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Modelling the Impact of Tides and Geothermal Heat Flux on the Climate of Early Earth

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Key Points:

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9	•	First spatially resolved modelling of both tides and their impact on early Earth's
10		climate
11	•	Impact of tidal heating is strongest when altering ocean surface currents close to

- the sea-ice edge
- Tidal heating has only local impacts, but tidal mixing and geothermal heat flux
 influence global climate state

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15 Abstract

On early Earth increased rates of tidal energy dissipation are likely, but depend on the 16 (unknown) distribution of continents. A stronger tidal heating could provide an addi-17 tional energy source during times of substantially lower solar input. So far, the problem 18 has been assessed in terms of the negligible contribution to Earth's global energy bud-19 get. Here we present a spatially resolved investigation of the impact of tidal heating, mix-20 ing, and geothermal heat on early Earth's climate. Using a random landmass distribu-21 tion, tidal heating is calculated for three different rotation periods (12, 18, 24 hours) and 22 fed into a climate model. For each rotation rate, three climate states with different at-23 mospheric CO_2 levels are simulated. We find that, depending on the climate state, tidal 24 heating can affect regional ocean dynamics and sea-ice cover. The impact is strongest 25 when tidal heating alters sea-ice dynamics and meridional heat transport close to the 26 sea-ice edge, but its global impact remains negligible with only small global mean changes 27 in ice cover ($\sim 0.3\%$) and temperature ($< 0.05^{\circ}$ C). Adding tidal mixing and geothermal 28 heat, however, leads to significant reduction in sea-ice cover of $\sim 11\%$ and $\sim 19\%$, respec-29 tively, and thus to larger global warming. As we do not consider the dynamical effects 30 of a higher rotation rate or different landmass distributions, this is only a first glimpse 31 at the importance of tides for the climate of early Earth. Nevertheless, our results sug-32 gest that tides and geothermal heat are important for understanding regional climates 33 and could have contributed to warming early Earth. 34

35 1 Introduction

Ocean tides influence a range of key Earth system processes (Green & Duarte, 2022). 36 These include sustaining ocean primary production (Sharples et al., 2007; Tuerena et al., 37 2019), providing parts of the energy needed to maintain the climate-regulating overturn-38 ing circulation in the ocean (Wunsch & Ferrari, 2004; Wilmes et al., 2021), and contribut-39 ing to key evolution events (Balbus, 2014; Byrne et al., 2020). The drag introduced on 40 Earth by the tides also forces Earth's spin to slow down, increasing day length, and caus-41 ing the Moon to recede to conserve angular momentum of the system (MacDonald, 1964; 42 Bills & Ray, 1999; Daher et al., 2021). Consequently, on early Earth the Moon was closer 43 and the days much shorter. The Archean Eon spans nearly 1/3 of Earth's history – from 44 4000–2500 Ma (million years ago) – and there is naturally uncertainty and a range of 45 estimates of daylength for the Eon, from as short as 13 hours at 3200 Ma (Eulenfeld & 46 Heubeck, 2023) to 18 hours at the end of the Eon (Farhat et al., 2022; Daher et al., 2021). 47 The tides are consequently the key controller of day length, which directly influence plan-48 etary habitability (Yang et al., 2014) as well as the timing of the oxygenation of the at-49 mosphere (Klatt et al., 2021). An understanding of the evolution of tides and tidal drag 50 through Earth's history is therefore key to understand other Earth system events. 51

The first order controller of tidal energetics is continental configuration, as for a 52 fixed lunar distance, and hence tidal forcing, the tidal dissipation rates can vary up to 53 three orders of magnitudes depending on continental configuration (Green et al., 2017; 54 Blackledge et al., 2020). Consequently, even though the forcing was larger during the Archean, 55 the tides may not have been more energetic. This is because the tides become resonant 56 for specific basin geometries (Green, 2010), e.g., the present day North Atlantic (Egbert 57 et al., 2004; Green, 2010), and that effect can lead to larger amplification of the tide than 58 the increase in forcing alone can do (Green et al., 2020). 59

The influence of the tides on the climate of Early Earth has largely been assessed in terms of their contribution to the planetary energy balance (Barnes et al., 2008) in light of the Faint Young Sun Problem (FYSP) (Feulner, 2012). Even considering the larger tidal forcing on early Earth and the uncertainty with respect to continental configuration, however, the heating from tidal dissipation is expected to be negligible compared to the energy input from the faint young Sun. Nevertheless, recent investigations of the

geological imprint of ancient tides and the long-term evolution of the Earth-Moon dis-66 tance have led to a renewed interest in the potential influence of tidal heating on the Hadean 67 and Archean climate. For example, based on simple models for the long-term evolution 68 of Earth's tides, Heller et al. (2021) argue that the global tidal dissipation, and hence 69 heating, rate on early Earth may have been underestimated. They conclude that tides 70 might have helped to keep surface water liquid for 220 Myr and could thus have been 71 an important contribution to the planetary energy budget during the early Hadean. How-72 ever, their approach necessarily neglects both tidally driven ocean dynamics and the in-73 fluence of continental configuration on tidal dissipation (and is thus mainly a valid rep-74 resentation of deep aquaplanets), and the climatic consequences are assessed with a zero-75 dimensional grey atmosphere model ignoring internal climate dynamics and the ice-albedo 76 feedback. 77

