trends. *Geophys. Res. Lett.* 45, 1086–1097. doi:10.1002/2017GL074964.
- Kostov, Y., Marshall, J., Hausmann, U., Armour, K. C., Ferreira, D., and Holland, M. M. (2017). Fast and slow responses of Southern Ocean sea surface temperature to SAM in coupled climate models. *Clim. Dyn.* 48, 1595–

1609. doi:10.1007/s00382-016-3162-z.
- Koven, C. D., Riley, W. J., and Stern, A. (2013). Analysis of permafrost thermal dynamics and response to climate change in the CMIP5 earth system models. *J. Clim.* 26, 1877–1900. doi:10.1175/JCLI-D-12-00228.1.
- Kraaijenbrink, P. D. A., Bierkens, M. F. P., Lutz, A. F., and Immerzeel, W. W. (2017). Impact of a global temperature rise of 1.5 degrees Celsius on Asia's glaciers. *Nature* 549, 257–260. doi:10.1038/nature23878.
- Krasting, J. P., Stouffer, R. J., Griffies, S. M., Hallberg, R. W., Malyshev, S. L., Samuels, B. L., et al. (2018). Role of Ocean Model Formulation in Climate Response Uncertainty. *J. Clim.* 31, 9313–9333. doi:10.1175/JCLI-D-18-0035.1.
- Kriegler, E., and Held, H. (2005). Utilizing belief functions for the estimation of future climate change. *Int. J. Approx. Reason.* 39, 185–209. doi:10.1016/j.ijar.2004.10.005.
- Krinner, G., Langeron, C., Ménégoz, M., Agosta, C., and Brutel-Vuilmet, C. (2014). Oceanic forcing of Antarctic climate change: A study using a stretched-grid atmospheric general circulation model. *J. Clim.* 27, 5786–5800. doi:10.1175/JCLI-D-13-00367.1.
- Krug, J., Weiss, J., Gagliardini, O., and Durand, G. (2014). Combining damage and fracture mechanics to model calving. *Cryosph.* 8, 2101–2117. doi:10.5194/tc-8-2101-2014.
- Kucharski, F., Ikram, F., Molteni, F., Farneti, R., Kang, I. S., No, H. H., et al. (2016). Atlantic forcing of Pacific decadal variability. *Clim. Dyn.* doi:10.1007/s00382-015-2705-z.
- Kucharski, F., Kang, I.-S. S., Farneti, R., and Feudale, L. (2011). Tropical Pacific response to 20th century Atlantic warming. *Geophys. Res. Lett.* 38, n/a-n/a. doi:10.1029/2010GL046248.
- Kucharski, F., Syed, F. S., Burhan, A., Farah, I., and Gohar, A. (2014). Tropical Atlantic influence on Pacific variability and mean state in the twentieth century in observations and CMIP5. *Clim. Dyn.* doi:10.1007/s00382-014-2228-z.
- Kuhlbrodt, T., and Gregory, J. M. (2012). Ocean heat uptake and its consequences for the magnitude of sea level rise and climate change. *Geophys. Res. Lett.* doi:10.1029/2012GL052952.
- Kuhlbrodt, T., Griesel, A., Montoya, M., Levermann, A., Hofmann, M., and Rahmstorf, S. (2007). On the driving processes of the Atlantic meridional overturning circulation. *Rev. Geophys.* 45, RG2001. doi:10.1029/2004RG000166.
- Kumar, P., Min, S. K., Weller, E., Lee, H., and Wang, X. L. (2016). Influence of climate variability on extreme ocean surface wave heights assessed from ERA-interim and ERA-20C. *J. Clim.* doi:10.1175/JCLI-D-15-0580.1.
- Kurtz, N. T., and Markus, T. (2012). Satellite observations of Antarctic sea ice thickness and volume. *J. Geophys. Res. Ocean.* 117, C08025. doi:10.1029/2012JC008141.
- Kusahara, K., Hasumi, H., and Williams, G. D. (2011). Impact of the Mertz Glacier Tongue calving on dense water formation and export. *Nat. Commun.* 2, 159. Available at: <https://doi.org/10.1038/ncomms1156>.
- Kwok, R. (2018). Arctic sea ice thickness, volume, and multiyear ice coverage: losses and coupled variability (1958–2018). *Environ. Res. Lett.* 13, 105005. doi:10.1088/1748-9326/aae3ec.
- Kwok, R., Comiso, J. C., Lee, T., and Holland, P. R. (2016). Linked trends in the South Pacific sea ice edge and Southern Oscillation Index. *Geophys. Res. Lett.* 43, 10,210-295,302. doi:10.1002/2016GL070655.
- Kwok, R., and Cunningham, G. F. (2015). Variability of Arctic sea ice thickness and volume from CryoSat-2. *Philos. Trans. R. Soc. A Math. Phys. Eng. Sci.* 373, 20140157. doi:10.1098/rsta.2014.0157.
- Kwok, R., and Kacimi, S. (2018). Three years of sea ice freeboard, snow depth, and ice thickness of the weddell sea from operation icebridge and cryosat-2. *Cryosphere*. doi:10.5194/tc-12-2789-2018.
- Kwok, R., Pang, S. S., and Kacimi, S. (2017). Sea ice drift in the Southern Ocean: Regional patterns, variability, and trends. *Elem Sci Anth* 5, 32. doi:10.1525/elementa.226.
- Kwok, R., Spreen, G., and Pang, S. (2013). Arctic sea ice circulation and drift speed: Decadal trends and ocean currents. *J. Geophys. Res. Ocean.* doi:10.1002/jgrc.20191.
- Lafaysse, M., Cluzet, B., Dumont, M., Lejeune, Y., Vionnet, V., and Morin, S. (2017). A multiphysical ensemble system of numerical snow modelling. *Cryosph.* 11, 1173–1198. doi:10.5194/tc-11-1173-2017.
- Lago, V., and England, M. H. (2019). Projected Slowdown of Antarctic Bottom Water Formation in Response to Amplified Meltwater Contributions. *J. Clim.* 32, 6319–6335. doi:10.1175/JCLI-D-18-0622.1.
- Lago, V., Wijffels, S. E., Durack, P. J., Church, J. A., Bindoff, N. L., and Marsland, S. J. (2016). Simulating the Role of Surface Forcing on Observed Multidecadal Upper-Ocean Salinity Changes. *J. Clim.* 29, 5575–5588. doi:10.1175/JCLI-D-15-0519.1.
- Laliberté, F., Howell, S. E. L., and Kushner, P. J. (2016). Regional variability of a projected sea ice-free Arctic during the summer months. *Geophys. Res. Lett.* 43, 256–263. doi:10.1002/2015GL066855.
- Laliberté, F., Howell, S. E. L., Lemieux, J.-F., Dupont, F., and Lei, J. (2018). What historical landfast ice observations tell us about projected ice conditions in Arctic archipelagoes and marginal seas under anthropogenic forcing. *Cryosph.* 12, 3577–3588. doi:10.5194/tc-12-3577-2018.
- Lambeck, K., Rouby, H., Purcell, A., Sun, Y., and Sambridge, M. (2014). Sea level and global ice volumes from the Last Glacial Maximum to the Holocene. *Proc. Natl. Acad. Sci.*, 201411762. doi:10.1073/pnas.1411762111.
- Landais, A., Masson-Delmotte, V., Capron, E., Langebroek, P. M., Bakker, P., Stone, E. J., et al. (2016). How warm was Greenland during the last interglacial period? *Clim. Past* 12, 1933–1948. doi:10.5194/cp-12-1933-2016.
- Landerer, F. W., Jungclauss, J. H., and Marotzke, J. (2007). Regional Dynamic and Steric Sea Level Change in

- Response to the IPCC-A1B Scenario. *J. Phys. Oceanogr.* 37, 296–312. doi:10.1175/JPO3013.1.
- Larour, E., Ivins, E. R., and Adhikari, S. (2017). Should coastal planners have concern over where land ice is melting? *Sci. Adv.* 3, e1700537.
- Larour, E., Seroussi, H., Adhikari, S., Ivins, E., Caron, L., Morlighem, M., et al. (2019). Slowdown in Antarctic mass loss from solid Earth and sea-level feedbacks. *Science* 364. doi:10.1126/science.aav7908.
- Larour, E., Utke, J., Bovin, A., Morlighem, M., and Perez, G. (2016). An approach to computing discrete adjoints for MPI-parallelized models applied to Ice Sheet System Model 4.11. *Geosci. Model Dev.* 9, 3907–3918. doi:10.5194/gmd-9-3907-2016.
- Larour, E., Utke, J., Csatho, B., Schenk, A., Seroussi, H., Morlighem, M., et al. (2014). Inferred basal friction and surface mass balance of the Northeast Greenland Ice Stream using data assimilation of ICESat (Ice Cloud and land Elevation Satellite) surface altimetry and ISSM (Ice Sheet System Model). *Cryosph.* 8, 2335–2351. doi:10.5194/tc-8-2335-2014.
- Larsen, N. K., Funder, S., Kjær, K. H., Kjeldsen, K. K., Knudsen, M. F., and Linge, H. (2014). Rapid early Holocene ice retreat in West Greenland. *Quat. Sci. Rev.* 92, 310–323.
- Larsen, N. K., Kjær, K. H., Lecavalier, B., Bjørk, A. A., Colding, S., Huybrechts, P., et al. (2015). The response of the southern Greenland ice sheet to the Holocene thermal maximum. *Geology* 43, 291–294.
- Larsen, N. K., Levy, L. B., Strunk, A., Søndergaard, A. S., Olsen, J., and Lauridsen, T. L. (2019). Local ice caps in Finderup Land, North Greenland, survived the Holocene Thermal Maximum. *Boreas* 48, 551–562. doi:10.1111/bor.12384.
- Latif, M., Martin, T., Reintges, A., and Park, W. (2017). Southern Ocean Decadal Variability and Predictability. *Curr. Clim. Chang. Reports* 3, 163–173. doi:10.1007/s40641-017-0068-8.
- Laufkotter, C., Frolicher, T. L., and Zscheischler, J. (submitted). High-impact marine heatwaves attributable to human-induced global warming. *Science* (submitted).
- Lauvset, S. K., Key, R. M., Olsen, A., van Heuven, S., Velo, A., Lin, X., et al. (2016). A new global interior ocean mapped climatology: the 1° × 1° GLODAP version 2. *Earth Syst. Sci. Data* 8, 325–340. doi:10.5194/essd-8-325-2016.
- Lavergne, T., Macdonald Sørensen, A., Kern, S., Tonboe, R., Notz, D., Aaboe, S., et al. (2019). Version 2 of the EUMETSAT OSI SAF and ESA CCI sea-ice concentration climate data records. *Cryosph.* 13, 49–78. doi:10.5194/tc-13-49-2019.
- Laxon, S. W., Giles, K. A., Ridout, A. L., Wingham, D. J., Willatt, R., Cullen, R., et al. (2013). CryoSat-2 estimates of Arctic sea ice thickness and volume. *Geophys. Res. Lett.* 40, 732–737. doi:10.1002/grl.50193.
- Lazeroms, W. M. J., Jenkins, A., Hilmar Gudmundsson, G., and Van De Wal, R. S. W. (2018). Modelling present-day basal melt rates for Antarctic ice shelves using a parametrization of buoyant meltwater plumes. *Cryosphere*. doi:10.5194/tc-12-49-2018.
- Le Bars, D., Drijfhout, S., and De Vries, H. (2017). A high-end sea level rise probabilistic projection including rapid Antarctic ice sheet mass loss. *Environ. Res. Lett.* 12. doi:10.1088/1748-9326/aa6512.
- Le clec'h, S., Charbit, S., Quiquet, A., Fettweis, X., Dumas, C., Kageyama, M., et al. (2019). Assessment of the Greenland ice sheet–atmosphere feedbacks for the next century with a regional atmospheric model coupled to an ice sheet model. *Cryosph.* 13, 373–395. doi:10.5194/tc-13-373-2019.
- Le Cozannet, G., Bulteau, T., Castelle, B., Ranasinghe, R., Wöppelmann, G., Rohmer, J., et al. (2019). Quantifying uncertainties of sandy shoreline change projections as sea level rises. *Sci. Rep.* 9, 42.
- Le Cozannet, G., Manceau, J. C., and Rohmer, J. (2017). Bounding probabilistic sea-level projections within the framework of the possibility theory. *Environ. Res. Lett.* doi:10.1088/1748-9326/aa5528.
- Lea, J. M., Mair, D. W. F., Nick, F. M., Rea, B. R., van As, D., Morlighem, M., et al. (2014a). Fluctuations of a Greenlandic tidewater glacier driven by changes in atmospheric forcing: observations and modelling of Kangeriata Nunaata Sermia, 1859–present. *Cryosph.* 8, 2031–2045. doi:10.5194/tc-8-2031-2014.
- Lea, J. M., Mair, D. W. F., Nick, F. M., Rea, B. R., Weidick, A., Kjær, K. H., et al. (2014b). Terminus-driven retreat of a major southwest Greenland tidewater glacier during the early 19th century: insights from glacier reconstructions and numerical modelling. *J. Glaciol.* 60, 333–344. doi:DOI: 10.3189/2014JoG13J163.
- Lecavalier, B. S., Fisher, D. A., Milne, G. A., Vinther, B. M., Tarasov, L., Huybrechts, P., et al. (2017). High Arctic Holocene temperature record from the Agassiz ice cap and Greenland ice sheet evolution. *Proc. Natl. Acad. Sci.* doi:10.1016/j.phpro.2012.10.059.
- Lecavalier, B. S., Milne, G. A., Simpson, M. J. R., Wake, L., Huybrechts, P., Tarasov, L., et al. (2014). A model of Greenland ice sheet deglaciation constrained by observations of relative sea level and ice extent. *Quat. Sci. Rev.* 102, 54–84. doi:10.1016/J.QUASCIREV.2014.07.018.
- Lecomte, O., Goosse, H., Fichet, T., de Lavergne, C., Barthelemy, A., and Zunz, V. (2017). Vertical ocean heat redistribution sustaining sea-ice concentration trends in the Ross Sea. *Nat. Commun.* 8. doi:10.1038/s41467-017-00347-4.
- Lecomte, O., Goosse, H., Fichet, T., Holland, P. R., Uotila, P., Zunz, V., et al. (2016). Impact of surface wind biases on the Antarctic sea ice concentration budget in climate models. *Ocean Model.* 105, 60–70. doi:10.1016/j.ocemod.2016.08.001.

- 1 Lee, B. S., Haran, M., and Keller, K. (2017a). Multidecadal Scale Detection Time for Potentially Increasing Atlantic
2 Storm Surges in a Warming Climate. *Geophys. Res. Lett.* 44, 10,610–617,623. doi:10.1002/2017GL074606.
- 3 Lee, H. J., Kwon, M. O., Yeh, S. W., Kwon, Y. O., Park, W., Park, J. H., et al. (2017b). Impact of poleward moisture
4 transport from the North Pacific on the acceleration of sea ice loss in the Arctic since 2002. *J. Clim.* 30, 6757–
5 6769. doi:10.1175/JCLI-D-16-0461.1.
- 6 Lee, H., Swenson, S. C., Slater, A. G., and Lawrence, D. M. (2014). Effects of excess ground ice on projections of
7 permafrost in a warming climate. *Environ. Res. Lett.* 9, 124006. doi:10.1088/1748-9326/9/12/124006.
- 8 Lee, S., Im, J., Kim, J., Kim, M., Shin, M., Kim, H. cheol, et al. (2016). Arctic Sea Ice Thickness Estimation from
9 CryoSat-2 Satellite Data Using Machine Learning-Based Lead Detection. *Remote Sens.* 8, 698.
10 doi:10.3390/rs8090698.
- 11 Lee, V., Cornford, S. L., and Payne, A. J. (2015). Initialization of an ice-sheet model for present-day Greenland. *Ann.*
12 *Glaciol.* 56, 129–140. doi:10.3189/2015AoG70A121.
- 13 Legg, S., Briegleb, B., Chang, Y., Chassignet, E. P., Danabasoglu, G., Ezer, T., et al. (2009). Improving Oceanic
14 Overflow Representation in Climate Models: The Gravity Current Entrainment Climate Process Team. *Bull. Am.*
15 *Meteorol. Soc.* 90, 657–670. doi:10.1175/2008BAMS2667.1.
- 16 Lempert, R. J., Popper, S. W., and Bankes, S. C. (2003). *Shaping the {{Next One Hundred Years}}: {{New Methods}}*
17 *for {{Quantitative}}, {{Long}}-{{Term Policy Analysis}}.* Rand Corporation.
- 18 Lenaerts, J. T. M., Fyke, J., and Medley, B. (2018). The Signature of Ozone Depletion in Recent Antarctic Precipitation
19 Change: A Study With the Community Earth System Model. *Geophys. Res. Lett.* 45, 12931–12939.
20 doi:10.1029/2018GL078608.
- 21 Lenaerts, J. T. M., Le Bars, D., Van Kampenhout, L., Vizcaino, M., Enderlin, E. M., and Van Den Broeke, M. R.
22 (2015). Representing Greenland ice sheet freshwater fluxes in climate models. *Geophys. Res. Lett.* 42, 6373–
23 6381. doi:10.1002/2015GL064738.
- 24 Lenaerts, J. T. M., Lhermitte, S., Drews, R., Ligtenberg, S. R. M., Berger, S., Helm, V., et al. (2017). Meltwater
25 produced by wind-albedo interaction stored in an East Antarctic ice shelf. *Nat. Clim. Chang.* 7, 58–62.
26 doi:10.1038/nclimate3180.
- 27 Lenaerts, J. T. M., van den Broeke, M. R., Déry, S. J., van Meijgaard, E., van de Berg, W. J., Palm, S. P., et al. (2012).
28 Modeling drifting snow in Antarctica with a regional climate model: 1. Methods and model evaluation. *J.*
29 *Geophys. Res. Atmos.* 117. doi:10.1029/2011JD016145.
- 30 Lenaerts, J. T. M., Vizcaino, M., Fyke, J., van Kampenhout, L., and van den Broeke, M. R. (2016). Present-day and
31 future Antarctic ice sheet climate and surface mass balance in the Community Earth System Model. *Clim. Dyn.*
32 47, 1367–1381. doi:10.1007/s00382-015-2907-4.
- 33 Letcher, T. W., and Minder, J. R. (2015). Characterization of the simulated regional snow albedo feedback using a
34 regional climate model over complex terrain. *J. Clim.* 28, 7576–7595. doi:10.1175/JCLI-D-15-0166.1.
- 35 Letterly, A., Key, J., and Liu, Y. (2016). The influence of winter cloud on summer sea ice in the Arctic, 1983–2013. *J.*
36 *Geophys. Res.* 121, 2178–2187. doi:10.1002/2015JD024316.
- 37 Levermann, A., Clark, P. U. P. U., Marzeion, B., Milne, G. A. G. A. G. A. G. A., Pollard, D., Radic, V., et al. (2013).
38 The multimillennial sea-level commitment of global warming. *Proc. Natl. Acad. Sci. U. S. A.* 110, 13745–50.
39 doi:10.1073/pnas.1219414110.
- 40 Levermann, A., and Winkelmann, R. (2016). A simple equation for the melt elevation feedback of ice sheets.
41 *Cryosphere* 10, 1799–1807. doi:10.5194/tc-10-1799-2016.
- 42 Levermann, A., Winkelmann, R., Albrecht, T., Goelzer, H., Golledge, N. R., Greve, R., et al. (submitteda). Projecting
43 Antarctica’s contribution to future sea level rise from basal ice shelf melt using linear response functions of 16 ice
44 sheet models (LARMIP-2). *Earth Syst. Dyn. Discuss.* (submitted). doi:https://doi.org/10.5194/esd-11-1-2020.
- 45 Levermann, A., Winkelmann, R., Albrecht, T., Goelzer, H., Nicholas, R., Greve, R., et al. (submitted, b). Projecting
46 Antarctica’s contribution to future sea level rise from basal ice-shelf melt using linear response functions of 16 ice
47 sheet models (LARMIP-2). *Earth Syst. Dyn. Discuss.* (submitted).
- 48 Levermann, A., Winkelmann, R., Nowicki, S., Fastook, J. L. L., Frieler, K., Greve, R., et al. (2014). Projecting
49 Antarctic ice discharge using response functions from SeaRISE ice-sheet models. *Earth Syst. Dyn.* 5, 271–293.
50 doi:10.5194/esd-5-271-2014.
- 51 Levitus, S., Antonov, J. I., Boyer, T. P., Baranova, O. K., Garcia, H. E., Locarnini, R. A., et al. (2012). World ocean
52 heat content and thermocline sea level change (0–2000 m), 1955–2010. *Geophys. Res. Lett.* 39, n/a-n/a.
53 doi:10.1029/2012GL051106.
- 54 Levy, R. H., Meyers, S. R., Naish, T. R., Golledge, N. R., McKay, R. M., Crampton, J. S., et al. (2019). Antarctic ice-
55 sheet sensitivity to obliquity forcing enhanced through ocean connections. *Nat. Geosci.* 12, 132–137.
56 doi:10.1038/s41561-018-0284-4.
- 57 Lewis, M. J., Palmer, T., Hashemi, R., Robins, P., Saulter, A., Brown, J., et al. (2019). Wave-tide interaction modulates
58 nearshore wave height. *Ocean Dyn.* doi:10.1007/s10236-018-01245-z.
- 59 Li, C., Notz, D., Tietsche, S., and Marotzke, J. (2013a). The transient versus the equilibrium response of sea ice to
60 global warming. *J. Clim.* 26, 5624–5636. doi:10.1175/JCLI-D-12-00492.1.
- 61 Li, C., von Storch, J. S., and Marotzke, J. (2013b). Deep-ocean heat uptake and equilibrium climate response. *Clim.*

- Dyn.* doi:10.1007/s00382-012-1350-z.
- Li, G., Du, Y., Xu, H., and Ren, B. (2015a). An Intermodel Approach to Identify the Source of Excessive Equatorial Pacific Cold Tongue in CMIP5 Models and Uncertainty in Observational Datasets. *J. Clim.* 28, 7630–7640. doi:10.1175/JCLI-D-15-0168.1.
- Li, G., and Xie, S.-P. (2014). Tropical Biases in CMIP5 Multimodel Ensemble: The Excessive Equatorial Pacific Cold Tongue and Double ITCZ Problems*. *J. Clim.* 27, 1765–1780. doi:10.1175/JCLI-D-13-00337.1.
- Li, G., Xie, S.-P., Du, Y., and Luo, Y. (2016a). Effects of excessive equatorial cold tongue bias on the projections of tropical Pacific climate change. Part I: the warming pattern in CMIP5 multi-model ensemble. *Clim. Dyn.* 47, 3817–3831. doi:10.1007/s00382-016-3043-5.
- Li, L., Schmitt, R. W., Ummenhofer, C. C., and Karnauskas, K. B. (2016b). Implications of North Atlantic Sea Surface Salinity for Summer Precipitation over the U.S. Midwest: Mechanisms and Predictive Value. *J. Clim.* 29, 3143–3159. doi:10.1175/JCLI-D-15-0520.1.
- Li, Q., and Lee, S. (2017). A Mechanism of Mixed Layer Formation in the Indo–Western Pacific Southern Ocean: Preconditioning by an Eddy-Driven Jet-Scale Overturning Circulation. *J. Phys. Oceanogr.* 47, 2755–2772. doi:10.1175/JPO-D-17-0006.1.
- Li, Q., Webb, A., Fox-Kemper, B., Craig, A., Danabasoglu, G., Large, W. G. W. G., et al. (2016c). Langmuir mixing effects on global climate: WAVEWATCH III in CESM. *Ocean Model.* 103, 145–160. doi:10.1016/j.oceanmod.2015.07.020.
- Li, X., Rignot, E., Morlighem, M., Mouginot, J., and Scheuchl, B. (2015b). Grounding line retreat of Totten Glacier, East Antarctica, 1996 to 2013. *Geophys. Res. Lett.* 42, 8049–8056. doi:10.1002/2015GL065701.
- Li, X., Rignot, E., Mouginot, J., and Scheuchl, B. (2016d). Ice flow dynamics and mass loss of Totten Glacier, East Antarctica, from 1989 to 2015. *Geophys. Res. Lett.* 43, 6366–6373. doi:10.1002/2016GL069173.
- Li, X., Xie, S.-P. P., Gille, S. T., and Yoo, C. (2015c). Atlantic-induced pan-tropical climate change over the past three decades. *Nat. Clim. Chang.* doi:10.1038/nclimate2840.
- Lievens, H., Demuzere, M., Marshall, H. P., Reichle, R. H., Brucker, L., Brangers, I., et al. (2019). Snow depth variability in the Northern Hemisphere mountains observed from space. *Nat. Commun.* 10, 4629. doi:10.1038/s41467-019-12566-y.
- Ligtenberg, S. R. M., van de Berg, W. J., van den Broeke, M. R., Rae, J. G. L., and van Meijgaard, E. (2013). Future surface mass balance of the Antarctic ice sheet and its influence on sea level change, simulated by a regional atmospheric climate model. *Clim. Dyn.* 41, 867–884. doi:10.1007/s00382-013-1749-1.
- Lin, N., Emanuel, K., Oppenheimer, M., and Vanmarcke, E. (2012). Physically based assessment of hurricane surge threat under climate change. *Nat. Clim. Chang.* 2, 462.
- Lin, N., Lane, P., Emanuel, K. A., Sullivan, R. M., and Donnelly, J. P. (2014). Heightened hurricane surge risk in northwest Florida revealed from climatological-hydrodynamic modeling and paleorecord reconstruction. *J. Geophys. Res. Atmos.* 119, 8606–8623.
- Lindsay, R., and Schweiger, A. (2015). Arctic sea ice thickness loss determined using subsurface, aircraft, and satellite observations. *Cryosph.* 9, 269–283. doi:https://doi.org/10.5194/tc-9-269-2015.
- Lindstrom, E., Bryan, F., and Schmitt, R. (2015). SPURS: Salinity Processes in the Upper-ocean Regional Study— The North Atlantic Experiment. *Oceanography* 28. Available at: <https://doi.org/10.5670/oceanog.2015.01>.
- Lippold, J., Pöppelmeier, F., Süfke, F., Gutjahr, M., Goepfert, T. J., Blaser, P., et al. (2019). Constraining the Variability of the Atlantic Meridional Overturning Circulation During the Holocene. *Geophys. Res. Lett.* 46, 11338–11346. doi:10.1029/2019GL084988.
- Lipscomb, W. H., Fyke, J. G., Vizcaino, M., Sacks, W. J., Wolfe, J., Vertenstein, M., et al. (2013). Implementation and Initial Evaluation of the Glimmer Community Ice Sheet Model in the Community Earth System Model. *J. Clim.* 26, 7352–7371. doi:10.1175/JCLI-D-12-00557.1.
- Lipscomb, W. H., Leguy, G. R., Jourdain, N. C., Asay-davis, X., Seroussi, H., and Nowicki, S. (submitted). ISMIP6 projections of ocean-forced Antarctic Ice Sheet evolution using the Community Ice Sheet Model. *Cryosph.* (submitted). doi:10.5194/tc-2019-334.
- Lique, C., Johnson, H. L., and Plancherel, Y. (2018). Emergence of deep convection in the Arctic Ocean under a warming climate. *Clim. Dyn.* 50, 3833–3847. doi:10.1007/s00382-017-3849-9.
- Liston, G. E., and Hiemstra, C. A. (2011). The changing cryosphere: Pan-Arctic snow trends (1979–2009). *J. Clim.* 24, 5691–5712. doi:10.1175/JCLI-D-11-00081.1.
- Little, C. M., Horton, R. M., Kopp, R. E., Oppenheimer, M., Vecchi, G. A., and Villarini, G. (2015). Joint projections of US East Coast sea level and storm surge. *Nat. Clim. Chang.* doi:10.1038/nclimate2801.
- Liu, J., Milne, G. A., 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.* doi:10.1038/ngeo2616.
- Liu, T., Schmitt, R. W., and Li, L. (2018a). Global Search for Autumn-Lead Sea Surface Salinity Predictors of Winter Precipitation in Southwestern United States. *Geophys. Res. Lett.* 45, 8445–8454. doi:10.1029/2018GL079293.
- Liu, W., and Liu, Z. (2013). A Diagnostic Indicator of the Stability of the Atlantic Meridional Overturning Circulation in CCSM3. *J. Clim.* 26, 1926–1938. doi:10.1175/JCLI-D-11-00681.1.
- Liu, W., Lu, J., and Xie, S.-P. (2018b). Southern Ocean Heat Uptake, Redistribution, and Storage in a Warming

- Climate: The Role of Meridional Overturning Circulation. *J. Clim.* 31, 4727–4743. doi:10.1175/JCLI-D-17.
- Liu, W., Xie, S. P., Liu, Z., and Zhu, J. (2017). Overlooked possibility of a collapsed atlantic meridional overturning circulation in warming climate. *Sci. Adv.* doi:10.1126/sciadv.1601666.
- Liu, Y., Moore, J. C., Cheng, X., Gladstone, R. M., Bassis, J. N., Liu, H., et al. (2015). Ocean-driven thinning enhances iceberg calving and retreat of Antarctic ice shelves. *Proc. Natl. Acad. Sci.* 112, 3263–3268. doi:10.1073/pnas.1415137112.
- Llovel, W., Penduff, T., Meyssignac, B., Molines, J., Terray, L., Bessi eres, L., et al. (2018). Contributions of Atmospheric Forcing and Chaotic Ocean Variability to Regional Sea Level Trends Over 1993–2015. *Geophys. Res. Lett.* 45, 13,405–13,413. doi:10.1029/2018GL080838.
- Lo, L., Belt, S. T., Lattaud, J., Friedrich, T., Zeeden, C., Schouten, S., et al. (2018). Precession and atmospheric CO₂ modulated variability of sea ice in the central Okhotsk Sea since 130,000 years ago. *Earth Planet. Sci. Lett.* 488, 36–45. doi:10.1016/j.epsl.2018.02.005.
- Lohmann, G. (2019). Temperatures from Energy Balance Models: the effective heat capacity matters. *Earth Syst. Dyn. Discuss.* doi:https://doi.org/10.5194/esd-2019-35.
- Loranty, M. M., Berner, L. T., Goetz, S. J., Jin, Y., and Randerson, J. T. (2014). Vegetation controls on northern high latitude snow-albedo feedback: Observations and CMIP5 model simulations. *Glob. Chang. Biol.* 20, 594–606. doi:10.1111/gcb.12391.
- Lozier, M. S., Li, F., Bacon, S., Bahr, F., Bower, A. S., Cunningham, S. A., et al. (2019). A sea change in our view of overturning in the subpolar North Atlantic. *Science* 363, 516–521. doi:10.1126/science.aau6592.
- L ubbecke, J. F., Durgadoo, J. V., and Biastoch, A. (2015). Contribution of increased agulhas leakage to tropical Atlantic warming. *J. Clim.* doi:10.1175/JCLI-D-15-0258.1.
- Luckman, A., Benn, D. I., Cottier, F., Bevan, S., Nilsen, F., and Inall, M. (2015). Calving rates at tidewater glaciers vary strongly with ocean temperature. *Nat. Commun.* 6, 8566. doi:10.1038/ncomms9566.
- Luo, D., Wu, Q., Jin, H., Marchenko, S. S., L u, L., and Gao, S. (2016). Recent changes in the active layer thickness across the northern hemisphere. *Environ. Earth Sci.* doi:10.1007/s12665-015-5229-2.
- Luo, J.-J., Sasaki, W., and Masumoto, Y. (2012). Indian Ocean warming modulates Pacific climate change. *Proc. Natl. Acad. Sci.* 109, 18701–18706. doi:10.1073/pnas.1210239109.
- Luo, Y., Lu, J., Liu, F., and Garuba, O. (2017). The role of ocean dynamical thermostat in delaying the El Ni o-Like response over the equatorial Pacific to climate warming. *J. Clim.* doi:10.1175/JCLI-D-16-0454.1.
- Luo, Y., Lu, J., Liu, F., and Liu, W. (2015). Understanding the El Ni o-like oceanic response in the tropical Pacific to global warming. *Clim. Dyn.* doi:10.1007/s00382-014-2448-2.
- L uthi, M. P. (2009). Transient response of idealized glaciers to climate variations. *J. Glaciol.* 55, 918–930. doi:10.3189/002214309790152519.
- L uthi, M. P., and Bauder, A. (2010). Analysis of Alpine glacier length change records with a macroscopic glacier model. *Geogr. Helv.* 65, 92–102. doi:10.5194/gh-65-92-2010.
- L uthi, M. P., Bauder, A., and Funk, M. (2010). Volume change reconstruction of Swiss glaciers from length change data. *J. Geophys. Res.* 115, F04022. doi:10.1029/2010JF001695.
- L uthi, M. P., Vieli, A., Moreau, L., Joughin, I., Reisser, M., Small, D., et al. (2016). A century of geometry and velocity evolution at Ekip Sermia, West Greenland. *J. Glaciol.* 62, 640–654. doi:DOI: 10.1017/jog.2016.38.
- Lutz, S., Anesio, A. M., Raiswell, R., Edwards, A., Newton, R. J., Gill, F., et al. (2016). The biogeography of red snow microbiomes and their role in melting arctic glaciers. *Nat. Commun.* 7, 11968. doi:10.1038/ncomms11968.
- Lynch-Stieglitz, J. (2017). The Atlantic Meridional Overturning Circulation and Abrupt Climate Change. *Ann. Rev. Mar. Sci.* doi:10.1146/annurev-marine-010816-060415.
- Lynch, A. H., Serreze, M. C., Cassano, E. N., Crawford, A. D., and Stroeve, J. (2016). Linkages between Arctic summer circulation regimes and regional sea ice anomalies. *J. Geophys. Res.* 121, 7868–7880. doi:10.1002/2016JD025164.
- Lyu, K., Zhang, X., Church, J. A., Slangen, A. B. A., and Hu, J. (2014). Time of emergence for regional sea-level change. *Nat. Clim. Chang.* 4, 1006–1010. doi:10.1038/nclimate2397.
- Ma, X., Jing, Z., Chang, P., Liu, X., Montuoro, R., Small, R. J., et al. (2016). Western boundary currents regulated by interaction between ocean eddies and the atmosphere. *Nature* 535, 533–537. doi:10.1038/nature18640.
- MacFerrin, M., Machguth, H., As, D. van, Charalampidis, C., Stevens, C. M., Heilig, A., et al. (2019). Rapid expansion of Greenland’s low-permeability ice slabs. *Nature* 573, 403–407. doi:10.1038/s41586-019-1550-3.
- MacGregor, J. A., Fahnestock, M. A., Catania, G. A., Paden, J. D., Prasad Gogineni, S., Young, S. K., et al. (2015). Radiostratigraphy and age structure of the Greenland Ice Sheet. *J. Geophys. Res. Earth Surf.* 120, 212–241. doi:10.1002/2014JF003215.
- Machguth, H., MacFerrin, M., van As, D., Box, J. E., Charalampidis, C., Colgan, W., et al. (2016a). Greenland meltwater storage in firn limited by near-surface ice formation. *Nat. Clim. Chang.* 6, 390. doi:10.1038/nclimate2899.
- Machguth, H., Thomsen, H. H., Weidick, A., Ahlstr m, A. P., Abermann, J., Andersen, M. L., et al. (2016b). Greenland surface mass-balance observations from the ice-sheet ablation area and local glaciers. *J. Glaciol.* 62, 861–887. doi:10.1017/jog.2016.75.

- 1 Mackie, S., Smith, I. J., Ridley, J. K., Stevens, D. P., and Langhorne, P. J. (submitted). Climate response to increasing
2 Antarctic iceberg and ice shelf melt. *J. Clim.* (submitted).
- 3 Mackintosh, A., Golledge, N., Domack, E., Dunbar, R., Leventer, A., White, D., et al. (2011). Retreat of the East
4 Antarctic ice sheet during the last glacial termination. *Nat. Geosci.* 4, 195–202. doi:10.1038/ngeo1061.
- 5 Magalhães, N. de, Evangelista, H., Condom, T., Rabatel, A., and Ginot, P. (2019). Amazonian Biomass Burning
6 Enhances Tropical Andean Glaciers Melting. *Sci. Rep.* 9, 16914. doi:10.1038/s41598-019-53284-1.
- 7 Mahlstein, I., Gent, P. R., and Solomon, S. (2013). Historical Antarctic mean sea ice area, sea ice trends, and winds in
8 CMIP5 simulations. *J. Geophys. Res. Atmos.* 118, 5105–5110. doi:10.1002/jgrd.50443.
- 9 Maksym, T. (2019). Arctic and Antarctic Sea Ice Change: Contrasts, Commonalities, and Causes. *Ann. Rev. Mar. Sci.*
10 11, 187–213. doi:10.1146/annurev-marine-010816-060610.
- 11 Maksym, T., and Markus, T. (2008). Antarctic sea ice thickness and snow-to-ice conversion from atmospheric
12 reanalysis and passive microwave snow depth. *J. Geophys. Res.* 113, C02S12. doi:10.1029/2006JC004085.
- 13 Mankoff, K. D., Colgan, W., Solgaard, A., Karlsson, N. B., Ahlstrom, A. P., van As, D., et al. (2019). Greenland Ice
14 Sheet discharge from 2000 to 2018. *Earth Syst. Sci. Data Discuss.*, 1–31. doi:10.5194/essd-2019-29.
- 15 Manucharyan, G. E., and Thompson, A. F. (2017). Submesoscale Sea Ice-Ocean Interactions in Marginal Ice Zones. *J.*
16 *Geophys. Res. Ocean.* 122, 9455–9475. doi:10.1002/2017JC012895.
- 17 Marcer, M., Serrano, C., Brenning, A., Bodin, X., Goetz, J., and Schoeneich, P. (2019). Evaluating the destabilization
18 susceptibility of active rock glaciers in the French Alps. *Cryosph.* 13, 141–155. doi:10.5194/tc-13-141-2019.
- 19 Marchau, V. A. W. J., Walker, W. E., Bloemen, P. J. T. M., and Popper, S. W. (2019). “Introduction,” in *Decision*
20 *Making under Deep Uncertainty: From Theory to Practice*, eds. V. A. W. J. Marchau, W. E. Walker, P. J. T. M.
21 Bloemen, and S. W. Popper (Cham: Springer International Publishing), 1–20. doi:10.1007/978-3-030-05252-2_1.
- 22 Marchesiello, P., and Estrade, P. (2010). Upwelling limitation by onshore geostrophic flow. *J. Mar. Res.* 68, 37–62.
- 23 Marcos, M., and Amores, A. (2014). Quantifying anthropogenic and natural contributions to thermosteric sea level rise.
24 *Geophys Res Lett* 41, 2502–2507. doi:10.1002/2014GL059766.
- 25 Marcos, M., Calafat, F. M., Berihuete, Á., and Dangendorf, S. (2015). Long-term variations in global sea level
26 extremes. *J. Geophys. Res. Ocean.* 120, 8115–8134. doi:10.1002/2015JC011173.
- 27 Marcos, M., Jordà, G., Gomis, D., and Pérez, B. (2011). Changes in storm surges in southern Europe from a regional
28 model under climate change scenarios. *Glob. Planet. Change* 77, 116–128.
29 doi:10.1016/J.GLOPLACHA.2011.04.002.
- 30 Marcos, M., and Woodworth, P. L. (2017). Spatiotemporal changes in extreme sea levels along the coasts of the North
31 Atlantic and the Gulf of Mexico. *J. Geophys. Res. Ocean.* doi:10.1002/2017JC013065.
- 32 Marcos, M., Wöppelmann, G., Matthews, A., Ponte, R. M., Birol, F., Arduin, F., et al. (2019). Coastal Sea Level and
33 Related Fields from Existing Observing Systems. *Surv. Geophys.* doi:10.1007/s10712-019-09513-3.
- 34 Marcott, S. A., Shakun, J. D., Clark, P. U., and Mix, A. C. (2013). A Reconstruction of Regional and Global
35 Temperature for the Past 11,300 Years. *Science* 339, 1198–1201. doi:10.1126/science.1228026.
- 36 Maris, M., De Boer, B., Ligtnerberg, S., Crucifix, M., Van De Berg, W. J., and Oerlemans, J. (2014). Modelling the
37 evolution of the Antarctic ice sheet since the last interglacial. *Cryosph.* 8, 1347–1360.
- 38 Marshall, A. G., Hemer, M. A., Hendon, H. H., and McInnes, K. L. (2018). Southern annular mode impacts on global
39 ocean surface waves. *Ocean Model.* doi:10.1016/j.ocemod.2018.07.007.
- 40 Marshall, A. G., Hendon, H. H., Durrant, T. H., and Hemer, M. A. (2015a). Madden Julian Oscillation impacts on
41 global ocean surface waves. *Ocean Model.* doi:10.1016/j.ocemod.2015.06.002.
- 42 Marshall, J., Armour, K. C., Scott, J. R. J. R., Kostov, Y., Hausmann, U., Ferreira, D., et al. (2014). The ocean’s
43 role in polar climate change: asymmetric Arctic and Antarctic responses to greenhouse gas and ozone forcing.
44 *Philos. Trans. R. Soc. A Math. Phys. Eng. Sci.* doi:10.1098/rsta.2013.0040.
- 45 Marshall, J., Scott, J. R., Armour, K. C., Campin, J. M., Kelley, M., and Romanou, A. (2015b). The ocean’s role in the
46 transient response of climate to abrupt greenhouse gas forcing. *Clim. Dyn.* doi:10.1007/s00382-014-2308-0.
- 47 Marsooli, R., Lin, N., Emanuel, K., and Feng, K. (2019). Climate change exacerbates hurricane flood hazards along US
48 Atlantic and Gulf Coasts in spatially varying patterns. *Nat. Commun.* doi:10.1038/s41467-019-11755-z.
- 49 Martin, D. F., Cornford, S. L., and Payne, A. J. (2019). Millennial-scale Vulnerability of the Antarctic Ice Sheet to
50 Regional Ice Shelf Collapse. *Geophys. Res. Lett.* 46, 1467–1475. doi:10.1029/2018GL081229.
- 51 Martin, L. C. P., Blard, P.-H., Lavé, J., Condom, T., Prémaillon, M., Jomelli, V., et al. (2018). Lake Taucá highstand
52 (Heinrich Stadial 1a) driven by a southward shift of the Bolivian High. *Sci. Adv.* 4, eaar2514.
53 doi:10.1126/sciadv.aar2514.
- 54 Martin, T., Park, W., and Latif, M. (2013). Multi-centennial variability controlled by Southern Ocean convection in the
55 Kiel Climate Model. *Clim. Dyn.* 40, 2005–2022. doi:10.1007/s00382-012-1586-7.
- 56 Martinec, Z., Klemann, V., van der Wal, W., Riva, R. E. M., Spada, G., Sun, Y., et al. (2018). A benchmark study of
57 numerical implementations of the sea level equation in GIA modelling. *Geophys. J. Int.* doi:10.1093/gji/ggy280.
- 58 Martinson, D. G. (2012). Antarctic circumpolar current’s role in the Antarctic ice system: An overview. *Palaeogeogr.*
59 *Palaeoclimatol. Palaeoecol.* 335–336, 71–74. doi:10.1016/J.PALAEO.2011.04.007.
- 60 Marty, C., Schlögl, S., Bavay, M., and Lehning, M. (2017). How much can we save? Impact of different emission
61 scenarios on future snow cover in the Alps. *Cryosph.* 11, 527–529. doi:10.5194/tc-11-517-2017.

