Toward a physically plausible upper bound of sea-level rise projections

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Accepted to: Climatic Change Letters (September 27, 2012)

Key Words: Sea-Level Rise, Climate Dynamics, Uncertainty Quantification, Risk Analysis
Abstract

Anthropogenic sea-level rise (SLR) causes considerable risks. Designing a sound SLR risk-management strategy requires careful consideration of decision-relevant uncertainties such as the reasonable upper bound of future SLR. The recent Intergovernmental Panel on Climate Change’s (IPCC) Fourth Assessment reported a likely upper SLR bound in the year 2100 near 0.6 m (meter). More recent studies considering semi-empirical modeling approaches and kinematic constraints on glacial melting suggest a reasonable 2100 SLR upper bound of approximately 2 m. These recent studies have broken important new ground, but they largely neglect uncertainties surrounding thermal expansion (thermosteric SLR) and/or observational constraints on ocean heat uptake. Here we quantify the effects of key parametric uncertainties and observational constraints on thermosteric SLR projections using an Earth system model with a dynamic three-dimensional ocean, which provides a mechanistic representation of deep ocean processes and heat uptake. Considering these effects nearly doubles the contribution of thermosteric SLR compared to previous estimates and increases the reasonable upper bound of 2100 SLR projections by 0.25 m. As an illustrative example of the effect of overconfidence, we show how neglecting thermosteric uncertainty in projections of the SLR upper bound can considerably bias risk analysis and hence the design of adaptation strategies. For conditions close to the Port of Los Angeles, the 0.25 m increase in the reasonable upper bound can result in a flooding-risk increase by roughly three orders of magnitude. Results provide evidence that relatively minor underestimation of the upper bound of projected SLR can lead to major downward biases of future flooding risks.
1. Introduction

Anthropogenic climate forcings are projected to drive considerable sea-level rise (SLR) and climate risks (Milne et al. 2009; Vellinga et al. 2009; Nicholls et al. 2011). The design of SLR risk-management strategies hinges critically on SLR projections (Hunter 2011; Lempert et al. 2012). One important characteristic of SLR projections is the plausible upper bound (Pfeffer et al. 2008). This metric can be an important factor for the design of risk-management strategies, for example in situations where the costs of coastal defense are relatively small compared to the negative impacts of flooding (Lempert et al. 2012). The need for strategies to manage risks associated with low probability/high consequence scenarios such as these was recently highlighted as a major scientific challenge in assessing climate change impacts and implications for adaptation (NAS 2010).

Estimates of plausible upper bounds are uncertain (Pfeffer et al. 2008; Milne et al. 2009). This uncertainty is caused by factors such as: (i) the inability of the current generation of Earth system models to simulate important processes, (ii) uncertainty in observations used to constrain these projections, and (iii) uncertainties in future anthropogenic forcings (Meehl et al. 2007; Moss et al. 2010). Estimates of the plausible upper bound have changed historically as a result of the improvements in Earth system models, the availability of more observations, and increasingly powerful methods to use these observations to constrain SLR projections (Titus et al. 1995; Meehl et al. 2007; Pfeffer et al. 2008). Titus et al. (1995), for example, estimates the 99th percentile of SLR in the year 2100 as approximately 1 m. The
most recent Intergovernmental Panel on Climate Change (IPCC) report provides an upper
range of approximately 0.6 m, but this estimate excludes “future rapid dynamical changes in
ice-flow” (Alley et al. 2007). More recent estimates using semi-empirical approaches suggest
a 2100 upper bound near 2 m (Vermeer and Rahmstorf 2009; Jevrejeva et al. 2010). A novel
study by Pfeffer et al. (2008) provides kinematic constraints on the rapid dynamical changes
in ice flow and reports SLR in 2100 is likely to be on the order of 1 m, but extreme scenarios
with SLR up to 2 m may be possible. The Pfeffer et al. (2008) study breaks important new
ground by considering effects of melting land ice, but it is silent on the effects of uncertainties
surrounding oceanic thermal expansion, or thermosteric SLR.