Here, we take the first step towards a more comprehensive assessment of the im-78 pact of tides on early Earth's climate by computing tidal energy dissipation rates from 79 a spatially resolved tidal model for one particular continental configuration (broadly rep-80 resentative of an early Earth) and feeding the additional energy input as heat into a three-81 dimensional coupled climate model. This way, we can also investigate the effect of lo-82 cal tidal heating on ocean and sea-ice dynamics beyond the negligible impact on the global 83 climate. Furthermore, we run additional sets of simulations including the effects of tides 84 on vertical mixing in the ocean and with geothermal heat flux. 85

86 2 Methods

2.1 Tidal Modelling

The tide model simulations were done using OTIS, the Oregon State Tidal Inversion Software (e.g., Egbert et al., 2004; Green & Nycander, 2013). It has been used extensively for paleotidal simulations as far back as the Proterozoic (Green & Huber, 2013; Green et al., 2017; Green et al., 2020; Green & Duarte, 2022). It solves the linearised shallow water equations:

$$\frac{\partial \mathbf{U}}{\partial t} + \mathbf{f} \times \mathbf{U} = -gH\nabla(\zeta - \zeta_{EQ} - \zeta_{SAL}) - \mathbf{F}$$
(1)

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$$\frac{\partial \zeta}{\partial t} = -\nabla \cdot \mathbf{U} \tag{2}$$

⁹⁵ Here, **U** is the depth integrated volume transport, H is water depth, **f** the Coriolis vec-⁹⁶ tor, g the gravitational constant, ζ the tidal elevation, ζ_{EQ} the equilibrium tidal eleva-⁹⁷ tion, and ζ_{SAL} the tidal elevation due to self-attraction and loading. The tidal drag is ⁹⁸ parameterised through two parts: $\mathbf{F} = \mathbf{F}_B + \mathbf{F}_{IT}$, where \mathbf{F}_B is frictional losses due to ⁹⁹ bed friction and \mathbf{F}_{IT} energy losses due to tidal conversion. The former is represented by ¹⁰⁰ the standard quadratic law:

$$\mathbf{F}_B = C_d \mathbf{u} |\mathbf{u}| \tag{3}$$

where $C_d = 0.003$ is a drag coefficient, and **u** tidal velocity. The conversion is written as $\mathbf{F}_{IT} = C_{IT}\mathbf{U}$, with the conversion coefficient C_{IT} given by (Zaron & Egbert, 2006; Green & Nycander, 2013)

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$$Y_{IT}(x,y) = \gamma \frac{(\nabla H)^2 N_b \bar{N}}{8\pi\omega}$$
(4)

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where $\gamma = 50$ is a scaling factor (see Zaron & Egbert, 2006, for more details), N_b the buoyancy frequency at the sea-bed, \bar{N} the vertical average of the buoyancy frequency, and ω is the frequency of the tidal constituent under evaluation. The buoyancy frequency is computed from a statistical fit based on observations from present day Earth, or N(x, y) =0.00524 exp(-z/1300), where z is the vertical coordinate, and the constants 0.00524 and 1300 have units of s⁻¹ and m, respectively. Whilst the abyssal stratification is known from the climate model simulations (see section 2.3), the tides are relatively insensitive to changes in stratification within a factor 2 (Egbert et al., 2004). Furthermore, itera-

114 tive studies, where stratification has been implemented in the tidal model and the re-

sulting dissipation returned to a climate model, showed only minor impact on the results

(Schmittner et al., 2015).

2.2 Simulations and Computations

The tidal dissipation rates, D, were computed using the energy balance method (Egbert & Ray, 2001)

$$D = W - \nabla \cdot \mathbf{P} \tag{5}$$

The work rate done by the tide-generating force, W, and the energy flux, \mathbf{P} , are given by

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$$W = g\rho_0 \langle \mathbf{U} \cdot \nabla(\eta_{eq} + \eta_{SAL}) \rangle \tag{6}$$

$$\mathbf{P} = g\rho_0 \langle \mathbf{U}\eta \rangle \tag{7}$$

in which $\langle \rangle$ denote time-averages over a tidal period. As the end result of tidal energy dissipation is heat, we add the tidal dissipation as tidal heating in the climate model.

Tidal model simulations were done for three rotation rates – 12, 18, and 24 hours – using one topography – denoted "225" – from Blackledge et al. (2020). This was chosen because it consists of a series of smaller continents scattered around the planet, much like early Earth has been proposed to have looked like (Hawkesworth et al., 2020; Sawada, 2020; Cawood et al., 2022). The associated changes in tidal forcing due the closer proximity of the Moon for each investigated rotation rate was taken from Daher et al. (2021). The resulting tidal energy dissipation fields are shown in Fig. 1.

