Spatial Variability of Sea Level Rise in Twenty-First Century Projections

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ABSTRACT

A set of state-of-the-science climate models are used to investigate global sea level rise (SLR) patterns induced by ocean dynamics in twenty-first-century climate projections. The identified robust features include bipolar and bighemisphere seesaws in the basin-wide SLR, dipole patterns in the North Atlantic and North Pacific, and a beltlike pattern in the Southern Ocean. The physical and dynamical mechanisms that cause these patterns are investigated in detail using version 2.1 of the Geophysical Fluid Dynamics Laboratory (GFDL) Coupled Model (CM2.1). Under the Intergovernmental Panel on Climate Change’s (IPCC) Special Report on Emissions Scenarios (SRES) A1B scenario, the steric sea level changes relative to the global mean (the local part) in different ocean basins are attributed to differential heating and salinity changes of various ocean layers and associated physical processes. As a result of these changes, water tends to move from the ocean interior to continental shelves. In the North Atlantic, sea level rises north of the Gulf Stream but falls to the south. The dipole pattern is induced by a weakening of the meridional overturning circulation. This weakening leads to a local steric SLR east of North America, which drives more waters toward the shelf, directly impacting northeastern North America. An opposite dipole occurs in the North Pacific. The dynamic SLR east of Japan is linked to a strong steric effect in the upper ocean and a poleward expansion of the subtropical gyre. In the Southern Ocean, the beltlike pattern is dominated by the baroclinic process during the twenty-first century, while the barotropic response of sea level to wind stress anomalies is significantly delayed.

1. Introduction

Sea level rise (SLR) is a central issue in climate change research. Tide gauge records show that the global mean SLR is about 1.7 mm yr\(^{-1}\) during the twentieth century (Church and White 2006; Bindoff et al. 2007), with a possible acceleration indicated by recent altimetry data (Willis et al. 2008; Leuliette and Miller 2009). The global SLR involves a significant anthropogenic component, and is mainly induced by global ocean thermal expansion and the melting of land ice. It is widely believed that SLR is a serious threat to coastal communities and is closely linked to various coastal hazards, including storm surge, inundation of low-lying areas, beach erosion, and damages to coastal infrastructures and ecosystems (Nicholls et al. 2007). Given the environmental and socioeconomic impacts of SLR, it is very important to understand past and present SLRs and generate reliable future SLR projections, thereby providing valuable information for policy makers and coastal communities.

Global mean SLR has been studied extensively (Bindoff et al. 2007). In response to a changing climate, however, SLR will not be spatially uniform, but show complex patterns, as indicated by available observations (Douglas 2001; Bindoff et al. 2007). As a result, some regions could experience local SLRs considerably faster and larger than the global mean, whereas the local SLR elsewhere may be well below the global mean or even negative. This situation suggests that each coastal community must develop its own strategy to tackle SLR according to its spatial variation. Although the SLR pattern is very important, it suffers from an insufficient amount of study to date and was simply attributed to natural geological
processes. With the recent progress in this field (Gregory et al. 2001; Levermann et al. 2005; Landerer et al. 2007; Yin et al. 2009; Mitrovica et al. 2009; Körper et al. 2009; Pardaens et al. 2010), a better understanding of the SLR patterns in past, present, and future climates, and their underlying mechanisms, have been identified as research targets with high priority in the coming decade (Milne et al. 2009).

Early simulations and projections of SLR employed simple climate models, such as those including an upwelling-diffusion oceanic model (Wigley and Raper 1993). Simple climate models are computationally cheap and able to provide insights into how global SLR is modified through various mechanisms, and under different greenhouse gas (GHG) emission scenarios. However, such simple climate models cannot simulate and project the SLR pattern. In contrast, the realistic global atmosphere–ocean general circulation models (AOGCMs) have the ability to simulate the spatial variation of SLR (Gregory et al. 2001). Application of AOGCMs to the study on SLR patterns lags other model applications because of limitations in model formulation, such as the rigid-lid approximation, Boussinesq approximation, virtual salt flux formulation (Yin et al. 2010), model resolution, lack of a dynamic land ice component, etc. In the latest generation of AOGCMs, many of these limitations have been alleviated or eliminated by some models, and SLR projections with more detailed structures and higher levels of confidence are possible.

Multiple factors can impact the absolute SLR pattern, such as ocean dynamics (e.g., Yin et al. 2009) and the impacts of the gravitational effect that modifies the geoid (Mitrovica et al. 2009). SLR patterns relative to land are also influenced by geological processes such as glacial isostatic adjustment (Peltier 2001). Modern global positioning system technology facilitates a direct detection of land motion in addition to model simulation (Wöppelmann et al. 2009). Nonetheless, even in the latest climate models, only the dynamic factor is represented, with other impacts typically associated with different scientific disciplines whose input to climate models has only recently been explored (R. Kopp 2010, personal communication). In particular, the incorporation of gravitational effects requires a dynamic ice sheet model and changes in the ocean component formulation to allow for a changing land–sea boundary, with such activities under intensive development. Presently, the projection of the land ice melting and the resultant SLR patterns have large uncertainties. In contrast, the SLR pattern induced by ocean circulation is, in principle, represented with high fidelity in AOGCMs. Namely, the undulation of sea level is a fundamental variable in free-surface ocean models, with SLR patterns reflecting aspects of ocean circulation. Nonetheless, differences do appear between models, due to differences in the water mass structure and its associated ocean circulation features. A preliminary examination of such differences is one focus of the present study.

In this study, we focus on the twenty-first century projection of the SLR pattern induced by ocean dynamics. We examine simulations of the present-day dynamic sea level (DSL) by 17 state-of-the-science AOGCMs included in the Intergovernmental Panel on Climate Change’s (IPCC) Fourth Assessment Report (AR4). The DSL indicates the deviation of sea surface from the global mean, which is closely related to ocean currents through geostrophic balance, 3D density structure, and mass distribution of the ocean. Hence, fluxes of momentum, heat, and water at the ocean surface can cause changes in the DSL (Bryan 1996; Knutti and Stocker 2000; Lowe and Gregory 2006; Levermann et al. 2005; Sakamoto et al. 2005; Landerer et al. 2007). Data from the IPCC AR4 models were obtained from phase 3 of the Coupled Model Intercomparison Project (CMIP3) database (information online at http://www-pcmdi.llnl.gov/). Based on their skills at simulating the present-day DSL, we choose a subset of models to develop an ensemble projection, in which robust features in the spatial structure of SLR are revealed. Mechanisms that cause the SLR pattern are further explored using version 2.1 of the Geophysical Fluid Dynamics Laboratory’s (GFDL) Coupled Model (CM2.1).

