# The solid Earth's influence on sea level

INVITED REVIEW

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## ABSTRACT

**Because it lies at the intersection of Earth's solid, liquid, and gaseous components, sea level links the dynamics of the fluid part of the planet with those of the solid part of the planet. Here, I review the past quarter century of sea-level research and show that the solid components of Earth exert a controlling influence on the amplitudes and patterns of sea-level change across time scales ranging from years to billions of years. On the shortest time scales ($10^0$–$10^2$ yr), elastic deformation causes the ground surface to uplift instantaneously near deglaciating areas while the sea surface depresses due to diminished gravitational attraction. This produces spatial variations in rates of relative sea-level change (measured relative to the ground surface), with amplitudes of several millimeters per year. These sea-level "fingerprints" are characteristic of (and may help identify) the deglaciation source, and they can have significant societal importance because they will control rates of coastal inundation in the coming century. On time scales of $10^3$–$10^5$ yr, the solid Earth's time-dependent viscous response to deglaciation also produces spatially varying patterns of relative sea-level change, with centimeters-per-year amplitude, that depend on the time-history of deglaciation. These variations, on average, cause net seafloor subsidence and therefore global sea-level drop. On time scales of $10^6$–$10^8$ yr, convection of Earth's mantle also supports long-wavelength topographic relief that changes as continents migrate and mantle flow patterns evolve. This changing "dynamic topography" causes meters per millions of years of relative sea-level change, even along seemingly "stable" continental margins, which affects all stratigraphic records of Phanerozoic sea level. Nevertheless, several such records indicate sea-level drop of ~230 m since a mid-Cretaceous highstand, when continental transgressions were occurring worldwide. This global drop results from several factors that combine to expand the "container" volume of the ocean basins. Most importantly, ridge volume decrease since the mid-Cretaceous, caused by an ~50% slowdown in seafloor spreading rate documented by tectonic reconstructions, explains ~250 m of sea-level fall. These tectonic changes have been accompanied by a decline in the volume of volcanic edifices on Pacific seafloor, continental convergence above the former Tethys Ocean, and the onset of glaciation, which dropped sea level by ~40, ~20, and ~60 m, respectively. These drops were approximately offset by an increase in the volume of Atlantic sediments and net seafloor uplift by dynamic topography, which each elevated sea level by ~60 m. Across supercontinental cycles, expected variations in ridge volume, dynamic topography, and continental compression together roughly explain observed sea-level variations throughout Pangean assembly and dispersal. On the longest time scales ($10^9$ yr), sea level may change as ocean water is exchanged with reservoirs stored by hydrous minerals within the mantle interior. Mantle cooling during the past few billion years may have accelerated drainage down subduction zones and decreased degassing at mid-ocean ridges, causing enough sea-level drop to impact the Phanerozoic sea-level budget. For all time scales, future advances in the study of sea-level change will result from improved observations of lateral variations in sea-level change, and a better understanding of the solid Earth deformations that cause them.**

## INTRODUCTION

Sea level has remained a fundamental boundary on Earth's surface for as long as oceans have existed on our planet. Most of Earth's species, including humans, have adapted to life on one side of this barrier or the other; geological processes such as erosion and sedimentation operate in completely different ways across sea level. The fundamental nature of the sea surface implies that changes in sea level drive first-order shifts in the landscape of our planet and dictate changes in the ranges of its biological inhabitants. For example, the ongoing 1–2 mm/yr rise of sea level during the past century (e.g., Church et al., 2004; Douglas, 1991, 1992; Ray and Douglas, 2011) and its recent acceleration (e.g., Church and White, 2006, 2011) affect human activities (e.g., Houghton et al., 2010; Pilkey and Cooper, 2004) and will do so even more dramatically if sea level rises another ~0.8 m by 2100, as forecast (Meehl et al., 2007). Even more drastic scenarios that predict 1.3–2.0 m of global rise in the next century (Grinsted et al., 2009; Jevrejeva et al., 2010; Pfeffer et al., 2008; Rahmstorf, 2007; Vermeer and Rahmstorf, 2009) make sea level one of the most tangible and dramatic consequences of future climate change.

Although the sea surface divides the disparate subaerial and submarine realms, sea level itself is sensitive to a variety of external forcing factors that act on the whole Earth system. From above, warming of Earth's surface climate induces sea-level rise by melting landed ice and by thermally expanding seawater (e.g., Bindoff et al., 2007; Cazenave and Nerem, 2004; Hock et al., 2009; Lombard et al., 2005; Meier et al., 2007; Nerem et al., 2006; Rignot et al., 2011; Wigley and Raper, 1987). From below, the solid Earth drives sea-level change by altering the shape and volume of the basins that contain the oceans, and by deflecting the ocean surface in response to changes in Earth's gravitational field (Fig. 1). On relatively short time scales (instantaneous to $10^5$ yr), these boundary deflections occur as a rheological (elastic or viscous) response of the solid Earth to the redistribution of hydrological loads on Earth's surface (Figs. 1A and 1B). On longer time scales ($10^6$ yr and longer), these deflections are associated with surface tectonics and the convective dynamics

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*GSA Bulletin*; July/August 2013; v. 125; no. 7/8; p. 1027–1052; doi: 10.1130/B30764.1; 11 figures.

of the mantle (Fig. 1C). Sea-level change, which has been documented for all of these time scales (Fig. 2), thus links a wide variety of Earth processes, ranging from the climatic and hydrological systems that operate above Earth's surface to the plate-tectonic and mantle convective systems operating below it.

The past 25 yr have seen a growing recognition of the fact that observations of sea level (both geological and historical) are made relative to the solid Earth, which moves vertically on all time scales (Fig. 1). Thus, rates of observed sea-level change may vary between coastal locations, depending on vertical ground motions. Such local observations of sea level, whether obtained from tide gauges, land records, beach terraces, or stratigraphy, measure relative sea level (sometimes referred to as regional sea level), defined as the local elevation of the sea surface measured relative to the solid surface of Earth. Here, I will be careful to

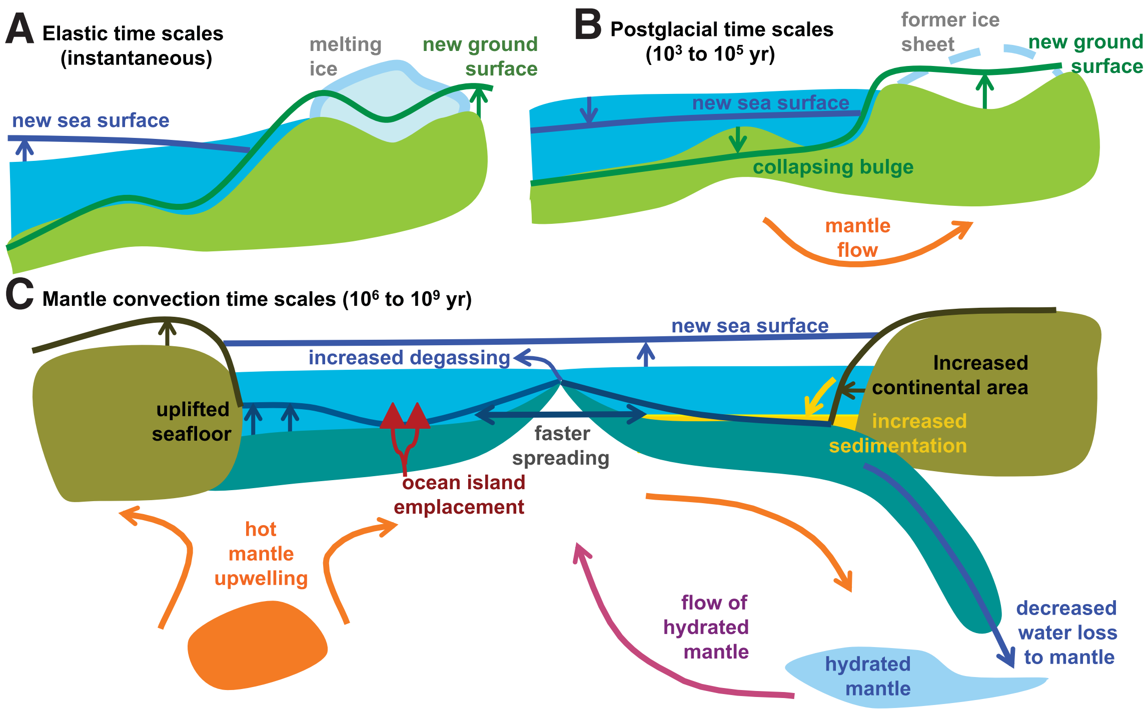


**Figure 1. Mechanisms by which the solid Earth affects sea level, separated by operative time scale. Shown are the solid Earth and oceans (filled areas) and their surfaces after an applied change to the system (lines). (A) On the shortest time scales, the solid Earth deforms elastically in response to an imposed load (Farrell, 1972). Here, melting of an ice sheet uplifts the ground near areas of mass loss and depresses the ocean basins, which gain mass. The sea surface drops near the mass loss because the diminished ice sheet gravitationally attracts less seawater (i.e., the geoid depresses). Relative to the ground surface, sea-level drops near melting ice but rises faster than average over the rest of the ocean (Clark and Primus, 1987; Farrell and Clark, 1976). Observations of spatial variations in sea-level change can be used to "fingerprint" the source of the change (e.g., Blewitt and Clarke, 2003; Conrad and Hager, 1997; Mitrovica et al., 2001; Tamisiea et al., 2001). (B) Following glacial unloading, Earth deforms viscously on time scales of $10^3$–$10^5$ yr as the mantle flows back into the depressed region. This uplifts the region near the former ice sheet (locally causing relative sea-level drop) and depresses the surrounding peripheral forebulge (Clark et al., 1978; Davis and Mitrovica, 1996; Farrell and Clark, 1976). If the forebulge collapses beneath the sea surface, the added basin volume causes far-field (eustatic) sea-level drop, which Mitrovica and Peltier (1991) termed "equatorial ocean siphoning." Ocean loading causes a similar viscous response along coastlines that also drops eustatic sea level (Mitrovica and Milne, 2002). (C) On time scales of $10^6$ yr and longer, solid Earth processes associated with plate tectonics and mantle dynamics dominate sea-level change (Harrison, 1990; Miller et al., 2005). Shown here are the major processes that can elevate global average (eustatic) sea level (and depress it when acting oppositely). Global sea level rises when the "container" volume of the ocean basins decreases due to: ridge expansion associated with faster spreading (Cogné et al., 2006; Hays and Pitman, 1973; Kominz, 1984; Müller et al., 2008b; Pitman, 1978; Xu et al., 2006), expansion of continental area (Harrison, 1990; Kirschner et al., 2010), growth of submarine volumes of sediment cover or volcanic debris (Harrison, 1999; Müller et al., 2008b), and net dynamic uplift of the seafloor by mantle flow (Conrad and Husson, 2009). This last process (dynamic topography) may also induce lateral variations in sea-level change by deflecting the ground surface locally (e.g., Moucha et al., 2008). Sea level also rises if water exchange with the deep mantle becomes imbalanced (either via increased outgassing or diminished loss via subduction) (Crowley et al., 2011; Korenaga, 2011; McGovern and Schubert, 1989; Sandu et al., 2011).**

distinguish relative sea level from eustatic sea level (also referred to as global sea level), which is defined as the globally averaged level of the sea surface measured relative to a fixed reference such as Earth's center of mass (e.g., Eriksson, 1999). Changes in eustatic sea level result from processes (Fig. 1) that change either the volume of seawater in the oceans or the containing capacity of the ocean basins (e.g., Worsley et al., 1984). Water volume changes generally result from climatological processes but may involve water exchange with the solid Earth on long time scales (see section on "Ocean-Mantle Water Exchange and Sea-Level Change over Earth History"). Basin volume changes always result from solid Earth processes (rock deformation, tectonics, volcanism, sedimentation, and mantle convection) and can occur on all time scales. However, it is important to remember that all observations of relative sea level are sensitive to both changes in eustatic sea level as well as to local uplift or subsidence at the measurement location. As a result, an understanding of solid Earth deformation is necessary to understand how observational constraints on relative sea level (Fig. 2) are related to constraints on the geological and climatological processes (Fig. 1) that cause both eustatic changes in sea level and vertical motion of Earth's solid surface.

As an example of the interplay between the solid Earth and climatologically induced changes in sea level, consider the melting of water mass from landed ice into the oceans. This load redistribution induces an instantaneous elastic rebound of the bedrock beneath the deglaciated area (Fig. 1A) that is later followed by a viscous deformation of the mantle, which also deflects the land surface (Fig. 1B) (Chappell, 1974). Accompanying both deformations is a deflection of the geoid (Farrell and Clark, 1976), which is the gravitational equipotential surface upon which the sea surface rests. The resulting time-dependent deflections of the solid and sea surfaces, which occur everywhere but are largest in the vicinity of deglaciation, induce spatial variations in relative sea level (Clark et al., 1978). When averaged over the ocean area, these changes in relative sea level are usually nonzero, which means that the solid Earth's response to deglaciation can also induce eustatic change (e.g., Walcott, 1972). (Technically, the sea surface [geoid] must shift to another, parallel, equipotential surface to preserve water mass.) Although the amplitude of eustatic change associated with solid Earth deformation is only a fraction of that associated with the initial melting event, this solid Earth–induced change can continue for tens of thousands of years after melting due to the solid Earth's viscous response (Mitrovica and Milne, 2002; Mitrovica and Peltier, 1991). Furthermore, relative sea-level change in the vicinity of deglaciation can be larger in amplitude, and of opposite sign, compared to the eustatic change associated with the initial melting (e.g., Clark and Primus, 1987). Thus, the solid Earth exerts a first-order and time-dependent influence on both the measurement and interpretation of eustatic change.

Although much of the theoretical background needed to compute spatial variations in relative sea level was developed in the 1970s (Clark et al., 1978; Farrell, 1972; Farrell and Clark, 1976), only in the past 25 yr has this theory been used to help understand and predict patterns of sea-level change (e.g., Clark and Primus, 1987; Nakiboglu and Lambeck, 1991). In fact, several recent studies have computed the spatial variations in relative sea level associated with various continental sources of new ocean water (e.g., Conrad and Hager, 1997; Fiedler and Conrad, 2010; Riva et al., 2010; Tamisiea et al., 2001). In principle, these "fingerprints" can be used to identify the sources of recent sea-level change (Blewitt and Clarke, 2003; Mitrovica et al., 2001). As sea-level rise continues in the coming decades, we will see an enhanced effort to exploit this link between climate and the solid Earth to understand observed patterns of sea-level change (Douglas, 2008; Mitrovica et al., 2011), and to predict future patterns (Gomez et al., 2010; Mitrovica et al., 2009; Willis and Church, 2012).