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2.3 Modelling Tidal Impacts on Early Earth's Climate

To investigate the impact of tidal energetics on the climate on early Earth, we em-134 ploy the deep-time version of the intermediate-complexity climate model CLIMBER- 3α 135 (Montoya et al., 2005). The model is based on a modified version of the ocean general 136 circulation model MOM3 (Pacanowski & Griffies, 1999; Hofmann & Morales Maqueda, 137 2006) run at a horizontal resolution of $3.75^{\circ} \times 3.75^{\circ}$ and using 24 levels vertically, a sea-138 ice model capturing both thermodynamics and dynamics (Fichefet & Morales Maqueda, 139 1997) at the same horizontal resolution as the ocean model, and a fast statistical-dynamical 140 atmosphere model (Petoukhov et al., 2000) with a coarse horizontal resolution of 22.5° 141 in longitude and 7.5° in latitude. The deep-time version of CLIMBER-3 α entails the re-142 moval of parameterisations relating to the present-day climate system, additional diag-143 nostics, and bug fixes. This version has been used in a range of studies relating to plan-144 etary habitability, in particular for the climate on early Earth (Kienert et al., 2012, 2013) 145 and major glaciations in Earth's history (Feulner & Kienert, 2014; Feulner et al., 2015; 146 Feulner, 2017; Eberhard et al., 2023; Feulner et al., 2023). 147

All climate model simulations are run for the continental configuration "225" from 148 Blackledge et al. (2020) as described above, re-gridded to the resolution of the ocean model 149 (see Fig. 2). The solar constant values corresponding to the three tidal scenarios with 150 12, 18, and 24 hours rotation rate are computed based on the relation between day length 151 and time from Farhat et al. (2022) and the stellar evolution model MESA (Jermyn et 152 al., 2023). For each rotation rate, simulations with three different atmospheric CO_2 lev-153 els are performed which are characterised by a low, medium, and high fraction of global 154 sea-ice cover (see Table 1). We run four sets of nine simulations each: 155

 Control simulations for the three tidal scenarios and three climate states without any tidal effects and ignoring geothermal heat flow. Key ocean and sea-ice characteristics of these simulations are shown in Figure B1 in Appendix B.



Figure 1. Maps (left column) and zonal means (right column) of modelled tidal dissipation rates for all three rotation rates of 12 h (top panels), 18 h (middle panels), and 24 h (bottom panels). Note the logarithmic scale.



Figure 2. Ocean bathymetry and land orography based on continental configuration "225" from Blackledge et al. (2020) at the $3.75^{\circ} \times 3.75^{\circ}$ resolution of CLIMBER-3 α 's ocean model.

2. Same as 1., but including the effects of *tidal heating*. In this case, the energy input by the tides (computed as described in Sections 2.1 and 2.2, and shown in Fig. 1) is implemented as additional bottom heat flux into the ocean.

3. Same as 2., but taking the effects of tides on vertical mixing into account. The 162 vertical diffusivity is represented in the model by a constant background diffusiv-163 ity, afkph, and an amplitude of the spatially varying diffusivity, dfkph. In the con-164 trol simulations these are tuned to present day values of 0.8×10^{-4} and $1.05 \times$ 165 $10^{-4} \,\mathrm{m^2 \, s^{-1}}$, respectively. At present day the energy required to sustain abyssal 166 diffusivities is provided in equal parts by the wind and the tide (Wunsch & Fer-167 rari, 2004), and we assumed the same ratio here. For simplicity, we computed the 168 ratio, R, of the globally integrated abyssal tidal dissipation rates (i.e., in waters 169 deeper than 500 m) from the simulations presented here to a present-day tidal model 170 simulation (Green et al., 2017), i.e., $R = D_{\text{Archaean}}/D_{\text{PD}}$. We then modify the 171 tidally driven half of the afkph and dfkph by multiplying the control value with 172 R so that $afkph = 0.8 \times (0.5 + 0.5R)$ and $dfkph = 1.05 \times (0.5 + 0.5R)$. The re-173 sulting values are shown in Table 2 and further highlight the existing knowledge 174 that tidal dissipation rates, and hence vertical mixing, are to first order controlled 175 by continental configuration (Blackledge et al., 2020; Green et al., 2018). The dis-176 sipation and associated diffusivity change by around 20% when we move from a 177 present day topography (24 hour control) to our chosen Archean one (24 hour Archean). 178 The reason for this is tidal resonances developing over large areas in the Archean 179 topography. As we then change rotation rate, the Archean basins go out of res-180 onances for an 18 hour day length; coincidentally the background diffusivity is then 181 the same as for present day. At 12 hours, the tides become more energetic and the 182 diffusivity is 80% higher than at present. Note that Farhat et al. (2022) propose 183 a rapid change in day length between 12–15 hours due to potential orbital reso-184 nances and our results support this. Also note that because the diffusivity is com-185 puted in waters deeper than 500 m and the tidal heating s the total dissipation, 186 including shallow waters we can have the same total dissipation but different dif-187 fusivities and vice versa. 188 189

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4. Same as 3., but *adding geothermal heat flux*. Time evolution of global geothermal heat flux from radioactive decay follows Turcotte and Schubert (2002), with the

Table 1. Overview of tidal scenarios and climate model simulations. Tidal simulations are performed for three different day lengths; the corresponding point in time is estimated from day length following Farhat et al. (2022), the value of the solar constant S_0 at these times are obtained from stellar evolution model MESA (Jermyn et al., 2023). The corresponding global mean values for tidal heating (from the tidal model) and geothermal heating (see text for details) are indicated as well. For each of the three tidal scenarios, climate model simulations are performed at three atmospheric CO₂ concentrations for which the climate model simulations yield low, intermediate, and high global sea-ice cover.