In all model projections, the external GHG forcing in the twenty-first century follows the IPCC’s Special Report on Emissions Scenarios (SRES) A1B scenario (Nakićenović and Swart 2000). It is found that the SLR pattern is similar in other scenario runs and only the exact magnitude differs. For comparison, we will also present the results from idealized “water hosing” experiments (Stouffer et al. 2006b; Yin and Stouffer 2007), in which an external freshwater flux of 0.1 or 1.0 Sv (Sv = 10⁶ m³ s⁻¹) is input into 50°–70°N in the North Atlantic. Water-hosing experiments are typically designed to study the impacts of the Atlantic meridional overturning circulation (AMOC) changes on climate including global and regional sea levels (Yin et al. 2010). With the analysis and comparison, our goal is to understand the mechanisms responsible for the projected spatial variability of SLR during the twenty-first century. The results also guide detection of any future changes in ocean circulation based on the DSL observations.

The paper is organized as follows. Section 2 briefly describes the models and summarizes the analysis methods. Section 3 presents the results from the simulations and projections, and section 4 offers a discussion and conclusions.
TABLE 1. The 17 IPCC AR4 models used in the present study. Terms used include the following: FS, free surface; RL, rigid lid; BO, Boussinesq; NB, non-Boussinesq; FWF, freshwater flux; and VSF, virtual salt flux. The resolution is in latitude × longitude, and boldface indicates the type of vertical coordinate.

<table>
<thead>
<tr>
<th>Model</th>
<th>Institution</th>
<th>Oceanic model</th>
<th>Sea level representation</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCCR Coupled Model, version 2.0 (BCM2.0)</td>
<td>Bjerknes Centre for Climate Research (BCCR), Bergen, Norway</td>
<td>0.5°–1.5° × 1.5°, L35, ρ</td>
<td>FS, NB, VSF</td>
</tr>
<tr>
<td>Third-generation Coupled Global Climate Model (CGCM3.1)</td>
<td>Canadian Centre for Climate Modeling Analysis (CCCMMA), Victoria, BC, Canada</td>
<td>1.9° × 1.9°, L29, Z</td>
<td>RL, BO, VSF</td>
</tr>
<tr>
<td>Coupled Model, version 2.0 (CM2.0)</td>
<td>GFDL, Princeton, NJ</td>
<td>0.3°–1° × 1°, L50, Z</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>CM2.1</td>
<td>GFDL</td>
<td>0.3°–1° × 1°, L50, Z</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>Atmosphere–Ocean Model (AOM)</td>
<td>Goddard Institute for Space Studies (GISS), New York, NY</td>
<td>3° × 4°, L16, P</td>
<td>FS, NB, FWF</td>
</tr>
<tr>
<td>GISS-EH</td>
<td>GISS</td>
<td>2° × 2°, L16, ρ</td>
<td>FS, NB, VSF</td>
</tr>
<tr>
<td>GISS-ER</td>
<td>GISS</td>
<td>4° × 5°, L13, P</td>
<td>FS, NB, FWF</td>
</tr>
<tr>
<td>Coupled Model, version 4 (CM4)</td>
<td>Institut Pierre Simon Laplace, Paris, France</td>
<td>2° × 2°, L31, Z</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>Model for Interdisciplinary Research on Climate 3.2, high-resolution version (MIROC3.2 HIRES)</td>
<td>University of Tokyo, Tokyo, Japan</td>
<td>0.2° × 0.3°, L47, Zls</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>MIROC3.2, medium-resolution version (MEDRES)</td>
<td>University of Tokyo</td>
<td>0.5°–1.4° × 1.4°, L43, Zls</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>ECHAM and the global Hamburg Ocean Primitive Equation (ECHO G)</td>
<td>Meteorological Institute, University of Bonn (MUIB), Bonn, Germany; Korea Meteorological Administration, Seoul, South Korea</td>
<td>0.5°–2.8° × 2.8°, L20, Z</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>ECHAM5</td>
<td>Max Planck Institute for Meteorology (MPI), Hamburg, Germany</td>
<td>1.5° × 1.5°, L40, Z</td>
<td>FS, BO, FWF</td>
</tr>
<tr>
<td>CGCM2.3.2</td>
<td>Meteorological Research Institute, Tsukuba, Japan</td>
<td>0.5°–2° × 2.5°, L23, Z</td>
<td>RL, BO, VSF</td>
</tr>
<tr>
<td>Community Climate System Model, version 3 (CCSM3)</td>
<td>National Center for Atmospheric Research (NCAR), Boulder, CO</td>
<td>0.3°–1° × 1°, L40, Z</td>
<td>FS, BO, VSF</td>
</tr>
<tr>
<td>Parallel Climate Model (PCM)</td>
<td>NCAR</td>
<td>0.5°–0.7° × 1.1°, L40, Z</td>
<td>FS, BO, VSF</td>
</tr>
<tr>
<td>Third climate configuration of the Met Office Unified Model (HADCM3)</td>
<td>Hadley Centre for Climate Prediction and Research, Bracknell, United Kingdom</td>
<td>1.25° × 1.25°, L20, Z</td>
<td>RL, BO, VSF</td>
</tr>
<tr>
<td>Hadley Centre Global Environmental Model, version 1 (HADGEM1)</td>
<td>Hadley Centre for Climate Prediction and Research</td>
<td>0.3°–1° × 1°, L40, Z</td>
<td>FS, BO, FWF</td>
</tr>
</tbody>
</table>

2. Model description and analysis methods

a. Model description

Table 1 lists the 17 IPCC AR4 models with the DSL simulations and projections available at the CMIP3 archive used by the IPCC AR4. Detailed information for each model has been described by Randall et al. (2007) and references therein. Configuration details related to the DSL representation differ among the models (Table 1). Notable details include grid resolution, vertical coordinate, algorithms to compute sea surface height, use of Boussinesq or non-Boussinesq formulation, and real water fluxes versus virtual salt fluxes. Some of the differences in sea level patterns between the models presented in this paper may be related to differences in these configuration details, though it is beyond the scope of this paper to unravel such dependencies.