- Marzeion, B., Cogley, J. G., Richter, K., and Parkes, D. (2014). Glaciers. Attribution of global glacier mass loss to anthropogenic and natural causes. *Science* 345, 919–921. doi:10.1126/science.1254702.
- Marzeion, B., Hock, R., Anderson, B., Andrew, B., Champollion, N., Fujita, K., et al. (submitted). Partitioning the Uncertainty of Ensemble Projections of Global Glacier Mass Change. *Earth's Futur.* (submitted).
- Marzeion, B., Jarosch, A. H., and Hofer, M. (2012). Past and future sea-level change from the surface mass balance of glaciers. *Cryosph.* 6, 1295–1322. doi:10.5194/tc-6-1295-2012.
- Marzeion, B., Kaser, G., Maussion, F., and Champollion, N. (2018). Limited influence of climate change mitigation on short-term glacier mass loss. *Nat. Clim. Chang.* 8, 305–308. doi:10.1038/s41558-018-0093-1.
- Marzeion, B., Leclercq, P. W., Cogley, J. G., and Jarosch, A. H. (2015a). Brief Communication: Global reconstructions of glacier mass change during the 20th century are consistent. *Cryosph.* 9, 2399–2404. doi:10.5194/tc-9-2399-2015.
- Marzeion, B., Leclercq, P. W., Cogley, J. G., and Jarosch, A. H. (2015b). Brief Communication: Global reconstructions of glacier mass change during the 20th century are consistent. *Cryosph.* 9, 2399–2404. doi:10.5194/tc-9-2399-2015.
- Mason, E., Molemaker, J., Shchepetkin, A. F., Colas, F., McWilliams, J. C., and Sangrà, P. (2010). Procedures for offline grid nesting in regional ocean models. *Ocean Model.* 35, 1–15. doi:10.1016/j.ocemod.2010.05.007.
- Massom, R. A., Eicken, H., Haas, C., Jeffries, M. O., Drinkwater, M. R., Sturm, M., et al. (2001). Snow on Antarctic sea ice. *Rev. Geophys.* 39, 413–445. doi:10.1029/2000RG000085.
- Massom, R. A., Scambos, T. A., Bennetts, L. G., Reid, P., Squire, V. A., and Stammerjohn, S. E. (2018). Antarctic ice shelf disintegration triggered by sea ice loss and ocean swell. *Nature* 558, 383–389. doi:10.1038/s41586-018-0212-1.
- Massonnet, F., Guemas, V., Fučkar, N. S., and Doblas-Reyes, F. J. (2015). The 2014 High Record of Antarctic Sea Ice Extent [in “Explaining Extremes of 2014 from a Climate Perspective”]. *Bull. Am. Meteorol. Soc.* 96, 163–167. doi:10.1175/BAMS-D-15-00093.1.
- Massonnet, F., Mathiot, P., Fichet, T., Goosse, H., König Beatty, C., Vancoppenolle, M., et al. (2013). A model reconstruction of the Antarctic sea ice thickness and volume changes over 1980–2008 using data assimilation. *Ocean Model.* 64, 67–75. doi:10.1016/j.ocemod.2013.01.003.
- Mathiot, P., Jenkins, A., Harris, C., and Madec, G. (2017). Explicit representation and parametrised impacts of under ice shelf seas in the z*- coordinate ocean model NEMO 3.6. *Geosci. Model Dev.* 10, 2849–2874. doi:10.5194/gmd-10-2849-2017.
- Mathis, M., Elizalde, A., and Mikolajewicz, U. (2018). Which complexity of regional climate system models is essential for downscaling anthropogenic climate change in the Northwest European Shelf? *Clim. Dyn.* 50, 2637–2659. doi:10.1007/s00382-017-3761-3.
- Mathis, M., and Pohlmann, T. (2014). Projection of physical conditions in the North Sea for the 21st century. *Clim. Res.* 61, 1–17. doi:10.3354/cr01232.
- Mauritzen, C., Melsom, A., and Sutton, R. T. (2012). Importance of density-compensated temperature change for deep North Atlantic Ocean heat uptake. *Nat. Geosci.* 5, 905–910. doi:10.1038/ngeo1639.
- Maussion, F., Butenko, A., Champollion, N., Dusch, M., Eis, J., Fourteau, K., et al. (2019). The Open Global Glacier Model (OGGM) v1.1. *Geosci. Model Dev.* 12, 909–931. doi:10.5194/gmd-12-909-2019.
- Maussion, F., Butenko, A., Eis, J., Fourteau, K., Jarosch, A. H., Landmann, J., et al. (2018). The Open Global Glacier Model (OGGM) v1.0. *Geosci. Model Dev. Discuss.* 2018, 1–33. doi:10.5194/gmd-2018-9.
- Mawdsley, R. J., and Haigh, I. D. (2016). Spatial and temporal variability and long-term trends in skew surges globally. *Front. Mar. Sci.* 3, 29.
- Mawdsley, R. J., Haigh, I. D., and Wells, N. C. (2015). Global secular changes in different tidal high water, low water and range levels. *Earth's Futur.* doi:10.1002/2014EF000282.
- McCarthy, G. D., Joyce, T. M., and Josey, S. A. (2018). Gulf Stream Variability in the Context of Quasi-Decadal and Multidecadal Atlantic Climate Variability. *Geophys. Res. Lett.* 45. doi:10.1029/2018GL079336.
- McGregor, S., Timmermann, A., Stuecker, M. F., England, M. H., Merrifield, M., Jin, F.-F. F., et al. (2014). Recent Walker circulation strengthening and Pacific cooling amplified by Atlantic warming. *Nat. Clim. Chang.* 4, 888–892. doi:10.1038/nclimate2330.
- McGuire, A. D., Koven, C., Lawrence, D. M., Klein, J. S., Xia, J., Beer, C., et al. (2016). Variability in the sensitivity among model simulations of permafrost and carbon dynamics in the permafrost region between 1960 and 2009. *Global Biogeochem. Cycles.* doi:10.1002/2016GB005405.
- McInnes, K. L., Walsh, K. J. E., Hoeke, R. K., O'Grady, J. G., Colberg, F., and Hubbert, G. D. (2014). Quantifying storm tide risk in Fiji due to climate variability and change. *Glob. Planet. Change* 116, 115–129.
- McKay, N. P., Overpeck, J. T., and Otto-Bliesner, B. L. (2011). The Role of Ocean Thermal Expansion in {{Last Interglacial}} Sea Level Rise. *Geophys. Res. Lett.* 38, L14605. doi:10.1029/2011GL048280.
- McKay, R., Golledge, N. R., Maas, S., Naish, T., Levy, R., Dunbar, G., et al. (2016). Antarctic marine ice-sheet retreat in the Ross Sea during the early Holocene. *Geology* 44, 7–10. doi:10.1130/G37315.1.
- McMillan, M., Leeson, A., Shepherd, A., Briggs, K., Armitage, T. W. K., Hogg, A., et al. (2016). A high-resolution record of Greenland mass balance. *Geophys. Res. Lett.* 43, 7002–7010. doi:10.1002/2016GL069666.

- McMillan, M., Shepherd, A., Sundal, A., Briggs, K., Muir, A., Ridout, A., et al. (2014). Increased ice losses from Antarctica detected by CryoSat-2. *Geophys. Res. Lett.* 41, 3899–3905. doi:10.1002/2014GL060111.
- Mecking, J. V., Drijfhout, S. S., Jackson, L. C., and Andrews, M. B. (2017). The effect of model bias on Atlantic freshwater transport and implications for AMOC bi-stability. *Tellus, Ser. A Dyn. Meteorol. Oceanogr.* 69. doi:10.1080/16000870.2017.1299910.
- Medley, B., and Thomas, E. R. (2019). Increased snowfall over the Antarctic Ice Sheet mitigated twentieth-century sea-level rise. *Nat. Clim. Chang.* doi:10.1038/s41558-018-0356-x.
- Meehl, G. A., Arblaster, J. M., Bitz, C. M., Chung, C. T. Y., and Teng, H. (2016). Antarctic sea-ice expansion between 2000 and 2014 driven by tropical Pacific decadal climate variability. *Nat. Geosci.* 9, 590–595. doi:10.1038/ngeo2751.
- Meehl, G. A., Arblaster, J. M., Chung, C. T. Y., Holland, M. M., DuVivier, A., Thompson, L., et al. (2019). Sustained ocean changes contributed to sudden Antarctic sea ice retreat in late 2016. *Nat. Commun.* 10, 14. doi:10.1038/s41467-018-07865-9.
- Meehl, G. A., Goddard, L., Murphy, J., Stouffer, R. J., Boer, G., Danabasoglu, G., et al. (2009). Decadal prediction: Can it be skillful? *Bull. Am. Meteorol. Soc.* doi:10.1175/2009BAMS2778.1.
- Meehl, G. A., and Washington, W. M. (1996). El Niño-like climate change in a model with increased atmospheric CO₂ concentrations. *Nature*. doi:10.1038/382056a0.
- Méheust, M., Stein, R., Fahl, K., and Gersonde, R. (2018). Sea-ice variability in the subarctic North Pacific and adjacent Bering Sea during the past 25 ka: new insights from IP25 and Uk'37 proxy records. *Arktos* 4, 8. doi:10.1007/s41063-018-0043-1.
- Meier, H. E. M. (2015). “Projected Change—Marine Physics,” in (Springer, Cham), 243–252. doi:10.1007/978-3-319-16006-1_13.
- Meier, W. N., Gallaher, D., and Campbell, G. G. (2013). New estimates of Arctic and Antarctic sea ice extent during September 1964 from recovered Nimbus I satellite imagery. *Cryosph.* 7, 699–705. doi:10.5194/tc-7-699-2013.
- Meier, W. N., and Stewart, J. S. (2019). Assessing uncertainties in sea ice extent climate indicators. *Environ. Res. Lett.* 14, 035005. doi:10.1088/1748-9326/aaf52c.
- Meijers, A. J. S. S., Shuckburgh, E., Bruneau, N., Sallee, J. B., Bracegirdle, T. J., Wang, Z., et al. (2012). Representation of the Antarctic Circumpolar Current in the CMIP5 climate models and future changes under warming scenarios. *J. Geophys. Res.* 117, C12008. doi:10.1029/2012JC008412.
- Melet, A., Meyssignac, B., Almar, R., and Le Cozannet, G. (2018). Under-estimated wave contribution to coastal sea-level rise. *Nat. Clim. Chang.* doi:10.1038/s41558-018-0088-y.
- Melia, N., Haines, K., and Hawkins, E. (2016). Sea ice decline and 21st century trans-Arctic shipping routes. *Geophys. Res. Lett.* 43, 9720–9728. doi:10.1002/2016GL069315.
- Melzer, B. A., and Subrahmanyam, B. (2017). Decadal changes in salinity in the oceanic subtropical gyres. *J. Geophys. Res. Ocean.* 122, 336–354. doi:10.1002/2016JC012243.
- Menary, M. B., Hodson, D. L. R., Robson, J. I., Sutton, R. T., Wood, R. A., and Hunt, J. A. (2015). Exploring the impact of CMIP5 model biases on the simulation of North Atlantic decadal variability. *Geophys. Res. Lett.* 42, 5926–5934. doi:10.1002/2015GL064360.
- Menary, M. B., Park, W., Lohmann, K., Vellinga, M., Palmer, M. D., Latif, M., et al. (2012). A multimodel comparison of centennial Atlantic meridional overturning circulation variability. *Clim. Dyn.* 38, 2377–2388. doi:10.1007/s00382-011-1172-4.
- Menary, M. B., and Wood, R. A. (2018). An anatomy of the projected North Atlantic warming hole in CMIP5 models. *Clim. Dyn.* doi:10.1007/s00382-017-3793-8.
- Menary, M., Robson, J., Allan, R., Booth, B., Cassou, C., Gastineau, G., et al. (submitted). Aerosol-forced AMOC change in CMIP6 historical simulations. *Nat. Geosci.* (submitted).
- Mendoza, P. A., Clark, M. P., Mizukami, N., Newman, A. J., Barlage, M., Gutmann, E. D., et al. (2015). Effects of Hydrologic Model Choice and Calibration on the Portrayal of Climate Change Impacts. *J. Hydrometeorol.* 16, 762–780. doi:10.1175/JHM-D-14-0104.1.
- Ménégou, M., Krinner, G., Balkanski, Y., Boucher, O., Cozic, A., Lim, S., et al. (2014). Snow cover sensitivity to black carbon deposition in the Himalayas: From atmospheric and ice core measurements to regional climate simulations. *Atmos. Chem. Phys.* 14, 4237–4249. doi:10.5194/acp-14-4237-2014.
- Menéndez, M., and Woodworth, P. L. (2010). Changes in extreme high water levels based on a quasi-global tide-gauge data set. *J. Geophys. Res. Ocean.* 115.
- Menezes, V. V., Macdonald, A. M., and Schatzman, C. (2017). Accelerated freshening of Antarctic Bottom Water over the last decade in the Southern Indian Ocean. *Sci. Adv.* doi:10.1126/sciadv.1601426.
- Mengel, M., Levermann, A., Frieler, K., Robinson, A., Marzeion, B., and Winkelmann, R. (2016). Future sea level rise constrained by observations and long-term commitment. *Proc. Natl. Acad. Sci.* 113, 2597–2602. doi:10.1073/pnas.1500515113.
- Menounos, B., Hugonnet, R., Shean, D., Gardner, A., Howat, I., Berthier, E., et al. (2019). Heterogeneous Changes in Western North American Glaciers Linked to Decadal Variability in Zonal Wind Strength. *Geophys. Res. Lett.* 46, 200–209. doi:10.1029/2018GL080942.

- 1 Menviel, L., Timmermann, A., Timm, O. E., and Mouchet, A. (2010). Climate and biogeochemical response to a rapid
2 melting of the West Antarctic Ice sheet during interglacials and implications for future climate.
3 *Paleoceanography* 25, PA4321. doi:10.1029/2009PA001892.
- 4 Mercenier, R., Lüthi, M. P., and Vieli, A. (2018). Calving relation for tidewater glaciers based on detailed stress field
5 analysis. *Cryosph.* 12, 721–739. doi:10.5194/tc-12-721-2018.
- 6 Mercer, J. H. (1978). West Antarctic ice sheet and CO₂ greenhouse effect: a threat of disaster. *Nature* 271, 321.
7 doi:10.1038/271321a0.
- 8 Meredith, M., Sommerkorn, M., Cassotta, S., Derksen, C., Ekaykin, A., Hollowed, A., et al. (2019). “Polar Regions,” in
9 *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*, eds. H.-O. Pörtner, D. C. Roberts, V.
10 Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, et al.
- 11 Merino, N., Jourdain, N. C., Le Sommer, J., Goosse, H., Mathiot, P., and Durand, G. (2018). Impact of increasing
12 antarctic glacial freshwater release on regional sea-ice cover in the Southern Ocean. *Ocean Model.* 121, 76–89.
13 doi:10.1016/j.ocemod.2017.11.009.
- 14 Mernild, S. H., Lipscomb, W. H., Bahr, D. B., Radić, V., and Zemp, M. (2013). Global glacier changes: a revised
15 assessment of committed mass losses and sampling uncertainties. *Cryosph.* 7, 1565–1577. doi:10.5194/tc-7-1565-
16 2013.
- 17 Meyssignac, B., Salas Y Melia, D., Becker, M., Llovel, W., and Cazenave, A. (2012). Tropical Pacific spatial trend
18 patterns in observed sea level: Internal variability and/or anthropogenic signature? *Clim. Past* 8, 787–802.
19 doi:10.5194/cp-8-787-2012.
- 20 Meyssignac, B., Slangen, A. B. A., Melet, A., Church, J. A., Fettweis, X., Marzeion, B., et al. (2017). Evaluating model
21 simulations of twentieth-century sea-level rise. Part II: Regional sea-level changes. *J. Clim.* 30, 8565–8593.
22 doi:10.1175/JCLI-D-17-0112.1.
- 23 Miège, C., Forster, R. R., Brucker, L., Koenig, L. S., Solomon, D. K., Paden, J. D., et al. (2016). Spatial extent and
24 temporal variability of Greenland firn aquifers detected by ground and airborne radars. *J. Geophys. Res. Earth*
25 *Surf.* 121, 2381–2398. doi:10.1002/2016JF003869.
- 26 Miles, B. W. J., Stokes, C. R., and Jamieson, S. S. R. (2016). Pan-ice-sheet glacier terminus change in East Antarctica
27 reveals sensitivity of Wilkes Land to sea-ice changes. *Sci. Adv.* 2, e1501350. doi:10.1126/sciadv.1501350.
- 28 Miles, B. W. J., Stokes, C. R., and Jamieson, S. S. R. (2017). Simultaneous disintegration of outlet glaciers in Porpoise
29 Bay (Wilkes Land), East Antarctica, driven by sea ice break-up. *Cryosph.* 11, 427–442. doi:10.5194/tc-11-427-
30 2017.
- 31 Miles, B. W. J., Stokes, C. R., Vieli, A., and Cox, N. J. (2013). Rapid, climate-driven changes in outlet glaciers on the
32 Pacific coast of East Antarctica. *Nature* 500, 563–566. doi:10.1038/nature12382.
- 33 Milillo, P., Rignot, E., Mouginot, J., Scheuchl, B., Morlighem, M., Li, X., et al. (2017). On the Short-term Grounding
34 Zone Dynamics of Pine Island Glacier, West Antarctica, Observed With COSMO-SkyMed Interferometric Data.
35 *Geophys. Res. Lett.* 44, 10,436–10,444. doi:10.1002/2017GL074320.
- 36 Milillo, P., Rignot, E., Rizzoli, P., Scheuchl, B., Mouginot, J., Bueso-Bello, J., et al. (2019). Heterogeneous retreat and
37 ice melt of thwaites glacier, West Antarctica. *Sci. Adv.* 5, eaau3433. doi:10.1126/sciadv.aau3433.
- 38 Millan, R., Mouginot, J., Rabatel, A., Jeong, S., Cusicanqui, D., Derkacheva, A., et al. (2019a). Mapping Surface Flow
39 Velocity of Glaciers at Regional Scale Using a Multiple Sensors Approach. *Remote Sens.* 11, 2498.
40 doi:10.3390/rs11212498.
- 41 Millan, R., Rignot, E., Rivera, A., Martineau, V., Mouginot, J., Zamora, R., et al. (2019b). Ice Thickness and Bed
42 Elevation of the Northern and Southern Patagonian Icefields. *Geophys. Res. Lett.* 46, 6626–6635.
43 doi:10.1029/2019GL082485.
- 44 Miller, G. H., Landvik, J. Y., Lehman, S. J., and Southon, J. R. (2017). Episodic Neoglacial snowline descent and
45 glacier expansion on Svalbard reconstructed from the 14C ages of ice-entombed plants. *Quat. Sci. Rev.* 155, 67–
46 78. doi:https://doi.org/10.1016/j.quascirev.2016.10.023.
- 47 Miller, K. G., Kopp, R. E., Horton, B. P., Browning, J. V., and Kemp, A. C. (2013). A geological perspective on sea-
48 level rise and its impacts along the U.S. mid-Atlantic coast. *Earth's Futur.* 1, 3–18. doi:10.1002/2013EF000135.
- 49 Miller, K. G., Wright, J. D., Browning, J. V., Kulpecz, A., Kominz, M., Naish, T. R., et al. (2012). High tide of the
50 warm Pliocene: Implications of global sea level for Antarctic deglaciation. *Geology* 40, 407–410.
51 doi:10.1130/G32869.1.
- 52 Mills, K. E., Pershing, A. J., Brown, C. J., Chen, Y., Chiang, F.-S., Holland, D. S., et al. (2014). Fisheries management
53 in a changing climate: Lessons from the 2012 ocean heat wave in the Northwest Atlantic. *Oceanography* 26, 191–
54 195. doi:10.5670/oceanog.2013.27.
- 55 Minowa, M., Sugiyama, S., Sakakibara, D., and Skvarca, P. (2017). Seasonal Variations in Ice-Front Position
56 Controlled by Frontal Ablation at Glacier Perito Moreno, the Southern Patagonia Icefield. *Front. Earth Sci.* 5.
57 doi:10.3389/feart.2017.00001.
- 58 Mitrovica, J. X., Hay, C. C., Kopp, R. E., Harig, C., and Latychev, K. (2018). Quantifying the sensitivity of sea level
59 change in coastal localities to the geometry of polar ice mass flux. *J. Clim.* 31. doi:10.1175/JCLI-D-17-0465.1.
- 60 Mitrovica, J. X., Tamisiea, M. E., Davis, J. L., and Milne, G. A. (2001). Recent mass balance of polar ice sheets
61 inferred from patterns of global sea-level change. *Nature*. doi:10.1038/35059054.

- Moffa-Sánchez, P., and Hall, I. R. (2017). North Atlantic variability and its links to European climate over the last 3000 years. *Nat. Commun.* 8, 1726. doi:10.1038/s41467-017-01884-8.
- Moftakhari, H. R., Salvadori, G., AghaKouchak, A., Sanders, B. F., and Matthew, R. A. (2017). Compounding effects of sea level rise and fluvial flooding. *Proc. Natl. Acad. Sci.* 114, 9785–9790. doi:10.1073/pnas.1620325114.
- Mohajerani, Y., Velicogna, I., and Rignot, E. (2018). Mass Loss of Totten and Moscow University Glaciers, East Antarctica, Using Regionally Optimized GRACE Mascons. *Geophys. Res. Lett.* 45, 7010–7018. doi:10.1029/2018GL078173.
- Mollaret, C., Hilbich, C., Pellet, C., Flores-Orozco, A., Delaloye, R., and Hauck, C. (2019). Mountain permafrost degradation documented through a network of permanent electrical resistivity tomography sites. *Cryosph.* 13, 2557–2578. doi:10.5194/tc-13-2557-2019.
- Monaghan, A. J., Bromwich, D. H., Fogt, R. L., Wang, S. H., Mayewski, P. A., Dixon, D. A., et al. (2006). Insignificant change in antarctic snowfall since the international geophysical year. *Science* 313, 827–831. doi:10.1126/science.1128243.
- Monselesan, D. P., O’Kane, T. J., Risbey, J. S., and Church, J. (2015). Internal climate memory in observations and models. *Geophys. Res. Lett.* doi:10.1002/2014GL062765.
- Moon, T., Joughin, I., and Smith, B. (2015). Seasonal to multiyear variability of glacier surface velocity, terminus position, and sea ice/ice mélange in northwest Greenland. *J. Geophys. Res. Earth Surf.* 120, 818–833. doi:10.1002/2015JF003494.
- Moore, G. W. K., and McNeil, K. (2018). The Early Collapse of the 2017 Lincoln Sea Ice Arch in Response to Anomalous Sea Ice and Wind Forcing. *Geophys. Res. Lett.* 45, 8343–8351. doi:10.1029/2018GL078428.
- Moorman, R., Morrison, A. K., and Hogg, A. M. (submitted). Positive and negative feedbacks to Antarctic ice shelf melt in an eddy rich global ocean–sea-ice model. (submitted).
- Mori, N., Kato, M., Kim, S., Mase, H., Shibutani, Y., Takemi, T., et al. (2014). Local amplification of storm surge by Super Typhoon Haiyan in Leyte Gulf. *Geophys. Res. Lett.* 41, 5106–5113. doi:10.1002/2014GL060689.
- Mori, N., Shimura, T., Yoshida, K., Mizuta, R., Okada, Y., Fujita, M., et al. (2019). Future changes in extreme storm surges based on mega-ensemble projection using 60-km resolution atmospheric global circulation model. *Coast. Eng. J.* 61, 295–307. doi:10.1080/21664250.2019.1586290.
- Morim, J., Hemer, M., Cartwright, N., Strauss, D., and Andutta, F. (2018). On the concordance of 21st century wind-wave climate projections. *Glob. Planet. Change.* doi:10.1016/j.gloplacha.2018.05.005.
- Morim, J., Hemer, M., Wang, X. L., Cartwright, N., Trenham, C., Semedo, A., et al. (2019). Robustness and uncertainties in multivariate wind-wave climate projections. *Nat. Clim. Chang.* 9, 711–718.
- Morlighem, M., Bondzio, J., Seroussi, H., Rignot, E., Larour, E., Humbert, A., et al. (2016a). Modeling of Store Gletscher’s calving dynamics, West Greenland, in response to ocean thermal forcing. *Geophys. Res. Lett.* 43, 2659–2666. doi:10.1002/2016GL067695.
- MORLIGHEM, M., RIGNOT, E., BINDER, T., BLANKENSHIP, D., DREWS, R., EAGLES, G., et al. Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the Antarctic ice sheet.
- Morlighem, M., Rignot, E., Mouginot, J., Seroussi, H., and Larour, E. (2014). Deeply incised submarine glacial valleys beneath the Greenland ice sheet. *Nat. Geosci.* 7, 418. Available at: <https://doi.org/10.1038/ngeo2167>.
- Morlighem, M., Rignot, E., and Willis, J. K. (2016b). Improving Bed Topography Mapping of Greenland Glaciers Using NASA’s Oceans Melting Greenland (OMG) Data. *Oceanography* 29, 62–71. doi:10.5670/oceanog.2016.99.
- Morlighem, M., Williams, C. N., Rignot, E., An, L., Arndt, J. E., Bamber, J. L., 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, 11,11-51,61. doi:10.1002/2017GL074954.
- 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. *Cryosph.* 13, 723–734. doi:10.5194/tc-13-723-2019.
- Morrison, A. K., England, M. H., and Hogg, A. M. (2015). Response of Southern Ocean Convection and Abyssal Overturning to Surface Buoyancy Perturbations. *J. Clim.* 28, 4263–4278. doi:10.1175/JCLI-D-14-00110.1.
- Morrison, A. K., Griffies, S. M., Winton, M., Anderson, W. G., and Sarmiento, J. L. (2016). Mechanisms of Southern Ocean heat uptake and transport in a global eddying climate model. *J. Clim.* 29, 2059–2075. doi:10.1175/JCLI-D-15-0579.1.
- Morrison, A. K., and Hogg, A. M. C. (2013). On the relationship between southern ocean overturning and ACC transport. *J. Phys. Oceanogr.* 43, 140–148. doi:10.1175/JPO-D-12-057.1.
- Morrow, R., and Kestenare, E. (2017). 22-year surface salinity changes in the Seasonal Ice Zone near 140°E off Antarctica. *J. Mar. Syst.* 175, 46–62. doi:https://doi.org/10.1016/j.jmarsys.2017.07.003.
- Mortimer, C., Mudryk, L., Derksen, C., Luoju, K., Brown, R., Kelly, R., et al. (submitted). Evaluation of long term Northern Hemisphere snow water equivalent products. *Cryosph. Discuss.* (submitted).
- Mortin, J., Svensson, G., Graversen, R. G., Kapsch, M. L., Stroeve, J. C., and Boisvert, L. N. (2016). Melt onset over Arctic sea ice controlled by atmospheric moisture transport. *Geophys. Res. Lett.* 43, 6636–6642. doi:10.1002/2016GL069330.

- Mosbeux, C., Gillet-Chaulet, F., and Gagliardini, O. (2016). Comparison of adjoint and nudging methods to initialise ice sheet model basal conditions. *Geosci. Model Dev.* 9, 2549–2562. doi:10.5194/gmd-9-2549-2016.
- Mouginot, J., Rignot, E., Björk, A. A., van den Broeke, M., Millan, R., Morlighem, M., et al. (2019). Forty-six years of Greenland Ice Sheet mass balance from 1972 to 2018. *Proc. Natl. Acad. Sci.* 116, 9239–9244. doi:10.1073/pnas.1904242116.
- Mouginot, J., Rignot, E., and Scheuchl, B. (2014). Sustained increase in ice discharge from the Amundsen Sea Embayment, West Antarctica, from 1973 to 2013. *Geophys. Res. Lett.* 41, 1576–1584. doi:10.1002/2013GL059069.
- Mouginot, J., Rignot, E., Scheuchl, B., Fenty, I., Khazendar, A., Morlighem, M., et al. (2015). Fast retreat of Zachariae Isstrøm, northeast Greenland. *Science* 350, 1357–1361. doi:10.1126/science.aac7111.
- Mouginot, J., Rignot, E., Scheuchl, B., and Millan, R. (2017). Comprehensive Annual Ice Sheet Velocity Mapping Using Landsat-8, Sentinel-1, and RADARSAT-2 Data. *Remote Sens.* 9. doi:10.3390/rs9040364.
- Msadek, R., and Frankignoul, C. (2009). Atlantic multidecadal oceanic variability and its influence on the atmosphere in a climate model. *Clim. Dyn.* 33, 45–62. doi:10.1007/s00382-008-0452-0.
- Muckenhuber, S., Nilsen, F., Korosov, A., and Sandven, S. (2016). Sea ice cover in Isfjorden and Hornsund, Svalbard (2000–2014) from remote sensing data. *Cryosphere*. doi:10.5194/tc-10-149-2016.
- Mudryk, L. R., Derksen, C., Howell, S., Laliberté, F., Thackeray, C., Sospedra-Alfonso, R., et al. (2018). Canadian snow and sea ice: historical trends and projections. *Cryosph.* 12, 1157–1176. doi:10.5194/tc-12-1157-2018.
- Mudryk, L. R., Kushner, P. J., and Derksen, C. (2014). Interpreting observed northern hemisphere snow trends with large ensembles of climate simulations. *Clim. Dyn.* 43, 345–359. doi:10.1007/s00382-013-1954-y.
- Mudryk, L. R., Kushner, P. J., Derksen, C., and Thackeray, C. (2017). Snow cover response to temperature in observational and climate model ensembles. *Geophys. Res. Lett.* 44, 919–926. doi:10.1002/2016GL071789.
- Mudryk, L., Santolaria Otin, M., Krinner, G., Derksen, C., Brutel-Vuilmet, C., Menegoz, M., et al. (submitted). Historical snow cover and projected changes in the CMIP6 multi-model ensemble. *Cryosph. Discuss.* (submitted).
- Muilwijk, M., Smedsrud, L. H., Ilıcak, M., and Drange, H. (2018). Atlantic Water Heat Transport Variability in the 20th Century Arctic Ocean From a Global Ocean Model and Observations. *J. Geophys. Res. Ocean.* 123, 8159–8179. doi:10.1029/2018JC014327.
- Muis, S., Verlaan, M., Winsemius, H. C., Aerts, J. C. J. H., and Ward, P. J. (2016). A global reanalysis of storm surges and extreme sea levels. *Nat. Commun.* 7, 11969. doi:10.1038/ncomms11969.
- Müller, M., Arbic, B. K., and Mitrovica, J. X. (2011). Secular trends in ocean tides: Observations and model results. *J. Geophys. Res. Ocean.* doi:10.1029/2010JC006387.
- Munday, D. R., Johnson, H. L., and Marshall, D. P. (2013). Eddy Saturation of Equilibrated Circumpolar Currents. *J. Phys. Oceanogr.* 43, 507–532. doi:10.1175/JPO-D-12-095.1.
- Muntjewerf, L., Petrini, M., Vizcaino, M., Ernani da Silva, C., and Sellevold, R., Scherrenberg, M. D. W. (submitteda). Greenland ice sheet contribution to 21st century sea level rise as simulated by the coupled cesm2.1-cism2.1. *Geophys. Res. Lett.* (submitted).
- Muntjewerf, L., Sellevold, R., Vizcaino, M., Ernani da Silva, C., Petrini, M., and Thayer-Calder, K. (submitted, b). Accelerated greenland ice sheet mass loss under high greenhouse gas forcing as simulated by the coupled cesm2.1-cism2.1. *J. Adv. Model. Earth Syst.* (submitted).
- Murari, M. K., Owen, L., Dortch, J., Caffee, M., Dietsch, C., Fuchs, M., et al. (2014). *Timing and climatic drivers for glaciation across monsoon-influenced regions of the Himalayan–Tibetan orogen.* doi:10.1016/j.quascirev.2014.01.013.
- Muresan, I. S., Khan, S. A., Aschwanden, A., Khroulev, C., Van Dam, T., Bamber, J., et al. (2016). Modelled glacier dynamics over the last quarter of a century at Jakobshavn Isbræ. *Cryosph.* 10, 597–611. doi:10.5194/tc-10-597-2016.
- Musilova, M., Tranter, M., Bamber, J. L., Takeuchi, N., and Anesio, A. (2016). Experimental evidence that microbial activity lowers the albedo of glaciers. *Geochemical Perspect. Lett.* 2, 106–116. doi:10.7185/geochemlet.1611.
- Najafi, M. R., Zwiers, F. W., and Gillett, N. P. (2015). Attribution of Arctic temperature change to greenhouse-gas and aerosol influences. *Nat. Clim. Chang.* 5. doi:10.1038/nclimate2524.
- Nakanowatari, T., Sato, K., and Inoue, J. (2014). Predictability of the barents sea ice in early winter: Remote effects of oceanic and atmospheric thermal conditions from the North Atlantic. *J. Clim.* 27, 8884–8901. doi:10.1175/JCLI-D-14-00125.1.
- Nauels, A., Gütschow, J., Mengel, M., Meinshausen, M., Clark, P. U., and Schleussner, C.-F. (2019). Attributing long-term sea-level rise to Paris Agreement emission pledges. *Proc. Natl. Acad. Sci.* 116, 23487–23492.
- Nauels, A., Meinshausen, M., Mengel, M., Lorbacher, K., and Wigley, T. M. L. (2017a). Synthesizing long-term sea level rise projections -- the MAGICC sea level model v2.0. *Geosci. Model Dev.* 10, 2495–2524. doi:10.5194/gmd-10-2495-2017.
- Nauels, A., Rogelj, J., Schleussner, C.-F., Meinshausen, M., and Mengel, M. (2017b). Linking sea level rise and socioeconomic indicators under the Shared Socioeconomic Pathways. *Environ. Res. Lett.* 12, 114002. doi:10.1088/1748-9326/aa92b6.
- Naughten, K. A., Jenkins, A., Holland, P. R., Mugford, R. I., Nicholls, K. W., and Munday, D. R. (2019). Modeling the

- Influence of the Weddell Polynya on the Filchner–Ronne Ice Shelf Cavity. *J. Clim.* 32, 5289–5303. doi:10.1175/JCLI-D-19-0203.1.
- Naughten, K. A., Meissner, K. J., Galton-Fenzi, B. K., England, M. H., Timmermann, R., Hellmer, H. H., et al. (2018). Future Projections of Antarctic Ice Shelf Melting Based on CMIP5 Scenarios. *J. Clim.* 31, 5243–5261. doi:10.1175/JCLI-D-17-0854.1.
- Naveira Garabato, A. C., Polzin, K. L., Ferrari, R., Zika, J. D., and Forryan, A. (2016). A Microscale View of Mixing and Overturning across the Antarctic Circumpolar Current. *J. Phys. Oceanogr.* 46, 233–254. doi:10.1175/JPO-D-15-0025.1.
- Naveira Garabato, A. C., Williams, A. P., and Bacon, S. (2014). The three-dimensional overturning circulation of the Southern Ocean during the WOCE era. *Prog. Oceanogr.* 120, 41–78. doi:https://doi.org/10.1016/j.pocean.2013.07.018.
- Ndoye, S., Capet, X., Estrade, P., Sow, B., Machu, E., Brochier, T., et al. (2017). Dynamics of a “low-enrichment high-retention” upwelling center over the southern Senegal shelf. *Geophys. Res. Lett.* 44, 5034–5043.
- Nerem, R. S., Ablain, M., Cazenave, A., Church, J., Leuliette, E., Ablain, M., et al. (2017). “A 25-Year Satellite Altimetry-Based Global Mean Sea Level Record,” in *Satellite Altimetry over Oceans and Land Surfaces* (Boca Raton, FL : Taylor & Francis, 2017.: CRC Press), 187–210. doi:10.1201/9781315151779-5.
- 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.* 115, 201717312. doi:10.1073/pnas.1717312115.
- Nesje, A., Pilø, L. H., Finstad, E., Solli, B., Wangen, V., Ødegård, R. S., et al. (2012). The climatic significance of artefacts related to prehistoric reindeer hunting exposed at melting ice patches in southern Norway. *The Holocene* 22, 485–496. doi:10.1177/0959683611425552.
- Newsom, E. R., Bitz, C. M., Bryan, F. O., Abernathy, R., and Gent, P. R. (2016). Southern Ocean Deep Circulation and Heat Uptake in a High-Resolution Climate Model. *J. Clim.* 29, 2597–2619. doi:10.1175/JCLI-D-15-0513.1.
- Nghiem, S. V., Hall, D. K., Mote, T. L., Tedesco, M., Albert, M. R., Keegan, K., et al. (2012). The extreme melt across the Greenland ice sheet in 2012. *Geophys. Res. Lett.* 39. doi:10.1029/2012GL053611.
- Nghiem, S. V., Rigor, I. G., Clemente-Colón, P., Neumann, G., and Li, P. P. (2016). Geophysical constraints on the Antarctic sea ice cover. *Remote Sens. Environ.* 181, 281–292. doi:10.1016/j.rse.2016.04.005.
- Nias, I. J., Cornford, S. L., and Payne, A. J. (2016). Contrasting the modelled sensitivity of the Amundsen Sea Embayment ice streams. *J. Glaciol.* 62, 552–562. doi:10.1017/jog.2016.40.
- Nick, F. M., Vieli, A., Andersen, M. L., Joughin, I., Payne, A., Edwards, T. L., et al. (2013). Future sea-level rise from Greenland’s main outlet glaciers in a warming climate. *Nature*. doi:10.1038/nature12068.
- Nicolas, J. P., and Bromwich, D. H. (2014). New reconstruction of antarctic near-surface temperatures: Multidecadal trends and reliability of global reanalyses. *J. Clim.* 27, 8070–8093. doi:10.1175/JCLI-D-13-00733.1.
- Nicolaus, M., Tian-Kunze, X., Lique, C., Krumpen, T., Hendricks, S., Ricker, R., et al. (2017). Satellite-observed drop of Arctic sea ice growth in winter 2015–2016. *Geophys. Res. Lett.* 44, 3236–3245. doi:10.1002/2016gl072244.
- Niederrenk, A. L., and Notz, D. (2018). Arctic Sea Ice in a 1.5°C Warmer World. *Geophys. Res. Lett.* 45, 1963–1971. doi:10.1002/2017GL076159.
- Nielsen, L. T., Aðalgeirsdóttir, G., Gkinis, V., Nuterman, R., and Hvidberg, C. S. (2018). The effect of a Holocene climatic optimum on the evolution of the Greenland ice sheet during the last 10 kyr. *J. Glaciol.* 64, 477–488. doi:10.1017/jog.2018.40.
- Nienow, P. W., Sole, A. J., Slater, D. A., and Cowton, T. R. (2017). Recent Advances in Our Understanding of the Role of Meltwater in the Greenland Ice Sheet System. *Curr. Clim. Chang. Reports* 3, 330–344. doi:10.1007/s40641-017-0083-9.
- Nieves, V., Willis, J. K., and Patzert, W. C. (2015). Recent hiatus caused by decadal shift in Indo-Pacific heating. *Science* 349, 532–535. doi:10.1126/science.aaa4521.
- Nilsson, J., Vallenga, P., Simonsen, S. B., Sørensen, L. S., Forsberg, R., Dahl-Jensen, D., et al. (2015). Greenland 2012 melt event effects on CryoSat-2 radar altimetry. *Geophys. Res. Lett.* 42, 3919–3926. doi:10.1002/2015GL063296.
- Noël, B., van de Berg, W. J., Lhermitte, S., Wouters, B., Machguth, H., Howat, I., et al. (2017). A tipping point in refreezing accelerates mass loss of Greenland’s glaciers and ice caps. *Nat. Commun.* 8, 14730.
- Noël, B., van de Berg, W. J., Machguth, H., Lhermitte, S., Howat, I., Fettweis, X., et al. (2016). A daily, 1km resolution data set of downscaled Greenland ice sheet surface mass balance (1958–2015). *Cryosph.* 10, 2361–2377. doi:10.5194/tc-10-2361-2016.
- Noël, B., Van De Berg, W. J., Van Meijgaard, E., Kuipers Munneke, P., Van De Wal, R. S. W., and Van Den Broeke, M. R. (2015). Evaluation of the updated regional climate model RACMO2. 3: summer snowfall impact on the Greenland Ice Sheet. *Cryosph.* 9, 1831–1844.
- Noetzi, J., Biskaborn, B. K., Christiansen, H. H., Isaksen, K., Schoeneich, P., Smith, S., et al. (2019). Permafrost thermal state. *Bull. Am. Meteorol. Soc.* 100, S21–S22. doi:https://doi.org/10.1175/2019BAMSSStateoftheClimate.1.
- Notz, D. (2015). How well must climate models agree with observations? *Philos. Trans. R. Soc. A Math. Phys. Eng.*

- Sci.* 373. doi:10.1098/rsta.2014.0164.
- Notz, D., and Marotzke, J. (2012). Observations reveal external driver for Arctic sea-ice retreat. *Geophys. Res. Lett.* 39. doi:10.1029/2012GL051094.
- Notz, D., and Stroeve, J. (2016). Observed Arctic sea-ice loss directly follows anthropogenic CO₂ emission. *Science* 354, 747–750. doi:10.1126/science.aag2345.
- Notz, D., and Stroeve, J. (2018). The Trajectory Towards a Seasonally Ice-Free Arctic Ocean. *Curr. Clim. Chang. Reports* 4, 407–416. doi:10.1007/s40641-018-0113-2.
- Nowicki, S., Bindschadler, R. A., Abe-Ouchi, A., Aschwanden, A., Bueler, E., Choi, H., et al. (2013). Insights into spatial sensitivities of ice mass response to environmental change from the SeaRISE ice sheet modeling project I: Antarctica. *J. Geophys. Res. Earth Surf.* 118, 1002–1024. doi:10.1002/jgrf.20076.
- Nowicki, S. M. J., Payne, T., Larour, E., Seroussi, H., Goelzer, H., Lipscomb, W., et al. (2016). Ice Sheet Model Intercomparison Project (ISMIP6) contribution to CMIP6. *Geosci. Model Dev. Discuss.* 9, 1–42. doi:10.5194/gmd-2016-105.
- Nowicki, S., Payne, A. J., Abe-ouchi, A., Alexander, P., Albrecht, T., Asay-davis, X., et al. (submitteda). Contrasting contributions to future sea level under CMIP5 and CMIP6 scenarios from the Greenland and Antarctic ice sheets. *Geophys. Res. Lett.* (submitted), 1–12.
- Nowicki, S., Payne, A. J., Goelzer, H., Seroussi, H., William H. Lipscomb, A. A., and Abe-Ouchi Cecile, et al. (submitted, b). Experimental protocol for sea level projections from ISMIP6 standalone ice sheet models. *Cryosph.* (submitted). doi:10.1029/2005JE002414.
- Nummelin, A., Li, C., and Hezel, P. J. (2017). Connecting ocean heat transport changes from the midlatitudes to the Arctic Ocean. *Geophys. Res. Lett.* 44, 1899–1908. doi:10.1002/2016GL071333.
- Nussbaumer, S. U., and Zumbühl, H. J. (2012). The Little Ice Age history of the Glacier des Bossons (Mont Blanc massif, France): a new high-resolution glacier length curve based on historical documents. *Clim. Change* 111, 301–334. doi:10.1007/s10584-011-0130-9.
- O’Brien, C. L., Foster, G. L., Martínez-Botí, M. A., Abell, R., Rae, J. W. B., and Pancost, R. D. (2014). High sea surface temperatures in tropical warm pools during the Pliocene. *Nat. Geosci.* 7, 606–611. doi:10.1038/ngeo2194.
- O’Leary, M., and Christoffersen, P. (2013). Calving on tidewater glaciers amplified by submarine frontal melting. *Cryosphere* 7, 119–128. doi:10.5194/tc-7-119-2013.
- O’Neill, B. C., Kriegler, E., Ebi, K. L., Kemp-Benedict, E., Riahi, K., Rothman, D. S., et al. (2017). The roads ahead: narratives for shared socioeconomic pathways describing world futures in the 21st century. *Glob. Environ. Chang.* 42, 169–180.
- Obu, J., Westermann, S., Bartsch, A., Berdnikov, N., Christiansen, H. H., Dashtseren, A., et al. (2019). Northern Hemisphere permafrost map based on TTOP modelling for 2000–2016 at 1 km² scale. *Earth-Science Rev.* 193, 299–316. doi:10.1016/j.earscirev.2019.04.023.
- Obu, J., Westermann, S., Kääh, A., and Bartsch, A. (2018). Ground Temperature Map, 2000–2016, Northern Hemisphere Permafrost. doi:10.1594/PANGAEA.888600.
- Oerder, V., Colas, F., Echevin, V., Codron, F., Tam, J., and Belmadani, A. (2015a). Peru-Chile upwelling dynamics under climate change. *J. Geophys. Res. Ocean.* 120, 1152–1172.
- Oerder, V., Colas, F., Echevin, V., Codron, F., Tam, J., and Belmadani, A. (2015b). Peru-Chile upwelling dynamics under climate change. *J. Geophys. Res. Ocean.* 120, 1152–1172. doi:10.1002/2014JC010299.
- Oerder, V., Colas, F., Echevin, V., Masson, S., and Lemarié, F. (2018). Impacts of the Mesoscale Ocean-Atmosphere Coupling on the Peru-Chile Ocean Dynamics: The Current-Induced Wind Stress Modulation. *J. Geophys. Res. Ocean.* 123, 812–833. doi:10.1002/2017JC013294.
- Oka, E., Katsura, S., Inoue, H., Kojima, A., Kitamoto, M., Nakano, T., et al. (2017). Long-term change and variation of salinity in the western North Pacific subtropical gyre revealed by 50-year long observations along 137°E. *J. Oceanogr.* 73, 479–490. doi:10.1007/s10872-017-0416-2.
- Olason, E., and Notz, D. (2014). Drivers of variability in Arctic sea-ice drift speed. *J. Geophys. Res. C Ocean.* 119, 5755–5775. doi:10.1002/2014JC009897.
- Oldenburg, D., Armour, K. C., Thompson, L. A., and Bitz, C. M. (2018). Distinct Mechanisms of Ocean Heat Transport Into the Arctic Under Internal Variability and Climate Change. *Geophys. Res. Lett.* 45, 7692–7700. doi:10.1029/2018GL078719.
- Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., et al. (2016). Circumpolar distribution and carbon storage of thermokarst landscapes. *Nat. Commun.* doi:10.1038/ncomms13043.
- Oliver, E. C. J. (2019). Mean warming not variability drives marine heatwave trends. *Clim. Dyn.* 53, 1653–1659. doi:10.1007/s00382-019-04707-2.
- Oliver, E. C. J., Donat, M. G., Burrows, M. T., Moore, P. J., Smale, D. A., Alexander, L. V., et al. (2018). Longer and more frequent marine heatwaves over the past century. *Nat. Commun.* 9, 1–12. doi:10.1038/s41467-018-03732-9.
- Olonscheck, D., and Notz, D. (2017). Consistently estimating internal climate variability from climate model simulations. *J. Clim.* 30, 9555–9573. doi:10.1175/JCLI-D-16-0428.1.
- Onarheim, I. H., Eldevik, T., Smedsrud, L. H., and Stroeve, J. C. (2018). Seasonal and regional manifestation of Arctic sea ice loss. *J. Clim.*, JCLI-D-17-0427.1. doi:10.1175/JCLI-D-17-0427.1.

