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Simulating or prescribing the influence of tides on the Amundsen Sea ice shelves

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Abstract

The representation of tides in regional ocean simulations of the Amundsen Sea enhances ice-shelf melting, with weakest effects for Pine Island and Thwaites ($< +10\%$) and strongest effects for Dotson, Cosgrove and Abbot ($> +30\%$). Tides increase vertical mixing throughout the water column along the continental shelf break. Diurnal tides induce topographically trapped vorticity waves along the continental shelf break, likely underpinning the tidal rectification (residual circulation) simulated in the Dotson-Getz Trough. However, the primary effect by which tides affect ice-shelf melting is the increase of ice/ocean exchanges, rather than the modification of water masses on the continental shelf. Tide-induced velocities strengthen turbulent heat fluxes at the ice/ocean interface, thereby increasing melt rates. Approximately a third of this effect is counterbalanced by the resulting release of cold melt water that reduces melt downstream along the meltwater flow. The relatively weak tide-induced melting underneath Pine Island and Thwaites could be partly related to their particularly thick water column, which limits the presence of quarter wavelength tidal resonance. No sensitivity to the position of Pine Island and Thwaites with respect to the M_2 critical latitude is found. We refine and evaluate existing methodologies to prescribe the effect of tides on ice-shelf melt rates in ocean models that do not explicitly

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include tidal forcing. The best results are obtained by prescribing spatially-dependent tidal top-boundary-layer velocities in the melt equations. These velocities can be approximated as a linear function of existing barotropic tidal solutions. A correction factor needs to be applied to account for the additional melt-induced circulation associated with tides and to reproduce the relative importance of dynamical and thermodynamical processes.

Keywords: Amundsen Sea, tides, ice shelf melt, ice shelf cavity, NEMO, Pine Island, Thwaites

1. Introduction

The interactions between the Southern Ocean and the Antarctic Ice Sheet remain poorly understood and simulated in spite of their importance for the ice sheet stability and associated global sea level rise (e.g., Jacobs et al., 2012; Asay-Davis et al., 2017; Turner et al., 2017). Part of the difficulty in simulating oceanic melt rates lies in the absence or poor representation of tides in the ocean models that are used for climate projections (e.g., Dinniman et al., 2016). Ice shelf cavities are usually represented in barotropic tide models (e.g., Padman et al., 2002; Carrère et al., 2012), but with no thermodynamical ice/ocean interactions. Several regional ocean models resolving the primitive equations and enabling thermodynamical ice/ocean interactions have been used to estimate tide-induced melting (e.g., Makinson et al., 2011; Mueller et al., 2012, 2018; Galton-Fenzi et al., 2012; Robertson, 2013; Mack et al., 2017). This was done by imposing the solution of one of the aforementioned barotropic tide models at the regional model lateral boundaries. However, these studies were limited to individual ice shelves and their close vicinity, and some did not account for variable atmospheric forcing. Several Ocean General Circulation Models (OGCMs) have the capability to represent aspects of the tidal signal in global ocean simulations (e.g., Savage et al., 2017; Stewart et al., 2018) but some processes are missing (see below), and so far, tides have not been used in OGCMs that represent ocean/ice-shelf interactions. In fact, most OGCMs or large-extent ocean regional models that include thermodynamical ocean/ice-shelf interactions rely on parameterizations to account for melting modulation by tides (see section 2), yet an assessment of these parameterizations is lacking.

In their review of tidal effects on ice sheets, Padman et al. (2018) distinguish tidal processes occurring seaward of the ice shelf, such as tidal vertical

28 mixing and residual currents, from those directly affecting heat exchanges at
29 the ocean/ice-shelf interface. Tidal vertical mixing is caused by (i) the ver-
30 tical shear as barotropic tidal currents rub upon the seafloor (in particular
31 in shallow areas where these currents tend to be relatively swift); and by (ii)
32 the breaking of internal tidal waves (also called baroclinic tides or internal
33 tides) generated by the interaction of barotropic tidal currents with steep
34 topography. Internal tides freely propagate equatorward of a critical latitude
35 at which the tidal frequency equals the inertial frequency. Critical latitudes
36 are 74.5° for M_2 (principal lunar, semi-diurnal), 85.7° for S_2 (principal so-
37 lar, semi-diurnal), and near 27 to 30° for diurnal constituents dominated
38 by O_1 (principal lunar declinational) and K_1 (luni-solar declinational) (e.g.,
39 Cartwright, 1977; Furevik and Foldvik, 1996). Poleward of the critical lati-
40 tude, internal tides are trapped, i.e. they cannot propagate away from the
41 topography where they are generated, and therefore, they only induce ver-
42 tical mixing close to their generation site. Tidal vertical mixing has been
43 suggested to mix the relatively cold ice-shelf melt water with the underlying
44 and relatively warmer HSSW (High Salinity Shelf Water) in the Ross and
45 Weddell Seas, thereby increasing ice-shelf melt (MacAyeal, 1984b; Scheduikat
46 and Olbers, 1990; Makinson and Nicholls, 1999). Tides not only induce mix-
47 ing but also generate a mean residual circulation through the Stokes drift and
48 non-linear dynamics (Longuet-Higgins, 1969; Zimmerman, 1979). Residual
49 transports of a few tenths of a Sverdrup ($1 \text{ Sv} = 10^6 \text{ m}^3 \cdot \text{s}^{-1}$) affect heat and
50 salt exchanges across the Ross and Weddell Sea continental shelves (Makin-
51 son and Nicholls, 1999; Padman et al., 2009; Wang et al., 2013), but are
52 thought to be very small in all the other Antarctic Seas (Bessières et al.,
53 2008). Heat and salt exchange across ice-shelf edges can also be significantly
54 influenced by the tidal residual circulation (e.g., Makinson and Nicholls, 1999,
55 for the Filchner-Ronne ice shelf).

56 Within ice shelf cavities, tides primarily affect ice/ocean interactions by
57 increasing velocities and therefore turbulent exchanges along the ice base.
58 This effect is relatively more important for large and cold cavities, such as
59 Ross and Filchner-Ronne, where tidal currents can be significantly stronger
60 than buoyancy-driven currents (e.g. MacAyeal, 1984a; Padman et al., 2003;
61 Makinson et al., 2011), in contrast to other Antarctic ice shelves (e.g., Hemer
62 et al., 2006; Robertson, 2013). Finally, the extra melting caused by tides
63 induces an additional buoyancy-driven residual circulation, which in turn in-
64 creases ice-shelf melting (MacAyeal, 1984b; Makinson and Nicholls, 1999).
65 Estimating the relative importance of each of these tidal processes is a pre-

66 requisite for better prescribing or parameterizing the effect of tides on ice
67 shelf cavities.

