1 Sensitivity of contemporary sea level trends in a global 2 ocean state estimate to effects of geothermal fluxes

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7 **Abstract**

Geothermal fluxes constitute a sizable fraction of the present-day Earth net radiative imbalance and corresponding ocean heat uptake. Model simulations of contemporary sea level that impose a geothermal flux boundary condition are becoming increasingly common. To quantify the impact of geothermal fluxes on model estimates of contemporary (1993–2010) sea level changes, two ocean circulation model experiments are compared. The two simulations are based on a global ocean state estimate, produced by the Estimating the Circulation and Climate of the Ocean (ECCO) consortium, and differ only with regard to whether geothermal forcing is applied as a boundary condition. Geothermal forcing raises the global-mean sea level trend by 0.11 mm yr^{-1} in the perturbation experiment by suppressing a cooling trend present in the baseline solution below 2000 m. The imposed forcing also affects regional sea level trends. The Southern Ocean is particularly sensitive. In this region, anomalous heat redistribution due to geothermal fluxes results in steric height trends of up to $\pm 1 \text{ mm yr}^{-1}$ in the perturbation experiment relative to the baseline simulation. Analysis of a passive tracer experiment suggests that the geothermal input itself is transported by horizontal diffusion, resulting

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in more thermal expansion over deeper ocean basins. Thermal expansion in the perturbation simulation gives rise to bottom pressure increase over shallower regions and decrease over deeper areas relative to the baseline run, consistent with mass redistribution expected for deep ocean warming. These results elucidate the influence of geothermal fluxes on sea level rise and global heat budgets in model simulations of contemporary ocean circulation and climate.

8 *Keywords:* Ocean general circulation model, ocean state estimation, solid
9 Earth surface heat flow, geothermal fluxes, sea level rise, passive tracer

10 **1. Introduction**

11 Due to the conductive cooling of the oceanic lithosphere and turbu-
12 lent convection over hydrothermal vents, the solid Earth imparts heat to
13 the ocean at a rate of 35–36 TW ($1 \text{ TW} \equiv 10^{12} \text{ W}$) or $\sim 100 \text{ mW m}^{-2}$
14 ($1 \text{ mW} \equiv 10^{-3} \text{ W}$) on average (Pollack et al. 1993; Davies 2013). Whereas
15 the oceanographic literature is replete with studies of the oceanic response to
16 various types of surface forcing (e.g., Forget and Ponte 2015), there have been
17 far fewer works on the ocean’s adjustment to the geothermal fluxes along its
18 bottom boundary, which is surprising, since it is not small compared to the
19 energy imbalance of the ocean (Wunsch and Heimbach 2014) or the planet
20 (Allan et al. 2014).

21 Earlier investigations focus on regional ocean processes about localized
22 geothermal sources. Stommel (1982), Joyce and Speer (1987), and Speer
23 (1989) consider the oceanic response to geothermal forcing by hydrothermal
24 vents along the East Pacific Rise. They establish under what conditions

25 the anomalous hydrothermal input behaves as a passive tracer, advected by
26 background flow, or an active one, spreading westward as a “ β plume.”

27 Later papers target larger scales, considering geothermal forcing in the
28 context of circulation and climate more broadly. Adcroft et al. (2001), Scott
29 et al. (2001), and Emile-Geay and Madec (2009) use circulation models to
30 explore the influence of geothermal flow on the abyssal circulation. These
31 authors detail how geothermal fluxes impact the steady state meridional
32 overturning circulation, deep stratification, and meridional transport of heat.

33 Motivated by the need to understand current trends and anticipate future
34 changes in global sea level (important for adaptation efforts), recent studies
35 have begun to account for geothermal fluxes in estimates of the global-mean
36 steric sea level balance. Griffies and Greatbatch (2012), Hieronymus and
37 Nycander (2013), and Palter et al. (2014) suggest that geothermal fluxes
38 contribute $0.08\text{--}0.11\text{ mm yr}^{-1}$ to the steady state global sea level budget.
39 These estimates are derived by averaging the product of the local thermal
40 expansion coefficient and the rate of geothermal flow over the global ocean
41 floor. Such values are not negligible compared to the rate of global-mean
42 sea level rise ($2.6\text{--}2.9\text{ mm yr}^{-1}$) observed by satellite altimeters over the last
43 couple decades (Watson et al. 2015).

