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A Review of Recent Updates of Sea-Level Projections at Global and Regional Scales

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Abstract Sea-level change (SLC) is a much-studied topic in the area of climate research, integrating a range of climate science disciplines, and is expected to impact coastal communities around the world. As a result, this field is rapidly moving, and the knowledge and understanding of processes contributing to SLC is increasing. Here, we discuss noteworthy recent developments in the projection of SLC contributions and in the global mean and regional sea-level projections. For the Greenland Ice Sheet contribution to SLC, earlier estimates have been confirmed in recent research, but part of the source of this contribution has shifted from dynamics to surface melting. New insights into dynamic discharge processes and the onset of marine ice sheet instability increase the projected range for the Antarctic contribution by the end of the century. The contribution from both ice sheets is projected to increase further in the coming centuries to millennia. Recent updates of the global glacier outline database and new global glacier models have led to slightly lower projections for the glacier contribution to SLC (7–17 cm by 2100), but still

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project the glaciers to be an important contribution. For global mean sea-level projections, the focus has shifted to better estimating the uncertainty distributions of the projection time series, which may not necessarily follow a normal distribution. Instead, recent studies use skewed distributions with longer tails to higher uncertainties. Regional projections have been used to study regional uncertainty distributions, and regional projections are increasingly being applied to specific regions, countries, and coastal areas.

Keywords Sea-level change · Regional sea-level change · Sea-level projections · Ice sheets · Glaciers · Terrestrial water storage · Mediterranean

1 Introduction

As one of the most well-known consequences of climate change, sea-level change (SLC) is relevant to coastal communities and stakeholders around the world. A large number of the world's population ($\sim$10 %, McGranahan et al. 2007) lives and works near the coast and depends on the ocean as their primary source of food and livelihood. An increase in mean sea level can increase the impacts of storm surges and the risk of flooding events in coastal zones (Wong et al. 2013). To make well-informed decisions about protective or adaptive measures, it is crucial that decision makers are provided with the best possible projections of SLC. Projecting future SLC and understanding the physical processes that contribute to SLC is therefore an important and rapidly evolving research topic.

SLC is a result of changes in many different parts of the climate system and can therefore be seen as an integrative measure of climate change. Over 90 % of the energy that is stored in the climate system ends up in the ocean (Rhein et al. 2013), causing thermal expansion and sea-level rise. In addition, ice sheets and glaciers lose mass due to increasing temperatures (both atmospheric and in the ocean, Vaughan et al. 2013) and reservoirs of water on land change due to human intervention (Church et al. 2013), which not only changes the amount of water in the oceans, but also the Earth's gravitational field. The solid Earth also responds to the redistribution of mass on the Earth surface both for present-day and for distant past (last glacial maximum, $\sim$20,000 years ago) mass variations, changing the height of the ocean floor. In the past century, global mean sea level has already increased by 19 $\pm$ 2 cm (1901–2010, Church et al. 2013), a rise that is expected to continue and accelerate in the coming centuries.

The Intergovernmental Panel on Climate Change Fifth Assessment Report (IPCC AR5) chapter on sea-level rise (Church et al. 2013) presented a comprehensive assessment of papers up to the IPCC working group 1 cut-off date of March 2013. In the chapter, important strides were made compared to the IPCC Fourth Assessment report (AR4) by progress in closing the twentieth century sea-level budget, the addition of an assessment of the ice sheet dynamical contribution to SLC and by making regional sea-level projections for the twenty-first century. However, a lot of research has been completed since IPCC AR5, and the lead authors of the chapter on SLC have recently published an update of their work (Clark et al. 2015). Here, we also focus on work that has been published since AR5 and aim to complement the review by Clark et al. (2015) by including more recent publications for the different contributions where available and by presenting overviews of research on the terrestrial water storage (TWS) contribution and the Mediterranean region, which were not discussed in Clark et al. (2015). The case of the Mediterranean is chosen because it is an area that is vulnerable to SLC due to the high population densities around the basin, and a lot of sea-level research is done specifically for this region.

First, we present an overview of recent work on contributions to SLC due to mass changes in glaciers and ice sheets (Sect. 2) and TWS changes (Sect. 3). Then, we will discuss global mean sea-level projections and new ways to treat the uncertainties thereof (Sect. 4). Recent advances in and uses of regional sea-level projections are presented in Sect. 5. Thermal expansion and dynamical ocean fields are not discussed in a separate section but are included in Sects. 4 and 5, as the most up-to-date projections are based on climate model output which has not changed since IPCC AR5. Finally, Sect. 6 presents research on sea-level projections in the Mediterranean region.

2 Land Ice Mass Change Projections

2.1 Ice Sheet Projections

The ice sheets on Greenland and Antarctica are by far the largest potential source of future SLC, storing approximately 65 m sea-level equivalent (SLE, Vaughan et al. 2013; Clark et al. 2015). Both ice sheets have increasingly lost mass in the past decades (Rignot et al. 2011) and are expected to dominate the sea-level budget on the long term (Church et al. 2013).

2.1.1 Greenland

Mass loss from Greenland is controlled by changes in surface mass balance (SMB) and dynamic discharge, including the effects of basal lubrication and ocean warming. IPCC AR5 (Church et al. 2013) estimated that Greenland would contribute between 0.04 and 0.10 m for the RCP2.6 (Representative Concentration Pathway, Moss et al. 2010) scenario and 0.07–0.21 m for the RCP8.5 scenario by the end of this century. For the lower emission scenarios, surface melting and dynamic discharge were expected to contribute equally to the overall mass loss. For RCP8.5, the mass balance was projected to be dominated by increased melting at the surface.

If a certain threshold is passed, the feedback between the lowering surface elevation and increasing surface melting can lead to additional ice loss and eventually even the complete loss of the Greenland Ice Sheet (Ridley et al. 2010; Robinson et al. 2012). Edwards et al. (2014) found that this positive feedback might be less significant for this century than previously expected. They estimate the surface elevation feedback to account for at most an additional 6.9 % ice loss from Greenland as opposed to the 15 % estimated in AR5. Other recent studies confirm the conclusion from AR5 that basal lubrication will likely not have a significant effect on Greenland mass loss within this century (Shannon et al. 2013).

Based on a higher-order ice sheet model driven by temperature changes from Atmosphere-Ocean General Circulation Models (AOGCMs) results, Fürst et al. (2015) project a Greenland contribution of 0.01–0.17 m to SLC within the twenty-first century in response to both atmospheric and oceanic warming. In contrast to previous studies, they conclude that future ice loss will be dominated by surface melting rather than dynamic discharge because the marine ice margin will retreat over time, reducing the contact area between ice and ocean water and thus limiting dynamic discharge. For the two lower emission scenarios, RCP2.6 and RCP4.5, simulations yield a contribution to SLC between 0.03 and

0.32 m by 2300 (Fig. 1). These new projections thus fall within the AR5 likely ranges but with a higher contribution from surface melting as opposed to dynamic discharge (Clark et al. 2015).