68 While some models have the capability to explicitly simulate tides, pre-
69 scribing or parameterizing their effects can be desirable for a range of ap-
70 plications. First, parameterizing their effects obviates the need for filter-
71 ing model outputs for diagnostics over periods shorter than a few months
72 (which has to be done to avoid aliasing if an explicit representation of tides
73 is used). Second, even in OGCMs capable of representing tides, several
74 processes such as loading and self-attraction are sometimes not accounted
75 for (by contrast with some barotropic tide models). Finally, vertical mix-
76 ing due to internal tides is usually not adequately simulated because ocean
77 models do not resolve the cascade of energy conducive to the breaking of
78 internal tides (Müller, 2013), and spurious diapycnal mixing can be caused
79 by internal tides in level-coordinate models (Leclair and Madec, 2011). Sev-
80 eral kinds of tidal mixing parameterization have been proposed so far for
81 three-dimensional ocean models. Some of them enhance vertical diffusivity
82 to account for the dissipation of barotropic (Lee et al., 2006; Holloway and
83 Proshutinsky, 2007) or baroclinic (Simmons et al., 2004; Olbers and Eden,
84 2013) tides. It is also possible to parameterize the effect of residual tidal
85 currents in a three-dimensional ocean model by adding velocities from the
86 solution of a barotropic tidal model to the Eulerian velocities used in the
87 equations of tracer advection (Bessières et al., 2008). So far, most modelling
88 studies dedicated to ice-shelf/ocean interactions have, instead, parameterized
89 the influence of tides within the formulation of the ice/ocean heat and salt
90 fluxes in the top boundary layer of ice shelf cavities (Timmermann et al.,
91 2002; Jenkins et al., 2010; Hattermann et al., 2014; Asay-Davis et al., 2016).

92 In this paper, we focus on the Amundsen Sea Embayment where the rela-
93 tively warm ocean has a high potential to trigger marine ice sheet instabilities
94 (e.g., Weertman, 1974; Schoof, 2007; Durand et al., 2009; Favier et al., 2014;
95 Joughin et al., 2014; Mougnot et al., 2014), and where melt rates therefore
96 need to be accurately simulated. In the Amundsen Sea, relatively warm Cir-
97 cumpolar Deep Water (CDW) penetrates into the ice shelf cavities, leading to
98 the highest melt rates in Antarctica (e.g., Jacobs et al., 2012; Turner et al.,
99 2017). As a consequence, the melt-induced circulation is particularly strong
100 within ice-shelf cavities and in their vicinity (e.g., Jenkins, 1999; Dutrieux
101 et al., 2014; Jourdain et al., 2017). As tides are also weaker than in the Ross
102 and Weddell Seas (e.g., Padman et al., 2018), their effect on the Amundsen
103 Sea circulation and ice shelf melt is often neglected. The inclusion of tides in

104 an ocean model nonetheless indicated a significant increase in melt rates un-
 105 derneath some ice shelves, by 15% and 52% for Getz and Dotson ice shelves
 106 respectively (Robertson, 2013, , hereafter R2013). By contrast, R2013 did
 107 not find a strong effect on melt rates beneath Pine Island ice shelf (see ice
 108 shelf locations in Fig. 1). R2013 argued that the small effect of tides on Pine
 109 Island was due to its location poleward of the M_2 critical latitude (red line
 110 in Fig. 1), inducing a more vertically uniform tidal flow and therefore weaker
 111 tidal currents near the ice shelf base. Here we propose a new estimation of
 112 tide-induced ice-shelf melting in the Amundsen Sea sector, accounting for
 113 synoptic and seasonal atmosphere and sea ice variability, and considering a
 114 large regional domain in order to analyze tidal processes both seaward and
 115 underneath ice shelves. We identify the most important impacts of tides on
 116 ice shelf melting in our simulations, and use this as a basis for evaluating and
 117 improving methods to prescribe the effects of tides in an ocean model (e.g.,
 118 OGCM) that would not simulate them explicitly.

119 2. Model experiments

120 We make use of NEMO-3.6 (Nucleus for European Modelling of the
 121 Ocean; Madec and NEMO-team, 2016) that includes the ocean model OPA
 122 (Océan Parallélisé) and the Louvain-la-Neuve sea-ice model LIM-3.6 (Rous-
 123 set et al., 2015) with a single ice category. We use the same AMU12.L75
 124 regional configuration as Jourdain et al. (2017), with z-coordinates and an
 125 isotropic horizontal resolution of ~ 3 km along the continental shelf break,
 126 thus resolving the first few vertical normal modes of internal tides. Our set-
 127 up includes a split-explicit free surface formulation and a representation of
 128 ice-shelf cavities (Mathiot et al., 2017). The depth of the ocean/ice-shelf
 129 interface is calculated through the free surface formulation and fluctuates
 130 about the hydrostatic equilibrium position of ice within a reference ocean
 131 density profile. It is thus assumed that the ice flexural rigidity does not
 132 affect vertical motions of the ice/ocean interface. In our configuration, we
 133 assume a constant top-boundary-layer (TBL) thickness of 20 m and the heat
 134 flux through the TBL is expressed as in McPhee et al. (2008) and Jenkins
 135 et al. (2010):

$$136 \quad Q = \rho_w c_{pw} \Gamma_T u_* (T_{\text{TBL}} - T_f) \quad (1)$$

137 where Γ_T is a constant and uniform heat exchange coefficient, u_* the friction
 138 velocity, ρ_w and c_{pw} the density and heat capacity of sea water, and $(T_{\text{TBL}} - T_f)$

139 is the difference between the TBL temperature and the freezing point at
140 the ice draft depth. A similar expression is used for salinity. Simulated
141 temperatures, salinities and velocities are averaged over the imposed TBL
142 thickness to calculate the heat flux in Eq. (1) or in the equivalent expression
143 for salinity as in Losch (2008).

144 All the experiments used in this paper are summarized in Tab. 1. To
145 assess the representation of tides in our regional model configuration and
146 to describe the characteristics of barotropic tides, we first run a pseudo-
147 barotropic simulation (referred to as BTP-07) in which the sea ice model
148 is switched off, heat fluxes are set to zero at the ocean/ice-shelf interface,
149 no atmospheric forcing is used, and lateral boundaries other than tidal con-
150 stituents are set to zero velocities, with constant and uniform temperature
151 and salinity profiles (equal to the domain-mean initial state values), in a
152 similar way as Maraldi et al. (2013) and Mueller et al. (2012, 2018). The
153 only external forcing consists of the amplitude and phase of seven tidal con-
154 stituents (M_2 , S_2 , N_2 , K_1 , O_1 , Q_1 , M_4), which are used as lateral boundary
155 conditions for sea surface height and barotropic velocities. It is therefore
156 assumed that the tidal signal is forced remotely, and that the gravitational
157 forcing within the regional domain is negligible. The tidal constituents used
158 as lateral boundary conditions are interpolated from the global Finite El-
159 ement Solution FES2012 (Carrère et al., 2012; Lyard et al., 2006). The
160 latter comes from the resolution of the tidal barotropic equations in the
161 spectral domain with assimilation of long-term altimetry data. To com-
162 pare our results to an independent tide dataset, we also use the Circum-
163 Antarctic Tidal Simulation CATS2008, which is an inverse model that as-
164 simulates altimetry data and in-situ tide records (updated version of Pad-
165 man et al., 2002, [https://www.esr.org/research/polar-tide-models/
list-of-polar-tide-models/cats2008](https://www.esr.org/research/polar-tide-models/list-of-polar-tide-models/cats2008)). Both CATS2008 and a former
166 version of FES were found to be relatively accurate near Dotson ice shelf
167 despite the lack of assimilated data in the Amundsen Sea (McMillan et al.,
168 2011). Our harmonic analyses are done after 2 years of spin-up, and are
169 based on 15-minute samples analyzed over a 190-day window. With these
170 characteristics, our harmonic analyses can accurately separate at least 7 tidal
171 constituents.
172

173 The second set of experiments is designed to represent more realistic
174 conditions, with fully-coupled sea-ice and thermodynamically-coupled ice-
175 shelf cavities. Lateral boundaries (temperature, salinity, velocities) are from
176 the global 0.25° simulation described by Spence et al. (2014). Importantly,

