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Revisiting tectonic corrections applied to Pleistocene sea-level highstands

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ABSTRACT

Tectonic displacement contaminates estimates of peak eustatic sea level (and, equivalently, minimum continental ice volumes) determined from the elevation of Quaternary interglacial highstand markers. For sites at which a stratigraphic or geomorphic marker of peak Marine Isotope Stage (MIS) 5e sea level exists, the standard approach for estimating local tectonic uplift (or subsidence) rates takes the difference between the elevation of the local highstand marker and a reference MIS 5e eustatic value, commonly chosen as +6 m, and divides by the age of the marker. The resulting rate is then applied to correct the elevation of all other local observed sea-level markers for tectonic displacement, including peak highstands of different ages (e.g., MIS 5a, MIS 5c and MIS 11), under the assumption that the tectonic rate remained constant over those periods. This approach introduces two potentially significant errors. First, the peak eustatic value adopted for MIS 5e in most previous studies (i.e., +6 m) is likely incorrect. Second, local peak sea level during MIS 5e is characterized by significant departures from eustasy due to glacial isostatic adjustment in response to both successive glacial–interglacial cycles and excess polar ice-sheet melt relative to present day values. We use numerical models of glacial isostatic adjustment that incorporate both of these effects to quantify the plausible range of the combined error and show that, even at sites far from melting ice sheets, local peak sea level during MIS 5e may depart from eustasy by 2–4 m, or more. We also demonstrate that the associated error in the estimated tectonic rates can significantly alter previous estimates of peak eustatic sea level during Quaternary highstands, notably those associated with earlier interglacials (e.g., MIS 11).

1. Introduction

The relationship between temperature change and ice-volume fluctuations during Pleistocene glacial–interglacial cycles provides an important measure of the sensitivity of the Earth system to climate change (Shackleton, 2000). Inferences of past ice volume derive from analysis of marine and ice-core oxygen isotopic depth series (Shackleton, 2000; Siddall et al., 2003) and various stratigraphic and geomorphic markers of local sea level. Local reconstructions of peak sea level during past interglacials, including MIS 5e (~125 ka) and MIS 11 (~410 ka), suggest at least partial collapse of both the Greenland Ice Sheet and the West Antarctic Ice Sheet during these periods of ice-age warmth (Neumann and Hearty, 1996; Hearty et al., 2007; Blanchon et al., 2009; Kopp et al., 2009, 2013; Muhs et al., 2011; Dutton and Lambeck, 2012; Raymo and Mitrovica, 2012; Roberts et al., 2012; O'Leary et al., 2013). The robustness of stratigraphic- and geomorphic-based sea-level inferences, however, depends on the accuracy with which these markers can be corrected for the complicating effect of tectonic uplift or subsidence.

A variety of processes contribute to local tectonic uplift and subsidence. At convergent plate margins, accretion, thrusting and/or thickening of the crust drive local uplift (Cawood et al., 2009); similar mechanisms occur along strike-slip boundaries oblique to plate motion (Basile and Allemand, 2002). Hotspots variably drive uplift or subsidence depending on the buoyancy force relative to

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the surrounding mantle. Likewise, passive margins undergo uplift or subsidence driven by ambient mantle convective forces, though typically (but not necessarily) on timescales longer than a glacial–interglacial cycle (Moucha et al., 2008; Müller et al., 2008; Spasojević et al., 2008; Rowley et al., 2013).

For sites that preserve an MIS 5e sea-level highstand marker, a standard method for estimating the rate of local tectonic uplift (or subsidence) computes the difference between the observed elevation of the MIS 5e marker and a reference eustatic (i.e., global mean) sea-level height for this time, typically chosen as ~6 m, and then divides this height difference by the age of the MIS 5e highstand (e.g., Broecker et al., 1968; Bloom et al., 1974; Dodge et al., 1983; Chappell and Shackleton, 1986; Ota and Omura, 1992; Gallup et al., 2002; Hearty, 2002; Speed and Cheng, 2004; Muhs et al., 2012a, among many others). Assuming that this rate remains constant through time, it is then applied to correct the elevation of all other observed sea-level markers at this site for tectonic uplift.

The precise origin of +6 m as an accepted value for MIS 5e peak eustatic sea level remains a topic of debate (see Murray-Wallace and Woodroffe, 2014 for a recent discussion). Hearty et al. (2007) attributed the correction to Neumann and Moore (1975) who reported a 5.9 m notch at a site in the northern Bahamas that they assumed was tectonically stable. However, earlier workers adopted the +6 m correction to estimate uplift rates, including both Broecker et al. (1968) in their study of Barbados and Bloom et al. (1974) in their analysis of records from Huon Peninsula (see also Chappell, 1974). Murray-Wallace and Belperio (1991) argued that the +6 m correction dates to the work of Veeh (1966), who estimated highstand values in the range +2–9 m on the basis of coral records from (purportedly) tectonically stable sites in both the Pacific and Indian Oceans. While Veeh (1966) was clearly an influential study, the adoption of +6 m by Bloom et al. (1974) and others may also have been influenced by a number of studies dating to the same period that inferred a MIS 5e peak eustatic sea level between 2 and 10 m (e.g., Broecker and Thurber, 1965; Thurber et al., 1965; Veeh, 1966; Land et al., 1967; Broecker and van Donk, 1970). Thus, the specific value of +6 m, originally adopted as representative of the above range of MIS 5e eustatic values, has been reinforced by numerous subsequent field studies at sites thought to be tectonically stable (e.g., Harmon et al., 1981; Brasier and Donahue, 1985; Jones and Hunter, 1990; Muhs et al., 2011), and it has evolved into a reference value commonly invoked without the associated uncertainty identified in earlier studies.