We use the DSL representation in the GFDL CM2.1 model (Delworth et al. 2006) for in-depth analysis of the causes of certain features found in the ensemble SLR pattern. Systematic assessment indicates that CM2.1 realistically simulates many features of the climate system such as ocean temperature and salinity (Gleckler et al. 2008). The horizontal resolution of the atmospheric model and the land model in CM2.1 is 2° latitude × 2.5° longitude. The atmospheric model has 24 vertical levels. The oceanic component is a Boussinesq free-surface general circulation model, which explicitly represents the freshwater flux at the ocean surface (Griffies et al. 2005). It employs 1° horizontal resolution with the meridional resolution refined to 1/3° in the tropics. It uses a depth vertical coordinate and has 50 levels with 22 levels in the upper 220 m. The dynamical–thermodynamical sea ice model calculates ice internal stresses using the elastic–viscous–plastic technique (Delworth et al. 2006).
b. Analysis methods

CM2.1 and other AR4 models used in this study employ the hydrostatic balance, in which case the bottom pressure \( p_b \) is given by the sum of the applied pressure at the ocean surface \( p_a \) plus the weight per area of seawater in an ocean column:

\[
p_b = p_a + g \int_{-H}^{h} \rho \, dz,
\]

where \( \rho \) is the in situ seawater density, \( z = \eta(x, y, t) \) the ocean free-surface deviation from the geoid at \( z = 0 \), and \( z = -H(x, y) \) the solid-earth lower boundary. Following Gill and Niiler (1973), who studied seasonal variations in the ocean, as well as Landerer et al. (2007) and Yin et al. (2009), who studied sea level trends under global warming, we use the time tendency of the hydrostatic balance to interpret spatial pattern changes in sea level:

\[
\frac{\partial \eta}{\partial t} = \left( \frac{1}{\rho_0 g} \right) \frac{\partial (p_b - p_a)}{\partial t} - \frac{1}{\rho_0} \int_{-H}^{h} \frac{\partial \rho}{\partial t} \, dz,
\]

where we approximate the surface density \( \rho(z = \eta) \) with the constant reference density \( \rho_0 \). The first term on the right-hand side is associated with changes in mass within a seawater column (the mass term), and the second term arises from changes in density (the steric term).

The surface of liquid water \( \eta_{\text{water}} \) computed by an ocean model feels the effects from the applied pressure of sea ice. An expression of sea level for use in climate studies includes the inverse barometer effect from sea ice, in which case we define an effective DSL:

\[
\eta(x, y, t) = \eta_{\text{water}}(x, y, t) + \frac{M_{\text{ice}}(x, y, t)}{\rho_0},
\]

with \( M_{\text{ice}}(x, y) \) the mass of sea ice per unit area. Use of Eq. (2) and the replacement of time tendencies with temporal increments relative to an initial reference state (e.g., the start of a global warming simulation) renders

\[
\Delta \eta(x, y, t) = \frac{\Delta p_b(x, y, t)}{g \rho_0} + \Delta h_a(x, y, t) + \Delta h_s(x, y, t),
\]

where \( \Delta h_a = -\Delta p_a/\rho_0 \) is the atmospheric inverse barometer and \( \Delta h_s = -(1/\rho_0) \int_{-H}^{h} \Delta \rho \, dz \) is the steric contribution. Note that most of the coupled models examined here ignore the effects of anomalous atmospheric pressure on the ocean. In addition, because the focus of the present study is on the pattern of SLR, all terms in Eq. (4) are defined as the deviation from their global means. In particular, the steric SLR \( (\Delta h_s) \) can be decomposed into a global mean \( (\bar{\Delta h}_s) \) and a local deviation \( (\Delta h'_s) \). It is the latter that is important to our understanding of the SLR pattern [Eq. (4)]. We use “global” and “local” steric SLR to distinguish these two parts hereafter.

Many of the ocean models in Table 1, including CM2.1, are based on the Boussinesq approximation. In the absence of net surface water fluxes, a Boussinesq model conserves volume, not mass. As pointed out by Grebacz (1994), such models are unable to capture the global steric SLR associated with changes in the global mean density. Also, the bottom pressure is corrupted due to spurious mass sources required to conserve volume rather than mass. The commonly used modification to sea level, though approximate, is thought to be sufficient for our purposes. So the total SLR \( (\Delta h) \) in CM2.1 is the sum of the dynamic SLR and a global uniform correction term:

\[
\Delta h = \Delta \eta + \frac{V^0}{A} \left( 1 - \frac{\langle \rho \rangle}{\langle \rho^0 \rangle} \right),
\]

where \( \langle \rho \rangle \) is the global mean ocean density with an initial value \( \langle \rho^0 \rangle \), \( V \) is the global ocean volume with an initial value of \( V^0 \), and \( A \) is the horizontal area of the ocean surface, assumed to be a constant in all of the ocean models considered here (i.e., no model has an algorithm for evolving shorelines). As the mean density decreases, such as when the ocean warms with a positive thermal expansion coefficient, then the corrected sea level increases. This effect is one of the dominant reasons for the observed changes in sea level during the twentieth century, making this global steric correction a critical element in the use of Boussinesq ocean models for sea level studies.