177 that global simulation does not represent tides. Both the global simulation
178 and the regional experiments are forced by atmospheric fields from version 2
179 of the Coordinated Ocean-ice Reference Experiments (CORE-2) Normal Year
180 Forcing (NYF) (Griffies et al., 2009; Large and Yeager, 2009). The reference
181 experiment (REF), which includes no tidal constituents, is the one described
182 and evaluated with respect to CTD, ice-shelf melt rates, and sea ice cover
183 by Jourdain et al. (2017). The mean barotropic circulation in REF is shown
184 in Fig. 1. We then ran three additional simulations accounting for one, four,
185 and 18 tidal constituents. The latter are imposed through lateral boundary
186 conditions as for BTP-07, but on top of other oceanic boundary conditions
187 from the global 0.25° simulation. The TIDE-M2 simulation only includes
188 the M₂ tidal constituent because its importance was emphasized in R2013.
189 The TIDE-04 simulation additionally includes S₂, K₁, and O₁. Finally, the
190 TIDE-18 simulation includes the four aforementioned constituents and N₂,
191 K₂, P₁, Q₁, N₁, 2N₂, μ₂, ν₂, L₂, T₂, M₄, M_f, M_m and M_{tm} (e.g. Schureman,
192 1958).

193 A third set of experiments is designed to propose a method to prescribe
194 the effect of tides on ice shelf melt. It is based on our most realistic configu-
195 ration (i.e. with stratification, sea ice, atmospheric forcing), but has no tides
196 prescribed at its boundaries. A first possibility to account for non-resolved
197 tides in (1) is to use a velocity-independent formulation of Q :

$$198 \quad Q = \rho_w c_{pw} \gamma_{T0} (T_{TBL} - T_f) \quad (2)$$

199 where γ_{T0} is a constant referred to as heat exchange velocity. The γ_{T0} value
200 can be chosen to represent non-resolved tidal velocities in cavities where tides
201 dominate the circulation (Timmermann et al., 2002), and can also represent
202 the poorly captured buoyancy-driven circulation in coarse-resolution mod-
203 els. However, a majority of recent simulations have preferred a velocity-
204 dependent formulation of ice-shelf/ocean heat fluxes (Eq. 1), because it
205 better accounts for non-uniform circulation and melt-circulation feedbacks
206 (Dansereau et al., 2014; Asay-Davis et al., 2017; Donat-Magnin et al., 2017).
207 We therefore opt for another method in which the expression of u_\star in (1) is
208 modified as follows:

$$209 \quad u_\star = \sqrt{C_d (u_{TBL}^2 + u_{tide}^2)} \quad (3)$$

210 where C_d is the drag coefficient, u_{TBL} the TBL velocity simulated by NEMO
211 (averaged within the imposed TBL thickness), and u_{tide} is a prescribed “tidal
212 TBL velocity” (e.g. Jenkins et al., 2010; Hattermann et al., 2014; Asay-Davis

213 et al., 2016). To our knowledge, such parameterization of tidal effects has
 214 always been used at a single point or with a uniform tidal TBL velocity,
 215 although Jenkins et al. (2010) noted that such velocity would ideally be
 216 a spatially varying quantity. It is presently unclear to what extent such
 217 prescription of mean-square tidal TBL velocity adequately accounts for the
 218 various processes involved in the overall interactions between tides and ice-
 219 shelves. Here we assess three different definitions of the tidal TBL velocity:
 220 uniform velocity for all the ice shelves in the domain (Utide-UNIF), uniform
 221 velocity under each ice shelf, but with a value specific to each ice shelf (Utide-
 222 PERISF), and 2-dimensional local velocities (Utide(x,y)). The methods used
 223 to calculate these prescribed velocities are discussed along with the results.

224 All the analyses performed on the second and third sets of experiments
 225 are done after a spin up of 6 years, which is sufficient to reach a steady state
 226 (Jourdain et al., 2017). We analyze yearly averages over the 7th year. One-
 227 year average is enough to avoid any significant aliasing by the modulation of
 228 the four main tidal harmonics by the other harmonics represented here (as
 229 indicated by Tab. 4-7 of Müller, 2013).

230 3. Results

231 3.1. Harmonic decomposition of barotropic tides

232 We start with a description of the main characteristics of barotropic tides
 233 in the Amundsen Sea. A Sea Surface Height (SSH) harmonic analysis of
 234 the BTP-07 simulation (Fig. 2) indicates a stronger amplitude for the diur-
 235 nal (O_1 and K_1) than for the semi-diurnal (M_2 , S_2 , N_2) harmonics in the
 236 Amundsen Sea (Fig. 2). The M_4 constituent, resulting from an interaction
 237 of M_2 with itself, is significantly weaker than the other six constituents. The
 238 diurnal constituents generate topographically trapped vorticity waves along
 239 the continental shelf break (stronger amplitude and closed cotidal lines in
 240 Fig. 2a,b), as previously described for other locations poleward of the diurnal
 241 critical latitudes (e.g. Middleton et al., 1987; Padman and Kottmeier, 2000;
 242 Padman et al., 2003). The semi-diurnal constituents experience a stronger
 243 signal in the cavities than on the continental shelf seaward of the ice fronts
 244 (Fig. 2c,d,e), which is reminiscent of tide resonance in semi-enclosed bays.
 245 Such resonance can occur for bays of characteristic size close to the tidal
 246 quarter wavelength which can be estimated as $\sqrt{gH}/\nu_{\text{tide}}$, where g is the
 247 gravity acceleration, H the water column thickness, and ν_{tide} the tidal fre-
 248 quency (Fig. 3). The qualitative match between the semi-diurnal quarter

249 wave lengths and the cavity sizes (Fig. 3b) supports the similarity between
250 the resonance in semi-enclosed bays and the resonance in ice-shelf cavities.
251 For diurnal constituents, quarter wave lengths are longer than most typical
252 ice-shelf sizes, so that cavities appear less resonant. There is weak resonance
253 in Thwaites and in the core of Pine Island: the cavities are deep and there-
254 fore present a quarter wave length much longer than the cavity size. Finally,
255 considering the entire Amundsen Sea Embayment in Fig. 3, the continental
256 shelf appears too narrow or too deep to favour a widespread resonance of
257 diurnal and semi-diurnal tides in this region.

258 The harmonic analysis of barotropic velocities in the BTP-07 simulation
259 indicates strong amplitude along the continental shelf break for the diur-
260 nal constituents (up to 25 cm.s^{-1} in Fig. 4a,b). These patterns are consis-
261 tent with the aforementioned topographically trapped vorticity waves and
262 are expected to induce vertical mixing as they dissipate locally. For most
263 constituents, strong barotropic velocities are also generated within ice shelf
264 cavities, with the exception of Thwaites and the main trunk of Pine Is-
265 land cavity (Fig. 4), likely because of the aforementioned reduced quarter
266 wavelength resonance. Such strong tidal velocities under the ice shelves are
267 expected to increase TBL velocities, and therefore melt rates.