2.1.2 Antarctica

The mass balance of the Antarctic Ice Sheet is determined by changes in the SMB as well as changes in the ice flux across the grounding line resulting from enhanced basal sliding, calving or sub-shelf melting. Since surface melting will remain negligible within the twenty-first century (Vizcaino et al. 2010; Huybrechts et al. 2011), the SMB is predominantly determined by snow accumulation. Evidence from paleodata and projections from global and regional climate models show that snowfall in Antarctica is very likely to increase with future atmospheric warming (Frieler et al. 2015) by 5 ± 1 % per degree warming. The resulting mass gain might however be compensated or even overcompensated by dynamic effects (Winkelmann et al. 2012).

In AR5, the overall contribution of Antarctica to SLC was estimated to range from −0.03 to 0.14 m for RCP2.6 and −0.06 to 0.12 m for RCP8.5 by 2100 compared to 1986–2005 (Church et al. 2013). The SLC arising from rapid dynamics was projected to be −0.01 to 0.16 m, deduced from a combination of model results, expert judgement, and statistical extrapolation of current trends. Due to insufficient understanding of the underlying processes, scenario dependence for rapid dynamics could not be established in IPCC AR5.

Significant progress has been made since IPCC AR5 to understand the dynamic processes and quantify their effect on Antarctic ice loss for the twenty-first century and beyond. Pollard et al. (2015) found that crevasse-induced ice shelf loss can lead to the onset of rapid ice discharge from several Antarctic drainage basins. Based on the results from the SeaRISE model intercomparison project (Nowicki et al. 2013), Levermann et al. (2014) developed a probabilistic approach to estimate the future sea-level contribution from Antarctica, combining uncertainty in the climatic boundary conditions, the oceanic response and the ice sheet response. The results, based on linear response theory, correspond with the recently observed mass loss from the Antarctic Ice Sheet (Shepherd et al.

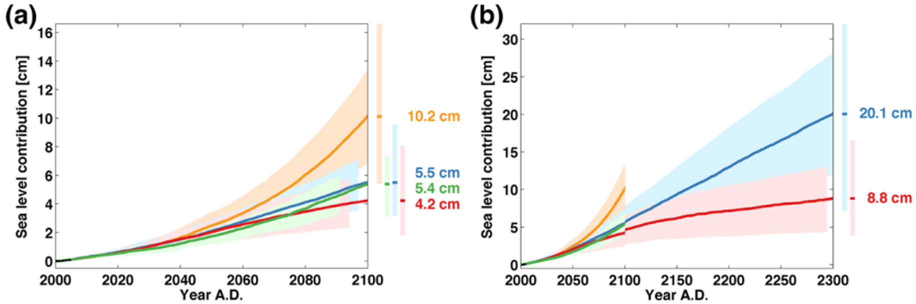

Fig. 1 Projected global mean sea-level contribution (cm) from the Greenland Ice Sheet (surface mass balance and dynamics) using a three-dimensional ice flow model driven by output from 10 atmosphere–ocean general circulation models **a** for four RCP climate scenarios over the twenty-first century and **b** for two RCP climate scenarios until 2300 [reproduced from Fürst et al. (2015)]. The shaded area indicates the ensemble mean ±1σ, while the vertical bars show the spread (all climate models) at the end of 2100 and 2300, respectively

2012). Levermann et al. (2014) find that the 90 % uncertainty associated with the contribution from Antarctica, mainly due to ocean warming, reaches up to 0.23 m (median 0.07 m; 90 % 0.0–0.23 m) by 2100 for RCP2.6 and up to 0.37 m (median 0.09 m; 90 % 0.01–0.37 m) for RCP8.5 (Fig. 2).

IPCC AR5 concluded that the collapse of marine ice sheet basins could cause additional SLC above the likely range of ‘up to several tenths of a metre’ (Church et al. 2013), but the timing could not be quantified. The mechanism underlying such a potential collapse is the marine ice sheet instability (MISI, Weertman 1974; Mercer 1978): several Antarctic basins are partly grounded below sea level, on bedrock generally sloping downwards towards the interior of the ice sheet. If the grounding line retreats into such an area, it could become unstable.

Shortly after the release of AR5, several studies were published showing with increasing certainty that parts of West Antarctica might in fact already be undergoing unstable grounding line retreat (Favier et al. 2014; Joughin et al. 2014; Rignot et al. 2014). The retreat was most likely caused by warm circumpolar deep water reaching the ice shelf cavities in recent years—whether this process was influenced by anthropogenic climate change is not yet clear.

Using a process-based statistical approach, Ritz et al. (2015) derived probability estimates for exceeding particular thresholds in the marine basins of Antarctica as a function of time if MISI is triggered. Their results suggest that particularly in the Amundsen Sea Sector, large and rapid ice loss due to the marine ice sheet instability could be initiated within this century. By 2100, the total ice loss from such rapid dynamics is estimated to contribute up to 0.3 m to global SLC, quantifying and narrowing down the IPCC AR5 uncertainties, and 0.72 m by 2200 (95 % quantiles). Large uncertainties remain, especially with respect to the effect of basal sliding on the ice flux (Ritz et al. 2015).

These advances made in estimating both the gradual response to oceanic warming and the possibly abrupt onset of self-sustained grounding line retreat can be consolidated into a new uncertainty range for Antarctic ice loss. It contains the IPCC likely range but leads to an overall larger spread for the twenty-first century sea-level projections.

However, a recent paper by Pollard and DeConto (2016) includes a number of processes in their model simulation that were not included in models before, such as the hydrofracturing of Antarctic ice shelves due to atmospheric warming and subsequent ice cliff instabilities. The model is found to be in relatively good agreement with geological

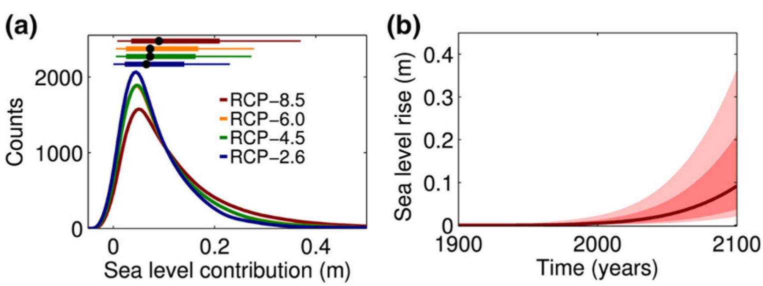

Fig. 2 Projected global mean sea-level contribution (m) from the Antarctic Ice Sheet in the twenty-first century. **a** Uncertainty range including climate, ocean, and ice dynamic uncertainty for the year 2100 (*top: thick line* is 66 % range, *thin line* is 90 % range). *Different colours* represent different climate scenarios used to drive three Antarctic Ice Sheet models. **b** Time series of future SLC from Antarctica (median 66 and 90 % uncertainty ranges) [reproduced from Levermann et al. (2014)]

estimates of the Pliocene ($\sim$3 million years ago) and the last interglacial (130,000–115,000 years ago). Depending on the geologic criteria used, they find possible contributions up to 1.05 $\pm$ 0.30 m (1σ) by 2100 under the RCP8.5 scenario. This means that the possibility of 1 m SLC from the Antarctic Ice Sheet by 2100 still cannot be excluded.