Interpretations of sea-level change over geologic time scales have also been modified by

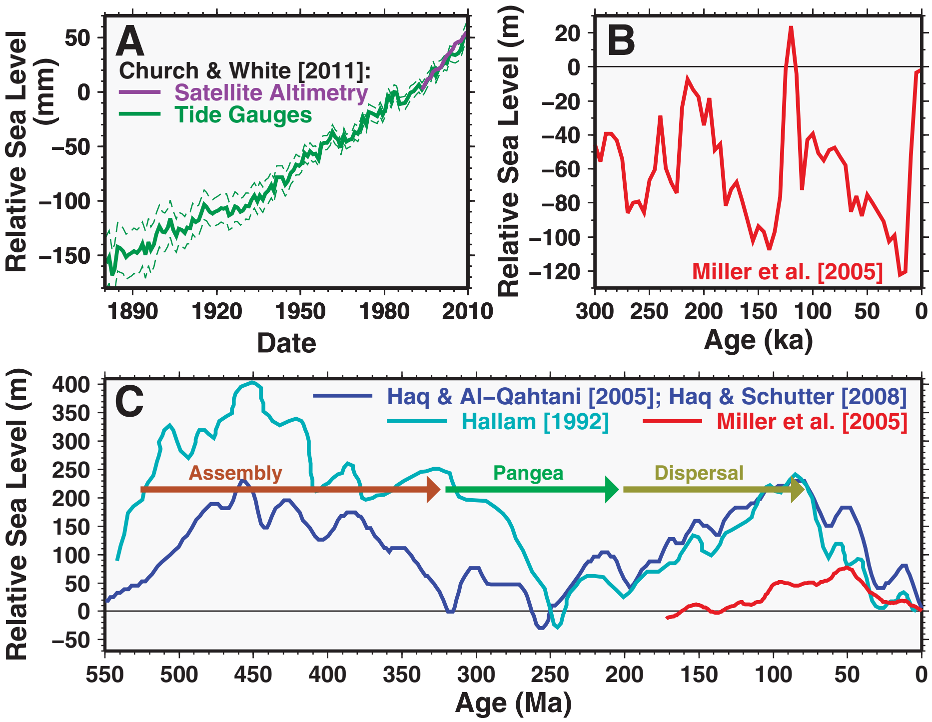


**Figure 2. Observations of sea-level change, separated by relevant time scale. (A) A reconstruction of global average sea-level change by Church and White (2011), based on tide gauge (green) and satellite altimetry (purple) data. This reconstruction suggests that sea level rose 1.7 ± 0.2 mm/yr from 1900 to 2009 and accelerated to 3.2 ± 0.4 mm/yr from 1993 to 2009. (B) Late Quaternary sea-level change is primarily associated with climate-induced growth and shrinkage of continental ice sheets, and the solid Earth's viscoelastic response. Shown are stratigraphic constraints on relative sea level in New Jersey (Miller et al., 2005). The peak at ca. 125 ka represents the Last Interglacial, a warm period corresponding to the last time sea levels were at least as high as they are today (McCulloch and Esat, 2000). (C) Phanerozoic sea level periodically transgressed continental boundaries (Bond, 1978). Shown are stratigraphic constraints on relative sea level for Arabia since the Mesozoic (Haq and Al-Qahtani, 2005) and New Jersey since the Cretaceous (Miller et al., 2005), and a synthesis of Phanerozoic (Hallam, 1992) and Paleozoic records (Haq and Schutter, 2008). Phanerozoic sea-level trends correlate with supercontinent tectonics, denoted by arrows showing the approximate timing of Pangean assembly, duration, and dispersal (after Cogné and Humler, 2008; Li and Zhong, 2009).**

our understanding of solid Earth dynamics during the past quarter century. In particular, the idea that mantle convection can support up to ~1 km of long-wavelength topographic relief was proposed ~25 yr ago (Hager, 1984; Hager et al., 1985; Mitrovica et al., 1989). The fact that this "dynamic topography" may evolve over time scales of millions of years and longer (e.g., Gurnis, 1990c, 1993; Lithgow-Bertelloni and Gurnis, 1997) has led to the realization that stratigraphic observations of sea level (e.g., Hallam, 1992; Hallam and Cohen, 1989; Haq and Al-Qahtani, 2005; Haq et al., 1987; Haq and Schutter, 2008; Miller et al., 2005) simultaneously reflect not only eustatic change, but also local geologic uplift and subsidence patterns at the observation locations (Fig. 1C). Dynamic topography thus provides geodynamicists with a new sea-level constraint on deep Earth dynamics (e.g., Conrad and Gurnis, 2003; Conrad et al., 2004; Gurnis, 1990a; Gurnis et al., 2000; Liu et al., 2008; Spasojevic et al., 2009), but it has also required records of sea-level change over geologic time scales (Fig. 2C) to be reinterpreted as partly due to local vertical motions (Gurnis, 1993; Gurnis et al., 1998; Müller et al., 2008b; Spasojevic et al., 2008). The stratigraphic constraint on sea level is additionally complicated by the fact that vertical motions for past continental shorelines are difficult to constrain and yet may be locally larger than global eustatic change (Braun, 2010). In fact, the interconnectedness of eustatic and relative sea-level change has recently led some to question whether eustasy is a useful concept (Mitrovica, 2009; Moucha et al., 2008).

Despite the realization that solid Earth deformation significantly complicates measurements of eustatic sea-level change (Jones et al., 2012; Lovell, 2010; Ruban et al., 2010b, 2012), the past quarter century has seen significantly improved predictions of eustatic change based on new tectonic constraints on the time-evolution of the ocean basin "container" volume, which controls eustatic change over geologic time scales (e.g., Harrison, 1990; Miller et al., 2005). New geophysical databases, such as those quantifying present-day patterns of seafloor age (e.g., Müller et al., 2008a), bathymetry (e.g., Smith and Sandwell, 1997), sediment cover (e.g., Divins, 2003), and seamount locations (e.g., Kim and Wessel, 2011), have allowed us to better understand a variety of seafloor processes and quantify their temporal impact on sea level. New tectonic reconstructions of the seafloor, based on an accumulated plethora of geological constraints, allow us to quantify the influence of Cretaceous and Cenozoic seafloor bathymetry changes on eustatic sea level (Müller et al., 2008b; Xu et al., 2006). Numerical models of global mantle convection allow us to estimate how changing dynamic topography of the seafloor affects eustatic sea level (Conrad and Husson, 2009; Gurnis, 1993; Moucha et al., 2008; Spasojevic and Gurnis, 2012), and they help us to relate regional observations of sea level to eustatic change (Müller et al., 2008b; Spasojevic et al., 2008, 2009). Most recently, new ideas about hydrological transfer between Earth's surface oceans and hydrated minerals in the mantle interior have allowed us to explore ideas about sea-level change due to water exchange with hydrated minerals in the mantle (Crowley et al., 2011; Korenaga, 2011; Sandu et al., 2011).

In what follows, I review our current understanding of the dynamic interaction between the solid Earth and sea level across time scales ranging from the life span of a human to that of a supercontinent (Fig. 1). In particular, I focus on the past quarter century of scientific progress, during which we have (1) recognized and quantified the significant contribution of vertical ground motions to geologic and historical observations of relative sea level, and (2) placed new constraints on rates of eustatic sea-level change caused by a variety of solid Earth processes. The next quarter century should see improved constraints on the time-dependent history of solid Earth deformation (Braun, 2010), which will allow us to relate observations of relative sea-level change more directly to improved constraints on eustatic sea-level change. Prediction of spatial variations will become even more important in the next quarter century as accelerated climate change induces greater hydrological mass redistributions, and therefore greater solid Earth deformations (Mitrovica et al., 2009; Slangen et al., 2012).

## ELASTIC DEFORMATION AND ANTHROPOGENIC SEA-LEVEL CHANGE

More than half of the 2–3 mm/yr of observed sea-level rise (Fig. 2A) during the past 10–100 yr (Church and White, 2011), and perhaps longer (Jevrejeva et al., 2008), has been attributed to the melting of continental ice into the oceans (Leuliette and Miller, 2009; Miller and Douglas, 2004). This redistribution of hydrological loads on Earth's surface, which is accelerating (Church et al., 2011; Meier et al., 2007), interacts with the solid Earth on both global and regional scales, and it does so in ways that are geodetically detectable. Exploitation of this interaction has allowed us to develop new constraints on patterns and rates of mass transfer between the cryosphere and oceans during the past quarter century.

Observed changes to Earth's oblate (flattened) shape and to its center of mass have been directly linked to redistributions of hydrological masses on Earth's surface during the past few decades. Specifically, measurements of Earth's oblateness (specifically, its $J_2$ gravitational parameter) during the past several decades had shown a steady ~30 yr decrease until the mid-1990s (Cox and Chao, 2002). This decrease in oblateness was attributed to viscous adjustments of the solid Earth to past deglaciations, which move solid Earth mass toward the polar regions (Mitrovica and Peltier, 1993). This long-term decline reversed itself in the mid-1990s (Roy and Peltier, 2011), presumably because melting of the Greenland and Antarctic ice sheets began to move mass away from the poles at rates fast enough to increase Earth's oblateness (Nerem and Wahr, 2011). An imbalance between the rates of melting in the Northern and Southern Hemispheres causes a net southward or northward motion of water mass, which must be compensated by an equivalent translation of the solid Earth in the opposite direction (Conrad and Hager, 1997). Such translation of the solid Earth relative to the center of mass of the entire Earth system can in principle be detected geodetically, and such measurements could be used to constrain climate change–induced mass movements on Earth's surface (Métivier et al., 2010; Rietbroek et al., 2012). Indeed, accounting for geocenter motion helps to reconcile hemispheric variations in the tide-gauge record of sea-level rise over the past century (Collilieux and Wöppelmann, 2011; Wöppelmann et al., 2009).

Redistribution of water mass on Earth's surface additionally drives an instantaneous elastic deformation within the solid Earth and deflects the gravitational equipotential (geoid) surface that defines sea level (Farrell, 1972). Within ~30° of the deglaciating area, Earth's surface uplifts, and the diminished gravitational attraction of seawater to the remaining ice causes the sea surface to fall (Fig. 1A). Taken together, these two effects cause a contemporaneous *fall* in relative sea level near the deglaciating area (Farrell and Clark, 1976). This local effect is balanced by enhanced relative sea-level *rise* in portions of the ocean away from the deglaciating area, where the geoid elevates and the seafloor depresses due to the addition of water mass to the oceans (Fig. 1A). Because each continental source of mass loss induces a different geographic pattern of sea-level response, several authors have proposed that observations of spatial variations in sea-level change (e.g., as measured by tide gauges) could be used as "fingerprints" to constrain the continental locations of mass wastage (Blewitt and Clarke, 2003; Conrad and Hager, 1997; Mitrovica et al., 2001; Nakiboglu and Lambeck, 1991; Tamisiea et al., 2001).

To estimate the sea-level "fingerprint" of present-day deglaciation, I computed the deflections of the solid Earth and geoid surfaces based on the most recent constraints on mass loss from glaciers and ice sheets during the past decade. These estimates, which are based on satellite observations of temporal variations of the geoid (GRACE satellite mission), indicate that the Greenland and Antarctic ice sheets, and mountain glaciers, are currently losing water to the oceans at rates of ~222, 165, and 151 Gt/yr, respectively, which together corresponds to an average eustatic sea-level rise of 1.48 mm/yr (Jacob et al., 2012; Rignot et al., 2011; Velicogna, 2009). I discretized this mass redistribution onto a 0.5° × 0.5° grid by distributing the mass loss for each glacial system of Jacob et al. (2012) into negative point loads for continental grid points within 100 km of each glacial system, and distributing corresponding positive point loads over all ocean areas. By convolving these loads with Greens functions for the elastic response of the solid Earth and the geoid defined by Farrell (1972), and iterating the ocean load as deflections of the land and sea surfaces redistribute seawater (following the procedure of Conrad and Hager, 1997), I computed rates of relative sea-level change associated with the solid Earth's instantaneous elastic response (Fig. 3A).

The elastic response of the solid Earth to presently occurring deglaciation (Fig. 3A) predicts negative sea-level change across most of the North American and Eurasian coastlines, and around Antarctica and southern South America. This sea-level drop is caused by mass loss on nearby continents, which uplifts the land surface and decreases the gravitational attraction of seawater. Close to major deglaciation sources (Antarctica and especially Greenland), the rate of sea-level drop induced by solid Earth deflections is faster than the average global rise rate (1.48 mm/yr). These areas thus experience net relative sea-level drop, which may stabilize some continental ice sources (Gomez et al., 2012). By contrast, equatorial regions, particularly in the center of large ocean basins, experience rates of relative sea-level rise that are up to ~20%–30% faster than the global average because these areas gain seawater mass, and thus experience sea-level behavior opposite to regions of mass wastage. Although the average relative sea level induced by these solid Earth and geoid deflections is (necessarily) zero, the sea surface actually drops by 0.09 mm/yr (6% of the global average rise rate) relative to Earth's center of mass because the additional seawater in the oceans compresses rocks beneath the seafloor. Because of solid Earth and geoid changes, this sea-surface height, which is measured by satellites, differs from the average sea-surface height measured relative to the land surface (Conrad and Hager, 1997). This difference may complicate efforts to compare sea-level observations made by satellite altimetry to those made using tide gauges (e.g., Church and White, 2011).

Observations of spatial variations in rates of sea-level rise associated with the response of the solid Earth to deglaciation (e.g., Fig. 3A) can potentially place constraints on the continental sources of mass loss (e.g., Blewitt and Clarke, 2003; Conrad and Hager, 1997; Tamisiea et al., 2001). For example, Mitrovica et al. (2001) suggested that anomalously slow European rates

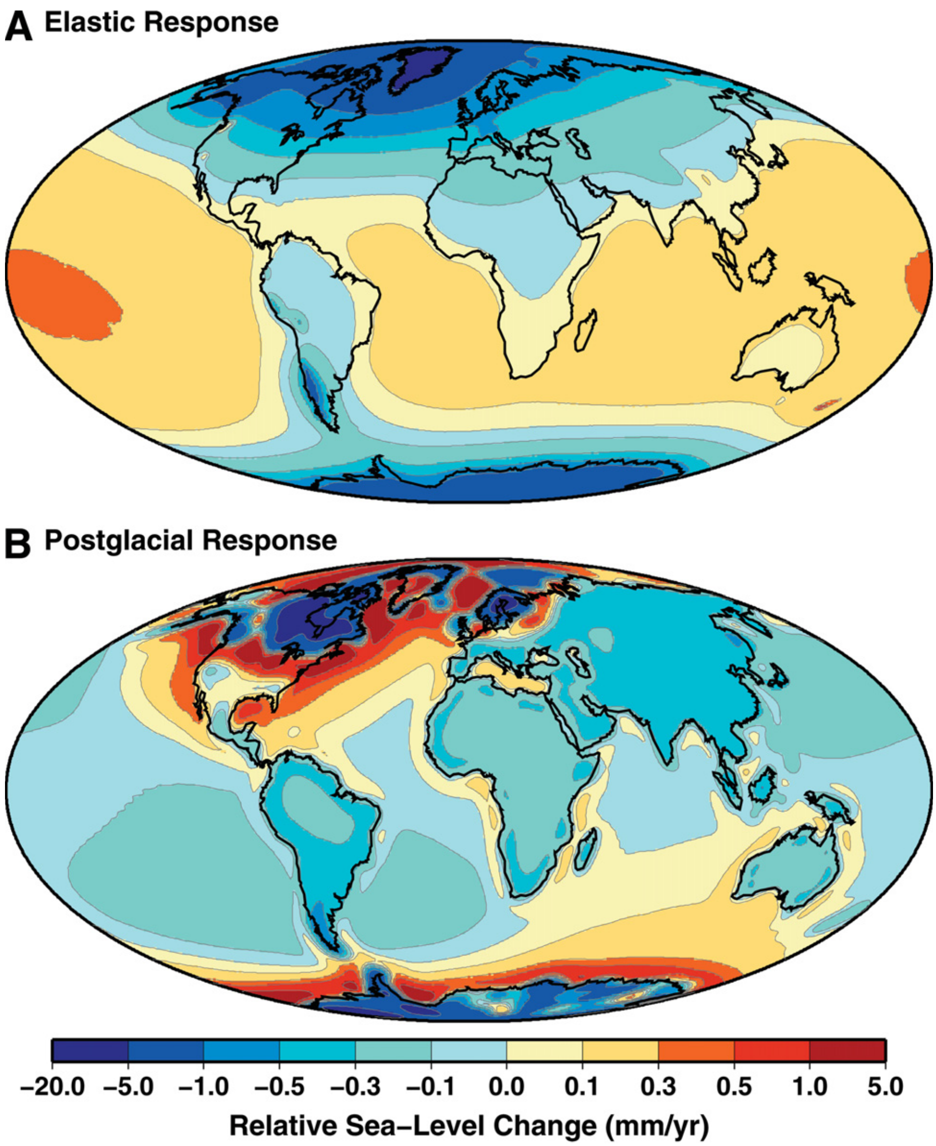


**Figure 3. Estimated rates of relative sea-level change for the present day due to (A) elastic deformation of the solid Earth caused by present-day deglaciation of glaciers and ice sheets and (B) viscous deformation of the solid Earth in response to past deglaciation. The elastic response (A) was computed following Conrad and Hager (1997) using the deglaciation rates given by Jacob et al. (2012), which add 1.48 mm/yr to sea level. The postglacial response (B) was computed by Paulson et al. (2007a) using the ICE-5G model (Peltier, 2004) of past degaciation. Relative sea level is computed everywhere (including land areas) as the difference between computed geoid and ground surface motions, where the geoid is offset to preserve total water mass. Eustatic sea level (relative to Earth's center of mass, and not including water transfer to the ocean) drops by 0.09 and 0.23 mm/yr in A and B, respectively, due to net subsidence of the seafloor.**

of relative sea-level rise imply that mass loss in Greenland caused a large component of twentieth-century sea-level rise. Other efforts to use sea-level "fingerprints" to constrain continental mass loss have generally proven less successful (e.g., Douglas, 2008; Plag and Jüttner, 2001; Plag, 2006), likely for a variety of reasons. Primarily, interpretation of sea-level observations (both satellite and tide gauge) is complicated by significant spatial and temporal variability in sea level associated with climatological influences such as seawater temperature and salinity (e.g., Llovel et al., 2011b), wind patterns (e.g., Bromirski et al., 2011; Merrifield and Maltrud, 2011), and ocean currents (e.g., Lorbacher et al., 2012; Piecuch and Ponte, 2011; Sallenger et al., 2012). In some areas, relative sea level may be influenced, or even dominated, by vertical ground motion associated with earthquake-cycle tectonic deformation (e.g., Larsen et al., 2003), volcanic loading (e.g., Moore, 1987), or fluid extraction from the subsurface (e.g., Ericson et al., 2006); not all of these effects are sufficiently well constrained to confidently correct records of relative sea-level change (Douglas, 2008; Nerem and Mitchum, 2002). Additionally, sea-level fingerprints of deglaciation are sensitive to the spatial distribution of mass loss within the deglaciating region (Mitrovica et al., 2011), which may be uncertain. This is particularly true close to regions of deglaciation, where elastic uplift rates can be significant; geodetically determined rock uplift rates of 5–15 mm/yr have been attributed to deglaciation in Alaska (Sauber et al., 2000) and Greenland (Khan et al., 2007), where even seasonal variations in the ice load have been detected (Bevis et al., 2012).

