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Uncertainties in Long-Term Twenty-First Century Process-Based Coastal Sea-Level Projections

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Abstract

Many processes affect sea level near the coast. In this paper, we discuss the major uncertainties in coastal sea-level projections from a process-based perspective, at different spatial and temporal scales, and provide an outlook on how these uncertainties may be reduced. Uncertainty in centennial global sea-level rise is dominated by the ice sheet contributions. Geographical variations in projected sea-level change arise mainly from dynamical patterns in the ocean response and other geophysical processes. Finally, the uncertainties in the short-duration extreme sea-level events are controlled by near coastal processes, storms and tides.

Keywords Uncertainties · Coastal sea level

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1 Introduction

Sea-level projections have a long tradition in climate research, and it has long been known that both the observed and projected rates of sea-level change vary regionally. As one moves from global to regional scales, additional processes have the potential to affect sea level, often increasing the uncertainty in future projections. In this paper, we consider global scale projections as our starting point, and then highlight additional uncertainty associated with processes influencing coastal and regional sea level. We define regional sea-level projections as those at a scale of approximately 100-km resolution—the same ocean horizontal scales as represented in the Coupled Model Intercomparison Project Phase 5 (CMIP5) experiments. CMIP5 data are also used to provide lateral boundary conditions for dynamical downscaling experiments that can provide sea-level projections at local coastal scales less than 10 km. Uncertainties in projections usually increase in time, as for most climate variables. In this paper, we are mainly concerned with a typical time scale of 100 years. The uncertainties with sea-level changes on geological time scales are not discussed. It is important to note that the uncertainty of the "open ocean" sea-level projections represented in CMIP5 climate models increases as we approach the coast as more processes need to be considered (Carson et al. 2019). While climate models explicitly represent sea-level rise associated with global thermal expansion and local changes in ocean circulation and density, the ice-related and land water storage components are calculated offline. This implies that the sea-level rise information is restricted to the climate model resolution and physical processes associated with climate models. Consequently, they are unable to represent important coastal processes, like erosion, sedimentation changes associated with changes in waves and tides, etc.

Figure 1 shows the key contributions to sea-level changes at the different operational scales used in this paper. The contribution of glaciers, ice sheets and land water storage changes are calculated using offline models forced with boundary conditions derived from climate model projections (mainly temperature and precipitation), implying that feedbacks between freshwater flux and ocean circulation are not captured. This furthermore implies that the uncertainty caused by the covariance of the different components is often not fully captured both at global and regional scales, hindering a proper uncertainty estimate (Le Bars 2018). One of the main reasons for this approach is the limited resolution of climate
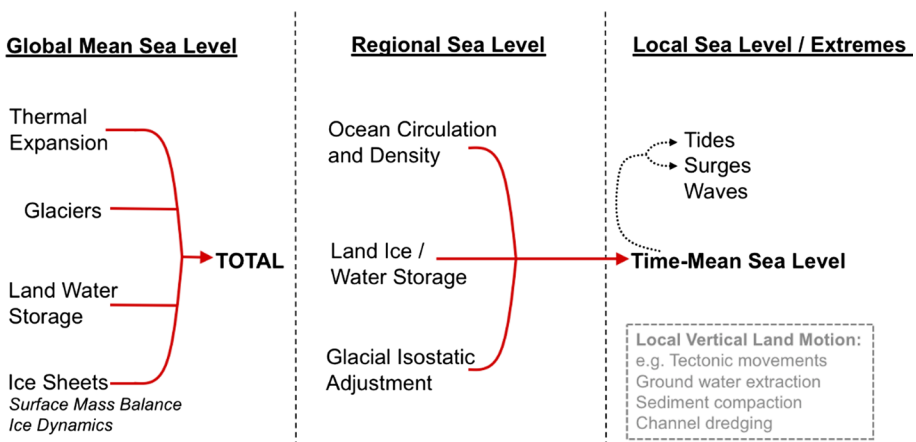

Fig. 1 Components of global mean sea level (GMSL), regional sea level (RSL) and local sea level (LSL)

models, which itself is a result of finite computing resources. For the calculation of changes in glacier volume and area, one would ideally use climate information at 1-km scale, which cannot be provided on 100-year time scales by global climate models (GCMs); hence, downscaling techniques are required. Similarly, high resolution (~ 10 km) is required to accurately calculate changes for the Greenland and Antarctic ice sheets in order to simulate the surface mass balance and ice dynamical processes. Further complications arise in Earth System Models for the inclusion of coupled ice sheet models, due to their long response time to environmental changes and our limited physical understanding of ice sheet dynamics (e.g., Schoof 2007; Pattyn 2018). Finally, land water storage changes are not resolved in climate models as they require non-RCP-driven[1] information on human behaviour and geological information to make projections of terrestrial water changes.