- Oppenheimer, M., Glavovic, B., and et al. (2019). “Chapter 4: Sea Level Rise and Implications for Low Lying Islands, Coasts and Communities,” in *IPCC SR Ocean and Cryosphere*.
- Orton, P., Georgas, N., Blumberg, A., and Pullen, J. (2012). Detailed modeling of recent severe storm tides in estuaries of the New York City region. *J. Geophys. Res. Ocean.* 117.
- Orton, P. M., Conticello, F. R., Cioffi, F., Hall, T. M., Georgas, N., Lall, U., et al. (2018). Flood hazard assessment from storm tides, rain and sea level rise for a tidal river estuary. *Nat. Hazards*, 1–29. doi:10.1007/s11069-018-3251-x.
- Otto-Bliesner, B. L., Jahn, A., Feng, R., Brady, E. C., Hu, A., and Löfverström, M. (2017). Amplified North Atlantic warming in the late Pliocene by changes in Arctic gateways. *Geophys. Res. Lett.* 44, 957–964. doi:10.1002/2016GL071805.
- Overly, T. B., Hawley, R. L., Helm, V., Morris, E. M., and Chaudhary, R. N. (2016). Greenland annual accumulation along the EGIS line, 1959–2004, from ASIRAS airborne radar and neutron-probe density measurements. *Cryosph.* 10, 1679–1694. doi:10.5194/tc-10-1679-2016.
- Owen, L., and Dortch, J. (2014). *Nature and timing of Quaternary glaciation in the Himalayan–Tibetan orogen*. doi:10.1016/j.quascirev.2013.11.016.
- Oyarzún, D., and Brierley, C. M. (2018). The future of coastal upwelling in the Humboldt current from model projections. *Clim. Dyn.*, 1–17.
- Palanisamy, H., Meyssignac, B., Cazenave, A., and Delcroix, T. (2015). Is anthropogenic sea level fingerprint already detectable in the Pacific Ocean? *Environ. Res. Lett.* 10, 084024. doi:10.1088/1748-9326/10/8/084024.
- Palerme, C., Claud, C., Dufour, A., Genthon, C., Wood, N. B., and L’Ecuyer, T. (2017a). Evaluation of Antarctic snowfall in global meteorological reanalyses. *Atmos. Res.* 190, 104–112. doi:10.1016/j.atmosres.2017.02.015.
- Palerme, C., Genthon, C., Claud, C., Kay, J. E., Wood, N. B., and L’Ecuyer, T. (2017b). Evaluation of current and projected Antarctic precipitation in CMIP5 models. *Clim. Dyn.* 48, 225–239. doi:10.1007/s00382-016-3071-1.
- Palerme, C., Kay, J. E., Genthon, C., L’Ecuyer, T., Wood, N. B., and Claud, C. (2014). How much snow falls on the Antarctic ice sheet? *Cryosphere* 8, 1577–1587. doi:10.5194/tc-8-1577-2014.
- Palmer, M. D. (2017). Reconciling Estimates of Ocean Heating and Earth’s Radiation Budget. *Curr. Clim. Chang. Reports* 3, 78–86. doi:10.1007/s40641-016-0053-7.
- Palmer, M. D., Gregory, J. M., Bagge, M., Calvert, D., Hagedoorn, J. M., Howard, T., et al. Exploring the Drivers of Global and Regional Sea-Level Change over the 21st Century and Beyond.
- Palmer, M. D., Gregory, J. M., Bagge, M., Calvert, D., Hagedoorn, J. M., Howard, T., et al. (submitted). Exploring the Drivers of Global and Regional Sea - Level Change over the 21st Century and Beyond. *Earths Futur.* (submitted).
- Palmer, M. D., Harris, G. R., and Gregory, J. M. (2018). Extending CMIP5 projections of global mean temperature change and sea level rise due to thermal expansion using a physically-based emulator. *Environ. Res. Lett.* 13, 84003.
- Palmer, M. D., and McNeall, D. J. (2014). Internal variability of Earth’s energy budget simulated by CMIP5 climate models. *Environ. Res. Lett.* 9, 34016.
- Palmer, M. D., Roberts, C. D., Balmaseda, M., Chang, Y. S., Chepurin, G., Ferry, N., et al. (2017). Ocean heat content variability and change in an ensemble of ocean reanalyses. *Clim. Dyn.* 49, 909–930. doi:10.1007/s00382-015-2801-0.
- Paolo, F. S., Fricker, H. A., and Padman, L. (2015). Volume loss from Antarctic ice shelves is accelerating. *Science* 348, 327–331. doi:10.1126/science.aaa0940.
- Paquin, J. P., and Sushama, L. (2014). On the Arctic near-surface permafrost and climate sensitivities to soil and snow model formulations in climate models. *Clim. Dyn.* doi:10.1007/s00382-014-2185-6.
- Pardaens, A. K., Gregory, J. M., and Lowe, J. A. (2011). A model study of factors influencing projected changes in regional sea level over the twenty-first century. *Clim. Dyn.* 36, 2015–2033. doi:10.1007/s00382-009-0738-x.
- Park, H. S., Lee, S., Son, S. W., Feldstein, S. B., and Kosaka, Y. (2015). The impact of poleward moisture and sensible heat flux on arctic winter sea ice variability. *J. Clim.* 28, 5030–5040. doi:10.1175/JCLI-D-15-0074.1.
- Park, W., and Latif, M. (2018). Ensemble global warming simulations with idealized Antarctic meltwater input. *Clim. Dyn.* doi:10.1007/s00382-018-4319-8.
- Parkes, D., and Goosse, H. (2019). Modelling regional glacier length changes over the last millennium using the Open Global Glacier Model. *Cryosph. Discuss.* 2019, 1–19. doi:10.5194/tc-2019-275.
- Parkes, D., and Marzeion, B. (2018). Twentieth-century contribution to sea-level rise from uncharted glaciers. *Nature* 563, 551–554. doi:10.1038/s41586-018-0687-9.
- Parkinson, C. L., and Cavalieri, D. J. (2012). Antarctic sea ice variability and trends, 1979–2010. *Cryosphere* 6, 871–880. doi:10.5194/tc-6-871-2012.
- Patara, L., and Böning, C. W. (2014). Abyssal ocean warming around Antarctica strengthens the Atlantic overturning circulation. *Geophys. Res. Lett.* 41, 3972–3978. doi:10.1002/2014GL059923.
- Patterson, M. O., McKay, R., Naish, T., ... C. E.-N., 2014, undefined, Escutia, C., et al. (2014). Orbital forcing of the East Antarctic ice sheet during the Pliocene and Early Pleistocene. *Nat. Geosci* 7, 841–847. doi:10.1038/ngeo2273.
- Pattyn, F. (2017). Sea-level response to melting of Antarctic ice shelves on multi-centennial timescales with the fast

- Elementary Thermomechanical Ice Sheet model (f.ETISh v1.0). *Cryosph.* 11, 1851–1878. doi:10.5194/tc-11-1851-2017.
- Pattyn, F., and Durand, G. (2013). Why marine ice sheet model predictions may diverge in estimating future sea level rise. *Geophys. Res. Lett.* 40, 4316–4320. doi:10.1002/grl.50824.
- Pattyn, F., Favier, L., Sun, S., and Durand, G. (2017). Progress in Numerical Modeling of Antarctic Ice-Sheet Dynamics. *Curr. Clim. Chang. Reports* 3, 174–184. doi:10.1007/s40641-017-0069-7.
- Pattyn, F., Ritz, C., Hanna, E., Asay-Davis, X., DeConto, R., Durand, G., et al. (2018). The Greenland and Antarctic ice sheets under 1.5 °C global warming. *Nat. Clim. Chang.* doi:10.1038/s41558-018-0305-8.
- Pattyn, F., Schoof, C., Perichon, L., Hindmarsh, R. C. A., Bueler, E., de Fleurian, B., et al. (2012). Results of the Marine Ice Sheet Model Intercomparison Project, MISIP. *Cryosph.* 6, 573–588. doi:10.5194/tc-6-573-2012.
- Pauling, A. G., Smith, I. J., Langhorne, P. J., and Bitz, C. M. (2017). Time-Dependent Freshwater Input From Ice Shelves: Impacts on Antarctic Sea Ice and the Southern Ocean in an Earth System Model. *Geophys. Res. Lett.* 44, 10,454–10,461. doi:10.1002/2017GL075017.
- Pavlova, O., Gerland, S., and Hop, H. (2019). “Changes in Sea-Ice Extent and Thickness in Kongsfjorden, Svalbard (2003–2016),” in, 105–136. doi:10.1007/978-3-319-46425-1_4.
- Pederson, G. T., Gray, S. T., Woodhouse, C. A., Betancourt, J. L., Fagre, D. B., Littell, J. S., et al. (2011). The unusual nature of recent snowpack declines in the North American Cordillera. *Science* 333, 332–335. doi:10.1126/science.1201570.
- Pedro, J. B., Jochum, M., Buizert, C., He, F., Barker, S., and Rasmussen, S. O. (2018). Beyond the bipolar seesaw: Toward a process understanding of interhemispheric coupling. *Quat. Sci. Rev.* 192, 27–46. doi:10.1016/J.QUASCIREV.2018.05.005.
- Pegler, S. S. (2018). Marine ice sheet dynamics: The impacts of ice-shelf buttressing. *J. Fluid Mech.* 857, 605–647. doi:10.1017/jfm.2018.741.
- Pelle, T., Morlighem, M., and Bondzio, J. (2019). Brief communication: PICOP, a new ocean melt parameterization under ice shelves combining PICO and a plume model. *Cryosphere* 13, 1043–1049. doi:10.5194/tc-13-1043-2019.
- Pelling, H. E., Mattias Green, J. A., and Ward, S. L. (2013). Modelling tides and sea-level rise: To flood or not to flood. *Ocean Model.* 63, 21–29. doi:10.1016/J.OCEMOD.2012.12.004.
- Peltier, W. R., Argus, D. F., and Drummond, R. (2015). Space geodesy constrains ice age terminal deglaciation: The global ICE-6G_C (VM5a) model. *J. Geophys. Res. Solid Earth* 120, 450–487.
- Pendleton, S. L., Miller, G. H., Lifton, N., Lehman, S. J., Southon, J., Crump, S. E., et al. (2019). Rapidly receding Arctic Canada glaciers revealing landscapes continuously ice-covered for more than 40,000 years. *Nat. Commun.* 10, 445. doi:10.1038/s41467-019-08307-w.
- Peng, G., and Meier, W. N. (2018). Temporal and regional variability of Arctic sea-ice coverage from satellite data. *Ann. Glaciol.* 59, 191–200. doi:10.1017/aog.2017.32.
- Perego, M., Price, S., and Stadler, G. (2014). Optimal initial conditions for coupling ice sheet models to Earth system models. *J. Geophys. Res. Surf.* 119, 1894–1917. doi:10.1002/2014jfr003181.
- Perket, J., Flanner, M. G., and Kay, J. E. (2014). Diagnosing shortwave cryosphere radiative effect and its 21st century evolution in CESM. *J. Geophys. Res.* 119, 1356–1362. doi:10.1002/2013JD021139.
- Perkins-Kirkpatrick, S. E., King, A. D., Cougnon, E. A., Grose, M. R., Oliver, E. C. J., Holbrook, N. J., et al. (2018). The role of natural variability and anthropogenic climate change in the 2017/2018 Tasman Sea marine heatwave. *Bull. Am. Meteorol. Soc.*, 1–4. doi:10.1175/BAMS-D-18-0116.1.
- PERMOS (2019). Permafrost in Switzerland 2014/2015 to 2017/2018. , eds. J. Noetzli, C. Pellet, and B. Staub Cryospheric Commission of the Swiss Academy of Sciences doi:10.13093/permos-rep-2019-16-19.
- Petty, A. A., Holland, M. M., Bailey, D. A., and Kurtz, N. T. (2018). Warm Arctic, Increased Winter Sea Ice Growth? *Geophys. Res. Lett.* 45, 12,922–12,930. doi:10.1029/2018GL079223.
- Pickering, M. D. D., Horsburgh, K. J. J., Blundell, J. R. R., Hirschi, J. J.-M. J. M., Nicholls, R. J. J., Verlaan, M., et al. (2017). The impact of future sea-level rise on the global tides. *Cont. Shelf Res.* 142, 50–68. doi:10.1016/j.csr.2017.02.004.
- Pickering, M. D., Wells, N. C., Horsburgh, K. J., and Green, J. A. M. (2012). The impact of future sea-level rise on the European Shelf tides. *Cont. Shelf Res.* doi:10.1016/j.csr.2011.11.011.
- Pilo, G. S., Holbrook, N. J., Kiss, A. E., and Hogg, A. M. (2019). Sensitivity of Marine Heatwave Metrics to Ocean Model Resolution. *Geophys. Res. Lett.* 46. doi:10.1029/2019GL084928.
- Pollard, D., and DeConto, R. M. (2012). Description of a hybrid ice sheet-shelf model, and application to Antarctica. *Geosci. Model Dev. Discuss.* 5, 1077–1134.
- Pollard, D., DeConto, R. M., and Alley, R. B. (2015). Potential Antarctic Ice Sheet retreat driven by hydrofracturing and ice cliff failure. *Earth Planet. Sci. Lett.* 412, 112–121. doi:10.1016/j.epsl.2014.12.035.
- Pollard, D., Gomez, N., and Deconto, R. M. (2017). Variations of the Antarctic Ice Sheet in a Coupled Ice Sheet-Earth-Sea Level Model: Sensitivity to Viscoelastic Earth Properties. *J. Geophys. Res. Earth Surf.* 11, 2124–2138. doi:10.1002/2017JF004371.
- Polyakov, I. V., Pnyushkov, A. V., Alkire, M. B., Ashik, I. M., Baumann, T. M., Carmack, E. C., et al. (2017). Greater

- role for Atlantic inflows on sea-ice loss in the Eurasian Basin of the Arctic Ocean. *Science* 356, 285–291. doi:10.1126/science.aai8204.
- Pope, J. O., Holland, P. R., Orr, A., Marshall, G. J., and Phillips, T. (2017). The impacts of El Niño on the observed sea ice budget of West Antarctica. *Geophys. Res. Lett.* 44, 6200–6208. doi:10.1002/2017GL073414.
- Portela, E., Kolodziejczyk, N., Maes, C., and Thierry, V. (2019). Interior water-mass variability in the Southern-Hemisphere oceans during the last decade. *J. Phys. Oceanogr.* 0, null. doi:10.1175/JPO-D-19-0128.1.
- Porter, S. E., Parkinson, C. L., and Mosley-Thompson, E. (2016). Bellingshausen Sea ice extent recorded in an Antarctic Peninsula ice core. *J. Geophys. Res. Atmos.* 121, 13,813–886,900. doi:10.1002/2016JD025626.
- Portnov, A., Smith, A. J., Mienert, J., Cherkashov, G., Rekant, P., Semenov, P., et al. (2013). Offshore permafrost decay and massive seabed methane escape in water depths >20 m at the South Kara Sea shelf. *Geophys. Res. Lett.* 40, 3962–3967. doi:10.1002/grl.50735.
- Poulsen, M. B., Jochum, M., and Nuterman, R. (2018). Parameterized and resolved Southern Ocean eddy compensation. *Ocean Model.* 124, 1–15. doi:https://doi.org/10.1016/j.ocemod.2018.01.008.
- Pratap, B., Dobhal, D. P., Mehta, M., and Bhambri, R. (2015). Influence of debris cover and altitude on glacier surface melting: A case study on Dokriani Glacier, central Himalaya, India. *Ann. Glaciol.* 56, 9–16. doi:10.3189/2015AoG70A971.
- Previdi, M., and Polvani, L. M. (2016). Anthropogenic impact on Antarctic surface mass balance, currently masked by natural variability, to emerge by mid-century. *Environ. Res. Lett.* 11, 094001. doi:10.1088/1748-9326/11/9/094001.
- Pritchard, H. D. (2019). Asia's shrinking glaciers protect large populations from drought stress. *Nature* 569, 649–654. doi:10.1038/s41586-019-1240-1.
- Pritchard, H. D., Ligtenberg, S. R. M., Fricker, H. A., Vaughan, D. G., Van Den Broeke, M. R., and Padman, L. (2012). Antarctic ice-sheet loss driven by basal melting of ice shelves. *Nature* 484, 502–505. doi:10.1038/nature10968.
- Punge, H. J., Gallée, H., Kageyama, M., and Krinner, G. (2012). Modelling snow accumulation on Greenland in Eemian, glacial inception, and modern climates in a GCM. *Clim. Past* 8, 1801–1819. doi:10.5194/cp-8-1801-2012.
- Purich, A., Cai, W., England, M. H., and Cowan, T. (2016a). Evidence for link between modelled trends in Antarctic sea ice and underestimated westerly wind changes. *Nat. Commun.* 7, 10409. doi:10.1038/ncomms10409.
- Purich, A., England, M. H., Cai, W., Chikamoto, Y., Timmermann, A., Fyfe, J. C., et al. (2016b). Tropical pacific SST drivers of recent antarctic sea ice trends. *J. Clim.* 29, 8931–8948. doi:10.1175/JCLI-D-16-0440.1.
- Purich, A., England, M. H., Cai, W., Sullivan, A., and Durack, P. J. (2018). Impacts of broad-scale surface freshening of the Southern Ocean in a coupled climate model. *J. Clim.* 31, 2613–2632. doi:10.1175/JCLI-D-17-0092.1.
- Purkey, S. G., and Johnson, G. C. (2010). Warming of Global Abyssal and Deep Southern Ocean Waters between the 1990s and 2000s: Contributions to Global Heat and Sea Level Rise Budgets. *J. Clim.* 23, 6336–6351. doi:10.1175/2010JCLI3682.1.
- Purkey, S. G., and Johnson, G. C. (2012). Global Contraction of Antarctic Bottom Water between the 1980s and 2000s. *J. Clim.* 25, 5830–5844. doi:10.1175/JCLI-D-11-00612.1.
- Purkey, S. G., and Johnson, G. C. (2013). Antarctic Bottom Water Warming and Freshening: Contributions to Sea Level Rise, Ocean Freshwater Budgets, and Global Heat Gain. *J. Clim.* 26, 6105–6122. doi:10.1175/JCLI-D-12-00834.1.
- Purkey, S. G., Johnson, G. C., Talley, L. D., Sloyan, B. M., Wijffels, S. E., Smethie, W., et al. (2019). Unabated Bottom Water Warming and Freshening in the South Pacific Ocean. *J. Geophys. Res. Ocean.* 124, 1778–1794. doi:10.1029/2018JC014775.
- Qiao, F., Yuan, Y., Deng, J., Dai, D., and Song, Z. (2016). Wave–turbulence interaction-induced vertical mixing and its effects in ocean and climate models. *Philos. Trans. R. Soc. A Math. Phys. Eng. Sci.* 374, 20150201. doi:10.1098/rsta.2015.0201.
- Qiu, B., and Chen, S. (2012). Multidecadal Sea Level and Gyre Circulation Variability in the Northwestern Tropical Pacific Ocean. *J. Phys. Oceanogr.* 42, 193–206. doi:10.1175/JPO-D-11-061.1.
- Qiu, B., Chen, S., Qiu, B., and Chen, S. (2005). Variability of the Kuroshio Extension Jet, Recirculation Gyre, and Mesoscale Eddies on Decadal Time Scales. *J. Phys. Oceanogr.* 35, 2090–2103. doi:10.1175/JPO2807.1.
- Qu, X., and Hall, A. (2014). On the persistent spread in snow-albedo feedback. *Clim. Dyn.* 42, 69–81. doi:10.1007/s00382-013-1774-0.
- Quante, M., and Colijn, F. eds. (2016). *North Sea Region Climate Change Assessment*. Cham: Springer International Publishing doi:10.1007/978-3-319-39745-0.
- Quiquet, A., Ritz, C., Punge, H. J., and y Méliá, D. (2013). Greenland ice sheet contribution to sea level rise during the last interglacial period: a modelling study driven and constrained by ice core data. *Clim. Past* 9, 353–366.
- Rabatel, A., Francou, B., Soruco, A., Gomez, J., Cáceres, B., Ceballos, J. L., et al. (2013). Current state of glaciers in the tropical Andes: a multi-century perspective on glacier evolution and climate change. *Cryosph.* 7, 81–102. doi:10.5194/tc-7-81-2013.
- Rabe, B., Karcher, M., Kauker, F., Schauer, U., Toole, J. M., Krishfield, R. A., et al. (2014). Arctic Ocean basin liquid freshwater storage trend 1992–2012. *Geophys. Res. Lett.* 41, 961–968. doi:10.1002/2013GL058121.

- 1 Radić, V., Bliss, A., Beedlow, A. C., Hock, R., Miles, E., and Cogley, J. G. (2014). Regional and global projections of
2 twenty-first century glacier mass changes in response to climate scenarios from global climate models. *Clim. Dyn.*
3 42, 37–58. doi:10.1007/s00382-013-1719-7.
- 4 Rahmstorf, S., Box, J. E., Feulner, G., Mann, M. E., Robinson, A., Rutherford, S., et al. (2015). Exceptional twentieth-
5 century slowdown in Atlantic Ocean overturning circulation. *Nat. Clim. Chang.* doi:10.1038/nclimate2554.
- 6 Rahmstorf, S., Perrette, M., and Vermeer, M. (2012). Testing the robustness of semi-empirical sea level projections.
7 *Clim. Dyn.* 39, 861–875. doi:10.1007/s00382-011-1226-7.
- 8 Rampal, P., Weiss, J., and Marsan, D. (2009). Positive trend in the mean speed and deformation rate of Arctic sea ice,
9 1979–2007. *J. Geophys. Res. Ocean.* 114. doi:10.1029/2008JC005066.
- 10 Rao, S. A., Dhakate, A. R., Saha, S. K., Mahapatra, S., Chaudhari, H. S., Pokhrel, S., et al. (2012). Why is Indian Ocean
11 warming consistently? *Clim. Change* 110, 709–719. doi:10.1007/s10584-011-0121-x.
- 12 Raphael, M. N., Marshall, G. J., Turner, J., Fogt, R. L., Schneider, D., Dixon, D. A., et al. (2016). The Amundsen Sea
13 Low: Variability, Change, and Impact on Antarctic Climate. *Bull. Am. Meteorol. Soc.* 97, 111–121.
14 doi:10.1175/BAMS-D-14-00018.1.
- 15 Rasmussen, D. J., Bittermann, K., Buchanan, M. K., Kulp, S., Strauss, B. H., Kopp, R. E., et al. (2018). Extreme sea
16 level implications of 1.5 °C, 2.0 °C, and 2.5 °C temperature stabilization targets in the 21st and 22nd centuries.
17 *Environ. Res. Lett.* 13. doi:10.1088/1748-9326/aaac87.
- 18 Ray, R. D. (2009). Secular changes in the solar semidiurnal tide of the western North Atlantic Ocean. *Geophys. Res.*
19 *Lett.* doi:10.1029/2009GL040217.
- 20 Raymo, M. E., Kozdon, R., Evans, D., Lisiecki, L., and Ford, H. L. (2018). The accuracy of mid-Pliocene $\delta^{18}\text{O}$ -based
21 ice volume and sea level reconstructions. *Earth-Science Rev.* doi:10.1016/j.earscirev.2017.11.022.
- 22 Reed, A. J., Mann, M. E., Emanuel, K. A., Lin, N., Horton, B. P., Kemp, A. C., et al. (2015). Increased threat of tropical
23 cyclones and coastal flooding to New York City during the anthropogenic era. *Proc. Natl. Acad. Sci.* 112, 12610–
24 12615.
- 25 Reese, R., Gudmundsson, G. H. H., Levermann, A., Winkelmann, R., Reese, A. R., Gudmundsson, G. H. H., et al.
26 (2018). The far reach of ice-shelf thinning in Antarctica. *Nat. Clim. Chang.* 8, 53–57. doi:10.1038/s41558-017-
27 0020-x.
- 28 Reese, R., Levermann, A., Albrecht, T., Seroussi, H., and Winkelmann, R. (submitted). The role of history and strength
29 of the oceanic forcing in sea-level projections from Antarctica with the Parallel Ice Sheet Model. *Cryosph.*
30 (submitted), 1–21.
- 31 Reguero, B. G., Losada, I. J., and Méndez, F. J. (2019). A recent increase in global wave power as a consequence of
32 oceanic warming. *Nat. Commun.* doi:10.1038/s41467-018-08066-0.
- 33 Reichl, B. G., and Hallberg, R. (2018). A simplified energetics based planetary boundary layer (ePBL) approach for
34 ocean climate simulations. *Ocean Model.* 132, 112–129. doi:10.1016/J.OCEMOD.2018.10.004.
- 35 Reintges, A., Martin, T., Latif, M., and Keenlyside, N. S. (2017). Uncertainty in twenty-first century projections of the
36 Atlantic Meridional Overturning Circulation in CMIP3 and CMIP5 models. *Clim. Dyn.* 49, 1495–1511.
37 doi:10.1007/s00382-016-3180-x.
- 38 Reinthaler, J., Paul, F., Granados, H. D., Rivera, A., and Huggel, C. (2019). Area changes of glaciers on active
39 volcanoes in Latin America between 1986 and 2015 observed from multi-temporal satellite imagery. *J. Glaciol.*,
40 1–15. doi:DOI: 10.1017/jog.2019.30.
- 41 Renault, L., Molemaker, M. J., McWilliams, J. C., Shchepetkin, A. F., Lemarié, F., Chelton, D., et al. (2016).
42 Modulation of Wind Work by Oceanic Current Interaction with the Atmosphere. *J. Phys. Oceanogr.* 46, 1685–
43 1704. doi:10.1175/JPO-D-15-0232.1.
- 44 Renner, A. H. H., Gerland, S., Haas, C., Spreen, G., Beckers, J. F., Hansen, E., et al. (2014). Evidence of Arctic sea ice
45 thinning from direct observations. *Geophys. Res. Lett.* doi:10.1002/2014GL060369.
- 46 Renner, A. H. H., Heywood, K. J., and Thorpe, S. E. (2009). Validation of three global ocean models in the Weddell
47 Sea. *Ocean Model.* 30, 1–15. doi:10.1016/J.OCEMOD.2009.05.007.
- 48 Réveillet, M., Six, D., Vincent, C., Rabatel, A., Dumont, M., Lafaysse, M., et al. (2018). Relative performance of
49 empirical and physical models in assessing the seasonal and annual glacier surface mass balance of Saint-Sorlin
50 Glacier (French Alps). *Cryosphere* 12, 1367–1386. doi:10.5194/tc-12-1367-2018.
- 51 Réveillet, M., Vincent, C., Six, D., and Rabatel, A. (2017). Which empirical model is best suited to simulate glacier
52 mass balances? *J. Glaciol.* 63, 39–54. doi:10.1017/jog.2016.110.
- 53 Reverdin, G. (2010). North Atlantic Subpolar Gyre Surface Variability (1895–2009). *J. Clim.* 23, 4571–4584.
54 doi:10.1175/2010JCLI3493.1.
- 55 RGI Consortium (2017a). Randolph Glacier Inventory—A Dataset of Global Glacier Outlines: Version 6.0: Technical
56 Report, Global Land Ice Measurements from Space. doi:https://doi.org/10.7265.
- 57 RGI Consortium (2017b). Randolph Glacier Inventory – A Dataset of Global Glacier Outlines: Version 6.0: Available
58 at: https://doi.org/10.7265/N5-RGI-60.
- 59 Rhein, M., Kieke, D., Hüttel-Kabus, S., Roessler, A., Mertens, C., Meissner, R., et al. (2011). Deep water formation, the
60 subpolar gyre, and the meridional overturning circulation in the subpolar North Atlantic. *Deep Sea Res. Part II*
61 *Top. Stud. Oceanogr.* 58, 1819–1832. doi:10.1016/j.dsr2.2010.10.061.

- Rhein, M., Rintoul, S. R., Aoki, S., Campos, E., Chambers, D., Feely, R. A., et al. (2013). Observations: Ocean. in *{Climate Change 2013}: The Physical Science Basis. {Contribution of Working Group I} to the {Fifth Assessment Report of the IPCC}*, eds. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, et al. (Cambridge University Press), 255–316. Available at: <http://bit.ly/2MLrFpL>.
- Ribeiro, A., Barbosa, S. M., Scotto, M. G., and Donner, R. V. (2014). Changes in extreme sea-levels in the Baltic Sea. *Tellus A Dyn. Meteorol. Oceanogr.* 66, 20921. doi:10.3402/tellusa.v66.20921.
- Richter, I., and Xie, S. P. (2008). On the origin of equatorial Atlantic biases in coupled general circulation models. *Clim. Dyn.* doi:10.1007/s00382-008-0364-z.
- Richter, K., and Marzeion, B. (2014). Earliest local emergence of forced dynamic and steric sea-level trends in climate models. *Environ. Res. Lett.* 9, 114009. doi:10.1088/1748-9326/9/11/114009.
- Richter, K., Marzeion, B., and Riva, R. (2017). The effect of spatial averaging and glacier melt on detecting a forced signal in regional sea level. *Environ. Res. Lett.* doi:10.1088/1748-9326/aa5967.
- Ricker, R., Hendricks, S., Perovich, D. K., Helm, V., and Gerdes, R. (2015). Impact of snow accumulation on CryoSat-2 range retrievals over Arctic sea ice: An observational approach with buoy data. *Geophys. Res. Lett.* 42, 4447–4455. doi:10.1002/2015GL064081.
- Ridley, J., Gregory, J. M., Huybrechts, P., and Lowe, J. (2010). Thresholds for irreversible decline of the Greenland ice sheet. *Clim. Dyn.* 35, 1065–1073. doi:10.1007/s00382-009-0646-0.
- Ridley, J. K., and Blockley, E. W. (2018). Brief communication: Solar radiation management not as effective as CO₂ mitigation for Arctic sea ice loss in hitting the 1.5 and 2°C COP climate targets. *Cryosphere* 12, 3355–3360. doi:10.5194/tc-12-3355-2018.
- Ridley, J. K., Lowe, J. A., and Hewitt, H. T. (2012). How reversible is sea ice loss? *Cryosphere* 6, 193–198. doi:10.5194/tc-6-193-2012.
- Rietbroek, R., Brunnabend, S.-E., Kusche, J., Schröter, J., and Dahle, C. (2016). Revisiting the contemporary sea-level budget on global and regional scales. *Proc. Natl. Acad. Sci. U. S. A.* 113, 1504–1509. doi:10.1073/pnas.1519132113.
- Riggs, G. A., Hall, D. K., and Román, M. O. (2017). Overview of NASA’s MODIS and Visible Infrared Imaging Radiometer Suite (VIIRS) snow-cover Earth System Data Records. *Earth Syst. Sci. Data* 9, 765–777. doi:10.5194/essd-9-765-2017.
- Rignot, E., Jacobs, S., Mouginot, J., and Scheuchl, B. (2013). Ice-shelf melting around Antarctica. *Science* 341, 266–270. doi:10.1126/science.1235798.
- 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. doi:10.1002/2014GL060140.
- Rignot, E., Mouginot, J., and Scheuchl, B. (2011). Ice flow of the antarctic ice sheet. *Science* 333, 1427–1430. doi:10.1126/science.1208336.
- Rignot, E., Mouginot, J., Scheuchl, B., van den Broeke, M. R., van Wessem, M. J., and Morlighem, M. (2019). Four decades of Antarctic Ice Sheet mass balance from 1979–2017. *Proc. Natl. Acad. Sci.* 116, 1–9. doi:10.1073/pnas.1812883116.
- Rind, D., Schmidt, G. A., Jonas, J., Miller, R., Nazarenko, L., Kelley, M., et al. (2018). Multicentury Instability of the Atlantic Meridional Circulation in Rapid Warming Simulations With GISS ModelE2. *J. Geophys. Res. Atmos.* 123, 6331–6355. doi:10.1029/2017JD027149.
- Ringler, T., Petersen, M., Higdon, R. L., Jacobsen, D., Jones, P. W., and Maltrud, M. (2013). A multi-resolution approach to global ocean modeling. *Ocean Model.* 69, 211–232. doi:10.1016/j.oceanmod.2013.04.010.
- Rintoul, S. R., Silvano, A., Pena-Molino, B., van Wijk, E., Rosenberg, M., Greenbaum, J. S., et al. (2016). Ocean heat drives rapid basal melt of the Totten Ice Shelf. *Sci. Adv.* 2. doi:10.1126/sciadv.1601610.
- Riser, S. C., Freeland, H. J., Roemmich, D., Wijffels, S., Troisi, A., Belbéoch, M., et al. (2016). Fifteen years of ocean observations with the global Argo array. *Nat. Clim. Chang.* 6, 145. Available at: <http://dx.doi.org/10.1038/nclimate2872>.
- Ritz, C., Edwards, T. L., Durand, G., Payne, A. J., Peyaud, V., and Hindmarsh, R. C. A. (2015). Potential sea-level rise from Antarctic ice-sheet instability constrained by observations. *Nature* 528, 115–118. doi:10.1038/nature16147.
- Rivera, A., Bown, F., Carrión, D., and Zenteno, P. (2012). Glacier responses to recent volcanic activity in Southern Chile. *Environ. Res. Lett.* 7, 14036. doi:10.1088/1748-9326/7/1/014036.
- Roach, L. A., Dean, S. M., and Renwick, J. A. (2018a). Consistent biases in Antarctic sea ice concentration simulated by climate models. *Cryosph.* 12, 365–383. doi:10.5194/tc-12-365-2018.
- Roach, L. A., Dörr, J., Holmes, C. R., Massonnet, F., Blockley, E. W., Notz, D., et al. (submitted). Antarctic sea ice in CMIP6. *Geophys. Res. Lett.* (submitted).
- Roach, L. A., Horvat, C., Dean, S. M., and Bitz, C. M. (2018b). An Emergent Sea Ice Floe Size Distribution in a Global Coupled Ocean-Sea Ice Model. *J. Geophys. Res. Ocean.* 123, 4322–4337. doi:10.1029/2017JC013692.
- Robel, A. A. (2017). Thinning sea ice weakens buttressing force of iceberg mélange and promotes calving. *Nat. Commun.* 8, 1–7. doi:10.1038/ncomms14596.
- Robel, A. A., and Banwell, A. F. (2019). A Speed Limit on Ice Shelf Collapse Through Hydrofracture. *Geophys. Res.*

- Lett.* 46, 12092–12100. doi:10.1029/2019GL084397.
- Robel, A. A., and Tziperman, E. (2016). The role of ice stream dynamics in deglaciation. *J. Geophys. Res.* 121, 1540–1554.
- Roberts, C. D., Jackson, L., and McNeall, D. (2014). Is the 2004–2012 reduction of the Atlantic meridional overturning circulation significant? *Geophys. Res. Lett.* 41, 3204–3210. doi:10.1002/2014GL059473.
- Roberts, M. J., Jackson, L. C., Roberts, C. D., Meccia, V., Docquier, D., Konigk, T., et al. (submitted). Sensitivity of the Atlantic Meridional Overturning Circulation to Model Resolution in CMIP6 HighResMIP Simulations and Implications for Future Changes. *JAMES* (submitted).
- Roberts, M. J., Baker, A., Blockley, E. W., Calvert, D., Coward, A., Hewitt, H. T., et al. (2019). Description of the resolution hierarchy of the global coupled HadGEM3-GC3.1 model as used in CMIP6 HighResMIP experiments. *Geosci. Model Dev.* 12, 4999–5028. doi:10.5194/gmd-12-4999-2019.
- Roberts, M. J., Vidale, P. L., Senior, C., Hewitt, H. T., Bates, C., Berthou, S., et al. (2018). The Benefits of Global High Resolution for Climate Simulation: Process Understanding and the Enabling of Stakeholder Decisions at the Regional Scale. *Bull. Am. Meteorol. Soc.* 99, 2341–2359. doi:10.1175/BAMS-D-15-00320.1.
- Robinson, A., Alvarez-Solas, J., Calov, R., Ganopolski, A., and Montoya, M. (2017). MIS-11 duration key to disappearance of the Greenland ice sheet. *Nat. Commun.* 8, 16008. Available at: <https://doi.org/10.1038/ncomms16008>.
- Robinson, A., Calov, R., and Ganopolski, A. (2011). Greenland ice sheet model parameters constrained using simulations of the Eemian Interglacial. *Clim. Past* 7, 381–396.
- Robinson, A., Calov, R., and Ganopolski, A. (2012). Multistability and critical thresholds of the Greenland ice sheet. 2, 429. doi:10.1038/nclimate1449 <https://www.nature.com/articles/nclimate1449#supplementary-information>.
- Robinson, A., and Goelzer, H. (2014). The importance of insolation changes for paleo ice sheet modeling. *Cryosphere* 8, 1419–1428. doi:10.5194/tc-8-1419-2014.
- Robson, J., Ortega, P., and Sutton, R. (2016). A reversal of climatic trends in the North Atlantic since 2005. *Nat. Geosci.* 9, 513–517. doi:10.1038/ngeo2727.
- Robson, J., Sutton, R., Lohmann, K., Smith, D., and Palmer, M. D. (2012). Causes of the Rapid Warming of the North Atlantic Ocean in the Mid-1990s. *J. Clim.* 25, 4116–4134. doi:10.1175/JCLI-D-11-00443.1.
- Roe, G. H., Baker, M. B., and Herla, F. (2017). Centennial glacier retreat as categorical evidence of regional climate change. *Nat. Geosci.* 10, 95–99. Available at: <https://doi.org/10.1038/ngeo2863>.
- Roemmich, D., Church, J., Gilson, J., Monselesan, D., Sutton, P., and Wijffels, S. (2015). Unabated planetary warming and its ocean structure since 2006. *Nat. Clim. Chang.* 5, 240. Available at: <http://dx.doi.org/10.1038/nclimate2513>.
- Roemmich, D., Gilson, J., Sutton, P., Wijffels, S., and Riser, S. (2007). Decadal spinup of the South Pacific subtropical gyre. *J. Phys. Ocean.* 37, 162–173.
- Roemmich, D., Gould, J. W., and Gilson, J. (2012). 135 years of global ocean warming between the Challenger expedition and the Argo Programme. *Nat. Clim. Chang.* doi:10.1038/nclimate1461.
- Rohling, E. J., Hibbert, F. D., Grant, K. M., Galaasen, E. V., Irvani, N., Kleiven, H. F., et al. (2019). Asynchronous Antarctic and Greenland ice-volume contributions to the last interglacial sea-level highstand. *Nat. Commun.* 10, 1–9.
- Romanovsky, V. E., Smith, S. L., Isaksen, K., Shiklomanov, N. I., Streletskiy, D. A., Kholodov, A. L., et al. (2018). “Terrestrial Permafrost,” in *State of the Climate in 2017*, S161–S165.
- Romanovsky, V., Isaksen, K., Drozdov, D., Anisimov, O., Instantes, A., Leibmann, M., et al. (2017). “Changing Permafrost and its Impacts,” in *Snow, Water, Ice and Permafrost in the Arctic (SWIPA) 2017* (Oslo, Norway: Arctic Monitoring and Assessment Programme (AMAP)), 65–136.
- Rösel, A., Itkin, P., King, J., Divine, D., Wang, C., Granskog, M. A., et al. (2018). Thin Sea Ice, Thick Snow, and Widespread Negative Freeboard Observed During N-ICE2015 North of Svalbard. *J. Geophys. Res. Ocean.* 123, 1156–1176. doi:10.1002/2017JC012865.
- Rosenblum, E., and Eisenman, I. (2017). Sea ice trends in climate models only accurate in runs with biased global warming. *J. Clim.* 30, 6265–6278. doi:10.1175/JCLI-D-16-0455.1.
- Rott, H., Abdel Jabber, W., Wuite, J., Scheiblaue, S., Floricioiu, D., Van Wessem, J. M., et al. (2018). Changing pattern of ice flow and mass balance for glaciers discharging into the Larsen A and B embayments, Antarctic Peninsula, 2011 to 2016. *Cryosphere* 12, 1273–1291. doi:10.5194/tc-12-1273-2018.
- Roussel, M.-L., Lemonnier, F., Genthon, C., and Krinner, G. (submitted). Brief communication: Evaluating Antarctic precipitation in ERA5 and CMIP6 against CloudSat observations. *Cryosph. Discuss.* (submitted).
- Roxy, M. K., Dasgupta, P., McPhaden, M. J., Suematsu, T., Zhang, C., and Kim, D. (2019). Twofold expansion of the Indo-Pacific warm pool warps the MJO life cycle. *Nature* 575, 647–651. doi:10.1038/s41586-019-1764-4.
- Roxy, M. K., Ritika, K., Terray, P., and Masson, S. (2014). The Curious Case of Indian Ocean Warming*,+. *J. Clim.* 27, 8501–8509. doi:10.1175/JCLI-D-14-00471.1.
- Rückamp, M., Falk, U., Frieler, K., Lange, S., and Humbert, A. (2018). The effect of overshooting 1.5°C global warming on the mass loss of the Greenland ice sheet. *Earth Syst. Dynam.* 9, 1169–1189. doi:10.5194/esd-9-1169-2018.