Day length [h]	Time [Myr]	Age [Ma]	$\begin{vmatrix} S_0 \\ [W/m^2] \end{vmatrix}$	Tidal heating $[mW/m^2]$	$ \begin{vmatrix} \text{Geothermal} \\ [mW/m^2] \end{vmatrix} $	CO_2 [ppm]	Ice cover
12	867	3700	1010	14.40	217.7	$\begin{array}{c} 176,000 \\ 200,000 \\ 220,000 \end{array}$	$ \begin{array}{c c} 16\% \\ 7.1\% \\ 3.0\% \end{array} $
18	2767	1800	1161	10.90	121.6	15,000 25,000 28,000	$\begin{array}{c} 15\% \\ 5.4\% \\ 3.2\% \end{array}$
24	4567	0	1361	2.758	90.42	$140 \\ 277 \\ 350$	$\begin{array}{c c} 14\% \\ 7.2\% \\ 4.5\% \end{array}$

surface heat flow due to mantle cooling adjusted to 29 mW/m^2 to yield a presentday flux of 87 mW/m^2 for the modern continental configuration (Pollack et al., 1993). Furthermore, values are multiplied by a factor of 1.03 to take the partitioning between continental and oceanic crust for the lower total land area of continental configuration "225" into account. As for tidal heating, geothermal heat flux is uniformly distributed over the globe and added to the lowermost ocean cells; input to land areas is ignored in the model.

We employ an idealised present day-like orbital configuration with an obliquity of 23.5° and a circular orbit. All climate model simulations were integrated for at least 4000 years until equilibrium was approached. Note that rotation rates in the climate model simulations are 24 hours throughout to isolate the effect of tidal dissipation on the climate from the changes in Earth-system dynamics due to higher rotation rates (which are beyond the scope of this paper).

²⁰⁴ 3 Results

In the following, we will look at the impact of tidal heating, tidal mixing, and geother-205 mal heat flux on the climate for the different sets of simulations described in Section 2.3 206 by comparing the changes in key climate variables, in particular surface air temperature 207 (SAT), sea-surface temperature (SST), sea-ice cover, and ocean surface velocities com-208 pared to the respective control simulations (the latter are shown in Figure B1). All di-209 agnostics have been averaged over the last 500 years of each simulation to minimise the 210 impact of inter-annual to centennial climate variability. With respect to the overall cli-211 mate state, the global and annual mean SATs and SSTs of all simulations are summarised 212 in Table A1. We structure the discussion of results of our simulation experiments fol-213 lowing three key messages summarising the dependence on rotation rate (Section 3.1, 214 Figure 3), climate state (Section 3.2, Figure 4), and forcing (Section 3.3, Figures 5 and 215 6).216



Figure 3. Tidal heating and climate-system response depend on rotation rate. Changes in sea-surface temperature, surface velocity, and sea-ice cover due to tidal heating for three different tidal scenarios with (a) 12 h, (b) 18 h, and (c) 24 h rotation rate and for intermediate CO_2 levels. The blue and red colors correspond to changes in SST when introducing tidal heating. Yellow areas are land; semi-transparent white areas correspond to regions that have at least 50% sea-ice fraction in the annual mean. Dotted and hatched areas represent regions of sea-ice gain or loss, respectively. Arrows indicate a change in ocean surface currents, with their length scaling with the change in speed. Maps showing the effects of tidal heating for all climate states and all rotation rates can be found in Figure B2 in Appendix B.

Table 2. Summary of the background (afkph) and amplitude (dfkph) of the vertical diffusivities used in the simulations; the 24 h values are the present-day default values used in the coupled climate model.

Day length [h]	afkph $[10^{-4} \text{ m}^2 \text{ s}^{-1}]$	$\big \; {\tt dfkph} \; [10^{-4} \; {\rm m}^2 \; {\rm s}^{-1}]$
12, Archean	1.44	1.90
18, Archean	0.80	1.05
24, Archean	0.96	1.26
24, control	0.80	1.05

3.1 Climate response to tidal heating depends on rotation rate

For a given continental configuration, tidal dissipation (and thus heating) strongly 218 depends on the assumed daylength (see Figure 1). These differences in tidal dissipation 219 are also reflected in the climate response to this heating shown in Figure 3 for the in-220 termediate climate state (the corresponding maps for all climate states can be found in 221 Figure B2 in Appendix B). Although the global effect of tidal heating on the climate is 222 very small (with global annual mean surface air temperature changes typically less than 223 0.1°C, see Table A1), marked regional impacts are visible at higher latitudes, varying 224 with the spatial distribution and magnitude of tidal heating for different rotation rates. 225

These changes in sea-surface temperature, sea-ice cover, and ocean surface veloc-226 ities in Figure 3 are largest close to the sea-ice edge and suggest that the response is (at 227 least partly) mediated through changes in ocean currents. In our simulations, tidal heat-228 ing impacts regional SST mostly by altering ocean surface currents and hence sea-ice dy-229 namics as well as meridional heat transport. Depending on the patterns of the ocean sur-230 face currents and the sea-ice extent, tidal heating can have varying regional impacts. In 231 particular, it is important to note that patterns of tidal heating do not necessarily cor-232 relate with regions of surface warming, as the tidal heat itself is too weak to penetrate 233 to the surface and cause substantial changes in sea-ice cover (see below). Instead, the 234 heating leads to changes in ocean surface currents which can trigger retreat or even ad-235 vance of sea ice and thus local warming or cooling, respectively. 236

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3.2 Climate response to tidal heating depends on background climate

The observation from Section 3.1 that tidal heating has the most significant effect on the climate if it can influence the sea-ice edge also leads to a strong dependence of the response to the background climate state, see Figure 4 for a rotation rate of 12 hours, and Figure B2 for all rotation rates and climate states.