44 The aforementioned studies elucidate the *steady state* ocean response to
45 geothermal forcing. However, increasingly common are *transient* model sim-
46 ulations of contemporary ocean climate and sea level change that include
47 geothermal forcing as a boundary condition (e.g., Griffies et al. 2014). Yet,
48 to the best of our knowledge, there has been no systematic study exploring
49 the impact of incorporating geothermal forcing in these relatively short (e.g.,

50 decadal) transient model runs, and so various questions remain, for example,

- 51 • What is the nature of the modeled ocean response to geothermal fluxes?
- 52 • Does the geothermal forcing affect regional patterns of sea level change?
- 53 • How is geothermal input redistributed by ocean transport processes?
- 54 • Is there any impact on surface heat and freshwater exchanges with the
- 55 atmosphere?

56 We address these questions using two 20-year model simulations that are
57 identical in all respects (initial conditions, atmospheric state, internal model
58 parameters, etc.) except that the first simulation (the *baseline*) does not
59 include a geothermal flux boundary condition, whereas the second one (the
60 *perturbation*) does incorporate this type of forcing along the ocean floor. We
61 infer the impact of the geothermal fluxes by taking the difference between
62 the two solutions. This study is distinct from previous work on the equilib-
63 rium response in that we consider transient behavior over a short interval.
64 While we acknowledge that our results may be sensitive to the particular
65 study period and that different conclusions might follow from consideration
66 of longer periods, our findings are relevant to efforts to simulate global and
67 regional sea level changes over the modern altimetric period (Griffies et al.
68 2014; Storto et al. 2015).

69 The remainder of this paper is structured as follows: in section 2, we
70 outline our numerical model framework, introducing the two solutions to be
71 compared; in section 3, we present our main findings from the model inter-
72 comparison, including results from a passive tracer simulation designed to

73 elucidate the transport pathways and the dynamical influence of geother-
74 mal fluxes; finally, in section 4, we summarize our findings and discuss their
75 implications.

76 **2. Material and methods**

77 *2.1. Ocean state estimate and the baseline solution*

78 Our framework is a dynamically and kinematically consistent estimate of
79 the global ocean circulation and sea ice state over 1992–2011 produced by
80 the Estimating the Circulation and Climate of the Ocean (ECCO) consor-
81 tium (Wunsch et al. 2009; Wunsch and Heimbach 2013; Forget et al. 2015a).
82 Broadly speaking, the ECCO state estimates represent numerical solutions
83 to an evolved formulation of the Massachusetts Institute of Technology gen-
84 eral circulation model (Marshall et al. 1997), adjusted to myriad ocean data
85 (e.g., Wunsch and Heimbach 2013). Based on the adjoint model (Marotzke
86 et al. 1999; Heimbach et al. 2005), iterative adjustments are made to ini-
87 tial conditions (i.e., temperature and salinity), surface boundary conditions
88 (i.e., the atmospheric state), as well as internal model parameters (e.g., spa-
89 tially varying diffusion coefficients; see Forget et al. 2015b) so as to minimize
90 the misfit (or cost or objective function) between the model and data.

91 The state estimate used in this study (ECCO version 4, release 1; here-
92 after simply referred to as the baseline solution) differs from earlier ECCO
93 configurations (e.g., Wunsch et al. 2009), for example, in terms of horizon-
94 tal grid, vertical coordinate, parameterization choices, and surface boundary
95 conditions. Forget et al. (2015a) and Forget and Ponte (2015) document
96 the innovations in ECCO version 4 and the baseline solution in great de-