- 1 Rückamp, M., Goelzer, H., and Humbert, A. (submitted). Sensitivity of Greenland ice sheet projections to spatial
2 resolution in higher-order simulations: the AWI contribution to ISMIP6-Greenland using ISSM. *Cryosph.*
3 (submitted).
- 4 Ruckert, K. L., Shaffer, G., Pollard, D., Guan, Y., Wong, T. E., Forest, C. E., et al. (2017). Assessing the Impact of
5 Retreat Mechanisms in a Simple Antarctic Ice Sheet Model Using Bayesian Calibration. *PLoS One* 12, e0170052.
6 doi:10.1371/journal.pone.0170052.
- 7 Rugenstein, M. A. A., Caldeira, K., and Knutti, R. (2016). Dependence of global radiative feedbacks on evolving
8 patterns of surface heat fluxes. *Geophys. Res. Lett.* doi:10.1002/2016GL070907.
- 9 Rugenstein, M. A. A., Winton, M., Stouffer, R. J., Griffies, S. M., and Hallberg, R. (2013). Northern high-latitude heat
10 budget decomposition and transient warming. *J. Clim.* doi:10.1175/JCLI-D-11-00695.1.
- 11 Ruppel, C. D., Herman, B. M., Brothers, L. L., and Hart, P. E. (2016). Subsea ice-bearing permafrost on the U.S.
12 Beaufort Margin: 2. Borehole constraints. *Geochemistry, Geophys. Geosystems* 17, 4333–4353.
13 doi:10.1002/2016GC006582.
- 14 Ruprich-Robert, Y., Danabasoglu, G., Msadek, R., Castruccio, F. S., Delworth, T. L., and Yeager, S. G. (2018).
15 Modulation of Arctic Sea Ice Loss by Atmospheric Teleconnections from Atlantic Multidecadal Variability. *J.*
16 *Clim.* 32, 1419–1441. doi:10.1175/jcli-d-18-0307.1.
- 17 Russell, J. L., Kamenkovich, I., Bitz, C., Ferrari, R., Gille, S. T., Goodman, P. J., et al. (2018). Metrics for the
18 Evaluation of the Southern Ocean in Coupled Climate Models and Earth System Models. *J. Geophys. Res. Ocean.*
19 123, 3120–3143. doi:10.1002/2017JC013461.
- 20 Ryan, J. C., Hubbard, A., Stibal, M., Irvine-Fynn, T. D., Cook, J., Smith, L. C., et al. (2018). Dark zone of the
21 Greenland Ice Sheet controlled by distributed biologically-active impurities. *Nat. Commun.* 9, 1065.
22 doi:10.1038/s41467-018-03353-2.
- 23 Ryan, J. C., Smith, L. C., van As, D., Cooley, S. W., Cooper, M. G., Pitcher, L. H., et al. (2019). Greenland Ice Sheet
24 surface melt amplified by snowline migration and bare ice exposure. *Sci. Adv.* 5. doi:10.1126/sciadv.aav3738.
- 25 Rykaczewski, R. R., Dunne, J. P., Sydeman, W. J., Garc'ia-Reyes, M., Black, B. A., and Bograd, S. J. (2015). Poleward
26 displacement of coastal upwelling-favorable winds in the ocean's eastern boundary currents through the 21st
27 century. *Geophys. Res. Lett.* 42, 6424–6431.
- 28 Saha, S., Owen, L., Orr, E., and Caffee, M. (2018). *Timing and nature of Holocene glacier advances at the*
29 *northwestern end of the Himalayan-Tibetan orogen.* doi:10.1016/j.quascirev.2018.03.009.
- 30 Saito, F., Abe-Ouchi, A., Takahashi, K., and Blatter, H. (2016). SeaRISE experiments revisited: potential sources of
31 spread in multi-model projections of the Greenland ice sheet. *Cryosph.* 10, 43–63. doi:10.5194/tc-10-43-2016.
- 32 Sakai, A., and Fujita, K. (2017). Contrasting glacier responses to recent climate change in high-mountain Asia. *Sci. Rep.*
33 7, 13717. doi:10.1038/s41598-017-14256-5.
- 34 Sakamoto, T., Komuro, Y., Nishimura, T., Ishii, M., Tatebe, H., Shiogama, H., et al. (2012). MIROC4h: A New High-
35 Resolution Atmosphere-Ocean Coupled General Circulation Model. *J. Meteorol. Soc. Japan. Ser. II* 90, 325–359.
36 doi:10.2151/jmsj.2012-301.
- 37 Sallée, J.-B. (2018). Southern Ocean Warming. *Oceanography* 31. doi:https://doi.org/10.5670/oceanog.2018.215.
- 38 Sallée, J.-B., Pellichero, V., Akhondas, C., Pauthenet, E., Vignes, L., Schmidtke, S., et al. (submitted). Fifty-year
39 changes of the world ocean's surface layer in response to climate change. *Nature* (submitted).
- 40 Sallée, J.-B., Shuckburgh, E., Bruneau, N., Meijers, A. J. S., Bracegirdle, T. J., and Wang, Z. (2013a). Assessment of
41 Southern Ocean mixed-layer depths in CMIP5 models: Historical bias and forcing response. *J. Geophys. Res.*
42 *Ocean.* 118, 1845–1862. doi:10.1002/jgrc.20157.
- 43 Sallée, J.-B., Shuckburgh, E., Bruneau, N., Meijers, A. J. S., Bracegirdle, T. J., Wang, Z., et al. (2013b). Assessment of
44 Southern Ocean water mass circulation and characteristics in CMIP5 models: Historical bias and forcing
45 response. *J. Geophys. Res. Ocean.* 118, 1830–1844. doi:10.1002/jgrc.20135.
- 46 Sandberg Sørensen, L., Simonsen, S. B., Forsberg, R., Khvorostovsky, K., Meister, R., and Engdahl, M. E. (2018). 25
47 years of elevation changes of the Greenland Ice Sheet from ERS, Envisat, and CryoSat-2 radar altimetry. *Earth*
48 *Planet. Sci. Lett.* 495, 234–241. doi:https://doi.org/10.1016/j.epsl.2018.05.015.
- 49 Sangiorgi, F., Bijl, P. K., Passchier, S., Salzmann, U., Schouten, S., McKay, R., et al. (2018). Southern Ocean warming
50 and Wilkes Land ice sheet retreat during the mid-Miocene. *Nat. Commun.* 9, 1–11. doi:10.1038/s41467-017-
51 02609-7.
- 52 Santoso, A., Mcphaden, M. J., and Cai, W. (2017). The Defining Characteristics of ENSO Extremes and the Strong
53 2015/2016 El Niño. *Rev. Geophys.* 55, 1079–1129. doi:10.1002/2017RG000560.
- 54 Sasaki, H., Sasai, Y., Kawahara, S., Furuichi, M., Araki, F., Ishida, A., et al. (2004). A series of eddy-resolving ocean
55 simulations in the world ocean - OFES (OGCM for the Earth Simulator) project. in *Oceans '04 MTS/IEEE*
56 *Techno-Ocean '04 (IEEE Cat. No.04CH37600)* (IEEE), 1535–1541. doi:10.1109/OCEANS.2004.1406350.
- 57 Scapozza, C., Lambiel, C., Bozzini, C., Mari, S., and Conedera, M. (2014). Assessing the rock glacier kinematics on
58 three different timescales: a case study from the southern Swiss Alps. *Earth Surf. Process. Landforms* 39, 2056–
59 2069. doi:10.1002/esp.3599.
- 60 Schemm, S. (2018). Regional Trends in Weather Systems Help Explain Antarctic Sea Ice Trends. *Geophys. Res. Lett.*
61 45, 7165–7175. doi:10.1029/2018GL079109.

- 1 Schenk, T., Csatho, B., van der Veen, C., and McCormick, D. (2014). Fusion of multi-sensor surface elevation data for
2 improved characterization of rapidly changing outlet glaciers in Greenland. *Remote Sens. Environ.* 149, 239—
3 251. doi:10.1016/j.rse.2014.04.005.
- 4 Scherler, D., Wulf, H., and Gorelick, N. (2018). Global Assessment of Supraglacial Debris-Cover Extents. *Geophys.*
5 *Res. Lett.* 45, 11,798–11,805. doi:10.1029/2018GL080158.
- 6 Schindelegger, M., Green, J. A. M., Wilmes, S. B., and Haigh, I. D. (2018). Can We Model the Effect of Observed Sea
7 Level Rise on Tides? *J. Geophys. Res. Ocean.* doi:10.1029/2018JC013959.
- 8 Schlegel, N.-J., Seroussi, H., Schodlok, M. P., Larour, E. Y., Boening, C., Limonadi, D., et al. (2018). Exploration of
9 Antarctic Ice Sheet 100-year contribution to sea level rise and associated model uncertainties using the ISSM
10 framework. *Cryosph. Discuss.*
- 11 Schlegel, N. J., Larour, E., Seroussi, H., Morlighem, M., and Box, J. E. (2015). Ice discharge uncertainties in Northeast
12 Greenland from boundary conditions and climate forcing of an ice flow model. *J. Geophys. Res. Surf.* 120, 29–54.
13 doi:10.1002/2014jfg003359.
- 14 Schleussner, C. F., Lissner, T. K., Fischer, E. M., Wohland, J., Perrette, M., Golly, A., et al. (2016). Differential climate
15 impacts for policy-relevant limits to global warming: The case of 1.5 °C and 2 °C. *Earth Syst. Dyn.*
16 doi:10.5194/esd-7-327-2016.
- 17 Schloesser, F., Friedrich, T., Timmermann, A., DeConto, R. M., and Pollard, D. (2019a). Antarctic iceberg impacts on
18 future Southern Hemisphere climate. *Nat. Clim. Chang.* 9, 672–677. doi:10.1038/s41558-019-0546-1.
- 19 Schloesser, F., Friedrich, T., Timmermann, A., DeConto, R. M., and Pollard, D. (2019b). Antarctic iceberg impacts on
20 future Southern Hemisphere climate. *Nat. Clim. Chang.* 9, 672–677. doi:10.1038/s41558-019-0546-1.
- 21 Schlosser, E., Haumann, F. A., and Raphael, M. N. (2018). Atmospheric influences on the anomalous 2016 Antarctic
22 sea ice decay. *Cryosph.* 12, 1103–1119. doi:10.5194/tc-12-1103-2018.
- 23 Schmidt, L. S., Aðalgeirsdóttir, G., Pálsson, F., Langen, P. L., Gudmundsson, S., and Björnsson, H. (2019). Dynamic
24 simulations of Vatnajökull ice cap from 1980 to 2300. *J. Glaciol.*, 1–16. doi:DOI: 10.1017/jog.2019.90.
- 25 Schmidtko, S., Heywood, K. J., Thompson, A. F., and Aoki, S. (2014). Multidecadal warming of Antarctic waters.
26 *Science* 346, 1227–1231. doi:10.1126/science.1256117.
- 27 Schmidtko, S., and Johnson, G. C. (2012). Multidecadal warming and shoaling of antarctic intermediate water. *J. Clim.*
28 doi:10.1175/JCLI-D-11-00021.1.
- 29 Schoof, C. (2007). Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *J. Geophys. Res. Earth*
30 *Surf.* 112, n/a–n/a. doi:10.1029/2006JF000664.
- 31 Schoof, C. (2010). Ice-sheet acceleration driven by melt supply variability. *Nature* 468, 803–806.
32 doi:10.1038/nature09618.
- 33 Schröder, L., Horwath, M., Dietrich, R., Helm, V., van den Broeke, M. R., and Ligtenberg, S. R. M. (2019). Four
34 decades of Antarctic surface elevation changes from multi-mission satellite altimetry. *Cryosph.* 13, 427–449.
35 doi:10.5194/tc-13-427-2019.
- 36 Schroeder, D. M., Dowdeswell, J. A., Siegert, M. J., Bingham, R. G., Chu, W., Mackie, E. J., et al. (2019).
37 Multidecadal observations of the Antarctic ice sheet from restored analog radar records. *Atmos. Planet. Sci.* 116.
38 doi:10.25740/ykq4-9345.
- 39 Schroeter, S., Hobbs, W., and Bindoff, N. L. (2017). Interactions between Antarctic sea ice and large-scale atmospheric
40 modes in CMIP5 models. *Cryosph.* 11, 789–803. doi:10.5194/tc-11-789-2017.
- 41 Schroeter, S., Hobbs, W., Bindoff, N. L., Massom, R., and Matear, R. (2018). Drivers of Antarctic sea ice volume
42 change in CMIP5 models. *J. Geophys. Res. Ocean.* 123, 1–25. doi:10.1029/2018JC014177.
- 43 Schweiger, A. J., Wood, K. R., and Zhang, J. (2019). Arctic Sea Ice Volume Variability over 1901–2010: A Model-
44 Based Reconstruction. *J. Clim.* 32, 4731–4752. doi:10.1175/jcli-d-19-0008.1.
- 45 Screen, J. A., and Williamson, D. (2017). Ice-free Arctic at 1.5 °C? *Nat. Clim. Chang.* 7, 230–231.
46 doi:10.1038/nclimate3248.
- 47 Seehaus, T., Cook, A. J., Silva, A. B., and Braun, M. (2018). Changes in glacier dynamics in the northern Antarctic
48 Peninsula since 1985. *Cryosphere* 12, 577–594. doi:10.5194/tc-12-577-2018.
- 49 Seidov, D. (2001). Meltwater and the global ocean conveyor: northern versus southern connections. *Glob. Planet.*
50 *Change* 30, 257–270. doi:10.1016/S0921-8181(00)00087-4.
- 51 Sellevold, R., Kampenhout, L., Lenaerts, J., Noël, B., Lipscomb, W., and Vizcaíno, M. (2019). Surface mass balance
52 downscaling through elevation classes in an Earth system model: application to the Greenland ice sheet. *Cryosph.*
53 13, 3193–3208. doi:10.5194/tc-13-3193-2019.
- 54 Sen Gupta, A., McGregor, S., van Sebille, E., Ganachaud, A., Brown, J. N., and Santoso, A. (2016). Future changes to
55 the Indonesian Throughflow and Pacific circulation: The differing role of wind and deep circulation changes.
56 *Geophys. Res. Lett.* 43, 1669–1678. doi:10.1002/2016GL067757.
- 57 Sen Gupta, A., Santoso, A., and Taschetto, A. S. (2009a). AMS Journals Online - Projected Changes to the Southern
58 Hemisphere Ocean and Sea Ice in the IPCC AR4 Climate Models. *J.*
- 59 Sen Gupta, A., Santoso, A., Taschetto, A. S., Ummenhofer, C. C., Trevena, J., and England, M. H. (2009b). Projected
60 Changes to the Southern Hemisphere Ocean and Sea Ice in the IPCC AR4 Climate Models. *J. Clim.* 22, 3047–
61 3078. doi:10.1175/2008JCLI2827.1.

- 1 Sen Gupta, A., Stellema, A., Pontes, G., Taschetto, A. S., Verges, A., and Rossi, V. (submitted). Future Changes to the
2 upper ocean Western Boundary Currents across two generations of climate models. *Environ. Res. Lett.*
3 (submitted).
- 4 Seo, H., Miller, A. J., Roads, J. O., Seo, H., Miller, A. J., and Roads, J. O. (2007). The Scripps Coupled Ocean–
5 Atmosphere Regional (SCOAR) Model, with Applications in the Eastern Pacific Sector. *J. Clim.* 20, 381–402.
6 doi:10.1175/JCLI4016.1.
- 7 Seo, H., Murtugudde, R., Jochum, M., and Miller, A. J. (2008). Modeling of mesoscale coupled ocean–atmosphere
8 interaction and its feedback to ocean in the western Arabian Sea. *Ocean Model.* 25, 120–131.
9 doi:10.1016/J.OCEMOD.2008.07.003.
- 10 Seo, K.-W., Wilson, C. R., Scambos, T., Kim, B.-M., Waliser, D. E., Tian, B., et al. (2015). Surface mass balance
11 contributions to acceleration of Antarctic ice mass loss during 2003–2013. *J. Geophys. Res. Solid Earth* 120,
12 3617–3627. doi:10.1002/2014JB011755.
- 13 Sérazin, G., Meyssignac, B., Penduff, T., Terray, L., Barnier, B., and Molines, J. M. (2016). Quantifying uncertainties
14 on regional sea level change induced by multidecadal intrinsic oceanic variability. *Geophys. Res. Lett.*
15 doi:10.1002/2016GL069273.
- 16 Seroussi, H. et al. (submitted). ISMIP6 Antarctica: a multi-model ensemble of the Antarctic ice sheet evolution over the
17 21st century. (submitted).
- 18 Seroussi, H., Morlighem, M., Rignot, E., Mouginot, J., Larour, E., Schodlok, M., et al. (2014). Sensitivity of the
19 dynamics of Pine Island Glacier, West Antarctica, to climate forcing for the next 50 years. *Cryosphere* 8, 1699–
20 1710. doi:10.5194/tc-8-1699-2014.
- 21 Seroussi, H., Nakayama, Y., Larour, E., Menemenlis, D., Morlighem, M., Rignot, E., et al. (2017). Continued retreat of
22 Thwaites Glacier, West Antarctica, controlled by bed topography and ocean circulation. *Geophys. Res. Lett.* 44,
23 6191–6199. doi:10.1002/2017GL072910.
- 24 Seroussi, H., Nowicki, S., Payne, A. J., Goelzer, H., Lipscomb, W. H., Ouchi, A. A., et al. (submitted). ISMIP6
25 Antarctica: a multi-model ensemble of the Antarctic ice sheet evolution over the 21 st century. *Cryosph.*
26 (submitted).
- 27 Seroussi, H., Nowicki, S., Simon, E., Abe Ouchi, A., Albrecht, T., Brondex, J., et al. (2019). initMIP-Antarctica: An ice
28 sheet model initialization experiment of ISMIP6. *Cryosph. Discuss.*, 1–44. doi:10.5194/tc-2018-271.
- 29 Serreze, M. C., and Stroeve, J. (2015). Arctic sea ice trends, variability and implications for seasonal ice forecasting.
30 *Philos. Trans. R. Soc. A Math. Phys. Eng. Sci.* 373. doi:10.1098/rsta.2014.0159.
- 31 Sgubin, G., Swingedouw, D., Drijfhout, S., Mary, Y., and Bennabi, A. (2017). Abrupt cooling over the North Atlantic
32 in modern climate models. *Nat. Commun.* 8. doi:10.1038/ncomms14375.
- 33 Shackleton, S., Bereiter, B., Baggenstos, D., Bauska, T. K., Brook, E. J., Marcott, S. A., et al. (2019). Is the Noble Gas-
34 Based Rate of Ocean Warming During the Younger Dryas Overestimated? *Geophys. Res. Lett.* 46, 5928–5936.
35 doi:10.1029/2019GL082971.
- 36 Shakhova, N., Semiletov, I., Gustafsson, O., Sergienko, V., Lobkovsky, L., Dudarev, O., et al. (2017). Current rates and
37 mechanisms of subsea permafrost degradation in the East Siberian Arctic Shelf. *Nat. Commun.* 8, 15872.
38 doi:10.1038/ncomms15872.
- 39 Shakun, J. D., Clark, P. U., He, F., Lifton, N. A., Liu, Z., and Otto-Bliesner, B. L. (2015). Regional and global forcing
40 of glacier retreat during the last deglaciation. *Nat. Commun.* 6, 8059. doi:10.1038/ncomms9059.
- 41 Shakun, J. D., Clark, P. U., He, F., Marcott, S. A., Mix, A. C., Liu, Z., et al. (2012). Global warming preceded by
42 increasing carbon dioxide concentrations during the last deglaciation. doi:10.1038/nature10915.
- 43 Shannon, S. R., Payne, A. J., Bartholomew, I. D., van den Broeke, M. R., Edwards, T. L., Fettweis, X., et al. (2013).
44 Enhanced basal lubrication and the contribution of the Greenland ice sheet to future sea-level rise. *Proc. Natl.*
45 *Acad. Sci.* 110, 14156–14161. doi:10.1073/pnas.1212647110.
- 46 Shannon, S., Smith, R., Wiltshire, A., Payne, T., Huss, M., Betts, R., et al. (2019). Global glacier volume projections
47 under high-end climate change scenarios. *Cryosph.* 13, 325–350. doi:10.5194/tc-13-325-2019.
- 48 Sheen, K. L., Garabato, A. C. N., Brearley, J. A., Meredith, M. P., Polzin, K. L., Smeed, D. A., et al. (2014). Eddy-
49 induced variability in Southern Ocean abyssal mixing on climatic timescales. *Nat. Geosci.* 7.
50 doi:10.1038/ngeo2200.
- 51 Shepherd, A., Gilbert, L., Muir, A. S., Konrad, H., McMillan, M., Slater, T., et al. (2019). Trends in Antarctic Ice Sheet
52 Elevation and Mass. *Geophys. Res. Lett.* 46, 8174–8183. doi:10.1029/2019gl082182.
- 53 Shepherd, A., Ivins, E. R., Geruo, A., Barletta, V. R., Bentley, M. J., Bettadpur, S., et al. (2012). A reconciled estimate
54 of ice-sheet mass balance. *Science* 338, 1183–1189. doi:10.1126/science.1228102.
- 55 Shu, Q., Song, Z., and Qiao, F. (2015). Assessment of sea ice simulations in the CMIP5 models. *Cryosphere* 9, 399–
56 409. doi:10.5194/tc-9-399-2015.
- 57 Sidorenko, D., Goessling, H. F., Koldunov, N. V., Scholz, P., Danilov, S., Barbi, D., et al. (2019). Evaluation of
58 FESOM2.0 coupled to ECHAM6.3: Pre-industrial and HighResMIP simulations. *J. Adv. Model. Earth Syst.* 11.
59 doi:10.1029/2019MS001696.
- 60 Sigmond, M., Fyfe, J. C., and Swart, N. C. (2018). Ice-free Arctic projections under the Paris Agreement. *Nat. Clim.*
61 *Chang.* 8, 404–408. doi:10.1038/s41558-018-0124-y.

- 1 Silvano, A., Rintoul, S. R., Peña-Molino, B., Hobbs, W. R., van Wijk, E., Aoki, S., et al. (2018). Freshening by glacial
2 meltwater enhances melting of ice shelves and reduces formation of Antarctic Bottom Water. *Sci. Adv.* 4.
3 doi:10.1126/sciadv.aap9467.
- 4 Silvy, Y., Guilyardi, E., Sallée, J.-B., and Durack, P. J. (submitted). Human-induced changes to the World Ocean water-
5 masses and their time of emergence. *Nat. Clim. Chang.* (submitted).
- 6 SIMIP Community (submitted). Arctic Sea Ice in CMIP6. *Geophys. Res. Lett.* (submitted).
- 7 Simmonds, I. (2015). Comparing and contrasting the behaviour of Arctic and Antarctic sea ice over the 35 year period
8 1979–2013. *Ann. Glaciol.* 56, 18–28. doi:10.3189/2015AoG69A909.
- 9 Simms, A. R., Lisiecki, L., Gebbie, G., Whitehouse, P. L., and Clark, J. F. (2019). Balancing the last glacial maximum
10 (LGM) sea-level budget. *Quat. Sci. Rev.* 205, 143–153.
- 11 Simpkins, G. R., Ciasto, L. M., and England, M. H. (2013). Observed variations in multidecadal Antarctic sea ice trends
12 during 1979–2012. *Geophys. Res. Lett.* 40, 3643–3648. doi:10.1002/grl.50715.
- 13 Simpkins, G. R., Ciasto, L. M., Thompson, D. W. J., and England, M. H. (2012). Seasonal relationships between large-
14 scale climate variability and antarctic sea ice concentration. *J. Clim.* 25, 5451–5469. doi:10.1175/JCLI-D-11-
15 00367.1.
- 16 Simpson, M. J. R., Milne, G. A., Huybrechts, P., and Long, A. J. (2009). Calibrating a glaciological model of the
17 Greenland ice sheet from the Last Glacial Maximum to present-day using field observations of relative sea level
18 and ice extent. *Quat. Sci. Rev.* 28, 1631–1657.
- 19 Sinclair, K. E., Bertler, N. A. N., Bowen, M. M., and Arrigo, K. R. (2014). Twentieth century sea-ice trends in the Ross
20 Sea from a high-resolution, coastal ice-core record. *Geophys. Res. Lett.* 41, 3510–3516.
21 doi:10.1002/2014GL059821.
- 22 Singh, D., Flanner, M. G., and Perket, J. (2015). The global land shortwave cryosphere radiative effect during the
23 MODIS era. *Cryosph.* 9, 2057–2070. doi:10.5194/tc-9-2057-2015.
- 24 Skliris, N., Marsh, R., Josey, S. A., Good, S. A., Liu, C., and Allan, R. P. (2014). Salinity changes in the World Ocean
25 since 1950 in relation to changing surface freshwater fluxes. *Clim. Dyn.* doi:10.1007/s00382-014-2131-7.
- 26 Skliris, N., Zika, J. D., Nurser, G., Josey, S. A., and Marsh, R. (2016). Global water cycle amplifying at less than the
27 Clausius-Clapeyron rate. *Sci. Rep.* doi:10.1038/srep38752.
- 28 Slangen, A. B. A., van de Wal, R. S. W., Reerink, T. J. T., de Winter, R. R. C., Hunter, J. R. J., R. J. R.,
29 Woodworth, P. L. P., et al. (2017a). The Impact of Uncertainties in Ice Sheet Dynamics on Sea-Level
30 Allowances at Tide Gauge Locations. *J. Mar. Sci. Eng.* 5, 1–24. doi:10.3390/jmse5020021.
- 31 Slangen, A. B. A., Carson, M., Katsman, C. A., van de Wal, R. S. W., Köhl, A., Vermeersen, L. L. A., et al. (2014a).
32 Projecting twenty-first century regional sea-level changes. *Clim. Change* 124, 317–332. doi:10.1007/s10584-014-
33 1080-9.
- 34 Slangen, A. B. A., Church, J. A., Agosta, C., Fettweis, X., Marzeion, B., and Richter, K. (2016). Anthropogenic forcing
35 dominates global mean sea-level rise since 1970. *Nat. Clim. Chang.* 6, 701–705. doi:10.1038/NCLIMATE2991.
- 36 Slangen, A. B. A., Church, J. A., Zhang, X., and Monselesan, D. (2014b). Detection and attribution of global mean
37 thermosteric sea-level change. *Geophys. Res. Lett.* 41, 5951–5959. doi:10.1002/2014GL061356.
- 38 Slangen, A. B. A., Church, J. A., Zhang, X., and Monselesan, D. P. (2015). The sea level response to external forcings
39 in historical simulations of CMIP5 climate models. *J. Clim.* 28, 8521–8539. doi:10.1175/JCLI-D-15-0376.1.
- 40 Slangen, A. B. A., Katsman, C. A., van de Wal, R. S. W., Vermeersen, L. L. A., and Riva, R. E. M. (2012). Towards
41 regional projections of twenty-first century sea-level change based on IPCC SRES scenarios. *Clim. Dyn.* 38,
42 1191–1209. doi:10.1007/s00382-011-1057-6.
- 43 Slangen, A. B. A., Meyssignac, B., Agosta, C., Champollion, N., Church, J. A., Fettweis, X., et al. (2017b). Evaluating
44 Model Simulations of Twentieth-Century Sea Level Rise. Part I: Global Mean Sea Level Change. *J. Clim.* 30,
45 8539–8563. doi:10.1175/JCLI-D-17-0110.1.
- 46 Slangen, A. B. A., Van De Wal, R. S. W., Wada, Y., and Vermeersen, L. L. A. (2014c). Comparing tide gauge
47 observations to regional patterns of sea-level change (1961–2003). *Earth Syst. Dyn.* 5. doi:10.5194/esd-5-243-
48 2014.
- 49 Slater, A. G., and Lawrence, D. M. (2013). Diagnosing present and future permafrost from climate models. *J. Clim.* 26,
50 5608–5623. doi:10.1175/JCLI-D-12-00341.1.
- 51 Slater, A. G., Lawrence, D. M., and Koven, C. D. (2017). Process-level model evaluation: A snow and heat transfer
52 metric. *Cryosph.* 11, 989–996. doi:10.5194/tc-11-989-2017.
- 53 Slater, D. A., Goldberg, D. N., Nienow, P. W., and Cowton, T. R. (2016). Scalings for Submarine Melting at Tidewater
54 Glaciers from Buoyant Plume Theory. *J. Phys. Oceanogr.* 46, 1839–1855. doi:10.1175/JPO-D-15-0132.1.
- 55 Slater, D. A., Straneo, F., Das, S. B., Richards, C. G., Wagner, T. J. W., and Nienow, P. W. (2018). Localized Plumes
56 Drive Front-Wide Ocean Melting of A Greenlandic Tidewater Glacier. *Geophys. Res. Lett.* 45, 12,350–12,358.
57 doi:10.1029/2018GL080763.
- 58 Slater, D. A., Straneo, F., Felikson, D., Little, C. M., Goelzer, H., Fettweis, X., et al. (2019a). Estimating Greenland
59 tidewater glacier retreat driven by submarine melting. *Cryosph.* 13, 2489–2509. doi:10.5194/tc-13-2489-2019.
- 60 Slater, D., Straneo, F., Felikson, D., Little, C., Goelzer, H., Fettweis, X., et al. (2019b). Past and future response of
61 Greenland’s tidewater glaciers to submarine melting. *Cryosph. Discuss.*, 1–32. doi:10.5194/tc-2019-98.

- Sloyan, B. M., and O’Kane, T. J. (2015). Drivers of decadal variability in the Tasman Sea. *J. Geophys. Res. Ocean.* 120, 3193–3210. doi:10.1002/2014JC010550.
- Smale, D. A., Wernberg, T., Oliver, E. C. J., Thomsen, M., Harvey, B. P., Straub, S. C., et al. (2019). Marine heatwaves threaten global biodiversity and the provision of ecosystem services. *Nat. Clim. Chang.* 9, 306. doi:10.1038/s41558-019-0412-1.
- Small, J., Bacmeister, J., Bailey, D., Baker, A., Bishop, S., Bryan, F., et al. (2014a). A new high-resolution global climate simulation using {Community Atmosphere Model version 5} and an eddy resolving ocean model. *Jour. Adv. Model. Earth Syst.*
- Small, R. J., Bacmeister, J., Bailey, D., Baker, A., Bishop, S., Bryan, F., et al. (2014b). A new synoptic scale resolving global climate simulation using the Community Earth System Model. *J. Adv. Model. Earth Syst.* 6, 1065–1094. doi:10.1002/2014MS000363.
- Small, R. J., Curchitser, E., Hedstrom, K., Kauffman, B., Large, W. G., Small, R. J., et al. (2015). The Benguela Upwelling System: Quantifying the Sensitivity to Resolution and Coastal Wind Representation in a Global Climate Model*. *J. Clim.* 28, 9409–9432. doi:10.1175/JCLI-D-15-0192.1.
- Smedsrud, L. H., Halvorsen, M. H., Stroeve, J. C., Zhang, R., and Kloster, K. (2017). Fram Strait sea ice export variability and September Arctic sea ice extent over the last 80 years. *Cryosphere* 11, 65–79. doi:10.5194/tc-11-65-2017.
- Smeed, D. A., Josey, S. A., Beaulieu, C., Johns, W. E., Moat, B. I., Frajka-Williams, E., et al. (2018). The North Atlantic Ocean Is in a State of Reduced Overturning. *Geophys. Res. Lett.* 45, 1527–1533. doi:10.1002/2017GL076350.
- Smeed, D. A., McCarthy, G. D., Cunningham, S. A., Frajka-Williams, E., Rayner, D., Johns, W. E., et al. (2014). Observed decline of the Atlantic meridional overturning circulation 2004–2012. *Ocean Sci.* 10, 29–38. doi:10.5194/os-10-29-2014.
- Smith, C. J., Forster, P. M., Allen, M., Leach, N., Millar, R. J., Passerello, G. A., et al. (2018). FAIR v1. 3: A simple emissions-based impulse response and carbon cycle model. *Geosci. Model Dev.* 11, 2273–2297.
- Smith, K. L., Polvani, L. M., and Marsh, D. R. (2012). Mitigation of 21st century Antarctic sea ice loss by stratospheric ozone recovery. *Geophys. Res. Lett.* 39, L20701–. doi:10.1029/2012GL053325.
- Smith, M. W., Quincey, D. J., Dixon, T., Bingham, R. G., Carrivick, J. L., Irvine-Fynn, T. D. L., et al. (2016). Aerodynamic roughness of glacial ice surfaces derived from high-resolution topographic data. *J. Geophys. Res. Earth Surf.* 121, 748–766. doi:10.1002/2015JF003759.
- Snauffer, A. M., Hsieh, W. W., and Cannon, A. J. (2016). Comparison of gridded snow water equivalent products with in situ measurements in British Columbia, Canada. *J. Hydrol.* 541, 714–726. doi:10.1016/j.jhydrol.2016.07.027.
- Snow, K., Hogg, A. M., Downes, S. M., Sloyan, B. M., Bates, M. L., and Griffies, S. M. (2015). Sensitivity of abyssal water masses to overflow parameterisations. *Ocean Model.* 89, 84–103. doi:https://doi.org/10.1016/j.ocemod.2015.03.004.
- Solomina, O. N., Bradley, R. S., Hodgson, D. A., Ivy-Ochs, S., Jomelli, V., Mackintosh, A. N., et al. (2015). Holocene glacier fluctuations. *Quat. Sci. Rev.* doi:10.1016/j.quascirev.2014.11.018.
- Solomina, O. N., Bradley, R. S., Jomelli, V., Geirsdottir, A., Kaufman, D. S., Koch, J., et al. (2016). Glacier fluctuations during the past 2000 years. *Quat. Sci. Rev.* doi:10.1016/j.quascirev.2016.04.008.
- Sommers, A. N., Rajaram, H., Weber, E. P., MacFerrin, M. J., Colgan, W. T., and Stevens, C. M. (2017). Inferring Firm Permeability from Pneumatic Testing: A Case Study on the Greenland Ice Sheet. *Front. Earth Sci.* 5, 20. doi:10.3389/feart.2017.00020.
- Somot, S., Sevault, F., Déqué, M., and Crépon, M. (2008). 21st century climate change scenario for the Mediterranean using a coupled atmosphere–ocean regional climate model. *Glob. Planet. Change* 63, 112–126. doi:10.1016/J.GLOPLACHA.2007.10.003.
- Song, Z., Latif, M., Park, W., and Zhang, Y. (2018). Influence of Model Bias on Simulating North Atlantic Sea Surface Temperature During the Mid-Pliocene. *Paleoceanogr. Paleoclimatology* 33, 884–893. doi:10.1029/2018PA003397.
- Sospedra-Alfonso, R., Melton, J. R., and Merryfield, W. J. (2015). Effects of temperature and precipitation on snowpack variability in the Central Rocky Mountains as a function of elevation. *Geophys. Res. Lett.* 42, 4429–4438. doi:10.1002/2015GL063898.
- Sospedra-Alfonso, R., and Merryfield, W. J. (2017). Influences of temperature and precipitation on historical and future snowpack variability over the Northern Hemisphere in the second generation Canadian Earth System Model. *J. Clim.* 30, 4633–4656. doi:10.1175/JCLI-D-16-0612.1.
- Soudzilovskaia, N. A., van Bodegom, P. M., and Cornelissen, J. H. C. (2013). Dominant bryophyte control over high-latitude soil temperature fluctuations predicted by heat transfer traits, field moisture regime and laws of thermal insulation. *Funct. Ecol.* 27, 1442–1454. doi:10.1111/1365-2435.12127.
- Sousa, M. C., Alvarez, I., Gomez-Gesteira, M., Dias, J. M., and others (2017). Why coastal upwelling is expected to increase along the western Iberian Peninsula over the next century? *Sci. Total Environ.* 592, 243–251.
- Spall, M. A., and Pickart, R. S. (2001). Where Does Dense Water Sink? A Subpolar Gyre Example*. *J. Phys. Oceanogr.* 31, 810–826. doi:10.1175/1520-0485(2001)031<0810:WDDWSA>2.0.CO;2.

- 1 Spence, P., Griffies, S. M., England, M. H., Hogg, A. M. C., Saenko, O. A., and Jourdain, N. C. (2014). Rapid
2 subsurface warming and circulation changes of Antarctic coastal waters by poleward shifting winds. *Geophys.*
3 *Res. Lett.* 41, 4601–4610. doi:10.1002/2014GL060613.
- 4 Spence, P., Holmes, R. M., Hogg, A. M., Griffies, S. M., Stewart, K. D., and England, M. H. (2017). Localized rapid
5 warming of West Antarctic subsurface waters by remote winds. *Nat. Clim. Chang.* 7, 595–603.
6 doi:10.1038/nclimate3335.
- 7 Spolaor, A., Vallelonga, P., Turetta, C., Maffezzoli, N., Cozzi, G., Gabrieli, J., et al. (2016). Canadian arctic sea ice
8 reconstructed from bromine in the Greenland NEEM ice core. *Sci. Rep.* 6, 33925. doi:10.1038/srep33925.
- 9 Spreen, G., Kwok, R., and Menemenlis, D. (2011). Trends in Arctic sea ice drift and role of wind forcing: 1992–2009.
10 *Geophys. Res. Lett.* 38. doi:10.1029/2011GL048970.
- 11 Stabeno, P. J., and Bell, S. W. (2019). Extreme Conditions in the Bering Sea (2017–2018): Record-Breaking Low Sea-
12 Ice Extent. *Geophys. Res. Lett.* doi:10.1029/2019GL083816.
- 13 Stammer, D., Bracco, A., AchutaRao, K., Beal, L., Bindoff, N. L., Braconnot, P., et al. (2019). Ocean Climate
14 Observing Requirements in Support of Climate Research and Climate Information. *Front. Mar. Sci.* 6, 444.
15 doi:10.3389/fmars.2019.00444.
- 16 Stammerjohn, S., and Maksym, T. (2016). “Gaining (and losing) Antarctic sea ice: variability, trends and mechanisms,”
17 in *Sea Ice* (Chichester, UK: John Wiley & Sons, Ltd), 261–289. doi:10.1002/9781118778371.ch10.
- 18 Stammerjohn, S., Massom, R., Rind, D., and Martinson, D. (2012). Regions of rapid sea ice change: An inter-
19 hemispheric seasonal comparison. *Geophys. Res. Lett.* 39, L06501. doi:10.1029/2012GL050874.
- 20 Stap, L. B., van de Wal, R. S. W., de Boer, B., Köhler, P., Hoencamp, J. H., Lohmann, G., et al. (2018). Modeled
21 Influence of Land Ice and CO₂ on Polar Amplification and Paleoclimate Sensitivity During the Past 5 Million
22 Years. *Paleoceanogr. Paleoclimatology* 33, 381–394. doi:10.1002/2017PA003313.
- 23 Stap, L. B., Van De Wal, R. S. W., De Boer, B., Bintanja, R., and Lourens, L. J. (2017). The influence of ice sheets on
24 temperature during the past 38 million years inferred from a one-dimensional ice sheet-climate model. *Clim. Past*
25 13, 1243–1257. doi:10.5194/cp-13-1243-2017.
- 26 Staten, P. W., Lu, J., Grise, K. M., Davis, S. M., and Birner, T. (2018). Re-examining tropical expansion. *Nat. Clim.*
27 *Chang.* 8, 768–775. doi:10.1038/s41558-018-0246-2.
- 28 Steele, M., and Ermold, W. (2015). Loitering of the retreating sea ice edge in the Arctic Seas. *J. Geophys. Res. Ocean.*
29 120, 7699–7721. doi:10.1002/2015JC011182.
- 30 Steger, C. R., Reijmer, C. H., and van den Broeke, M. R. (2017). The modelled liquid water balance of the Greenland
31 Ice Sheet. *Cryosph.* 11, 2507–2526. doi:10.5194/tc-11-2507-2017.
- 32 Stein, R., Fahl, K., Gierz, P., Niessen, F., and Lohmann, G. (2017a). Arctic Ocean sea ice cover during the penultimate
33 glacial and the last interglacial. *Nat. Commun.* 8. doi:10.1038/s41467-017-00552-1.
- 34 Stein, R., Fahl, K., Schade, I., Manerung, A., Wassmuth, S., Niessen, F., et al. (2017b). Holocene variability in sea ice
35 cover, primary production, and Pacific-Water inflow and climate change in the Chukchi and East Siberian Seas
36 (Arctic Ocean). *J. Quat. Sci.* 32, 362–379. doi:10.1002/jqs.2929.
- 37 Stephenson, S. R., and Smith, L. C. (2015). Influence of climate model variability on projected Arctic shipping futures.
38 *Earth's Futur.* 3, 331–343. doi:10.1002/2015EF000317.
- 39 Stewart, A. L., and Hogg, A. M. (2017). Reshaping the Antarctic Circumpolar Current via Antarctic Bottom Water
40 Export. *J. Phys. Oceanogr.* doi:10.1175/JPO-D-17-0007.1.
- 41 Stewart, A. L., Klocker, A., and Menemenlis, D. (2018). Circum-Antarctic Shoreward Heat Transport Derived From an
42 Eddy- and Tide-Resolving Simulation. *Geophys. Res. Lett.* 45, 834–845. doi:10.1002/2017GL075677.
- 43 Stewart, A. L., and Thompson, A. F. (2013). Connecting Antarctic Cross-Slope Exchange with Southern Ocean
44 Overturning. *J. Phys. Oceanogr.* 43, 1453–1471. doi:10.1175/JPO-D-12-0205.1.
- 45 Stewart, A. L., and Thompson, A. F. (2015). Eddy-mediated transport of warm Circumpolar Deep Water across the
46 Antarctic Shelf Break. *Geophys. Res. Lett.* 42, 432–440. doi:10.1002/2014GL062281.
- 47 Stibal, M., Box, J. E., Cameron, K. A., Langen, P. L., Yallop, M. L., Mottram, R. H., et al. (2017). Algae Drive
48 Enhanced Darkening of Bare Ice on the Greenland Ice Sheet. *Geophys. Res. Lett.* 44, 11,463–11,471.
49 doi:10.1002/2017GL075958.
- 50 Stocker, T. F., and Johnsen, S. J. (2003). A minimum thermodynamic model for the bipolar seesaw. *Paleoceanography*
51 18. doi:10.1029/2003PA000920.
- 52 Stokes, C. R., Margold, M., Clark, C. D., and Tarasov, L. (2016). Ice stream activity scaled to ice sheet volume during
53 Laurentide Ice Sheet deglaciation. *Nature* 530, 322–326. doi:10.1038/nature16947.
- 54 Stone, E. J., Lunt, D. J., Annan, J. D., and Hargreaves, J. C. (2013). Quantification of the Greenland ice sheet
55 contribution to Last Interglacial sea level rise. *Clim. Past* 9, 621–639. doi:10.5194/cp-9-621-2013.
- 56 Stopa, J. E., Arduin, F., and Girard-Arduin, F. (2016). Wave climate in the Arctic 1992–2014: seasonality and trends.
57 *Cryosph.* 10, 1605–1629. doi:10.5194/tc-10-1605-2016.
- 58 Storlazzi, C. D., Gingerich, S. B., Van Dongeren, A., Cheriton, O. M., Swarzenski, P. W., Quataert, E., et al. (2018).
59 Most atolls will be uninhabitable by the mid-21st century because of sea-level rise exacerbating wave-driven
60 flooding. *Sci. Adv.* doi:10.1126/sciadv.aap9741.
- 61 Stouffer, R. J., Yin, J., Gregory, J. M., Dixon, K. W., Spelman, M. J., Hurlin, W., et al. (2006). Investigating the Causes

- of the Response of the Thermohaline Circulation to Past and Future Climate Changes. *J. Clim.* 19, 1365–1387. doi:10.1175/JCLI3689.1.
- Straneo, F., and Cenedese, C. (2015). The Dynamics of Greenland's Glacial Fjords and Their Role in Climate. *Ann. Rev. Mar. Sci.* 7, 89–112. doi:10.1146/annurev-marine-010213-135133.
- Straneo, F., Hamilton, G. S., Sutherland, D. A., Stearns, L. A., Davidson, F., Hammill, M. O., et al. (2010). Rapid circulation of warm subtropical waters in a major glacial fjord in East Greenland. *Nat. Geosci.* doi:10.1038/ngeo764.
- Straneo, F., and Heimbach, P. (2013). North Atlantic warming and the retreat of Greenland's outlet glaciers. *Nature* 504, 36. Available at: <https://doi.org/10.1038/nature12854>.
- Straneo, F., Heimbach, P., Sergienko, O., Hamilton, G., Catania, G., Griffies, S., et al. (2013). Challenges to understanding the dynamic response of Greenland's marine terminating glaciers to oceanic and atmospheric forcing. *Bull. Am. Meteorol. Soc.* doi:10.1175/BAMS-D-12-00100.1.
- Strauss, B. H., Kopp, R. E., Sweet, W. V., and Bittermann, K. (2016). Unnatural coastal floods: Sea level rise and the human fingerprint on U.S. floods since 1950.
- Streletskiy, D. A., Shiklomanov, N. I., Little, J. D., Nelson, F. E., Brown, J., Nyland, K. E., et al. (2017). Thaw Subsidence in Undisturbed Tundra Landscapes, Barrow, Alaska, 1962–2015. *Permafrost Periglacial Process.* 28, 566–572. doi:10.1002/ppp.1918.
- Stroeve, J., Barrett, A., Serreze, M., and Schweiger, A. (2014). Using records from submarine, aircraft and satellites to evaluate climate model simulations of Arctic sea ice thickness. *Cryosphere* 8, 1839–1854. doi:10.5194/tc-8-1839-2014.
- Stroeve, J. C., Kattsov, V., Barrett, A., Serreze, M., Pavlova, T., Holland, M., et al. (2012). Trends in Arctic sea ice extent from CMIP5, CMIP3 and observations. *Geophys. Res. Lett.* 39. doi:10.1029/2012GL052676.
- Stroeve, J., and Notz, D. (2015). Insights on past and future sea-ice evolution from combining observations and models. *Glob. Planet. Change* 135, 119–132. doi:10.1016/j.gloplacha.2015.10.011.
- Stroeve, J., and Notz, D. (2018). Changing state of Arctic sea ice across all seasons. *Environ. Res. Lett.* 13, 103001. doi:10.1088/1748-9326/aade56.
- Stuecker, M. F., Bitz, C. M., and Armour, K. C. (2017). Conditions leading to the unprecedented low Antarctic sea ice extent during the 2016 austral spring season. *Geophys. Res. Lett.* 44, 9008–9019. doi:10.1002/2017GL074691.
- Stuhne, G. R., and Peltier, W. R. (2015). Reconciling the ICE-6G_C reconstruction of glacial chronology with ice sheet dynamics: The cases of Greenland and Antarctica. *J. Geophys. Res.* 120, 1841–1865.
- Su, Z. (2017). Preconditioning of Antarctic maximum sea ice extent by upper ocean stratification on a seasonal timescale. *Geophys. Res. Lett.* 44, 6307–6315. doi:10.1002/2017GL073236.
- Sugimoto, S., Hanawa, K., Watanabe, T., Suga, T., and Xie, S.-P. P. (2017). Enhanced warming of the subtropical mode water in the North Pacific and North Atlantic. *Nat. Clim. Change* 7, 656. doi:10.1038/nclimate3371.
- Sullivan, P. P., and McWilliams, J. C. (2010). Dynamics of Winds and Currents Coupled to Surface Waves. *Annu. Rev. Fluid Mech.* 42, 19–42. doi:10.1146/annurev-fluid-121108-145541.
- Sun, C., Kucharski, F., Li, J., Jin, F. F., Kang, I. S., and Ding, R. (2017a). Western tropical Pacific multidecadal variability forced by the Atlantic multidecadal oscillation. *Nat. Commun.* doi:10.1038/ncomms15998.
- Sun, Q., Whitney, M. M., Bryan, F. O., and Tseng, Y. (2017b). A box model for representing estuarine physical processes in Earth system models. *Ocean Model.* 112, 139–153.
- Sun, S., Pattyn, F., Simon, E. G., Albrecht, T., Stephen Cornford, R. C., Dumas, C., et al. (submitted). Antarctic ice sheet response to sudden and sustained ice 2 shelf collapse (ABUMIP). *J. Glaciol.* (submitted), 4–11.
- Supply, A., Boutin, J., Vergely, J.-L., Martin, N., Hasson, A., Reverdin, G., et al. (2018). Precipitation Estimates from SMOS Sea-Surface Salinity. *Q. J. R. Meteorol. Soc.* 144, 103–119. doi:10.1002/qj.3110.
- Sutter, J., Gierz, P., Grosfeld, K., Thoma, M., and Lohmann, G. (2016). Ocean temperature thresholds for Last Interglacial West Antarctic Ice Sheet collapse. *Geophys. Res. Lett.* 43, 2675–2682. doi:10.1002/2016GL067818.
- Sutterley, T. C., Velicogna, I., Rignot, E., Mouginot, J., Flament, T., Van Den Broeke, M. R., et al. (2014). Mass loss of the Amundsen Sea Embayment of West Antarctica from four independent techniques. *Geophys. Res. Lett.* 41, 8421–8428. doi:10.1002/2014GL061940.
- Svensson, C., and Jones, D. A. (2004). Dependence between sea surge, river flow and precipitation in south and west Britain. *Hydrol. Earth Syst. Sci.* 8, 973–992.
- Swart, N. C., Fyfe, J. C., Hawkins, E., Kay, J. E., and Jahn, A. (2015). Influence of internal variability on Arctic sea-ice trends. *Nat. Clim. Change* 5, 86–89. doi:10.1038/nclimate2483.
- Swart, N. C., Gille, S. T., Fyfe, J. C., and Gillett, N. P. (2018). Recent Southern Ocean warming and freshening driven by greenhouse gas emissions and ozone depletion. *Nat. Geosci.* 11, 836–841. doi:10.1038/s41561-018-0226-1.
- Swingedouw, D., Braconnot, P., Delecluse, P., Guilyardi, E., and Marti, O. (2006). The impact of global freshwater forcing on the thermohaline circulation: adjustment of North Atlantic convection sites in a CGCM. *Clim. Dyn.* 28, 291–305. doi:10.1007/s00382-006-0171-3.
- Swingedouw, D., Rodehacke, C. B., Behrens, E., Menary, M., Olsen, S. M., Gao, Y., et al. (2013). Decadal fingerprints of freshwater discharge around Greenland in a multi-model ensemble. *Clim. Dyn.* doi:10.1007/s00382-012-1479-9.