Figure 1 shows that, for regional sea-level changes, additional processes need to be considered. This includes ocean circulation and density changes that can be derived directly from the GCMs, as well as the regional imprint of land water storage and land ice changes. These two latter terms lead to a redistribution of mass between land and ocean and, consequently, the sea-level equation (Farrell and Clark 1976) needs to be resolved to calculate the pattern of the sea-level change (e.g., Mitrovica et al. 2001; Slangen et al. 2012, 2014; Perrette et al. 2013) to account for gravitational and rotational effects related to the mass redistribution. At local scales, one also needs to consider the contribution from vertical land motion induced by changes in loading of the crust due to ice mass changes. Together, these effects are referred to as "GRD" (gravitation, rotation, deformation) (Gregory et al. 2019).

IPCC AR5 (Church et al. 2013) presented projections for regional sea-level changes for the next century based on CMIP5 model simulations. For a given RCP scenario, these represent the change in the mean sea level over time due to climate change. However, it is important to note that local non-climatic processes, such as subsidence, can be much larger than the climate-driven component of relative sea-level change. In addition to the "*likely* range" projections presented in IPCC AR5, probabilistic approaches covering regional sea-level have been developed that aim to more fully represent the distribution of projected changes [e.g., Kopp et al. 2014; Jackson and Jevrejeva 2016; De Winter et al. (2017)]. Those results can be used by stakeholders who need more information than the *likely* range provides.

Figure 1 also indicates what is needed to resolve not only the sea-level trend, but also in addition the coastal high water levels associated with short-duration extreme events (e.g., surges and waves) on time scales ranging from hours to decades. For this reason, we discuss the uncertainties in global and regional projections as well as the uncertainties associated with shorter time scales particularly near the coast. We will focus on the extreme water-level variability rather than the morphodynamical processes that also play a role in the coastal zone and are particularly important in delta regions. Finally, we will briefly discuss the challenges of using sea-level rise projections in an adaptation framework (Stammer et al. 2019; Hinkel et al. 2019).

2 Sources of Uncertainties in Global Projections

Global mean sea-level (GMSL) rise occurs due to global thermal expansion (associated with ocean warming), and gain of ocean mass from glaciers, ice sheets and changes in the terrestrial water storage. Figure 2 shows a compilation of our understanding of the

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[1] RCP is Representative Concentration Pathway, i.e. the greenhouse gas concentration trajectory adopted by the IPCC for its fifth Assessment Report (AR5).

Fig. 2 A compilation of uncertainties in GMSL over the period 1900–2100 based on historical sea-level reconstructions, see text and projections presented by Church et al. (2013). The shading indicates the likely range around the mean reconstruction or projection
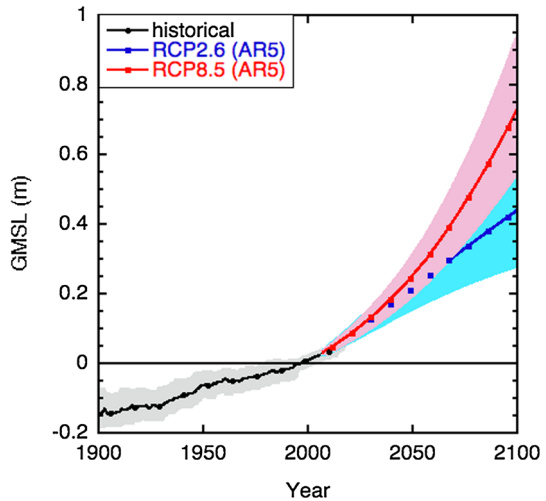

uncertainties of global sea-level reconstructions over the last century and projections over the twenty-first century. The reconstruction over the last century is based on equally weighting the reconstructions by Ray and Douglas (2011), Church and White (2011), Jevrejeva et al. (2014), Hay et al. (2015) and Dangendorf et al. (2017). The uncertainties for the projections are based on the assessed "*likely* range" (17–83‰) presented in IPCC AR5 (Church et al. 2013). Uncertainties decrease from the past to present and increase again further ahead in time, related to the chosen reference period in the present (e.g., 1986–2005). Typically, uncertainties in projections for the second half of the twenty-first century are larger than the uncertainty over the historical period. The uncertainty in global mean sea-level projections strongly depends on the emission scenario as shown in Fig. 2. The uncertainty does not include the potential collapse of the marine-based sectors of the Antarctic ice sheet (Church et al. 2013). Hence, uncertainties are possibly larger than shown.