There are, of course, important exceptions. Some studies applied the same methodology for estimating tectonic uplift rates, but adopted a different value, or range of values, for the peak eustatic sea level during MIS 5e (Murray-Wallace, 2002; Schellmann and Radtke, 2004; Omura et al., 2004; Potter et al., 2004; Schellmann et al., 2004; Dumas et al., 2006; Bowen, 2010; Muhs et al., 2012b, 2014). These studies up to and including 2010 all considered values less than or equal to +6 m, reflecting the prevailing view that +6 m was an upper bound on MIS 5e peak eustatic sea level. Recent studies by two independent groups, however, concluded that MIS 5e peak eustatic sea level was ~6–9 m (Kopp et al., 2009, 2013; Dutton and Lambeck, 2012). Kopp et al. (2013) performed a statistical analysis of globally distributed MIS 5e geologic markers and marine oxygen isotope records and concluded that peak eustatic sea level during MIS 5e was extremely likely (95% probability) greater than 6.4 m but was unlikely (33% probability) to have exceeded 8.8 m. Dutton and Lambeck (2012) focused on a small set of high quality MIS 5e sea-level histories from Western Australia and the Seychelles and estimated peak eustatic sea level in the range 5.5–9 m. However, Hay et al. (2014) pointed out that the upper bound on the range cited by Dutton and Lambeck (2012), derived from the Seychelles record, does not account for the geographic variability in sea-level change associated with ice-sheet collapse (see below), and revised this bound downward to 7.5 m. The emerging view that +6 m represents a *lower* bound on MIS 5e peak eustatic sea level motivated Muhs et al. (2012b, 2014) to consider values of 6 and 9 m in estimating tectonic uplift rates for records from Curaçao and the Canary Islands, respectively.

Two potentially important errors are introduced in estimating a tectonic signal by subtracting a reference eustatic value from a local MIS 5e highstand marker. First, as discussed above, the adopted reference value may be in error. Second, an observed (local) MIS 5e highstand marker should not be corrected using the peak global eustatic value, but rather for a value that reflects the combined effect of eustasy and glacial isostatic adjustment (GIA) specific to that location. GIA encompasses the full deformational, gravitational, and rotational perturbation in sea level driven by the redistribution of ice and ocean mass, and introduces significant geographic variability (i.e., departures from eustasy) into the correction (Milne and Mitrovica, 2008; Dutton and Lambeck, 2012; Lambeck et al., 2012; O'Leary et al., 2013; Hay et al., 2014). The incorrect assumption that a globally averaged sea-level value can be used to correct local sea-level records or, equivalently, that the difference between the two is negligible, pervades the analyses cited above.

In this paper, we adopt geophysical models of GIA consistent with the observationally inferred bounds on peak eustatic sea level during MIS 5e to explore the potential error incurred in estimating tectonic effects on the basis of a simple eustatic correction to local MIS 5e highstand markers. We begin with a brief discussion of the basic physics underlying the departure of local, post-glacial sea level from eustasy. We then consider the impact of the assumption of eustasy on estimates of tectonic rates at the specific sites considered in the above analyses.

2. Methods

Our numerical predictions adopt a gravitationally self-consistent sea-level theory that accounts for the viscoelastic deformation of a spherically symmetric, Maxwell viscoelastic Earth, time-dependent shoreline migration, and the effects of Earth rotation changes on sea level (Kendall et al., 2005). The calculations require, as input, models for both Earth's viscoelastic structure and the evolving ice cover. For the former, we adopt the seismic model PREM to prescribe the elastic and density structure (Dziewonski and Anderson, 1981). We also adopt a radial profile of mantle viscosity defined by a very high viscosity (essentially elastic) lithosphere of thickness 95 km, an upper mantle viscosity of 0.5×10^{21}Pa s, and a lower mantle viscosity of 10^{22}Pa s. This model of mantle viscosity is consistent with a suite of inferences based on GIA data (Lambeck et al., 1998; Mitrovica and Forte, 2004).

Our synthetic ice history is comprised of two components. The first component of the ice history, which assumes that there is no difference in ice volume between the Last Interglacial (LIG; MIS 5e) and the present interglacial, isolates what one might call a background ice-age (i.e., glacial–interglacial) cycle signal. The geographic variability in sea-level change associated with this signal has been the focus of several recent GIA studies of MIS 5e (Dutton and Lambeck, 2012; Lambeck et al., 2012; O'Leary et al., 2013). (See also Raymo and Mitrovica, 2012, for analogous modeling of the background glacial–interglacial cycle signal associated with MIS 11). We adopt the ICE-5G version 1.2 model of the last glacial–interglacial cycle (Peltier, 2004), but slightly modify this ice history such that the cycle extends from 120 ka to present day. We then construct a penultimate glacial–interglacial cycle by duplicating the ICE-5G history and shifting it back in time so that