268 The tidal residual circulation is now estimated using the same methodol-
269 ogy as Bessières et al. (2008), i.e. through the annual mean barotropic stream
270 function (Fig. 5). The main feature is a strengthening of the Antarctic Slope
271 Current (westward jet along the continental shelf break, e.g. Heywood et al.,
272 1998; Mathiot et al., 2011) by up to 0.8 Sv (maximum barotropic stream
273 function difference between 2 points). Locally, the residual circulation can be
274 more important than the background ocean circulation, e.g. 0.70 Sv residual
275 transport through the green section in Fig. 5 vs 0.49 Sv in the REF simula-
276 tion (Fig. 1). This effect is locally important, but remains small compared to
277 the total barotropic transport by the Slope Current which is approximately
278 12 Sv along the entire slope of the continental shelf (e.g., 0.12 Sv residual
279 in Fig. 5 vs 11.7 Sv through the cyan section in Fig. 1a). There is also a
280 southward residual flow on the eastern flank of Dotson-Getz Trough. It reach
281 approximately 0.2 Sv near the continental shelf break, 0.17 Sv through the
282 magenta section, and gradually decreases to zero near the ice shelves. The
283 associated transport has a similar magnitude as the transport in the absence
284 of tides at this location (e.g., 0.13 Sv through the magenta section in Fig. 1b).
285 These residual circulations were not found in Bessières et al. (2008), possi-
286 bly due to their different set-up (e.g. their coarser horizontal resolution).

287 Underneath most ice shelves, the residual tidal circulation is typically a few
288 mSV, which represents only a few percent of the circulation in the presence
289 of ice shelf melt (Tab. 2). This effect is slightly stronger for Getz and Abbot,
290 where the residual tidal circulation reaches ~ 20 mSv, representing $\sim 10\%$ of
291 the circulation in the presence of ice shelf melt (Tab. 2).

292 We now compare the tide characteristics in the NEMO pseudo-barotropic
293 simulation to the barotropic tide models, in order to assess NEMO's ability to
294 rebuild the tidal signals from the lateral boundary conditions. Overall, there
295 is a fair agreement between the SSH harmonics from NEMO and those from
296 FES2012 and CATS2008, both in terms of phase and amplitude (Fig. S1-
297 S3). An exception is the phase of M_2 that is quite different between NEMO
298 and the tide models, but also between FES2012 and CATS2008. A possible
299 explanation is a particularly strong sensitivity to the different bathymetries
300 used in these models near the critical latitude. The amplitude of diurnal
301 SSH harmonics is slightly weaker in NEMO than in FES2012 and CATS2008
302 (Fig. S1ab-S3ab), which could be related to the absence of direct tide influ-
303 ence (gravitational potential) on the water mass located within our simula-
304 tion domain. It is unclear why the resonance of semi-diurnal constituents
305 within the ice shelf cavities is significantly less prominent in FES2012 and
306 CATS2008 than in NEMO (Fig. S1cde-S3cde). The harmonics of barotropic
307 velocities in NEMO have a pattern that better matches CATS2008 than
308 FES2012, but the maximum velocity amplitude along the continental shelf
309 break and within ice-shelf cavities in NEMO is more similar to FES2012
310 (~ 25 cm.s $^{-1}$, not shown) than to CATS2008 (~ 16 cm.s $^{-1}$, not shown).

311 *3.2. Impact of tides on ice-shelf melt*

312 R2013 has been the only study so far to estimate the effects of tides on ice-
313 shelf melt in the Amundsen Sea sector. Besides a different ocean model used,
314 that study did not account for sea ice or for variability in the atmospheric
315 forcing. We therefore revisit the influence of tides in this sector with a more
316 realistic model set-up including sea ice and an atmosphere that varies at
317 seasonal and synoptic scales. As R2013 emphasized the importance of the
318 presence of the M_2 critical latitude near the ice shelves of the Amundsen
319 Sea, we first run a simulation that is only forced by M_2 . Then, we add three
320 additional constituents, K_1 , O_1 , and S_2 , which have stronger amplitudes than
321 M_2 , to get the same set of four harmonics as used by R2013. Finally, we use
322 a total of 18 harmonics (see section 2) to estimate the influence of a more
323 complete tidal signal on melt rates.

324 The effect of 18 tidal harmonics is shown in Fig. 6a. Tides generally
 325 increase melt rates (Tab. 3), although a weak decrease is found at a few lo-
 326 cations, mostly in the Getz cavity. However, even with 18 tidal constituents,
 327 the increase is negligible for Pine Island (Tab. 3) and small for Thwaites
 328 (+7.8%), which is consistent with the weak tidal amplitude in these two
 329 cavities. By contrast, the relative increase is larger than 30% for Dotson,
 330 Cosgrove and Abbot. These results have to be considered carefully because
 331 of the high uncertainty on the bathymetry under these ice shelves. Using
 332 four tidal constituents instead of 18 induces errors below 6%, so restrict-
 333 ing a tidal analysis to four constituents as in R2013 seems very reasonable.
 334 Tide-induced melt is generally not located near grounding lines (Fig. 6a),
 335 probably because tidal currents often become weak in the shallowest parts
 336 of small cavities.

337 Our results with four constituents are in qualitative agreement with R2013
 338 on the relative importance of tides for three different cavities, although we
 339 find slightly weaker effects. R2013 found that tides were responsible for
 340 moderate change in the Getz cavity (+24 Gt.yr⁻¹, +15%), stronger change
 341 in Dotson cavity (+11 Gt.yr⁻¹, +52%), and little change in Pine Island cav-
 342 ity (-2 Gt.yr⁻¹, -3%). R2013 also emphasized the importance of M₂ and
 343 the associated critical latitude. Our simulation in which only M₂ is rep-
 344 resented indicates that this constituent only accounts for a limited part of
 345 the tidal influence on melt rates. This was expected from the dominance
 346 of diurnal tides and of S₂ over M₂ revealed by the pseudo-barotropic simu-
 347 lations. Further, we repeated R2013's experiment where the latitudes were
 348 shifted by 1°S (i.e. bathymetry and ice drafts are shifted by 1°N). We ran
 349 shifted-latitude experiments with and without tides to isolate the effect on
 350 the geostrophic circulation from the critical latitude effects. In the absence
 351 of tides, shifting latitudes enhances mean melt rates in all the cavities, but
 352 the increase is weaker than 2.5% everywhere (Fig. 7a). In the presence of
 353 tides, shifting latitudes affect melt rates by less than 1% in all the cavities
 354 (Fig. 7b). This implies that the M₂ critical latitude is of limited importance
 355 to the ice-shelf melting in this region. A possible explanation for the opposite
 356 conclusion of R2013 is that our atmospheric forcing varies at synoptic and
 357 seasonal scales, which may lead to a large variability of the effective critical
 358 latitude (accounting for the ocean relative vorticity). This result is reminis-
 359 cent of Richet et al. (2017) who analyzed simulations with weakened critical
 360 latitude effects on tidal energy dissipation in the presence of a background
 361 ocean circulation.