Cold climate states with a large sea-ice extent experience hardly any impact of tidal 242 heating on the ice cover. This is not too surprising, because the minute tidal warming 243 mostly occurs at depth (see also Figure 7a), and because the extended sea-ice cover im-244 pedes strong meridional surface currents and heat exchange between the ocean and the 245 atmosphere in the polar regions. Instead sea-ice cover reaches down to latitudes of roughly 246 50° N/S, respectively, where ocean surface currents are mostly zonal for the continental 247 configuration used in our study (see Figure B1 in Appendix B) and thus do not greatly 248 contribute to meridional heat transport towards the sea-ice edge. 249

Intermediate and warm climate states, on the other hand, show more pronounced impacts of tidal heating on sea-ice cover, as tidal heating influences ocean surface currents close to the sea-ice edge in this case, leading to changes in sea-ice cover and thus



Figure 4. Response to tidal heating depends on background climate state. Changes in seasurface temperature, surface velocity, and sea-ice cover due to tidal heating for a (a) cold, (b) intermediate, and (c) warm climate state and for a rotation rate of 12 h. See caption of Figure 3 for explanation of colors and shading. Maps showing the effects of tidal heating for all rotation rates can be found in Figure B2 in Appendix B.

local climate, amplified by the ice-albedo feedback. Despite the fact that tidal heating 253 is a source of energy to the ocean, the effects on SST and sea-ice extent by tidal heat-254 ing can lead to either regional warming or cooling depending on how ocean surface cur-255 rents are affected. This also strongly depends on the continental configuration, of course. 256 For example, the shape, position, and orientation of the islands at high southern lati-257 tudes leads to flow of warmer surface waters towards the polar region, significantly af-258 fecting local sea-ice cover. The distribution of continents is also responsible for the North-259 South asymmetry in the response to tidal heating: There is slightly more land in the South-260 ern hemisphere leading to an enhanced dissipation and tidal heating closer to the ocean 261 surface (see Figure 7a) and to an offset in the global overturning circulation (see Fig-262 ure 8a). It is therefore worth noting that even a small asymmetry in continental config-263 uration can lead to significant differences in the state of the ocean. In general, climate 264 states with strong meridional surface currents near the sea-ice edge are more vulnera-265 ble to changes in ocean circulation. The impact of tidal heating is therefore strongest 266 in these cases. 267

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3.3 Tidal mixing and geothermal heat are more important than tidal heating

Despite the regional variations described above, the global impact of tidal heating 270 on surface air temperatures, SSTs, and sea-ice cover is almost negligible, at least for the 271 investigated continental configuration, tidal scenarios, and climate states. Tidally induced 272 changes in global mean surface air temperatures vary between -0.022° C and $+0.035^{\circ}$ C. 273 depending on tidal scenario and baseline state (for comparison, the standard deviation 274 of the global mean temperature due to natural climate variability in the control simu-275 lations ranges from 0.003°C to 0.048°C). The mean difference in global sea-ice over all 276 scenarios and states is about $+8.6 \times 10^4$ km² equivalent to a very small gain of 0.25% 277 of the total sea-ice area. 278

As we have seen in Section 3.2, for a cold climate state and 12 hour rotation rate, 279 for example, the influence of tidal heating is insignificant in the case of the cold climate 280 state (Figures 4a and 5a) with the global and annual mean surface air temperature re-281 maining the same as for the control simulation $(9.2^{\circ}C)$. The reason becomes obvious when 282 looking at the temperature distribution with depth and latitude (Figure 7a). As the tidal 283 heating is small (fractions of a Wm^{-2}), the heat itself is not large enough to penetrate 284 to the ocean surface and directly influence SSTs unless the energy input occurs close to 285 the ocean surface. In some regions, the energy input from tidal heating leads to a small 286 warming at depth, other regions show slight cooling due to changes in ocean circulation. 287 In general, these temperature changes are small $(|\Delta T| < 0.05^{\circ} \text{C})$ and cannot signifi-288 cantly affect the surface climate. Furthermore, tidal heating does not significantly change 289 the amplitude or structure of the global overturning (see Figure 8a–b). 290

The response of the climate to tides changes significantly when taking the effect 291 of tidal mixing into account in addition to just tidal heating (see Figures 5b and 6b for 292 the cold and warm climate state, respectively). In this case, the global climate is signif-293 icantly warmer, with a global and annual mean surface air temperature of 10°C com-294 pared to 9.2°C in the control simulation for the cold climate state. Generally speaking, 295 tidally enhanced ocean mixing will transport the tidal heat from the deeper ocean lay-296 ers towards the surface and can thus affect the surface climate either directly or via changes 297 in ocean currents or sea-ice cover, amplified by the ice-albedo feedback. This effect can 298 clearly be seen in the depth-latitude temperature section (Figure 7b). For all climate states 299 and rotation rates, the changes in global and annual mean surface air temperatures com-300 pared to the respective control simulations range from -0.013° C to $+0.72^{\circ}$ C when tidal 301 mixing is included. Global sea-ice area changes by about -3.8×10^6 km², equivalent to 302 a loss of 11% of the total sea-ice area of the corresponding control simulation. 303