Here we use the University of Victoria’s Earth System Climate model (UVic) (Weaver
et al. 2001; Olson et al. 2012) to estimate the thermosteric contribution to 2100 SLR,
including the effect of heat uptake by the deep ocean. We constrain the projections using the
observational record and we quantify the model’s parametric uncertainties, in order to
estimate a physically plausible upper bound of the thermosteric contribution to 2100 SLR.
We demonstrate how accounting for these effects increases the reasonable upper bound of
projected SLR in 2100, using the high SLR scenario of Pfeffer et al. (2008) as the foundation
of our analysis. We highlight the Pfeffer et al. (2008) result, because it reflects a
comprehensive recent estimate of uncertainties controlling SLR except for thermosteric
effects, but the results and interpretations apply for any SLR scenario neglecting thermosteric
uncertainty.
The aim of this study is to examine how the inclusion of uncertainties related to thermal expansion affects upper bound estimates of 2100 SLR and to explore implications for assessing future flooding risks. Using observations of intra-annual sea-level variability close to the Port of Los Angeles, we show that relatively small (~10%) increases in the upper bound of SLR can increase local flooding risks by roughly three orders of magnitude. Our study is silent on potentially important additional uncertainties, such as changes in future storm surge characteristics (Lin et al., 2012). We discuss this and other caveats in the conclusions.

2. Methods

2.1 Model Description

We use a modified version 2.8 of the University of Victoria Earth System Climate Model (UVic) (Weaver et al. 2001). The model consists of a three-dimensional ocean general circulation component (MOM2) coupled to a single-layer energy-moisture balance atmosphere component. It includes thermodynamic/dynamic sea ice and thermomechanical land ice models. The ocean resolution is 1.8° latitude by 3.6° longitude with 19 unevenly spaced vertical levels. The model was spun up for 3500 years using pre-industrial atmospheric forcing fields (calendar year 1800). We performed five spin-ups, corresponding to the five distinct values of the ocean background diffusivity parameter (see next section for discussion on parameters). The transient simulations are performed from 1800 to 2100 using
the RCP 8.5 pathway (Moss et al. 2010). The specified time-varying RCP8.5 forcings include: atmospheric carbon dioxide concentrations, radiative forcings from all other non-CO$_2$ greenhouse gases, and spatially resolved changes in albedo associated with direct effects of sulfate aerosols. In addition, the model includes historic volcanic radiative forcings and updated solar forcing, as documented in (Olson et al. 2012). We have also updated the model’s Schmidt number parameterization (Wanninkof 1992), relating tracer diffusion over the ocean to surface temperature, from a polynomial to exponential expression to ensure stability for scenarios with significant warming.

2.2 Ensemble Design

We construct a 250-member perturbed physics ensemble experiment using the UVic model, varying three model parameters that influence thermosteric SLR hindcasts and projections. Each ensemble member represents a unique combination of the following three model parameters: vertical ocean background diffusivity ($K_v$), aerosol scaling (AS), and climate sensitivity (CS). The parameter $K_v$ affects vertical mixing in the ocean and the rate of heat uptake. The parameter AS represents a scale factor applied to sulfate albedos modulating the radiative effects of anthropogenic sulfate aerosols. The parameter CS is the model’s equilibrium climate sensitivity (CS), defined as the equilibrium surface temperature response to a doubling of atmospheric carbon dioxide relative to pre-industrial concentrations.

The model ensemble samples parameter ranges based on previous analyses using expert assessment, along with instrumental and paleo-derived observations (Knutti and Hegerl...
Kv is varied on a uniform grid with values of (0.1, 0.2, 0.3, 0.4, 0.5) cm/s². This range is consistent with previous UVic parameter estimation analyses using observational constraints (Goes et al. 2010; Olson et al. 2012). AS is also varied on a uniform grid with values of (0, 0.75, 1.5, 2.25, 3), consistent with the range outlined in the IPCC’s Fourth Assessment (Hegerl et al. 2007) and a separate UVic parameter estimation study (Olson et al. 2012). The model’s equilibrium climate sensitivity (CS) is a diagnosed parameter. We vary CS by adjusting the local outgoing longwave radiation at the top of the atmosphere using a global parameter (f*) as in Olson et al (2012):

\[ OLR^*(t) = OLR(t) + f^*(T(t) - T_o) \]