The components that explain the sea-level change over the last century (Gregory et al. 2012), which was dominated by global thermal expansion and glacier mass loss, will continue to contribute to the sea-level change for the next century, but not in the same ratios. This is already emerging from the observations of current rates of sea-level change which shows an acceleration of sea-level rise (Cazenave et al. 2018), that, for a large part, can be attributed to an increase in the mass loss of the ice sheets (Shepherd et al. 2018; Mouginot et al. 2019; Rignot et al. 2019). The projections presented by Church et al. (2013) provide an indication of the contribution of each GMSL term to the total variance in projected sea-level change over the twenty-first century (Fig. 3).

As shown in Fig. 2, the projections do not account for the potential collapse of the marine-based sectors of Antarctica (Church et al. 2013). Twenty-first century sea-level rise substantially larger than 1 m is only thought to be possible with a significant contribution from the major ice sheets of Greenland and Antarctica. As such, there is a wide consensus that for centennial projections of sea-level change, the major source of uncertainty is related to ice sheet processes, since the largest uncertainty in any other component is only at the 10 cm level. The large uncertainty in estimates of mass loss in Antarctica is due to our limited physical understanding of the dynamic response of the ice sheet (Pattyn 2018).

Fig. 3 Evolution of the relative contribution of thermal expansion (TE), glaciers (GL), the Greenland ice sheet (GIS), Antarctica (Ant) and land water storage changes (LWS) as presented by Church et al. (2013) to the total variance. The dotted blue line indicates qualitatively the increase in the dynamic contribution of the Antarctic ice sheet if marine-based sectors of Antarctica collapse
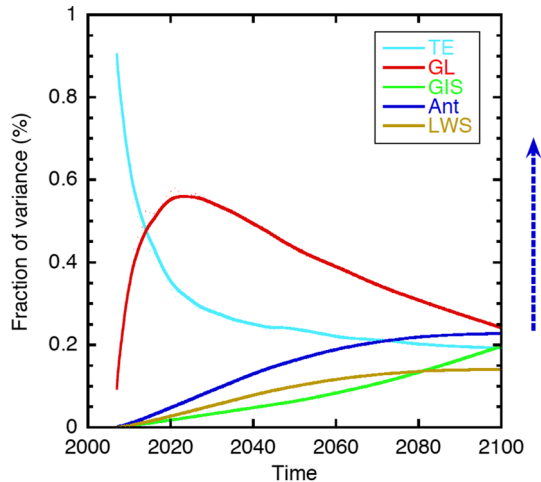
For Greenland, sea-level projections are driven largely by surface mass balance changes, which have a smaller uncertainty. The uncertainty in the ice dynamical component is related to the limited observational evidence and our ability to understand and simulate key ice dynamic processes (Edwards et al. 2019). Improved and prolonged satellite-based techniques are rapidly changing the field (e.g., Shepherd et al. 2018).

Different mechanisms can lead to mass changes of the Antarctic ice sheet. Traditionally, emphasis has been on the possibility of changes in accumulation, whereby, in a warmer climate, accumulation rates increase because warmer air can contain more moisture, promoting a gain of ice sheet mass and a decrease in the sea-level contribution. Currently, emphasis is more on interplay of the dynamical processes and the environmental changes whereby the environmental changes are no longer restricted to atmospheric processes but apply also to the ocean changes surrounding the ice sheet. Currently, the ice sheet is losing mass (Joughin et al. 2014; Rignot et al. 2014) in restricted areas in west Antarctica, and the general understanding is that these changes are driven by changes in the ocean water temperature which controls the basal melt rates below the ice shelves. Whether these changed water temperatures are due to ocean variability or are driven by climate change remains to be seen (Jenkins et al. 2018). As a result, it is at present difficult to assess, which term in the mass balance is driving future changes of the ice sheet.