Figure 5. Tidal mixing and geothermal heat flow are more important than tidal heating (cold climate state). Changes in sea-surface temperature, surface velocity, and sea-ice cover for a cold climate state with 12 h rotation rate and due to (a) tidal heating only, (b) tidal heating and mixing, and (c) tidal heating, mixing, and geothermal heat flow. See caption of Figure 3 for explanation of colors and shading. Appendix B contains all maps showing the effects of tidal heating (Figure B2), tidal heating and mixing (Figure B3), and tidal heating, mixing, and geothermal heat flux (Figure B4), respectively.



Figure 6. Tidal mixing and geothermal heat flow are more important than tidal heating (warm climate state). Same as Figure 5 but for a warm climate state. Appendix B contains all maps showing the effects of tidal heating (Figure B2), tidal heating and mixing (Figure B3), and tidal heating, mixing, and geothermal heat flux (Figure B4), respectively.

Adding geothermal heat flux on top of tidal heating and mixing leads to even more 304 pronounced changes in the surface climate (see Figures 5c and 6c for the cold and warm 305 climate state, respectively), with global mean surface air temperature changes relative 306 to the control simulations in the range of $+0.13^{\circ}$ C to $+1.7^{\circ}$ C depending on the tidal sce-307 nario and climate state. The mean change in sea-ice area is also more pronounced in this 308 case, about -6.9×10^6 km², equivalent to a loss of 19% of the total sea-ice area in the 309 control simulations. The more pronounced response is hardly surprising given that the 310 geothermal heat flux is one order of magnitude larger than the tidal heating (see Table 1). 311 For the cold climate case, for example, the global and annual mean surface air temper-312 ature increases to 11°C, and thus 1.0°C higher than for the simulation with tidal heat-313 ing plus mixing or 1.8° C higher than the control simulation. Compared to the simula-314 tion with tidal heating, the input of heat is more uniform and generally at larger depths, 315 leading to an overall warming of the ocean on top of the changes induced by tidal heat-316 ing and mixing (see Figure 7c). 317

The global overturning streamfunction (Figure 8) shows a northern hemisphere clock-318 wise cell with a maximum of about 25 Sv between 60-70°N. This shows that deepwater 319 forms in the north and spreads out at depth, much like in the present day North Atlantic. 320 There is a corresponding counter-clockwise cell in the southern hemisphere, but it is much 321 weaker than the cell in the north in the control and tidal heating runs (Figure 8a-b). As 322 tidal mixing is included, the southern cell becomes more energetic, and with all tidal processes and geothermal heat switched on (Figure 8d) there is a deep cell in the southern 324 hemisphere with a maximum of some $20 \,\text{Sv}$ between $30-70^{\circ}\text{S}$. It has been known for some 325 time that the strength of the overturning circulation depends on the level of vertical mix-326 ing (Wunsch & Ferrari, 2004) and in this configuration, that leads to the establishment of the southern hemisphere cell. 328

Compared to the cold climate state (Figure 5), the effects get more pronounced for the warm climate state (Figure 6) as already discussed in Section 3.2 for the effects of tidal heating alone.

332 4 Discussion

To help resolve the FYSP through tidal heating, the tidal heating rate would be 333 required to contribute significantly to Earth's energy balance. Based on global numbers, 334 Archean tidal heating is generally considered to be too low to counteract the low solar 335 input (Feulner, 2012), as even geothermal heat flux is greater by one order of magnitude. 336 Although Heller et al. (2021) argue that tidal heating rates may be underestimated, they 337 also conclude that tidal heating helped to maintain liquid surface water for the first 220 338 Myr of Earth's existence at most, and hence is most important during the Hadean rather 339 than the Archean. 340

Our spatially-resolved simulations of tidal dissipation and its impact on climate con-341 firm that the tidal heat itself does not contribute to a measurable increase in the sur-342 face energy budget. Instead, tidal heating induces local changes in ocean currents and 343 thus sea-ice dynamics and meridional heat transport. These secondary effects then may 344 lead to regional cooling or warming of the ocean surface and associated changes in sea-345 ice cover, which amplify the changes in temperature via the ice-albedo feedback. On the 346 other hand, our results also indicate that a large ice extent makes the climate more ro-347 bust against impacts from tidal heating due to the insulating effect of sea-ice cover. Fur-348 thermore, the impact of tidal heating on the global climate on early Earth is negligible. 349

However, including the effect of tidal dissipation on vertical mixing in the ocean leads to more pronounced warming. Combined with the heating from the higher geothermal heat flux on early Earth, tides can thus contribute to warming the climate under the faint young Sun.