In addition to mapping the steric increment \( (\Delta h_s) \), it is of interest to partition this increment into contributions from temperature and salinity changes. For this purpose, we introduce the thermosteric and halosteric sea level increments:

\[
\Delta h_s^{\text{thermo}} = \frac{1}{\rho_0} \int_{-H}^{h} \left[ \rho(T, S, P) - \rho(T, S, P_0) \right] \, dz \quad \text{and} \quad \Delta h_s^{\text{halo}} = \frac{1}{\rho_0} \int_{-H}^{h} \left[ \rho(T, S, P) - \rho(T, S, P_0) \right] \, dz,
\]

where \( T, S, P, \) and \( \eta_{\text{R}} \) are the temperature, salinity, pressure, and sea level for the reference state. Notably, since ocean density is a nonlinear function of temperature, salinity, and pressure, any such partitioning of the
Fig. 1. (top left) Observed and (remaining panels) simulated present-day (1992–2002) DSL (m). The data from 17 AR4 models are available at the CMIP3 Web site. The global mean DSL should be zero by definition and has been subtracted in case it is nonzero in some models. The observation is based on altimetry data (Maximenko and Niiler 2005).
density change will be approximate. Additionally, in an ocean with roughly constant salt content, the dominant effect on the global mean density change arises from temperature changes [i.e., thermosteric SLR; Lowe and Gregory (2006)]. However, climate change is associated with increased sea ice melt, land ice melt, changing precipitation and evaporation patterns, and salt redistribution in the ocean, thus making the halosteric sea level change important regionally, such as in the Arctic.

3. Results

a. Ensemble projections of IPCC AR4 climate models

In the current climate, the DSL shows both inter- and intrabasin scale features (Fig. 1; Maximenko and Niiler 2005). The DSL is very low in the Labrador Sea and the Nordic Seas, which is associated with the cyclonic circulation and deep water formation, and in the Southern Ocean, which reflects the strong Antarctic Circumpolar Current (ACC). In contrast, DSL is high in the tropical and subtropical Pacific and Indian Oceans. On average, the DSL in the Pacific and Indian Oceans is about 0.4 m higher than the global mean, whereas in the Atlantic, Arctic, and Southern Ocean, it is about 0.1, 0.4, and 0.6 m lower than the global mean, respectively. Hence, there is a pronounced Atlantic–Pacific DSL asymmetry and an equator-to-pole DSL gradient. The interbasin DSL difference drives the Bering Strait and Indonesian Throughflows, facilitating water exchanges between different ocean basins. It is also closely linked to the formation and propagation of different water masses, and to the global ocean conveyor belt (Seidov and Haupt 2005).

In the Atlantic and Pacific, the DSL patterns are skewed toward the western basin and the center of the subtropical gyres, with the highest DSL value found in the western North Pacific (Fig. 1). In addition, the DSL exhibits a sharp gradient across the western boundary currents such as the Gulf Stream and Kuroshio. The associated pressure gradient force is required by geostrophy to maintain the narrow and strong currents. Along the equatorial Pacific where the Coriolis force approaches zero, the east–west DSL difference of about 0.5 m drives the strong Pacific equatorial undercurrent, which is a critical element in ENSO. For example, the GFDL CM2.0 with the smallest $\varepsilon$ has limitations when applied for regional DSL projection in the North Atlantic, because no deep convection is simulated in the Labrador Sea (Yin et al. 2009).

We exclude the 5 models with $\varepsilon$ greater than 0.3 m, and treat the other 12 models equally in calculating the model ensemble mean. The ensemble of models outperforms any individual model in terms of twentieth-century DSL simulations (Fig. 2). The ensemble mean projection of the SLR pattern in the twenty-first century shows several robust inter- and intrabasin-scale features (Figs. 3, 4, and 5b), including (a) bihemisphere and bipolar seesaw patterns in the basin-wide dynamic SLR (Fig. 5b), (b) a dynamic SLR along the northeast coast of North America and on the Arctic coast (Fig. 4), (c) a dynamic SLR in the North Pacific subtropical gyre region, and (d) beltlike changes in the Southern Ocean. Compared to the previous projections (Gregory et al. 2001), current climate models still show disagreement, especially in the Arctic and Southern Oceans (Fig. 4b).
Fig. 3. Projections of the DSL anomalies (m) during 2091–2100 relative to 1981–2000 by 17 AR4 models. The results are from the A1B scenario run. The ensemble mean is calculated using the 12 models with relatively small RMSEs. The global mean value of the anomalies is subtracted in all panels.
The projection from CM2.1 is qualitatively similar to the multimodel ensemble mean. Given that the full dataset of CM2.1 is accessible and extra experiments are available, we use CM2.1 to further investigate the SLR pattern and its related mechanisms.

b. Interbasin sea level changes in CM2.1

The DSL simulation by CM2.1 is realistic during the twentieth century (Figs. 5a and 6). Under the SRES A1B scenario, CM2.1 projects different mean DSL changes in different oceans. The DSL rises in the North Atlantic, North Pacific, Arctic, and Indian Oceans, whereas it falls in the South Atlantic, South Pacific, and Southern Oceans (Figs. 5b, 6d, and 7b). The magnitude of the DSL change increases toward the polar regions (Fig. 6f), so that the largest changes by the end of the twenty-first century occur in the Arctic (0.09 m) and Southern Ocean (−0.04 m). Therefore, the DSL anomalies show pronounced bipolar seesaw patterns (Fig. 6f). Due to the changes in the thermohaline forcing at the ocean surface,
the AMOC weakens by 39% over the twenty-first century (Fig. 8). But this weakening does not alter the Atlantic–Pacific DSL asymmetry. The DSL slightly increases in the North Atlantic and North Pacific, but decreases in the South Atlantic and South Pacific. As a result, the pronounced Atlantic–Pacific DSL difference (Fig. 5a) and the Bering Strait Throughflow (Fig. 8) remain basically unchanged over the twenty-first century. This is in contrast to previous studies (Levermann et al. 2005; Hu et al. 2008), which showed that a significant weakening of the AMOC induced by external freshwater addition in the deep-water-formation region could cause a large dynamic SLR in the Atlantic (Fig. 5a), largely eliminating the Atlantic–Pacific asymmetry and altering the Bering Strait Throughflow. For semi-enclosed and inland seas, the DSL falls in the regions equatorward of 45°N (Mediterranean Sea, Black Sea) due to the increase in excess evaporation, whereas it rises poleward of 45°N (Hudson Bay, Baltic Sea) due to the increase in excess precipitation (Fig. 6d). These seas connect to the World Ocean in CM2.1 via mixing processes at the connecting points. The SLR at coastal regions and on the shelf is slightly faster and larger than that in the ocean interior (Fig. 7b).