Though accumulation changes are likely to be too small to explain the observed mass loss, the dynamic implications of changes in Antarctic surface mass balance may be important. Increased ablation rates at the surrounding shelves of Antarctica have locally increased such that the firn layer of the snow is saturated and the remaining water penetrates the deeper parts of the ice shelf and destabilizes the shelves. This process of hydrofracturing was a contributor to the rapid break up of Larsen-B (Rott et al. 1996). Disappearance of the ice shelves does not lead directly to a sea-level contribution, but the shelves are crucial for the stability of the grounded ice. In the study by DeConto and Pollard (2016), the hydrofracturing mechanism combined with marine ice cliff instability plays an important role in the rapid deglaciation of Antarctica. However, results from a carefully calibrated regional climate model (Kuipers Munneke et al. 2014; Trusel et al. 2015) do suggest that the amounts of melt are too small on most ice shelves until the end of the century to cause large-scale hydrofracturing on the major ice shelves.

The point at which increased basal melt will trigger destabilization of ice shelves is largely unknown because of limitations in our knowledge of the topography of the ice cavities below the ice shelves and our understanding of future changes in ocean circulation on the continental shelves surrounding the ice shelves. Basal melt models including the local circulations in ice cavities are now emerging (Lazeroms et al. 2018; Reese et al. 2017). While they capture the changes in basal melt rates under changed geometrical conditions, the feedback from increased fresh water production on the ocean circulation itself is not explicitly included in ocean models. In addition, models describing the physics of calving of floating ice shelves at the open ocean side are only starting to be developed (e.g., Borstad et al. 2012), and the first direct measurements of basal melt rates are just emerging (Sutherland et al. 2019).

On top of the uncertainties in the forcing of the ice shelves, we have to consider the uncertainty in the dynamical response to the changes in the forcing. What matters for the ice sheet from a dynamical point of view is the loss of buttressing. The lower the buttressing, the faster the ice sheet will lose mass, and hence sea-level rise will occur. Traditionally, it was thought that, in particular, on retrograde slopes, an initial retreat of the grounding line could lead to an irreversible retreat (Marine Ice Sheet Instability, MISI). Recently, it has been suggested that even for prograding slopes, the ice sheet may experience a rapid retreat via marine ice cliff instability (MICI) (Bassis and Walker 2012). Currently, MISI is captured by state-of-the-art ice sheet models if the resolution is sufficiently detailed and the subglacial topography known, but MICI is only treated in a highly parameterized way (DeConto and Pollard 2016). In particular, the wastage term determines how much ice is lost to the ocean by ice cliff instability. A possible precursor to both instability mechanisms of ice mass loss is a destabilization of the ice shelves. Retreat or weakening of ice shelves can therefore act as a tipping point for ice sheet mass loss. Quantification is challenging as it depends both on the projections of increased atmospheric warming driving hydrofracturing as well as changes in ocean water circulation and temperature changes on the continental shelves driving the basal melting of the shelves.

3 Sources of Uncertainties in Regional Sea-Level Projections

3.1 Ocean Processes

Sea-level changes due to changes in ocean density and circulation with inverse barometer correction of sea-level pressure are here referred to as sterodynamic sea-level changes (Gregory et al. 2019), while sterodynamic sea-level change minus global mean thermosteric sea-level change is referred to as dynamic sea-level change (e.g., Yin 2012). Over the twenty-first century, regional dynamic sea-level changes can be as large as the global mean thermosteric sea-level change (Yin 2012), thereby doubling the respective regional ocean-driven sea-level change in some areas. Unlike the contributions from land ice and land water, sterodynamic and dynamic sea-level changes are directly simulated by GCMs, and uncertainties are commonly assessed as the spread over the available ensemble.

By the end of the twenty-first century, uncertainties associated with ocean processes are the largest in regions of high ensemble mean sterodynamic sea-level change such as the Arctic Ocean and the northern North Atlantic (Slangen et al. 2014; Little et al. 2015), particularly along the densely populated north-eastern North American coast (Carson et al. 2016), as shown in Fig. 4.
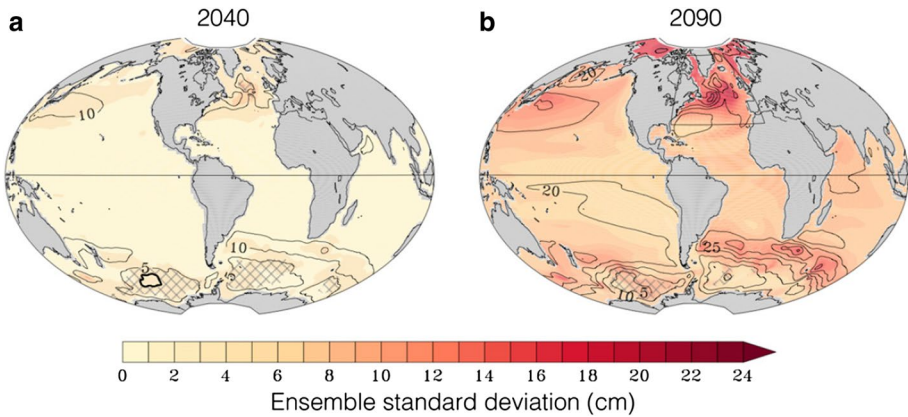
Fig. 4 CMIP5 sterodynamic component of sea-level change for (left) 2040 and (right) 2090 as multimodel, multiscenario, multirealization decadal mean (contours, in cm) and the ensemble standard deviation (colours). Cross-hatching indicates regions in which the ensemble mean divided by the ensemble standard deviation is less than one, figure from Little et al. (2015)