Figure 7. Depth-latitude section of potential temperature changes compared to the control simulation for a cold climate state and (a) tidal heating, (b) tidal heating plus mixing, and (c) tidal heating plus mixing plus geothermal heat flux (corresponding to the reults shown in Figure 5). Note the much smaller temperature changes for just tidal heating in panel (a). Asterisks at the ocean surface indicate areas of at least 50% annual-mean sea-ice cover. The filled triangles in panels (a) and (b) indicate the second quartile of the depth distribution of tidal heating, with the size of the symbol scaled according to the magnitude of maximum heating. The vertical lines indicate the first and third quartiles, respectively. The corresponding combined energy input from tidal heating and geothermal heat is indicated in a similar manner in panel (c).



Figure 8. Global overturning streamfunction for a cold climate state with 12 h rotation rate and (a) the control simulation, (b) tidal heating only, (c) tidal heating plus mixing, and (d) tidal heating, mixing, and geothermal heat flow (the simulations shown in Figures 5 and 7). Colours indicate the strength of the streamfunction. Contours are drawn in intervals of 15 Sv with negative values shown as dashed lines, the zero contour as dotted line, and positive values as solid lines.

Our approach of using a state-of-the-art tidal model and a coupled climate model to investigate the effects of tidal heating on early Earth's climate spatially resolved can certainly be considered an improvement compared to simple global assessments. Nevertheless, our study is just a first step towards a more complete assessment of the influence of tides on early Earth's climate, with a number of questions still to be investigated.

First, one key factor determining tidal dissipation is the distribution of landmasses, 359 which is unknown for early Earth. The continental configuration used in this study is 360 randomly generated due to the lack of knowledge about Hadean and Archean Earth. As 361 the landmasses strongly influence both the magnitude of tidal heating and its potential impact on meridional currents, the results of our study allow an assessment only for this 363 particular continental configuration. Systematic ensemble investigations of different ran-364 dom continental configurations may find that there are scenarios in which lunar tides have 365 a stronger influence on Earth's climate compared to the configuration investigated above, 366 considering that tidal dissipation can vary by three orders of magnitude for different land-367 ocean distributions (Blackledge et al., 2020) and that continental configuration can also 368 significantly affect the snowball bifurcation point (Eberhard et al., 2023). 369

Second, the tides are not implemented dynamically, but the tidal dissipation fields are used to provide an additional heat source at the bottom of the ocean in the tidal heating runs and to drive vertical mixing in the mixing runs. The change in the ocean temperature structure and overturning stream function as tidally driven mixing is implemented (Figures 7–8) show that using physically sound vertical mixing rates for the time period under consideration is necessary; present day mixing rates are poor proxy for past rates.

Third, despite the fact that the tidal heating and mixing are computed using a tidal 376 model with day lengths of 12, 18 and 24 hours, our coupled climate model only simulates 377 the climate dynamics of a present 24 hour length of day. Hence, any consequences of a 378 faster rotation (e.g. the narrowing or appearance of additional atmospheric circulation 379 cells) are not considered. Earlier work has shown that the snowball bifurcation is intri-380 cately linked to the large-scale circulation patterns in the atmosphere (Feulner et al., 2023). 381 In this context it is important to note that there is still a large uncertainty in the evo-382 lution of Earth's day length which translates also into a larger uncertainty in the rate 383 of tidal heating at our chosen points in time. 384

Finally, tides within Earth's mantle were dominant during the freezing of Earth's 385 surface in the first million years of the Hadean (Zahnle et al., 2007). After the mantle 386 became solid, oceanic tides probably were the key driver for tidal heating (Blackledge 387 et al., 2020), and hence our study focuses on ocean tides. More recent research (Daher 388 et al., 2021) also investigated the impact of atmospheric tides which may have helped 389 to stabilize Earth's rotation, but the direct effect of these on the ocean would have been 390 small. A more thorough investigation of tides in a magma ocean would be very inter-391 esting but is beyond our scope here. 392

5 Conclusions

In summary, our study uses spatially resolved maps of tidal dissipation from a stateof-the-art tidal model which represent more realistic tidal energy inputs compared to simple exponential models or the CPL model by Heller et al. (2021) that assume a globally uniform tidal heating. By forcing a coupled climate model with this spatially-resolved tidal heating, tidal mixing, and geothermal heat flux our study makes a step further towards understanding the impacts of tides onto the climate on early Earth and contributes to assessing what role tides could have played in context of the FYSP.

401 Our results show that the global effect of tidal heating is insignificant but that tidal 402 heating can affect sea-ice cover on local and regional scales and account for a substan-403 tial change in sea ice cover. In some scenarios the fraction of ice covered area of a grid

cell is altered by more than 20% due to the tidal heating. This impact is achieved by in-404 ducing changes in the ocean currents and thus in sea-ice dynamics as well as of heat trans-405 port towards the ice line. Tidal heating alone is thus certainly not the solution to the 406 FYSP, as it does not provide enough energy. However, our results show that tidal heat-407 ing can have regional impacts. The fact that tidal heating acts primarily via altering oceanic 408 currents underlines the importance of dynamical features when investigating tidal im-409 pacts. 410

Out study also shows that tidal mixing and geothermal heat flux are much more 411 important than tidal heating alone. In particular the effects of tidal mixing should be 412 included in future studies of deep-time paleoclimate states. With respect to the effect 413 of tides on the climate on early Earth, future studies will have to investigate the effects 414 of a shorter day length on the dynamics of the coupled Earth system under the faint young 415 Sun and subject to stronger tidal forcing. 416

7

7

Appendix A Global mean temperatures for all simulations 417

b) 90 a) Cold climate state & 12h Cold climate state & 18h c) Cold climate state & 24h 7 S 20 nº dì f) iate climate state & 12h ntermediate climate state & 18h Intermediate climate state & 24h G 7 n h) g) i) Warm climate state & 12h Warm climate state & 18h Warm climate state & 24h ٢. 5 150 180 210 Longitude [°] 150 180 210 Longitude [°] ongitude [° 25 SST [°C] → 5 cm s⁻¹ 🔲 land 15 cm s⁻¹ ice

Appendix B Additional figures 418

Figure B1. Control simulations. Absolute sea-surface temperatures (SSTs), surface velocities, and sea-ice cover (grid cells with annual-mean sea-ice fraction of at least 50%) for the nine control simulations without tides and geothermal heat flux.