We interpret the DSL changes in CM2.1 according to Eq. (4), in which two physical processes are identified: the steric effect, associated with changes in ocean density, and a mass effect, associated with a redistribution of ocean mass. We reemphasize that this interpretation follows from the hydrostatic approximation, and thus follows for both Boussinesq (as in CM2.1) and non-Boussinesq ocean formulations. The steric effect in CM2.1 causes a global mean SLR of 0.28 m over the twenty-first century [Fig. 7a and Eq. (5)]. This rise is dominated by the contribution from the upper 1000 m, which shows a roughly linear trend after the year 2040. With the gradual penetration of the warming signal into the deep ocean, the contribution of the ocean layer below 1000 m to the global steric SLR increases with time. Almost all of the global steric SLR is induced by the thermosteric effect.

The local steric SLR and its horizontal gradient are important for interpreting the SLR pattern [Eq. (4)]. The local steric sea level rises in the North Atlantic, South Atlantic, North Pacific, and Indian Oceans, whereas it lowers in the South Pacific, Arctic, and Southern Oceans (Figs. 5b and 7d). These local steric sea level changes are mainly attributed to different ocean layers and to different physical processes. Over the twenty-first century, the most pronounced local steric SLR in the upper 1000 m occurs in the North Pacific, with magnitude of up to 0.2 m in the subtropical gyre (Fig. 9b). It is caused by the halosteric effect and an upper-ocean freshening due to the increase in excess precipitation (Figs. 7f and 10b). The deep North Pacific is dominated by a thermosteric SLR smaller than the global mean, thereby contributing negatively to the local steric SLR (Fig. 9c).

In contrast, the local steric SLR in the Atlantic is attributable to the deep ocean below 1000 m (Fig. 9c) and to the thermosteric effect (Fig. 10a). Weakening of the formation and southward propagation of the cold North Atlantic Deep Water (NADW) causes a deep warming
in the entire Atlantic (Fig. 9c). With the slowdown of the upper-ocean currents, in addition, the northward heat transport in the upper Atlantic is reduced, resulting in a heat accumulation in the low and middle latitudes. These AMOC-induced thermosteric SLRs enhance the part induced by external radiative forcing, leading to a large local thermosteric SLR of up to 0.4 m in the Atlantic (Fig. 10a). On the other hand, the increase in excess evaporation causes a local halosteric sea level depression (Fig. 10b) and largely offsets the local thermosteric effect. The local thermosteric and halosteric SLRs show an opposite Atlantic–Pacific asymmetry, so that the local steric SLR is relatively small in each basin.

In a warming climate, the Arctic is projected to become fresher due to the increase in net precipitation and associated river runoff, and the melting of sea ice and the Greenland ice sheet (McPhee et al. 2009). This freshening causes a local halosteric SLR, which is compensated for and reversed by the local thermosteric effect (Figs. 7e, 7f, and 10). The reason is that the Arctic is a region of minimum ocean warming due to the melting of sea ice. This is in contrast to the warming of surface air temperature, which is characterized by polar amplification with the largest magnitude found in the Arctic (Meehl et al. 2007). In addition, thermal expansivity in polar regions is very small relative to the low latitudes (Lowe and Gregory 2006). The Arctic is also a region where the atmospheric inverse barometer effect is significant and contributes positively to the SLR due to the reduction in atmospheric mass loading on the ocean. Due to the polar amplification of the surface warming, the polar high weakens by up to 4 hPa (Fig. 11), which induces an SLR of about 0.03 m in the Arctic (Fig. 5b). The inverse barometer effect contributes negatively to the SLR in other ocean regions (Stammer and Hüttemann 2008).
In addition to the steric SLR and atmospheric pressure changes, a warming climate also causes a mass redistribution between ocean basins (Fig. 12). During the twenty-first century, mass loading decreases in the South Atlantic and Southern Oceans, but increases in the North Pacific and Arctic (Figs. 5b and 7c). The mass-induced SLR can reach 0.16 m in the Arctic and 0.05 m in the South Atlantic. The North Atlantic and South Pacific show little change in total ocean mass. Thus, ocean mass moves from the Southern Hemisphere to the Northern Hemisphere.

The mass-induced SLR is closely related to the mean depth of the ocean basins. On average, the Arctic and the South Atlantic are the shallowest and the deepest ocean, respectively, with mean depths of about 1100 and 4100 m. In response to ocean warming, the oceans with
greater depth expand more, provided that more heat penetrates into the deep ocean. This process leads to a movement of water mass between ocean basins and, more remarkably, from the ocean interior to the shelf (Figs. 7c and 12). Consequently, the local steric effect and the mass redistribution contribute together to the DSL change in the North Pacific and Southern Oceans, but oppositely in the South Atlantic, Arctic, and Indian Oceans (Figs. 5b and 7).

c. Intrabasin and local sea level changes in CM2.1

The DSL anomalies by the end of the twenty-first century show complex patterns in each ocean basin (Fig. 6d). Statistically significant differences include dipole-like patterns in the North Atlantic and North Pacific and beltlike patterns in the Southern Ocean. In the North Atlantic, the DSL rises north of the Gulf Stream and North Atlantic Current, but falls in the subtropical gyre. On the other hand, the DSL rises in the North Pacific subtropical gyre, but falls in the subpolar gyre. In the Southern Ocean, the DSL changes with opposite signs north and south of 50°S. The results from the 0.1-Sv water-hosing run help interpret the SLR pattern in the A1B scenario run, and reveal the role of the AMOC that undergoes similar weakening in the two runs (Figs. 6d and 6e). Weakening of the AMOC is responsible for the dipole pattern in the North Atlantic, and can partially explain the beltlike change in the Southern Ocean, but is not related to the DSL change in the North Pacific.

In addition to the spatial variability, the temporal variability of the DSL also changes significantly (Fig. 13). During the twentieth century, the interannual variability of the DSL is strong in the tropical Pacific and Indian Oceans, associated with ENSO and dynamics of the equatorial current system (Fig. 13a). This variability also shows a large variability in the North Atlantic Current. Due to lack of mesoscale eddies in CM2.1, the DSL variability in the Gulf Stream, Kuroshio, and ACC regions is weak. Over the twenty-first century, the interannual variability of the DSL weakens in the tropical Pacific and Indian Oceans (Fig. 13b), probably resulting from changes of ENSO, gyre circulation, and tropical precipitation (not shown). Significant enhancement is found only in the Greenland–Iceland–Norwegian Sea, which is induced by the AMOC weakening and a shift of deep convection sites. Because the focus of the present study is on the spatial variability of the SLR, a detailed investigation of its temporal properties is the subject of future research.