- Sylla, A., Mignot, J., Capet, X., and Gaye, A. T. (2019). Weakening of the Senegalo-Mauritania Upwelling System under climate change, Climate dynamics. *Clim. Dyn.* accepted.
- Takahashi, C., and Watanabe, M. (2016). Pacific trade winds accelerated by aerosol forcing over the past two decades. *Nat. Clim. Chang.* doi:10.1038/nclimate2996.
- Takayabu, I., Hibino, K., Sasaki, H., Shiogama, H., Mori, N., Shibutani, Y., et al. (2015). Climate change effects on the worst-case storm surge: A case study of Typhoon Haiyan. *Environ. Res. Lett.* doi:10.1088/1748-9326/10/6/064011.
- Talke, S. A., and Jay, D. A. (2013). Nineteenth century North American and Pacific tidal data: Lost or just forgotten? *J. Coast. Res.* 29, 118–127.
- Talke, S. A., and Jay, D. A. (2020). Changing Tides: The Role of Natural and Anthropogenic Factors. *Ann. Rev. Mar. Sci.* 12. doi:10.1146/annurev-marine-010419-010727.
- Talke, S. A., Orton, P., and Jay, D. A. (2014). Increasing storm tides in New York Harbor, 1844–2013. *Geophys. Res. Lett.* doi:10.1002/2014GL059574.
- Talley, L. D. (2013). Closure of the Global Overturning Circulation Through the Indian, Pacific, and Southern Oceans: Schematics and Transports. *Oceanography* 26. Available at: <https://doi.org/10.5670/oceanog.2013.07>.
- Tan, N., Ladant, J. B., Ramstein, G., Dumas, C., Bachem, P., and Jansen, E. (2018). Dynamic Greenland ice sheet driven by pCO₂ variations across the Pliocene Pleistocene transition. *Nat. Commun.* doi:10.1038/s41467-018-07206-w.
- Tandon, N. F., Kushner, P. J., Docquier, D., Wettstein, J. J., and Li, C. (2018). Reassessing Sea Ice Drift and Its Relationship to Long-Term Arctic Sea Ice Loss in Coupled Climate Models. *J. Geophys. Res. Ocean.* 123, 4338–4359. doi:10.1029/2017JC013697.
- Tarasov, L., Dyke, A. S., Neal, R. M., and Peltier, W. R. (2012). A data-calibrated distribution of deglacial chronologies for the North American ice complex from glaciological modeling. *Earth Planet. Sci. Lett.* 315, 30–40.
- Tedesco, M., Doherty, S., Fettweis, X., Alexander, P., Jeyaratnam, J., and Stroeve, J. (2016). The darkening of the Greenland ice sheet: trends, drivers, and projections (1981–2100). *Cryosph.* 10, 477–496. doi:10.5194/tc-10-477-2016.
- Tedesco, M., and Fettweis, X. (2019). Unprecedented atmospheric conditions (1948–2019) drive the 2019 exceptional melting season over the Greenland ice sheet. *Cryosph. Discuss.* 2019, 1–26. doi:10.5194/tc-2019-254.
- Tedesco, M., Fettweis, X., Mote, T., Wahr, J., Alexander, P., Box, J. E., et al. (2013). Evidence and analysis of 2012 Greenland records from spaceborne observations, a regional climate model and reanalysis data. *Cryosph.* 7, 615–630. doi:10.5194/tc-7-615-2013.
- Tedstone, A. J., Bamber, J. L., Cook, J. M., Williamson, C. J., Fettweis, X., Hodson, A. J., et al. (2017). Dark ice dynamics of the south-west Greenland Ice Sheet. *Cryosph.* 11, 2491–2506. doi:10.5194/tc-11-2491-2017.
- Terray, P., Masson, S., Prodhomme, C., Roxy, M. K., and Sooraj, K. P. (2016). Impacts of Indian and Atlantic oceans on ENSO in a comprehensive modeling framework. *Clim. Dyn.* 46, 2507–2533. doi:10.1007/s00382-015-2715-x.
- Terzago, S., von Hardenberg, J., Palazzi, E., and Provenzale, A. (2014). Snowpack Changes in the Hindu Kush–Karakoram–Himalaya from CMIP5 Global Climate Models. *J. Hydrometeorol.* 15, 2293–2313. doi:10.1175/JHM-D-13-0196.1.
- Thackeray, C. W., and Fletcher, C. G. (2015). Snow albedo feedback: Current knowledge, importance, outstanding issues and future directions. *Prog. Phys. Geogr.* 40, 392–408. doi:10.1177/0309133315620999.
- Thackeray, C. W., Fletcher, C. G., and Derksen, C. (2015). Quantifying the skill of CMIP5 models in simulating seasonal albedo and snow cover evolution. *J. Geophys. Res.* 120, 5831–5849. doi:10.1002/2015JD023325.
- Thackeray, C. W., Fletcher, C. G., Mudryk, L. R., and Derksen, C. (2016). Quantifying the uncertainty in historical and future simulations of Northern Hemisphere spring snow cover. *J. Clim.* 29, 8647–8663. doi:10.1175/JCLI-D-16-0341.1.
- Thackeray, C. W., Qu, X., and Hall, A. (2018). Why Do Models Produce Spread in Snow Albedo Feedback? *Geophys. Res. Lett.* doi:10.1029/2018GL078493.
- The IMBIE Team (2018). Mass balance of the Antarctic Ice Sheet from 1992 to 2017. *Nature* 558, 219–222. doi:10.1038/s41586-018-0179-y.
- The IMBIE Team (2019). Mass balance of the Greenland Ice Sheet from 1992 to 2018. *Nature*. doi:10.1038/s41586-019-1855-2.
- Thomas, E. R., and Abram, N. J. (2016). Ice core reconstruction of sea ice change in the Amundsen-Ross Seas since 1702 A.D. *Geophys. Res. Lett.* 43, 5309–5317. doi:10.1002/2016GL068130.
- Thomas, E. R., Hosking, J. S., Tuckwell, R. R., Warren, R. A., and Ludlow, E. C. (2015). Twentieth century increase in snowfall in coastal West Antarctica. *Geophys. Res. Lett.* 42, 9387–9393. doi:10.1002/2015GL065750.
- Thomas, R. H., and Bentley, C. R. (1978). A Model for Holocene Retreat of the West Antarctic Ice Sheet. *Quat. Res.* 10, 150–170. doi:10.1016/0033-5894(78)90098-4.
- Thompson, A. F., Stewart, A. L., Spence, P., and Heywood, K. J. (2018). The Antarctic Slope Current in a Changing Climate. *Rev. Geophys.* 56, 741–770. doi:10.1029/2018RG000624.
- Thompson, D. W. J. J., Solomon, S., Kushner, P. J., England, M. H., Grise, K. M., and Karoly, D. J. (2011). Signatures

- of the Antarctic ozone hole in Southern Hemisphere surface climate change. *Nat. Geosci.* 4, 741–749. doi:10.1038/ngeo1296.
- Thornalley, D. J. R., Oppo, D. W., Ortega, P., Robson, J. I., Brierley, C. M., Davis, R., et al. (2018). Anomalous weak Labrador Sea convection and Atlantic overturning during the past 150 years. *Nature*. doi:10.1038/s41586-018-0007-4.
- Tiel, M., Kohn, I., Van Loon, A. F., and Stahl, K. (2019). The compensating effect of glaciers: characterising the relation between interannual streamflow variability and glacier cover. *Hydrol. Process.* doi:10.1002/hyp.13603.
- Tietsche, S., Notz, D., Jungclaus, J. H., and Marotzke, J. (2011). Recovery mechanisms of Arctic summer sea ice. *Geophys. Res. Lett.* 38. doi:10.1029/2010GL045698.
- Tilling, R. L., Ridout, A., Shepherd, A., and Wingham, D. J. (2015). Increased Arctic sea ice volume after anomalously low melting in 2013. *Nat. Geosci.* 8, 643–646. doi:10.1038/ngeo2489.
- Timko, P. G., Arbic, B. K., Richman, J. G., Scott, R. B., Metzger, E. J., and Wallcraft, A. J. (2013). Skill testing a three-dimensional global tide model to historical current meter records. *J. Geophys. Res. Ocean.* 118, 6914–6933. doi:10.1002/2013JC009071.
- Timmermann, R., and Goeller, S. (2017). Response to Filchner–Ronne Ice Shelf cavity warming in a coupled ocean–ice sheet model -- Part 1: The ocean perspective. *Ocean Sci.* 13, 765–776. doi:10.5194/os-13-765-2017.
- Timmermann, R., and Hellmer, H. H. (2013). Southern Ocean warming and increased ice shelf basal melting in the twenty-first and twenty-second centuries based on coupled ice-ocean finite-element modelling. *Ocean Dyn.* 63, 1011–1026. doi:10.1007/s10236-013-0642-0.
- Timmermans, B., Stone, D., Wehner, M., and Krishnan, H. (2017). Impact of tropical cyclones on modeled extreme wind-wave climate. *Geophys. Res. Lett.* doi:10.1002/2016GL071681.
- Timmermans, M. L. (2015). The impact of stored solar heat on Arctic sea ice growth. *Geophys. Res. Lett.* 42, 6399–6406. doi:10.1002/2015GL064541.
- Timmermans, M. L., Proshutinsky, A., Golubeva, E., Jackson, J. M., Krishfield, R., McCall, M., et al. (2014). Mechanisms of Pacific Summer Water variability in the Arctic’s Central Canada Basin. *J. Geophys. Res. Ocean.* doi:10.1002/2014JC010273.
- Tinker, J., Lowe, J., Pardaens, A., Holt, J., and Barciela, R. (2016). Uncertainty in climate projections for the 21st century northwest European shelf seas. *Prog. Oceanogr.* 148, 56–73. doi:10.1016/J.POCEAN.2016.09.003.
- Titchner, H. A., and Rayner, N. A. (2014). The Met Office Hadley Centre sea ice and sea surface temperature data set, version 2: 1. Sea ice concentrations. *J. Geophys. Res. Atmos.* 119, 2864–2889. doi:10.1002/2013JD020316.
- Tomita, H., Hihara, T., Kako, S., Kubota, M., and Kutsuwada, K. (2019). An introduction to J-OFURO3, a third-generation Japanese ocean flux data set using remote-sensing observations. *J. Oceanogr.* 75, 171–194. doi:10.1007/s10872-018-0493-x.
- Treat, C. C., and Jones, M. C. (2018). Near-surface permafrost aggradation in Northern Hemisphere peatlands shows regional and global trends during the past 6000 years. *Holocene* 28, 998–1010. doi:10.1177/0959683617752858.
- Trenberth, K. E., Fasullo, J. T., von Schuckmann, K., and Cheng, L. (2016). Insights into Earth’s Energy Imbalance from Multiple Sources. *J. Clim.* 29, 7495–7505. doi:10.1175/JCLI-D-16-0339.1.
- Trenberth, K. E., Zhang, Y., Fasullo, J. T., and Cheng, L. (2019). Observation-Based Estimates of Global and Basin Ocean Meridional Heat Transport Time Series. *J. Clim.* 32, 4567–4583. doi:10.1175/JCLI-D-18-0872.1.
- Trombotto-Liaudat, D., and Bottegat, E. (2019). Recent evolution of the active layer in the Morenas Coloradas rock glacier, Central Andes, Mendoza, Argentina and its relation with kinematics. *Cuad. Investig. Geográfica.* doi:10.18172/cig.3946.
- Trotta, F., Pinardi, N., Fenu, E., Grandi, A., and Lyubartsev, V. (2017). Multi-nest high-resolution model of submesoscale circulation features in the Gulf of Taranto. *Ocean Dyn.* 67, 1609–1625. doi:10.1007/s10236-017-1110-z.
- Trusel, L. D., Frey, K. E., Das, S. B., Karnauskas, K. B., Kuipers Munneke, P., Van Meijgaard, E., et al. (2015). Divergent trajectories of Antarctic surface melt under two twenty-first-century climate scenarios. *Nat. Geosci.* 8, 927–932. doi:10.1038/ngeo2563.
- Tsai, V. C., Stewart, A. L., and Thompson, A. F. (2015). Marine ice-sheet profiles and stability under Coulomb basal conditions. *J. Glaciol.* 61, 205–215. doi:10.3189/2015JoG14J221.
- Tschudi, M. A., Stroeve, J. C., and Stewart, J. S. (2016). Relating the age of Arctic sea ice to its thickness, as measured during nasa’s ICESat and IceBridge campaigns. *Remote Sens.* doi:10.3390/rs8060457.
- Tsujino, H., Urakawa, L. S., Griffies, S. M., Danabasoglu, G., Adcroft, A. J., Amaral, A. E., et al. (submitted). Evaluation of global ocean–sea-ice model simulations based on the experimental protocols of the Ocean Model Intercomparison Project phase 2 (OMIP-2). *Geosci. Model Dev. Discuss.* (submitted).
- Turner, J., Bracegirdle, T., Phillips, T., Marshall, G., and Hosking, S. (2013). An Initial Assessment of Antarctic Sea Ice Extent in the CMIP5 Models. *J. Clim.* 26, 1473–1484.
- Turner, J., Hosking, J. S., Marshall, G. J., Phillips, T., and Bracegirdle, T. J. (2016). Antarctic sea ice increase consistent with intrinsic variability of the Amundsen Sea Low. *Clim. Dyn.* 46, 2391–2402. doi:10.1007/s00382-015-2708-9.
- Turner, J., Orr, A., Gudmundsson, G. H., Jenkins, A., Bingham, R. G., Hillenbrand, C. D., et al. (2017a). Atmosphere-

- ocean-ice interactions in the Amundsen Sea Embayment, West Antarctica. *Rev. Geophys.* 55, 235–276. doi:10.1002/2016RG000532.
- Turner, J., Phillips, T., Marshall, G. J., Hosking, J. S., Pope, J. O., Bracegirdle, T. J., et al. (2017b). Unprecedented springtime retreat of Antarctic sea ice in 2016. *Geophys. Res. Lett.* 44, 6868–6875. doi:10.1002/2017GL073656.
- Uehara, K., Scourse, J. D., Horsburgh, K. J., Lambeck, K., and Purcell, A. P. (2006). Tidal evolution of the northwest European shelf seas from the Last Glacial Maximum to the present. *J. Geophys. Res. Ocean.* doi:10.1029/2006JC003531.
- Uotila, P., Holland, P. R., Vihma, T., Marsland, S. J., and Kimura, N. (2014). Is realistic Antarctic sea-ice extent in climate models the result of excessive ice drift? *Ocean Model.* 79, 33–42. doi:10.1016/j.ocemod.2014.04.004.
- Vaks, A., Gutareva, O. S., Breitenbach, S. F. M., Avirmed, E., Mason, A. J., Thomas, A. L., et al. (2013). Speleothems reveal 500,000-year history of Siberian permafrost. *Science* 340, 183–186. doi:10.1126/science.1228729.
- Valdes, P. (2011). Built for stability. *Nat. Geosci.* 4, 414–416. doi:10.1038/ngeo1200.
- Valdivieso, M., Haines, K., Balmaseda, M., Chang, Y.-S., Drevillon, M., Ferry, N., et al. (2017). An assessment of air–sea heat fluxes from ocean and coupled reanalyses. *Clim. Dyn.* 49, 983–1008. doi:10.1007/s00382-015-2843-3.
- Välisuo, I., Vihma, T., Pirazzini, R., and Schäfer, M. (2018). Interannual Variability of Atmospheric Conditions and Surface Melt in Greenland in 2000–2014. *J. Geophys. Res. Atmos.* 123, 10,410–443,463. doi:10.1029/2018JD028445.
- van Angelen, J. H., van den Broeke, M. R., Wouters, B., and Lenaerts, J. T. M. (2014). Contemporary (1960–2012) Evolution of the Climate and Surface Mass Balance of the Greenland Ice Sheet. *Surv. Geophys.* 35, 1155–1174. doi:10.1007/s10712-013-9261-z.
- Van De Berg, W. J., Van Den Broeke, M., Ettema, J., Van Meijgaard, E., and Kaspar, F. (2011). Significant contribution of insolation to Eemian melting of the Greenland ice sheet. *Nat. Geosci.* doi:10.1038/ngeo1245.
- van den Berk, J., Drijfhout, S. S., and Hazeleger, W. (2019). Atlantic salinity budget in response to Northern and Southern Hemisphere ice sheet discharge. *Clim. Dyn.* 52, 5249–5267. doi:10.1007/s00382-018-4444-4.
- van den Broeke, M., Box, J., Fettweis, X., Hanna, E., Noël, B., Tedesco, M., et al. (2017). Greenland Ice Sheet Surface Mass Loss: Recent Developments in Observation and Modeling. *Curr. Clim. Chang. Reports* 3, 345–356. doi:10.1007/s40641-017-0084-8.
- Van Den Broeke, M. R., Enderlin, E. M., Howat, I. M., Kuipers Munneke, P., Noël, B. P. Y., Jan Van De Berg, W., et al. (2016). On the recent contribution of the Greenland ice sheet to sea level change. *Cryosph.* 10, 1933–1946. doi:https://doi.org/10.5194/tc-10-1933-2016.
- van den Hurk, B., Kim, H., Krinner, G., Seneviratne, S. I., Derksen, C., Oki, T., et al. (2016). LS3MIP (v1.0) contribution to CMIP6: The Land Surface, Snow and Soil moisture Model Intercomparison Project - Aims, setup and expected outcome. *Geosci. Model Dev.* 9, 2809–2832. doi:10.5194/gmd-9-2809-2016.
- van den Hurk, B., van Meijgaard, E., de Valk, P., van Heeringen, K.-J., and Gooijer, J. (2015). Analysis of a compounding surge and precipitation event in the Netherlands. *Environ. Res. Lett.* 10, 35001.
- van Kampenhout, L., Lenaerts, J. T. M., Lipscomb, W. H., Sacks, W. J., Lawrence, D. M., Slater, A. G., et al. (2017). Improving the Representation of Polar Snow and Firn in the Community Earth System Model. *J. Adv. Model. Earth Syst.* 9, 2583–2600. doi:10.1002/2017MS000988.
- van Kampenhout, L., Lenaerts, J., Lipscomb, W., Lhermitte, S., Noel, B., Vizcaino, M., et al. (submitted). Greenland ice sheet climate and surface mass balance in CESM2. *J. Geophys. Res. F Earth Surf.* (submitted).
- Van Tricht, K., Lhermitte, S., Lenaerts, J. T. M., Gorodetskaya, I. V., L'Ecuyer, T. S., Noël, B., et al. (2016). Clouds enhance Greenland ice sheet meltwater runoff. *Nat. Commun.* 7, 10266. doi:10.1038/ncomms10266.
- Van Wessem, J. M., Van De Berg, W. J., Noël, B. P. Y., Van Meijgaard, E., Amory, C., Birnbaum, G., et al. (2018). Modelling the climate and surface mass balance of polar ice sheets using RACMO2 - Part 2: Antarctica (1979–2016). *Cryosphere* 12, 1479–1498. doi:10.5194/tc-12-1479-2018.
- Vandecrux, B., MacFerrin, M., Machguth, H., Colgan, W. T., van As, D., Heilig, A., et al. (2019). Firn data compilation reveals widespread decrease of firn air content in western Greenland. *Cryosph.* 13, 845–859. doi:10.5194/tc-13-845-2019.
- Varela, R., Álvarez, I., Santos, F., Gómez-Gesteira, M., and others (2015). Has upwelling strengthened along worldwide coasts over 1982–2010? *Sci. Rep.* 5, 10016.
- Vecchi, G. A., and Soden, B. J. (2007). Global warming and the weakening of the tropical circulation. *J. Clim.* doi:10.1175/JCLI4258.1.
- Vecchi, G. A., Soden, B. J., Wittenberg, A. T., Held, I. M., Leetmaa, A., and Harrison, M. J. (2006). Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature* 441, 73–76. doi:10.1038/nature04744.
- Vecchi, G., Clement, A., and Soden, B. J. (2008). Examining the Tropical Pacific's Response to Global Warming. *Eos (Washington, DC)*.
- Veh, G., Korup, O., von Specht, S., Roessner, S., and Walz, A. (2019a). Unchanged frequency of moraine-dammed glacial lake outburst floods in the Himalaya. *Nat. Clim. Chang.* 9, 379–383. doi:10.1038/s41558-019-0437-5.
- Veh, G., Korup, O., and Walz, A. (2019b). Hazard from Himalayan glacier lake outburst floods. 1–6. doi:10.1073/pnas.1914898117.

- 1 Velicogna, I., Sutterley, T. C., and Van Den Broeke, M. R. (2014). Regional acceleration in ice mass loss from
2 Greenland and Antarctica using GRACE time-variable gravity data. *Geophys. Res. Lett.* 41, 8130–8137.
- 3 Verfaillie, D., Lafaysse, M., Déqué, M., Eckert, N., Lejeune, Y., and Morin, S. (2018). Multi-component ensembles of
4 future meteorological and natural snow conditions for 1500 m altitude in the Chartreuse mountain range,
5 Northern French Alps. *Cryosphere* 12, 1249–1271. doi:10.5194/tc-12-1249-2018.
- 6 Vihma, T., Tisler, P., and Uotila, P. (2012). Atmospheric forcing on the drift of Arctic sea ice in 1989–2009. *Geophys.*
7 *Res. Lett.* 39. doi:10.1029/2011GL050118.
- 8 Vijay, S., Khan, S. A., Kusk, A., Solgaard, A. M., Moon, T., and Bjørk, A. A. (2019). Resolving Seasonal Ice Velocity
9 of 45 Greenlandic Glaciers With Very High Temporal Details. *Geophys. Res. Lett.* 46, 1485–1495.
10 doi:10.1029/2018GL081503.
- 11 Vionnet, V., Brun, E., Morin, S., Boone, A., Faroux, S., Le Moigne, P., et al. (2012). The detailed snowpack scheme
12 Crocus and its implementation in SURFEX v7.2. *Geosci. Model Dev.* doi:10.5194/gmd-5-773-2012.
- 13 Vitousek, S., Barnard, P. L., Fletcher, C. H., Frazer, N., Erikson, L., and Storlazzi, C. D. (2017). Doubling of coastal
14 flooding frequency within decades due to sea-level rise. *Sci. Rep.* 7, 1–9. doi:10.1038/s41598-017-01362-7.
- 15 Vizcaino, M. (2014). Ice sheets as interactive components of Earth System Models: progress and challenges. *Wiley*
16 *Interdiscip. Rev. Clim. Chang.* 5, 557–568. doi:10.1002/wcc.285.
- 17 Vizcaino, M., Mikolajewicz, U., Gröger, M., Maier-Reimer, E., Schurgers, G., Winguth, A. M. E. E., et al. (2008).
18 Long-term ice sheet-climate interactions under anthropogenic greenhouse forcing simulated with a complex Earth
19 System Model. *Clim. Dyn.* 31, 665–690. doi:10.1007/s00382-008-0369-7.
- 20 Vizcaino, M., Mikolajewicz, U., Jungclaus, J., and Schurgers, G. (2010). Climate modification by future ice sheet
21 changes and consequences for ice sheet mass balance. *Clim. Dyn.* 34, 301–324. doi:10.1007/s00382-009-0591-y.
- 22 Vizcaino, M., Mikolajewicz, U., Ziemen, F., Rodehacke, C. B., Greve, R., Broeke, M. R., et al. (2015). Coupled
23 simulations of Greenland Ice Sheet and climate change up to A.D. 2300. *Geophys. Res. Lett.* 42, 3927–3935.
24 doi:10.1002/2014GL061142.
- 25 Voudoukas, M. I., Mentaschi, L., Voukouvalas, E., Verlaan, M., and Feyen, L. (2017). Extreme sea levels on the rise
26 along Europe’s coasts. *Earth’s Futur.* 5, 304–323.
- 27 Voudoukas, M. I., Mentaschi, L., Voukouvalas, E., Verlaan, M., Jevrejeva, S., Jackson, L. P., et al. (2018). Global
28 probabilistic projections of extreme sea levels show intensification of coastal flood hazard. *Nat. Commun.* 9,
29 2360. doi:10.1038/s41467-018-04692-w.
- 30 Wada, Y. (2016). Modeling groundwater depletion at regional and global scales: Present state and future prospects.
31 *Surv. Geophys.* 37, 419–451.
- 32 Wada, Y., Lo, M.-H., Yeh, P. J.-F., Reager, J. T., Famiglietti, J. S., Wu, R.-J., et al. (2016). Fate of water pumped from
33 underground and contributions to sea-level rise. *Nat. Clim. Chang.* doi:10.1038/NCLIMATE3001.
- 34 Wada, Y., van Beek, L. P. H., Sperna Weiland, F. C., Chao, B. F., Wu, Y.-H., and Bierkens, M. F. P. (2012). Past and
35 future contribution of global groundwater depletion to sea-level rise. *Geophys. Res. Lett.* 39, n/a-n/a.
36 doi:10.1029/2012GL051230.
- 37 Wagner, T. J. W., and Eisenman, I. (2015). How Climate Model Complexity Influences Sea Ice Stability. *J. Clim.* 28,
38 3998–4014. doi:10.1175/JCLI-D-14-00654.1.
- 39 Wahl, T., and Chambers, D. P. (2015). Evidence for multidecadal variability in US extreme sea level records. *J.*
40 *Geophys. Res. Ocean.* 120, 1527–1544.
- 41 Wahl, T., Haigh, I. D., Nicholls, R. J., Arns, A., Dangendorf, S., Hinkel, J., et al. (2017). Understanding extreme sea
42 levels for broad-scale coastal impact and adaptation analysis. *Nat. Commun.* 8. doi:10.1038/ncomms16075.
- 43 Walsh, J. E., Fetterer, F., Scott Stewart, J., and Chapman, W. L. (2017). A database for depicting Arctic sea ice
44 variations back to 1850. *Geogr. Rev.* 107, 89–107. doi:10.1111/j.1931-0846.2016.12195.x.
- 45 Wang, C., Zhang, L., Lee, S.-K. K., Wu, L., and Mechoso, C. R. (2014a). A global perspective on CMIP5 climate
46 model biases. *Nat. Clim. Chang.* 4, 201–205. doi:10.1038/nclimate2118.
- 47 Wang, G., Hendon, H. H., Arblaster, J. M., Lim, E.-P., Abhik, S., and van Rensch, P. (2019a). Compounding tropical
48 and stratospheric forcing of the record low Antarctic sea-ice in 2016. *Nat. Commun.* 10, 13. doi:10.1038/s41467-
49 018-07689-7.
- 50 Wang, G., Xie, S.-P. P., Huang, R. X., and Chen, C. (2015a). Robust Warming Pattern of Global Subtropical Oceans
51 and Its Mechanism. *J. Clim.* 28, 8574–8584. doi:10.1175/JCLI-D-14-00809.1.
- 52 Wang, H., Legg, S. A., and Hallberg, R. W. (2015b). Representations of the Nordic Seas overflows and their large scale
53 climate impact in coupled models. *Ocean Model.* 86, 76–92. doi:10.1016/j.ocemod.2014.12.005.
- 54 Wang, H., Legg, S. A., and Hallberg, R. W. (2015c). Representations of the Nordic Seas overflows and their large scale
55 climate impact in coupled models. *Ocean Model.* 86, 76–92. doi:10.1016/j.ocemod.2014.12.005.
- 56 Wang, J., Church, J. A., Zhang, X., and Chen, X. (submitted). Observations confirm projected global mean and regional
57 sea-level trends and accelerations. *Nature* (submitted).
- 58 Wang, L., Cole, J. N. S., Bartlett, P., Versegny, D., Derksen, C., Brown, R., et al. (2016a). Investigating the spread in
59 surface albedo for snow-covered forests in CMIP5 models. *J. Geophys. Res.* 121, 1104–1119.
60 doi:10.1002/2015JD023824.
- 61 Wang, L., Derksen, C., Brown, R., and Markus, T. (2013). Recent changes in pan-Arctic melt onset from satellite

- passive microwave measurements. *Geophys. Res. Lett.* 40, 522–528. doi:10.1002/grl.50098.
- Wang, M., and Overland, J. E. (2015). Projected future duration of the sea-ice-free season in the Alaskan Arctic. *Prog. Oceanogr.* 136, 50–59. doi:10.1016/j.pocean.2015.01.001.
- Wang, Q., Danilov, S., Sidorenko, D., Timmermann, R., Wekerle, C., Wang, X., et al. (2014b). The Finite Element Sea Ice-Ocean Model (FESOM) v.1.4: formulation of an ocean general circulation model. *Geosci. Model Dev.* 7, 663–693. doi:10.5194/gmd-7-663-2014.
- Wang, Q., Wang, X., Wekerle, C., Danilov, S., Jung, T., Koldunov, N., et al. (2019b). Ocean heat transport into the Barents Sea: Distinct controls on the upward trend and interannual variability. *Geophys. Res. Lett.* doi:10.1029/2019gl083837.
- Wang, W., Rinke, A., Moore, J. C., Ji, D., Cui, X., Peng, S., et al. (2016b). Evaluation of air-soil temperature relationships simulated by land surface models during winter across the permafrost region. *Cryosph.* 10, 1721–1737. doi:10.5194/tc-10-1721-2016.
- Wang, X., Key, J., Kwok, R., and Zhang, J. (2016c). Comparison of Arctic Sea Ice Thickness from Satellites, Aircraft, and PIOMAS Data. *Remote Sens.* 8, 713. doi:10.3390/rs8090713.
- Wang, Y.-L., and Wu, C.-R. (2018). Discordant multi-decadal trend in the intensity of the Kuroshio along its path during 1993–2013. *Sci. Rep.* 8, 14633. doi:10.1038/s41598-018-32843-y.
- Wang, Y.-L., Wu, C.-R., and Chao, S.-Y. (2016d). Warming and weakening trends of the Kuroshio during 1993–2013. *Geophys. Res. Lett.* 43, 9200–9207. doi:10.1002/2016GL069432.
- Wang, Y., Jiang, J. H., Su, H., Choi, Y.-S., Huang, L., Guo, J., et al. (2018). Elucidating the Role of Anthropogenic Aerosols in Arctic Sea Ice Variations. *J. Clim.* 31, 99–114. doi:10.1175/JCLI-D-17-0287.1.
- Wang, Z., Kuhlbrodt, T., and Meredith, M. P. (2011). On the response of the Antarctic Circumpolar Current transport to climate change in coupled climate models. *J. Geophys. Res. Ocean.* 116. doi:10.1029/2010JC006757.
- Ward, P. J., Couasnon, A., Eilander, D., Haigh, I. D., Hendry, A., Muis, S., et al. (2018). Dependence between high sea-level and high river discharge increases flood hazard in global deltas and estuaries. *Environ. Res. Lett.* 13, 084012. doi:10.1088/1748-9326/aad400.
- Watkins, M. M., Wiese, D. N., Yuan, D.-N., Boening, C., and Landerer, F. W. (2015). Improved methods for observing Earth's time variable mass distribution with GRACE using spherical cap mascons. *J. Geophys. Res. Solid Earth* 120, 2648–2671. doi:10.1002/2014JB011547.
- Waugh, D. W., McC. Hogg, A., Spence, P., England, M. H., and Haine, T. W. N. (2019). Response of Southern Ocean Ventilation to Changes in Midlatitude Westerly Winds. *J. Clim.* 32, 5345–5361. doi:10.1175/JCLI-D-19-0039.1.
- WCRP Global Sea Level Budget Group (2018). Global sea-level budget 1993–present. *Earth Syst. Sci. Data* 10, 1551–1590. doi:10.5194/essd-10-1551-2018.
- Webber, B. G. M., Heywood, K. J., Stevens, D. P., and Assmann, K. M. (2019). The Impact of Overturning and Horizontal Circulation in Pine Island Trough on Ice Shelf Melt in the Eastern Amundsen Sea. *J. Phys. Oceanogr.* 49, 63–83. doi:10.1175/JPO-D-17-0213.1.
- Weber, M. E., Clark, P. U., Kuhn, G., Timmermann, A., Spreng, D., Gladstone, R., et al. (2014). Millennial-scale variability in Antarctic ice-sheet discharge during the last deglaciation. *Nature* 510, 134–138. doi:10.1038/nature13397.
- Weertman, J. (1974). Stability of the junction of an ice sheet and an ice shelf. *J. Glaciol.* 13, 3–11. doi:10.3198/1974JoG13-67-3-11.
- Weijer, W., Cheng, W., Drijfhout, S. S., Fedorov, A. V., Hu, A., Jackson, L. C., et al. (2019). Stability of the Atlantic Meridional Overturning Circulation: A Review and Synthesis. *J. Geophys. Res. Ocean.*, 2019JC015083. doi:10.1029/2019JC015083.
- Weijer, W., and van Sebille, E. (2014). Impact of Agulhas Leakage on the Atlantic Overturning Circulation in the CCSM4. *J. Clim.* 27, 101–110. doi:10.1175/JCLI-D-12-00714.1.
- Weller, E., Min, S.-K., Cai, W., Zwiers, F. W., Kim, Y.-H., and Lee, D. (2016). Human-caused Indo-Pacific warm pool expansion. *Sci. Adv.* 2. doi:10.1126/sciadv.1501719.
- Wernberg, T., Bennett, S., Babcock, R. C., Bettignies, T. De, Cure, K., Depczynski, M., et al. (2016). Climate-driven regime shift of a temperate marine ecosystem. *Science* 353, 169–172. doi:10.1126/science.aad8745.
- Wernberg, T., Smale, D. A., Tuya, F., Thomsen, M. S., Langlois, T. J., de Bettignies, T., et al. (2013). An extreme climatic event alters marine ecosystem structure in a global biodiversity hotspot. *Nat. Clim. Chang.* 3, 78–82. doi:10.1038/nclimate1627.
- WGMS (2017). Global Glacier Change Bulletin No. 2 (2014–2015). , eds. M. Zemp, S. U. Nussbaumer, I. Gärtner-Roer, J. Huber, H. Machguth, F. Paul, et al. Zurich, Switzerland.: ICSU(WDS)/IUGG(IACS)/UNEP/UNESCO/WMO, World Glacier Monitoring Service doi:10.5904/wgms-fog-2017-10.
- Whitehouse, P. L. (2018). Glacial isostatic adjustment modelling: Historical perspectives, recent advances, and future directions. *Earth Surf. Dyn.* doi:10.5194/esurf-6-401-2018.
- Whitehouse, P. L., Bentley, M. J., and Le Brocq, A. M. (2012). A deglacial model for Antarctica: geological constraints and glaciological modelling as a basis for a new model of Antarctic glacial isostatic adjustment. *Quat. Sci. Rev.* 32, 1–24.

- Wiebe, E. C., and Weaver, A. J. (1999). On the sensitivity of global warming experiments to the parametrisation of sub-grid scale ocean mixing. *Clim. Dyn.* 15, 875–893. doi:10.1007/s003820050319.
- Wijffels, S., Roemmich, D., Monselesan, D., Church, J., and Gilson, J. (2016). Ocean temperatures chronicle the ongoing warming of Earth. *Nat. Clim. Chang.* 6, 116–118. doi:10.1038/nclimate2924.
- Williams, G., Maksym, T., Wilkinson, J., Kunz, C., Murphy, C., Kimball, P., et al. (2015). Thick and deformed Antarctic sea ice mapped with autonomous underwater vehicles. *Nat. Geosci.* 8, 61–67. doi:10.1038/ngeo2299.
- Wilmes, S. B., and Green, J. A. M. (2014). The evolution of tides and tidal dissipation over the past 21,000 years. *J. Geophys. Res. Ocean.* doi:10.1002/2013JC009605.
- Wilmes, S. B., Green, J. A. M., Gomez, N., Rippeth, T. P., and Lau, H. (2017). Global Tidal Impacts of Large-Scale Ice Sheet Collapses. *J. Geophys. Res. Ocean.* doi:10.1002/2017JC013109.
- Wilson, R., Carrión, D., and Rivera, A. (2016). Detailed dynamic, geometric and supraglacial moraine data for Glacier Pio XI, the only surge-type glacier of the Southern Patagonia Icefield. *Ann. Glaciol.* 57, 119–130. doi:10.1017/aog.2016.32.
- Wilton, D. J., Jowett, A., Hanna, E., Bigg, G. R., van den Broeke, M. R., Fettweis, X., et al. (2017). High resolution (1 km) positive degree-day modelling of Greenland ice sheet surface mass balance, 1870–2012 using reanalysis data. *J. Glaciol.* 63, 176–193. doi:DOI: 10.1017/jog.2016.133.
- Winkelmann, R., Levermann, A., Ridgwell, A., and Caldeira, K. (2015). Combustion of available fossil fuel resources sufficient to eliminate the Antarctic Ice Sheet. *Sci. Adv.* 1, e1500589–e1500589. doi:10.1126/sciadv.1500589.
- Winton, M. (2006). Does the Arctic sea ice have a tipping point? *Geophys. Res. Lett.* 33, L23504. doi:10.1029/2006GL028017.
- Winton, M., Griffies, S. M., Samuels, B. L., Sarmiento, J. L., and Licher, T. L. F. (2013). Connecting changing ocean circulation with changing climate. *J. Clim.* doi:10.1175/JCLI-D-12-00296.1.
- Wise, M. G., Dowdeswell, J. A., Jakobsson, M., and Larter, R. D. (2017). Evidence of marine ice-cliff instability in Pine Island Bay from iceberg-keel plough marks. *Nature*. doi:10.1038/nature24458.
- Wittmann, M., Zwaafink, C. D. G., Schmidt, L. S., Gumundsson, S., Pálsson, F., Arnalds, O., et al. (2017). Impact of dust deposition on the albedo of Vatnajökull ice cap, Iceland. *Cryosph.* 11, 741–754. doi:10.5194/tc-11-741-2017.
- Wolovick, M. J., and Moore, J. C. (2018). Stopping the flood: could we use targeted geoengineering to mitigate sea level rise? *Cryosph.* 12, 2955–2967.
- Wong, T. E., Bakker, A. M. R., and Keller, K. (2017). Impacts of Antarctic fast dynamics on sea-level projections and coastal flood defense. *Clim. Change* 144, 347–364. doi:10.1007/s10584-017-2039-4.
- Wood, M., Rignot, E., Fenty, I., Menemenlis, D., Millan, R., Morlighem, M., et al. (2018). Ocean-Induced Melt Triggers Glacier Retreat in Northwest Greenland. *Geophys. Res. Lett.* 45, 8334–8342. doi:10.1029/2018GL078024.
- Woods, C., and Caballero, R. (2016). The role of moist intrusions in winter Arctic warming and sea ice decline. *J. Clim.* 29, 4473–4485. doi:10.1175/JCLI-D-15-0773.1.
- Woodworth, P. L. (2010). A survey of recent changes in the main components of the ocean tide. *Cont. Shelf Res.* doi:10.1016/j.csr.2010.07.002.
- Woodworth, P. L., Hunter, J. R., Marcos, M., Caldwell, P., Menéndez, M., and Haigh, I. (2016). Towards a global higher-frequency sea level dataset. *Geosci. Data J.* 3, 50–59.
- Worby, A. P., Geiger, C. A., Paget, M. J., van Woert, M. L., Ackley, S. F., and DeLiberty, T. L. (2008). Thickness distribution of Antarctic sea ice. *J. Geophys. Res.* 113, C05S92. doi:10.1029/2007JC004254.
- Woth, K., Weisse, R., and von Storch, H. (2006). Climate change and North Sea storm surge extremes: an ensemble study of storm surge extremes expected in a changed climate projected by four different regional climate models. *Ocean Dyn.* 56, 3–15. doi:10.1007/s10236-005-0024-3.
- Wouters, B., Gardner, A. S., and Moholdt, G. (2019). Global Glacier Mass Loss During the GRACE Satellite Mission (2002–2016). *Front. Earth Sci.* 7, 96. doi:10.3389/feart.2019.00096.
- Wouters, B., Martín-Español, A., Helm, V., Flament, T., Van Wessem, J. M., Ligtenberg, S. R. M., et al. (2015). Dynamic thinning of glaciers on the Southern Antarctic Peninsula. *Science* 348, 899–903. doi:10.1126/science.aaa5727.
- Wu, L., Cai, W., Zhang, L., Nakamura, H., Timmermann, A., Joyce, T., et al. (2012). Enhanced warming over the global subtropical western boundary currents. *Nat. Clim. Chang.* 2, 161–166. doi:10.1038/nclimate1353.
- Wu, Q., Hou, Y., Yun, H., and Liu, Y. (2015). Changes in active-layer thickness and near-surface permafrost between 2002 and 2012 in alpine ecosystems, Qinghai-Xizang (Tibet) Plateau, China. *Glob. Planet. Change*. doi:10.1016/j.gloplacha.2014.09.002.
- Wullschleger, S. D., Epstein, H. E., Box, E. O., Euskirchen, E. S., Goswami, S., Iversen, C. M., et al. (2014). Plant functional types in Earth system models: past experiences and future directions for application of dynamic vegetation models in high-latitude ecosystems. *Ann. Bot.* 114, 1–16. doi:10.1093/aob/mcu077.
- Wüst, G. (1936). Oberflächensalzgehalt, verdunstung und niederschlag auf den weltmeere, Landerkundliche forschung. *Festschrift Norbert Krebs (Länderkundliche Res., 347–359.*
- Xiang, B., Zhao, M., Held, I. M., and Golaz, J.-C. (2017). Predicting the severity of spurious “double ITCZ” problem in CMIP5 coupled models from AMIP simulations. *Geophys. Res. Lett.* 44, 1520–1527.