30 cm s⁻¹

Acronyms 419

SST Sea Surface Temperature 420

Table A1. Global and annual mean surface air temperatures (SATs) and sea-surface temperatures (SSTs) for all simulations. All values are computed over the last 500 years of each simulation.

$ S_0$	CO_2	Simulation	SAT	SST
$[W/m^2]$	[ppm]		$[^{\circ}C]$	[°C]
1010	176000	Control	9.23	12.0
		Tidal heating	9.22	12.0
		Tidal heating $+$ mixing	9.95	12.4
		Tidal heating $+$ mixing $+$ geothermal	11.0	13.1
	200000	Control	14.3	15.3
		Tidal heating	14.2	15.3
		Tidal heating $+$ mixing	14.8	15.8
		Tidal heating $+$ mixing $+$ geothermal	15.0	16.0
	220000	Control	16.7	17.2
		Tidal heating	16.7	17.2
		Tidal heating $+$ mixing	17.0	17.4
		Tidal heating $+$ mixing $+$ geothermal	17.2	17.6
1161	15000	Control	10.7	13.4
		Tidal heating	10.7	13.4
		Tidal heating $+$ mixing	10.8	13.4
		Tidal heating $+$ mixing $+$ geothermal	10.9	13.5
	25000	Control	16.3	17.2
		Tidal heating	16.3	17.2
		Tidal heating $+$ mixing	16.3	17.2
		Tidal heating $+$ mixing $+$ geothermal	16.5	17.3
	28000	Control	17.5	18.0
		Tidal heating	17.5	18.0
		Tidal heating + mixing	17.5	18.0
		Tidal heating + mixing + geothermal	17.6	18.1
1361	140	Control	11.8	14.8
		Tidal heating	11.8	14.8
		Tidal heating $+$ mixing	11.9	14.8
		Tidal heating $+$ mixing $+$ geothermal	12.0	14.9
	277	Control	16.1	17.6
		Tidal heating	16.2	17.6
		Tidal heating $+$ mixing	16.2	17.7
		Tidal heating $+$ mixing $+$ geothermal	16.3	17.7
	350	Control	17.7	18.7
		Tidal heating	17.7	18.7
		Tidal heating $+$ mixing	18.0	18.9
		Tidal heating $+$ mixing $+$ geothermal	18.1	19.0

- 421 **CLIMBER** CLIMate-BiosphERe Model
- 422 **EMIC** Earth system Model of Intermediate Complexity
- ⁴²³ **FYSP** Faint Young Sun Problem
- 424 GCM General Circulation Model
- 425 AOGCM Atmospheric-Oceanic General Circulation Model



Figure B2. Simulations with tidal heating. Changes in sea-surface temperature, surface velocity, and sea-ice cover due to tidal heating for three different tidal scenarios with (a,d,g) 12 h, (b,e,h) 18 h, and (c,f,i) 24 h rotation rate and for (a,b,c) low, (d,e,f) intermediate, and (g,h,i)high CO₂ levels corresponding to cold, intermediate, and warm climate states, respectively. The blue and red colors correspond to changes in SST when introducing tidal heating. Yellow areas are land; semi-transparent white areas correspond to regions that are covered have at least 50% sea-ice fraction in the annual mean. Dotted and hatched areas represent regions of sea-ice loss or gain, respectively. Arrows indicate a change in ocean surface currents, with their length scaling with the change in speed. Figure 3 corresponds to the panels in the second row, Figure 4 to the ones in the first column.

426 Open Research Section

All model input and output files as well as the pre-processing and post-processing scripts used to generate model input and the figures in the paper are published with DOI 10.5880/PIK.2023.004 via the GFZ data portal. The DOI will be activated upon publication of the manuscript. The data is already available at http://www.pik-potsdam .de/data/doi/10.5880/PIK.2023.004/ (preview URL). The source code for the model used in this study is archived at the Potsdam Institute for Climate Impact Research and is made available upon request.

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Figure B3. Simulations with tidal heating and mixing. Same as Figure B2 but for the simulations also including the effects of tidal mixing. Note that the mixing parameters for 18 hours daylength are the same as the default model parameters.

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Figure B4. Simulations with tidal heating, mixing, and geothermal heat flux. Same as Figure B3 but for the simulations also including the effects of geothermal heat flux.

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Figure 1.



dissipation [W/m²]

Figure 2.



Figure 3.



Figure 4.



Figure 5.



Figure 6.



Figure 7.



Figure 8.



Figure B1.





Figure B2.





Figure B3.





Figure B4.