97 tail, and so a brief description is sufficient in this paper. The model setup
 98 solves the primitive equations with a nominal horizontal spacing of 1° and
 99 fifty vertical levels. Also used are interactive dynamic and thermodynamic
 100 models for sea ice and snow (Losch et al. 2010) along with parameterizations
 101 for effects occurring on small spatial scales (Redi 1982; Gaspar et al. 1990;
 102 Gent and McWilliams 1990; Duffy et al. 1999). This Boussinesq solution
 103 employs a nonlinear free surface and real freshwater exchanges along with a
 104 rescaled (r^*) vertical coordinate (Adcroft and Campin 2004; Campin et al.
 105 2004, 2008). Bulk formulae (Large and Yeager 2004) are used for the sur-
 106 face boundary conditions (except for the momentum equations where wind
 107 stress is specified directly). Atmospheric state variables are first taken from
 108 the Interim European Centre for Medium-Range Weather Forecasts Reanal-
 109 ysis (ERA-Interim) described by Dee et al. (2011) and adjusted based on
 110 procedures outlined above. Note that, while it has been adjusted to obser-
 111 vations through the iterative optimization, this estimate constitutes a freely
 112 running forward model solution. Also, given that the model is Boussinesq
 113 in formulation, we apply the spatially uniform, temporally varying Great-
 114 batch (1994) correction to the sea level and bottom pressure fields. This
 115 post hoc correction consists of “adding back” to the diagnosed sea level and
 116 bottom pressure fields, where bottom pressure is converted to units of water
 117 thickness, the time series of the global mean steric height inferred from the
 118 estimated density field.

119 Figure 1 shows a map of baseline sea level (η) trends over 1993–2010 as
 120 a context for what follows. While these trends and their correspondence to
 121 observations are discussed in detail elsewhere (Forget et al. 2015a; Forget and

122 Ponte 2015), we briefly consider some salient aspects. The baseline solution
 123 gives a global-mean η trend of 2.6 ± 0.4 mm yr⁻¹, which is in agreement
 124 with altimetric data that suggest that global-mean η rose at a rate of 2.6–
 125 2.9 mm yr⁻¹ from 1993 to mid-2014 (Watson et al. 2015). Mass addition
 126 and thermal expansion give 2.1 ± 0.4 and 0.5 ± 0.1 mm yr⁻¹ to the baseline
 127 trend, respectively (not shown); these contributions are similar to recent
 128 observational assessments for overlapping time periods (e.g., Church et al.
 129 2013; von Schuckmann et al. 2014; Llovel et al. 2014; Purkey et al. 2014).

130 The estimate also shows regional deviations from the global-mean η trend
 131 of up to ± 10 mm yr⁻¹ (Figure 1). These spatial variations owe mostly to
 132 steric height (η_ρ) changes rather than bottom pressure (p_b) changes (not
 133 shown), in accord with Stammer et al. (2013), who show strong correspon-
 134 dence between η and η_ρ trend maps over 1993–2010 based on altimetric data
 135 and *in situ* measurements, respectively. (For clarity, we will quote p_b val-
 136 ues in equivalent thickness units rather than the traditional pressure units.)
 137 Some of the most noteworthy regional η features along the western tropical
 138 Pacific and Indian Oceans are thought to reflect the baroclinic ocean response
 139 to wind stress by means of long Rossby waves (cf. Timmermann et al. 2010;
 140 Merrifield 2011; McGregor et al. 2012; Forget and Ponte 2015).

141 *2.2. Estimate of the solid Earth surface heat flow and the perturbation solu-* 142 *tion*

143 Davies (2013) presents a map of solid Earth surface heat flow on a $2^\circ \times 2^\circ$
 144 equal area grid based on a mixture of a half-space conduction model, ex-
 145 tant data, and an assumed relationship between heat flow and geology (his
 146 Figure 8). Higher heat flow over the ocean (≥ 130 mW m⁻²) is found in

147 areas of younger earth crust along mid-ocean ridges, including the East Pa-
 148 cific Rise and the Mid-Atlantic, Southeast Indian, Pacific-Antarctic, Juan
 149 de Fuca, and Reykjanes Ridges, while lower heat flow ($\leq 50 \text{ mW m}^{-2}$) is
 150 observed in areas of older ocean crust over abyssal plains, such as the Pa-
 151 cific, Iberian, Australian-Antarctic, Amundsen-Bellingshausen, Angola, and
 152 Argentine Basins.