2.1.3 Long-Term Projections

Sea level will continue to rise well beyond 2100, even under strong mitigation scenarios (Church et al. 2013). Due to the long lifetime of anthropogenic CO_2 in the atmosphere and the consequent slow decline in temperatures, greenhouse gas emissions within this century can induce a sea-level commitment of several metres for the next millennia (Levermann et al. 2013). On these timescales, the Greenland Ice Sheet shows critical threshold behaviour with respect to atmospheric warming due to the surface elevation feedback (Ridley et al. 2010; Robinson et al. 2012).

Long-term projections from different process-based model simulations are now also available for the Antarctic Ice Sheet (Golledge et al. 2015; Winkelmann et al. 2015). Since several ice basins in Antarctica are potentially preconditioned to become subject to MISI, the response of the ice sheet to global warming might also be highly nonlinear. Golledge et al. (2015) find that the irreversible retreat of major Antarctic drainage basins can only be avoided if greenhouse gas emissions do not exceed the RCP2.6 level. Winkelmann et al. (2015) studied the evolution of Antarctica on millennial timescales and showed that the West Antarctic Ice Sheet becomes unstable after 600–800 GtC of additional carbon emissions. They further concluded that, on a multi-millennial timescale, Antarctica could become essentially ice-free for a scenario in which all available fossil carbon resources are combusted (10,000 GtC). These new studies suggest that the rate of SLC for higher emission scenarios could reach values of up to a few metres per century beyond 2100.

2.2 Glacier Projections

Glacier mass loss constituted a large contributor to twentieth century SLC (Gregory et al. 2013). Despite accelerating mass loss of the ice sheets (Shepherd et al. 2012), glacier mass loss continues to be a main component of SLC (Church et al. 2011) and is likely to remain an important factor in the twenty-first century. The AR5 evaluation of projected glacier mass loss in 2081–2100 relative to 1986–2005 ranges from 0.04 to 0.23 m, based on the results of four process-based models across different forcing scenarios (Church et al. 2013).

There are five glacier models operating on a global scale which have published projections of glacier mass change under the RCP scenarios (Marzeion et al. 2012; Hirabayashi et al. 2013; Radić et al. 2014; Slangen et al. 2014; Huss and Hock 2015) and one study which uses the Special Report on Emission Scenarios (SRES) to drive their glacier model (Giesen and Oerlemans 2013). They all combine a glacier surface mass balance model with a model that accounts for the response of glacier geometry to changes in glacier mass. The calculation of both the glacier mass balance and geometry change varies across the different models. All models except Huss and Hock (2015) were used in IPCC AR5, but some have been updated since, as will be detailed below.

Slangen et al. (2012, 2014) calculate the glacier mass balance from the sensitivity of the surface mass balance to temperature change and changes in precipitation. This sensitivity is parameterised by relations that are calibrated on more detailed model studies for 12 individual glaciers (Zuo and Oerlemans 1997). The initial areas of the glaciers are based on WGI-XF (World Glacier Inventory, extended format, Cogley 2009), and the glacier volumes are based on volume–area scaling. The glacier projections are forced by 14 models from the CMIP5 database (Taylor et al. 2012a) for each of the RCP4.5 and RCP8.5 scenarios.

Radić et al. (2014) and Marzeion et al. (2012) both use an approach in which accumulation and ablation are modelled explicitly. Accumulation is calculated by summing the solid precipitation over the glacier characterised by an area distribution over elevation. Ablation is calculated with a temperature-index method in both studies. Following Radić and Hock (2011), Radić et al. (2014) calculate the surface mass balance for each glacier at different elevation bands, whereas Marzeion et al. (2012) calculate melt from the temperature at the glacier tongue elevation only. Both studies use mass balance observations to calibrate the modelled glacier mass balance. In order to account for glacier retreat to higher elevations and thus allow for new equilibrium in a different climate, Radić et al. (2014) remove, or add in case of modelled mass gain, mass in the lowest elevation bins of the modelled glaciers, based on volume–area scaling. Marzeion et al. (2012) combine volume–length scaling with the mean slope of the glacier surface to let the glacier terminus retreat to higher elevations, or advance to lower elevations. They also include a response time between volume changes on the one hand, and length and area changes on the other. Their model is validated against in situ and geodetic mass balance observations of individual glaciers. Marzeion et al. (2012) do not model peripheral glaciers (PGs) in Antarctica explicitly, but apply the global mean specific mass balance rate as a rough approximation.

The results of Radić et al. (2014) shown in Fig. 3 are from projections that are forced by 14 models from the CMIP5 database for each of the RCP4.5 and RCP8.5 scenarios. The results of Marzeion et al. (2012) were updated based on a more recent version of the Randolph Glacier Inventory (RGI). Their projections were forced by 13 CMIP5 models for the RCP2.6 scenario, 15 models for RCP4.5, 11 models for RCP6.0, and 15 models for RCP8.5.

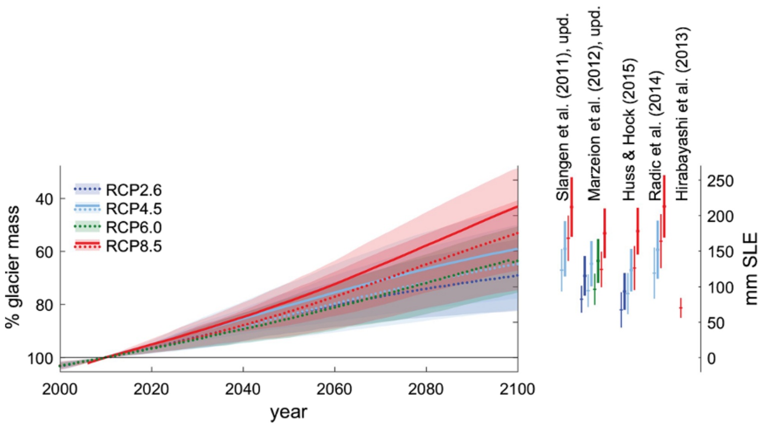

Fig. 3 Projected global mean sea-level contribution from glacier mass loss. *Left panel* percentage of glacier mass remaining (%), ensemble mean (*lines*) and 1σ spread (*shading*), *dashed lines* excluding, *full lines* including peripheral glaciers on Greenland and Antarctica. *Right panel* glacier contribution to SLC by 2100 (mm), ensemble mean and 1 ensemble standard deviation in 2100; *thick lines* including, *thin lines* excluding peripheral glaciers. All numbers are relative to 2010