- doi:10.1002/2016GL071992.
- Xiao, W., Esper, O., and Gersonde, R. (2016). Last Glacial - Holocene climate variability in the Atlantic sector of the Southern Ocean. *Quat. Sci. Rev.* 135, 115–137. doi:10.1016/j.quascirev.2016.01.023.
- Xiao, X., Zhao, M., Knudsen, K. L., Sha, L., Eiríksson, J., Gudmundsdóttir, E., et al. (2017). Deglacial and Holocene sea-ice variability north of Iceland and response to ocean circulation changes. *Earth Planet. Sci. Lett.* 472, 14–24. doi:10.1016/j.epsl.2017.05.006.
- Xie, S. P., Deser, C., Vecchi, G. A., Ma, J., Teng, H., and Wittenberg, A. T. (2010). Global warming pattern formation: Sea surface temperature and rainfall. *J. Clim.* doi:10.1175/2009JCLI3329.1.
- Xu, X., Chassignet, E. P., and Wang, F. (2019). On the variability of the Atlantic meridional overturning circulation transports in coupled CMIP5 simulations. *Clim. Dyn.* 52, 6511–6531. doi:10.1007/s00382-018-4529-0.
- Xu, Y., Rignot, E., Fenty, I., Menemenlis, D., and Flexas, M. M. (2013). Subaqueous melting of Store Glacier, west Greenland from three-dimensional, high-resolution numerical modeling and ocean observations. *Geophys. Res. Lett.* doi:10.1002/grl.50825.
- Xu, Y., Rignot, E., Menemenlis, D., and Koppes, M. (2012). Numerical experiments on subaqueous melting of Greenland tidewater glaciers in response to ocean warming and enhanced subglacial discharge. *Ann. Glaciol.* 53, 229–234. doi:DOI: 10.3189/2012AoG60A139.
- Yamaguchi, R., and Suga, T. (2019). Trend and Variability in Global Upper-Ocean Stratification Since the 1960s. *J. Geophys. Res. Ocean.* n/a. doi:10.1029/2019JC015439.
- Yan, Q., Zhang, Z., and Wang, H. (2016). Investigating uncertainty in the simulation of the Antarctic ice sheet during the mid-Piacenzian. *J. Geophys. Res. Atmos.* 121, 1559–1574. doi:10.1002/2015JD023900.
- Yan, Q., Zhang, Z., Wang, H., and Zhang, R. (2014). Simulation of Greenland ice sheet during the mid-Pliocene warm period. *Chinese Sci. Bull.* 59, 201–211.
- Yan, X., and Sun, C. (2015). An altimetric transport index for Kuroshio inflow northeast of Taiwan Island. *Sci. China Earth Sci.* 58, 697–706. doi:10.1007/s11430-014-5024-z.
- Yang, H., Lohmann, G., Wei, W., Dima, M., Ionita, M., and Liu, J. (2016a). Intensification and poleward shift of subtropical western boundary currents in a warming climate. *J. Geophys. Res. Ocean.* 121, 4928–4945. doi:10.1002/2015JC011513.
- Yang, J.-A., Kim, S., Mori, N., and Mase, H. (2018). Assessment of long-term impact of storm surges around the Korean Peninsula based on a large ensemble of climate projections. *Coast. Eng.* 142, 1–8. doi:10.1016/J.COASTALENG.2018.09.008.
- Yang, Q., Dixon, T. H., Myers, P. G., Bonin, J., Chambers, D., and Van Den Broeke, M. R. (2016b). Recent increases in Arctic freshwater flux affects Labrador Sea convection and Atlantic overturning circulation. *Nat. Commun.* 7, 10525. doi:10.1038/ncomms10525.
- Yau, A. M., Bender, M. L., Robinson, A., and Brook, E. J. (2016). Reconstructing the last interglacial at Summit, Greenland: Insights from GISP2. *Proc. Natl. Acad. Sci.* 113, 9710–9715. doi:10.1073/PNAS.1524766113.
- Yeager, S. (2015). Topographic Coupling of the Atlantic Overturning and Gyre Circulations. *J. Phys. Oceanogr.* 45, 1258–1284. doi:10.1175/JPO-D-14-0100.1.
- Yeh, S.-W., Cai, W., Min, S.-K., McPhaden, M. J., Dommenges, D., Dewitte, B., et al. (2018). ENSO Atmospheric Teleconnections and Their Response to Greenhouse Gas Forcing. *Rev. Geophys.* 56, 185–206. doi:10.1002/2017RG000568.
- Yi, S., Heki, K., and Qian, A. (2017). Acceleration in the Global Mean Sea Level Rise: 2005–2015. *Geophys. Res. Lett.* doi:10.1002/2017GL076129.
- Yin, J. (2012). Century to multi-century sea level rise projections from CMIP5 models. *Geophys. Res. Lett.* 39, n/a-n/a. doi:10.1029/2012GL052947.
- Yin, J., and Stouffer, R. J. (2007). Comparison of the Stability of the Atlantic Thermohaline Circulation in Two Coupled Atmosphere–Ocean General Circulation Models. *J. Clim.* 20, 4293–4315. doi:10.1175/JCLI4256.1.
- Yokoyama, Y., Esat, T. M., Thompson, W. G., Thomas, A. L., Webster, J. M., Miyairi, Y., et al. (2018). Rapid glaciation and a two-step sea level plunge into the Last Glacial Maximum. *Nature* 559, 603.
- Young, I. R., and Ribal, A. (2019). Multiplatform evaluation of global trends in wind speed and wave height. *Science* 364, 548–552. doi:10.1126/science.aav9527.
- Young, N. E., and Briner, J. P. (2015). Holocene evolution of the western Greenland Ice Sheet: Assessing geophysical ice-sheet models with geological reconstructions of ice-margin change. *Quat. Sci. Rev.* 114, 1–17.
- Yu, L. (2019). Global Air-Sea Fluxes of Heat, Fresh Water, and Momentum: Energy Budget Closure and Unanswered Questions. *Ann. Rev. Mar. Sci.* 11, 227–248. doi:10.1146/annurev-marine-010816.
- Yu, L., Jin, X., Josey, S. A., Lee, T., Kumar, A., Wen, C., et al. (2017). The global ocean water cycle in atmospheric reanalysis, satellite, and ocean salinity. *J. Clim.* 30, 3829–3852. doi:10.1175/JCLI-D-16-0479.1.
- Yu, Y., Liu, H., and Lin, P. (2012). A quasi-global 1/10° eddy-resolving ocean general circulation model and its preliminary results. *Chinese Sci. Bull.* 57, 3908–3916. doi:10.1007/s11434-012-5234-8.
- Yuan, N., Ding, M., Ludescher, J., Bunde, A., Parkinson, C. L., DeConto, R. M., et al. (2017). Increase of the Antarctic Sea Ice Extent is highly significant only in the Ross Sea. *Sci. Rep.* 7, 41096. doi:10.1038/srep41096.
- Zalazar, L., Ferri Hidalgo, L., Castro, M., Gargantini, H., Gimenez, M., Pitte, P., et al. (submitted). Spatial distribution

- and main characteristics of Andean ice masses in Argentina: results from the first National Glacier Inventory. *J. Glaciol.* (submitted).
- Zängl, G., Reinert, D., Ripodas, P., and Baldauf, M. (2015). The ICON (ICOsahedral Non-hydrostatic) modelling framework of DWD and MPI-M: Description of the non-hydrostatic dynamical core. *Q. J. R. Meteorol. Soc.* 141, 563–579. doi:10.1002/qj.2378.
- Zanna, L., Khatiwala, S., Gregory, J. M., Ison, J., and Heimbach, P. (2019). Global reconstruction of historical ocean heat storage and transport. *Proc. Natl. Acad. Sci.* 116, 201808838. doi:10.1073/pnas.1808838115.
- Zazulie, N., Rusticucci, M., and Raga, G. B. (2018). Regional climate of the Subtropical Central Andes using high-resolution CMIP5 models. Part II: future projections for the twenty-first century. *Clim. Dyn.* 51, 2913–2925. doi:10.1007/s00382-017-4056-4.
- Zekollari, H., Huss, M., and Farinotti, D. (2019). Modelling the future evolution of glaciers in the European Alps under the EURO-CORDEX RCM ensemble. *Cryosph.* 13, 1125–1146. doi:10.5194/tc-13-1125-2019.
- Zekollari, H., and Huybrechts, P. (2015). On the climate–geometry imbalance, response time and volume–area scaling of an alpine glacier: insights from a 3-D flow model applied to Vadret da Morteratsch, Switzerland. *Ann. Glaciol.* 56, 51–62. doi:DOI: 10.3189/2015AoG70A921.
- Zekollari, H., Huybrechts, P., Noël, B., van de Berg, W. J., and van den Broeke, M. R. (2017). Sensitivity, stability and future evolution of the world’s northernmost ice cap, Hans Tausen Iskappe (Greenland). *Cryosph.* 11, 805–825. doi:10.5194/tc-11-805-2017.
- Zemp, M., Frey, H., Gärtner-Roer, I., Nussbaumer, S. U., Hoelzle, M., Paul, F., et al. (2015). Historically unprecedented global glacier decline in the early 21st century. *J. Glaciol.* 61, 745–762. doi:10.3189/2015JoG15J017.
- Zemp, M., Huss, M., Thibert, E., Eckert, N., McNabb, R., Huber, J., et al. (2019). Global glacier mass changes and their contributions to sea-level rise from 1961 to 2016. *Nature* 568, 382–386. doi:10.1038/s41586-019-1071-0.
- Zhai, F., Hu, D., Wang, Q., and Wang, F. (2014). Long-term trend of Pacific South Equatorial Current bifurcation over 1950–2010. *Geophys. Res. Lett.* 41, 3172–3180. doi:10.1002/2014GL059934.
- Zhang, J. (2014). Modeling the impact of wind intensification on Antarctic sea ice volume. *J. Clim.* 27, 202–214. doi:10.1175/JCLI-D-12-00139.1.
- Zhang, L., Delworth, T. L., Cooke, W., and Yang, X. (2019a). Natural variability of Southern Ocean convection as a driver of observed climate trends. *Nat. Clim. Chang.* 9, 59–65. doi:10.1038/s41558-018-0350-3.
- Zhang, L., and Karnauskas, K. B. (2017). The Role of Tropical Interbasin SST Gradients in Forcing Walker Circulation Trends. *J. Clim.* 30, 499–508. doi:10.1175/JCLI-D-16-0349.1.
- Zhang, L., and Wang, C. (2013). Multidecadal North Atlantic sea surface temperature and Atlantic meridional overturning circulation variability in CMIP5 historical simulations. *J. Geophys. Res. Ocean.* 118, 5772–5791. doi:10.1002/jgrc.20390.
- Zhang, R., Sutton, R., Danabasoglu, G., Kwon, Y., Marsh, R., Yeager, S. G., et al. (2019b). A Review of the Role of the Atlantic Meridional Overturning Circulation in Atlantic Multidecadal Variability and Associated Climate Impacts. *Rev. Geophys.* 57, 316–375. doi:10.1029/2019RG000644.
- Zhang, T., Barry, R. G., Knowles, K., Heginbottom, J. A., and Brown, J. (1999). Statistics and characteristics of permafrost and ground-ice distribution in the Northern Hemisphere. *Polar Geogr.* 23, 132–154. doi:10.1080/10889379909377670.
- Zhang, T., Heginbottom, J. A., Barry, R. G., and Brown, J. (2000). Further statistics on the distribution of permafrost and ground ice in the Northern Hemisphere. *Polar Geogr.* 24, 126–131. doi:10.1080/10889370009377692.
- Zhang, X., and Church, J. A. (2012). Sea level trends, interannual and decadal variability in the Pacific Ocean. *Geophys. Res. Lett.* 39, n/a-n/a. doi:10.1029/2012GL053240.
- Zhang, Y., and Baptista, A. M. (2008). SELFE: A semi-implicit Eulerian–Lagrangian finite-element model for cross-scale ocean circulation. *Ocean Model.* 21, 71–96. doi:10.1016/J.OCEMOD.2007.11.005.
- Zhang, Y., Chen, C., Beardsley, R. C., Gao, G., Lai, Z., Curry, B., et al. (2016a). Studies of the Canadian Arctic Archipelago water transport and its relationship to basin-local forcings: Results from AO-FVCOM. *J. Geophys. Res. Ocean.* 121, 4392–4415. doi:10.1002/2016JC011634.
- Zhang, Y. J., Ye, F., Stanev, E. V., and Grashorn, S. (2016b). Seamless cross-scale modeling with SCHISM. *Ocean Model.* 102, 64–81. doi:10.1016/J.OCEMOD.2016.05.002.
- Zhang, Y., Renssen, H., Seppä, H., and Valdes, P. J. (2017). Holocene temperature evolution in the Northern Hemisphere high latitudes – Model-data comparisons. *Quat. Sci. Rev.* 173, 101–113. doi:10.1016/j.quascirev.2017.07.018.
- Zhang, Z., Tian, J., Qiu, B., Zhao, W., Chang, P., Wu, D., et al. (2016c). Observed 3D Structure, Generation, and Dissipation of Oceanic Mesoscale Eddies in the South China Sea. *Sci. Rep.* 6. doi:10.1038/srep24349.
- Zheng, X.-T. T., Xie, S.-P. P., Du, Y., Liu, L., Huang, G., and Liu, Q. (2013). Indian Ocean Dipole Response to Global Warming in the CMIP5 Multimodel Ensemble. *J. Clim.* 26, 6067–6080. doi:10.1175/JCLI-D-12-00638.1.
- Zhong, H., Overloop, P.-J. van, and Gelder, P. van (2013). A joint probability approach using a 1-D hydrodynamic model for estimating high water level frequencies in the Lower Rhine Delta. *Nat. Hazards Earth Syst. Sci.* 13, 1841–1852.

- 1 Zhou, Y., Li, Z., and Li, J. I. A. (2017). Slight glacier mass loss in the Karakoram region during the 1970s to 2000
2 revealed by KH-9 images and SRTM DEM. *J. Glaciol.* 63, 331–342. doi:DOI: 10.1017/jog.2016.142.
- 3 Zhu, D., Ciais, P., Krinner, G., Maignan, F., Jornet-Puig, A., and Hugelius, G. (2019). Controls of soil organic matter
4 on permafrost thermal and carbon dynamics. *Nat. Commun.* 10, 3172.
- 5 Zhu, J., Liu, Z., Zhang, J., and Liu, W. (2015). AMOC response to global warming: dependence on the background
6 climate and response timescale. *Clim. Dyn.* 44, 3449–3468. doi:10.1007/s00382-014-2165-x.
- 7 Zickfeld, K., Arora, V. K., and Gillett, N. P. (2012). Is the climate response to CO₂ emissions path dependent?
8 *Geophys. Res. Lett.* 39, n/a-n/a. doi:10.1029/2011GL050205.
- 9 Zickfeld, K., Solomon, S., and Gilford, D. M. (2017). Centuries of thermal sea-level rise due to anthropogenic
10 emissions of short-lived greenhouse gases. *Proc. Natl. Acad. Sci.* 114, 657–662. doi:10.1073/pnas.1612066114.
- 11 Ziemen, F. A., Hock, R., Aschwanden, A., Khroulev, C., Kienholz, C., Melkonian, A., et al. (2016). Modeling the
12 evolution of the Juneau Icefield between 1971 and 2100 using the Parallel Ice Sheet Model (PISM). *J. Glaciol.*
13 62, 199–214. doi:10.1017/jog.2016.13.
- 14 Zika, J. D., Gregory, J. M., McDonagh, E. L., Marzocchi, A., and Clement, L. (submitted). Rapid Southern Ocean
15 warming explained by changes in poleward heat transport. *Nat. Geosci.* (submitted).
- 16 Zika, J. D., Skliris, N., Blaker, A., Marsh, R., Nurser, A. J. G., and Josey, S. (2018). Improved estimates of water cycle
17 change from ocean salinity: the key role of ocean warming. *Environ. Res. Lett.* 13, 74036. doi:10.1088/1748-
18 9326/aace42.
- 19 Zika, J. D., Skliris, N., George, A. J., Josey, S. A., Mudryk, L., Laliberté, F., et al. (2015). Maintenance and broadening
20 of the ocean's salinity distribution by the water cycle. *J. Clim.* 20, 9550–9560. doi:10.1175/JCLI-D-15-0273.1.
- 21 Zorzut, V., Ruiz, L., Rivera, A., Pitte, P., and Villalba, R. (submitted). Slope estimation influences on ice thickness
22 inversion models: a case study for Monte Tronador glaciers, North Patagonian Andes. *J. Glaciol.* (submitted).
- 23 Zunz, V., Goosse, H., and Massonnet, F. (2013). How does internal variability influence the ability of CMIP5 models to
24 reproduce the recent trend in Southern Ocean sea ice extent? *Cryosph.* 7, 451–468. doi:10.5194/tc-7-451-2013.
- 25
- 26

Appendix Chapter 9.A

9.A.1 Details for Figure 9.10

Lower right panel: In all experiments an additional surface freshwater flux (hosing) is applied over the subpolar North Atlantic (regions vary between studies) for a limited time. The total amount of fresh water added is then this flux multiplied by the number of years applied (in Gt). While hosing is applied the AMOC weakens. In some models the AMOC recovers quickly, in other the AMOC hasn't recovered after 200 years, and in some the AMOC is starting to recover after 200 years. These are represented by circles, downward and upward triangles respectively which show the percentage AMOC change after 200 years. Sizes of symbols represent the magnitude of the hosing (from 0.1-1Sv) and colours indicate the different studies these results were taken from (de Vries and Weber, 2005; Stouffer et al., 2006; Yin and Stouffer, 2007; Meehl et al., 2009; Jackson, 2013; Liu and Liu, 2013; Timmermans et al., 2014; Hutchings et al., 2015; Jackson and Wood, 2018; Haskins et al., 2019).

9.A.2 Details for Figure 9.19

Upper left panel: This is a p-box based estimate of the range of values from a number of different estimates of paleo sea levels relative to present day and the median over all central estimates (Simpson et al., 2009; Argus and Peltier, 2010; Fyke et al., 2011; Robinson et al., 2011; Colville et al., 2011; Dolan et al., 2011; Born and Nisancioglu, 2012; Miller et al., 2012; Helsen et al., 2013; Nick et al., 2013; Quiquet et al., 2013; Stone et al., 2013; Dahl-Jensen et al., 2013; Lecavalier et al., 2014; Robinson and Goelzer, 2014; Colleoni et al., 2014; Koenig et al., 2015; Peltier et al., 2015; Stuhne and Peltier, 2015; Vizcaino et al., 2015; Calov R, Robinson A, Perrette M, 2015; Dutton et al., 2015b; Goelzer et al., 2016b; Khan et al., 2016; Yau et al., 2016; de Boer et al., 2017; Calov et al., 2018; Simms et al., 2019).

9.A.3 Details for Figure 9.20

Upper left panel: This is a p-box based estimate of the range of values from a number of different estimates of paleo sea levels relative to present day and the median over all central estimates (Bamber et al., 2009; Argus and Peltier, 2010; Mackintosh et al., 2011; Dolan et al., 2011; Golledge et al., 2012, 2013, 2014, 2015, 2017b; Miller et al., 2012; Whitehouse et al., 2012; Ivins et al., 2013; Argus et al., 2014; Maris et al., 2014; Briggs et al., 2014b; Pollard et al., 2015; De Boer et al., 2015; Dutton et al., 2015b; Goelzer et al., 2016b; Yan et al., 2016; Deconto and Pollard, 2016; Gasson et al., 2016b; Kopp et al., 2017; de Boer et al., 2017; Simms et al., 2019).

[START TABLE 9.A.1]

Table 9.A.1: Global mean sea-level contribution projections from ice sheets. Median values and (*likely*) [*very likely*] ranges of the process-based model ensemble, for 1995-2014 to 2100 in meters.

	SSP1-2.6	SSP2-4.5	SSP3-7.0	SSP5-8.5
	Antarctic ice sheet			
SROCC-consistent Kopp et al 2014	0.04 (0.01--0.12) [-0.01--0.24]	0.06 (0.01--0.16) [-0.01--0.31]		0.14 (0.03--0.35) [-0.01--0.66]

ISMIP6 (raw)	0.04 (0.01--0.10) [-0.09--0.18]			0.03 (-0.02--0.11) [-0.09--0.23]
ISMIP6 (emulated)	0.12 (0.01--0.25) [-0.07--0.39]	0.11 (-0.00--0.24) [-0.08--0.36]	0.12 (-0.00--0.24) [-0.07--0.36]	0.11 (0.01--0.26) [-0.05--0.37]
LARMIP2	0.12 (0.06--0.25) [0.04--0.38]	0.14 (0.07--0.28) [0.04--0.43]	0.15 (0.08--0.31) [0.05--0.48]	0.17 (0.08--0.36) [0.05--0.58]
Bamber et al. (2019) SEJ				0.21 (0.02-0.57) [-0.11-1.31]
	Greenland ice sheet			
SROCC-consistent Kopp et al 2014	0.06 (0.03--0.11) [0.02--0.17]	0.08 (0.03--0.15) [0.01--0.22]		0.14 (0.07--0.25) [0.05--0.38]
ISMIP6 (raw)	0.07 (0.07--0.08) [0.06--0.08]			0.14 (0.12--0.16) [0.10--0.18]
ISMIP6 (emulated)	0.07 (0.03--0.12) [- 0.00--0.19]	0.09 (0.05--0.14) [0.01--0.21]	0.12 (0.08--0.16) [0.05--0.22]	0.13 (0.09--0.19) [0.06--0.26]
Bamber et al. (2019) SEJ				0.23 (0.10-0.60) [0.03-0.99]

[END TABLE 9.A.1]

[START TABLE 9.A.2 HERE]

Table 9.A.2: Global mean sea-level projections (meters) for 5 SSP-RCP scenarios, for total change and individual contributions, median values and (*likely*) [*very likely*] ranges, from s 1995-2014 reference level, using alternative treatments of ice sheets.

	SROCC-consistent ice sheets			Bamber et al. (2019) SEJ
	SSP1-2.6	SSP2-4.5	SSP5-8.5	SSP5-8.5
Total medium term (2050)	0.19 (0.18--0.22) [0.17--0.25]	0.20 (0.18--0.24) [0.17--0.28]	0.25 (0.21--0.31) [0.19--0.40]	0.28 (0.21--0.41) [0.16--0.57]
Total long term (2090)	0.37 (0.31--0.47) [0.27--0.61]	0.45 (0.37--0.57) [0.32--0.75]	0.67 (0.54--0.90) [0.47--1.20]	0.85 (0.60--1.33) [0.47--1.83]

Total (2100)	0.41 (0.33--0.53) [0.29--0.71]	0.51 (0.41--0.68) [0.35--0.90]	0.80 (0.64--1.09) [0.56--1.47]	1.01 (0.71--1.63) [0.55--2.31]
Total rate long term (2080-2100)	4.1 (2.7--6.6) [1.7--10.2]	6.5 (4.5--9.9) [2.9--14.8]	13.0 (9.3--18.7) [8.2--26.3]	16.3 (10.0--29.6) [6.2--47.7]

1

	LARMIP2 AIS, SROCC-consistent GrIS		
	SSP1-2.6	SSP2-4.5	SSP5-8.5
Total medium term (2050)	0.21 (0.19--0.25) [0.17--0.30]	0.22 (0.19--0.25) [0.18--0.30]	0.24 (0.21--0.28) [0.18--0.35]
Total long term (2090)	0.44 (0.37--0.56) [0.32--0.70]	0.52 (0.44--0.64) [0.40--0.78]	0.68 (0.59--0.84) [0.54--1.03]
Total (2100)	0.50 (0.41--0.64) [0.36--0.81]	0.60 (0.51--0.75) [0.45--0.94]	0.84 (0.72--1.05) [0.66--1.29]
Total rate long term (2080-2100)	5.80 (4.40--8.09) [3.42--10.95]	8.46 (6.66--11.42) [5.17--15.35]	15.01 (12.38--19.41) [11.05--24.61]

* LARMIP2 does not account for surface mass balance changes in the Antarctic ice sheet, and so may overestimate total GMSL change.

[END TABLE 9.A.2 HERE]

[START TABLE 9.A.3 HERE]

Table 9.A.3: Global mean sea-level projections for exceedance of 5 GSAT temperature levels, by sorting GSAT change in 2081-2100 w.r.t. 1850-1900, for total change, median values and (*likely*) [*very likely*] range of the process-based model ensemble, for 1995-2014 to 2081-2100 in meters and the average rate in 2081-2100 in mm/yr.

	1.5 degrees	2 degrees	3 degrees	4 degrees	5 degrees
Thermosteric		0.15 (0.13--0.17) [0.07--0.22]	0.19 (0.17--0.22) [0.11--0.28]	0.26 (0.21--0.31) [0.15--0.37]	
GIC	0.08 (0.06--0.10)	0.09 (0.07--0.12) [0.06--0.14]	0.11 (0.09--0.14) [0.08--0.17]	0.13 (0.11--0.16) [0.09--0.19]	0.15 (0.12--0.19) [0.10--0.21]

	[0.05--0.12]				
AIS	0.12 (-0.00--0.25) [-0.07--0.34]	0.11 (0.00--0.25) [-0.09--0.39]	0.10 (-0.01--0.23) [-0.08--0.35]	0.12 (0.02--0.24) [-0.06--0.35]	0.10 (-0.01--0.27) [-0.08--0.40]
GrIS	0.07 (0.02--0.12) [-0.01--0.21]	0.08 (0.03--0.13) [0.01--0.19]	0.10 (0.06--0.15) [0.03--0.21]	0.12 (0.08--0.18) [0.05--0.26]	0.13 (0.10--0.18) [0.07--0.23]
TWS	0.04 (0.02--0.05) [0.01--0.06]	0.04 (0.02--0.05) [0.01--0.06]	0.04 (0.02--0.05) [0.01--0.06]	0.04 (0.02--0.05) [0.01--0.06]	0.04 (0.02--0.05) [0.01--0.06]
Total short term (2040)		0.13 (0.10--0.17) [0.08--0.21]	0.14 (0.11--0.17) [0.08--0.20]	0.14 (0.11--0.17) [0.09--0.20]	
Total medium term (2050)		0.19 (0.15--0.23) [0.12--0.27]	0.19 (0.15--0.24) [0.12--0.28]	0.20 (0.16--0.24) [0.13--0.27]	
Total long term (2090)		0.40 (0.29--0.52) [0.20--0.62]	0.46 (0.35--0.58) [0.27--0.72]	0.54 (0.43--0.66) [0.34--0.77]	
Total (2100)		0.48 (0.33--0.65) [0.21--0.83]	0.56 (0.41--0.71) [0.31--0.90]	0.69 (0.54--0.85) [0.44--1.00]	
Total rate medium term (2040-2060)		5.3 (4.2--6.3) [3.4--6.9]	5.8 (4.4--7.3) [3.6--8.6]	6.1 (5.1--7.2) [4.2--8.5]	
Total rate long term (2080-2100)		6.5 (3.9--9.9) [1.4--13.6]	8.2 (5.6--11.3) [3.4--15.6]	12.3 (9.7--15.5) [7.8--18.4]	

* The ISMIP6 emulator used to project ice sheet contributions does not account for temporal correlations, and so sample-wise uncertainties on rates would be overestimated. For the SOD, we report the trends in the 17th-83rd percentile and 5th-95th percentile estimates as the *likely* and *very likely* ranges. This approach effectively assumes perfect temporal correlation over 20 year periods, and may therefore underestimate uncertainties.

[END TABLE 9.A.3 HERE]

Figures

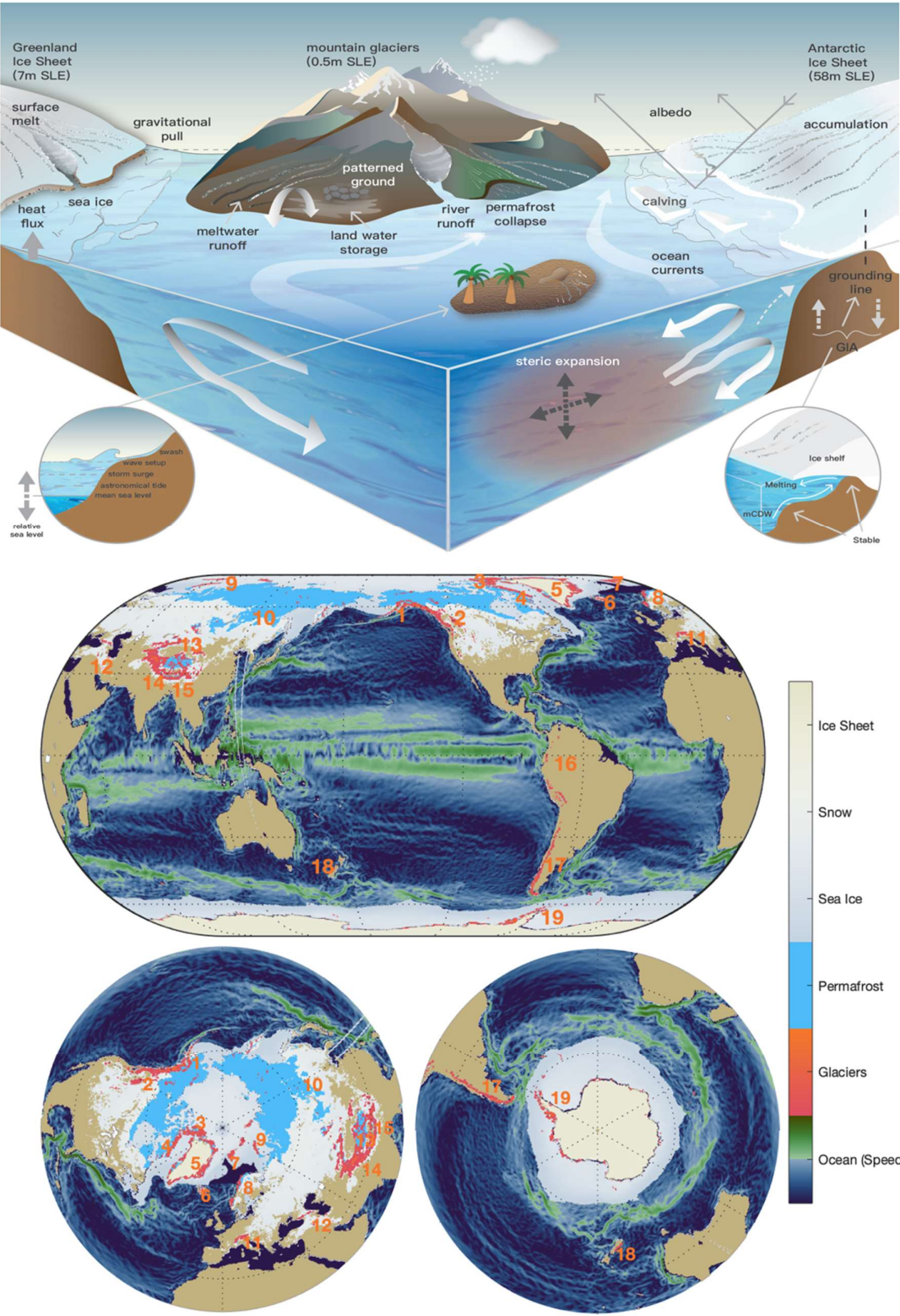


Figure 9.1: Components of ocean, cryosphere and sea level assessed in chapter. (a) Schematic of processes. (b) Geographic distribution of ocean and cryosphere components (numbers indicate glacierized regions).

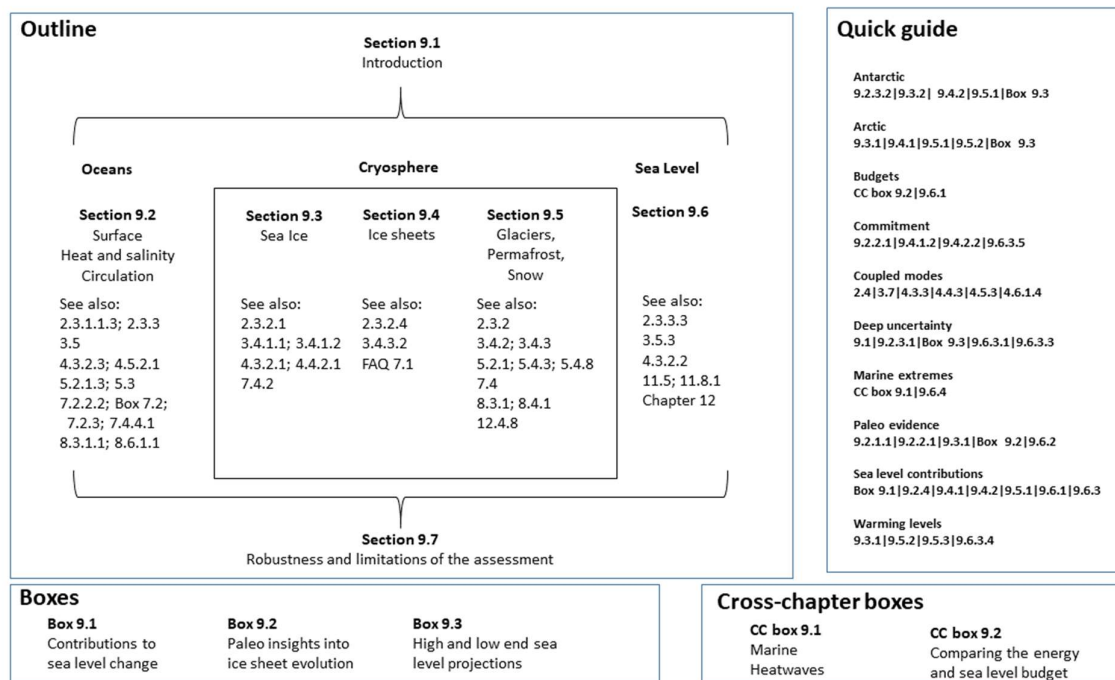


Figure 9.2: Visual guide to chapter 9 with links to relevant assessments in other chapters and common themes across the chapter.

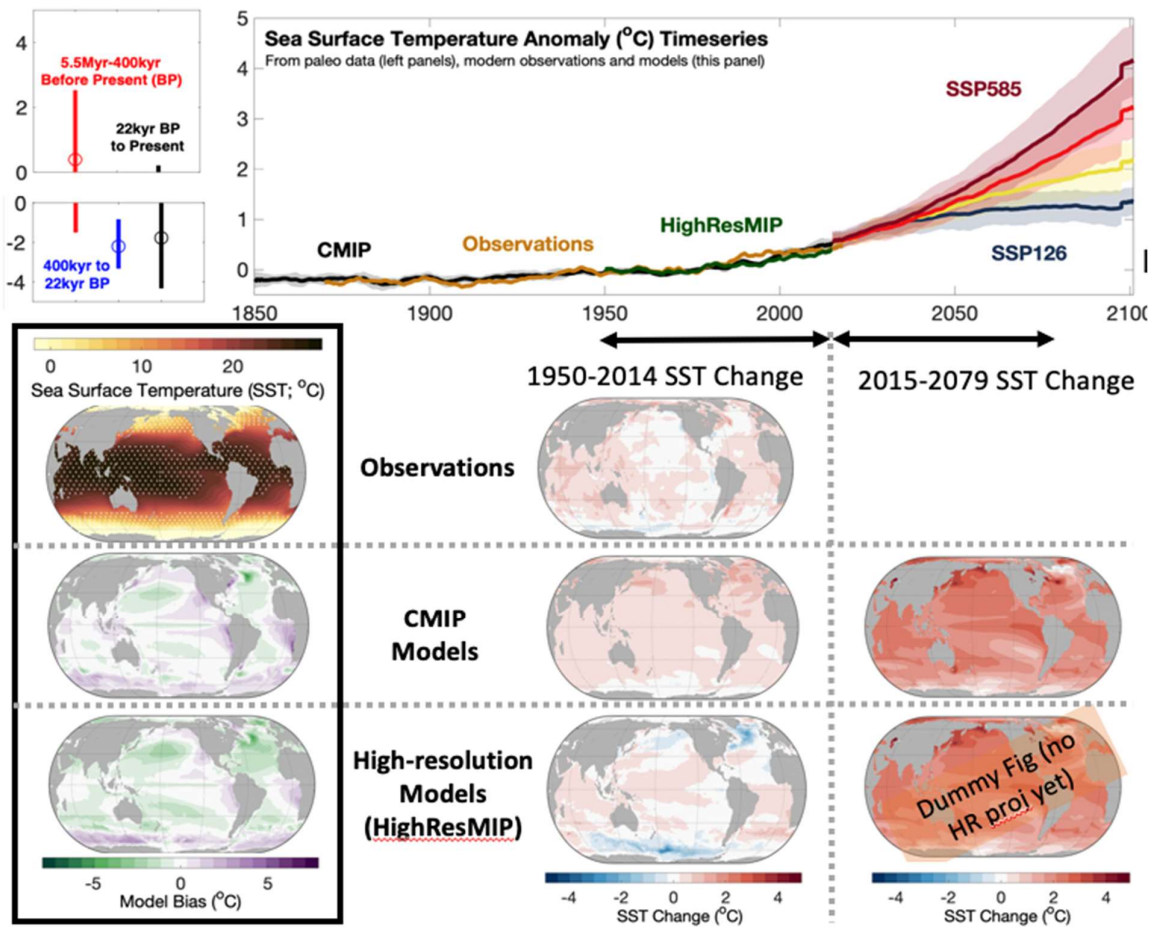


Figure 9.3: Sea Surface Temperature (SST) and its changes with time. (Top) Time series of global mean SST anomaly relative to 1970-1980 climatology. Shown are observations (HadISST), and multi-model means from the CMIP historical and SSP experiments, and from the HighResMIP experiment. (Bottom panels) Maps of observed SST (top left; 1995-2014 climatology HadISST), bias of CMIP (middle left) and HighResMIP (bottom left) over this period. Also shown are 1950-2014 historical SST changes from observations, CMIP and HighResMIP (middle column), and 2015-2079 projected changes under SSP5-8.5 for the CMIP and HighResMIP ensemble (right column).

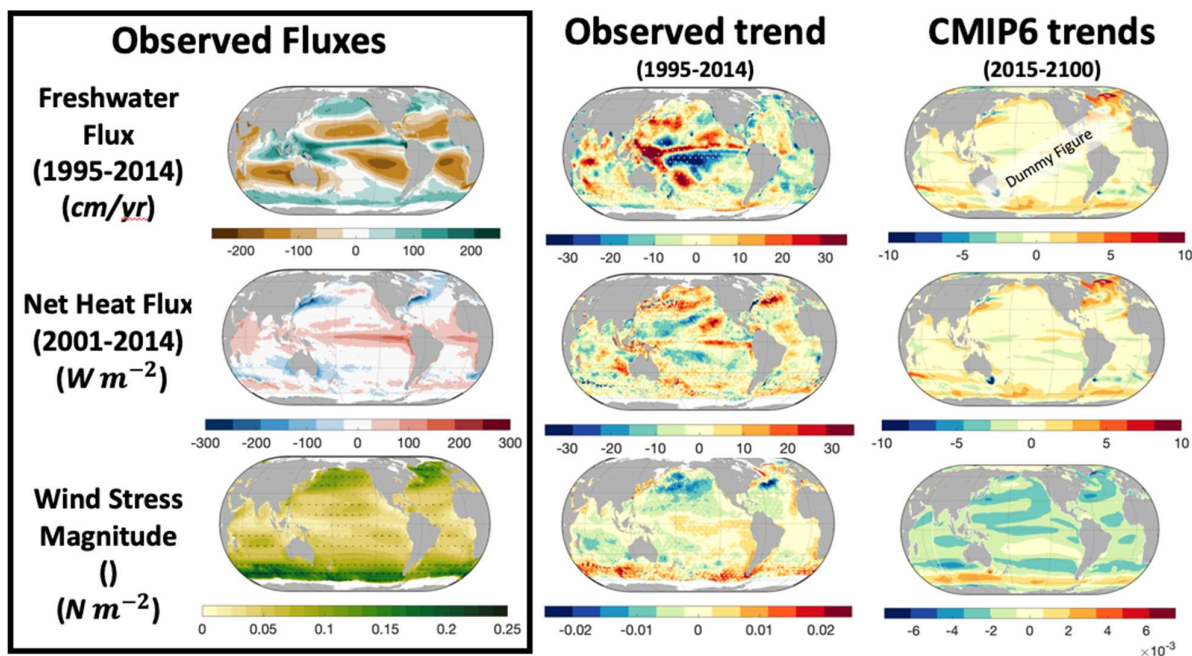


Figure 9.4: Global maps of (left) observed mean fluxes (left column), and the observed (net heat flux 2001-2014 and P-E and winds 1995-2014) and projected (2015-2100) trends of these fluxes (middle and right columns respectively). The fluxes shown are (top) freshwater flux, (middle row) net heat flux, and (bottom) wind stress magnitude, with positive numbers indicating ocean freshening, warming, accelerating, respectively. Objective interpolation from CERES EBAF v4 (Kato et al., 2018), OAFlux-HR (Yu, 2019), and GPCP (Adler et al., 2003) of fluxes and flux trends.

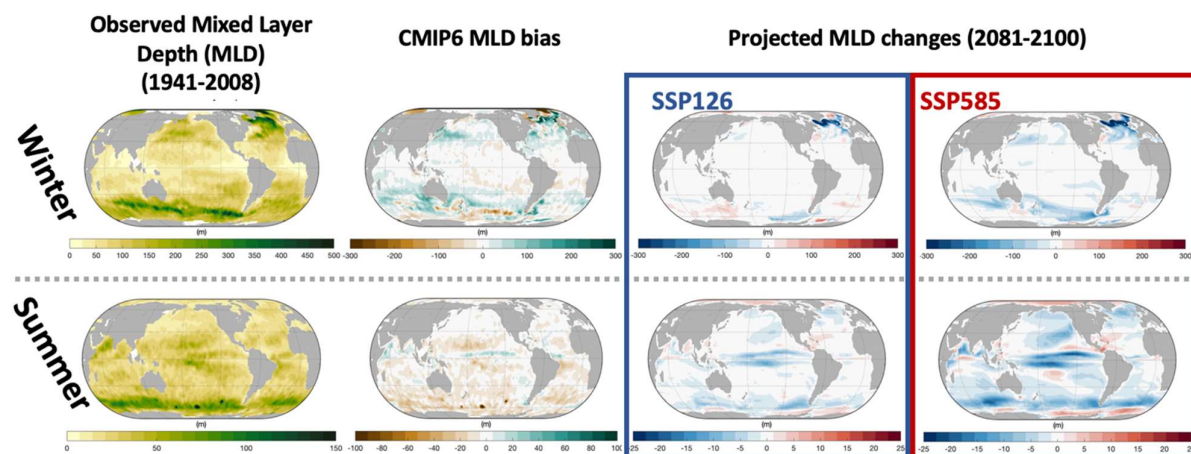
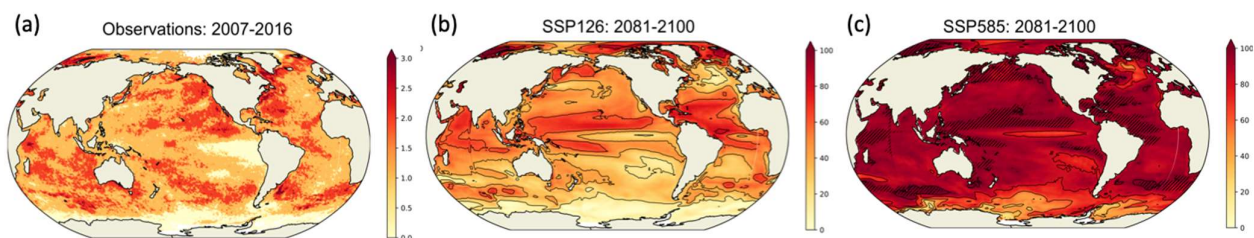


Figure 9.5: Observed climatological mean mixed layer depth from IFREMER/LOS Mixed Layer Depth (updated from de Boyer Montégut et al., 2004) (first column), bias between observation-based estimate and 1995-2014 CMIP6 climatological mean mixed layer depth (second column), projected MLD change from 1995-2014 to 2081-2100 under SSP1-2.6 (third column) and SSP5-8.5 (fourth column) scenarios, for the DJF/JJA in N. Hemisphere/S. Hemisphere (upper) and JJA/DJF in N. Hemisphere/S. Hemisphere (lower). The mixed layer depth is the depth where the potential density is 0.03 kg m^{-3} denser than at 10m. Units: m.



Cross-Chapter Box 9.1, Figure 1: Regional probability ratio of marine heatwaves, the proportion by which the number of marine heatwave days per year has increased. A marine heatwave is defined as a deviation beyond the daily 99th percentile (11-day window) in the deseasonalized SST. (a) The marine heatwave probability ratio of satellite observations during 2007-2016 versus the 1982-2016 period. (b,c) CMIP6 simulated multi-model mean probability ratio in 2081-2100 versus preindustrial for the (b) SSP1-2.6 and (c) SSP5-8.5 scenarios. The hatched areas indicate permanent marine heatwaves in 2081-2100 (>360 marine heatwave days per year). These models are included: MIROC6, NESM3, CanESM5, EC-Earth.