In the Arctic Ocean, this is caused by the sea ice—ocean interaction, whereas for the North American coast, deviations are driven by changes in the Atlantic Meridional Overturning Circulation (AMOC). The AMOC is projected to decline over the twenty-first century (Collins et al. 2013), resulting in a dynamic sea-level rise along the north-eastern American coast. Widely varying representation of the AMOC across models leads to a large uncertainty in this region (e.g., Yin et al. 2009; Little et al. 2019). This model or structural uncertainty was found to be the largest contributor to regional sea-level uncertainty by ocean processes throughout the twenty-first century, to varying degrees in different regions. Uncertainty due to internal variability and scenario uncertainty contributed significantly in the beginning and, towards the end of the century, respectively (Little et al. 2015).

Heat fluxes across the atmosphere–ocean interface cause changes in global and regional sterodynamic sea level. The simulated global mean thermosteric sea-level rise, and therefore also regional sterodynamic sea-level changes, depend on the climate feedback parameter and the ocean heat uptake efficiency, both of which vary across models due to different parameterizations and representations of oceanic and atmospheric processes (Melet and Meyssignac 2015 and references therein). They form an important source of variability in regional sea-level projections at present and will continue to do so in the future (Yin et al. 2010; Yin 2012; Church et al. 2013; Slangen et al. 2014; Gregory et al. 2016).

Where and to what depth the excess heat enters the ocean directly affects thermosteric sea-level change, one major component of sterodynamic sea-level change, as the thermal expansion coefficient depends on pressure, temperature and salinity. This effect contributes to locally high rates of thermosteric sea-level rise in regions where the ocean heat uptake is relatively large in the current climate, such as those associated with mode water formation (Suzuki and Ishii 2011; Terada and Minobe 2018). At the same time, local halosteric effects depend on the salinity versus freshwater input which is particularly important at high latitudes in the North Atlantic Ocean (Yin et al. 2010). Additionally, the projected warming that takes place in the deep ocean below the depth of shallow continental shelves drives an ocean mass redistribution that contributes strongly to

sea-level rise in coastal areas (Landerer et al. 2007; Yin et al. 2010; Richter et al. 2013). However, the consistency of the climate models with respect to shelf mass loading has not yet been evaluated, and is likely hindered by the limited resolution of GCMs.

The sea-level changes originating in the deep ocean affect coastal sea-level via westward propagating Rossby waves that subsequently travel equatorward along the western boundaries as coastally trapped waves (Minobe et al. 2017). However, interaction with currents and topography may affect this mechanism (Little et al. 2019). Additionally, as a consequence of the limited resolution (typically 100 km in CMIP5 climate models), the current generation of GCMs fails to capture and simulate future changes in strong and narrow western boundary currents such as the Gulf Stream and the Kuroshio, both of which are important for local sea-level changes. This can be addressed by dynamical downscaling using high-resolution ocean models (Liu et al. 2016; Zhang et al. 2017). For example, Liu et al. (2016) reported that over the twenty-first century, the sea-level rise difference is less than 10 cm along the coast of the main islands of Japan but can reach nearly 50 cm in the deep ocean east of Japan.

In contrast to atmosphere–ocean heat fluxes, momentum fluxes do not induce changes in global sea level. However, changes in the magnitude and location of major wind systems greatly affect regional sea level via the wind-driven ocean circulation. In particular, the poleward shift of mid-latitude westerly winds causes poleward expansion of the subtropical oceanic gyres, particularly in the Southern hemisphere (Saenko et al. 2005; Bouttes et al. 2012; Zhang et al. 2017). This poleward gyre-boundary migration is accompanied by sea-level rise around the gyre boundary as subtropical gyres lead to higher sea level than subpolar gyres. This mechanism partly contributes to the sea-level rise in the Kuroshio Extension region with a large inter-model spread (Zhang et al. 2014; Terada and Minobe 2018).

