Iceland, the Farallon slab, and dynamic topography of the North Atlantic

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ABSTRACT

Upwelling or downwelling flow in Earth's mantle is thought to elevate or depress Earth's surface on a continental scale. Direct observation of this "dynamic topography" on the seafloor, however, has remained elusive because it is obscured by isostatically supported topography caused by near-surface density variations. We calculate the nonisostatic topography of the North Atlantic by correcting seafloor depths for lithospheric cooling and sediment loading, and find that seafloor west of the Mid-Atlantic Ridge is an average of 0.5 km deeper than it is to the east. We are able to reproduce this basic observation in a model of mantle flow driven by tomographically inferred mantle densities. This model shows that the Farallon slab, currently in the lower mantle beneath the east coast of North America, induces downwelling flow that deepens the western North Atlantic relative to the east. Our model also predicts dynamic support of observed topographic highs near Iceland and the Azores, but suggests that the Icelandic high is due to local upper-mantle upwelling, while the Azores high is part of a plate-scale lower-mantle upwelling to the south. An anomalously deep area off the coast of Nova Scotia may be associated with the downwelling component of edge-driven convection at the continental boundary. Thus, several of the seafloor's topographic features can only be understood in terms of dynamic support from flow in Earth's mantle.

Keywords: dynamic topography, seafloor depth, mantle flow, North Atlantic, Iceland, Azores, Scotian Basin.

INTRODUCTION

Dynamically supported deflections of Earth's surface are predicted as a consequence of viscous mantle flow (e.g., Gurnis, 1993; Hager et al., 1985; Mitrovica et al., 1989), but poorly constrained and isostatically supported crustal thickness variations obscure this dynamic topography and make its detection difficult, particularly for the seafloor (Colin and Fleitout, 1990). Dynamically supported topography has been identified in southern Africa, where lower-mantle upwelling lifts the surface by ~1 km (Lithgow-Bertelloni and Silver, 1998). Continental-scale uplift and subsidence events in the geologic record have also been attributed to the motion of continents over regions of dynamic topography associated with mantle upwelling (Conrad and Gurnis, 2003) or slab-induced downwelling (e.g., Gurnis, 1993; Mitrovica et al., 1989; Pysklywec and Mitrovica, 1998). Although whole-mantle flow models predict more than 1 km of dynamic topography (e.g., Hager et al., 1985; Ricard et al., 1993), these predictions have been controversial because they are sensitive to the manner in which plate motions are implemented (Cadek and Fleitout, 1999; Thoraval and Richards, 1997) and to the way in which slabs are attached to plates (Zhong and Davies, 1999). These modeling uncertainties, coupled with uncertain estimates of Earth's nonisostatic topography field, have made comparisons between predicted and observed dynamic topography difficult.

Because they are younger than 180 Ma everywhere, Earth's ocean basins have been subjected to topography-producing tectonic and erosional processes for a shorter time than have the continents. As a result, estimation of nonisostatic topography is potentially more straightforward in the oceans, making them a natural place to look for dynamic topography. Although most of Earth's seafloor bathymetry is explained by cooling of the oceanic lithosphere with age (Stein and Stein, 1992), an isostatic correction for crustal and sediment thickness must also be made to completely remove isostatic contributions to topography (Crough, 1983). Previous estimates of the seafloor's residual topography (e.g., Colin and Fleitout, 1990; Kido and Seno, 1994) use coarse sediment thickness estimates that do not have the same resolution as recent bathymetric maps (Smith and Sandwell, 1997). Furthermore, estimates of residual seafloor topography have not been carefully compared to predictions of dynamic topography made using updated detailed models of the mantle's seismic velocity structure (e.g., Grand, 2003; Ritsema et al., 1999).

We compare new estimates of residual topography in the North Atlantic to predictions of dynamic topography made using two recent models for mantle density heterogeneity. The North Atlantic is a good place to make such a comparison for several reasons. First, new estimates of North Atlantic sediment thickness allow for improved estimates of residual topography (Louden et al., 2001). Second, the modeling uncertainties associated with subduction zones (Zhong and Davies, 1999) and surface plate motions (Cadek and Fleitout, 1999; Thoraval and Richards, 1997) should be minimal because the North Atlantic is subduction free and plate velocities are slow. Third, the Farallon slab, which may have caused dynamically induced subsidence of the Western Interior Seaway during the Cretaceous (Mitrovica et al., 1989), is currently in the lower mantle beneath eastern North America (Bunge and Grand, 2000). Downwelling caused by this slab should produce downward tilting of the western North Atlantic seafloor.