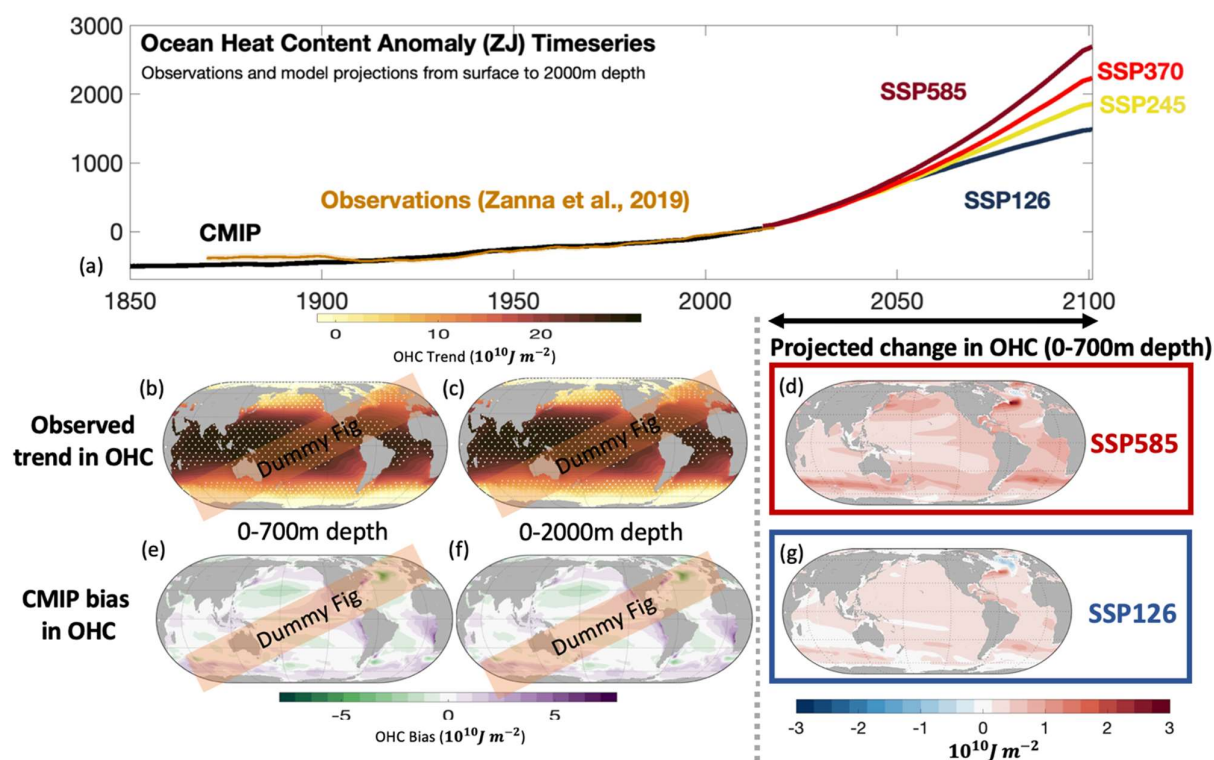


Figure 9.6: Ocean Heat Content and its changes with time. (a) Timeseries of global ocean heat content anomaly relative to 2005-2014 climatology in the upper 2000m of the ocean. Shown are observations (Zanna *et al.*, 2019), and multi-model means from the CMIP6 historical and SSP experiments. (b-g) Maps of Ocean Heat Content across different time periods, in different layers, and from different data sets/experiments. Maps show the observed trends of OHC for (b) 0-700m for the period 1971-2014, and (e) 0-2000m for the period 2005-2017, and the corresponding multi-model bias in these trends for CMIP6 ensembles versus observations (Zanna *et al.*, 2019). [PLACEHOLDER UNTIL THE SIMULATION DATA ARE AVAILABLE]. Also shown are the projected change in 0-700m OHC for (d) SSP1-2.6 and (g) SSP5-8.5 in the CMIP6 ensembles, for the period 2091-2100 versus 2005-2014.

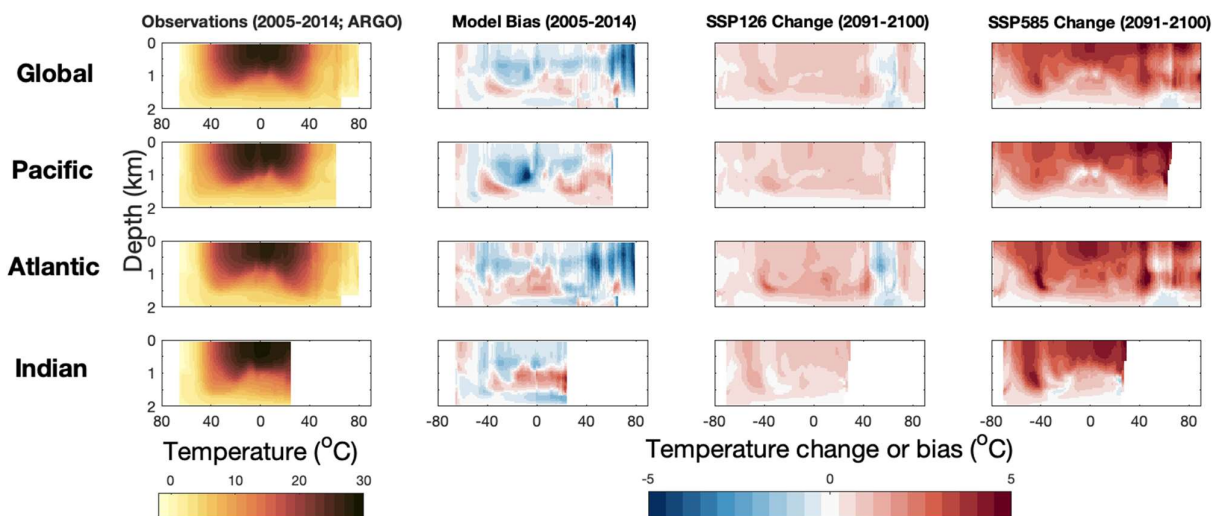


Figure 9.7: Meridional transects of potential temperature in the ocean and its changes with time in the upper 2000m of the Global, Pacific, Atlantic, and Indian Oceans. Shown are observed temperature (ARGO climatology 2005-2014), bias of the CMIP ensemble over this period, and future changes under SSP1-2.6 and SSP5-8.5.

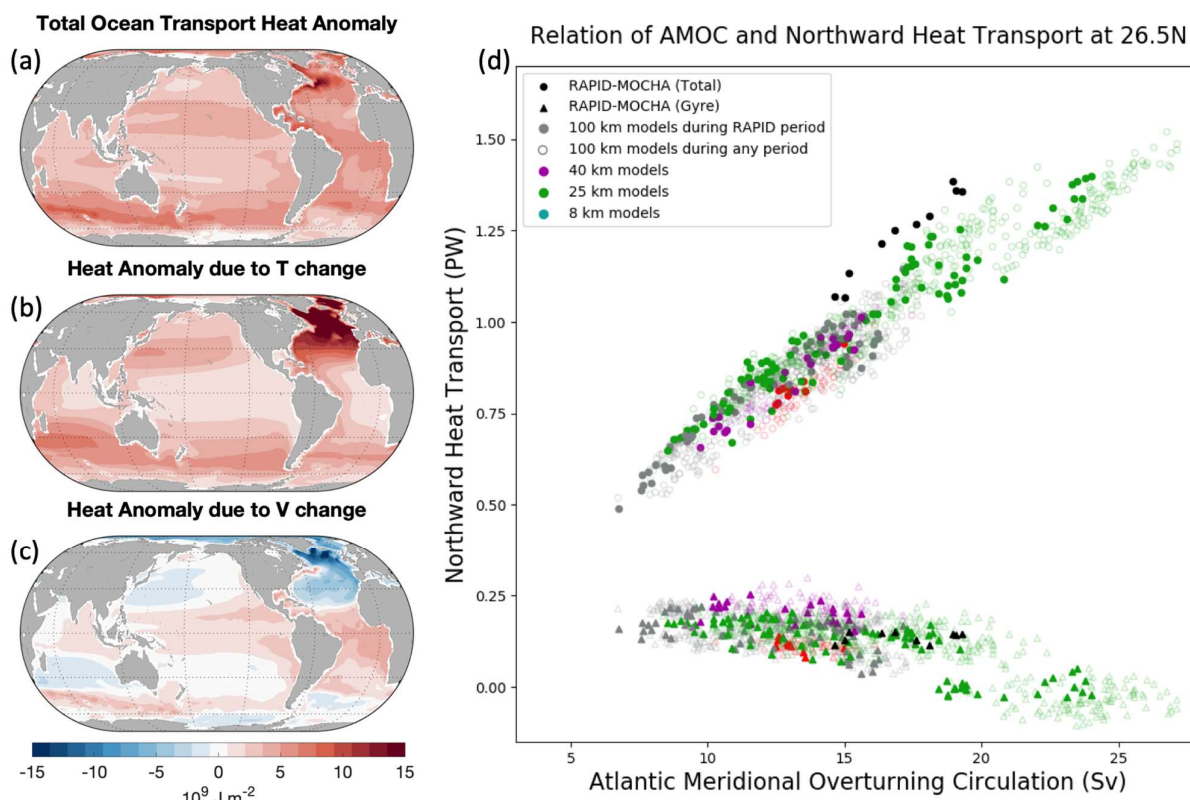


Figure 9.8: (a) Total ocean heat uptake after 100 years of simulation in an idealized simulation where atmospheric CO_2 increase by 1% per year as simulated by five Climate models (CanESM2, CanESM5, GFDL-ESM2M, 687 HadCM3, and MPI-ESM-LR), and associated decomposition of this total heat uptake into (b) passive uptake of added heat (“excess heat”; see text), and (c) pure redistribution of unperturbed heat (“redistributed heat”; see text) (Couldrey *et al.*, 2020). (d) Relationship between northward heat transport and Atlantic Meridional Overturning Circulation in HighResMIP models (1950-2050) and observations during the RAPID period (2004-2018).

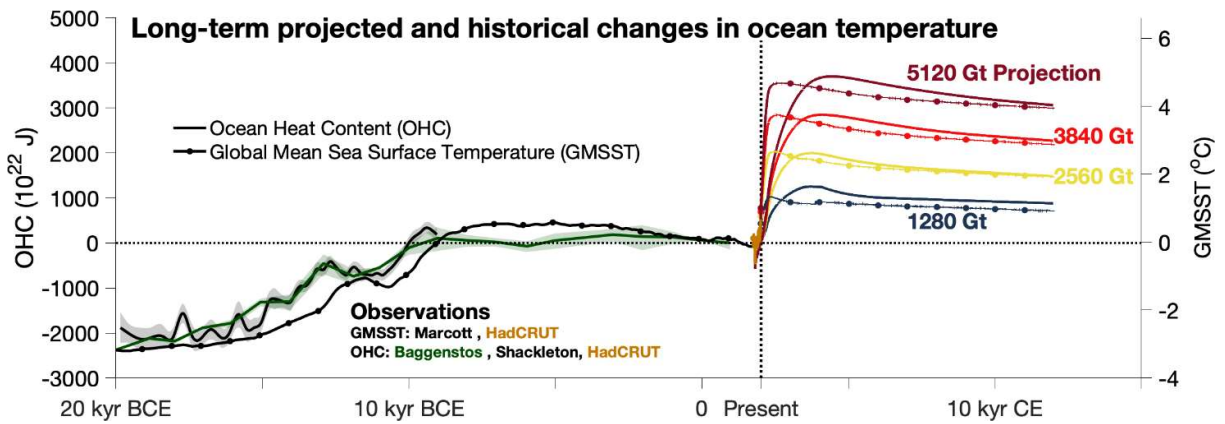


Figure 9.9: Measured changes in past mean ocean heat content OHC (20 kyr BCE to present) estimated from ice-core rare gases (Baggenstos *et al.*, 2019; Shackleton *et al.*, 2019), changes in global mean sea-surface temperature GMSST (Shakun *et al.*, 2012; Marcott *et al.*, 2013), and historical OHC changes from HadCM3 (Gregory, Lowe and Tett, 2006) and observations (Durack *et al.*, 2018; Johnson *et al.*, 2018). Future long-term projected changes in OHC and GMSST (present to 12 kyr CE) in response to four greenhouse gas emissions scenarios over multiple millennia (Clark *et al.*, 2016).

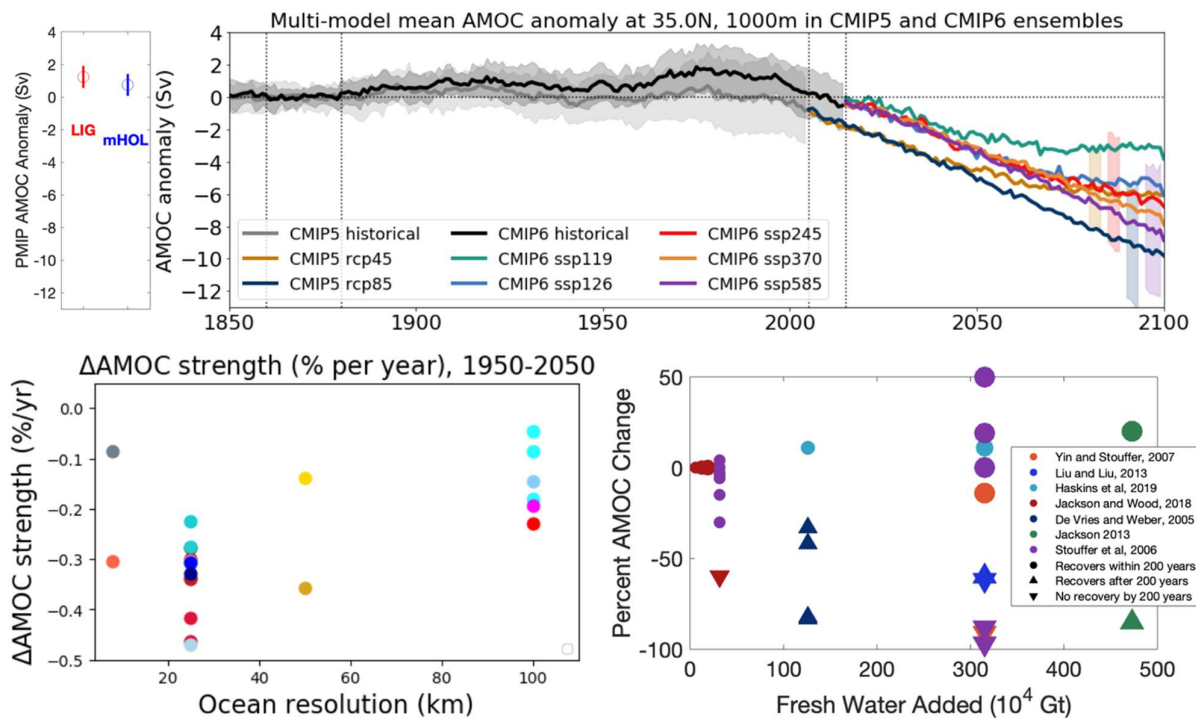


Figure 9.10: (Top left) AMOC magnitude in PMIP experiments. (Top right) Timeseries of AMOC from CMIP5 and CMIP6 based on (Menary et al., submitted). (Bottom left) % change in AMOC strength per year at different resolutions over 1950-2050 period with colours for model families (Roberts et al., submitted). (Bottom right) A compilation (Jackson and Wood, 2018) of percentage changes in the simulated AMOC after applying an additional freshwater flux in the subpolar North Atlantic at the surface for a limited time (de Vries and Weber, 2005; Haskins et al., 2019; Jackson, 2013; Jackson and Wood, 2018; Liu and Liu, 2013; Stouffer et al., 2006; Yin and Stouffer, 2007). Symbols indicate whether the AMOC recovers within 200 years (circles), is starting to recover (upwards arrow) or does not recover within 200 years (downwards arrow). Symbol size indicates rate of freshwater input. (See Supplementary material for further details).

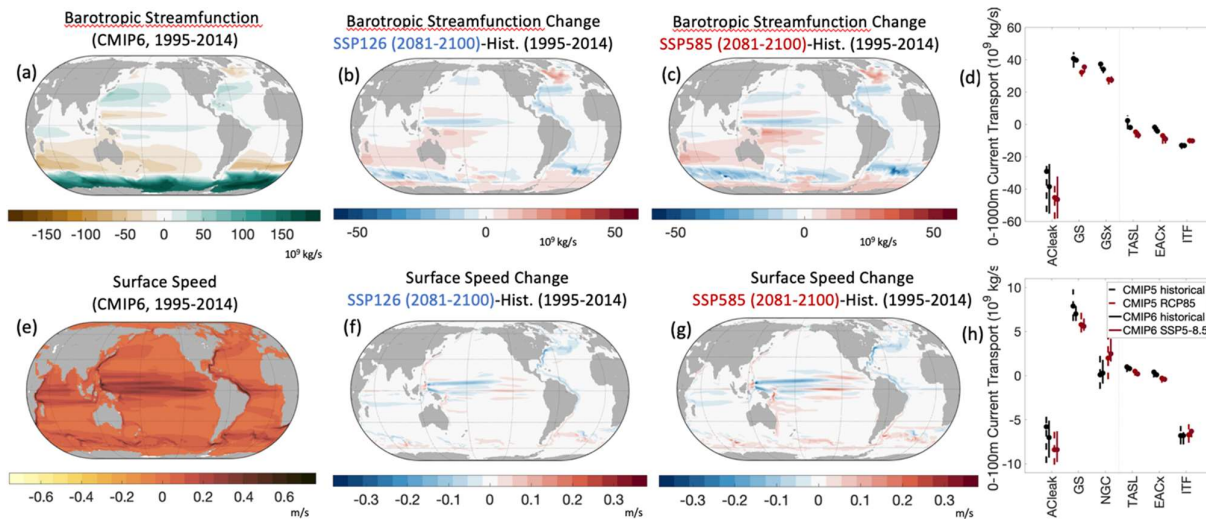


Figure 9.11: (a) Mean barotropic streamfunction 1995-2014 and projected barotropic streamfunction change 2081-2100 versus 1995-2014 under (b) SSP1-2.6 and (c) SSP5-8.5. Units are Sv. (d) Mean surface (0-100 m) speed and projected surface speed change 2081-2100 versus 1995-2014 under (f) SSP1-2.6 and (g) SSP5-8.5. The units are m s^{-1} . (d,h) Median and *likely* range of 1995-2014 and 2081-2100 transport of 6 currents with the largest transport change or largest fractional change (Sen Gupta et al., submitted). (d) Deep currents: Agulhas Leakage (ACleak), Gulf Stream (GS), Gulf Stream Extension (GSx), Tasman Leakage (TASL), East Australia Current Extension (EACx), and Indonesian Throughflow (ITF). (h) Shallow currents: as for deep but with New Guinea Current (NGC), and without GSx.

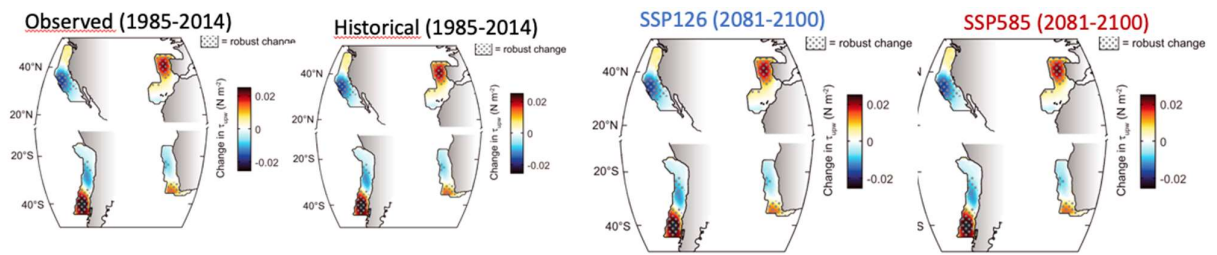


Figure 9.12: [PLACEHOLDER FOR REGENERATING USING REANALYSIS and CMIP6 MODELS INCLUDING HIGH-RES] Upwelling-favourable winds stress (τ_{upw}) in the four main upwelling systems (Rykaczewski et al., 2015). (a) 1985-2014 climatological mean intensity of τ_{upw} from atmospheric reanalysis; (b) 1985-2014 historical mean τ_{upw} as simulated by CMIP6 models. Change τ_{upw} from 1985-2014 to 2081–2100 predicted under (c) SSP126 scenario and (d) SSP585 scenario.

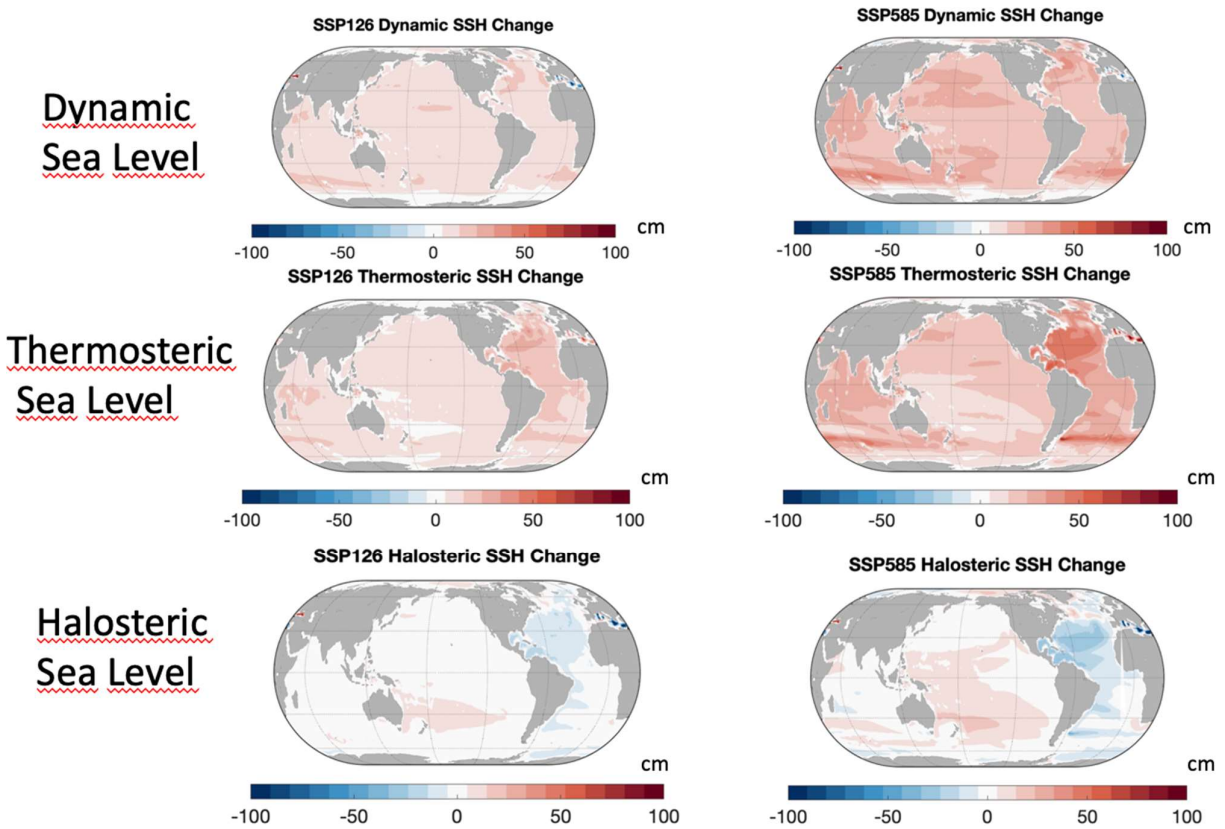


Figure 9.13: CMIP6 multi-model mean projected change in ocean dynamic, thermosteric, and halosteric sea level between 1984-2015 and 2081-2100 using the a method that does not require a reference level (Landerer et al., 2007). The unit is cm.

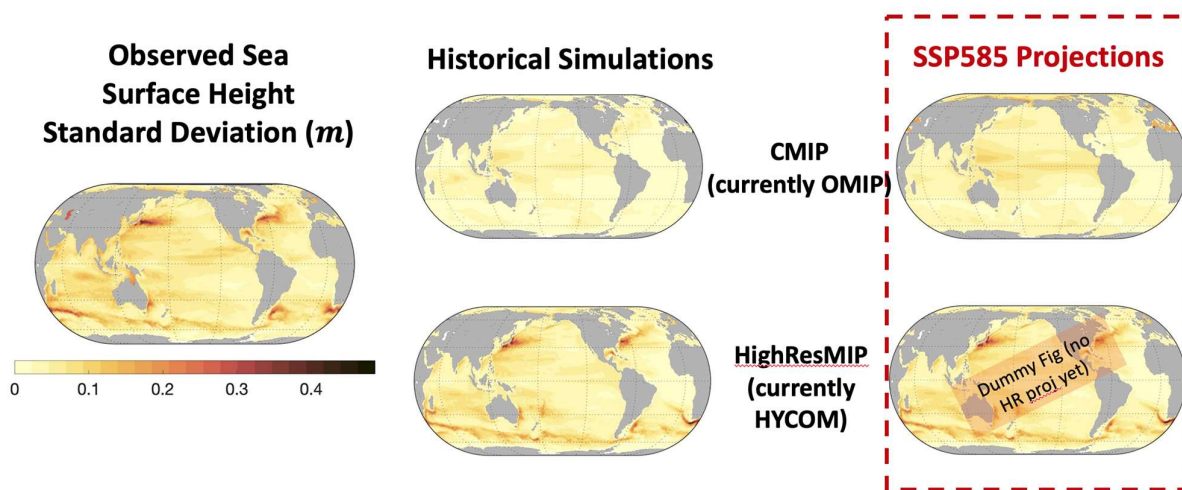


Figure 9.14: Standard deviation of SSH from (left) Aviso observations, and standard deviation of total (thermo/halo/dyn) SSH from (top middle) historical CMIP, (bottom middle) historical HighResMIP, (top right) SSP585 projections of CMIP, and (bottom right) SSP585 projections of HighResMIP. [PLACEHOLDER FOR UPDATE WHEN MORE MODELS ARE UPLOADED]

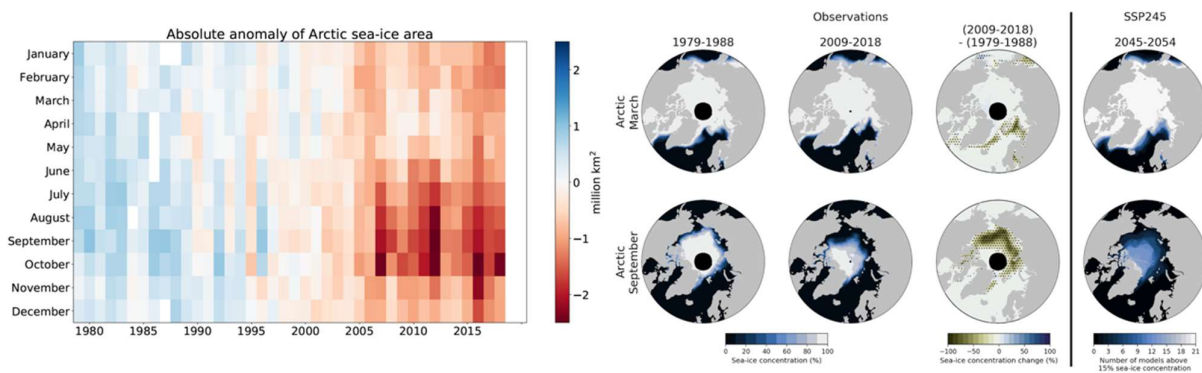


Figure 9.15: (Left): Absolute anomaly of monthly-mean Arctic sea-ice area during the period 1979 to 2018 relative to the average monthly-mean Arctic sea-ice area during the period 1979 to 2008. Based on the average sea-ice area derived from OSISAF/CCI (OSI-450 for 1979-2015, OSI-430b for 2016-2018)(Lavergne et al., 2019), NASA Team (version 1, 1979-2018) (Cavalieri et al., 1996) and Bootstrap (version 3, 1979-2018) (Comiso, 2017) sea-ice concentration estimates. (Right): Sea-ice coverage in the Arctic as given by the average of the three most widely used satellite-based estimates for March and September, which usually are the months of maximum and minimum sea-ice coverage, respectively. First column: Mean sea-ice coverage during the decade 1979-1988. Second column: Mean sea-ice coverage during the decade 2009-2018. Third column: Absolute change in sea-ice concentration between these two decades, with stippling indicating significant differences. Fourth column: number of available CMIP6 models that simulate a mean sea-ice concentration above 15 % for the decade 2045-2055. The observational records are based on the average sea-ice coverage of OSISAF/CCI (OSI-450 for 1979-2015, OSI-430b for 2016-2018)(Lavergne et al., 2019), NASA Team (version 1, 1979-2018)(Cavalieri et al., 1996) and Bootstrap (version 3, 1979-2018)(Comiso, 2017).ⁱ

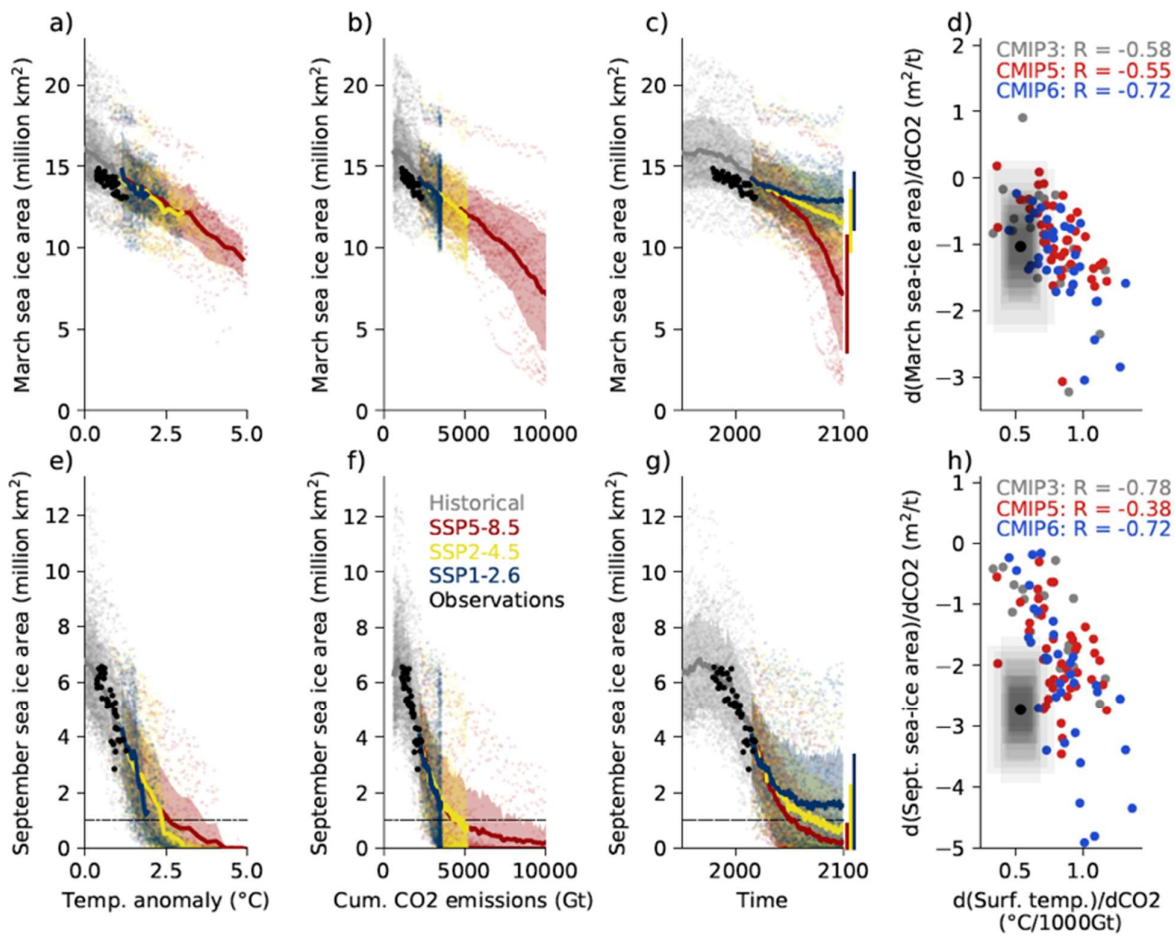


Figure 9.16: Monthly mean March (a-d) and September (e-h) sea-ice area as a function of global surface air temperature (GSAT) anomaly (a,e); cumulative anthropogenic CO₂ emissions (b,f); year (c,g) in CMIP6 model simulations (shading, ensemble mean as bold line) and in observations (black dots). Panels d and h show the sensitivity of sea-ice loss to anthropogenic CO₂ emissions as a function of the modelled sensitivity of GSAT to anthropogenic CO₂ emissions (after SIMIP Community, submitted).

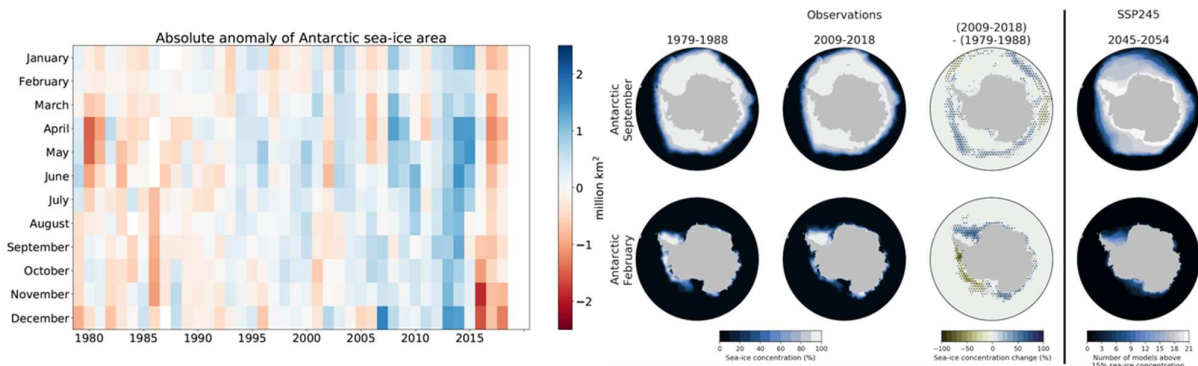


Figure 9.17: (Left): Absolute anomaly of monthly-mean Antarctic sea-ice area during the period 1979 to 2018 relative to the average monthly-mean Antarctic sea-ice area during the period 1979 to 2008. Based on the average sea-ice area derived from OSISAF/CCI (OSI-450 for 1979-2015, OSI-430b for 2016-2018)(Lavergne et al., 2019), NASA Team (version 1, 1979-2018)(Cavalieri et al., 1996) and Bootstrap (version 3, 1979-2018)(Comiso, 2017) sea-ice concentration estimates. (Right): Sea-ice coverage in the Antarctic as given by the average of the three most widely used satellite-based estimates for September and February, which usually are the months of maximum and minimum sea-ice coverage, respectively. First column: Mean sea-ice coverage during the decade 1979-1988. Second column: Mean sea-ice coverage during the decade 2009-2018. Third column: Absolute change in sea-ice concentration between these two decades, with stippling indicating significant differences. Fourth column: number of available CMIP6 models that simulate a mean sea-ice concentration above 15 % for the decade 2045-2055. The observational records are based on the average sea-ice coverage of OSISAF/CCI (OSI-450 for 1979-2015, OSI-430b for 2016-2018)(Lavergne et al., 2019), NASA Team (version 1, 1979-2018)(Cavalieri et al., 1996) and Bootstrap (version 3, 1979-2018)(Comiso, 2017).

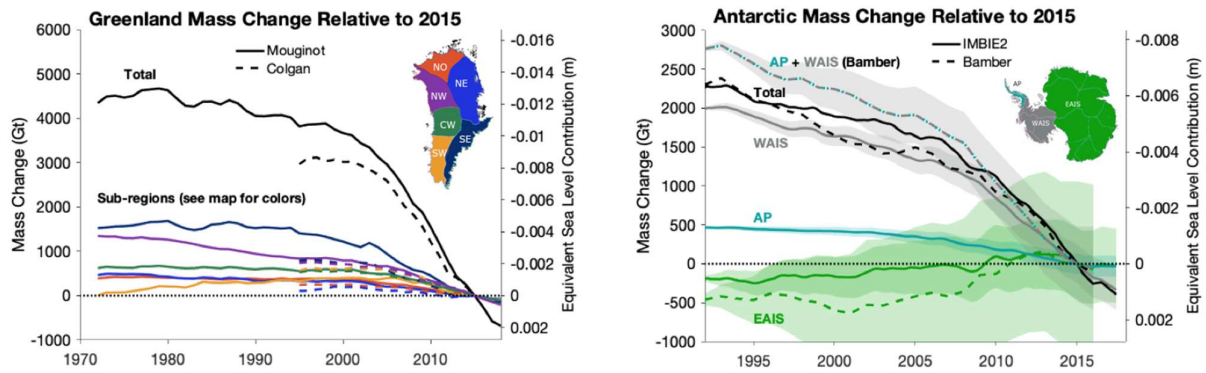


Figure 9.18: (a) Timeseries of mass changes in Greenland for each of the major drainage basins shown in the inset figure (Colgan et al., 2019; Mouginot et al., 2019) for the periods 1972 – 2018 and 1992–2018. (b) Timeseries of mass changes of three portions of Antarctica (Bamber et al., 2018a; The IMBIE Team, 2018) for the period 1992 – 2018

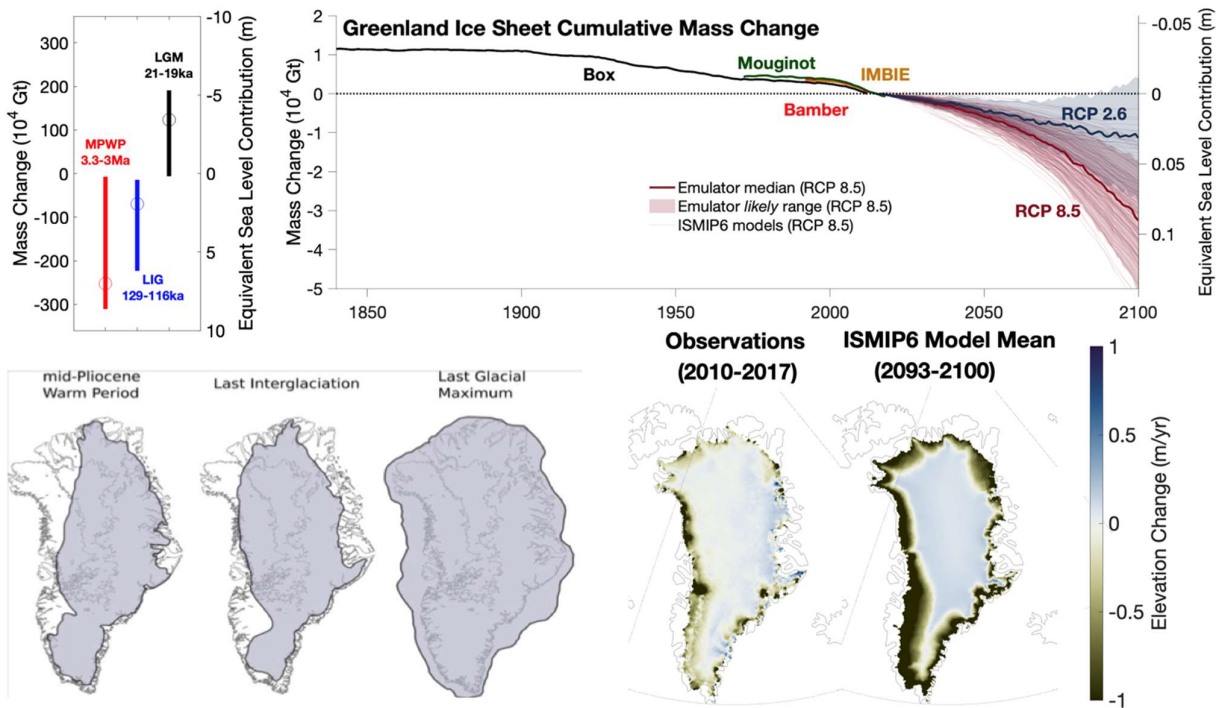


Figure 9.19: Estimated sea level equivalent volume relative to present day (top left panel, detail in supplementary material) and spatial extent (bottom left) of the Greenland ice sheet during the mid-Pliocene Warm Period, the Last Interglaciation, and the Last Glacial Maximum (see Annex II for details). Paleo extents in bottom panels are interpreted from individual reconstructions (Goelzer et al., 2017; Koenig et al., 2015; Lecavalier et al., 2014). Grey shading shows extent of grounded ice. (top right) Cumulative mass loss (sea level equivalent) from the Greenland ice sheet since 1840 (updated from (Box and Colgan, 2013; Kjeldsen et al., 2015) reconstruction), with satellite observations shown from 1972 (Mouginit et al., 2019) and 1993 (Bamber et al., 2018b; The IMBIE Team, 2019), and projections from ISMIP6 by 2100 under RCP8.5/SSP5-8.5 and RCP2.6/SSP1-2.6 scenarios (thin lines from (Goelzer et al., submitted; Nowicki et al., submitted, a) and likely range of the ISMIP6 emulation (shades and bold line (Edwards et al., submitted)). (bottom right) Maps of mean elevation changes 2010-2017 derived from CryoSat 2 radar altimetry (Bamber et al., 2018b) and ISMIP6 model mean (2061-2100) projected changes (Goelzer et al., submitted).

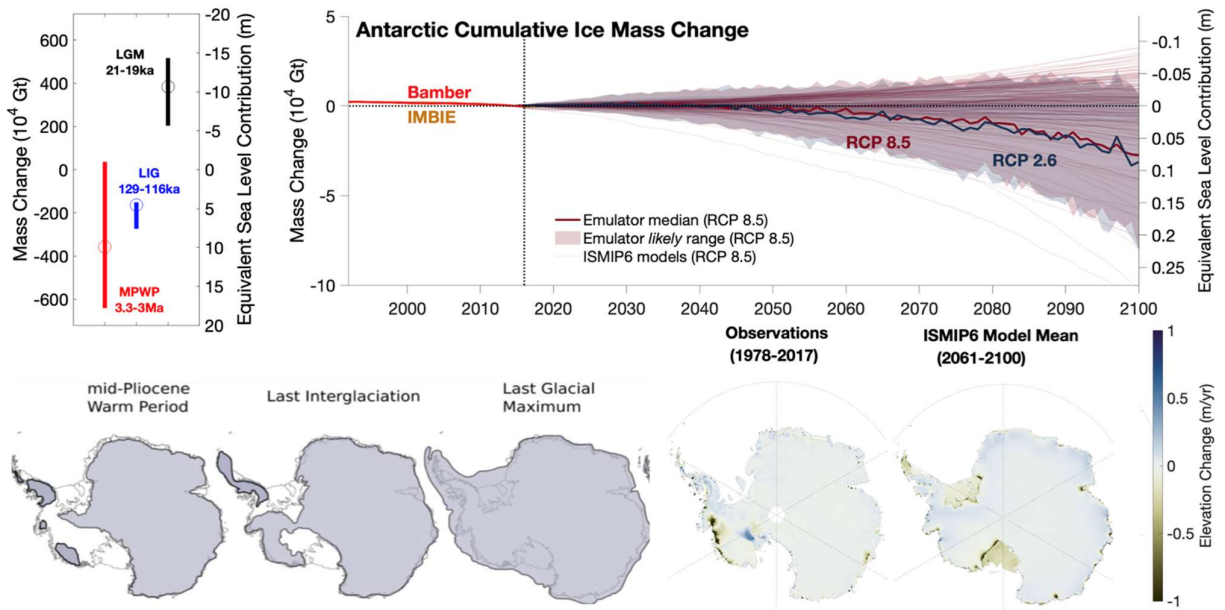


Figure 9.20: Estimated sea level equivalent volume relative to present day (top left panel, detail in supplementary material) and spatial extent (bottom left) of the Antarctic ice sheet during the mid-Pliocene Warm Period, the Last Interglaciation, and the Last Glacial Maximum (see Annex II for details) compared to present-day extent (Fretwell et al., 2013). Paleo extents are interpreted from individual reconstructions (Anderson et al., 2002; Bentley et al., 2014; De Boer et al., 2015; Goelzer et al., 2016a). Grey shading shows extent of grounded ice. (top right) Cumulative mass loss (sea level equivalent) from the Antarctic ice sheet since 2015, with satellite observations shown from 1993 (Bamber et al., 2018a; The IMBIE Team, 2018; WCRP Global Sea Level Budget Group, 2018), and ISMIP6 projected changes by 2100 under RCP8.5/SSP5-8.5 and RCP2.6/SSP1-2.6 scenarios (thin lines from (Seroussi et al., 2019) and (Nowicki et al., submitted, a)) and likely range of the ISMIP6 emulation (shaded line, (Edwards et al., submitted)). (bottom right) Maps of mean elevation changes 1978-2017 derived from restored analog radar records (Schroeder et al., 2019) and projected changes by 2100 under selected RCP 8.5 scenarios (Seroussi et al., 2019).