The interpretation of spatial variations in sea level is additionally complicated by other hydrologic mass redistributions that induce their own sea-level "fingerprints" with magnitudes comparable to those caused by deglaciation. For example, Fiedler and Conrad (2010) found that water impoundment in artificial reservoirs (which may have caused ~0.55 mm/yr of sea-level drop in the past 50 years; Chao et al., 2008) likely caused sea level to rise ~0.2 mm/yr faster on coastlines than in the global ocean (because coastlines are closer to the water mass added to these reservoirs). An opposite behavior may be expected from the ~0.5–0.8 mm/yr of sea-level rise thought to be associated with groundwater depletion (Pokhrel et al., 2012; Wada et al., 2012). Temporal variations in river basin water storage (Llovel et al., 2011a; Wang et al., 2013) may also complicate patterns of sea-level change. Although constraints on the temporal and spatial development of these large and widely dispersed hydrologic loads are improving (e.g., Lettenmaier and Milly, 2009; Schmidt et al., 2008), continued uncertainty makes computation of their sea-level fingerprints difficult and corrupts the sea-level signal associated with deglaciation.

Thus, the sea-level fingerprint of deglaciation (e.g., Fig. 3A) is difficult to observe because it is overwhelmed by climatologically induced sea-level variability and because the driving patterns of hydrologic load redistribution are poorly constrained. For a given melting history, however, the sea-level response to hydrologic mass movements can be computed robustly (Mitrovica et al., 2011), although such calculations are also sensitive to several complications. These include changes to Earth's rotation axis (which influence sea level by deflecting Earth's rotational bulge) (Milne and Mitrovica, 1998; Mitrovica et al., 2005), seawater inundation of continental areas (Gomez et al., 2010), spatial heterogeneity in Earth's elastic properties (Mitrovica et al., 2011), and the solid Earth's response to past climate change (Tamisiea and Mitrovica, 2011). Because all of these factors can be addressed (Mitrovica et al., 2011), the solid Earth's influence on sea level can be computed accurately if the hydrologic load is known. This is fortunate because the resulting spatial variations in sea-level change (e.g., Fig. 3A) are large enough to significantly impact human society in the coming century (Gomez et al., 2010; Mitrovica et al., 2009; Sallenger et al., 2012), and thus they need to be considered when planning for future sea-level rise (Willis and Church, 2012).

## VISCOUS DEFORMATION AND POSTGLACIAL SEA-LEVEL CHANGE

On time scales of $10^3$–$10^5$ yr, the solid Earth's viscous response to mass redistributions (Haskell, 1935) affects relative sea level by deflecting the seafloor (Fig. 1B). In particular, deglaciated areas rebound as viscous mantle flows toward them from collapsing peripheral "bulges" that surround them (Carlson and Clark, 2012; Chappell, 1974; Clark et al., 1978; Farrell and Clark, 1976). For example, late Pleistocene deglaciation in North America has caused the land surface to uplift near the center of the former Laurentide ice sheet, inducing sea-level fall near Hudson Bay (Mitrovica, 1996). By contrast, as the associated peripheral bulge along the U.S. east coast collapses, this region experiences a rise in relative sea level (Davis and Mitrovica, 1996). Such time-dependent deflections can be computed accurately using numerical models of viscous mantle deformation if effects such as the changing geometry of coastlines (Milne et al., 1999) and changes to Earth's rotation (Mitrovica et al., 2005) are included. Because these calculations depend on deglaciation history and mantle viscosity structure, observations of postglacial sea-level change provide a primary constraint on both (e.g., Tushingham and Peltier, 1991), although tradeoffs often accompany this constraint (Mitrovica et al., 1993; Paulson et al., 2007b).

Recent observations of ground motion from global positioning system (GPS) data (Sella et al., 2007) and geoid motion from satellite altimetry (GRACE mission) have placed new constraints on both deglaciation (Peltier, 2004) and mass displacements associated with present-day rebound (Paulson et al., 2007a). These new constraints, used in conjunction with geologic observations of sea-level change in various locations (Engelhart et al., 2011), inform models of viscous mantle deformation (Paulson et al., 2007a). Such models predict patterns of relative sea-level change (Fig. 3B) with amplitudes that are comparable to those associated with Earth's elastic response to present-day deglaciation (Fig. 3A). The patterns of change associated with these two modes of deformation are distinct, but their uncertainty and simultaneous occurrence heighten the difficulty of fully explaining spatial patterns of sea-level change (Ostanciaux et al., 2012). Additionally, patterns of postglacial sea-level change (Fig. 3B) are influenced by lateral variations in mantle viscosity (Zhong et al., 2003) and deformation-induced changes to Earth's rotation axis and associated rotational bulge (Mitrovica et al., 2005), both of which may complicate the use of such models to remove the postglacial component from present-day sea-level observations (Chambers et al., 2010; Kendall et al., 2006; King et al., 2012; Métivier et al., 2012; Spada et al., 2012; Whitehouse et al., 2012).

Paulson et al.'s (2007a) prediction of the sea-level response to viscous deformation (Fig. 3B) also demonstrates the response of the global ocean to postglacial rebound occurring in the polar regions. In particular, as the peripheral bulges, which are most often located in oceanic areas, collapse, the ocean basin volume expands, causing sea level to fall globally via a process that Mitrovica and Peltier (1991) termed "equatorial ocean siphoning." For Paulson et al.'s (2007a) calculation, the average seafloor falls by 0.23 mm/yr, primarily due to rapid subsidence (>0.5 mm/yr) around North America, Europe, and Antarctica (Fig. 3B). This global eustatic sea-level change is observed as a drop in sea level for (primarily equatorial) ocean areas away from the collapsing peripheral bulges, and it has been detected as 1–2 m of sea-level drop during the past ~3 k.y. at Pacific islands (Grossman et al., 1998). Such observations are consistent with predictions that rates of eustatic drop should decay exponentially with time (Mitrovica and

Peltier, 1991). An analogous effect occurs along the edges of continents (e.g., parallel to the African coastline in Fig. 3B), where isostatic compensation of ocean basin subsidence causes sublithospheric mantle to flow landward across continental boundaries. This causes relative sea-level rise for the ocean within a few hundred kilometers of shorelines, and a corresponding eustatic sea-level drop in the global ocean that is comparable in amplitude to the drop produced by the collapse of peripheral bulges (Mitrovica and Milne, 2002).

The sea-level record of the Quaternary (e.g., Fig. 2B) primarily records a climate-induced glaciation history that should be accompanied by spatial variations in sea level associated with the solid Earth's elastic and viscous responses to loading (Fig. 3). Observations of these spatial variations for past sea-level highstands can, in principle, be used to infer the patterns and amplitudes of past deglaciation events (Lambeck et al., 2012). For example, studies that account for sea-level variability during the Last Interglacial (ca. 120 ka; Fig. 2B) (Dutton and Lambeck, 2012; Kopp et al., 2009) show that global sea level during this period was likely ~6–8 m higher than today, a value higher than previous estimates based on direct observations made on seemingly "stable" shorelines (e.g., Muhs, 2002), and lower than rates inferred from global averaging of paleoshoreline data (e.g., Pedoja et al., 2011). Similar analyses for other interglacial periods (e.g., Raymo and Mitrovica, 2012) have exploited hemispheric variability in sea-level change to re-estimate past deglaciation volumes for individual ice sheets. Some of these analyses indicate more significant deglaciation of Antarctica or Greenland than previously thought (Raymo et al., 2011). Such constraints are important because the degree of melting that major ice sheets experienced during past warming events can be used to estimate the extent of future deglaciation (Overpeck et al., 2006) and associated sea-level rise (Rohling et al., 2008) that we may expect due to current and future warming.

## PLATE TECTONICS, VOLCANISM, AND SEA-LEVEL CHANGE SINCE THE CRETACEOUS

Sea-level change over geologic time scales, as inferred from sedimentary stratigraphy (e.g., Haq and Al-Qahtani, 2005; Haq et al., 1987; Haq and Schutter, 2008; Miller et al., 2005; Vail et al., 1977) and geological evidence of continental transgression and regression (e.g., Bond, 1976), exhibits variations on time scales ranging from millions to hundreds of millions of years. In particular, highstands in the Cretaceous and Paleozoic indicate sea level a few hundreds of meters higher than present levels (Fig. 2C). Because sea level on such long time scales should be fully compensated isostatically (i.e., full relaxation of ocean basin loading), a mass of mantle equal to that of the added water depth should become displaced from beneath oceanic lithosphere. Since seawater density is ~30% that of mantle rock, isostatic compensation of seawater (if fully completed) should cause observed sea-level change to be damped by a factor of 0.7 compared to changes in water-level thickness (Pitman, 1978). In what follows, I will account for isostatic compensation when reporting the impact of various driving mechanisms on sea level. However, when considering constraints on geologic sea-level change, one must keep in mind (and this will be discussed later herein) that all observations are made relative to a land surface that is in motion vertically (Jones et al., 2012; Lovell, 2010; Moucha et al., 2008), and that sea-level changes necessarily involve large hydrologic mass redistributions that induce significant spatial variations in sea level (Cramer et al., 2011; Mitrovica et al., 2009; Raymo et al., 2011), as discussed already. Thus, sea-level variations in geologic time should be considered within the context of solid and water surfaces that are moving relative to each other in a spatially heterogeneous way.

Sea-level variations of hundreds of meters cannot be explained by climatic effects alone. Earth's present-day ice sheets contain the seawater equivalent to 64 m of sea level (Lemke et al., 2007). When isostatically compensated, an ice-free world should thus exhibit eustatic sea level an average of ~45 m higher than present levels (although with accompanying spatial variations). A sea-level drop of this amplitude is thought to have occurred during the Cenozoic as climatic cooling induced ice-sheet growth (Gasson et al., 2012). Seawater cooling by ~12 °C (Lear et al., 2000) also likely caused an additional ~12 m of sea-level drop (Miller et al., 2009). Thus, climatic effects are likely responsible for ~57 m of Cenozoic sea-level drop, although much of this eustatic change may have taken place relatively abruptly at the Eocene-Oligocene boundary (ca. 34 Ma) (DeConto and Pollard, 2003; Gasson et al., 2012; Lear et al., 2000). Thus, pre-Cenozoic sea-level variations, and Cenozoic variations in excess of ~57 m, must result from geological changes in the "containing" volume of the ocean basins (Worsley et al., 1984). As a result, most studies (e.g., see review by Harrison, 1990) explain Phanerozoic sea-level change in terms of plate-tectonic mechanisms that affect the average area or depth of the ocean basins (Miller et al., 2005). In what follows, I will review the past quarter century of progress in constraining such mechanisms (Fig. 1C) and will quantify the contribution of each to a new sea-level budget for the Cenozoic and Cretaceous. While doing so, I will consider the implications of time-dependent solid Earth deformation for observations of sea-level change (Fig. 2C).

### Ridge Volume

Because seafloor depth increases rapidly as lithosphere spreads away from a mid-ocean ridge (e.g., Stein and Stein, 1992), changes in the volume of the global ridge system have long been considered to be the major contributor to eustatic sea-level change over geological time scales (Cogné et al., 2006; Hays and Pitman, 1973; Kominz, 1984; Müller et al., 2008b; Pitman, 1978; Xu et al., 2006). There are two types of tectonic changes to the global ridge system that may increase eustatic sea level globally. First, the global ridge system may lengthen if new ridges form within the ocean basins (Worsley et al., 1984). For example, the elongation of the Mid-Atlantic Ridge between 180 and 120 Ma is thought to have raised global sea level by ~30–50 m, with an additional ~70 m of associated rise due to passive-margin extension and sedimentation (Heller et al., 1996). Second, the average spreading rate of Earth's system of spreading ridges may increase. This acceleration of spreading rates tends to "fatten" Earth's ridge system, raising the average depth of the seafloor and elevating global sea level (Gaffin, 1987; Pitman, 1978). Constraints on both trends can be gained using carefully constrained tectonic reconstructions, which have recently improved significantly (e.g., Müller et al., 2008b; Seton et al., 2012; Torsvik et al., 2010).

To preserve mass, changes in plate-tectonic spreading rates must be accompanied by associated changes in Earth's internal density structure (Gurnis, 1990c). For example, an increase in spreading rate forces old lithosphere to subduct more rapidly into the mantle interior, where it enhances mantle downwelling. This downwelling dynamically depresses Earth's surface above it (see "Dynamic Topography" subsection), which locally amplifies relative sea-level rise (Gurnis, 1993). This subsidence, however, also expands ocean basin volume and thus partially opposes the eustatic rise associated with the initial acceleration of ridge spreading. The mantle's dynamic response should be smaller (because some of the associated boundary deflections may occur in continental regions or on the core-mantle boundary; Gurnis, 1990c) and delayed in onset by several tens of millions of years (Husson and Conrad, 2006). Thus, sea-level adjustments computed solely from spreading-rate variations can be considered as upper bounds on net sea-level change because they should be accompanied by some dynamic deflection of the seafloor. Fully self-consistent estimates of the sea-level response due to changes in the Cenozoic and Cretaceous ridge systems have been achieved only recently using numerical models that preserve mantle mass (Spasojevic and Gurnis, 2012).

To estimate how changes in global average spreading rate affect eustatic sea level, I computed the sea-level response to a linear change in global average plate velocity (half-spreading rate) from a given initial value to a final value of 2.6 cm/yr (corresponding to the present-day average value), occurring over a given time scale (Fig. 4B). I applied these changes (two examples are shown in Fig. 4A) to a single plate with a ridge-trench width of 3300 km, which corresponds to the average half-width of the present-day ocean basins (estimated by dividing the ocean area $360 \times 10^6$ km$^2$ by the double ridge length ~$110 \times 10^3$ km). The resulting changes in ridge volume, which can be computed by applying an empirical relationship between seafloor age and seafloor depth (Stein and Stein, 1992), can be interpreted as changes in global eustatic sea level (after isostatic compensation of the seawater) if we assume that all ocean basins exhibit tectonic behavior similar to the basin modeled here (Fig. 4B). This assumption simplifies tectonic change that varies between ocean basins, but it allows us to estimate the eustatic response to net changes in the global average spreading rate. Although these estimates (Fig. 4B) do not include mass preservation, which may further reduce eustatic change (Gurnis, 1990c), they do show that tens of meters of global change may be possible in only a few million years—but only if plate velocities vary globally by a few centimeters per year (factor of ~2) during this time period. Application of this calculation to the sea-level record (Fig. 2C) shows that second-order sea-level variations, with amplitudes of ~50 m and occurring over time scales of ~20 m.y. (Fig. 2C), require spreading rates to globally accelerate or decelerate by ~50% (from ~4 cm/yr or ~1.5 cm/yr to the present-day value of ~2.6 cm/yr; Fig. 4B). The larger first-order sea-level variations, with amplitudes of ~200 m occurring over ~100 to ~200 m.y., require similar ~50% changes in spreading rate, but occurring over much longer time scales of hundreds of millions of years (Fig. 4B).