ESTIMATING RESIDUAL SEAFLOOR TOPOGRAPHY

Estimated residual seafloor depths for the North Atlantic are taken from recent estimates made by Louden et al. (2001). In this study, the GDH1 model of seafloor depth as a function of age (Stein and Stein, 1992) is removed from the satellite-derived bathymetry of Smith and Sandwell (1997). Isostatically adjusted sediment thicknesses are removed (following Crough, 1983) and crustal thickness variations, which are typically <200 km in wavelength, are smoothed using a 200 km mean value filter. Although some of the resulting residual topography (Fig. 1A) is associated with locally supported ocean-island volcanism, the longer wavelength features (>300 km) are unexplained by isostatic processes. These include broad highs centered around Iceland and the Azores, and a low off the coast of Nova Scotia (Fig. 1A). On an ocean-wide scale, seafloor west of the Mid-Atlantic Ridge is deeper than seafloor to the east by an average of 557 m south of the Charlie-Gibbs fracture zone (52°N).

PREDICTING DYNAMIC TOPOGRAPHY

We determine dynamic topography from a degree 20 spherical harmonic solution to the equations of conservation of mass and momentum for incompressible fluid flow in the

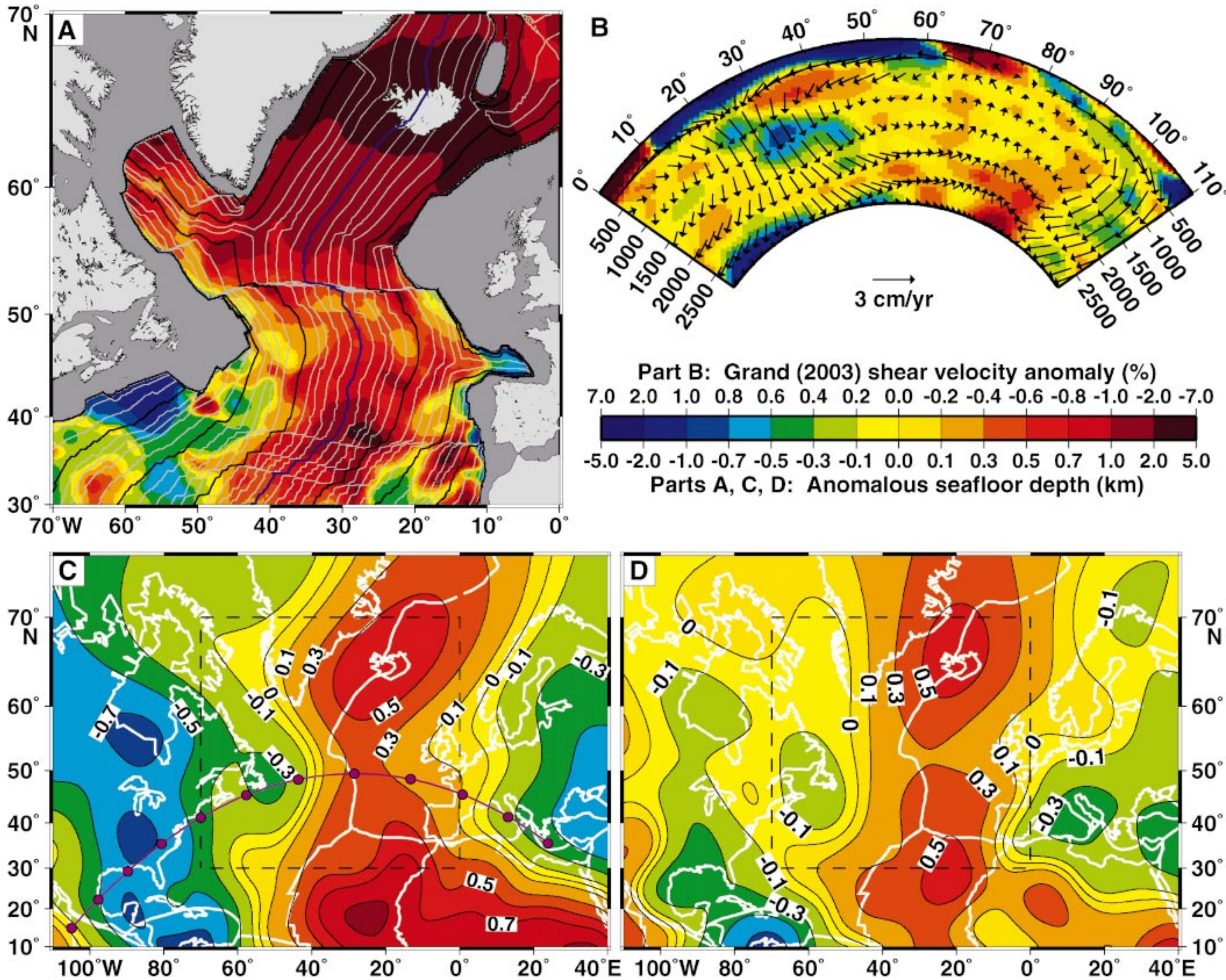

Figure 1. Comparison of (A) residual (nonisostatic) seafloor topography with dynamic topography predicted by models of mantle flow. In A, Mid-Atlantic Ridge (blue) and seafloor age contours at 10 and 50 m.y. intervals (gray and black) are shown for reference. Seismic shear velocity anomalies (colors) from Grand (2003) tomography that are used to drive mantle flow (arrows) are shown in cross section (B, path shown in C). Dynamic topography caused by this flow (C), and that caused by flow determined using Ritsema et al. (1999) tomography (D), can be compared directly with observed topography (A) by examining dashed areas of predicted fields (C and D).

mantle (Hager and O'Connell, 1981). Instantaneous flow is driven by density heterogeneities determined from the seismic shear-wave velocity models of Grand (2003) and Ritsema et al. (1999). We determine densities using a velocity to density conversion factor of 0.3 g cm^{-3} km^{-1} s, which is within the range indicated by mineral physics estimates (Karato and Karki, 2001) and used by other studies of the geoid and dynamic topography (e.g., Conrad and Gurnis, 2003; Lithgow-Bertelloni and Silver, 1998; Thoraval and Richards, 1997). Because