Climate models are not yet routinely coupled to glacier and ice sheet models; thus, the impact of freshwater fluxes from melting land ice on the ocean circulation is not simulated and represents an additional uncertainty, particularly in the northern North Atlantic and Arctic Ocean. This region is particularly sensitive to the input of meltwater from the surrounding ice-covered land masses in north-eastern Canada and Greenland that is associated with the strength of the AMOC (e.g., Stammer 2008; Yin et al. 2009). A study by Slangen and Lenaerts (2016) with a fully coupled version of the Community Earth System Model (CESM) shows that freshwater forcing from the ice sheets leads to an ocean-driven change of locally over 0.1 m over the twenty-first century in the Southern Ocean, the Arctic Ocean and the northern North Atlantic including the east coast of North America. Similarly, Bronselaer et al. (2018) used a CMIP5 model to study effects of meltwater from the ice sheets and ice shelves of Antarctica, and they found that meltwater can induce subsurface ocean warming, leading to a positive feedback in further ice mass loss. Golledge et al. (2019) argue that meltwater from Antarctica will be trapped near the sea surface and increases ice mass loss.

On shorter (decadal and interannual) time scales, sea-level changes are largely associated with natural variability and climate modes such as Pacific decadal oscillation (PDO)/ interdecadal Pacific oscillation (IPO), Atlantic multidecadal oscillation and El Nino and Southern oscillation. How these and other climate phenomena influenced sea-level are documented by Woodworth et al. (2019) and Han et al. (2018, 2019). In particular, the rapid sea-level rise observed in the western tropical Pacific during the satellite era is related to the PDO/IPO variability (Zhang and Church 2012; Hamlington et al. 2013). These modes are not predictable on time scales longer than a few years (Kim et al. 2012; Chikamoto et al. 2015; Meehl et al. 2016), and are therefore ignored in projections.

To account for natural variability in sea-level projections, it is important that climate models (1) represent the observed variability adequately and (2) capture potential future changes in frequency and magnitude of natural climate modes. Regarding (1), it has been suggested that the decadal variability in ocean models is too low in the tropical Pacific Ocean (Bilbao et al. 2015) over the altimetric record. Peyser and Yin (2017) showed that CMIP5 models underestimate sea-level variability associated with El Nino and the PDO. With respect to (2), some features of climate modes are expected to change in future (Hu and Bates 2018). Cai et al. (2014) reported that the frequency of extreme El Nino will increase, implying that extreme sea-level anomalies associated with El Nino may occur more frequently (Widlansky et al. 2015).

On somewhat longer time scales, Hu and Deser (2013) found that the decadal variability in sea level did not change between 2041–2060 and 1980–1999 based on analyses of 40 ensemble members of a single climate model, and Little et al. (2015) reported that decadal sea-level variability around New York City is expected to be roughly constant throughout the twenty-first century.

3.2 Other Geophysical Processes

Regional sea-level changes as experienced in coastal zones are not only the result of ocean processes, but also of changes in the solid earth topography and gravity field (Tamisiea and Mitrovica 2011), which are the subject of this section.

As the solid earth is not rigid, the height of the surface determining the relative sea level is a result of a balance between Earth dynamic processes, such as mantle convection and plate tectonics and buoyancy forces (Conrad 2013). When mass accumulates at a specific location, the additional weight depresses the surface and raises the geoid, causing sea level to rise in its proximity and to fall in the far field. This solid earth–water interaction is described by the sea-level equation (Farrell and Clark 1976; Kendall et al. 2005) and is mainly instantaneous. The major uncertainty in the pattern is the uncertainty in the loading changes by the ice.

The main geophysical process that contributes to regional sea-level change on century time scales is Glacial Isostatic Adjustment (GIA), which is the ongoing response of the solid earth to the loading–unloading effect of the last glacial cycle (Tamisiea 2011). Uncertainties from forward GIA models are usually not available, since they are mainly due to uncertainties in earth structure and glacial history that are very difficult to assess. The only systematic study of GIA uncertainties has been recently published by Melini and Spada (2019), where they have shown that both input uncertainties (ice history and Earth mechanical properties) and structural uncertainties (differences in GIA models and in observational constraints) are equally important contributors to the total error budget. At many locations, the solid Earth response is small, and signals at the cm/year level are only expected close to the former ice sheets (e.g., Hudson Bay Baltic Sea). Other regions where the GIA signals are important are those in the peripheral areas of the former ice sheets, like the US East coast, where vertical land motion rates are about 1–2 mm/year. As the solid Earth processes are slow, we can assume the GIA signal to be linear at the centennial scale; hence, if we assume an uncertainty of about 0.5 mm/year (Frederikse et al. 2017), it will contribute no more than 5 cm to the uncertainty near the end of the twenty-first century.