1) NORTH ATLANTIC

The present-day North Atlantic is characterized by a vigorous overturning circulation, which is susceptible to the increase in the GHG concentration (Stouffer et al. 1989; Schmittner et al. 2005; Gregory et al. 2005). The lowest sea level is found in the Labrador Sea and is associated with the cyclonic subpolar gyre and deep water formation. It extends southward and directly impacts the northeast coast of North America. By the end of the twenty-first century, the deep convection in the Labrador Sea and Nordic Seas weakens significantly and only that south of Iceland remains vigorous (Fig. 14). Although the weakening of the AMOC does not cause an overall dynamic SLR in the North Atlantic (Fig. 5b), it has great impacts on the local and coastal sea levels. The DSL rises rapidly at the northwestern corner of the North Atlantic, including the coastal region north of Cape Hatteras. The maximum rise of about 0.4 m occurs east of Newfoundland. By the end of the twenty-first century, the dynamic SLRs can reach 0.23, 0.23, and 0.15, respectively, in the densely populated Boston, New York City, and Washington, D.C., areas, compared to 0.04 m at Miami (Fig. 15). The superimposition of these dynamic SLRs on the global mean SLR exposes the northeastern North America to some of the fastest and largest SLRs during this century (Yin et al. 2009), and to more severe coastal hazards and potential damages such as elevated storm surges (Colle et al. 2008; Kirshen et al. 2008). Indeed, available observations show that the SLR...
Fig. 9. Local steric SLRs (m) during 2091–2100 relative to 1981–2000. The values show the A1B scenario run of CM2.1 for (a) the entire depth, (b) 0–1000 m, and (c) 1000–5500 m. Local steric SLR is calculated as $-(1/\rho_0) \int_H \Delta \rho \, dz$ with the global steric SLR subtracted.
along the northeastern U.S. coast is almost double the global mean during the twentieth century and suggest a local component of about 2 mm yr\(^{-1}\) that cannot be completely explained by isostatic adjustment (Cooper et al. 2008; Llovel et al. 2009; Wöppelmann et al. 2009).

The rapid dynamic SLR is induced by the weakening of the AMOC (Figs. 6d and 6e). Over the twenty-first century, the steric SLR south of Greenland and east of North America is much larger than the global mean (Fig. 9a). The local steric SLR is dominated by the halosteric effect north of 50\(^\circ\)N, due to an ocean freshening but cooling, and by the thermosteric effect south of 50\(^\circ\)N, due to an ocean warming but salinification (Fig. 10). The latter results from heat accumulation along the Gulf Stream and the North Atlantic Current in the upper ocean and a deep ocean warming, especially along the deep western boundary current. Both result from the weakening of the AMOC. The local steric SLR causes two sharp gradients: across the Gulf Stream and North Atlantic Current and across the shelf break (Fig. 9a). While the former is balanced by the slowing Gulf Stream and North Atlantic Current, the latter cannot be balanced by any geostrophic currents, thereby leading to an increase in the mass loading on the shelf near the
northeastern North America (Fig. 12). The ocean mass redistribution and local steric effect on the shelf contribute oppositely to the dynamic SLR, with the former dominating.

To better understand the extra steric SLR east of North America and along the deep western boundary, several vertical profiles are plotted (Fig. 16). It is evident that the ocean above 3000 m contributes positively to the absolute steric SLR at point A, which is on the route of the extra steric SLR. By contrast, the steric SLR in the subtropical gyre (point B) is smaller in the upper 800 m and from 1500 to 3000 m. So the steric SLR gradient between points A and B has an upper and a deep component accounting for the changes. In the upper ocean, the slowing Gulf Stream and North Atlantic Current is accompanied by the adjustment of the 3D density structure. In addition, there is a contribution from the lower NADW (1500–3000 m) associated with the Nordic Sea Overflow. The upper NADW associated with Labrador Seawater plays an important role in the basin-wide steric SLR (Fig. 9c), as indicated by the large steric SLR from 800 to 1500 m at both points A and B (Fig. 16).
In the subtropical gyre, the DSL falls, with a magnitude of about 0.15 m. This is consistent with a slowdown of the subtropical gyre by 14 Sv or 31% and an increase in the upper-ocean density (Figs. 8 and 9). As a result, the DSL anomalies display a remarkable dipole pattern (Bryan 1996; Gregory et al. 2001). In some other model projections, an additional DSL depression occurs in the subpolar gyre region possibly associated with a strengthening of the gyre circulation, so that the DSL anomalies show a tripole pattern (Landerer et al. 2007). In CM2.1, however, the DSL generally rises south of Greenland, suggesting a weakening (by 6 Sv or 17%; Fig. 8) and a northeastward shift of the subpolar gyre (Fig. 14) during the twenty-first century. The relative changes of the DSL with a rise along the North America coast and a fall along the European coast (Fig. 6d) are consistent with theoretical analysis, as a consequence of the slowdown of the AMOC (Bingham and Hughes 2008).