What constraints do tectonic reconstructions of plate motions place on ridge length and spreading-rate variations? Some authors have noted that information about the past seafloor is difficult or impossible to constrain given that some ancient seafloor has been lost to subduction (e.g., Rowley, 2002, 2008). However, spreading rates for lost seafloor can be inferred from rates on conjugate seafloor (the counterpart formed simultaneously on the opposite side of a spreading ridge) that has been preserved (e.g., Kominz, 1984), and by utilizing, and extrapolating from, additional geological constraints (Müller et al., 2008b; Seton et al., 2012). Using such methods, Xu et al. (2006) reconstructed the paleoseafloor through the Cenozoic, and Müller et al. (2008b) did so for the Cenozoic and Cretaceous. Both reconstructions suggest that the seafloor

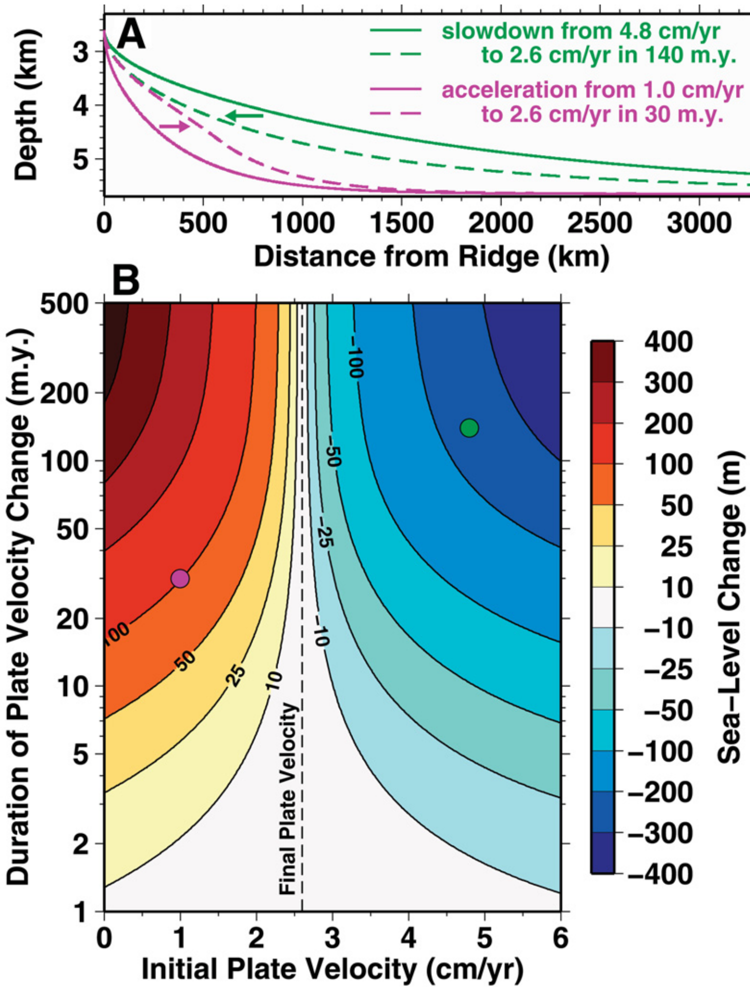


**Figure 4. Estimation of the impact of ridge volume change on global eustatic sea level, calculated by comparing the volume of a half-ridge on a 3300-km-wide plate that is initially moving at a given steady-state rate (e.g., solid lines in A) to the volume of the same ridge after its half-spreading rate has linearly increased or decreased to a rate of 2.6 cm/yr (the current global average half-spreading rate) over a given period of time (e.g., dashed lines in A). The depth of the basin is computed from Stein and Stein's (1992) relationship between seafloor age and depth, which is calibrated using the bathymetry of the modern seafloor (including marine sediments and volcanics). The change in sea level for this basin (assuming isostatic compensation of seawater) can be interpreted as global eustatic change if we assume that tectonic changes to this ridge system represent average changes to the global ridge system. Such changes are shown (in B) for a range of values for the initial plate velocity (*x*-axis) and the acceleration/deceleration duration (*y*-axis). The green and pink dots show sea-level change for the two spreading rate change scenarios exemplified in A. The green dot shows a sea-level scenario that approximates the one inferred from Müller et al.'s (2008b) reconstruction (Fig. 5A).**

production rate decreased by only 20%–30% during the Cenozoic, primarily due to a decrease in the rate of seafloor spreading as opposed to a change in the length of the ridge system (Fig. 5). This spreading-rate slowdown was a continuation of a long-term trend that started in the Early Cretaceous (Seton et al., 2009), when spreading rates were ~80% faster than they are today (also for a ridge system of similar length; Fig. 5). Corresponding spreading-rate slowdowns from ~3.3 cm/yr at 65 Ma and ~4.8 cm/yr at 140 Ma (Fig. 4A) yield global sea-level drops of ~60 m and ~240 m, respectively (Fig. 4B). To first order, these variations are comparable to those observed in the sea-level record (Fig. 2C). Shorter-duration fluctuations in spreading rate with an ~20–30 m.y. period and a 10%–20% amplitude are evident in the tectonic reconstructions (Fig. 5). Such fluctuations may lead to 10–20 m of eustatic sea-level change (Fig. 4B), which is somewhat smaller than the observed ~50 m amplitudes observed for second-order fluctuations in sea level (Fig. 2C). Thus, spreading-rate fluctuations may explain eustatic sea-level change on ~100 m.y. time scales (Müller et al., 2008b), but not on ~30 m.y. or shorter time scales because average spreading rate changes are not sufficiently rapid (Fig. 5).

Müller et al. (2008b) estimated the impact of ridge volume changes on eustatic sea level directly from tectonic reconstructions of paleoseafloor ages (Figs. 6A–6C). Their results show that changes in average seafloor basement depth (excluding bathymetry associated with sediments and volcanic edifices) have led to eustatic sea-level drop of ~240 m since the Early Cretaceous, consistent with rough estimates based on changes in spreading rate alone (Fig. 4B). The specific tectonic trends that are responsible for this sea-level drop can be deduced by analyzing Müller et al.'s (2008b) tectonic reconstructions. Currently, the area distribution of seafloor ages exhibits an approximately steady decrease in seafloor area with age (Fig. 6D) (Rowley, 2002), but this distribution has changed with time (Becker et al., 2009) and was significantly different at past times (Figs. 6E–6F). The convolution of these past distributions with Hillier and Watts' (2005) expression for the relationship between seafloor age and basement depth (calibrated for northern Pacific seafloor in the absence of sediments or volcanic edifices, which will be treated separately, and validated for the entire Pacific by Zhong et al. [2007a]) yields the volume of each ridge system relative to the 6120 m basement depth limit for the oldest seafloor. Dividing this volume by the area of the ocean basins ($361 \times 10^6$ km$^2$) yields the contribution of each basin's ridge system to sea level (Figs. 6G–6I). These calculations show that net aging of the seafloor dropped eustatic sea level by ~64 m and 187 m during the Cretaceous and Cenozoic, respectively, for a total of 251 m, which is consistent with Müller et al.'s (2008b) estimates. Two trends were responsible for this change. First, the opening of the Atlantic basin at the expense of the Pacific caused old Pacific seafloor to be replaced by younger seafloor in the Atlantic. Second, as the ridge system of the Pacific gradually shortened, the distance to the nearest ridge in that basin gradually increased (Loyd et al., 2007), which aged the Pacific seafloor.

Significant uncertainty is inherent to tectonic reconstructions of the seafloor (Rowley, 2008). Although Müller et al. (2008b) estimated uncertainty in seafloor ages as only ~10 m.y. in any given location, the major uncertainty for sea level is instead associated with the tectonic reconstruction itself. For example, the ridge system that subducted beneath the northwest Pacific at ~65 Ma (Fig. 6B) has since been completely destroyed, although geological observations along the Pacific western margin support early Cenozoic ridge subduction there (Agar et al., 1989; Whittaker et al., 2007). Similarly, seafloor ages for the Tethys basin are poorly constrained due to the fact that the seafloor of this basin has been entirely lost to subduction, but they can be inferred from geological observations of the former Tethys margin (Heine et al., 2004). Furthermore, Xu et al. (2006) showed that relatively minor changes to a tectonic reconstruction (e.g., differing treatments of early Cenozoic western Pacific tectonics) can significantly affect estimates of past sea level (by ~100 m or more) because they become amplified as the reconstruction is extrapolated backward in time. Thus, estimates of sea-level trends associated with past changes in ridge volume become significantly more uncertain as tectonic reconstructions of the seafloor are extended backward into the Cretaceous (Müller et al., 2008b) and especially into the Jurassic (Seton et al., 2012).

To obtain a new estimate of the uncertainty associated with tectonic reconstructions of ridge volume, I recalculated sea level after making extreme assumptions about the way in which seafloor ages may differ from those reconstructed by Müller et al. (2008b). In particular, I note that the Atlantic basin can be accurately reconstructed for past times because it is bounded by passive margins. Much of the seafloor in the Pacific and Indian basins, however, must be reconstructed because it has been lost to subduction. To accommodate error associated with this reconstruction, I reassigned the bathymetric height of seafloor that is added to these basins in the reconstruction. To estimate a minimum change in sea level, I assigned heights that decrease linearly from their present-day averages (724 m and 706 m above the 6120 m baseline in the Pacific and Indian basins, respectively) to their minima in the reconstruction (at 31 and 110 Ma), and then are constant at those values (at 718 and 648 m, respectively) from that point backward in time. Similarly, to estimate maximum sea-level change, I assigned heights that increase linearly to their maximum value in the reconstruction (at 119 and 47 Ma for the Pacific and Indian basins, respectively) and are then constant at those values (at 1091 and 885 m, respectively) backward through the reconstruction. This error estimate thus attempts to account for uncertainty in the tectonic reconstruction by assigning depths to reconstructed seafloor that are based on the range of depths in that basin. The result (Fig. 7) is a relatively tightly constrained eustatic sea-level drop of 100–200 m during the Cenozoic, with a best estimate close to the upper end of that range. Cretaceous sea-level change is significantly less well constrained, with 60–330 m of sea-level drop since 140 Ma being possible, with a best estimate of ~250 m higher eustatic sea level being maintained until ca. 80 Ma (Fig. 7). These estimates of the range in possible sea-level histories due to ridge volume changes are conservative, as they do not account for the additional geological and tectonic constraints that are included in Müller et al.'s (2008b) reconstruction. Nevertheless, they compare well with Müller et al.'s (2008b) error estimates, which are similar in amplitude (up to ~100 m), but are more uniformly distributed through the reconstruction period.

### Marine Sedimentation

Temporal changes to the average thickness of marine sediments affect eustatic sea level by changing the container volume of the ocean basins (Harrison, 1990; Müller et al., 2008b). Although we cannot measure either the time-history of sediment accumulation or its destruction by subduction, we can gain information about past sediment volumes by examining recent compilations of seismic constraints on the marine sediments (Divins, 2003; Winterbourne et al., 2009). Such constraints allow us to examine maps of marine sediment thickness (Divins, 2003), which, after corrections for isostatic compensation and sediment compaction (Sykes, 1996), range from nearly zero to ~5 km in thickness (Fig. 8A). These sediment piles represent the accumulation of sediments since the seafloor originally formed at a mid-ocean ridge. We can estimate how the average thickness of these piles may have changed by applying sediment distribution patterns estimated from

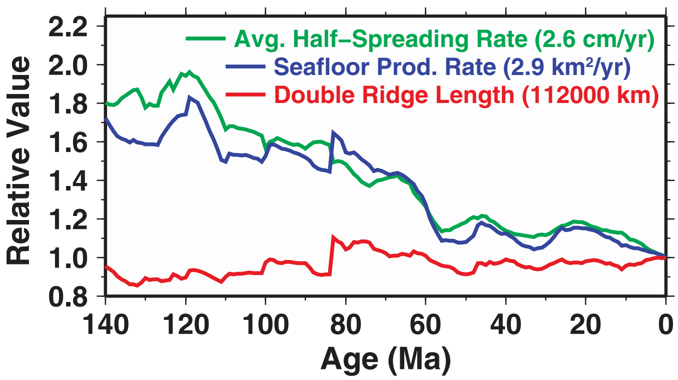


**Figure 5.** Temporal variations in seafloor half-spreading rate (green; computed following Becker et al., 2009), double ridge length (red; computed as the length of the 0 Ma isochron), and seafloor production rate (the product of half-spreading rate and double ridge length) during the past 140 m.y., computed based on the seafloor age reconstructions of Müller et al. (2008b). Variations shown are normalized by present-day value (noted in the legend). Note the relative stability of ridge length and the ~60%–80% slowdown in ridge spreading and seafloor production since the Early Cretaceous.

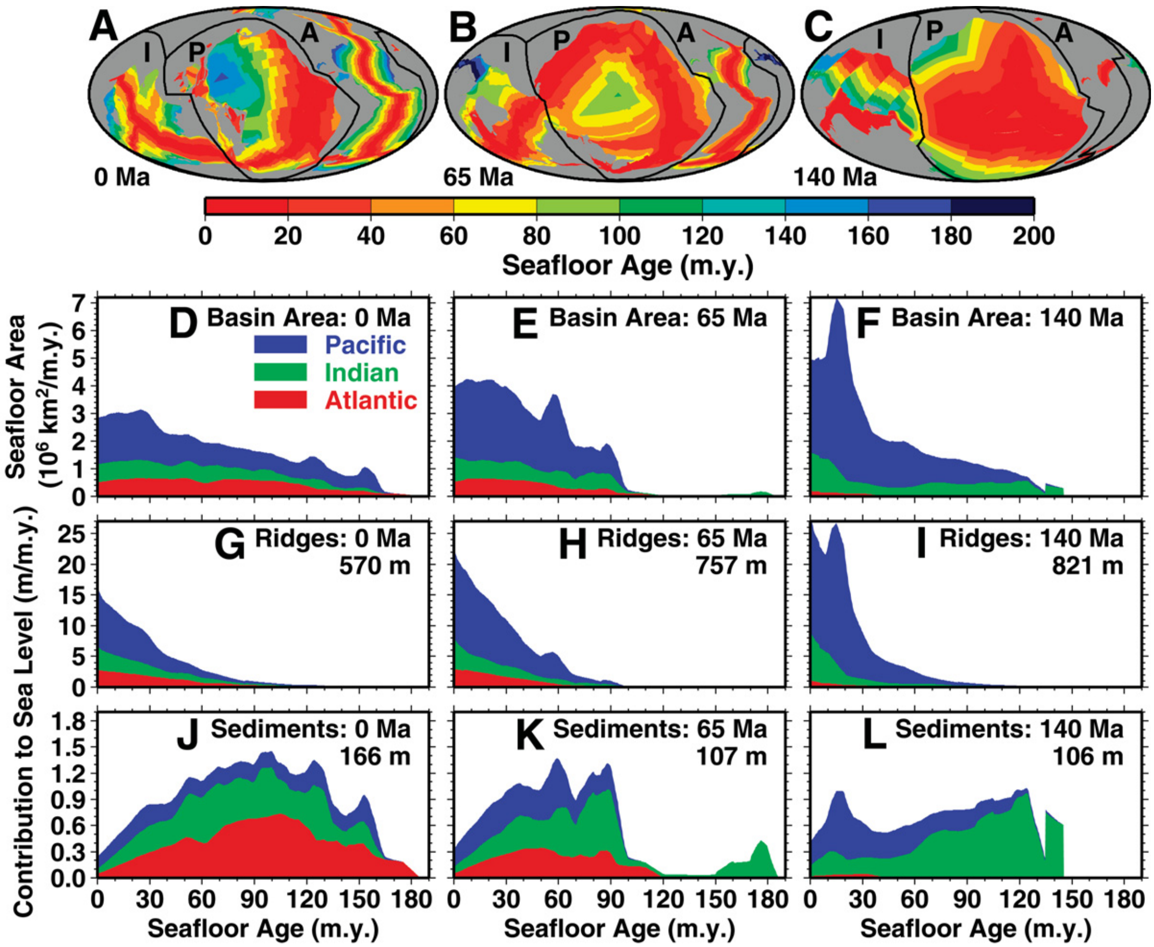


**Figure 6.** Tectonic reconstructions of seafloor age from Müller et al. (2008b) for (A) the present day (0 Ma), (B) the beginning of the Cenozoic (65 Ma), and (C) the beginning of the Cretaceous (140 Ma). (D–F) Corresponding seafloor area distributions for these reconstructions. (G–I) Computed contributions to eustatic sea level of ridge volume and (J–L) computed contributions to eustatic sea level of seafloor sediments. Area-age distributions (D–F) are computed from Müller et al.'s (2008b) age grids following Conrad and Lithgow-Bertelloni (2007). Ridge volumes (G–I) are estimated by applying Hillier and Watts' (2005) relation between seafloor age and basement depth, which excludes the bathymetric contributions of sediments and volcanism, to the area-age distributions (D–F) using the maximum seafloor depth (6120 m) as a baseline. Sediment thickness estimates (J–L) are estimated by applying the relation between sediment thickness and seafloor age determined in Figure 8B to the area-age distributions (D–F). For each distribution, the contribution of each of three basins (Pacific, Indian, and Atlantic, bounded and labeled as P, I, and A, respectively, in A–C) is denoted (cumulatively) by color (blue, green, and red).