where \( OLR^*(t) \) is the time-varying local perturbed outgoing longwave radiation, \( OLR(t) \) is the time-varying local unperturbed outgoing longwave radiation, \( f^* \) is the global input parameter, \( T(t) \) is the time-varying local surface temperature, and \( T_o \) is local equilibrium surface temperature (prior to onset of time-varying forcings). This technique relates the input parameter \( f^* \) to the local temperature anomalies at each grid point, thus preserving influences of regional variability on the global temperature response. As noted previously, CS is a diagnosed quantity in UVic which is sensitive only to \( f^* \) on sufficiently long (millennial) time scales, when the full ocean can be considered in a state of approximate dynamic equilibrium. In order to map \( f^* \) to CS, we analyze double CO₂ simulations starting from the equilibrated pre-industrial spin-up run with Kv set to the model’s standard value (Kv=0.1 cm/s²). Ten double CO₂ simulations were integrated for 3000 years for the following \( f^* \)
values: (2582.9, 1484.7, 851.4, 480.1, 203.2, -40.7, -359.8, -526.4, -681.3, -844.2). The corresponding diagnosed CS values are: (11.2, 8.2, 6.5, 5.4, 4.0, 3.1, 2.6, 2.2, 1.6, 1.1) °C, representing the ensemble grid design points shown in Figure 1.

2.3 Thermosteric SLR Calculation

Thermosteric SLR is calculated from the three-dimensional density field, following from previous methods (Yin et al. 2009):

\[
SLR = \frac{1}{A} \int_A \int_z \frac{\Delta \rho}{\rho} dz dA
\]

where \( \rho \) is seawater density, \( A \) is ocean surface area, and \( z \) is depth. Density is derived from the equation of state for seawater (McDougall et al. 2003). We show global average SLR integrated from the surface to two different reference depths: 700 meters (for data comparison) and full ocean depth (for projections). Modeled SLR hindcasts and projections are relative to the average during the years corresponding to the observations (1950-2003) (Domingues et al. 2008).

3. Results and Discussion

The ensemble includes CS values between roughly 1 and 11 °C. It has been argued that CS values near the upper limit of our range represent scenarios with small, but finite probabilities (Knutti and Hegerl 2008). In order to obtain a plausible upper bound of
thermosteric SLR, we only consider ensemble members with CS values which are within a probable range based on previous UVic analysis (Olson et al. 2012). We adopt as our prior a recently published posterior probability density function for CS (Olson et al. 2012), derived from a Bayesian data-model fusion of the UVic model and instrumental observations of surface temperature and ocean heat (Figure 1). This adopted prior reflects the current best-estimate of the CS range based on previous comprehensive UVic analysis utilizing expert elicitation, paleo-information, and the instrumental records of changes in surface air temperature and oceanic heat content (Olson et al, 2012). We sample CS parameters that fall within a 99% credible interval of the prior (Figure 1). The CS sampling filters 40% of the ensemble members, thus 100 of the original 250 ensemble members are omitted. The omitted ensemble members reflect CS values outside the range 1.5-6 deg C.

We further constrain the model projections using observations of thermosteric SLR in the top 700 meters (Domingues et al. 2008). Ocean temperature observations below ~1000 meters are relatively sparse. Previous efforts examining observed trends in global thermosteric SLR during the latter half of the 20th century typically rely on strong assumptions about the contribution of the deep ocean, such as assuming constant linear trends (Domingues et al. 2008). We improve on this heuristic approach by using the UVic Earth system model, which includes a three dimensional ocean general circulation model component. This allows us to simulate the global circulation and thus infer total thermosteric SLR mechanistically throughout the full ocean. Specifically, we partition the model's thermosteric SLR into two categories: the upper 700 meters and the full ocean. Results from the upper 700 meters are
used to compare modeled thermosteric SLR against the observational record (Domingues et al. 2008), while the full ocean results are used to make projections to 2100 (Figure 2). This technique assumes that the accuracy of the model’s representation of the full-ocean dynamics can be diagnosed based on the agreement with the global average thermosteric SLR within the uppermost 700 meters between 1950 and 2003. We test this assumption by analyzing the mapping between the modeled SLR due to the upper ocean compared to the full ocean in Figure 2C. The figure shows a general linear relationship between thermosteric SLR within the upper 700 meters and the full ocean, with a slope greater than 1. Thus, the deep ocean contributes to the total thermosteric SLR through enhanced ocean heat uptake, and the relative contribution increases for higher SLR scenarios. However, the modeled SLR signal is contained primarily within the upper 700 meters.