153 We create a forcing function by mapping the Davies (2013) values onto
 154 the model grid based on a scattered-data interpolant and nearest-neighbor
 155 interpolation. This gives a spatially weighted mean heat flow over the ocean
 156 of 92 mW m^{-2} (global ocean heating of 33 TW) with values ranging from
 157 4 to 790 mW m^{-2} (Figure 2a), similar to those used in previous modeling
 158 work (Adcroft et al. 2001; Scott et al. 2001; Emile-Geay and Madec 2009).

159 This forcing function facilitates the generation of a *perturbation* model
 160 simulation. This 20-year simulation uses the same setup that was described
 161 above and used to create the baseline solution (i.e., the same initial condi-
 162 tions, internal parameters, control variables, atmospheric state, etc.) but now
 163 with geothermal fluxes (Q) applied at the seafloor. While the atmospheric
 164 state used in the bulk-formulae surface forcing is the same in baseline and
 165 perturbation experiments, surface heat ($\mathcal{H}$) and freshwater ($\mathcal{F}$) fluxes need
 166 not be the same in the two runs, since any differences in sea surface temper-
 167 ature between baseline and perturbation experiments could lead to different
 168 upward radiation and turbulent exchanges with the atmosphere.

169 2.3. A word on notation

170 Since we will frequently compare derived quantities based on state vari-
 171 ables between the baseline and perturbation experiments, we introduce the

172 following notation for economy: Δx denotes the difference in some variable
 173 x between the two experiments; $\bar{x}$ is the time mean of x ; x' indicates least
 174 squares linear (temporal) trend of x . For instance, $\Delta\eta'$ would thus denote
 175 the difference in sea level trend between the two simulations.

176 3. Results

177 3.1. Comparing baseline and perturbation solutions

178 To gauge the influence of $\mathcal{Q}$ on η' we consider a map of $\Delta\eta'$ (Figure 3;
 179 cf. Figure 1, but note the difference in color scale). Including $\mathcal{Q}$ increases
 180 the global-mean η' by 0.11 mm yr^{-1} (Figure 3), which is consistent with
 181 an order-of-magnitude estimate $\eta' = \alpha\mathcal{Q}/(C_p\rho) \sim 0.1 \text{ mm yr}^{-1}$ computed
 182 using representative values for geothermal flux ($\mathcal{Q} \sim 10^{-1} \text{ W m}^{-2}$), thermal
 183 expansion coefficient ($\alpha \sim 10^{-4} \text{ }^\circ\text{C}^{-1}$), specific heat capacity ($C_p \sim 4 \times 10^3$
 184 $\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$), and ocean density ($\rho = 10^3 \text{ kg m}^{-3}$). It is also comparable
 185 to previously published estimates of global-mean η_ρ tendency owing to $\mathcal{Q}$
 186 (Griffies and Greatbatch 2012; Hieronymus and Nycander 2013; Palter et al.
 187 2014).

188 Regional $\Delta\eta'$ values can deviate substantially from this global-mean value,
 189 most notably in the Southern Ocean (Figure 3). Regional $\Delta\eta'$ patterns do
 190 not generally correspond to the spatial structure of $\mathcal{Q}$ (cf. Figures 2a, 3),
 191 confirming our expectation that ocean transports act to redistribute heat.
 192 The most prominent $\Delta\eta'$ features over the Southern Ocean are steric in na-
 193 ture (cf. Figures 3, 4a). Interestingly, spatially varying $\Delta\eta'_\rho$ distributions
 194 appear over most other ocean basins (Figure 4a), which are not reflected in
 195 $\Delta\eta'$ (cf. Figure 3). There appears to be a relation to the bathymetry such

196 that positive $\Delta\eta'_\rho$ values, which occur over deep-ocean areas (cf. Figure 2b),
 197 tend to be compensated by negative $\Delta p'_b$ values (Figure 4b).