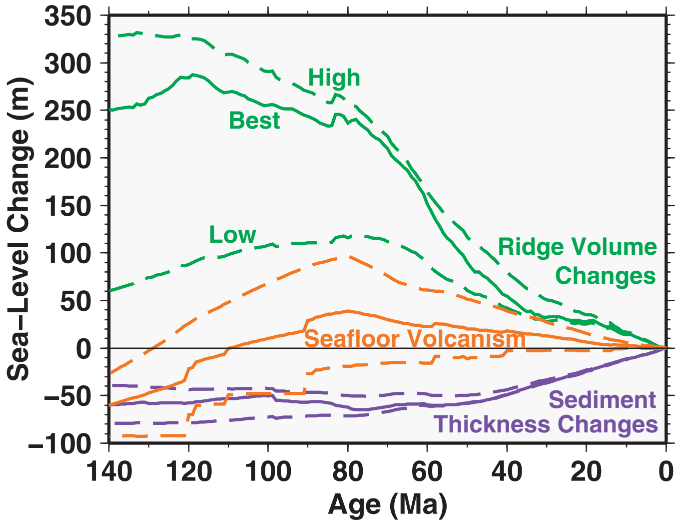


**Figure 7. Eustatic sea-level change due to global changes in ridge volume (green), sediment thickness (purple), and seafloor volcanism (orange). Ridge volume change is computed by summing the ridge volume in each ocean basin (Figs. 6G–6I) as a function of time and subtracting the present-day value. Best estimate follows directly from Müller et al.'s (2008b) reconstructions; high and low estimates are computed by reassigning bathymetric heights for reconstructed seafloor based on the historical range of average basin elevation (see text). Sediment thickness variations are also computed from Müller et al.'s (2008b) reconstructions (e.g., as in Figs. 6J–6L) by applying the estimated relation between sediment thickness and seafloor age for each basin (Fig. 8B) to the seafloor ages of the three (high, best, and low) ridge volume reconstructions. An upper bound on sea-level change caused by seafloor volcanism changes is estimated by assuming that all seafloor created in the Pacific basin during the mid-Cretaceous (140–80 Ma) hosted a volume of volcanism (large seamounts and flood basalts) that was 3.5 times that of non-Pacific or non-Cretaceous seafloor. A lower bound is based on Müller et al.'s (2008b) estimate, which assumes that Pacific Cretaceous seafloor that has been since lost to subduction had the same (low) volume of volcanism as non-Cretaceous seafloor. The best estimate curve for seafloor volcanism is midway between these upper and lower bounds.**

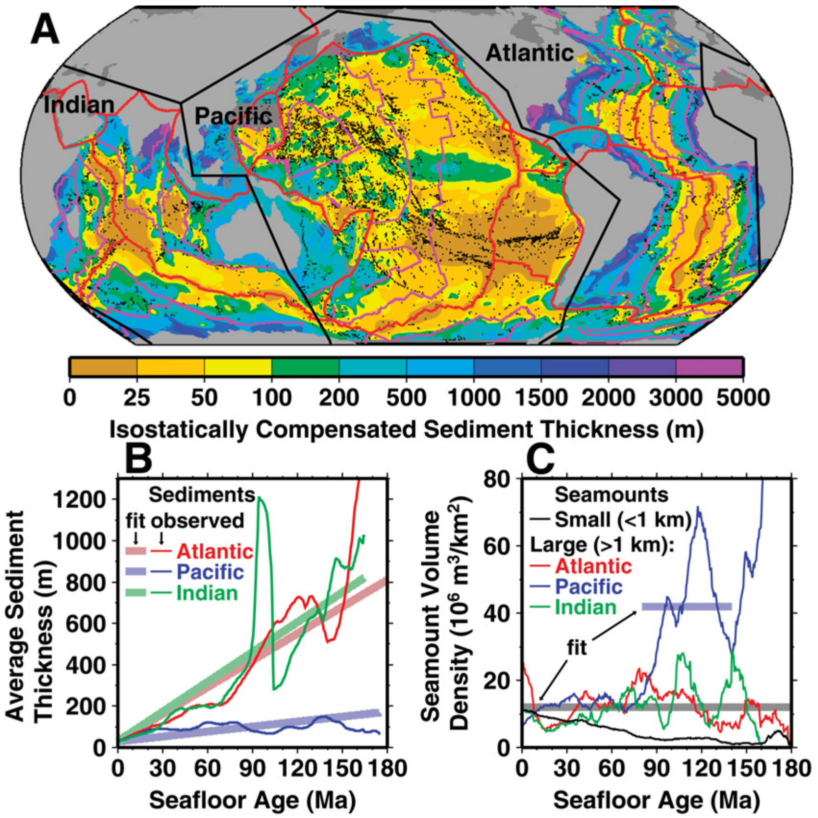


**Figure 8. Patterns of sedimentation and seamount emplacement on the present-day seafloor. (A) Map showing sediment thickness (colors) from Divins (2003), isostatically compensated following Sykes (1996), and locations of seamounts with heights greater than 1 km (black dots) from Kim and Wessel (2011). Plate boundaries (red lines) and the 50, 100, and 150 m.y. isochrons (pink lines) are shown for reference. (B) The variations of average isostatically compensated sediment thickness and (C) the average seamount volume density (both computed from the fields in A using a 10 m.y. running time window), with current seafloor age shown for the Atlantic, Pacific, and Indian basins (denoted by colors of thin lines; basin boundaries are shown in A). A linear fit to the currently observed sediment trends for each basin (thick lines in B) is used to compute the contribution of sediments to sea level (Fig. 7) from the area–age distribution of the seafloor at past times (e.g., Figs. 6D–6F). Note that because sediment thicknesses in B are isostatically compensated, they translate directly into the sediment impact on sea level. Seamount volume density, given in C in units that correspond to the thickness (in m) of a volume-equivalent uniform layer covering the seafloor, must be isostatically compensated (by multiplying by 0.7) before computing the impact on sea level. The seamount volume density for Pacific seafloor created in the mid-Cretaceous (42 m, thick blue line) is 3.5 times larger than it is for the rest of the seafloor (12 m, thick gray line). The decrease with seafloor age of the global seamount volume density for small (<1 km high) seamounts (black line in C) indicates that small seamounts may become more difficult to detect on older (and deeper) seafloor, and thus may be missing from Kim and Wessel's (2011) database.**

the present-day seafloor (Fig. 8A) to reconstructions of past seafloor (e.g., Figs. 6A–6C). This method assumes that the large spatial and temporal variations that characterize sediment accumulation on the seafloor can be smoothed by averaging spatially across an ocean basin and temporally over the lifetime of its seafloor. The resulting changes in average marine sediment thickness can be interpreted as driving eustatic sea-level change.

Two major patterns of average sediment thickness variations are noteworthy. First, sediments tend to get thicker with seafloor age (Fig. 8A), particularly in the Atlantic and Indian basins, because older seafloor has had more time to accumulate sediments. Second, the average sediment thickness in the Pacific is significantly thinner than it is in the Atlantic and Indian basins (Fig. 8A). This is because terrigenous sediments falling from the continents tend to get trapped and subducted at the trenches that ring the Pacific, and because the Pacific's greater average distance from land decreases bioproductivity, resulting in thinner pelagic sediments (Garrison, 2009). Both of these factors likely operated in the Pacific throughout the Cenozoic and Cretaceous, when the Pacific was also ringed by trenches and was even larger than it is today. Thus, reconstructions of the sediment load in the paleo-oceans must account for the fact that the age dependence of sediment thickness may vary between basins. To account for these differences, I computed the average (isostatically compensated) thickness of sediment in each basin as a function of seafloor age (thin lines, Fig. 8B). Linear fits to these curves (thick lines, Fig. 8B) show a similarly significant increase in sediment thickness with age in the Atlantic and Indian basins, while sediments in the Pacific basin are ~4 times thinner and increase with age more slowly.

Applying the sediment thickness versus seafloor age relations to the area-age distributions for each basin using Müller et al.'s (2008b) seafloor age reconstructions (Figs. 6D–6F), I estimated how changes in seafloor sedimentation affected sea level during the Cenozoic and Cretaceous. For example, the application of this analysis to the present-day seafloor (Fig. 6J) shows the largest sediment volumes (and thus the largest contribution to sea level) for middle-aged (ca. 60–120 Ma) seafloor that is still plentiful in area and has had time to accumulate large sediment thickness. As expected, the Pacific contribution to sea level is smaller than that of the Indian or Atlantic basins. Together, this analysis predicts that sea level is presently higher by 166 m due to sediments. At the beginning of the Cenozoic, however, the Atlantic contribution was much smaller (because that basin's seafloor was younger), and the Pacific's greater areal extent was balanced by its relative youth (Fig. 6K). These trends combined to cause only 107 m of sea-level elevation due to sediments. At the beginning of the Cretaceous, Atlantic sediments were minor, but old seafloor in the Indian basin accumulated thick sediments (Fig. 6L).

Together, these trends suggest that sediment thickness changes caused sea level to remain relatively constant during the Cretaceous, and then to rise ~60 m during the Cenozoic (Fig. 7). Potential uncertainty in Müller et al.'s (2008b) tectonic reconstruction permits deviations from these trends of up to ~30% (Fig. 7). By contrast, Müller et al. (2008b), using a similar method that accounted for latitudinal variations in sediment thickness but not differences between basins, predicted a similar sea-level trend with a slightly smaller ~50 m sea-level rise during the Cenozoic. This difference is likely due to Müller et al's (2008b) overprediction of sediment thicknesses in the Pacific. Neither model fully accounts for possible temporal variations in the tectonic, chemical, biological, and climatic conditions that control sedimentation (e.g., Harrison et al., 1981; Mackenzie and Morse, 1992; Meyers and Peters, 2011), many of which are poorly constrained for past times. For example, the large sediment thicknesses found on passive margins are not approximated well by either model, and yet may have affected sea level by tens of meters. Furthermore, evolutionary changes in marine planktonic biota, or changes in carbonate compensation depth, may affect rates of carbonate sedimentation or consumption over time, and thus may impact sea level (e.g., Mackenzie and Morse, 1992).

## Seafloor Volcanism

Seafloor volcanism produces seamounts, ocean islands, and flood basalts, all of which displace seawater and thus elevate eustatic sea level. Changes in the rate of production of these volcanic edifices, or in their rate of consumption at convergent margins, can induce sea-level change (Harrison, 1990; Müller et al., 2008b). Using a recent global database of seamount locations and sizes obtained from satellite altimetry (Kim and Wessel, 2011), I estimated basin-specific volume densities for large (heights >1 km) seamounts (Fig. 8A) as a function of seafloor age. The result (Fig. 8C) shows that seamount volume approximately equates to a 12-m-thick layer spread across the seafloor (thick gray line, Fig. 8C), except for Pacific seafloor produced during the Cretaceous, for which the corresponding layer is ~42 m thick. After isostatic compensation, the total volume of large seamounts in Kim and Wessel's (2011) database (~$5.8 \times 10^6$ km$^3$) elevates sea level by ~11 m, with ~70% of this value resulting from Pacific seamounts. Smaller seamounts (<1 km high) in this catalog raise sea level by only ~3 m (~1 m from each ocean basin), but this number is probably an underestimate because Kim and Wessel's (2011) database is incomplete due to detection limitations for small seamounts, especially in deep water (Wessel et al., 2010). Indeed the volume density of small seamounts on the youngest seafloor (shallowest waters) is enough to raise sea level by ~8 m (obtained by isostatically compensating an equivalent layer thickness of ~11 m; Fig. 8C). This is comparable to the sea-level impact of larger seamounts, but the lack of evidence for spatial or temporal clustering of small seamounts suggests that temporal changes to their total volume are unlikely.

Flood basalts from past episodes of extensive seafloor volcanism are preserved in large oceanic plateaus (Greene et al., 2010) that displace significant seawater. Large igneous provinces on the present-day seafloor currently elevate sea level by ~100 m (Müller et al., 2008b). However, the sea-level rise associated with the emplacement of this volcanic material since the Cretaceous must have been offset by sea-level drop associated with concurrent destruction of other provinces during this time period. For example, the Wrangellia province accreted onto northwestern North America during the Late Triassic or Early Cretaceous (Greene et al., 2010), and the Caribbean-Colombian Plateau accreted to South America and the Caribbean Arc at ca. 88 Ma (Kerr et al., 2000). Although such accretion may serve to grow continental crust, accreted plateaus may eventually become subducted (Saunders et al., 1996) or otherwise obliterated by secondary processes at subduction zones (Kerr et al., 2000). Indeed, continental deformation such as the Laramide orogeny has been attributed to subduction of conjugates to the Hess and Shatsky oceanic plateaux (Liu et al., 2010). As a result, the volume of igneous volcanism that was lost during the subduction of Cretaceous seafloor is not known. Therefore, it is difficult to estimate sea-level trends resulting from the emplacement and destruction of large igneous provinces, except to note that the Cretaceous was a period of intense igneous activity (Kerr et al., 2000), especially in the Pacific (Larson, 1991). Therefore, we can infer that Pacific Cretaceous seafloor was more likely than other seafloor to host oceanic plateaux that elevated sea level, as is the case for large seamounts (Fig. 8C).

To estimate an upper bound for the contribution of seafloor volcanism to sea-level change, I assumed that volcanic additions to the seafloor approximately balanced subtractions, except during the mid-Cretaceous in the Pacific basin, when volcanic products (seamounts, ocean islands, and flood basalts) were emplaced at an accelerated rate. By assuming that total seafloor volcanic products (plateaux and seamounts) were ten times more volumetric than large seamounts (thus together corresponding to the ~110 m sea-level offset of the present day), and assigning Cretaceous seafloor (formed between 140 and 80 Ma) in the Pacific basin a volcanic volume density that is 3.5 times that of other seafloor (based on the current ratio; Fig. 8C), I computed sea-level change due to volcanism using Müller et al.'s (2008b) seafloor age reconstruction. The result (Fig. 7, seafloor volcanism upper bound) shows a rise of ~130 m during the Cretaceous followed by an ~100 m sea-level drop since 80 Ma. This model assumes rapid volcanism (consistent with an ~420-m-thick volcanic layer) in the Pacific during the Cretaceous, and thus produces a rapid rise while this material was being emplaced and rapid fall during its removal. Müller et al.'s (2008b) estimate of ~100 m of sea-level rise, which we use as a lower bound on seafloor volcanism (Fig. 7), implicitly assumes a much lower rate of Cretaceous Pacific volcanic emplacement because it requires subducted seafloor to be devoid of igneous provinces. Because the actual impact of seafloor volcanism likely exists somewhere in between these two extremes, my "best estimate" is midway between these upper and lower bounds, producing ~100 m of sea-level rise during the Cretaceous and 40 m of sea-level drop since 80 Ma (Fig. 7).

## Ocean Basin Area

Because the geologic record of accretion and erosion at continental margins is often obscured, it is difficult to estimate how the areal extent of the ocean basins may have changed as a function of time. Nevertheless, the formation of passive margins by rifting is associated with crustal stretching (Heller et al., 1996; Kirschner et al., 2010), and continental collision involves orogeny and continental shortening. Because these processes respectively expand or contract continental areas at the expense of the oceans, they tend to raise or lower eustatic sea level. For example, Müller et al. (2008b) estimated that the ocean basins have decreased in area by $10^7$ km$^2$ since ca. 140 Ma due to continental rifting and new ocean basin formation. Assuming an average seafloor depth of 4000 m and isostatic compensation of seawater, this area decrease corresponds to eustatic sea-level rise of ~78 m. A more detailed study of passive-margin tectonics by Kirschner et al. (2010), however, suggested that continental extension has decreased ocean basin area by a significantly smaller amount, corresponding to eustatic rise of only ~21 m since Pangea breakup.

On the other hand, continental contraction may also be occurring, and indeed has been invoked to explain observations of recent coastal uplift worldwide (Pedoja et al., 2011). Harrison (1990) estimated that the Alpine-Tibetan orogeny between Africa-India and Eurasia reduced continental area by ~$3.2 \times 10^6$ km$^2$ since the collision began at ca. 50 Ma. Recent estimates suggest that the majority of this shortening, associated with ~1000 km of shortening along the ~2500-km-long Tibetan-Himalayan orogeny, occurred during a second stage of collision beginning at ca. 25 Ma (van Hinsbergen et al., 2012). This scenario suggests ~25 m of sea-level drop occurring since 50 Ma, with nearly 20 m occurring since 25 Ma. Some additional sea-level drop (~3 m) may have occurred since ca. 40 Ma due to shortening of South America by ~200 km along an ~2000 km length of the Central Andes (Oncken et al., 2006). Thus, although continental orogeny likely caused ~30 m of sea-level drop during the Cenozoic, this drop is likely approximately balanced over longer time scales by extension-induced sea-level rise associated with Pangea rifting and breakup since ca. 200 Ma.

## Dynamic Topography

Mantle flow is thought to support up to ~1 km of long-wavelength (thousands of kilometers) topographic relief on Earth's surface (Hager et al., 1985), with elevated topography occurring above mantle upwelling (e.g., Gurnis et al., 2000; Lithgow-Bertelloni and Richards, 1998) and depressed topography above downwellings (e.g., DiCaprio et al., 2011). This "dynamic topography" may evolve as patterns of mantle flow change with time and as continents and oceans move laterally with respect to the underlying mantle flow (for a recent review, see Flament et al., 2013). This dynamic topography deflects seafloor bathymetry (e.g., Conrad et al., 2004; Winterbourne et al., 2009) and may change the container volume of the ocean basins, and thus eustatic sea level, if its net bathymetric impact changes with time (Gurnis, 1990c, 1993; Spasojevic and Gurnis, 2012). Additionally, changing dynamic support of continental margins may lead to regional transgressions or regressions caused by local subsidence or uplift of individual coastlines (Moucha et al., 2008; Müller et al., 2008b; Spasojevic et al., 2008). Constraining the changing influence of dynamic topography on sea level (both eustatic and relative to a continental margin) requires knowledge of the ways in which mantle flow patterns have changed with time and the ways in which continents have moved relative to them.

To determine the impact of present-day dynamic topography on sea level, Conrad and Husson (2009) used a model of global mantle flow driven by density heterogeneity inferred from seismic tomography (Fig. 9). This model shows that dynamic topography currently uplifts the seafloor enough to positively offset sea level by ~90 ± 20 m. The offset is positive because subduction preferentially induces negative dynamic topography for continental areas (Gurnis, 1993), thus causing a positive average seafloor deflection for oceanic areas (Fig. 9). Additionally, recent mantle tomography shows large low-seismic-velocity anomalies in the lower mantle, which may be interpreted as low-density upwellings beneath the South Pacific and Africa (Fig. 9). In this case, the positive dynamic topography above these active upwellings also positively offsets eustatic sea level, and grows with time as the driving density heterogeneity approaches the surface (Gurnis et al., 2000). Using a numerical adjoint model of time-dependent mantle flow, Spasojevic and Gurnis (2012) estimated 100–200 m of eustatic rise since the Late Cretaceous due to dynamic topography, mostly due to growing seafloor uplift associated with the rise of low-density upwellings. On the other hand, recent seismological constraints suggest that these lower-mantle structures are chemically dense (Deschamps and Trampert, 2003; Ishii and Tromp, 1999; Kennett et al., 1998), in which case upwellings above these structures, and their associated dynamic topography, may be relatively static (Lassak et al., 2010; McNamara et al., 2010).