near-surface tomography may be partially determined by isostatically compensated compositional differences, we initially use only density heterogeneities below 325 km to drive flow (Lithgow-Bertelloni and Silver, 1998). This allows us to predict long-wavelength dynamic topography associated with structures in the deep mantle. We elevate this cutoff depth in tests to examine the local effect of near-surface density heterogeneity on short-wavelength dynamic topography, remembering that these predictions are only valid for isostatically uncompensated density anomalies. Following recent global models used to predict the geoid and dynamic topography, we use free slip boundary conditions at the surface and base of the mantle, a 130-km-thick lithosphere that is 10 times the upper-mantle viscosity, and a lower mantle below 670 km with a viscosity 50 times that of the upper mantle (e.g., Lithgow-Bertelloni and Silver, 1998; Ricard et al., 1993). The resulting solution for mantle flow (Fig. 1B) also predicts radial stresses at the surface. The dynamic topography produced by these stresses (Figs. 1C and 1D) can be obtained by dividing by the gravitational acceleration and the surface density contrast, which we take to be 2300 kg m^{-3} here for the water-lithosphere interface (e.g., Hager et al., 1985). Because dynamic topography contributes to the geoid, we calculate the predicted geoid using the Grand (2003) and Ritsema et al. (1999) models, and find variance reductions of 61% and 59%, respectively. Although better fits to the geoid can be obtained for layered flow models (e.g., Cadek and Fleitout, 1999), our model fits the geoid as well as the best-fitting unlayered flow models (e.g., Thoraval and Richards, 1997).

DYNAMIC TOPOGRAPHY OF THE NORTH ATLANTIC

We compare predictions of dynamic topography (Figs. 1C and 1D) to the residual seafloor topography (Fig. 1A).

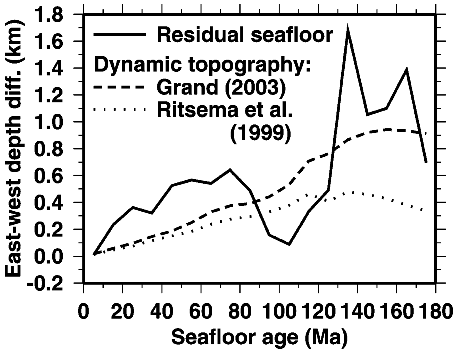

Figure 2. Depth differences between eastern and western sides of Mid-Atlantic Ridge for residual seafloor topography (solid line) and predictions of dynamic topography made using two tomographic models (dashed, dotted lines). We difference average elevation of equally old seafloor on both sides of ridge within 10 m.y. age bands (see Fig. 1A) and between latitudes of 30° and 52°N. Result shows significant deepening of North American side of Atlantic relative to European side for both residual topography and predicted dynamic topography.