Huss and Hock (2015) use a temperature-index model to calculate mass changes for every individual glacier, but their model approach is different from the earlier models discussed above. They do not use volume–area or volume–length scaling. Instead, they derive the initial glacier volume following Huss and Farinotti (2012). This method provides ice thickness, and therefore glacier volume and glacier bed elevation, distributed over 10-m elevation intervals for every glacier. Glacier geometry changes due to changes in calculated glacier mass are distributed over the glacier elevation following the parameterisation of Huss et al. (2010). Furthermore, Huss and Hock (2015) explicitly compute mass loss through calving using the simple model of Oerlemans and Nick (2005) that describes the calving rate as a linear function of the height of the calving front. Finally, they subtract the mass loss of glacier ice below sea level, which does not contribute to SLC, from the total of calculated glacier mass loss in their calculation of the glacier contribution to SLC (note that, for comparability, the loss of ice below sea level is also included in their numbers shown in Fig. 3). For calibration, Huss and Hock (2015) assume that mean specific balance rate of each individual glacier should equal the observed region-wide mean specific balance rate (Gardner et al. 2013) within a range of ± 0.1 m w.e.a^{-1} (m of water equivalent per year). The model is validated against in situ and geodetic mass balance observations, as well as observed area changes and calving rates, for individual glaciers. The results of Huss and Hock (2015) used here are from projections that are forced by 12 models from the CMIP5 database for the RCP2.6 scenario and 14 models for the RCP4.5 and RCP8.5 scenarios.

Hirabayashi et al. (2013) use a grid-based approach to modelling glacier mass change. Within each 0.5×0.5 degree grid cell, individual glaciers are lumped together as one glacier, while applying sub-gridscale elevation bands preserves the vertical elevation distribution of the ice area within each grid cell. Their model was calibrated against the observations of Dyurgerov and Meier (2005) and does not cover PGs on Greenland or Antarctica. The projections used here are forced by 10 models from the CMIP5 database for the RCP8.5 scenario only.

In each of the five global studies described above, the mass balance is calculated with a temperature-index model. Giesen and Oerlemans (2013) apply a more complex surface mass balance model that besides the dependence of glacier mass balance on temperature and precipitation also includes incoming solar radiation in the calculation of ablation. They calibrate this model to 89 glaciers with in situ observations of winter and summer mass balance and then upscale the results to all glaciers. Their projections for the twenty-first century are based on an ensemble of CMIP3 model runs for scenario A1B. They find a significant effect of projected decrease in incoming solar radiation in the Arctic region on the projected sea-level contribution. The twenty-first century global glacier mass loss found in Giesen and Oerlemans (2013) is significantly less than in other studies (Marzeion et al. 2012; Radić et al. 2014; Slangen et al. 2014) for comparable RCPs. In a regional study of future surface mass balance with the high-resolution regional climate model MAR, Lang et al. (2015) find significantly less mass loss for Svalbard than Marzeion et al. (2012) and Radić et al. (2014). Suggested explanations for this discrepancy are the coarse resolution of the global climate models that were used to force the global glacier models, and a better representation of the physical processes in the regional climate model compared to the empirical temperature-index mass balance models. Lang et al. (2015) also find a significant reduction in the incident solar radiation due to increased cloudiness, supporting the findings of Giesen and Oerlemans (2013) for the Arctic. Huss and Hock (2015) also find a 16–22 % lower projected glacier mass loss when they include incoming solar radiation (assumed to be constant in time) in a sensitivity experiment with their glacier mass balance model.

Figure 3 shows the projected glacier mass loss from the five global studies under RCP scenarios. They all show a large spread in the projected global glacier mass loss within the ensemble of different climate model runs for the same scenario. The ensemble standard deviation within each scenario is comparable to the differences between the ensemble means of different scenarios. Also, the differences between the different glacier models, but identical scenarios, are of comparable magnitude. The exception is the projection of Hirabayashi et al. (2013), which for the RCP8.5 scenario projects glacier mass loss comparable to the other models' projections for RCP2.6.

Updates of existing projections (Marzeion et al. 2012) and new models (Huss and Hock 2015) published after the IPCC AR5 have generally lead to slightly lower projected mass losses (Table 1). For the RCP8.5 scenario for instance, IPCC AR5 projected a contribution of 16 ± 7 cm, while Huss and Hock (2015) and the updated Marzeion et al. (2012) present projections around 12.5 cm for the same scenario. In the case of Marzeion et al. (2012), this is attributable to updates of the RGI; it is unclear for Huss and Hock (2015) since no previous estimate existed. On the other hand, the results of Slangen et al. (2012, 2014) are very similar to those of Radić et al. (2014). The results of Giesen and Oerlemans (2013) and Lang et al. (2015) suggest that a projected decrease in Arctic incoming solar radiation could lead to a lower projected mass loss than is given by the temperature-index models. However, a direct comparison of the individual studies is complicated through the differing compositions of the ensembles used for forcing the glacier models. Therefore, a coordinated glacier model intercomparison is currently underway to better understand the causes of the model and ensemble spread.

3 Terrestrial Water Storage Change Projections

Terrestrial water storage (TWS) change can result in a positive contribution to SLC due to a net transfer of water from long-term groundwater storage to the active hydrological cycle and eventually to ocean storage (Gornitz 1995; Taylor et al. 2012b). Other terrestrial components potentially influencing SLC include water impoundment behind dams (which can cause sea-level fall), drainage of endorheic lakes (mostly from the Aral Sea) and wetlands, deforestation, and changes in natural water storage (soil moisture, groundwater,

Table 1 Projected glacier contributions to SLC for 2010–2100 (mm, ensemble mean ± 1σ), for four different RCP scenarios, peripheral glaciers excluded (numbers in brackets include peripheral glaciers)

Study	RCP2.6	RCP4.5	RCP6.0	RCP8.5
Hirabayashi et al. (2013)	–	–	–	73 ± 14
Huss and Hock (2015)	67 ± 25 (93 ± 26)	90 ± 29 (123 ± 30)	–	126 ± 31 (178 ± 33)
Marzeion et al. (2012, updated)	82 ± 19 (115 ± 28)	94 ± 23 (132 ± 32)	96 ± 22 (136 ± 31)	124 ± 25 (175 ± 35)
Radić et al. (2014)	–	122 ± 36 (155 ± 41)	–	167 ± 38 (216 ± 44)
Slangen and van de Wal (2011, updated)	–	123 ± 30 (153 ± 39)	–	168 ± 32 (212 ± 42)

permafrost, and snow). Natural TWS change mostly varies with decadal climate variation with no significant trend.

Chao et al. (2008) found that the volume of water accumulated in dams up to 2010 was equivalent to a sea-level fall of $\sim$30 mm. However, Lettenmaier and Milly (2009) indicated that the volume of silt accumulated in dams should be removed from the estimate, which is equal to $\sim$4 mm less sea-level fall. Indeed, silting-up of existing dams may already be, or in coming decades may become, a larger effect on impoundment than construction of new dam capacity (Wisser et al. 2013).