2) NORTH PACIFIC

A notable feature in the current climate is the very high DSL in the western subtropical gyre. The steep DSL gradient at the western boundary is associated with strong and narrow Kuroshio and its extension—the Pacific

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**Fig. 13.** Interannual DSL variability (m) in CM2.1. (a) The RMS variability of the DSL during 1951–2000; (b) the variability changes during 2081–2100 in the A1B scenario run. Color indicates the regions where the change is statistically significant at the 95% confidence level through the F test.
counterpart of the Gulf Stream and North Atlantic Current. Unlike that in the North Atlantic, no deep convection and no deep water formation take place in the North Pacific. So the steep DSL gradient associated with the Kuroshio is predominantly wind driven, whereas that associated with the Gulf Stream and North Atlantic Current has a significant thermohaline component. Under the SRES A1B scenario, the DSL rises rapidly in the subtropical gyre east of Japan, with a magnitude of 0.2 m (Fig. 6d). However, the anomaly contours close in the ocean interior and the dynamic SLR do not directly impact the densely populated coastal regions of East Asia. This situation differs from that in the northwestern North Atlantic, where contours of the dynamic SLR intercept the coastline of North America (Fig. 6d). The dynamic SLRs at Shanghai and Tokyo are, respectively, 0.06 and 0.08 m (Fig. 15), much smaller than that at New York City (0.23 m). The DSL falls at the northern North Pacific around the Kamchatka Peninsula, which is associated with deepening of the Aleutian low (Fig. 11) and strengthening of the subpolar gyre. Changes in the eastern North Pacific are statistically insignificant. The

FIG. 14. Mixed layer depth (m) and barotropic streamfunction (Sv) in CM2.1 for (a) 1981–2000 and (b) 2091–2100 in the A1B scenario run. Contours show the barotropic streamfunction and color shows the mixed layer depth.
SLR at San Francisco generally follows the global mean (Fig. 15).

The dynamic SLR in the subtropical gyre does not necessarily mean a strengthening of the gyre circulation. In fact, the subtropical gyre slightly weakens by 8% over the twenty-first century. This discrepancy can be explained by the following balance:

$$g V \Delta \eta = f V \Delta \psi - \frac{g}{(H + \eta) \rho_0} \int_{-H}^{H} V \Delta \rho \, dz' \, dz, \quad (9)$$

where $\psi$ is the barotropic streamfunction associated with a rigid-lid model as used by Lowe and Gregory (2006). According to Eq. (9), the dynamic SLR can be decomposed into a barotropic and a baroclinic component. In

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**FIG. 15.** Dynamic and global steric SLRs at coastal cities worldwide in CM2.1. The SLRs are relative to 1981–2000. The cities are grouped according to their geographical location. Gray dashed line is the global steric SLR. The dashed color lines are the dynamic SLRs, and the solid color lines are the sum of the dynamic and global steric SLRs.
addition to the dependence of the Coriolis parameter on latitude, the baroclinic process also can lead to differences between $\Delta \eta$ and $\Delta \phi$. Due to strong stratification and a lack of deep convection in the North Pacific, the largest ocean warming during the twenty-first century occurs in this region (not shown). The ocean expands strongly but only in the upper 1000 m (point C in Fig. 16), whereas it does so in a much thicker layer down to 3000 m in the North Atlantic (points A and B in Fig. 16). The steric SLR is not spatially uniform but increases toward the center of the subtropical gyre. This baroclinic process can steepen the DSL gradient associated with the subtropical gyre even when its strength remains unchanged or decreases slightly [Eq. (9)].

In addition, the Hadley circulation expands toward the poles in response to a warming climate (Seidel et al. 2008). The western Pacific subtropical high and associated wind system shift poleward accordingly (Figs. 11 and 17a). The anomalous wind stress causes an anticyclonic circulation north of 35°N and a cyclonic circulation south of 35°N in the North Pacific, as shown by the Sverdrup streamfunction pattern (Fig. 17b). So the dynamic SLR center east of Japan reflects a poleward expansion rather than a strengthening of the subtropical gyre during the twenty-first century (Fig. 14).

3) SOUTHERN OCEAN

The beltlike pattern of the DSL changes is remarkable and fairly robust in the AR4 model projections (Fig. 3). Generally, the DSL rises by 0.07 m north of 50°S but falls by 0.16 m south of 50°S (Fig. 18a). These anomalies are mainly induced by the strengthening and poleward shift of the Southern Hemisphere westerlies (Fig. 18a; Saenko et al. 2005; Stouffer et al. 2006a), so that the westerlies are aligned more closely with the ACC (Toggweiler 2009). Small thermal expansivity may also be important in explaining the anomaly pattern (Lowe and Gregory 2006). The changes in wind stress have two effects on the Southern Ocean circulation: enhanced zonal momentum flux and enhanced northward Ekman transport. The former tends to speed up the ACC directly by providing more mechanical energy, while the latter can modify isopycnal surfaces and influence the ACC indirectly.

However, the transport at the Drake Passage shows a decrease by about 5 Sv during the twenty-first century (Fig. 8), followed by a strong reintensification by 25 Sv during 2100–2300 when the GHG concentration is stabilized (Russell et al. 2006). Consequently, the DSL change in the twenty-first century involves a significant baroclinic component [Eq. (9)]. Indeed, the local steric and dynamic SLRs display similar geographical patterns and magnitudes (Figs. 6d and 9a), whereas the bottom pressure anomalies related to the barotropic process are secondary in explaining the DSL anomalies (Fig. 12). Although wind stress changes little after 2100, the beltlike pattern of the DSL anomaly continues to strengthen, so that the depression south of 50°S can reach 0.45 m by 2300 (Fig. 18a). There is a corresponding large decrease in the ocean bottom pressure at about 60°S (not shown). The evolutionary patterns of the ACC and DSL indicate that there is a delayed barotropic response of the ACC to the wind stress anomalies.

In response to the enhanced Ekman pumping, the meridional overturning circulation in the Southern Ocean gradually strengthens by about 4 Sv or 10% over the twenty-first century (Fig. 8), but rapidly weakens to the twentieth-century value after 2100 (not shown). During the twenty-first century, isopycnal surfaces migrate downward and poleward, notably in the region north of 50°S, with little change in the tilt of the isopycnals (Fig. 18b). Due to large thermal inertia of the ocean, the migration of isopycnals continues after 2100 and significant changes occur south of 50°S (Fig. 18b), including a steepening of the isopycnal slope associated with a stronger baroclinicity. Consequently, the responses of the overturning and ACC to the increase in the GHG concentration and wind stress anomaly are very different. The former response is a fast process, whereas the latter is a slow one. The DSL anomaly pattern is dominated by the baroclinic process during the twenty-first century, and the barotropic contribution becomes significant after 2100.
4. Discussion and conclusions

The spatial variability of sea level rise in response to climate change is an important issue but suffers from an insufficient amount of study to date. Sea level in the present-day climate shows a remarkable topography. This sea level topography is projected to change dramatically in response to a warming climate and changing ocean circulation. Based on an ensemble projection by 12 state-of-the-science climate models, we identified robust features in the SLR pattern induced by ocean circulation during the twenty-first century. These inter- and intrabasin features include bipolar and bihemisphere seesaws of the basin-averaged dynamic SLR, dipole patterns in the North Atlantic and North Pacific, and a beltlike pattern in the Southern Ocean. Mechanisms associated with these SLR patterns are examined in detail using the GFDL CM2.1 climate model.