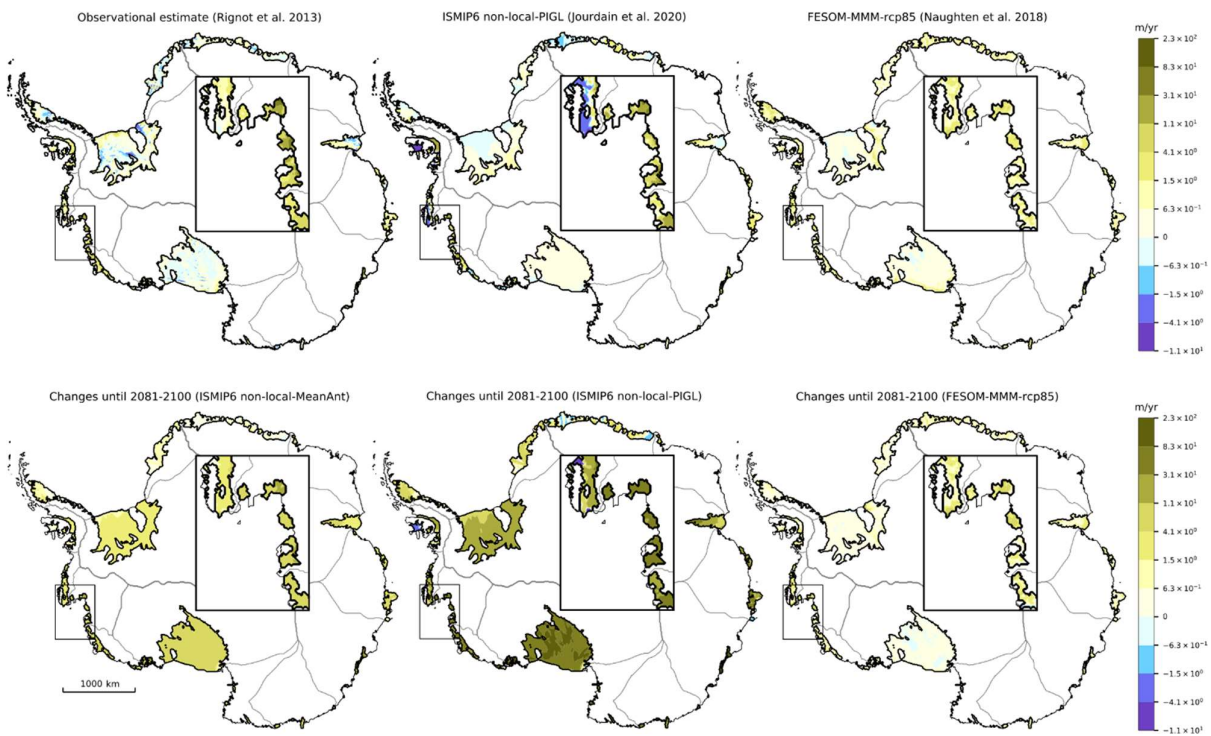


Figure 9.21: Ice shelf basal melt rates for present-day (upper panels) and changes from present-day to the end of the 21st century under the RCP8.5 scenario (lower panels). Present-day melt rates were estimated through: the input-output method constrained by satellite data and atmosphere/snow simulations (Rignot et al., 2013) and representative of 2003-2008 (upper left); the non-local-PIGL parameterization constrained by observation-based ocean properties (Jourdain et al., submitted) and representative of 1995-2014 (upper centre); the Finite Element Sea-ice/ice-shelf Ocean Model (FESOM) simulation over 2006-2015, forced by atmospheric conditions from a CMIP5 multi-model mean (MMM) under the RCP8.5 scenario (Naughten et al., 2018) upper right). Future anomalies are calculated as 2081-2100 minus present-day using the ISMIP6 non-local-MeanAnt and non-local-PIGL parameterizations ((Jourdain et al., submitted) lower left and centre respectively) based on projections from the NorESM1-M CMIP5 model, and the FESOM-MMM projection (lower right). Note the symmetric-log colour bar (linear around zero, logarithmic for stronger negative and positive values). Inset highlights the Amundsen Sea Region.

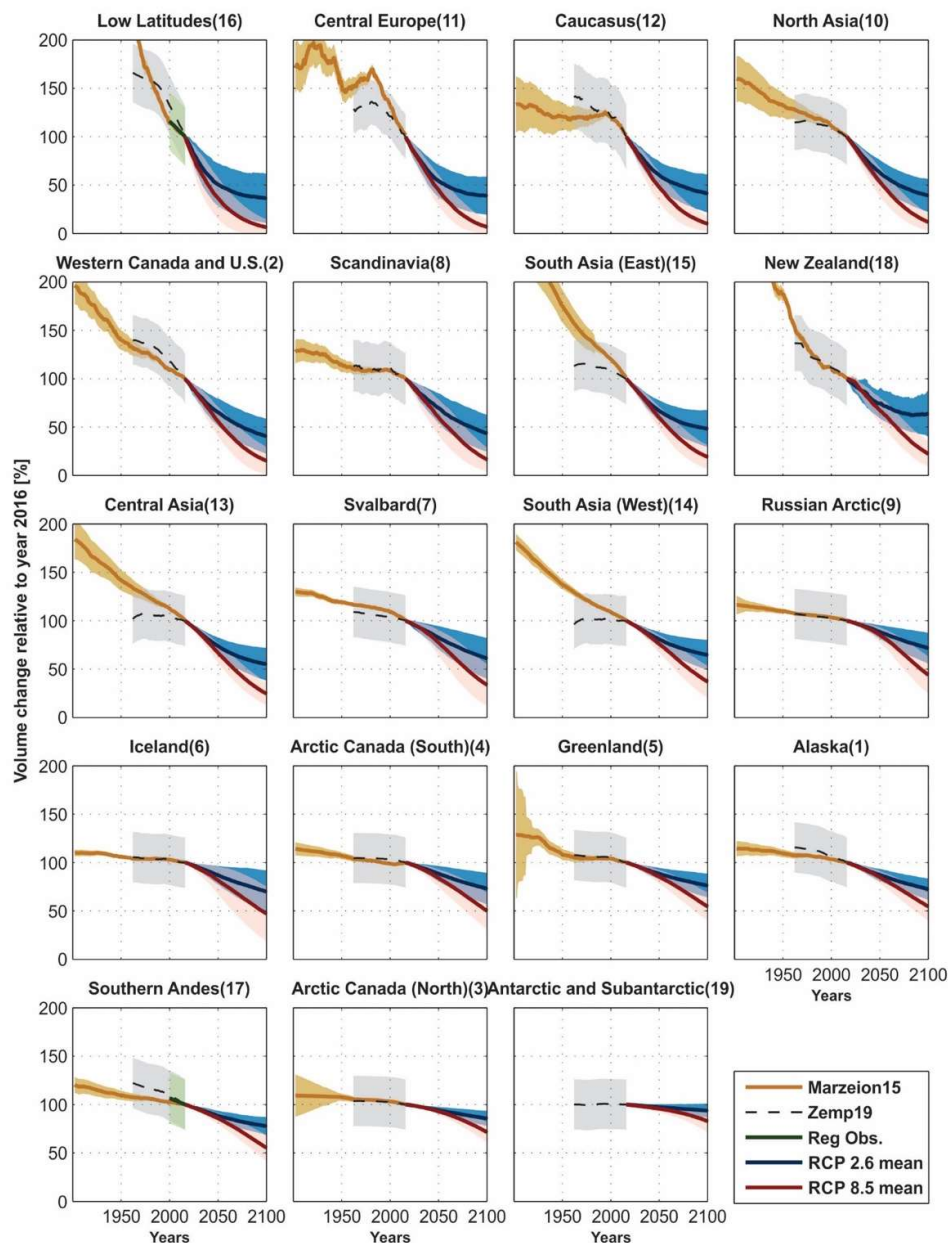


Figure 9.22: Modelled, observed and project volume change of glaciers in the 19 GTN-G regions from 1902 to 2100 (see Figure 9.1 for location) relative to the volume in year 2016. Regions are ordered from largest mass loss in 21st century in top left and the smallest mass loss in bottom right. Light brown lines and shaded areas indicate the regional glacier change through the 20th century and its respective uncertainty given by Marzeion et al. (2015b). Black lines and grey shaded areas indicate the regional glacier mass change between 1961 to 2016 and its respective uncertainty obtained by Zemp et al. (2019). Green lines and shaded areas (Region 16 and 19 [Placeholder for more regional studies]) indicate regional estimates where available (Dussailant et al., 2019). Two scenarios, RCP2.6 and RCP8.5 from GlacierMIP Phase 2 (Marzeion et al., submitted), are shown. See Table 9.3 for projected mass change in each region. For a better comparison between regions, the maximum relative volume was set to 200%, although for some regions the volume changes between 1902 and 2016 exceeded that value. For the low latitude, New Zealand and South Asia (East) glaciers the changes were larger than 1000%, 350%, and 250%, respectively.

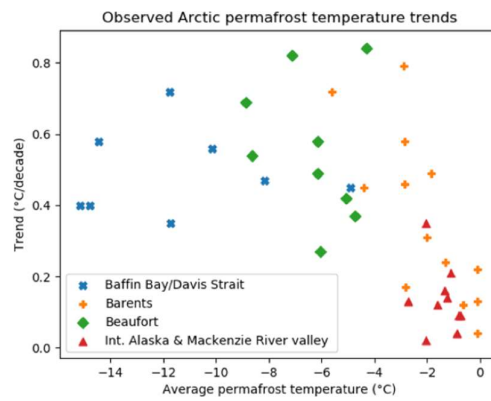


Figure 9.23: Observed recent rate of terrestrial Arctic permafrost temperature change (°C/decade) (updated from (Romanovsky et al., 2018)) as a function of local average permafrost temperature (°C). Coastal regions are identified by the name of the adjacent sea (Baffin Bay/Davis Strait, Barents and Beaufort Seas).

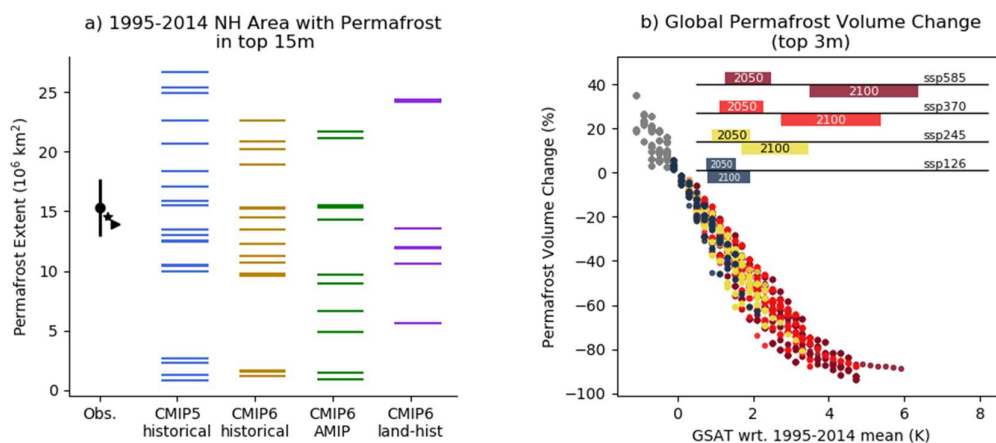


Figure 9.24: a) Diagnosed NH permafrost extent (area with perennially frozen ground at 15 m depth, or at the deepest model soil level if this is above 15 m) for 1979-1998, for available CMIP5 and CMIP6 models, from the first ensemble member of the historical coupled run, and for CMIP6 Amip (atmosphere+land surface, ocean fixed) and LS3MIP land-hist (land only, atmosphere prescribed) runs. Observed and reanalysis-based current permafrost extents are indicated as black symbols (triangle: Obu et al. (2018); star: Zhang et al. (1999); circle: central value and associated range from Gruber (2012)). b) Simulated global permafrost volume change between the surface and 3 m depth as a function of the simulated GSAT change, from the first ensemble members of a selection of scenarios, for available CMIP6 models.

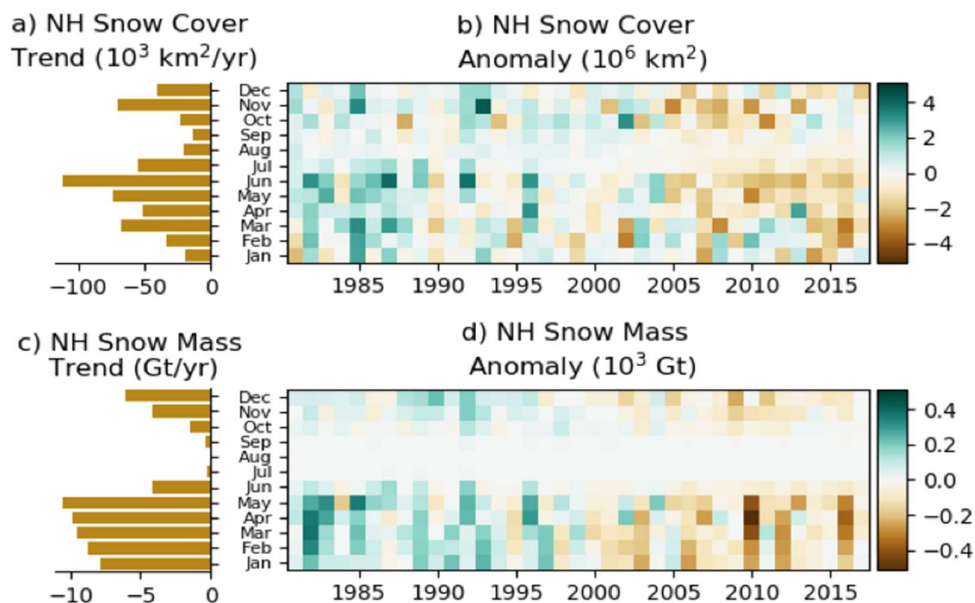


Figure 9.25: Observed monthly northern hemisphere snow cover trends (a) and anomalies (b), and snow mass trends (c) and anomalies (d) from the observation-based ensemble discussed in the text (Mudryk et al., submitted). Trends and anomalies are calculated over the 1981-2017 period.

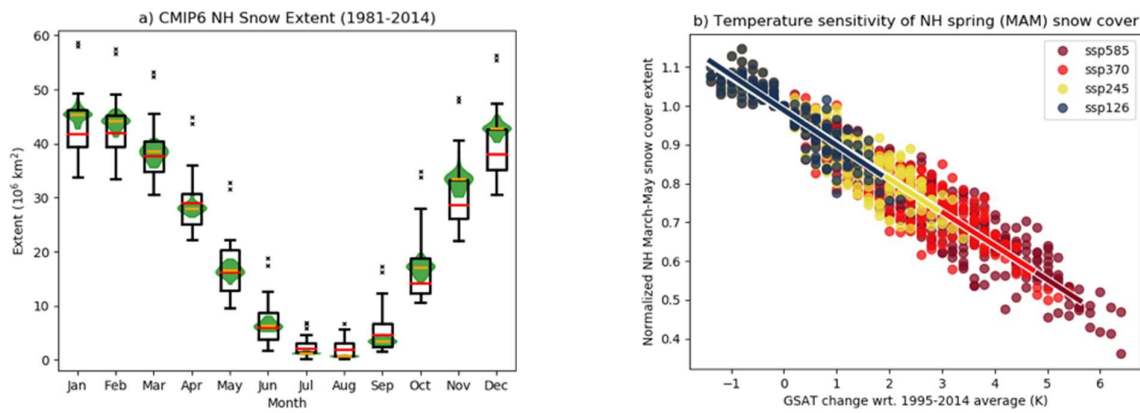
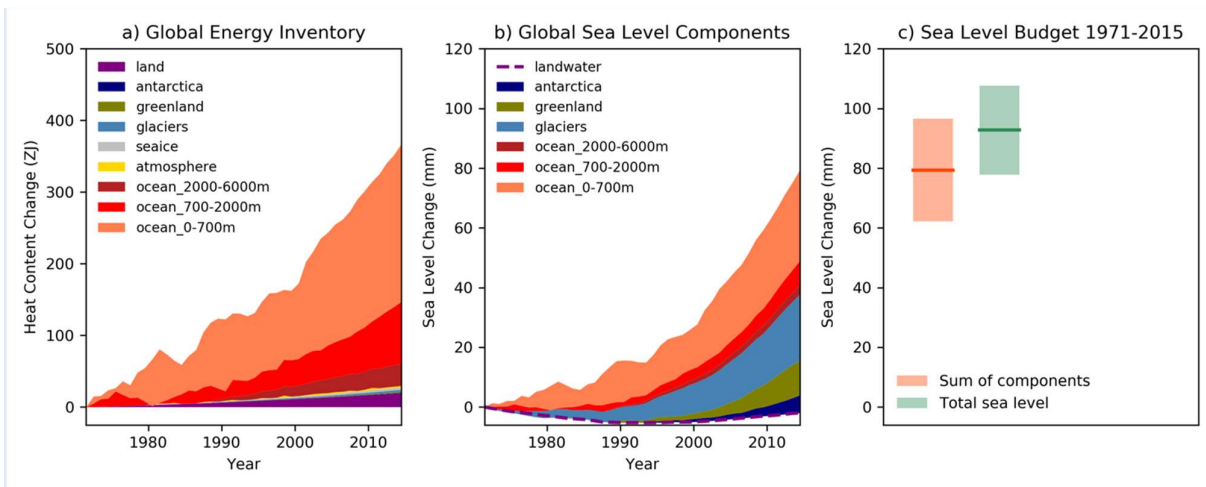


Figure 9.26: a) Simulated CMIP6 and observed (Mudryk et al., submitted) snow cover extent (in millions of km²) for 1981-2014. Boxes and whiskers with outliers represent monthly mean values for the individual CMIP6 models averaged over 1981-2014, with the red bar indicating median of the CMIP6 multi-model ensemble for that period. The observed interannual distribution over the period is represented in green, with the yellow bar indicating the median. b) Spring (March to May) NH snow cover extent against GSAT (relative to the 1995-2014 average) for the CMIP6 Tier 1 scenarios (SSP1-2.6, SSP2-4.5, SSP3-7.0 and SSP5-8.5), with linear regressions. Each data point is the mean for one CMIP6 simulation (first ensemble member for each available model) in the corresponding temperature bin.



Cross-Chapter 9.2, Figure 1: a) Observed changes in the global energy inventory components for 1971-2015; b) Observed changes in global mean sea level components for 1971-2015; c) Sea level budget for 1971-2015, comparing the sum of components in b) to the total reconstructed sea-level change (Dangendorf et al., 2019). a) and b) show individual components in shaded colours as indicated in the figure legends. Shading in c) indicates the 5-95 percentile range. Full details of the datasets and methods used available in the Annex I. Corresponding sea-level trends are in Table 9.4.

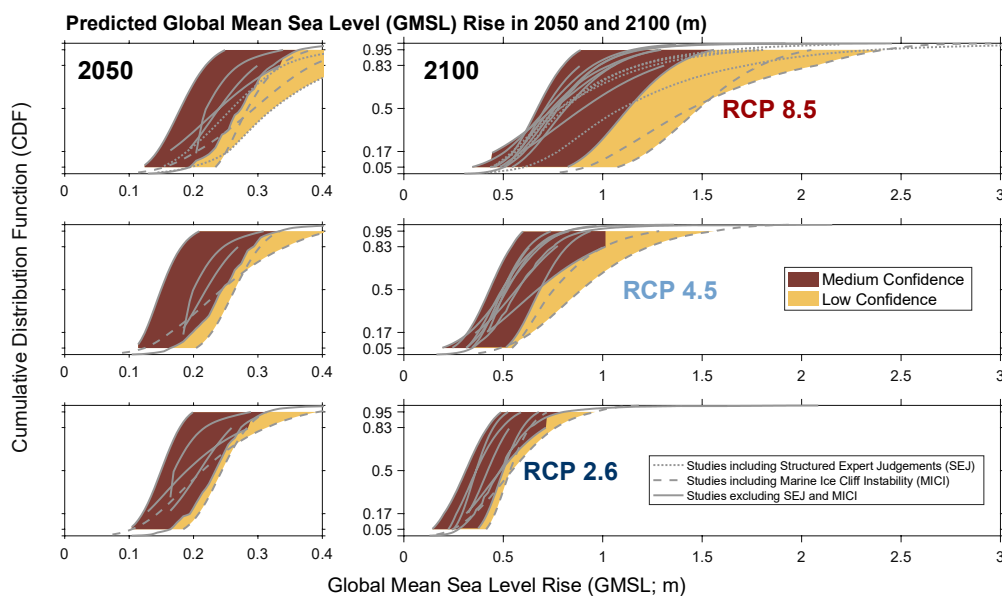


Figure 9.27: Partial cumulative distribution functions (CDFs) from literature global mean sea level (GMSL) projections (m) for 2050 (left) and 2100 (right), for RCP8.5 (upper), RCP4.5 (middle) and RCP4.5 (lower). For each study, the values along the x-axis represent different GMSLs (standardized to account for minor differences in time periods) and the y-axis represents the probability that each study ascribes to that level not being exceeded, conditional upon the scenario. Dark brown shading indicates the *medium confidence* p-box (excluding studies incorporating MICI or relying on structured expert judgement [SEJ]) and light brown shading indicates the *low confidence* p-box (including MICI and studies relying on SEJ). Because of limited availability of reported outcomes, the p-box is not shaded below the 5th or above the 95th percentile.

1

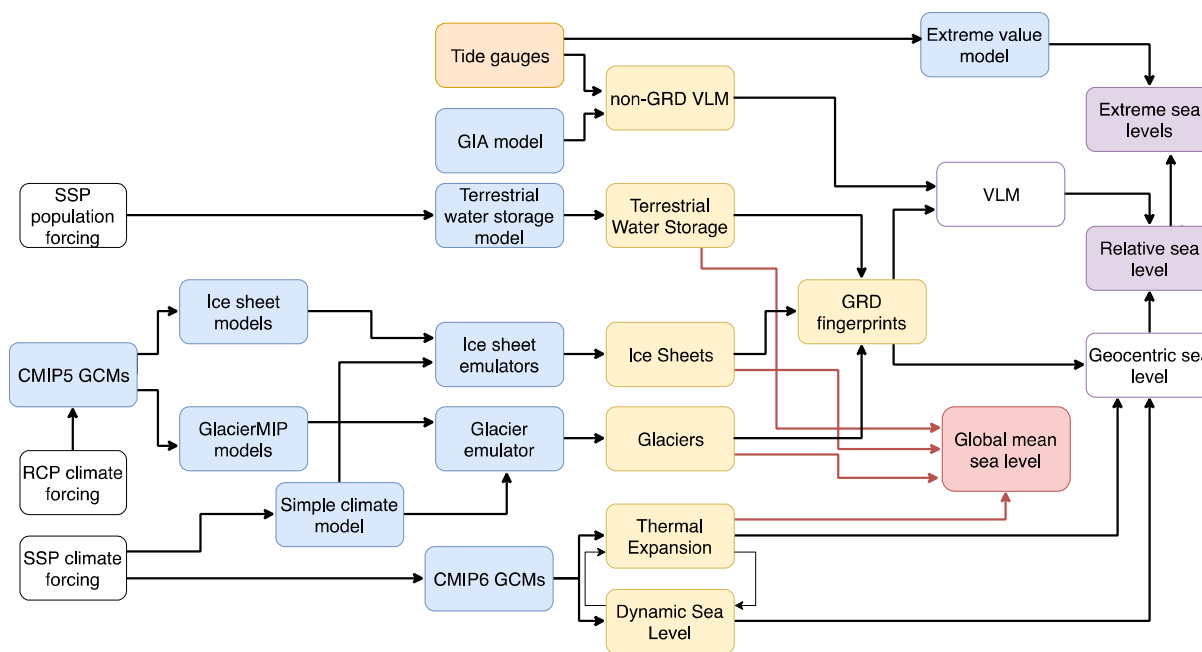


Figure 9.28: Flowchart illustrating the procedure for computing the global mean sea level change (red), regional relative sea level change (purple) and extreme sea level change (purple) projections framework. White blocks with black outline represent external forcing, blue blocks represent model options, orange block represents tide gauge data, yellow blocks represent sea-level contributors, white blocks with purple outline the corrections for vertical land movement.

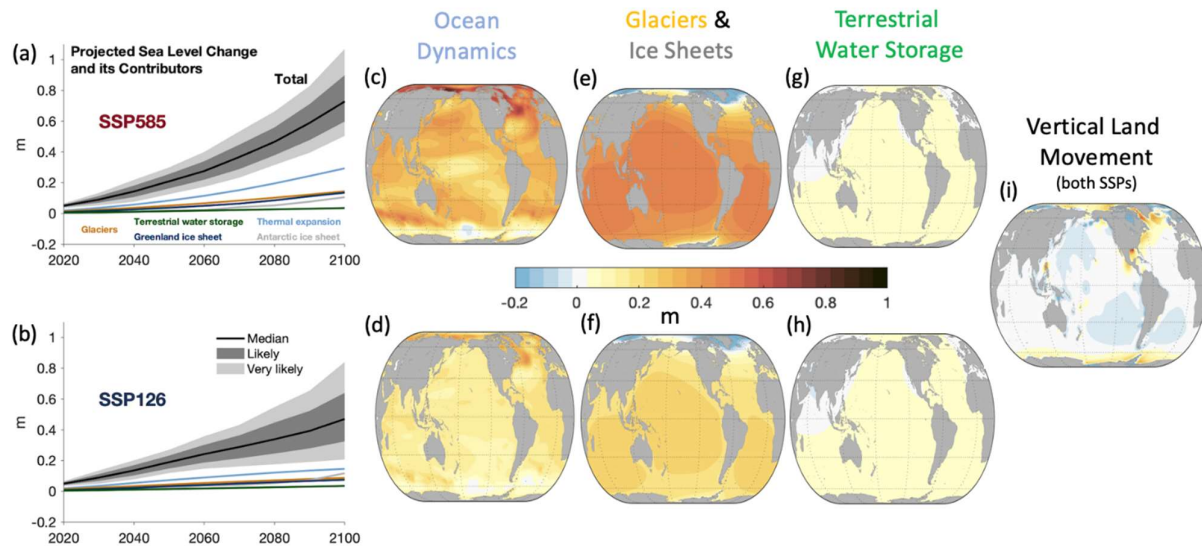


Figure 9.29: Global mean and regional relative sea-level projections (m) by contribution for the SSP5-8.5 and SSP1-2.6 scenarios. (a) and (b) global mean contributions to sea-level change as a function of time, relative to 1995-2014, (b)-(k) regional projections of the sea-level contributions in 2100 relative to 1995-2014 for SSP5-8.5 (c, e, g) and SSP1-2.6 (d, f, h). (c)-(d) Ocean contribution (includes ocean dynamics, steric change, inverse barometer), (d)-(f) ice sheets and glaciers, (g)-(h) Terrestrial Water Storage, (i) Vertical Land Movement (common to both SSPs).

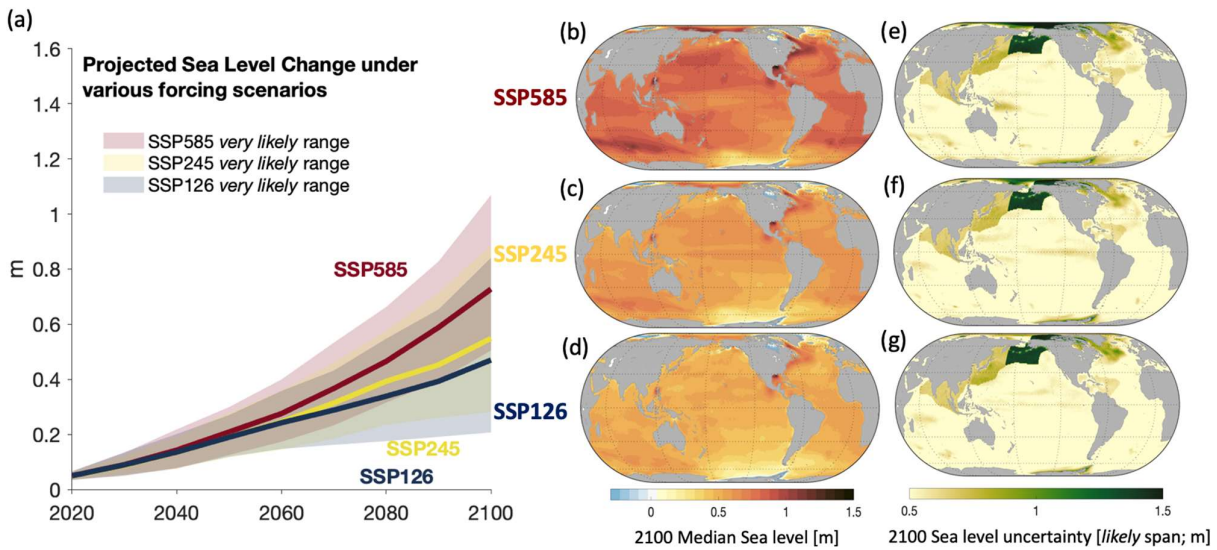


Figure 9.30: Global mean sea-level change for 3 SSP-RCP scenarios (a) and regional relative sea-level (b-g) change for 3 scenarios between 1995-2014 and 2100 (m): (b, e) SSP5-8.5, (c, f) SSP2-4.5, and (d, g) SSP1-2.6; projection median (b-d) and 17th-83rd percentile range (e-g). [PLACEHOLDER TO UPDATE TO 5 SCENARIOS FOR FGD WHEN DATA IS AVAILABLE]

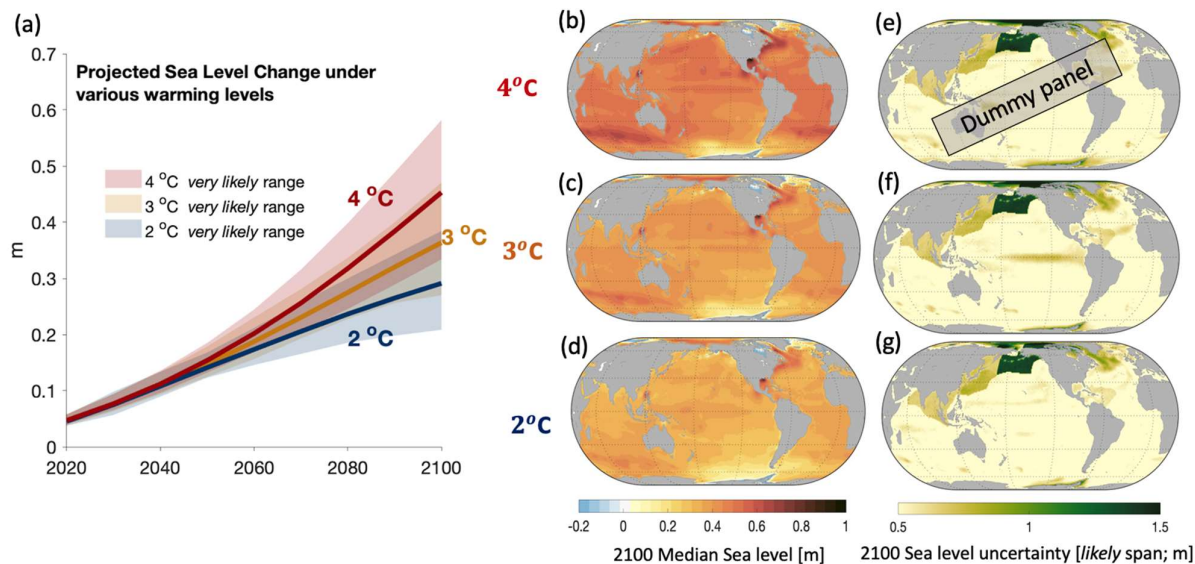


Figure 9.31: Global mean sea-level change for 3 GSAT temperature levels (a) and regional relative sea-level (b-g). The change for 3 temperature levels between 1850-1900 and 2081-2100: (b, e) 4 °C, (c, f) 3 °C, and (d, g) 2 °C. 1996-2014 relative to 2100 projection median (b-d) and 17th-83rd percentile range (e-g). [PLACEHOLDER PANEL e AWAITING DATA UPLOAD]

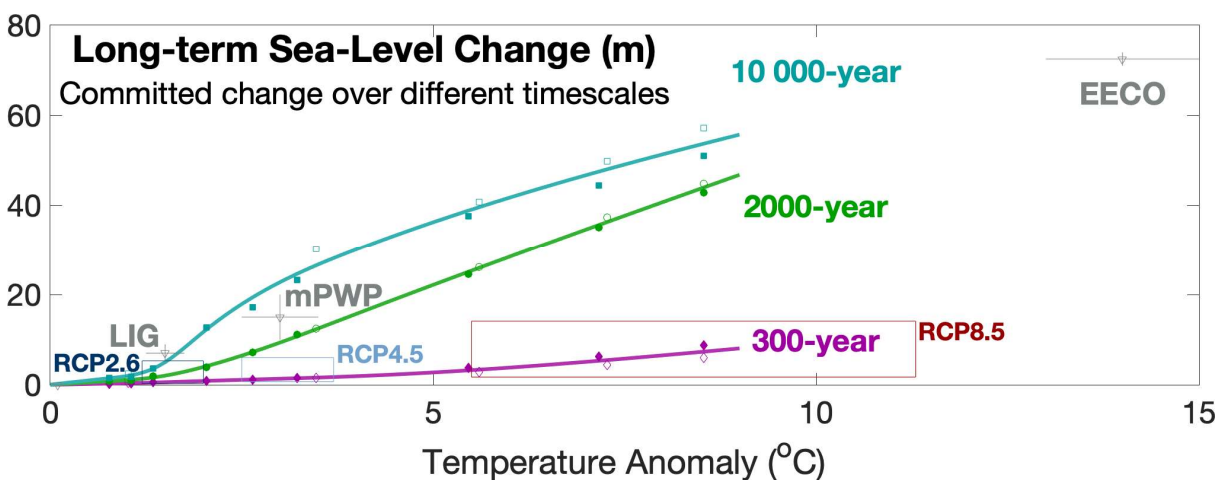


Figure 9.32: Global mean sea-level commitment as a function of global mean surface temperature from models and paleo data. Grey dots and ranges indicate constraints from the Common Era, the LIG, the MPWP, and the early Eocene Climatic Optimum. Blue, green and purple dots and fits indicate respectively the 10,000-year sea-level commitments, 2000-year sea-level commitments and 300-year sea-level commitments from (Clark et al., 2016b). Boxes indicate the scenario ranges from CMIP5.

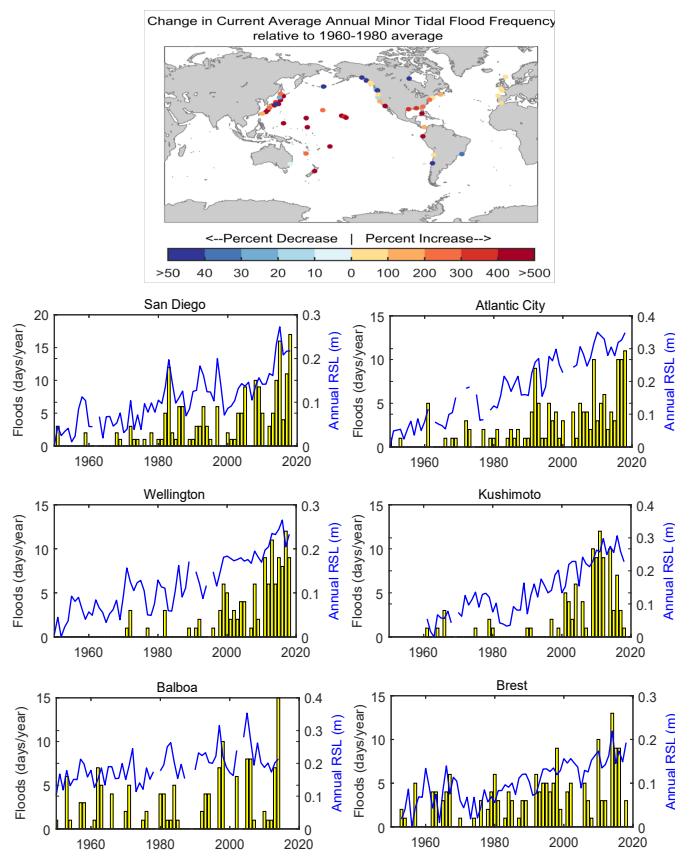


Figure 9.33: Historical occurrences of minor extreme sea levels, defined as the 99th percentile of daily observed water levels over 1995-2014. (a) Percent change in occurrences over 1995-2014 relative to those over 1960-1980. (b-f) Annual mean sea level (blue) and annual occurrences of extreme sea levels over the 1995-2014 99th percentile daily maximum (yellow) at six selected tide gauge locations. [PLACEHOLDER FIGURE FROM SROCC WILL BE REDRAWN]

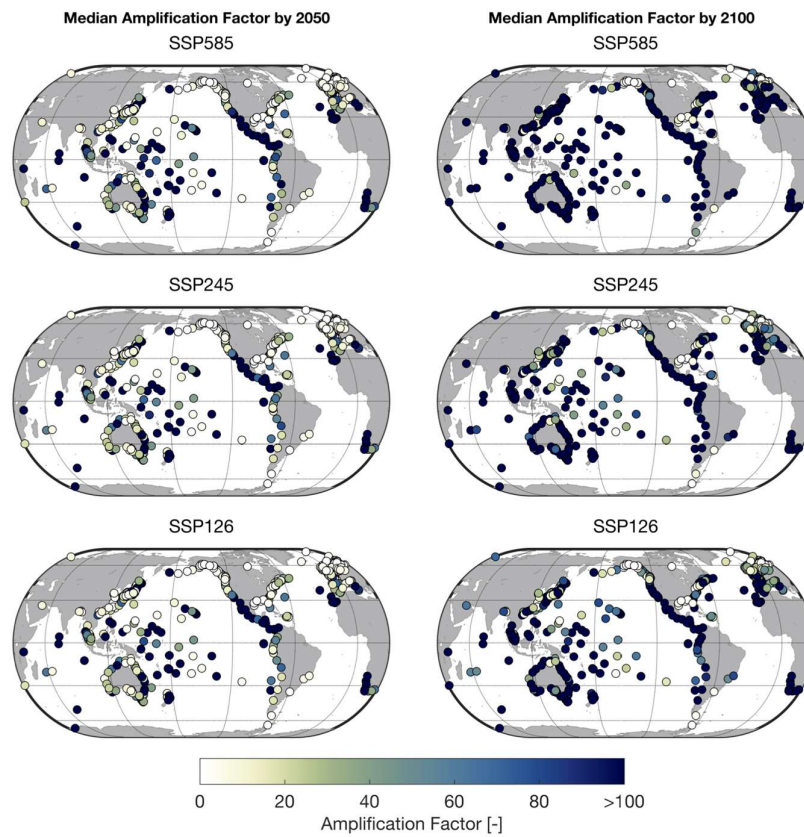


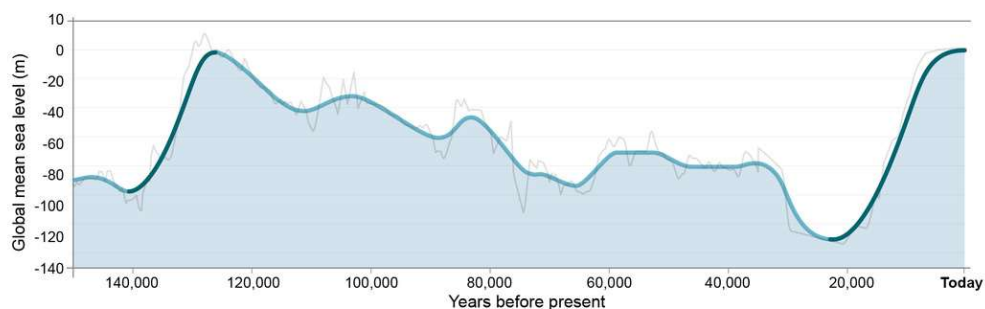
Figure 9.34: Projected frequency amplification factors for the 1% average annual probability coastal extreme water level in 2050 (left) and 2100 (right), based on a statistical analysis applied to historical extreme water level return period curves following SROCC, using the regional sea-level projections of this chapter (Section 9.6.3.3).

1

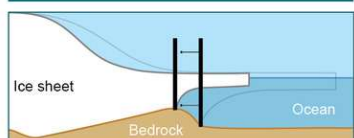
FAQ9.1: Will human-induced changes be reversible?

The past suggests that some consequences of climate change will continue even if temperature is stabilised.

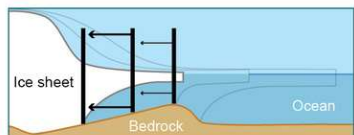
Sharp sea level rises are followed by **long gradual reductions**



Ice sheet retreat = sea level rise

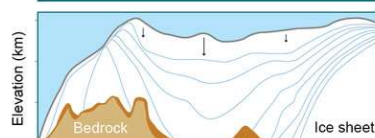


If bedrock is flat, the retreat stops when warming stops. As ice sheet retreats, **less ice** is released into ocean

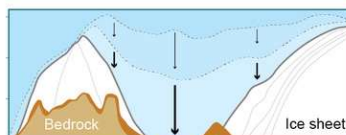


As bedrock dips landward the retreat is quick and self-sustained. As ice sheet retreats, **more ice is released into ocean** – ice sheets retreats further

Ice sheet retreat = sea level rise



The ice sheet is very thick therefore its surface is very high and the air at high altitude is very cold



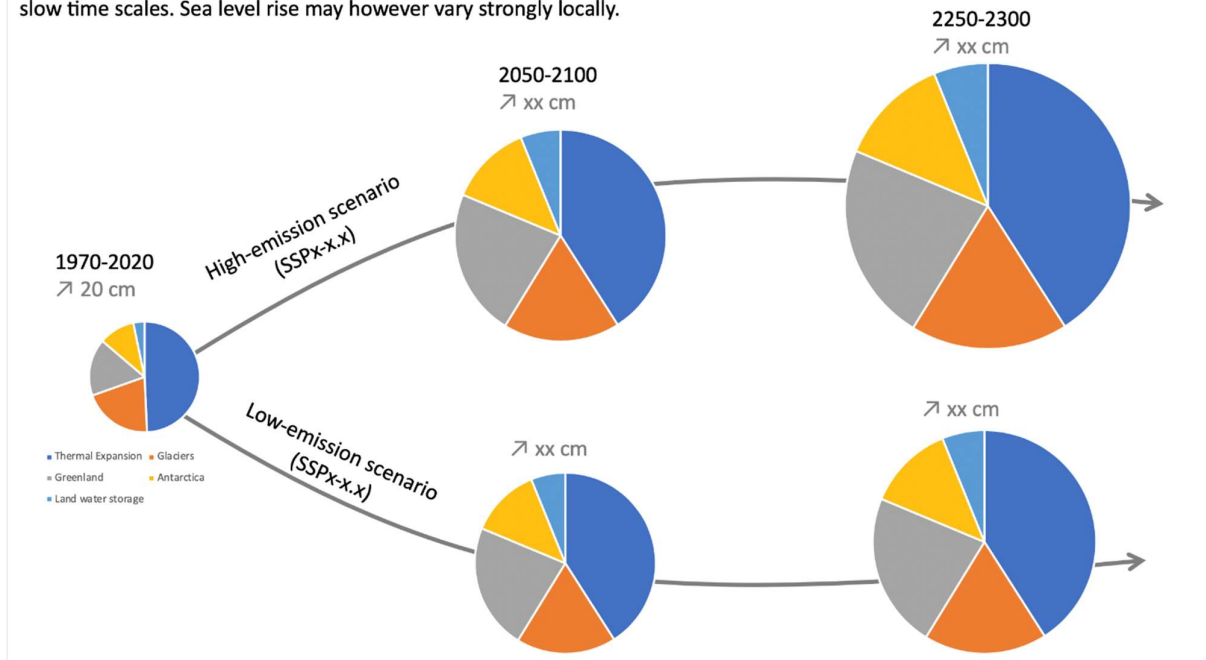
As the ice sheet melts, its **surface goes down** until it reaches a threshold, where the surrounding air is warmer and melts the ice even more quickly

FAQ 9.1, Figure 1: [PLACEHOLDER FOR FIGURE WITH ELEMENTS SIMILAR TO ABOVE.] Source: Nick Golledge

1

FAQ9.2: How much will sea level rise in the next few decades?

Global sea level will continue to rise in the next decades because its contributors (oceans, ice sheets, glaciers) respond on slow time scales. Sea level rise may however vary strongly locally.



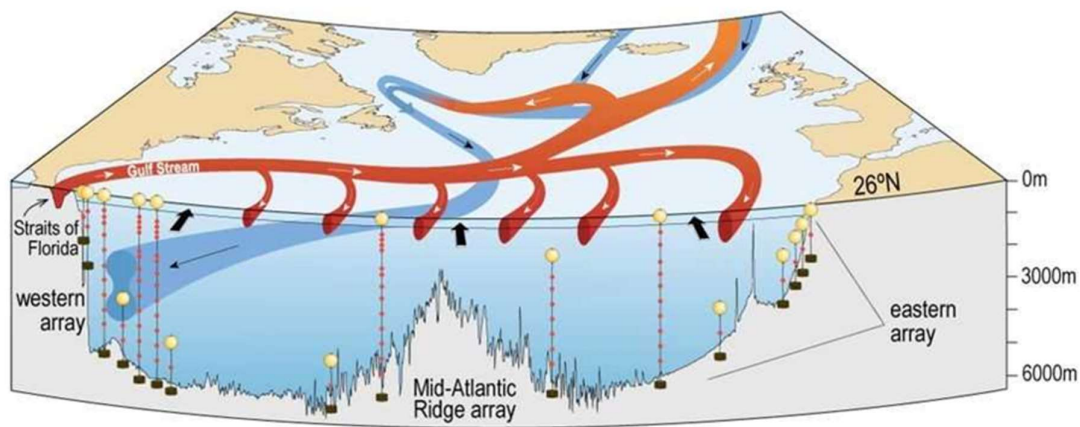
FAQ 9.2, Figure 1: [PLACEHOLDER FOR FIGURE WITH ELEMENTS SIMILAR TO SOME OF THE ABOVE].
Source: Sophie Berger & Nick Golledge & Aimée Slangen

2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31

1

Observing the Atlantic Circulation

RAPID-WATCH/MOCHA: A Purposefully Designed Array for Sustained Observations of the Meridional Overturning Circulation and Heat Flux



2

3 **FAQ 9.3, Figure 1:** [Placeholder for Gulf Stream Figure taken from RAPID/MOCHA website]

4