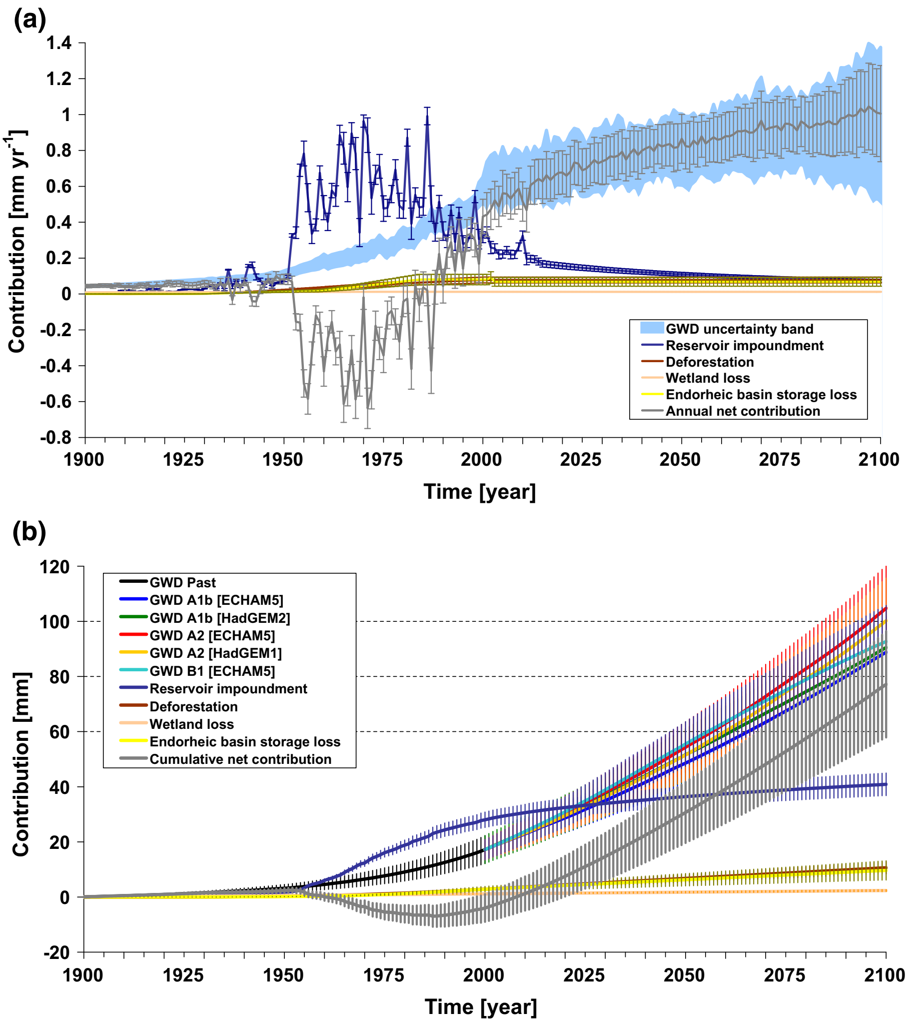

Fig. 4 Historical and projected terrestrial water contributions to SLC for a range of processes. **a** Yearly rates for 1900–2100 (mm year^{-1}) and **b** cumulative contribution to SLC since 1900 (mm). *Bars* indicate 1σ standard deviation. *Blue band* in (**a**) is based on 10,000 Monte Carlo realisations from five future projections of groundwater depletion, individual projections, and uncertainties shown in (**b**) [from Wada et al. (2012)]

Using a global hydrological model, Wada et al. (2012) estimated that the contribution of groundwater depletion (GWD) to SLC increased from 0.035 ± 0.009 mm year^{-1} in 1900 to 0.57 ± 0.09 mm year^{-1} in 2000 (Fig. 4). These figures were recently revised to lower values in Wada et al. (2016), who found a sea-level contribution of 0.12 ± 0.04 mm year^{-1} for the period 1993–2010 using a coupled climate–hydrological model. A volume-based study by Konikow (2011) also found slightly lower values than Wada et al. (2012) using direct groundwater observations, calibrated groundwater modelling, GRACE satellite data, and partly extrapolation for some regions. Also combining hydrological modelling with information from well observations and GRACE satellites, Döll et al. (2014) estimated the SLC contribution of GWD was 0.31 mm year^{-1} during 2000–2009. Another study (Pokhrel et al. 2012) used an integrated water resources assessment model to estimate all changes in TWS. However, their estimate is likely to overestimate the GWD contribution, because the model did not account for any physical constraints on the amount of groundwater pumping.

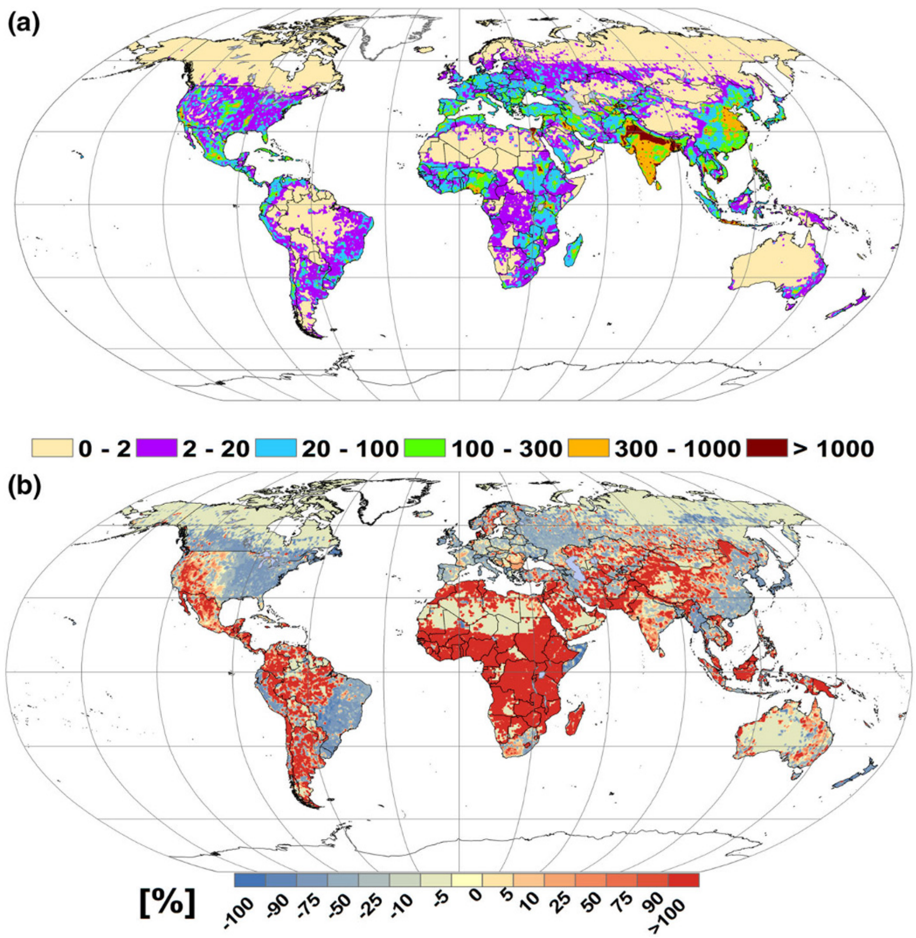

Fig. 5 **a** Projected global human water consumption in 2099 (million m^3 year^{-1}) and **b** the relative change (%) between 2010 and 2099 [from Wada and Bierkens (2014)]

Satellite observations have opened a path to monitor groundwater storage changes in data scarce regions (Famiglietti 2014). Since its launch in 2002, the GRACE satellite has been increasingly employed to quantify GWD at regional scales (Rodell et al. 2009; Famiglietti et al. 2011). GWD can be assessed after subtracting remaining TWS changes from GRACE-derived total TWS changes. However, coarse spatial resolution and noise contamination inherent in GRACE data hinder their global application in estimating GWD (Longuevergne et al. 2010).