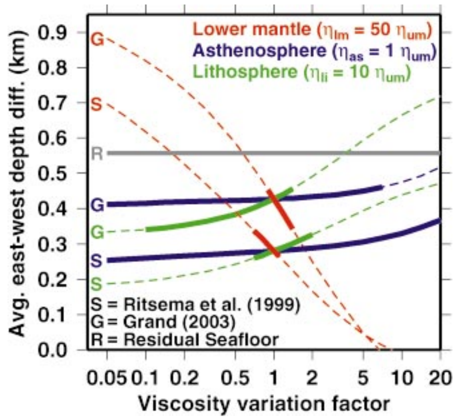

Figure 3. Average differences in predicted dynamic topography between eastern and western sides of Mid-Atlantic Ridge and between 30° and 52°N for different radial viscosity structures. Results for Grand (2003) and Ritsema et al. (1999) tomography models are compared to average residual depth difference of 557 m (gray line). For each line, viscosity (η) of one layer is varied by given factor (*x*-axis) while others are held at their base level. Base-level viscosities are given as multiples of upper mantle viscosity (upper right), which is used as reference viscosity because neither geoid nor dynamic topography is sensitive to absolute mantle viscosity. Thick solid segments show flow fields that also reproduce observed geoid with 40% variance reduction or better, while thin dashed lines do not fit geoid to this level.

East-West Depth Differences

For the North Atlantic seafloor between 30° and 52°N, residual seafloor depths are asymmetrical about the Mid-Atlantic Ridge. Seafloor west of the ridge and older than 50 Ma is anomalously deep by 100–500 m or more while seafloor east of the ridge is, for the most part, anomalously shallow (Fig. 1A). When comparing seafloor of the same age on either side of the ridge, we find that residual depths to the west are consistently deeper than those to the east (Fig. 2). Because we compare seafloor of equal age on either side of the ridge, this asymmetry is insensitive to the age-depth relationship used to remove the effect of lithospheric cooling. Because we have removed topography due to variations in sediment thickness and filtered out most of the crustal thickness variations, the east-west asymmetry must be due to mantle processes such as hotspot volcanism or dynamic topography.

Our models for mantle flow beneath the North Atlantic show mantle downwelling occurring beneath both eastern North America and southern Europe (Fig. 1B). These downwellings are driven by seismically slow, and presumably dense, slabs (Fig. 1B) that have been identified as the subducted Farallon plate to the east (Bunge and Grand, 2000) and the former Tethyan seafloor to the west (Van der Voo et al., 1999). Because the Farallon slab is now beneath the east coast of North America (Fig. 1B), the downwelling it produces (Fig. 1B) causes negative dynamic topography in both the continental interior and the western North Atlantic (Fig. 1C). Because the Tethyan slab sinks beneath eastern Europe and Africa (Fig. 1B), it produces negative dynamic topography in the Mediterranean and eastward, but not in the Atlantic (Fig. 1C). The result is negative dynamic topography over the western North Atlantic but not over the eastern side (Fig. 1C), which is the basic pattern we observe in the residual seafloor depths (Fig. 1A). Because these two downwellings are asymmetrical about the Mid-Atlantic Ridge, they generate upwelling flow (Fig. 1B), and resulting dynamic topography (Fig. 1C), that is centered to the east of the ridge. The same basic long-wavelength patterns of dynamic topography are produced if mantle flow is driven by the Ritsema et al. (1999) model (Fig. 1D). Test calculations in which flow is driven only by structures deeper than 670 km confirm that the lower mantle is the source of the North Atlantic's topographic asymmetry. This finding assumes similar velocity to density scaling ratios for the upper and lower mantles, which may or may not be the case (Karato and Karki, 2001).

The amplitude of these flow-driven asymmetries in predicted dynamic topography is smaller than is observed (Fig. 2) by an average of 19% for the Grand (2003) and 47% for the Ritsema et al. (1999) models. The predicted depth difference across the Atlantic can be increased if the viscosity of the lower mantle is decreased or if that of the lithosphere is increased, but the geoid predicted by these models does not fit observations (Fig. 3). The inclusion of a low-viscosity asthenosphere (130–230 km) does not significantly affect dynamic topography (Fig. 3). We note, however, that our underprediction of dynamic topography is small given the uncertainty of several model parameters. For example, the velocity to density scaling parameter is uncertain to at least 50% (Karato and Karki, 2001), and a laterally varying mantle viscosity structure may induce more complicated flow patterns than the ones studied here. Although uncertainties in these parameters may affect the amplitude and geometry of dynamic topography, our prediction of the North Atlantic's topographic asymmetry should remain robust because it is caused by asymmetry in the location of mantle slabs with respect to the Mid-Atlantic Ridge.