Under the SRES A1B scenario, large changes in the basin-averaged DSL occur in the Arctic and Southern Oceans with opposite signs. The rise in the Arctic is induced by ocean mass redistribution, which is partially compensated by the local steric effect (departure from the global mean). Mass redistribution and the local steric...
effect are of the same sign, contributing together to the DSL fall in the Southern Ocean. In the middle and low latitudes, the pronounced Atlantic–Pacific DSL asymmetry is insensitive to the increase in the greenhouse-gas concentration, even though the Atlantic meridional overturning circulation weakens significantly during the twenty-first century. This insensitivity results from compensation of the local thermosteric and halosteric effects in the Atlantic and Pacific. In previous research, elimination of the Atlantic–Pacific DSL asymmetry occurs only when a large external freshwater flux is imposed in the North Atlantic (Levermann et al. 2005; Yin et al. 2010; Fig. 5a). So we conclude that a significant change in the Atlantic–Pacific DSL asymmetry is unlikely but possible during this century, given the large uncertainty associated with the melting of the Greenland ice sheet, which could be a large external freshwater forcing.

In a warming climate, the oceans with greater depth can absorb more heat and therefore exhibit a larger thermal expansion. The corresponding sharp gradient of the steric SLR across a shelf break drives more waters toward the shelf, thereby significantly increasing the ocean bottom pressure near the coasts. This bottom pressure signature is striking and robust across the AR4 model projections (e.g., Figs. 7 and 12; Landerer et al. 2007), suggesting that it can be used for ocean warming detection. These modeling results provide guidance for the design of observation systems and the use of observation data. Compared to the direct measurement of ocean temperature by a global network that employs thousands of floats, a small network to measure the ocean bottom pressure on the shelf and near coasts may be efficient and consistent in detecting ocean change signals. In addition, with new data such as those from the Gravity Recovery and Climate Experiment starting to emerge, observational studies on ocean mass redistribution and bottom pressure changes can assist us in providing fingerprints of global warming signals and the evaluation of model projections.

The dynamic SLR dipole in the North Atlantic is thermohaline induced and mainly attributable to weakening of the AMOC. This weakening causes a local steric SLR east of North America, so that more waters are redistributed toward the shelf, leading to a rapid dynamic SLR along northeastern North America. It is found that this dynamic SLR can be expressed as a linear function of the weakening of the AMOC (Levermann et al. 2005; Yin et al. 2009; Bingham and Hughes 2009). However, the slope of roughly 2 cm Sv$^{-1}$ in transient climate projections (Yin et al. 2009) and climate variability simulations (Bingham and Hughes 2009) is significantly smaller than that of 5 cm Sv$^{-1}$ obtained by Levermann et al. (2005). In addition to model formulation, this difference is also caused by the differing experimental designs and forcings. The large rise in Levermann et al. is induced by large external freshwater forcing and is partially attributed to the basin-wide dynamic SLR in the Atlantic and the elimination of the Atlantic–Pacific DSL asymmetry. However, this elimination does not occur in the twenty-first century projections analyzed here. So the relationship of 2 cm Sv$^{-1}$ is likely more realistic to describe the AMOC and DSL anomalies during this century. Consequently, the SLR along the North America coast should be incorporated into an AMOC monitoring and early warning system.

The dipole pattern in the North Pacific has a distinctly different nature. The dipole is mainly wind induced and reflects the poleward expansion of the subtropical gyre. Due to ocean stratification, the steric effect is strong in the upper 1000 m. The strong steric SLR can steepen

\[ \text{FIG. 18. Zonal mean of the wind stress, DSL, and isopycnals in the Southern Ocean; (a) the wind stress and DSL and (b) the isopycnal surfaces (}s_2). \text{Black, red, and green lines in (b), respectively, indicate the periods of 1981–2000, 2091–2100, and 2291–2300. The results are from the A1B scenario run of CM2.1.} \]
the DSL gradient associated with the subtropical gyre, even when the gyre circulation weakens. The rapid dynamic SLR in the subtropical gyre does not impact the coastal regions of East Asia directly, in contrast to that in the northeastern North America.

The Southern Ocean is an eddy-filled ocean and mesoscale eddies play an important role in the momentum, heat, and freshwater budgets. Due to the coarse resolution in the ocean models considered here, mesoscale eddies are not resolved and their effects on the large-scale flows are parameterized. The absence of explicit eddies is demonstrated by the weak DSL variability in the ACC and western boundary current regions (Fig. 13). Recent studies suggest that in an eddy-permitting coupled model, the DSL, ACC, and overturning circulation in the Southern Ocean become relatively insensitive in the twenty-first century projection (Farneti et al. 2010). This is because stronger wind stress intensifies eddy activity, which tends to offset the changes induced by the mean flow (Hallberg and Gnanadesikan 2006). This “eddy saturation” regime is not captured by the coarse-resolution model such as CM2.1 and many other AR4 models. Given the importance of the Southern Ocean in the global climate system, the role of the mesoscale eddy in the SLR projections needs to be further explored.

The world’s deltas and ocean islands are among the most vulnerable regions to SLR. The former suffers from rapid subsidence (Syvitski et al. 2009), while the latter is low lying and has very limited capacity to adapt to SLR. We found that the dynamic SLRs at these regions are generally below 0.1 m over the twenty-first century, so that the total SLRs follow the global mean (Fig. 15). The sea level at Tuvalu shows some remarkable interannual variability, which is probably associated with ENSO.

In summary, we performed a comprehensive and detailed analysis on the projected spatial variability of SLR induced by ocean dynamics during the twenty-first century, and explored the underlying mechanisms for the robust features. The results facilitate methods of interpreting differing regional SLRs seen in observations and, therefore, provide important information for policy makers and coastal communities.

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