Future projections of the GWD contribution to SLC are subject to large uncertainties due to the combination of climate projections from Atmosphere-Ocean General Circulation Models (AOGCMs) with future socio-economic and land-use scenarios that are inherently uncertain. The TWS contribution to SLC is projected to be 38.7 $\pm$ 12.9 mm, based on CMIP3 climate model output (Wada et al. 2012; Church et al. 2013). Since IPCC AR5, the groundwater model simulation has been updated, based on the latest CMIP5 climate and IPCC AR5 socio-economic data sets [see Fig. 5 for the latest projection of human water consumption from Wada and Bierkens (2014)], but does not provide the GWD contribution to SLC yet. The existing twenty-first century projections indicate increasing GWD caused by (1) increasing water demand due to population growth, and (2) an increased evaporation projected in irrigated areas due to changes in precipitation variability and higher temperatures. Groundwater depletion will be limited by decreasing surface water availability and groundwater recharge, which may cause groundwater resources to become exhausted at some time in the coming century (Gleeson et al. 2015).

4 Global Mean Sea-Level Projections

Before we discuss total global mean sea-level projections, we briefly discuss thermal expansion, as this is one of the most important contributors to global mean sea-level change. The majority of the net energy increase in the Earth's climate system is stored in the ocean, increasing the ocean heat content, which leads to warming and expansion of the ocean water. The resulting global mean thermosteric SLC by 2100 is projected to be 0.14 m ($\pm$0.04 m) for the RCP2.6 scenario, up to 0.27 m ($\pm$0.06 m) for the RCP8.5 scenario in IPCC AR5 (Church et al. 2013). New results are expected when the output of the sixth Climate Modelling Intercomparison project is released from 2017 onwards.

Although the focus in sea-level science is gradually moving towards regional SLC projections, as this is more relevant for coastal adaptation, there are still lessons to be learnt from the global mean SLC. The signal-to-noise ratio is smaller in the global mean, allowing a focus on long-term changes rather than local, short-term variability. As a result, it can be used to focus on narrowing uncertainties in the projections.

A notable development in global mean sea-level projections since IPCC AR5 is the use of a probabilistic approach to explore uncertainties in sea-level projections beyond the likely range (Jevrejeva et al. 2014; Kopp et al. 2014; Grinsted et al. 2015). In this approach, the projections (as presented in IPCC AR5) are blended with expert assessments of the Greenland and Antarctic Ice Sheet contributions (Bamber and Aspinall 2013) or expert assessments of total SLC (Horton et al. 2014). Expert assessments of, for instance, the potential contribution from ice sheets can be a useful tool to assess the uncertainty ranges, because the ice sheet experts know which particular physical processes (e.g., calving, ice sheet–ocean interaction) are insufficiently represented in their ice sheet models. One should keep in mind, however, that the current changes in the climate system are unprecedented, and estimates based on intuition, such as expert assessments, should therefore be used with care.

Figure 6 demonstrates the difference between the conventional and probabilistic approaches for global mean sea-level projections. Probabilistic projections allow the selection of specific probability levels to estimate low-probability/high-risk SLC projections, which by definition are unlikely to be reached, but cannot be ruled out given paleoclimate proxy information and the limitations in process-based modelling (Jevrejeva et al. 2014). They also allow for the use of probability distributions that do not follow a Gaussian distribution, such as skewed probability distributions with a longer tail to high SLC projections (Fig. 6).

In addition to the studies focusing on uncertainties in the global mean, a new application of the semi-empirical approach was published recently by Mengel et al. (2016). Semi-empirical models were developed after IPCC AR4 to offer an alternative to more complicated physical models of SLC. They are based on the assumption that sea level in the future will respond to imposed climate forcing as it has in the past, which may not hold if potentially nonlinear physical processes, such as marine ice sheet instability or thermal expansion, do not scale in the future as they have in the past. Mengel et al. (2016) calibrate the semi-empirical model for each contribution separately, such that the timescales of each contribution are considered in the calibration of the model. Their projected global mean SLC by 2100 is 84.5 cm (57.4–131.2 cm; median, 5th and 95th percentile) for the RCP8.5 scenario. This brings the semi-empirical models closer to the process-based IPCC AR5 estimates of 74 cm (52–98 cm) than other, larger, semi-empirical estimates at the time of IPCC AR5 (Church et al. 2013, Table 13.6).

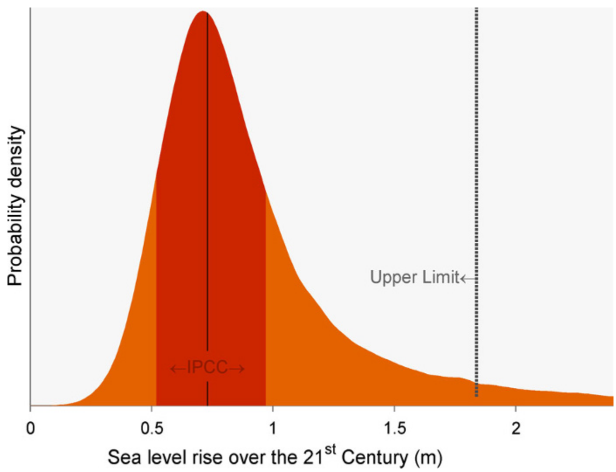

Fig. 6 Projected global mean sea-level rise by 2100 relative to 2000 for the RCP8.5 scenario and uncertainty (m). *Dark orange* represents the mean (*black line*) and likely range from IPCC AR5 (Church et al. 2013), *light orange* represents the probabilistic uncertainties from Jevrejeva et al. (2014). The vertical dotted black line represents the 95 % probability estimate of sea-level rise in 2100 (1.8 m) [from Jevrejeva et al. (2014)]

5 Regional Sea-Level Projections

Regional SLC can deviate substantially from the global mean due to a number of processes. Firstly, oceanic and atmospheric circulation changes and heat and salt redistribution in the ocean change the density of the water as well as redistribute mass within the oceans (Yin et al. 2010; Yin 2012). Secondly, any redistribution of mass between ocean and land, such as land ice mass change or TWS, affects the gravitational field of the Earth and causes viscoelastic deformation of the Earth's crust, the combination of which results in distinct sea-level patterns referred to as 'fingerprints' (Farrell and Clark 1976; Mitrovica et al. 2001). Thirdly, regional sea level can be influenced by vertical land motion, such as tectonic activity or glacial isostatic adjustment (GIA). GIA is the present-day viscous deformation of the Earth's crust as a result of ice melt after the last glacial maximum, which in turn also affects the gravitational field (Peltier 2004). GIA can have large local effects, while on a global mean scale the effect is negligible.