We apply a simple model calibration technique (Knutti et al. 2002) to exclude ensemble members that do not generally reproduce interannual variability in observed upper ocean thermosteric SLR (Domingues et al. 2008). This windowing technique rejects ensemble members that, on average, do not fall within two standard deviations of the observations (Figure 2). The windowing approach filters an additional 29 ensemble members from the SLR analysis, however, it does not affect the projected range of thermosteric SLR compared to the range based only on the CS sampling. This result provides an important cross-check on the CS prior probability density function used in our sampling, in that instrument-based records of ocean heat content and thermosteric SLR generally yield similar ranges of CS for the UVic model. This consistency reflects the close relationship between
The upper bound of our projected thermosteric SLR in 2100 is around 0.55 m, nearly twice the projected thermosteric contribution of 0.3 m used in Pfeffer et al. (2008) that neglects uncertainty about thermosteric SLR. The range is roughly comparable with the likely range of projected thermosteric SLR published in the IPCC’s Fourth Assessment Report (Meehl et al. 2007), though our upper bound is around 10 cm larger (Figure 2). Direct comparison of SLR ranges is challenging due to the mixture of parametric uncertainties, shown here using the UVic model, and structural uncertainties arising from the inter-model comparison of the IPCC’s Fourth Assessment.

As a useful starting point of how thermosteric uncertainties affects the upper bound of SLR projections, we use the high SLR scenario of 2 m by Pfeffer et al. (2008) as the foundation of our analysis. Our result shows that this upper bound would need to be increased from 2 to 2.25 m when considering thermosteric uncertainties. While this change in the upper bound is relatively modest (~12.5% increase), the underestimation of uncertainty (or overconfidence) has considerable implications for flooding risk projections and the design of risk-management strategies.

Projected SLR has become an important factor in assessments of future flooding-risk as well as risk-management strategies (Yohe et al. 1996; Purvis et al. 2008; Lempert et al. 2012). The plausible upper limit of SLR plays an important role in these assessments in at
least two situations. First, the plausible upper bound is often used as a worst-case scenario in vulnerability assessments (e.g., Tol et al. 2006). Second, in situations where the costs of heightening a coastal defense system are very small compared to the negative impacts of flooding, designing a coastal defense system for the plausible upper bound of SLR can be a reasonable approximation to an economically efficient strategy (Lempert et al. 2012).

Overconfidence in the plausible upper bound of SLR can lead to considerable downwards biases in the projected flooding risks and the design-height of coastal defenses. This effect is illustrated and quantified for an example location close to the Port of Los Angeles (Figure 3). At this location, the past observed intra-annual sea-level variations (Caldwell 2010) result in an approximately one-in-50 year flooding at approximately 1.5 m above mean sea-level (red circles and the fitted statistical model shown by the black line in Figure 3). A coastal defense system designed to maintain the same flooding frequency and assuming a SLR upper bound of 2 m (Pfeffer et al. 2008) would be located at 3.5 m (equal to 1.5 m for the flooding frequency at the current sea-level plus the assumed SLR upper limit of 2 m). If SLR follows the revised plausible upper limit of 2.25 m derived in this study (Figure 3, green line), which considers parametric uncertainties related to thermosteric contribution, then the flooding frequency would be nearly three orders of magnitude higher. This result demonstrates how small changes in the projected upper bound can lead to large downward biases in risk assessments and under-investments in risk-reduction strategies.
4. Conclusions and Caveats

We present results from a perturbed physics ensemble experiment using an Earth system model with a dynamic ocean component to quantify a physically plausible range in thermosteric SLR projections under parametric uncertainty. The intention of this analysis is primarily to demonstrate the potential for Earth system models to help inform risk- and decision-analysis. While there are also considerable uncertainties related to the dynamics and future evolution of land ice (Alley and Joughin 2012; Willis and Church 2012), these results are generally applicable to any upper bound which neglects thermosteric uncertainties. Recent observations indicate Greenland glacier accelerations are relatively small compared to previous estimates (Jacob et al. 2012; Moon et al. 2012). If this continues, the projected 2100 SLR may be considerably less than the high-end scenario proposed by Pfeffer et al. (2008). Hence, the contribution of thermosteric uncertainty would become relatively more significant if SLR upper bounds based on kinematic constraints of glacier outlet velocities are revised downward.

Because the model robustly represents global ocean circulations, our methodology also implicitly includes SLR uncertainties related to possible changes in global ocean circulation under anthropogenic climate change (e.g. Knutti and Stocker 2000). For example, the model’s projected changes in meridional overturning strength are sensitive to the parameter combinations, with the Kv and CS being the main factors. This result suggests the ensemble experiment is covering a broad range of thermosteric SLR associated with significant
dynamical changes in the global-scale meridional structure of ocean transport.