A second process affecting the uncertainty in regional and coastal sea-level changes is the accumulation of sediments at the estuaries of major rivers, which induces both bathymetry changes and gravitational effects. The accumulation of large amounts of sediments

loads the crust, which causes subsidence, and attracts ocean waters; both processes result in regional sea-level rise. A few studies have considered the Mississippi (Ivins et al. 2007) and the Indus (Ferrier et al. 2015) basins, where relative sea-level rates can locally amount to a few mm/year due to sediment accumulation.

Changes in the amount of freshwater stored on land will also affect relative sea level in both a direct way (water exchange between continents and oceans), as well as through induced changes in the solid earth. Examples are the effects of natural processes such as droughts (Borsa et al. 2014), intensified periods of rainfall on land (Boening et al. 2012), but also anthropogenic changes like dam building (Fiedler and Conrad 2010) and groundwater extraction (Veit and Conrad 2016). With respect to sea-level changes at centennial scales, the main uncertainty will likely arise from groundwater extraction, which depends strongly on socio-economic factors. Regional sea-level changes could amount to several millimetres per year, especially in California, Pakistan and NW India. In addition, local groundwater extraction is causing some megacities to subside by several centimetres per year (e.g., Abidin et al. 2015); whether this trend will continue in the future is dependent on political and socio-economic factors (Nicholls 2011).

Finally, the occurrence of earthquakes can affect local sea level due to both instantaneous elastic deformation of the crust, mostly affecting sea level very close to the earthquake location, and the subsequent viscoelastic adjustment in the earth curst and upper mantle, which might affect regional changes (Broerse et al. 2014), with rates largely dependent on the earthquake size (Melini et al. 2004). However, it is almost impossible to make projections of the effect of earthquakes, since recurrence times of large events are usually several centuries. Of all the processes listed above, only GIA and the effect of freshwater changes on land are usually incorporated into regional sea-level projections.

4 Sources of Uncertainties in Extreme Sea Level

Extreme sea level (ESL) is defined here as an event of extremely high sea level along the coast occurring over a relatively short period of one to a few days (sometimes also called "extreme water level"), triggered by storm surges, waves, tides or a combination of these processes. ESL events take place on top of the long-term trends in sea level and play an important role in flood risk. Flood risks may increase due to the long-term trend in mean sea level, but also by a change in the frequency or magnitude of extreme events (often expressed in terms of return periods). Changes in historical flood risk have been dominated by long-term sea-level rise, but whether that also holds for changes in future ESLs is uncertain.

Beside changes in climate modes as discussed in Sect. 3, ESL characteristics might also be affected by changes in frequency or magnitude of cyclones, waves and tides. Most AOGCM models show that the frequency of cyclones will either decrease or remain unchanged, but the intensity is expected to increase (Christensen et al. 2013). How this affects the extreme sea-level characteristics for those coastlines vulnerable to cyclone activity remains to be seen.

Close to the coast, many processes may influence ESL characteristics. Arns et al. (2017) showed that an increase in mean sea level reduces the depth limitation of waves in shallow seas yielding larger waves and higher run-up of the waves. Melet et al. (2018) demonstrate, by analysing coastal sea-level observations, that decadal changes in wave climate have an important role in dampening or enhancing local rates of the sea-level change. Coupled

climate and hydrodynamical models are needed to resolve the importance of these local-scale processes for local sea-level changes, as pioneered by Vousdoukas et al. (2018).