IPCC AR5 (Church et al. 2013) adopted the approach from Slangen et al. (2012, 2014) to compute regional sea-level projections by combining climate model results for thermal expansion and circulation changes with offline models to compute gravitational fingerprints as a result of mass change and GIA. Using this approach, both IPCC AR5 and Slangen et al. (2014) project regional sea-level values up to 20 % larger than the global mean in equatorial regions (Fig. 7), while close to regions of ice mass loss the values can be as small as 50 % of the global mean, mainly as a result of the gravitational effect. The meridional dipole in the Southern Ocean and the dipole in the North Atlantic are associated with the response of dynamic sea level (DSL) to increasing greenhouse gas forcing (Bilbao et al. 2015; Slangen et al. 2015), through wind stress and surface heat flux changes (Bouttes and Gregory 2014).

Carson et al. (2015) used the regional projections from Slangen et al. (2014) to study coastal SLC and found that coastal deviations from the global mean by 2100 can be up to 20 cm. The same regional sea-level projections were also used for a number of national assessments, such as Simpson et al. (2014) in Norway and Han et al. (2014, 2015) in Canada, where the global GIA model estimates were corrected or replaced by more accurate local GIA models or GPS measurements. Other regional assessments were done in Australia (CSIRO and Bureau of Meteorology 2015; Mcinnes et al. 2015) and the Netherlands (de

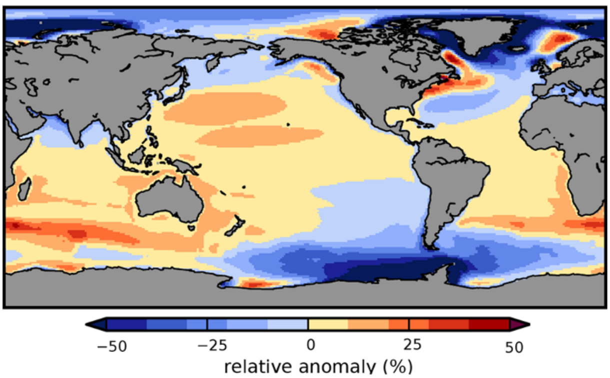

Fig. 7 Relative regional sea-level anomaly from the global mean change (between 1986–2005 and 2081–2100, %), based on the CMIP5-RCP4.5 scenario [from Slangen et al. (2014)]

Vries et al. 2014), which build on the IPCC-type regional sea-level projections. However, to really make a step forward in these national assessments, finer grid resolutions will be required to improve the model representation of ocean dynamic processes.

Using a probabilistic approach, Kopp et al. (2014) combined climate model information with an expert elicitation of the ice sheet contributions (Bamber and Aspinall 2013) to provide complete probability distributions of regional SLC projections. While the mean SLC is similar to IPCC AR5, Kopp et al. (2014) present high-end estimates which can be of particular interest and relevance for coastal management purposes. Following onto this, Little et al. (2015a) combined probability distributions with statistical models to estimate coastal flooding risk due to storm surges and SLC. They found that the risk of floods at the US East coast substantially increases as a result of SLC and changes in the frequency and intensity of tropical cyclones. However, these results were based on SLC from climate models only and do not include the SLC as a result of land-ice melt or TWS, which could lead to even larger flood risks.

To study the sources of uncertainty in sea level from climate models, Little et al. (2015b) decomposed the uncertainty into several components: model uncertainty, internal variability, scenario uncertainty, and a model–scenario interaction component. They found that, in the global mean, model uncertainty is the dominant term in the variance, whereas the variance due to scenario uncertainty increases in the twenty-first century and variance due to internal variability is initially large but decreases quickly. Locally, the contribution of each source of uncertainty can be very different, depending on the local magnitude of internal variability versus the response to external climate forcings. Both Hu and Deser (2013) and Bordbar et al. (2015) showed that internal variability in some locations can even be sufficiently large to be the main source of uncertainty all through the twenty-first century. As a result of the large internal variability, the time of emergence of SLC for DSL only (Lyu et al. 2014, their Fig. 2a) is beyond 2100 in over 80 % of the ocean area. The area with an emerging signal increases significantly (to almost 100 % by 2080) when thermal expansion, land ice, GIA, and GWD are included. For a further discussion of the literature on the effect of unforced variability on sea level and detection and attribution of SLC, see Han et al. and Marcos et al. in this issue, respectively.

The effect of freshwater input into the ocean as a result of land ice mass loss has been discussed in a number of studies, which have produced climate projections with integrated realistic estimates for glacier and ice sheet melt water run-off (Howard et al. 2014; Agarwal et al. 2015; Lenaerts et al. 2015). The first two studies focus on the impact of the freshwater forcing on DSL and find, using different models and different scenarios, that the impact is small (in the order of several cm) compared to the total SLC projected for the twenty-first century. However, both Howard et al. (2014) and Lenaerts et al. (2015) find that adding ice sheet freshwater forcings leads to a slight weakening of the Atlantic Meridional Overturning Circulation, indicating that it is important to include the freshwater forcing in climate models.

6 Mediterranean Sea-Level Projections

The Mediterranean is a semi-enclosed basin, linked to the open ocean through the Strait of Gibraltar. The high population density at the coast makes this basin particularly vulnerable to future SLC. Mediterranean sea level is influenced by various complex processes such as mass fluctuations (e.g., additional water input), variation in the density structure (steric effect), changes in circulation, waves, atmospheric pressure variations, and changes in the hydrographic conditions of incoming Atlantic water. These different components contribute to SLC at different timescales, from daily to interdecadal.

So far, global climate modelling attempts to assess future SLC in the Mediterranean did not deliver a consistent signal. Marcos and Tsimplis (2008) used projections from IPCC models to assess the interannual variation in steric sea level averaged for the Mediterranean, under the SRES A1B scenario, and found that global models did not agree on a trend. Indeed, their coarse resolution does not enable an accurate representation of important small-scale processes acting in the Mediterranean region, which are important to represent the water masses of the basin accurately. Additionally, AOGCMs have difficulties to simulate a reasonable water exchange at Gibraltar, which strongly influences the circulation and the changes in sea level in the Mediterranean Sea.

High-resolution regional climate modelling is thus needed to answer the question of ongoing Mediterranean SLC (Calafat et al. 2012). In addition to the thermosteric component, the contribution from changes in salinity has to be taken into account for the Mediterranean, since climate projections predict that the basin will become saltier in the future. Jordà and Gomis (2013) underlined that the saltening of the Mediterranean has two counteracting effects on sea level. Firstly, the halosteric effect leads to contraction of the water and thus a sea-level fall (-0.10 mm year^{-1} for 1960–2000). In contrast, the addition of salt to the basin in terms of mass leads to a sea-level rise ($+0.12$ mm year^{-1} for 1960–2000). As a simplification, these two contradicting effects can be neglected and Mediterranean mean SLC can be restricted to its thermosteric component.