362 We now look further into the physical mechanisms through which tides
 363 affect ice-shelf melt rates. Neglecting heat diffusion in ice (which yields no
 364 more than 10-20% error according to Dinniman et al. (2016) and Arzeno et al.
 365 (2014)), the melt rate m (in meters of ice per second) can be approximated
 366 as:

$$\rho_i m \simeq \rho_w c_{pw} \gamma_T (T_{\text{TBL}} - T_f) / L_f \quad (4)$$

367 where ρ_i is the ice density, L_f the latent heat associated with melting/freezing,
 368 $\gamma_T = \Gamma_T u^*$, and other variables have been introduced in (1). Expressing γ_T in
 369 the presence of explicit tides as $(\gamma_T)_{\text{tide}} = (\gamma_T)_{\text{no tide}} + \Delta\gamma_T$ in (4), and similarly
 370 for the thermal forcing, we can define a dynamical/thermodynamical decom-
 371 position that explains the differences in melt rates between the simulations
 372 with and without tides:

$$\left\{ \begin{array}{l} \rho_i \Delta m = \rho_i (m_{\text{tide}} - m_{\text{no tide}}) \\ \simeq \rho_w c_{pw} (T_{\text{TBL}} - T_f) \Delta\gamma_T / L_f \quad (\text{dynamical}) \\ + \rho_w c_{pw} \gamma_T \Delta(T_{\text{TBL}} - T_f) / L_f \quad (\text{thermodynamical}) \\ + \rho_w c_{pw} \Delta\gamma_T \Delta(T_{\text{TBL}} - T_f) / L_f \quad (\text{covariational}) \end{array} \right. \quad (5)$$

373 The decomposition for the simulation with 18 tidal harmonics indicates
 374 that tidal velocities in the TBL of ice-shelf cavities are the main driver of in-
 375 creased melt rates (the dynamical term dominates in Fig. 6). This dynamical
 376 term includes the mean effect of tidal velocities but also the effect of the ad-
 377 ditional buoyancy-driven circulation caused by tide-induced melting. About
 378 a third of the dynamical component is compensated by the thermodynamical
 379 and covariational components (Fig. 6c,d). The negative covariational compo-
 380 nent can be understood as a decrease of thermal forcing (through latent heat
 381 and injection of water at the freezing point) occurring where and when the
 382 TBL velocities are increased. The thermodynamical component accounts for
 383 a tide-induced reduction of thermal forcing in the cavities. It could a priori
 384 either come from a tide-induced cooling on the continental shelf (resulting
 385 from increased vertical mixing or residual currents), from a non-local thermal
 386 effect of increased melt rates (the local part being in the covariational term),
 387 or from modified vertical mixing within ice shelf cavities (outside the TBL).
 388 Similar tide-induced reduction in thermal forcing was reported by Gwyther
 389 et al. (2016). Our simulations indicate locally increased vertical diffusivity

390 (K_z) in ice shelf cavities (Fig. 8), but this tends to bring heat from below
 391 and therefore increase temperatures in the TBL, whereas a decrease is found.
 392 We also find that subsurface (200-1000 m) temperatures over the continen-
 393 tal shelf tend to get warmer when tides are included, by up to 0.15°C and
 394 0.45°C in front of Abbot and Dotson-Getz respectively (not shown), which
 395 likely results from a combination of both increased vertical mixing associated
 396 with topographically-trapped vorticity waves (Fig. 8) and the tidal residual
 397 circulation depicted in Fig. 5. However, this coastal warming cannot explain
 398 the decreased thermal forcing within ice shelf cavities. We conclude that the
 399 negative thermal component is explained by non-local thermal effect related
 400 to the injection and transport of cold meltwater in the TBL.

401 The prominence of the dynamical component in our decomposition sup-
 402 ports the concept of parameterizing the tides effect on melt rates through
 403 a prescribed tidal velocity, although the thermodynamical and covariational
 404 components are non negligible. Such parameterization is explored in the
 405 following sub-section.

406 3.3. Prescribing the effect of tides

407 We now attempt to represent the tidal influence on melt rates without
 408 explicitly representing tides in our simulations, but rather by prescribing a
 409 “tidal TBL velocity” $u_{\text{tide}}(x, y)$ in (3). Ideally u_{tide} would be deduced from
 410 a tidal model such as FES2012 or CATS2008, but we start with a proof
 411 of concept, assuming that the tidal influence on TBL velocities is perfectly
 412 known, and trying to reproduce the effect of tides on melt rates through the
 413 use of (3) in the melt equations of NEMO. A first challenge is to deduce
 414 $u_{\text{tide}}(x, y)$ from the simulation with explicit tides. A first possibility is to
 415 define $u_{\text{tide}}(x, y)$ as the local time-RMS difference between TBL velocities
 416 with and without tides:

$$u_{\text{tide}}(x, y) = \sqrt{\overline{u_{\text{TBL, TIDE18}}^2(x, y)} - \overline{u_{\text{TBL, REF}}^2(x, y)}} \quad (6)$$

417 where u are velocity amplitudes and overbars are 1-year averages (exact
 418 calculation of mean square TBL velocities during the simulations). Such
 419 definition leads to an overestimation of melt fluxes by 63 Gt.yr⁻¹, i.e. by
 420 68% (Fig. 9a). This strong overestimation is due to the melt-induced circu-
 421 lation that is included in our calculation of $u_{\text{tide}}(x, y)$. Indeed, by prescribing
 422 $u_{\text{tide}}(x, y)$ in (3), we increase melt rates, which creates an additional melt-
 423 induced circulation, which, in turns, further amplifies melt rates. In other

424 words, this definition of $u_{\text{tide}}(x, y)$ leads to double the feedback of the melt-
 425 induced circulation on melt rates.

426 A better way to estimate $u_{\text{tide}}(x, y)$ is to perform a harmonic analysis
 427 on TBL velocities in our realistic simulation with tides. As in the pseudo-
 428 barotropic case, we analyze 15-minute samples over a 190-day window, and
 429 extract the seven main harmonics k , i.e. the amplitude (u_k, v_k) and phase
 430 (ϕ_k, ψ_k) of zonal and meridional TBL velocities. From this, we reconstruct
 431 the RMS tidal TBL velocity as:

$$u_{\text{tide}}^2(x, y) = \overline{\left(\sum_{k=1}^7 u_k(x, y) \cos(\omega_k t + \phi_k(x, y)) \right)^2 + \left(\sum_{k=1}^7 v_k(x, y) \cos(\omega_k t + \psi_k(x, y)) \right)^2}^{190d} \quad (7)$$

432 where ω_k is the angular frequency of harmonic k . This definition of $u_{\text{tide}}(x, y)$
 433 gives weaker velocities than the previous definition (see last columns in
 434 Tab. 3), indicating that tidal currents create a residual melt-induced circula-
 435 tion that does not appear in the harmonic analysis. Such residual circulation
 436 is specific to ice shelf cavities and is distinct from the one due to the Stokes
 437 drift and non-linear dynamics (see Introduction). In terms of melt rates, the
 438 bias is reduced by half compared to the previous definition, with an overes-
 439 timation by 31 Gt.yr⁻¹, i.e. 34% (Fig. 9b). The remaining overestimation
 440 suggests that there is also a melt-induced circulation varying at tidal fre-
 441 quencies. The latter is included in our calculation of $u_{\text{tide}}(x, y)$ so that the
 442 feedback of the melt-induced circulation to melt rates is still overestimated.

443 To get rid of the overestimated melt-induced circulation at tidal frequen-
 444 cies, we multiply $u_{\text{tide}}(x, y)$ by a correction factor α , i.e. the simulated value
 445 of u_{\star} is calculated as:

$$u_{\star}(x, y) = \sqrt{C_d (u_{\text{TBL}}^2(x, y) + \alpha^2 u_{\text{tide}}^2(x, y))} \quad (8)$$

446 where u_{TBL} is the TBL velocity resolved by the ocean model. Empirically,
 447 we find that $\alpha = 0.777$ reduces the melt bias to nearly zero. Applying the
 448 decomposition defined in (5) to the case of prescribed $\alpha u_{\text{tide}}(x, y)$, we find
 449 that the dynamical/thermodynamical/covariational contributions are 148/-
 450 34/-23 Gt.yr⁻¹ respectively versus 148/-31/-26 Gt.yr⁻¹ in the simulation
 451 with explicit tides (Fig. 6). This shows that this method correctly replicates
 452 the effect of tides on melt rates. This result also supports the interpretation
 453 that tides mostly affect ice shelf melting through tidal velocities along the
 454 ice draft and subsequent turbulent exchange rather than through residual

455 circulation or tidal mixing farther from the ice shelf base. Prescribed tidal
 456 TBL velocities have no marked effect on K_z in ice-shelf cavities and along the
 457 continental shelf break (Fig. 8d), which indicates that tide-induced changes in
 458 K_z (Fig. 8c) are mostly related to tidal currents and not to the melt-induced
 459 circulation. It also confirms the negligible effect of tidal vertical mixing on
 460 melt rates in our simulations.