198 While the baseline and perturbation runs show similar warming and ex-
 199 pansion in the upper ocean, first-order differences are apparent in the deep
 200 sea (Figure 5). Whereas the baseline shows gradual cooling below 2000 m
 201 (cf. Wunsch and Heimbach 2014), the perturbation run manifests slight ocean
 202 warming (of ~ 1.2 TW or 1.3×10^{-5} °C yr $^{-1}$ on average) over this depth
 203 range (Figure 5c). The net warming of the deep ocean in the perturbation
 204 run indicates heating between 2000–4000 m that outweighs cooling below
 205 4000 m (Figure 5a). Overall, approximately 80% of the global ocean heat
 206 content increase due to $\mathcal{Q}$ comprises warming of waters deeper than 2000
 207 m (Figure 5b). These findings are consistent with there being a source of
 208 heat along the seafloor, which is largely concentrated at ocean ridges situated
 209 ~ 3000 m below the sea surface (cf. Figures 2b, 5b), in the case that vertical
 210 redistribution of geothermal input is unimportant on these timescales.

211 Spatial patterns of surface heat and freshwater flux changes ($\mathcal{H}$ and $\mathcal{F}$)
 212 are similar to the regional distributions of $\Delta\eta'_\rho$ along the Southern Ocean
 213 (Figure 6). The signs of $\Delta\overline{\mathcal{H}}$ and $\Delta\overline{\mathcal{F}}$ tend to be opposite to those of $\Delta\eta'_\rho$
 214 (cf. Figures 4a, 6), which suggests that $\Delta\overline{\mathcal{H}}$ and $\Delta\overline{\mathcal{F}}$ act to damp $\Delta\eta'_\rho$. For
 215 example, assuming that positive $\Delta\eta'_\rho$ values correspond to warmer surface
 216 waters in the perturbation solution, turbulent surface exchanges would indeed
 217 arise due to the bulk formulae forcing, such that the ocean would lose heat
 218 and freshwater to the atmosphere.

219 3.2. Passive tracer simulation

220 What underlies the $\Delta\eta'$, $\Delta\eta'_\rho$, $\Delta p'_b$, $\Delta\overline{\mathcal{H}}$, and $\Delta\overline{\mathcal{F}}$ patterns (Figures 3, 4, 6)?
 221 To what extent are they due to passive transport of geothermal input by
 222 background flows or to dynamic responses to the geothermal fluxes? To
 223 distinguish between the roles of passive and active responses, we perform
 224 a passive tracer simulation; we revert back to the baseline run, releasing a
 225 geothermal “tracer” or “dye” at the seafloor. This tracer is inert, having no
 226 influence on dynamics or thermodynamics; rather, it is passively transported
 227 *via* circulation and mixing processes.

228 Figure 7 shows maps of vertically integrated linear trends in the tempera-
 229 ture difference between the perturbation and baseline simulations ($\Delta\theta'$) and
 230 passive tracer content (C'). (For clarity, and in keeping with the notation
 231 introduced in section 2.3, the symbols θ and C should be read as *vertical inte-*
 232 *grals* of potential temperature and passive tracer.) Regional patterns of $\Delta\theta'$
 233 are analogous to spatial distributions of $\Delta\eta'_\rho$ (cf. Figures 4a, 7a), indicating
 234 the latter’s predominantly thermosteric nature. Spatial patterns of $\Delta\theta'$ and
 235 C' are qualitatively similar in many places (Figures 7a, 7b), with both show-
 236 ing positive trends over deeper ocean basins (cf. Figure 2b). Such is not the
 237 case, however, in the Southern Ocean, where the $\Delta\theta'$ features (Figure 7a) are
 238 not seen in the C' distributions (Figure 7b). This can be seen more clearly
 239 through examination of the residual difference $\Delta\theta' - C'$ (Figure 7c), which
 240 can be interpreted as dynamic response to geothermal forcing; signatures of
 241 dynamic response are seen not only along the Southern Ocean but also in
 242 other regions, including the North Atlantic, Wharton (Indian Ocean), and
 243 Pacific Basins. This dynamic response in the Southern Ocean (Figure 7c)

could reflect instabilities in the system: Losch et al. (2004) show that ocean models are sensitive to small perturbations in the Southern Ocean, due to the strong horizontal gradients and steep isopycnal surfaces; in such a case, circulation changes due to $\mathcal{Q}$ redistribute pre-existing (baseline) ocean heat content, introducing warm or cool surface water anomalies, leading to $\mathcal{H}$ and $\mathcal{F}$ anomalies.