Continental motion relative to a background pattern of dynamic topography may also impact eustatic sea level (Conrad and Husson, 2009; Gurnis, 1993). For example, continental motions in the no-net-rotation plate motion reference frame (DeMets et al., 1994) currently induce ~0.2 m/m.y. of sea-level rise because continents are on average moving toward depressed seafloor (Fig. 9). This estimate, however, depends on the reference frame for continental motion; when measured relative to Pacific hotspots (Gripp and Gordon, 2002), westward rotation of the lithosphere at ~0.44°/m.y. causes continents to move, on average, away from depressed regions, thus inducing sea-level drop (Conrad and Husson, 2009). Although the appropriate reference frame for plate motions is debated, most constraints currently suggest that net rotation is less than ~0.26°/m.y. (for a review, see Conrad and Behn, 2010), which implies that continental motion relative to dynamic topography is currently causing at least some sea-level rise. Indeed, most intracontinental basins experienced dynamic subsidence during the Cenozoic, consistent with motion toward regions of negative dynamic topography (Heine et al., 2008).

Given current uncertainty regarding the time dependence of mantle flow patterns and the absolute motions of continents relative to the deep mantle, placing constraints on the time-history of sea-level change due to dynamic topography necessarily incorporates significant uncertainty. To develop a conservative estimate, I used the Torsvik et al. (2010) reconstruction of absolute plate motions to move continental areas over the Conrad and Husson (2009) model for present-day dynamic deflection of the seafloor (Fig. 9). This exercise results in a relatively constant ~0.5 m/m.y. of eustatic sea-level rise since 140 Ma, as continents (primarily the Americas) on average move toward the downward-deflected seafloor. This value assumes that long-wavelength dynamic topography, and the lower-mantle flow field that generates it, remains relatively stationary with time despite changing geometries of the surface plates. This is a significant simplification of the complexity of large-scale mantle flow, but it is one that is supported by evidence that the major mantle upwelling structures beneath Africa and the South Pacific remain relatively stationary with time (Dziewonski et al., 2010; Li and Zhong, 2009; Torsvik et al., 2008a, 2008b). On the other hand, if subduction and associated depressed topography migrate with the continents, changes in average dynamic support of the seafloor may be smaller than the 0.5 m/m.y. assumed here. At the other extreme, propagation of active upwelling toward the surface may amplify seafloor dynamic topography, which would cause eustatic sea level to rise faster than 0.5 m/m.y. (Spasojevic and Gurnis, 2012). Thus, 0.5 m/m.y. of eustatic sea-level rise since 140 Ma represents an intermediate estimate of

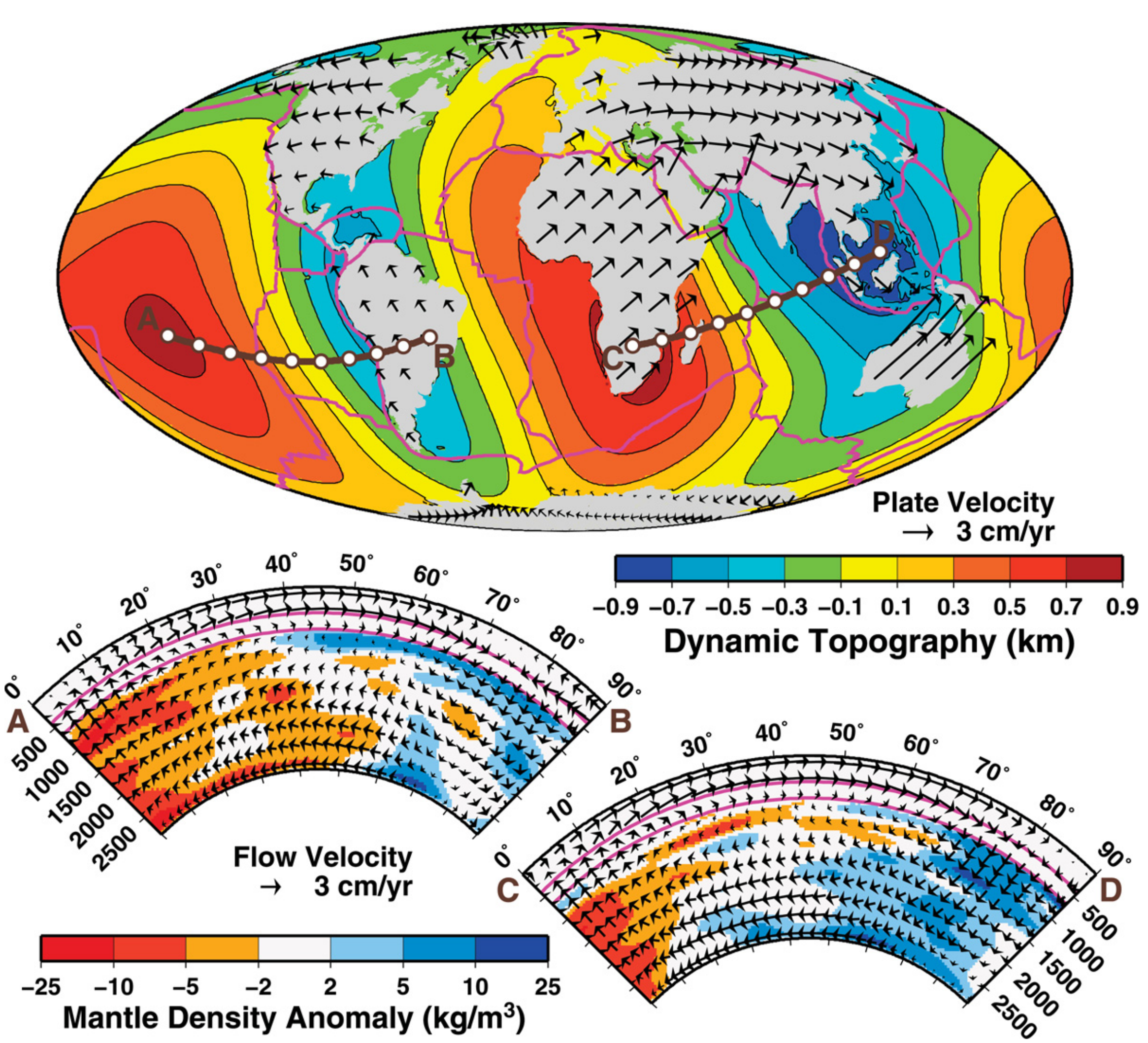


**Figure 9. Dynamically supported topographic deflection of the seafloor ("dynamic topography," shown as colors in top panel) is supported by global mantle flow (arrows in cross sections A-B and C-D, which cut the mantle beneath the corresponding paths in the top panel), shown here from the model of Conrad and Husson (2009), which is driven by density heterogeneity in the lower mantle (colors in cross sections). Dynamic topography is on average positive over the ocean basins and elevates eustatic sea level by ~48 m. This offset changes as the flow field advects density heterogeneity within the mantle, and as continents move laterally (arrows on top panel; from DeMets et al., 1994). Application of the Torsvik et al. (2010) reconstruction of continental motions to this topography field (see text) produces ~50 m of eustatic sea-level rise during the past 100 m.y. (Fig. 10).**

the impact of dynamic topography on eustatic sea level (Fig. 10), although one that is accompanied by significant uncertainty.

Because dynamic topography deflects the land surface against which relative sea level is measured, its change with time may significantly impact observations of sea level (e.g., Moucha et al., 2008). For example, numerical models of mantle flow suggest that the east coast of North America may have subsided 100–400 m during the Cenozoic, due to the westward motion of North America over negative topography associated with downwelling of the Farallon slab in the mid-mantle (Kominz et al., 2008; Müller et al., 2008b; Spasojevic et al., 2008). Indeed, some observations of apparent sea-level change may be primarily sensing vertical motion of the land surface caused by dynamic topography (Lovell, 2010; Petersen et al., 2010). The time-dependent contribution of dynamic topography to individual sea-level records can be estimated from numerical models of mantle flow, but uncertainty in the rheological properties of the mantle and the difficulty of advecting mantle density heterogeneity backward in time can make such models difficult to constrain (Spasojevic et al., 2009). Indeed, the corruption of the sea-level record by dynamic topography has led some to call into question the utility of the concept of eustasy (Mitrovica, 2009), and others to suggest constraining eustatic change using detailed intercomparisons of sea-level records from geographically dispersed regions (Ruban et al., 2010b, 2012; Zorina et al., 2008).

## Sea-Level Budget for the Cretaceous and Cenozoic

By summing the sea-level contributions from climate change and the five different solid Earth mechanisms described in the previous subsections, I computed a prediction of net sea-level change since the Cretaceous (Fig. 10). Of the six contributions, climatic cooling (~60 m), ridge volume decrease (~250 m), and ocean area increase (~10 m) caused ~320 m of net sea-level fall since the Early Cretaceous, with most of this drop (~200 m) occurring as ridge volume decrease during the Cenozoic (Fig. 7). Increases in average sediment thickness (60 m) and average seafloor dynamic topography (~70 m) have raised sea level by ~130 m since the Early Cretaceous. In this summation, the rapid seafloor accumulation of volcanic plateaux and seamounts during the Cretaceous first elevated sea level by ~100 m and then dropped it by ~40 m during the Cenozoic as some of this material began to subduct (Fig. 7). Together, these contributions predict an ~100 m sea-level increase between 140 and 80 Ma, followed by a relatively steady drop of ~250 m since ca. 80 Ma (Fig. 10). It is essential to note, however, the significant uncertainty in the magnitude and timing of several of these predicted sea-level trends. For some contributions, the uncertainty in sea-level change since the beginning of the

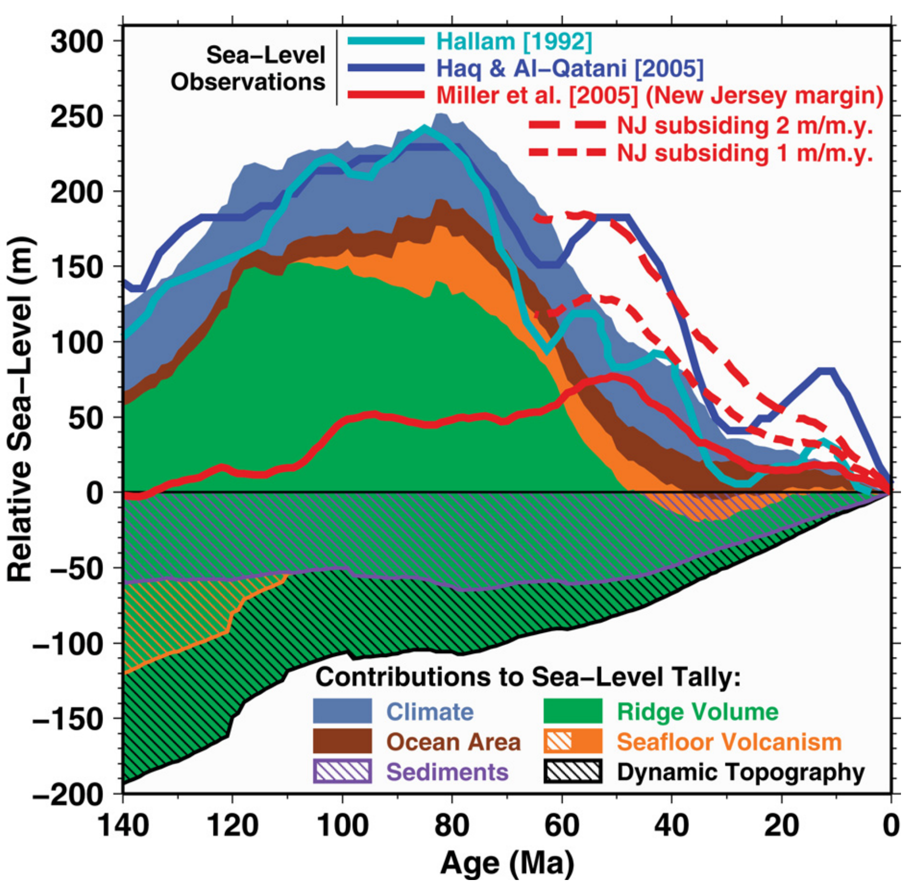


**Figure 10. Comparison of sea-level observations (solid lines, based on curves of Fig. 2C; Hallam, 1992; Haq and Al-Qahtani, 2005; Haq and Schutter, 2008; Miller et al., 2005) to a summation of best estimates for Cenozoic and Cretaceous drivers of eustatic sea-level change (colored and hatched areas). Among observations of sea level, Miller et al.'s (2005) record for New Jersey (red solid line) is also shown corrected for Cenozoic subsidence of eastern North America at rates of 1 and 2 m/m.y. (red dashed lines) due to regional dynamic topography associated with Farallon slab descent in the mid-mantle. The cumulative sea-level budget is shown by summing each of six contributions shown by filled areas: areas with solid colors show negative contributions to eustatic sea-level change (i.e., causing net global sea-level drop relative to the present day), while hatched areas show positive contributions (i.e., causing net global sea-level rise). Changes to ridge volume (solid green area), sediment thickness (purple hatched area) and seafloor volcanism (which, relative to current levels, depresses sea level before 110 Ma and elevates it after; orange hatched and solid areas, respectively) are taken from "best estimates" of Figure 7. Growth of the ice sheets and seawater expansion due to cooling of the Cenozoic climate (solid gray-blue area) are assigned 57 m of sea-level rise, most of which occurred near the Eocene-Oligocene transition (Cramer et al., 2011); here 4 m/m.y. of drop is assigned for 8 m.y. centered on 34 Ma and 0.5 m/m.y. afterward. Sea-level change associated with ocean basin area change (brown area) assumes 30 m of sea-level drop due to continental orogeny since 30 Ma and 20 m of sea-level rise since 200 Ma associated with passive-margin extension (Kirschner et al., 2010). Eustatic change driven by dynamic topography (black hatched area) is associated with net motion of continents toward regions of negative dynamic topography since the Cretaceous, and it is estimated here as the average (isostatically compensated) dynamic topography over ocean areas from Torsvik et al.'s (2010) plate-tectonic reconstruction.**

Cretaceous is 50%–100%, especially for those factors (e.g., seafloor age, sedimentation, and volcanism) that rely on assumptions about seafloor that is now subducted (Fig. 7). Furthermore, beyond the application of modeling constraints to individual tectonic and epeirogenic events recorded in the geologic record, there are few good constraints on the total ocean area and average dynamic topography of the seafloor for past times; even the basic trends of these curves therefore involve some speculation.

Despite its inherent uncertainty, the sea-level budget provides a relatively good match to sedimentary constraints on sea-level change (Hallam, 1992; Haq and Al-Qahtani, 2005; Haq and Schutter, 2008) throughout the Cretaceous and Cenozoic (Fig. 10). This tally accurately predicts the ~200–250 m sea-level drop since ca. 80 Ma, and a smaller rise of ~100–150 m from 140 to 80 Ma; these trends are seen in both the Hallam (1992) and Haq and Al-Qahtani (2005) curves. Second-order peaks in sea level at ca. 10 Ma and ca. 50 Ma are not predicted by the tally and cannot be explained by any of the processes described here. However, both Hallam (1992) and Haq and Al-Qahtani (2005) constrained rapid sea-level drop during 40–30 Ma, which may have been partially due to rapid climate cooling at the Eocene-Oligocene boundary (Cramer et al., 2011). The sea-level budget also includes a rapid drop at this time for the same reason, but its magnitude is too small to explain the observed magnitudes (Fig. 10).

The record of sea-level change along the New Jersey margin (Miller et al., 2005) suggests only ~70 m of sea-level drop since 50 Ma, i.e., significantly less than suggested by Hallam (1992) and Haq and Al-Qahtani (2005), and by the sea-level budget (Fig. 10). This discrepancy has been attributed to changing dynamic support of the New Jersey margin by mantle flow. For example, Moucha et al. (2008) attributed the observed drop along the New Jersey margin to regional dynamic uplift of ~150–200 m occurring as eustatic sea level also rose by ~100 m since ca. 30 Ma. This scenario explains Miller et al.'s (2005) observation of relative sea level, but it is inconsistent with other constraints discussed here that suggest eustatic drop since the Oligocene (Fig. 10). Alternatively, several authors have noted that motion of New Jersey over the sinking Farallon slab in the mid-mantle must have induced dynamic subsidence of the New Jersey margin during the Cenozoic; this subsidence should cause observations of relative sea level at New Jersey to underestimate any eustatic sea-level drop. Müller et al. (2008b) estimated this subsidence to be 100–400 m (corresponding to 1.4–5.7 m/m.y.), while Spasojevic et al. (2008) suggested 300–600 m, depending on assumptions about mantle density and viscosity structures. However, as noted by Kominz et al. (2008), such subsidence rates imply Eocene sea level higher than indicated by other measurements. Adding 2 m/m.y. of New Jersey subsidence to the Miller et al.'s (2005) curve produces a sea-level record that approximately aligns with that of Haq and Al-Qahtani (2005) but produces faster sea-level drop than indicated by the sea-level budget calculated here (Fig. 10). The sea-level record of Haq and Al-Qahtani (2005), however, was measured on the Arabian plate, which may also experience (unknown) vertical motion due to dynamic topography. Reducing New Jersey subsidence to 1 m/m.y. causes the Miller et al. (2005) curve to approximately align with both the Hallam (1992) curve and the sea-level budget during the Cenozoic presented here (Fig. 10). This may suggest that subsidence rates in New Jersey were slower than indicated by geodynamic models (Kominz et al., 2008), but it is important to remember that all observations of long-term sea-level change may be influenced by dynamic topography to some degree, and that the sea-level budget has its own uncertainties. Nevertheless, the fact that our best estimates for the various sea-level change contributors combine to approximately predict the available observations of sea-level change provides of some confidence that we understand the basic processes that have driven sea-level change since the Cretaceous.