These analyses include several simplifying assumptions. For example, we quantify the effects of key parametric uncertainties using a single Earth system model (i.e., we neglect structural uncertainties arising from different models). In contrast, the most recent IPCC report (Meehl et al. 2007) is primarily considering structural uncertainty and is mostly silent on parametric uncertainty. It is not clear how considering the combined effects of parametric and structural uncertainties, in combination with more observational constraints and improved parameter estimates (e.g. CS), will affect the projections, but this is an area of active research and beyond the scope of this preliminary study. Furthermore, our assimilation method considers only equilibrium CS estimates based on 2xCO2 model simulations relative to pre-industrial conditions. The current methodology does not account for the effect of different reference states in estimating CS, nor does it include possible changes in long-time scale feedbacks (e.g. ice-albedo feedback). In addition, the flooding risk analysis neglects, for simplicity, potential changes in the distribution of the intra-annual variability associated with changing extreme weather events (e.g. tropical cyclones) (Lin et al. 2012).

Given the aforementioned caveats, we show that considering the effects of key parametric uncertainties and observational constraints increases the physically plausible upper bound of total SLR in 2100 by roughly 0.25 m. Neglecting this effect of parametric uncertainties can lead to drastic underestimation of future flooding risks. Because of the inherent linkages between projections of surface temperature and ocean heat uptake, these
results may serve as a useful starting point for constraining the uncertainty of full-ocean
contributions to thermosteric SLR for different future forcing scenarios.

Acknowledgements

This study was partially supported by the National Science Foundation (SES-1049208) and
the Penn State Center for Climate Risk Management. The authors acknowledge helpful
discussions with Robert Lempert, Nancy Tuana, Patrick Applegate, Michael Oppenheimer,
Gary Yohe, Richard Alley, and Don Wuebbles. All errors and opinions are ours.
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Figure 1. Prior range and sampling for the climate sensitivity model parameter. The black curve represents an estimate based on surface air temperature and ocean heat uptake observations and information from the last glacial maximum (Olson et al. 2012). The 99% credible interval for this estimate is shown by the horizontal black line. The horizontal red line illustrates the range of climate sensitivities in the high resolution climate models of the most recent Intergovernmental Panel on Climate Change report (Randall et al. 2007, table 8.2). The closed circles along the x axis represent the climate sensitivity values used in our ensemble used to derive the plausible range of thermosteric sea-level rise (Figure 2). The
calibration technique excludes samples with climate sensitivities outside the prior’s 99% range (open circles).
Figure 2. Hindcasts and projections of global thermosteric sea level rise (SLR) derived from the climate model ensemble. A. Comparison between estimated thermosteric SLR (in centimeters) from observations (Domingues et al. 2008) (black line) referenced over the data period, and the modeled range for the uppermost 700 meters. Gray shading represents the full ensemble range. Blue shading indicates the calibrated range, which excludes ensemble members with climate sensitivities outside the 99% range of our prior distribution (Figure 1). We apply an additional constraint that restricts the ensemble to members that are, on average, positioned within two standard deviations error for the observed time series (dashed black lines). B. Time series of the modeled global average thermosteric SLR integrated over the full ocean. Gray and blue shading is consistent with Figure 2A. The black vertical bar represents the IPCC range of thermosteric SLR projections for the A1FI scenario (Meehl et al. 2007, Table 10.7), and the red horizontal line denotes the thermosteric component in Pfeffer et al.
(2008). C. Mapping between 2100 thermosteric SLR for the upper and full ocean for all
ensemble members. Shading is consistent with Figures 2A and 2B. The black solid line
marks a linear regression for the windowed ensemble members (blue circles), and the black
dotted line denotes the one-to-one ratio.
Figure 3. Frequency of observed and projected flooding close to the Port of Los Angeles (PoLA). The red circles are the observed PoLA hourly sea-level anomalies (with respect to the annual mean) from 1920 to 2008 (Lempert et al. 2012), plotted as a survival function (1 minus the cumulative density function). The black curve is a statistical model fit to anomalies greater than 1200 mm, using a Generalized Paretal Distribution (GPD). The green and blue curves represent the projected shifts in the GPD fit for 2100 following increases in global mean sea-levels of 2 and 2.25 m, respectively. The dashed black vertical line denotes the height of a hypothetical sea wall constructed to protect against 2 meters of mean sea-level
rise. The frequency of occurrence of the 50-year flooding event (denoted by the horizontal black line) increases by roughly three orders of magnitude (horizontal dashed gray line) once the uncertainty about thermosteric SLR is considered in the upper bound estimate.