Tracer trend patterns bear little resemblance to the distributions expected from the $\mathcal{Q}$ forcing function (Figures 2a, 7b). In fact, C' is smaller along ocean ridges with higher $\mathcal{Q}$ values and larger over abyssal plains with lower $\mathcal{Q}$ rates (Figures 2, 7b). What leads to such distributions? One possibility is lateral diffusion. For example, tracer trend at 2729 m in the South Pacific Ocean, immediately above the topography, suggests diffusive horizontal spreading from the source along the East Pacific Rise (Figure 8a, cf. Figure 2). To garner a rough sense of whether C' spatial patterns could result from horizontal diffusion, we take the three-dimensional passive tracer forcing function and, at each vertical level, apply a horizontal Gaussian spatial smoother with a standard deviation of 1000 km. (While this length scale is simply a crude choice based on visual inspection of Figure 8a, a similar order of magnitude follows from dimensional analysis, assuming a balance between time tendency and horizontal diffusion over 20 years with diffusivity on the order of $100 \text{ m}^2 \text{ s}^{-1}$.) The vertically integrated, spatially smoothed forcing function (Figure 8b) shows a close resemblance to the C' patterns (Figure 7b), hinting that isotropic horizontal diffusion contributes importantly to evolution of the tracer.

The $\Delta p'_b$ patterns remain to be understood (Figure 4b). Landerer et

269 al. (2007) consider the case of a horizontally uniform decrease in density
 270 (or relatedly increase in temperature) over the global ocean at some depth.
 271 They reason that the resulting thermal expansion over regions as deep as, or
 272 deeper than, that depth would induce an isostatic ocean adjustment (cf. Gill
 273 1982, section 9.9), which would lead to mass decrease over ocean regions
 274 deeper than the depth of the density decrease and mass increase elsewhere.
 275 Generalizing to an arbitrary number of vertical levels, the Landerer et al.
 276 (2007) model predicts $\Delta p'_b$ at vertical level i ($\Delta p'_{b,i}$) given $\Delta \eta'_\rho$ at all levels j
 277 ($\Delta \eta'_{\rho,j}$) as

$$\Delta p'_{b,i} = \left[\sum_{j=1}^i \left(\frac{A_j}{A_s} - 1 \right) \Delta \eta'_{\rho,j} + \sum_{j=i+1}^N \left(\frac{A_j}{A_s} \right) \Delta \eta'_{\rho,j} \right], \quad (1)$$

278 where A_s and A_j represent the areas of the ocean surface and ocean layer j ,
 279 respectively, while N is the total number of vertical levels ($N = 50$ in our
 280 case). Using the perturbation and baseline experiments to compute $\Delta \eta'_{\rho,j}$
 281 at each level (not shown), we find that this simple redistribution model ex-
 282 plains 63% of the area-weighted variance in $\Delta p'_b$ (Figure 9). This isostatic
 283 redistribution model thus grossly explains the mass redistribution caused by
 284 the geothermal forcing as it diffuses horizontally.

285 4. Summary and discussion

286 Motivated by recent investigations of abyssal ocean warming and its im-
 287 pact on global heat and sea level budgets (e.g., Purkey and Johnson 2010;
 288 Wunsch and Heimbach 2014), we studied the effect of imposing geothermal
 289 fluxes in transient simulations of contemporary ocean climate and sea level
 290 change. We considered two ocean model simulations, based on a physically

291 consistent global ocean state estimate produced by the ECCO consortium—
 292 the first without geothermal fluxes (baseline) and the second with this forcing
 293 (perturbation). Geothermal forcing increases the global-mean sea level trend
 294 over 1993–2010 by 0.11 mm yr^{-1} (or $\sim 20\%$ of the baseline global-mean steric
 295 height trend) in the perturbation relative to the baseline (Figure 3) owing to
 296 the suppression of a cooling trend in the baseline below 2000 m (Figure 5).
 297 This global-mean sea level trend increase is consistent with simple predictions
 298 based on the average geothermal flux along the global seafloor. Furthermore,
 299 the imposed forcing has an impact on regional trends in sea level (Figure 3),
 300 steric height (Figure 4a), and bottom pressure (Figure 4b), as well as on
 301 regional surface heat and freshwater fluxes (Figure 6). The Southern Ocean
 302 is especially sensitive—differences in regional sea level trends between the
 303 two runs can be as much as $\pm 1 \text{ mm yr}^{-1}$ along the path of the Antarctic
 304 Circumpolar Current (Figure 3).