## MANTLE CONVECTION AND SEA-LEVEL CHANGE DURING SUPERCONTINENTAL CYCLES

Global compilations of sedimentary stratigraphy provide evidence for global-scale cycles of sea-level change throughout the Phanerozoic (Hallam, 1992; Haq et al., 1987; Haq and Schutter, 2008; Miller et al., 2005). These records show long-term periods of sea-level drop approximately coincident with episodes supercontinental aggregation, and sea-level rise during periods of dispersal (e.g., Fig. 2C). Worsley et al. (1984) suggested that these correlations result from a combination of three long-term tectonic trends during the supercontinental cycles, and that these trends have largely dominated sea-level change during the past 2 b.y. of Earth history. In particular, Worsley et al. (1984) proposed that continental aggregation should (1) decrease the average age of the seafloor (thus rising sea level), (2) increase the ocean basin area (via continental collision, thus decreasing sea level), and (3) thermally elevate continents (by trapping heat beneath them, thus decreasing sea level). Our understanding of solid Earth interior dynamics, however, has evolved during the past three decades and has caused us to reevaluate the mechanisms by which the solid Earth impacts sea level during supercontinental cycles (Figs. 11A–11B).

### Average Age of the Seafloor

Although Worsley et al. (1984) suggested a cycle of supercontinental dispersal causing seafloor aging (and the reverse for aggregation), these trends assume that the opposing Panthalassic Ocean (i.e., the Pacific in today's world) retained a constant average age throughout. We now know that the Pacific basin does not maintain constant seafloor age, but instead exhibits its own tectonic evolution as it shrinks (Loyd et al., 2007). In particular, both the ridge length and spreading velocity of the Pacific have decreased since the Cretaceous (Becker et al., 2009) (Figs. 5 and 6A–6C). If we take the early Cenozoic as the time of maximum continental dispersal (i.e., prior to the closure of the Tethys Ocean), then the average seafloor age was actually minimized at the time of maximum dispersal (Figs. 6B and 6E). In fact, the most recent tectonic evolution of the Pacific, as reconstructed by Müller et al. (2008b), exhibited an opposite trend to that proposed by Worsley et al. (1984). Indeed, Cogné and Humler (2008) calibrated a relationship between a measure of the degree of continental dispersal and the average seafloor age during the past 180 m.y. and found a roughly inverse relationship. Thus, at least for the most recent (observable) phase of the supercontinental cycle, average seafloor ages were younger when continents were more dispersed and older when they were less dispersed. This relationship suggests that supercontinental dispersal is associated with an increased ridge volume that causes sea-level rise, and that supercontinental assembly should be associated with sea-level drop (Fig. 11C). Based on the most recent cycle, the amplitude of these sea-level changes is about ~200 m. It should be noted, however, that the dynamics of plate tectonics for Panthalassic-type ocean basins are not well understood and may manifest themselves differently in different supercontinental cycles. Indeed, Cogné and Humler's (2008) study is based on seafloor age reconstructions for the past ~180 m.y., which may be poorly constrained for Cretaceous and younger times (Rowley, 2008). Thus, although the inferred variation of sea level within the supercontinental cycle (Fig. 11C) accurately predicts sea-level rise during the most recent (Mesozoic and Cretaceous) assembly and sea-level drop during the previous (mid-Paleozoic) dispersal (Fig. 2C), we cannot be sure that the tectonic variations that predict this relationship operated in the same way for past times.

### Continental Area

Supercontinent formation necessarily involves continental collision and associated orogeny (Fig. 11A), which lower sea level by expanding ocean basin area (Worsley et al., 1984). The most recent continental collision (between India and Eurasia) likely dropped sea level ~20–30 m (see "Ocean Basin Area" subsection), which roughly balances the estimated long-term sea-level rise caused by continental stretching during Pangea breakup (Kirschner et al., 2010). Thus, ocean area changes during the supercontinental cycle should cause sea-level drop during assembly and rise during dispersal (Fig. 11D). Continental contraction and expansion thus bolster the expected sea-level impact of seafloor age changes (Fig. 11C), but they do so at smaller amplitudes of several tens of meters.

### Dynamic and Thermal Support of Continents and Seafloor

Supercontinent breakup and dispersal are surface manifestations of the link between plate tectonics and convection in Earth's mantle, and vertical motions of continents and seafloor are reflective of this link (Zhang et al., 2012). Worsley et al. (1984) suggested that heat accumulates in the mantle beneath continents that are not moving laterally, and indeed this accumulation of heat may eventually contribute to

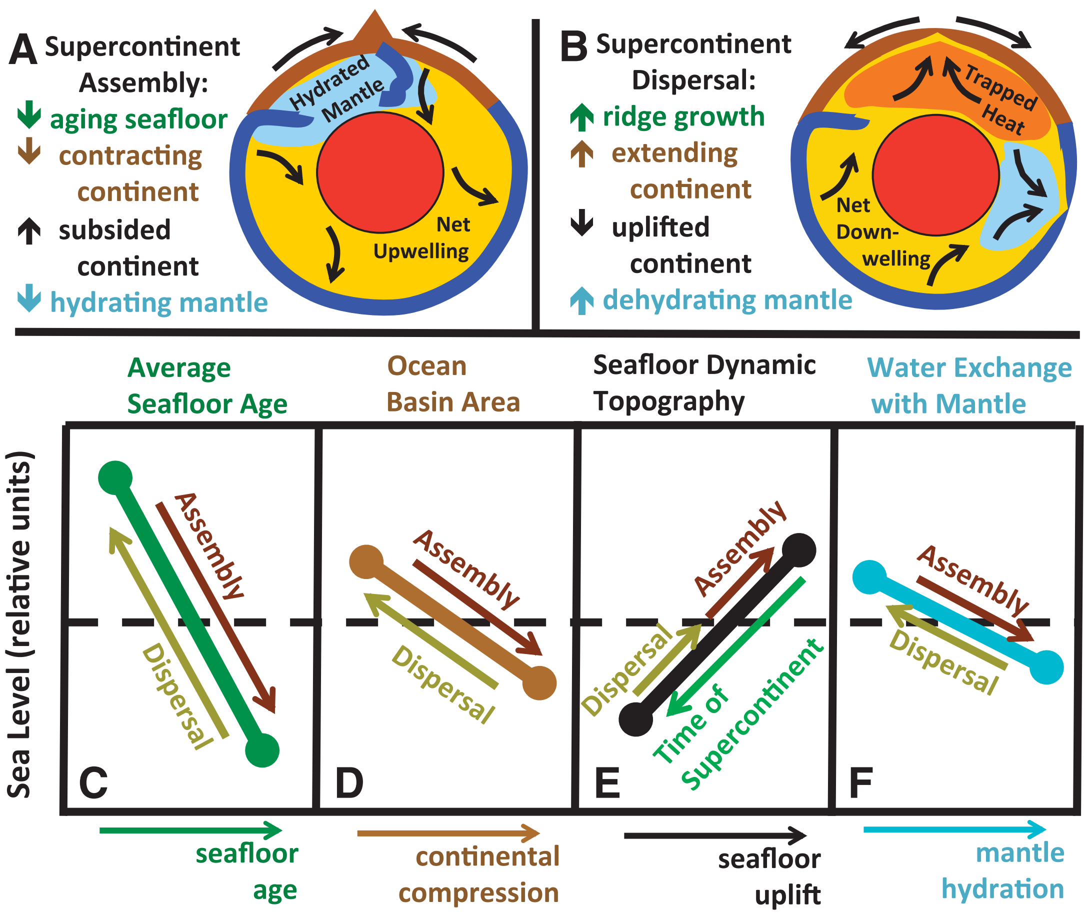


**Figure 11. Cycles of plate tectonics and mantle convection during the assembly, stability, and dispersal of supercontinents, and the eustatic sea-level trends thought to accompany them. Supercontinent assembly may be associated with older-than-average seafloor (Cogné and Humler, 2008), continents that are shortened by collision, and mantle that is becoming hydrated by subduction of old oceanic lithosphere (A). All of these factors tend to induce sea-level drop (to varying degrees) during the assembly phase of the supercontinental cycles, and their reversal during dispersal causes sea-level rise (C, D, F). Because supercontinents form over subduction and downwelling (A), net upwelling occurs beneath the seafloor, elevating sea level (Conrad and Husson, 2009). By contrast, a fragmenting supercontinent is elevated at the end of its life (B), while the seafloor and sea level are depressed (E). These sea-level factors are illustrated in cartoon diagrams of supercontinent formation (A) and dispersal (B), with their sea-level implications indicated by an upward arrow for sea-level rise and a downward arrow for sea-level drop. The responses of sea level to variations in these four factors are illustrated in relative terms in C–F, with the approximated phase of the supercontinent (assembly, time of supercontinent, and dispersal) indicated by arrows. The relative amplitudes of these sea-level responses, although poorly constrained, are indicated by the vertical extent of the trend lines shown.**

supercontinental breakup (Gurnis, 1988, 1990b; Lowman and Jarvis, 1999). Isostatic compensation of this heated subcontinental mantle may thus elevate a supercontinent relative to the surrounding oceans, lowering sea level during periods of supercontinental stability. However, Worsley et al.'s (1984) ideas about thermal support of continents do not account for flow of the viscous mantle, which must accompany these predicted redistributions of mantle heat. As discussed earlier herein, and first demonstrated for supercontinental breakup by Gurnis (1990b), this induced mantle flow dynamically deflects Earth's surface, resulting in net uplift or subsidence of the seafloor (Fig. 9), and thus sea-level change (Conrad and Husson, 2009). Indeed, recent numerical models of linked mantle flow and plate tectonics have begun to predict vertical motions of continents over time scales relevant to supercontinental cycles (e.g., Flowers et al., 2012).

Conrad and Husson (2009) suggested that the supercontinental cycle creates time-varying dynamic topography on the seafloor that drives eustatic sea-level change. This process is illustrated by comparing the mantle structures expected beneath an aggregating supercontinent (Fig. 11A) with those expected beneath a dispersing supercontinent (Fig. 11B). In particular, we note that subduction is necessary for ocean basin closure (Zhong et al., 2007b), which means that supercontinent formation occurs above mantle downwelling, which on average tends to depress continental elevations. Because the net dynamic deflection of Earth's surface must be zero, uplifted seafloor must accompany depressed continents, which implies elevated sea level at the time of continental aggregation. By contrast, mantle upwelling is expected beneath supercontinental breakup (e.g., Gurnis, 1988; Lowman and Jarvis, 1999), which implies positive dynamic topography over a dispersing supercontinent. This net uplift of a supercontinent occurring during its lifetime must be balanced by net subsidence of the seafloor, resulting in sea-level drop (Fig. 11E). The reverse process drives sea-level rise during the dispersal and assembly phases, when continental motion down dynamic topography gradients exposes more of the seafloor to regions of positive dynamic topography, as noted previously for recent trends. The amplitude of sea-level changes during this cycle may be estimated from the current offset of sea level by dynamic topography, which Conrad and Husson (2009) estimated to be ~100 m for the current relatively dispersed configuration of continents. Thus, dynamic topography may impact sea level nearly as much as ridge volume changes during a supercontinental cycle, but it should induce a different timing of highstands and lowstands within that cycle (Fig. 11).

### Combined Contributions over Supercontinental Cycles

Conrad and Husson's (2009) prediction that dynamic topography should cause sea-level rise during supercontinent dispersal and assembly, and sea-level drop during stability (Fig. 11E), correlates with some of the observed Phanerozoic sea-level trends (Fig. 2C). In particular, it predicts the period of sea-level rise during the most recent Pangean dispersal (ca. 200 Ma to ca. 75 Ma), but much of the corresponding sea-level drop occurred while Pangea was being assembled (ca. 450–325 Ma) instead of solely during Pangean stability (Fig. 2C). However, ridge volume and ocean basin area changes should cause sea level to drop during supercontinent assembly (Figs. 11C–11D), which may mask the sea-level impact of dynamic topography. Indeed, the combination of all three factors (seafloor age, ocean area, and dynamic topography) should concentrate sea-level rise into the dispersal phase (when all three factors contribute; Figs. 11C–11E) but spread the corresponding sea-level drop over both the supercontinent assembly and stability phases. Indeed, sea level may have risen more rapidly during dispersal (~2 m/m.y. between 200 and 75 Ma; Fig. 2C) than it fell during assembly and stability (~1–2 m/m.y. between 450 and 250 Ma; Fig. 2C), although it is difficult to compare such long-term rates of sea-level change for different periods of the stratigraphic record. Additionally, the mantle dynamics and surface tectonics of supercontinental cycles are complicated enough (Collins, 2003) that their sea-level impact may differ between cycles. Furthermore, supercontinental cycles are long enough that factors affecting Earth's thermal and chemical evolution (such as ocean-mantle water cycling, see "Water Exchange over Supercontinental Cycles" subsection in following section) may affect sea-level trends during supercontinental cycles.

## OCEAN-MANTLE WATER EXCHANGE AND SEA-LEVEL CHANGE OVER EARTH HISTORY

Although the changing "container" volume of the ocean basins dominated eustatic sea-level change since the Cretaceous (Fig. 10), and likely did so during the past supercontinental cycle (Fig. 11), the solid Earth's control on Earth's seawater volume may dominate sea level over the longest time scales associated with Earth's planetary evolution. Much of Earth's water may be currently stored in hydrated minerals within Earth's mantle (Hirschmann, 2006; Ohtani, 2005; Williams and Hemley, 2001), particularly within the transition zone (Inoue et al., 2010; Karato, 2011). Indeed, the mantle interior likely contains a volume of water that exceeds that of the surface oceans (Hirschmann and Kohlstedt, 2012), possibly by an order of magnitude (Marty, 2012). This mantle water directly influences mantle dynamics, and thus sea level as well via dynamic topography or subduction initiation, by decreasing mantle viscosity (Hirth and Kohlstedt, 1996). Indeed, some mantle upwellings may be arising from regions of elevated water content above ancient subducted slabs (Spasojevic et al., 2010; van der Lee et al., 2008). Water from mantle rocks is released into the surface environment by degassing at mid-ocean ridges (Jambon and Zimmermann, 1990; Ligi et al., 2005; Michael, 1995; Shaw et al., 2010) and is replenished by subduction of hydrated minerals (Faccenda et al., 2012; Hacker, 2008; Jarrard, 2003; Rüpke et al., 2004; Savage, 2012; Shaw et al., 2008; van Keken et al., 2011). If these two processes are not balanced, then the volume of water in Earth's surface environment will change (Crowley et al., 2011; Franck and Bounama, 1995; Korenaga, 2011; McGovern and Schubert, 1989; Sandu et al., 2011), as will sea level (Bounama et al., 2001; Hirschmann, 2006; Kasting and Holm, 1992).

### Water Exchange over Supercontinental Cycles

Because degassing rates at spreading ridges depend on the lithosphere production rate and on the depth of melting (Hirose and Kawamoto, 1995; Hirschmann et al., 2009), periods of anomalously rapid seafloor spreading (e.g., the Cretaceous; Fig. 5) should correspond to periods of increased water release from the mantle interior. Although the rate of seafloor loss to subduction also increases during such periods (to preserve seafloor area), the rate of water transfer back into the deep interior will not always balance degassing. This is because only a small fraction of subducted water is actually transported beyond the subduction-zone backarc and into the deep mantle (Hacker, 2008; Jarrard, 2003; Schmidt and Poli, 2003). Instead, van Keken et al. (2011) showed that most water transfer into the deep mantle occurs via rapid subduction of old lithosphere (e.g., in the western Pacific today; Iwamori, 2007), while young, hotter seafloor becomes dehydrated in the mantle wedge. Thus, time periods with significant subduction of old lithosphere (e.g., recently [Fig. 6D] and the Early Cretaceous [Fig. 6F]) should see unusually high rates of water transfer into the deep interior and thus should represent periods of eustatic sea-level drop. By contrast, time periods dominated by subduction of primarily young lithosphere (e.g., the Late Cretaceous [Fig. 6E]) should see a net transfer of water out of the mantle and eustatic sea-level rise. Thus, within the supercontinental cycle (Figs. 11A and 11B), we expect a net flux of water out of the mantle during periods of younger average seafloor age (dispersal phase; Fig. 11C) and flux back into the mantle during periods of older seafloor age (assembly phase; Fig. 11C). Water exchange between the mantle and the oceans should thus result in a supercontinental cycle of rise and drop (Fig. 11F) that amplifies the cycle expected from ridge volume change (Fig. 11C).