Iceland

Our estimate of nonisostatic topography shows more than 1–2 km of positive residual topography centered around Iceland (Fig. 1A). This topographic high is also predicted by our models of dynamic topography, although predicted amplitudes are only ~0.6 km (Figs. 1C and 1D). Our underprediction of the Icelandic topographic high may be caused by our exclusive use of seismic structures deeper than 325 km. Tests show that the inclusion of the extremely slow velocities beneath Iceland between 175 km and 325 km can increase the Iceland topographic high to more than 1 km. Tests also show that although seismically slow anomalies are continuous beneath Iceland from the surface to the lower mantle (Ritsema et al., 1999), it is structure in the upper mantle that produces this topographic high. The fact that nearly half of the observed topography is due to upper-mantle structure below 325 km suggests that mantle flow associated with upwelling beneath Iceland extends through the upper mantle.

Azores

A second area of high residual topography occurs near the Azores Islands at ~38°N, east of the triple junction on the Mid-Atlantic Ridge, and is distinct from the Icelandic high to the north (Fig. 1A). The Azores high is close to, but slightly north of, similar local highs of dynamic topography predicted using both tomography models (Figs. 1C and 1D). In contrast to the Icelandic high, tests show that while the locations of these local high points are determined by upper-mantle density heterogeneity, their amplitudes arise from a broad, seismically slow, structure in the lower mantle that uplifts much of the African plate (Lithgow-Bertelloni and Silver, 1998). Several ocean island chains, including the Azores, Madeira, Canary, and Cape Verde Islands, have been hypothesized to result from plumes rising from this larger upwelling (Courtillot et al., 2003). Thus, the locally high seafloor associated with these ocean islands (Fig. 1A) may be uplift from shallow plumes that arise from broad upwelling flow in the deep mantle.

Scotian Basin

The seafloor in an ~800-km-wide region off the coast of Nova Scotia is anomalously deep by more than 1 km (Fig. 1A). Some of this anomaly may be caused by a cooler or thicker lithosphere than assumed in the average model of Stein and Stein (1992), although this is not in agreement with heat flow data (Louden et al., 1987). A similar, but shallower, basin is predicted for dynamic topography from both tomography models (Figs. 1C and 1D) and results from downwelling flow of a shallow seismically fast region in the upper mantle northeast of Nova Scotia (Fig. 1B, near 60°). This topographic low increases to more than 1 km depth in tests that include seismic structure between 325 and 175 km. Downwelling flow of this type may be driven by edge-driven convection of lateral temperature gradients at the continent-ocean boundary (King and Ritsema, 2000) that could contribute to the initiation of subduction (Zheng and Arkani-Hamed, 2002). Thus, dynamic topography associated with edge-driven convection may be a possible explanation for the Scotian basin.

CONCLUSIONS

Previous studies have suggested that whole-mantle flow models that successfully predict the geoid also predict dynamic topography with amplitudes greater than observed, indicating that some of this topography must be hidden on a density interface at depth (e.g., Cadek and Fleitout, 1999; Wen and Anderson, 1997). In contrast, our results show an underprediction of North Atlantic dynamic topography for whole-mantle flow models that successfully predict the geoid (Fig. 3). This suggests that, at least for downwelling flow beneath the North Atlantic, both the geoid and dynamic topography constraints can be satisfied without mantle layering. Global comparisons of residual seafloor depths and dynamic topography predicted by whole and partially layered mantle flow models should reveal other useful correlations. Such comparisons may require a more careful treatment of plate motions than performed here because most oceanic plates are moving faster than those in the North Atlantic and these plate motions directly affect predictions of the geoid and dynamic topography (Cadek and Fleitout, 1999; Thoraval and Richards, 1997).

We have shown that dynamic topography induced by mantle flow can explain many topographic features of the seafloor that cannot be explained by sediment loading or lithospheric cooling. Short-wavelength features (<1000 km) can be associated with convective flow of upper-mantle density anomalies that drive shallow convective processes such as localized upwelling (near Iceland) or edge-driven convective downwelling (offshore Nova Scotia). Dynamic topography may also affect the shape of entire ocean basins, which we observe as an east-west depth difference across the North Atlantic that is caused by downwelling of the Farallon slab beneath the North American side of the Atlantic.

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