305 A passive tracer simulation was investigated to separate contributions
 306 from active and passive ocean responses to the imposed geothermal forcing.
 307 The vertical structures of the geothermal fluxes and linear trends in vertically
 308 integrated passive tracer and temperature difference between the baseline and
 309 perturbation simulations (Figure 5b) suggest that there are negligible differ-
 310 ences between the two experiments in terms of net vertical heat transport.
 311 Passive transport of the geothermal input by lateral eddy diffusion, however,
 312 contributes importantly to the differences between the experiments in terms
 313 of horizontal patterns of steric expansion (Figure 4a, 7b, 8b). The ocean
 314 bottom pressure field responds to these steric expansion patterns; ocean
 315 mass is redistributed horizontally from areas where the water column is deep

316 (e.g., overlying abyssal plains) to where it is shallow (e.g., in marginal seas;
 317 Figures 4b, 9). Dynamic ocean adjustment, involving circulation changes
 318 brought about by the geothermal forcing, underlies the most prominent dif-
 319 ferences between the two simulations in terms of sea level, steric height, and
 320 surface exchange of heat and freshwater along the Southern Ocean (e.g., Fig-
 321 ure 7c), most likely reflecting the inherent instabilities of strong current sys-
 322 tems such as the Antarctic Circumpolar Current (e.g., Losch et al. 2004).

323 Purkey and Johnson (2010) examine repeat synoptic hydrographic sec-
 324 tion data and highlight an apparent warming of the global deep ocean below
 325 2000 m during the 1990s and 2000s, thought to be due to a slowdown in deep
 326 water production around Antarctica, equivalent to a heat flux of 68 ± 61 mW
 327 m^{-2} averaged over the surface of the planet. Wunsch and Heimbach (2014)
 328 use an earlier iteration of the global ocean state estimate and find a slight
 329 cooling over these depths and this period. Imposing geothermal fluxes in the
 330 state estimate leads to a marginal warming of 2 mW m^{-2} averaged over the
 331 planet in the deep waters (Figure 5c). This is insufficient to achieve agree-
 332 ment with Purkey and Johnson (2010) and it is not immediately obvious
 333 what is responsible for the abiding discrepancy. However, ongoing iterative
 334 optimization of the state estimate solution (explicitly including geothermal
 335 forcing and also bringing in more observations, e.g., from the Gravity Recov-
 336 ery and Climate Experiment, as suggested by Piecuch and Ponte 2014), could
 337 lead to improvements in the state estimate, such as to the initial conditions,
 338 that bring the solution into consistency with the Purkey and Johnson (2010)
 339 values.¹

¹It is also worth mentioning that the uncertainties on the Purkey and Johnson (2010)

340 This investigation has mainly focused on differences in trends between the
341 baseline and perturbation experiments. Ancillary discussion is warranted of
342 how much these trends reflect overall sensitivity of the model simulations to
343 geothermal fluxes. Over a majority of the ocean, differences in sea level time
344 series between the two runs are indeed largely explained by linear trends
345 (not shown), meaning that the trends are in fact representative of the model
346 sensitivity to geothermal fluxes over the period of integration. However, there
347 are a few exceptions to this rule; linear trends do not capture the overall
348 sensitivity in some regions, for example, along strong current systems (not
349 shown), hinting at some dependence of the regional response to geothermal
350 fluxes on the background state.