How rapidly can mantle dehydration and rehydration change sea level? van Keken et al. (2011) estimated, with admitted uncertainty, that slabs currently transfer ~$3.2 \times 10^8$ Tg/m.y. of water into the deep mantle, which corresponds to roughly ~1 m/m.y. of eustatic sea-level drop if unbalanced by mantle degassing. This estimate is within a factor of ~2 of other recent estimates (Hacker, 2008; Savage, 2012), although some authors have estimated significantly smaller (Green et al., 2010) and greater (Rüpke et al., 2004; Schmidt and Poli, 2003) rates of mantle water transport. Because only the coldest subduction zones carry water deeper than ~230 km, van Keken et al.'s (2011) estimates indicate that over half (54%) of all water transport into the deep mantle occurs via subduction of oceanic lithosphere older than 100 m.y. Because subduction of such old lithosphere was negligible in the Late Cretaceous (Fig. 6E), we can infer that the rate of water loss per unit area of subducted seafloor was ~50% slower at 65 Ma than it is today. If seafloor production (and destruction by subduction) was ~20% faster at that time (Fig. 5), then the rate of water loss to the mantle at 65 Ma would correspond to ~$1.9 \times 10^8$ Tg/m.y., or ~60% of the present-day rate. These estimates for the rate of subduction water loss correspond to deviations of ~30% about their mean value of $2.9 \times 10^8$ Tg/m.y., or ~0.25 m/m.y. Based on the time dependence of the seafloor age distribution (Figs. 6D–6F), these rates imply eustatic sea-level rise of ~5–10 m during the first half of the Cenozoic, followed by a comparable sea-level drop in the second half. Such values would not significantly affect the total sea-level budget since the Cretaceous (Fig. 10). Within an ~400 m.y. supercontinental cycle, however, the total variations may be up to ~20–40 m (Fig. 11F), or comparable to the variations associated with changes in seafloor area (Fig. 11D). However, this estimate depends on the assumed rate of water loss to the mantle and its dependence on subducting seafloor age, both of which are uncertain.

### Water Exchange over Earth History

The rate of water outgassing at ridges might not balance the rate of water loss to the mantle over the longest geologic time scales. For example, as Earth cools during its billion-year thermal evolution, water transport into the mantle should accelerate as subducted slabs get colder (Franck and Bounama, 1995; McGovern and Schubert, 1989). Additionally, mid-ocean-ridge melting zones should become shallower (Rüpke et al., 2004), causing the degassing rate to slow (Sandu et al., 2011). Both mechanisms tend to cause a net drain of ocean water into the mantle interior as Earth cools. The resulting hydration of the mantle leads to a decrease in mantle viscosity (Hirth and Kohlstedt, 1996), which counterbalances the increase in viscosity associated with mantle cooling (Crowley et al., 2011). Thus, a slow water loss to the mantle tends to stabilize plate tectonics over geologic time (Korenaga, 2011), and may help to reconcile Earth's present-day heat flow with thermal evolution models (Crowley et al., 2011). These links between the thermal and hydration states of the mantle are consistent with those expected to occur within the supercontinental cycle; water accumulates within a cooling mantle during supercontinent assembly (Fig. 11A), and a warming mantle dehydrates during supercontinent dispersal (Fig. 11B).

Because the balance between mantle degassing and regassing is intimately connected with Earth's thermal evolution (Crowley et al., 2011; Korenaga, 2011), the time-history of water exchange with the mantle depends on a variety of uncertain parameters. These include factors that control plate-tectonic spreading rates, melting rates and water partitioning into melt at ridges, slab dehydration patterns at subduction zones, the dependence of mantle rheology on temperature and hydration (Sandu et al., 2011), and the dynamics of water circulation within the mantle (Fujita and Ogawa, 2013). As a result, the rate and even the duration of long-term eustatic sea-level drop associated with Earth's thermal evolution remain poorly constrained. Thermal evolution models by Rüpke et al. (2004), for example, estimated that the mantle has been rehydrating for the past 1–3 b.y., depending on the degree of slab serpentinization, and constrained Phanerozoic sea-level drop to 0.5–1.7 m/m.y. Korenaga (2011) suggested slower sea-level drop of ~0.5 m/m.y., but occurring since the Archean. Sandu et al. (2011) estimated that the mantle has been rehydrating for a few billion years, but that sea-level change depends on the total volume of water in the Earth-mantle system.

For the Phanerozoic, sea-level change predictions based on thermal evolution models are likely more poorly constrained than the sea-level observations themselves. For example, some authors have noted the apparent ~100 m drop in sea level between the last two Phanerozoic highstands (e.g., Fig. 2C) (Hallam, 1992) and used 0.2 m/m.y. as an upper bound on Phanerozoic sea-level drop in thermal evolution models (Parai and Mukhopadhyay, 2012). This mantle hydration rate yields at most 20 m of sea-level drop since the mid-Cretaceous, and thus does not contribute significantly to the sea-level budget since that time (Fig. 10). Prior to the Phanerozoic, however, geological constraints on sea level are more ambiguous. Maruyama and Liou (2005) appealed to a variety of observations to suggest several hundreds of meters of sea-level drop immediately prior to the Phanerozoic. On the other hand, Eriksson (1999) suggested that crustal volume, continental isostasy, erosion, and plate-tectonic processes interact to maintain relatively constant freeboard, and Eriksson et al. (2006) found only minor deviations from this hypothesis by examining the volcanic-sedimentary record of several cratons. Although such observations seem to imply long-term stability of Earth's surface reservoirs (Hirschmann, 2006), they are not inconsistent with ~0.2 m/m.y. of sea-level change if this change occurred first as Archean degassing followed by Proterozoic regassing, as some models suggest (Maruyama and Liou, 2005; Rüpke et al., 2004; Sandu et al., 2011). On the other hand, more rapid and sustained water loss to the mantle may be consistent with constant freeboard if the associated sea-level drop is approximately balanced by eustatic sea-level rise caused by growth in the area of continental lithosphere (Eriksson et al., 2006). However, because epeirogeny has likely affected continents since their origin, observations of Precambrian relative sea-level change suffer the same uncertain relationship with eustatic change (Flowers et al., 2012) as observations of change at all other time scales. Such arguments highlight the difficulty of constraining sea-level change occurring prior to the Phanerozoic.

## CONCLUDING REMARKS

Earth's water originally accumulated during planetary accretion (Wood et al., 2010) and/or soon afterward (Albarede, 2009) by delivery from asteroids (Morbidelli et al., 2000) and comets (Hartogh et al., 2011). Ever since, a changing portion of this water has been stored within our planet's surface oceans. The upper surface of these oceans defines sea level, and its elevation relative to our planet's land surface is constantly changing. We have observed sea-level change occurring at rates as slow as meters per million years for the Phanerozoic (Fig. 2C), and three orders of magnitude faster (mm/yr) during the past century (Fig. 2A). Here, I have shown that the solid Earth exerts a primary influence on this observed sea-level change on all time scales via a variety of processes associated with planetary cooling, mantle convection, plate tectonics, volcanism, sedimentation, and rock deformation (Fig. 1). In this work, I have discussed how advances in our knowledge of these solid Earth processes during the past quarter century have led to greater understanding of sea-level change throughout Earth history. Despite these advances, there remains ample room for progress during the coming decades.

On the longest time scales ($10^9$ yr and longer), sea level is governed by changes to the partitioning of Earth's total water reserves between the surface and the mantle interior. In particular, water exchange with the mantle, which is modulated by Earth's evolving thermal state, can cause seawater to drain into, or become exhumed from, the mantle interior (Fig. 1C). Thus, the portion of Earth's water that resides outside of the planetary interior is ultimately controlled by convection within the solid Earth (Crowley et al., 2011; Sandu et al., 2011), and changes to this seawater volume can sustain eustatic sea-level change of a few meters per million years for billions of years (Korenaga, 2011; Rüpke et al., 2004). Such rates are large enough to drain the ocean basins completely in the future (Bounama et al., 2001). However, geologic constraints on sea-level change over billions of years are few and disputed, and independent constraints on current fluxes of water into the subduction zones and out of the ridges are uncertain enough that it is difficult to constrain the net direction of flow even for the present day (Sandu et al., 2011). Thermal evolution models that include hydrologic cycling tend to predict sea-level drop during the past few billion years, but even these are sensitive to parameters with that are not well constrained. Although it seems unlikely that new constraints on Archean or Proterozoic sea level will become available, more advanced understanding of Earth's thermal evolution, both from modeling and additional constraints on Earth's internal structure and material properties, should help advance our understanding of sea-level change over the longest time scales relevant to Earth history.

On intermediate time scales—$10^8$ to $10^6$ yr—solid Earth processes also govern sea-level change, but they do so by affecting the "container" volume of the ocean basins (Fig. 1C). Indeed sea-level observations for the past ~100 m.y. can be matched rather well by estimating the influence of several of these solid Earth influences and summing them (Fig. 10). Most important to this summation is water displacement resulting from changes to the volume of Earth's mid-ocean-ridge system. In particular, plate-tectonic changes to the average age of the seafloor can raise or lower eustatic sea level by hundreds of meters over time scales as short as ~10 m.y. (Fig. 4), but only if spreading rates change globally and dramatically. However, recent advances to the sophistication of plate-tectonic reconstructions (Müller et al., 2008b; Torsvik et al., 2010) suggest that, at least for the Cenozoic and Cretaceous, Earth's ridge system evolves over slower time scales of ~100 m.y. (Figs. 5 and 6). Moreover, there is little evidence in the plate-tectonic record that "second-order" variations in sea level occurring on time scales of ~10–50 m.y. (Figs. 2C and 10) are caused by short-term variations in plate-tectonic spreading rates. Indeed, these short-term variations with amplitudes of 30–50 m are difficult to explain in terms of solid Earth processes that drive eustatic change. These time scales and magnitudes are, however, relevant for ground motion due to changing dynamic topography (Lovell, 2010), which suggests that some of these second-order variations may be due to local uplift or subsidence at individual measurement locations.

Changes in ridge volume represent changes in mantle heat flow, and are therefore likely tied to supercontinental cycles of assembly and dispersal, which are a tectonic response to convective cycling of mantle heat. Therefore, I have inferred a relationship between sea level and supercontinental phase (Fig. 11C) based on the most recent supercontinental dispersal for which there is a reliable tectonic reconstruction. We cannot, however, fully verify this relationship even for the most recent cycle, because, although tectonic reconstructions of the seafloor have recently been extended back through the Jurassic (Seton et al., 2012), we do not have a reliable tectonic reconstruction of the seafloor during times of Pangea assembly or stability. Although the tectonic history of this seafloor has been lost, information about it may still reside in the continental margins under which it subducted, or in the density structures of the mantle into which it descended. Thus, further constraints on the Panthalassic seafloor, and the sea-level response to supercontinental cycling, may be gleaned from future geological and seismological studies of these regions. First attempts indicate evidence for intra-Panthalassic subduction zones during Pangean stability (van der Meer et al., 2012), which is consistent with a smaller ridge system and lower sea level prior to supercontinent dispersal (Fig. 11C).

Accumulated sediments and volcanic debris on the seafloor also displace seawater, and changes in the volume of this material affect sea level (Fig. 1C). Compared to ridge volume changes, however, it is even more difficult to estimate past volumes for this material because emplacement patterns (Fig. 8) are more complicated and less predictable. As a result, the sea-level impact of sediments and (especially) volcanism for Cretaceous and earlier times is poorly constrained. Fortunately, volume changes for such material are likely smaller than those for the ridges by a factor of three or more (Fig. 7), which implies that their long-term impact on sea level may be similarly less important. However, it is not clear how the processes that drive sedimentation and intraplate volcanism might respond to a dramatically different configuration of continents, and we cannot rule out the possibility that systematic differences prior to the Cretaceous might have impacted sea level significantly. For example, I estimated that oceanic plateau emplacement resulted in 100 m of sea-level rise during the Cretaceous (Fig. 7), which would be the largest contributor to sea-level change during this period (Fig. 10). This estimate, however, is admittedly rather uncertain (Fig. 7), and it is not known whether the volcanic events that produced this rise are unusual within supercontinental cycles. Such questions can only be answered by gaining a more thorough understanding of the driving mechanisms behind sedimentation and intraplate volcanism, so that modeling efforts for past times can predict long-term trends more confidently.

Modeling efforts to understand how changes in rates of intraplate volcanism and seafloor spreading, and even continental expansion and contraction, respond to convection in Earth's mantle must examine the link between the mantle dynamics and patterns of plate tectonics at the surface (Zhang et al., 2010; Zhong et al., 2007b). Such efforts are beginning to also predict patterns of epeirogenic uplift and subsidence that results from this flow (e.g., Fig. 9), and as a result can be used to estimate the influence of net deflection of the seafloor on sea level as a function of time (Liu et al., 2008; Spasojevic and Gurnis, 2012; Spasojevic et al., 2009). Using a kinematic model of Cretaceous and Cenozoic continental motions (Torsvik et al., 2010), I have introduced an estimate of the sea-level impact of such deflection here (Fig. 10), but more sophisticated future modeling that includes full dynamics should yield better predictions over longer stretches of Earth history in the coming decades, and indeed are already beginning to do so (Spasojevic and Gurnis, 2012). Furthermore, such models make predictions about vertical continental motion, and thus can be tested against, and constrained by, geological observations of past uplift and subsidence based on sediment stratigraphy and thermochronometry (e.g., Flowers et al., 2012; Spasojevic et al., 2009; Zhang et al., 2012). Such models will become more sophisticated and more useful in the next few decades.

The recognition that continents move vertically in response to changing dynamic support of topography by mantle convection has significantly complicated the interpretation of sea-level change curves (e.g., Moucha et al., 2008). Stratigraphic constraints on sea-level change thus cannot be interpreted directly as arising directly from eustatic change, but instead they must be recognized as a measure of relative sea-level change between the sea surface and the land surface, both of which may be moving vertically (Gurnis, 1992). Indeed, some observational records (e.g., Miller et al.'s [2005] record from the New Jersey margin) have been shown to contain several meters per million years of vertical continental motion (Müller et al., 2008b; Spasojevic et al., 2008) in addition to eustatic change (Fig. 10). Because all continents are likely to experience at least some vertical motion at all times, it is becoming increasingly clear that records of eustatic change should not be considered in isolation. Instead, records of transgressions and regressions from different regions of the world should be intercompared, together with additional observations and models of continental vertical motion, in order to constrain global eustatic change (Ruban et al., 2010b, 2012; Zorina et al., 2008). Such efforts necessarily involve interdisciplinary and collaborative efforts among the fields of sedimentology, paleomagnetics, paleobiology, geodynamics, and tectonics, and they remain a major challenge for the future (Ruban et al., 2010a).

On the shortest time scales of $10^5$ yr and shorter, water exchange with the cryosphere, which has been modulated by climatic influence since the onset of major glaciation at the Eocene-Oligocene transition (and likely during previous cool periods of Earth history), dominates eustatic sea-level change (e.g., Miller et al., 2005). Yet, even this change is intimately linked to solid Earth dynamics: The unloading of continental ice masses into the oceans causes both instantaneous (elastic; Fig. 1A) and continuing (viscous; Fig. 1B) deformations of the solid Earth that deflect both the land and sea surfaces vertically (e.g., Clark and Lingle, 1979; Farrell and Clark, 1976). Like the sea-level variations induced on much longer time scales by dynamic support of topography by mantle flow, these deformation-induced spatial variations in sea-level change can be larger in magnitude than globally averaged eustatic change (Fig. 3). The result is a solid Earth "fingerprint" imprinted on the surface expression of sea-level change (e.g., Bamber and Riva, 2010; Clark et al., 2002; Douglas, 2008; Mitrovica et al., 2011). Interpreting these fingerprints will become increasing important in the coming decades, both for constraining the magnitude and patterns of Pleistocene deglaciations (Kopp et al., 2009; Lambeck et al., 2012; Raymo et al., 2011), and for predicting the spatial variability of current and future sea-level change to which human society must adapt (Gomez et al., 2010; Mitrovica et al., 2009; Sallenger et al., 2012; Willis and Church, 2012).

The solid Earth thus impacts sea level across all time scales (Fig. 1), and future advances in our understanding of past—and future—sea-level change will be gained by considering the influence of solid Earth dynamics. On the longest and shortest time scales, sea-level change reflects the changing volume of water in the ocean basins, and solid Earth dynamics either govern this volume or adjust to its mass redistribution on Earth's surface. On geological time scales, solid Earth deformations govern sea level by changing the shape of the ocean basins that contain Earth's water. On all time scales, the solid surface of Earth—against which sea level is measured—is in constant vertical motion. Although this fact complicates the interpretation of sea-level observations, it also provides us with new opportunities to use observations of the spatial variability of sea-level change to constrain a variety of climatological, geological, tectonic, and volcanic processes that affect both sea level and ground surface motions. For all time scales, the exploitation of this spatially varying record of sea level will present both opportunities and challenges to future sea-level scientists.

## ACKNOWLEDGMENTS

I thank Brendan Murphy for his invitation to contribute this review article to the Geological Society of America's 125th anniversary celebration, and Thorsten Becker, Michael Gurnis, Laurent Husson, Carolina Lithgow-Bertelloni, Jerry Mitrovica, Laurent Métivier, Dietmar Müller, Greg Ravizza, Dmitry Ruban, Trond Torsvik, Paul Wessel, and Shijie Zhong for discussions that helped inspire portions of this project. I especially thank Michael Gurnis and Shijie Zhong for constructive reviews that significantly improved the manuscript. This work was supported by National Science Foundation grant EAR-1151241.

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Science Editor: J. Brendan Murphy

Manuscript Received 22 July 2012
Revised Manuscript Received 25 February 2013
Manuscript Accepted 25 February 2013

Printed in the USA