In addition to waves, tidal characteristics may change over time. Historically, ocean tides have been considered stationary because of their close relationship to astronomical motions (Cartwright and Tayler 1971) including a small nodal cycle with a length of 18.6 years. Nevertheless, long-term tidal evolution has been reported for some locations (Cartwright 1972; Amin 1983; Bowen 1972; Familkhalili and Talke 2016), mainly associated with harbour modifications and changes in water depth. In addition, recent studies, however, provide evidence that ocean tides are changing in some ocean basins at diverse rates without any relationship to astronomical forcing (Mawdsley et al. 2015; Muller 2011). Changes in tides have been observed in the Gulf of Maine, (Ray 2006) in the North Atlantic (Muller 2011) and in China (Feng et al. 2015). Analysis for the Pacific Ocean (Devlin et al. 2017) suggests that sea-level variability is correlated with interannual tidal variability at most of the tide gauges in the Pacific. These tidal anomalies, while influenced by basin-scale climate processes and sea-level changes, appear to be locally forced and not coherent over amphidromic points (i.e. tidal nodes) or basin-scale. A further possibility for tides to change is related to increased water depths associated with sea-level rise (Pickering et al. 2017; Howard et al. 2019) leading to shifts of the amphidromic points and local tidal characteristics. Detailed hydrodynamical models may capture these uncertainties in near future simulations.

Seasonal changes to stratification and thermocline depth caused by variations in water temperature and wind produce a steric sea-level signal and alter the surface manifestation of internal tides, creating a detectable change in tide gauge records (Colosi and Munk 2006). Seasonal changes in wave speed in shallow coastal areas also modify propagation and dissipation of tides (Arbic and Garrett 2010). Furthermore, river flows and changes in run off directly influence bottom friction and modify stratification, leading to changes in tidal dynamics (Guo et al. 2015).

A recent study by Vousdoukas et al. (2018) demonstrates that future changes in tides are driven by sea-level trends, and projected changes in tides are higher for the RCP8.5 scenario compared to the RCP4.5. Increase in tidal range of several cm is projected along parts of north Australia, east Patagonia and the Sea of Okhotsk. However, for most of the world regional changes in tides are insignificant in comparison with the other components of sea-level rise at centennial time scales.

5 Discussion and Conclusions

As sea-level is expected to rise for over the twenty-first century and beyond (Oppenheimer et al. 2019), and likely at an accelerated pace, there is a need for more detailed information near the coasts. CMIP5 generation climate models (GCMs) are unable to represent important processes relating to extreme sea level (ESL) events and coastal processes. This poses challenges from an adaptation point of view, as sometimes a full probability density function is needed including high-end scenarios, whereas climate models as used in IPCC context (Church et al. 2013) only offer, for good reasons, a likely range. Risk adverse users of this information may wish to take a more conservative approach that explicitly considers information on the "tail risks" (Stammer et al. 2019; Hinkel et al. 2019; Jevrejeva et al. 2019). A key limitation is our understanding of the future response of the ice sheets, particularly from ice dynamic processes, and the potential for accelerated sea-level rise. As

a result, conditional probability distributions are provided in the literature which assume a certain Antarctic contribution based on expert judgement or on a single ice dynamical study (e.g., Bamber and Aspinall 2013; Kopp et al. 2014; Jackson and Jevrejeva 2016; Slangen and Lenaerts 2016; Bamber et al. 2019).

However, in order to get to the local scale where adaptation measures are to be taken, many other issues must also be resolved. Variations in ESL are among the factors to be considered; for this reason, adaptation measures are very local and need to evolve over time and are therefore sometimes framed as adaptation pathways (Haasnoot et al. 2011). One possibility to inform adaptation strategies is to include more detailed climate models near the coast, which help to resolve the shallowing of the bathymetry near the coast and its effect on local water level. Such models may also be combined with hydrodynamical models to better represent small-scale coastal processes. Around the Antarctic ice sheet, higher resolution in climate models may help to resolve processes associated with coastal water masses and/or circulation on the continental shelf and result in improved simulations of future ice sheet response. Nevertheless, improved resolution is not a panacea for the problems arising from a lack of small-scale physical processes in climate models. Particular challenges are the representation of the AMOC in models for patterns of sea-level change in the North Atlantic region and more generally the ocean mass distribution near the continental shelves, which is critical for coastal sea-level change.

Inclusion of small-scale processes like ice shelf hydrofracturing will likely need to be parameterized for the foreseeable future, and GCMs are yet to include the effects of tides and wave set-up on simulated sea level. Hence, to answer questions with respect to local coastal sea level, we have to improve the physics of climate models and combine them with hydrodynamical and/or regional ocean models which capture changes in local bathymetry, the interaction between surges, waves and tides, such that we can provide more information than only the RSL at 100-km resolution and move towards projections of high water levels at 10-km resolution. A more detailed outline for model improvements is provided by Ponte et al. (2019), but combining hydrodynamical and regional ocean models is already feasible and data for validation are available. Combining this with new data assimilation techniques will promote more realistic coastal information which is better suited for coastal decision making.

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