461 So far, the studies parameterizing the effect of tides have prescribed uni-
 462 form tidal velocities (e.g. Hattermann et al., 2014; Asay-Davis et al., 2016).
 463 We therefore evaluate this approach by prescribing uniform u_{tide} obtained
 464 through an average of $u_{\text{tide}}(x, y)$ all over the domain ($3.5 \text{ cm}\cdot\text{s}^{-1}$) or for each
 465 individual ice shelf (values in the last column of Tab. 3). The bias is approx-
 466 imately 1.5 and 2 times larger with per-ice-shelf and domain-wide uniform
 467 velocities respectively than with spatially-dependent velocities (Fig. 9c,d).
 468 The strongest biases are found in the vicinity of deep grounding lines, par-
 469 ticularly for Thwaites and Pine Island, where thermal forcing is high due to
 470 the presence of CDW, but where tides do not produce strong velocities (see
 471 subsection 3.2 and Mueller et al., 2012 and Gwyther et al., 2016). Given
 472 the sensitivity of the ice sheet dynamics to melting at the grounding line, we
 473 conclude that prescribing uniform tidal TBL velocities is best avoided.

474 As mentioned earlier, a tide parameterization would be more useful if it
 475 can be directly based on the barotropic velocities derived from tide models
 476 such as FES2012 or CATS2008. As shown in Fig. 10, $\alpha u_{\text{tide}}(x, y)$ is signif-
 477 icantly correlated with $\langle U_{\text{btp}} \rangle(x, y)$ calculated as in (7) but with amplitude
 478 and phases related to the barotropic velocities in the pseudo-barotropic ex-
 479 periment (BTP-07) rather than the TBL velocities in the realistic experiment
 480 (TIDE-18). It is therefore a good approximation to define $\alpha u_{\text{tide}}(x, y)$ as a
 481 linear function of $\langle U_{\text{btp}} \rangle(x, y)$ calculated from the output of a tide model.

482 In summary, we suggest the following method to parameterize the effect

483 of tides on ice-shelf melt in the melt equations:

$$\left\{ \begin{array}{l}
 \gamma_T(x, y) = \Gamma_T u_\star(x, y) \\
 u_\star(x, y) = \sqrt{C_d (u_{\text{TBL}}^2(x, y) + \alpha^2 u_{\text{tide}}^2(x, y))} \\
 \alpha u_{\text{tide}}(x, y) = A_0 \langle U_{\text{btp}} \rangle(x, y) + U_0 \\
 \langle U_{\text{btp}} \rangle^2(x, y) = \overline{\left(\sum_k U_k(x, y) \cos(\omega_k t + \Phi_k(x, y)) \right)^2} + \overline{\left(\sum_k V_k(x, y) \cos(\omega_k t + \Psi_k(x, y)) \right)^2}
 \end{array} \right. \quad (9)$$

484 in which u_{TBL} is the ocean model solution, (U_k, V_k) and (Φ_k, Ψ_k) are the am-
 485 plitudes and phases of zonal and meridional barotropic velocities provided by
 486 the tide model for each harmonic of angular frequency ω_k , and Γ_T , C_d , α , A_0 ,
 487 and U_0 are constant scalars with values summarized in Tab. 4. The overline
 488 represents an average over a period that is long enough to correctly sample
 489 the interaction between harmonics (e.g. 6 months for 7 harmonics). The
 490 parameters A_0 and U_0 account for the shape of the vertical velocity profile in
 491 ice shelf cavities (their values are shown in Fig. 10). A non-zero U_0 ensures
 492 a background heat transfer due to molecular diffusivity in the case of zero
 493 TBL velocity and zero tidal velocity, although we obtain U_0 from an empiri-
 494 cal fit with no knowledge of the role of molecular diffusivity. Gwyther et al.
 495 (2016) estimated that u_\star should have a minimum value of approximately
 496 $2.0 \times 10^{-5} \text{ m.s}^{-1}$ due to molecular diffusion, which would be equivalent to
 497 $U_0 = 6.3 \times 10^{-4} \text{ m.s}^{-1}$ in (9) i.e. ~ 5 times less than our value.

498 4. Discussion and Conclusion

499 In this paper, we have undertaken harmonic analyses of pseudo-barotropic
 500 simulations to show that our regional model configuration is able to pro-
 501 duce tides with similar characteristics as in the barotropic tide simulation
 502 imposed at the domain lateral boundaries. Diurnal tides induce topographi-
 503 cally trapped vorticity waves along the continental shelf break in the vicinity
 504 of the Dotson-Getz Trough. This slightly strengthens vertical mixing in that
 505 area, and is likely responsible for the residual circulation of 0.2 Sv that flows
 506 southward on the eastern flank of the Dotson-Getz Trough. While diurnal
 507 tides have a larger amplitude than semi-diurnal tides over the Amundsen

508 Sea continental shelf, semi-diurnal tides are found to resonate in the shal-
509 lowest ice-shelf cavities, so that both diurnal and semi-diurnal tides produce
510 strong velocities at the base of the ice shelves. Tides increase mean melt
511 rates in all the simulated cavities of the Amundsen Sea, with weakest effects
512 for Pine Island and Thwaites ($< +10\%$) and strongest effects for Dotson,
513 Cosgrove and Abbot ($> +30\%$). There is a large uncertainty on these esti-
514 mates, in particular because of the poorly known bathymetry underneath ice
515 shelves, and recent studies have proposed alternative bathymetry datasets in
516 the Amundsen Sea sector (Schaffer et al., 2016; Millan et al., 2017).

517 We suggest that the weak tide-induced melting underneath Pine Island
518 and Thwaites is partly related to the particularly thick water column for these
519 cavities, which makes resonant quarter wavelengths much larger than the
520 cavity size and therefore limits the presence of tidal resonance. By contrast
521 with Robertson (2013), we do not find a significant influence of the position
522 of Pine Island and Thwaites with respect to the M_2 critical latitude. The
523 importance of tidal resonances suggests that the characteristics of tides may
524 evolve in a warmer climate with thinner ice shelves and retreated grounding
525 lines. Proximity to resonance is also affected by large uncertainties in the
526 bathymetry underneath ice shelves (e.g., Schaffer et al., 2016; Millan et al.,
527 2017).

528 A dynamical/thermodynamical decomposition for all the simulated ice
529 shelves indicates that enhanced melting when tides are explicitly added is
530 mostly due to tide-induced velocities in the top boundary layer that enhance
531 heat fluxes at the ice/ocean interface. Approximately a third of this effect
532 is counterbalanced by the resulting additional release of cold melt water.
533 We also present evidence for a positive feedback whereby tide-induced cir-
534 culation produces more melt, which strengthens the buoyancy-driven ocean
535 circulation underneath ice shelves, in turn producing stronger melting. Such
536 positive feedback was previously reported by Makinson et al. (2011) in the
537 case of the Filchner-Ronne Ice Shelf. We find that tide-induced vertical
538 mixing does not significantly affect melt rates. This result has to be taken
539 cautiously because (1) our model resolution only allows the first few vertical
540 normal modes of internal tides to be resolved, and (2) the simulated vertical
541 mixing is inaccurate because the model is not able to explicitly represent
542 internal wave breaking (and the energy cascade) associated with the resolved
543 modes. These are, however, general caveats of large-scale ocean models, and
544 the first few baroclinic modes still contain most of the energy of internal
545 tides.