351 It is worth emphasizing the transient nature and short time duration of
352 the experiments considered in this study. Motivated by the availability of a
353 relatively large number of ocean reanalyses, state estimates, and circulation
354 models covering the modern altimetric period (e.g., Griffies et al. 2014; Storto
355 et al. 2015), our target was to elucidate the influence of including geother-
356 mal forcing in numerical simulations of contemporary ocean climate and sea
357 level change. Differences between perturbation and baseline do not reflect
358 an ocean in equilibrium with geothermal fluxes. Were we to have considered
359 a different duration (say 200 or 2000 instead of 20 years), spatial patterns of
360 trend differences would likely have differed. Nevertheless, the type of large-
361 scale adjustments of shallow versus deep regions, and dynamical features
362 over strong current regions offer valuable physical insights. Therefore, while

estimates, based as they are on sparse sampling, could be larger than the formal error bars suggest, due to possible spatiotemporal aliasing (cf. Ponte 2012).

our results are not directly comparable to previous findings on the impact of geothermal forcing on the steady state general circulation and ocean climate (Adcroft et al. 2001; Scott et al. 2001), our study complements these previous works by shedding light on the nature of the transient processes involved in the adjustment of the abyssal ocean to geothermal fluxes.

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Figure 1: Eighteen-year (1993–2010) trends in sea level η (mm yr^{-1}) from the baseline solution without geothermal forcing. Since the ocean model conserves volume rather than mass, the spatially uniform correction due to Greatbatch (1994) has been applied.

Figure 2: (a) Geothermal flux Q over the ocean (mW m^{-2} ; $1 \text{ mW} \equiv 10^{-3} \text{ W}$) from Davies (2013) and used as forcing in the perturbation experiment. (b) Ocean depth (m) from the model.

Figure 3: Eighteen-year (1993–2010) trends (mm yr^{-1}) in the sea-level difference between the perturbation and baseline solutions ($\Delta\eta'$).

Figure 4: Eighteen-year (1993–2010) trends (mm yr^{-1}) in the difference between (a) steric sea level ($\Delta\eta'_\rho$) and (b) ocean bottom pressure ($\Delta p'_b$) from the perturbation and baseline solutions. The $\Delta p'_b$ values have been scaled by $1 \times 10^4 \text{ kg m}^{-2} \text{ s}^{-2}$ to convert from pressure units to equivalent water thickness. The correction due to Greatbatch (1994) has also been applied to the $\Delta p'_b$ values. To highlight compensation, respective global-mean values of $\Delta\eta'_\rho$ and $\Delta p'_b$ are contoured in white.

Figure 5: (a) Area-weighted horizontal-mean 1993–2010 temperature trend integrated above a given depth z from baseline (blue solid) and perturbation (red dashed) solutions ($^{\circ}\text{C m yr}^{-1}$). (b) Difference (perturbation minus baseline) between the two curves in (a) (yellow). Also shown are the corresponding horizontally averaged and cumulatively integrated 1993–2010 trends in passive tracer content (green dash-dotted) and local geothermal forcing (purple dashed). (c) Time series (1992–2011) of the volume integral of potential temperature times heat capacity and density below 1994 m (J) from the baseline (cyan) and the perturbation (magenta) solutions.

Figure 6: Difference (perturbation minus baseline) between time means (over the 1993–2010 period) of surface (a) heat ($\Delta\overline{\mathcal{H}}$; W m^{-2}) and (b) freshwater ($\Delta\overline{\mathcal{F}}$; mm yr^{-1}) exchange. Sign convention is such that positive values increase temperature in (a) and decrease salinity in (b).

Figure 7: Eighteen-year (1993–2010) trends in vertical integrals ($^{\circ}\text{C m yr}^{-1}$) of (a) temperature difference between the two solutions ($\Delta\theta'$) and (b) passive tracer (C'). (c) The difference $\Delta\theta' - C'$ between panels (a) and (b) which reflects the dynamic ocean response to the geothermal forcing.

Figure 8: Eighteen-year (1993–2010) trends in (a) passive tracer ($^{\circ}\text{C yr}^{-1}$) at 2729 m and (b) the vertical integral of the horizontally smoothed 3-D geothermal forcing function ($^{\circ}\text{C m yr}^{-1}$).

Figure 9: Depth versus inter-model difference in 18-year (1993–2010) bottom pressure trends ($\Delta p'_b$; blue) along with the prediction—given the difference in area-weighted spatial-mean steric height trends as a function of depth—of $\Delta p'_b$ based on the Landerer et al. (2007) model (orange).