Two recent studies have analysed Mediterranean SLC in future scenarios with regional models. Carillo et al. (2012) projected a thermosteric sea-level rise from 5 to 7 cm by 2050 (vs. 1951–2000) for the A1B scenario. With a six-member ensemble of scenario simulations, Adloff et al. (2015) found a larger sea-level rise of 10–20 cm in 2050 and 45–60 cm in 2099 (with respect to 1961–1990). In both studies, a large source of uncertainty is attributed to the hydrographic characteristics of the Atlantic boundary conditions prescribed in the Mediterranean model. Using the ensemble of Adloff et al. (2015), Fig. 8 shows the comparison of the spread of thermosteric sea-level response of the Mediterranean linked to (1) the choice of hydrographic conditions of Atlantic waters prescribed at the western boundary of the Mediterranean, and (2) the choice of the socio-economic

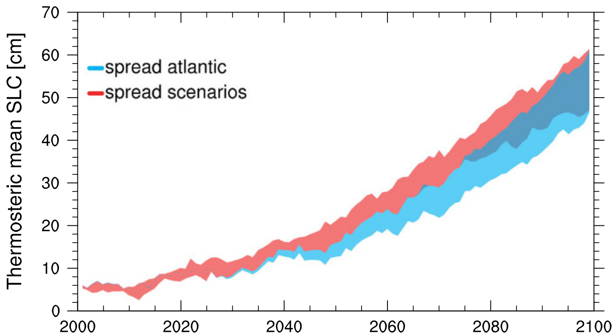

Fig. 8 Cumulative thermosteric sea-level change w.r.t. 1961–1990 (cm), averaged over the Mediterranean Sea from the six-member ensemble scenario simulations from Adloff et al. (2015). In *blue*, the uncertainties linked to the choice of the prescribed hydrographic conditions of Atlantic waters west of Gibraltar, and in *red*, the uncertainties linked to the choice of the socio-economic scenario

scenario. These results confirm how much the Mediterranean response in the future is driven by the Atlantic behaviour and raises the importance of the data set (mostly AOGCMs-derived) used to force the regional model at the boundary with the open ocean. Keeping in mind that the range of changes in near-Atlantic hydrography explored in the study by Adloff et al. (2015) is much smaller than the spread among CMIP models, it only gives a lower bound for the range of uncertainties in Mediterranean sea-level projections.

In comparison with the significant progress at the global scale, the advances at the Mediterranean scale remain small in terms of sea-level representation in regional ocean models. There is a significant lack of regional studies dealing with Mediterranean sea level, for hindcast periods as well as for projections, and none of them accounts for a proper Atlantic sea-level signal. The next step would be to include this missing feature and prescribe the complete sea-level signal at the Atlantic western boundary of Mediterranean regional models. This would allow the correct evolution of the Atlantic Ocean, which determines a part of the Mediterranean's behaviour, to be accounted for.

7 Synthesis

The field of sea-level research and all of its contributions is moving quickly, and a lot of work has been done since IPCC AR5. Here, we have reviewed the recent literature of projected sea-level contributions of ice sheets, glaciers, and terrestrial water storage to sea-level change. Furthermore, we discussed recent advances in global, regional, and Mediterranean sea-level projections. We did not discuss contributions that have seen little progress since IPCC AR5, most notably the thermal expansion and ocean dynamics components. However, these components are expected to be updated once the new model runs of the sixth phase of the Climate Model Intercomparison Project (CMIP6) become available.

The most recent sea-level projections for the Greenland Ice Sheet of 0.01–0.17 m by 2100 largely fall within the IPCC AR5 likely range for the twenty-first century. However, the contribution of surface melting is larger and the contribution of dynamic discharge is smaller than in IPCC AR5. Most projections for the Antarctic Ice Sheet since IPCC AR5 limit the sea-level contribution as a result of dynamic discharge and the potential onset of the marine ice sheet instability to 0.3 m by the end of this century. From the response to ocean warming, which is likely to dominate the dynamic Antarctic contribution, the 90 % uncertainty reaches up to 0.37 m by 2100 under the RCP8.5 scenario. However, Pollard and DeConto (2016) challenge this and project changes of well over 1 m by 2100 under the RCP8.5 scenario. All publications project that the bulk of SLC from Greenland and Antarctica will, however, occur after 2100 and might surpass several metres within the next centuries to millennia.

Glacier mass loss has been one of the main contributors to sea-level rise in the twentieth century and is expected to remain an important contributor in the next century. The latest findings, based on updates of glacier outlines used in existing projections and also new glacier models, project slightly lower contributions to sea-level rise from glaciers compared to IPCC AR5: from projections around $\sim$0.16 m in IPCC AR5 to the order of $\sim$0.12 m for the RCP8.5 scenario in more recent publications.

The sea-level contribution of changes in terrestrial water storage (TWS) has been difficult to estimate from observations in the past, but satellite observations now allow for better monitoring of changes in land water storage. Groundwater depletion is projected to increase due to growing water demand as a result of population growth and increasing evaporation. The projected contribution of TWS is 38.7 ± 12.9 mm for the period 2010–2100 (ensemble mean ± 1σ).

In projecting global mean SLC, the focus has turned towards providing better uncertainty estimates by using probabilistic methods and skewed uncertainty distributions. This may lead to better estimates of the low-probability/high-risk events in a changing climate. So far, these improved uncertainty distributions are based on expert elicitations, but as models evolve hopefully the uncertainty estimates will be based on modelling of physical processes in the near future.

Although significant advances have been made in recent years in projecting regional SLC, there are still a number of challenges that remain. The modelling and understanding of the ocean dynamical processes and incorporation of freshwater forcing as a result of ice sheet melt in climate models is an ongoing process. Ideally, the surface mass balance modelling of the ice sheets and glaciers would become part of the AOGCMs to obtain consistent results and include feedbacks between the ice sheets and glaciers with the rest of the climate system.

Ideally, sea-level change should be estimated on a national level, which is what coastal planners are interested in, but the spatial resolution of the current sea-level projections is still relatively coarse. To provide decision makers with better local estimates, models will need to use finer grid resolutions to account for local effects, such as coastal evolution and sediment transport. The increasing number of GPS measurements is also useful for local cases, as they give better estimates of vertical land motion, which can be large locally. In addition, new approaches now offer possibilities to link changes in flood risk and sea-level extremes to regional SLC.

Recent regional modelling studies in the Mediterranean have pointed out the relevance of the Atlantic signal, which largely contributes to the Mediterranean sea-level variability and represents one of the main sources of uncertainty in sea-level projections of the basin. Ongoing regional simulations are starting to tackle this issue and show that the prescription of sea-level information from the near-Atlantic at the lateral boundary significantly improves the Mediterranean sea-level representation at basin scale for hindcast periods. This will be added in future scenario simulations of the Mediterranean Sea.

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