546 As tides mostly affect simulated melt rates through increased velocities
547 in the top boundary layer (TBL) of ice-shelf cavities, it is possible to pre-
548 scribe their effect by including a “tidal TBL velocity” in the calculation of
549 the friction/exchange velocity along the ice draft. In the Amundsen Sea, pre-
550 scribing a spatially-uniform tidal TBL velocity leads to overestimated melt
551 rates near deep grounding lines where the thermal forcing is high due to the
552 presence of modified CDW, but where simulated tidal currents are weak. In
553 the absence of spatially-distributed observations of tidal currents along the
554 ice draft, we therefore recommend to derive spatially-varying tidal velocities
555 from the outputs of tide models following Eq. (9). In our simulations, TBL
556 velocities can be prescribed as 66% (A_0 in Tab. 4) of the barotropic velocities
557 from a tide model to account for the vertical profile due to the interaction
558 with the ice draft and for the melt-induced circulation. Furthermore, if it
559 was possible to observe spatially distributed currents underneath an ice shelf
560 and to perform a harmonic analysis, we would still have to apply a correction
561 factor of ~ 0.777 before prescribing the observed mean-square TBL velocity
562 into the TBL equations of an ocean model. Indeed, tide-induced melting en-
563 hances observed TBL velocities, so that directly applying observed velocities
564 would effectively apply these velocities twice, due to the feedback related to
565 the buoyancy-driven ocean circulation.

566 The proposed methodology to represent the effects of tides on melt rates
567 is not a standalone parameterization in the sense that it relies on a tide
568 model. For example, our methodology would remain valid in a warmer ocean,
569 with a different coastal circulation, but would not allow adapting the tide
570 characteristics (e.g. resonance) for evolving cavity shapes (see Mueller et al.,
571 2018) or when new bathymetry datasets become available. In this respect,
572 the proposed methodology can be considered as bulk formula for the interface
573 with barotropic tide models, in a similar way as the bulk formula used to
574 calculate air-sea fluxes from atmospheric models or reanalyses. Relying on
575 barotropic models specially developed for tides allows more accuracy in the
576 tide solution (e.g. through data assimilation) and a higher level of complexity
577 in the representation of tidal processes (e.g. self attraction and loading)
578 compared to what is usually implemented in OGCMs.

579 We suggest that our methodology to prescribe tidal TBL velocities could
580 be applied to other sectors of Antarctica, but with care with regards to two
581 aspects. First, although vertical mixing and residual circulation near the con-
582 tinental shelf break have no major effects on ice-shelf melting in the Amund-
583 sen Sea, this may not be true elsewhere in Antarctica. For example, water

584 masses in cold cavities such as Ross and Filchner-Ronne could hypotheti-
585 cally be more sensitive to the tide-induced exchanges across the continental
586 shelf and ice-shelf front. Second, our parameterization includes a factor α to
587 account for the buoyancy-driven TBL velocity associated with tide-induced
588 melting. According to the theoretical considerations in Appendix C of Jour-
589 dain et al. (2017), this factor is likely non-uniform around Antarctica, and
590 should depend on the background ocean density and on the depth at which
591 tides enhance melt rates. We nonetheless suggest to use $\alpha < 1$ everywhere, as
592 buoyancy-driven circulation resulting from tide-induced melting is expected
593 to occur in any ice shelf cavity. For example, such buoyancy-driven circula-
594 tion was simulated under Filchner-Ronne by Makinson et al. (2011).

595 Our concluding remark is for melt rate parameterizations used in ice sheet
596 models (see review in Asay-Davis et al., 2017). The methodology proposed
597 in this paper to prescribe tides could be applied to these parameterizations in
598 order to better distinguish melt-induced and tide-induced circulation. The
599 background melt-induced circulation depends on the thermal forcing and
600 makes melt rates proportional to the square thermal forcing (Holland et al.,
601 2008). In contrast, tide-induced melt rates are expected to depend more
602 linearly on the thermal forcing, although a weak non-linearity may also be
603 expected due to the additional circulation induced by melting.

604 **Data availability**

605 The model version, customizations, and parameters used to run the eight
606 main experiments presented in this paper are provided on [http://doi.org/
607 10.5281/zenodo.1067647](http://doi.org/10.5281/zenodo.1067647). The output and forcing files are available on
608 request.

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619 **References**

- 620 Arzeno, I. B., Beardsley, R. C., Limeburner, R., Owens, B., Padman, L.,
621 Springer, S. R., Stewart, C. L., Williams, M. J. M., 2014. Ocean variability
622 contributing to basal melt rate near the ice front of Ross Ice Shelf,
623 Antarctica. *J. Geophys. Res. Oceans* 119 (7), 4214–4233.
- 624 Asay-Davis, X. S., Cornford, S. L., Durand, G., Galton-Fenzi, B. K., Glad-
625 stone, R. M., Gudmundsson, G. H., Hattermann, T., Holland, D. M.,
626 Holland, D., Holland, P. R., et al., 2016. Experimental design for three
627 interrelated Marine Ice-Sheet and Ocean Model Intercomparison Projects.
628 *Geoscientific Model Development* 9, 2471–2497.
- 629 Asay-Davis, X. S., Jourdain, N. C., Nakayama, Y., 2017. Developments in
630 Simulating and Parameterizing Interactions Between the Southern Ocean
631 and the Antarctic Ice Sheet. *Current Climate Change Reports* 3 (4), 316–
632 329.
- 633 Bessières, L., Madec, G., Lyard, F., 2008. Global tidal residual mean circu-
634 lation: Does it affect a climate OGCM? *Geophys. Res. Lett.* 35 (3).
- 635 Carrère, L., Lyard, F., Cancet, M., Guillot, A., Roblou, L., 2012. FES2012: A
636 new global tidal model taking advantage of nearly 20 years of altimetry. In:
637 Proceedings of the "20 Years of Progress in Radar Altimetry" Symposium,
638 Venice. pp. 1–20.
- 639 Cartwright, D. E., 1977. Oceanic tides. *Reports on Progress in Physics* 40 (6),
640 665.
- 641 Dansereau, V., Heimbach, P., Losch, M., 2014. Simulation of subice shelf
642 melt rates in a general circulation model: Velocity-dependent transfer and
643 the role of friction. *J. Geophys. Res.* 119 (3), 1765–1790.
- 644 Dinniman, M. S., Asay-Davis, X. S., Galton-Fenzi, B. K., Holland, P. R.,
645 Jenkins, A., Timmermann, R., 2016. Modeling ice shelf/ocean interaction
646 in Antarctica: A review. *Oceanography* 29 (4), 144–153.
- 647 Donat-Magnin, M., Jourdain, N. C., Spence, P., Le Sommer, J., Gallée, H.,
648 Durand, G., 2017. Ice-Shelf Melt Response to Changing Winds and Glacier
649 Dynamics in the Amundsen Sea Sector, Antarctica. *J. Geophys. Res.*

- 650 Durand, G., Gagliardini, O., De Fleurian, B., Zwinger, T., Le Meur, E.,
651 2009. Marine ice sheet dynamics: Hysteresis and neutral equilibrium. *J.*
652 *Geophys. Res.* 114 (F3).
- 653 Dutrieux, P., De Rydt, J., Jenkins, A., Holland, P. R., Ha, H. K., Lee, S. H.,
654 Steig, E. J., Ding, Q., Abrahamsen, E. P., Schröder, M., 2014. Strong
655 sensitivity of Pine Island ice-shelf melting to climatic variability. *Science*
656 343 (6167), 174–178.
- 657 Favier, L., Durand, G., Cornford, S. L., Gudmundsson, G. H., Gagliardini,
658 O., Gillet-Chaulet, F., Zwinger, T., Payne, A. J., Le Brocq, A. M., 2014.
659 Retreat of Pine Island Glacier controlled by marine ice-sheet instability.
660 *Nature Climate Change* 4 (2), 117–121.
- 661 Furevik, T., Foldvik, A., 1996. Stability at M2 critical latitude in the Barents
662 Sea. *J. Geophys. Res.* 101 (C4), 8823–8837.
- 663 Galton-Fenzi, B. K., Hunter, J. R., Coleman, R., Marsland, S. J., Warner,
664 R. C., 2012. Modeling the basal melting and marine ice accretion of the
665 Amery Ice Shelf. *J. Geophys. Res.* 117 (C9).
- 666 Griffies, S. M., Biastoch, A., Böning, C., Bryan, F., Danabasoglu, G., Chas-
667 signet, E. P., England, M. H., Gerdes, R., Haak, H., Hallberg, R. W., et al.,
668 2009. Coordinated ocean-ice reference experiments (COREs). *Ocean Mod-*
669 *ell.* 26 (1), 1–46.
- 670 Gwyther, D. E., Cougnon, E. A., Galton-Fenzi, B. K., Roberts, J. L., Hunter,
671 J. R., Dinniman, M. S., 2016. Modelling the response of ice shelf basal melt-
672 ing to different ocean cavity environmental regimes. *Annals of Glaciology*,
673 1–11.
- 674 Hattermann, T., Smedsrud, L. H., Nøst, O. A., Lilly, J. M., Galton-Fenzi,
675 B. K., 2014. Eddy-resolving simulations of the Fimbul Ice Shelf cavity
676 circulation: Basal melting and exchange with open ocean. *Ocean Modelling*
677 82, 28–44.
- 678 Hemer, M. A., Hunter, J. R., Coleman, R., 2006. Barotropic tides beneath
679 the Amery Ice Shelf. *J. Geophys. Res.* 111 (C11).

- 680 Heywood, K. J., Locarnini, R. A., Frew, R. D., Dennis, P. F., King, B. A.,
681 1998. Transport and water masses of the Antarctic Slope Front system in
682 the eastern Weddell Sea. *Antarctic Research Series* 75, 203–214.
- 683 Holland, P. R., Jenkins, A., Holland, D. M., 2008. The response of ice shelf
684 basal melting to variations in ocean temperature. *J. Climate* 21 (11), 2558–
685 2572.
- 686 Holloway, G., Proshutinsky, A., 2007. Role of tides in Arctic ocean/ice cli-
687 mate. *J. Geophys. Res.* 112 (C4).
- 688 Jacobs, S., Jenkins, A., Hellmer, H., Giulivi, C., Nitsche, F., Huber, B., Guer-
689 rero, R., 2012. The Amundsen Sea and the Antarctic ice sheet. *Oceanog-
690 raphy* 25 (3), 154–163.
- 691 Jenkins, A., 1999. The impact of melting ice on ocean waters. *J. Phys.
692 Oceanogr.* 29 (9), 2370–2381.
- 693 Jenkins, A., Nicholls, K. W., Corr, H. F. J., 2010. Observation and parame-
694 terization of ablation at the base of Ronne Ice Shelf, Antarctica. *J. Phys.
695 Oceanogr.* 40 (10), 2298–2312.
- 696 Joughin, I., Smith, B. E., Medley, B., 2014. Marine ice sheet collapse poten-
697 tially under way for the Thwaites Glacier Basin, West Antarctica. *Science*
698 344 (6185), 735–738.
- 699 Jourdain, N. C., Mathiot, P., Merino, N., Durand, G., Le Sommer, J., Spence,
700 P., Dutrieux, P., Madec, G., 2017. Ocean circulation and sea-ice thinning
701 induced by melting ice shelves in the Amundsen Sea. *J. Geophys. Res.
702 Oceans* 122 (3), 2550–2573.
- 703 Large, W. G., Yeager, S. G., 2009. The global climatology of an interannually
704 varying air–sea flux data set. *Clim. Dynam.* 33 (2-3), 341–364.
- 705 Leclair, M., Madec, G., 2011. z-Coordinate, an Arbitrary Lagrangian–
706 Eulerian coordinate separating high and low frequency motions. *Ocean
707 Modelling* 37 (3-4), 139–152.
- 708 Lee, H.-C., Rosati, A., Spelman, M. J., 2006. Barotropic tidal mixing effects
709 in a coupled climate model: Oceanic conditions in the northern Atlantic.
710 *Ocean Model.* 11 (3), 464–477.

- 711 Longuet-Higgins, M. S., 1969. On the transport of mass by time-varying
712 ocean currents 16 (5), 431–447.
- 713 Losch, M., 2008. Modeling ice shelf cavities in az coordinate ocean general
714 circulation model. *J. Geophys. Res.* 113 (C8).
- 715 Lyard, F., Lefevre, F., Letellier, T., Francis, O., 2006. Modelling the global
716 ocean tides: modern insights from FES2004. *Ocean Dynamics* 56 (5-6),
717 394–415.
- 718 MacAyeal, D. R., 1984a. Numerical simulations of the Ross Sea tides. *J.*
719 *Geophys. Res.* 89 (C1), 607–615.
- 720 MacAyeal, D. R., 1984b. Thermohaline circulation below the Ross Ice Shelf:
721 A consequence of tidally induced vertical mixing and basal melting. *J.*
722 *Geophys. Res.* 89 (C1), 597–606.
- 723 Mack, S. L., Dinniman, M. S., McGillicuddy Jr, D. J., Sedwick, P. N., Klinck,
724 J. M., 2017. Dissolved iron transport pathways in the Ross Sea: Influence
725 of tides and horizontal resolution in a regional ocean model. *Journal of*
726 *Marine Systems* 166, 73–86.
- 727 Madec, G., NEMO-team, 2016. NEMO ocean engine, version 3.6 stable, Note
728 du Pôle de modélisation de l’Institut Pierre-Simon Laplace No 27, ISSN
729 No 1288-1619. Tech. rep., IPSL, France.
- 730 Makinson, K., Holland, P. R., Jenkins, A., Nicholls, K. W., Holland, D. M.,
731 2011. Influence of tides on melting and freezing beneath Filchner-Ronne
732 Ice Shelf, Antarctica. *Geophys. Res. Lett.* 38 (6).
- 733 Makinson, K., Nicholls, K. W., 1999. Modeling tidal currents beneath
734 Filchner-Ronne Ice Shelf and on the adjacent continental shelf: Their effect
735 on mixing and transport. *J. Geophys. Res.* 104 (C6), 13449–13465.
- 736 Maraldi, C., Chanut, J., Levier, B., Ayoub, N., De Mey, P., Reffray, G.,
737 Lyard, F., Cailleau, S., Drévilion, M., Fanjul, E., et al., 2013. NEMO
738 on the shelf: assessment of the Iberia-Biscay-Ireland configuration. *Ocean*
739 *Science* 9, 745–771.
- 740 Mathiot, P., Goosse, H., Fichefet, T., Barnier, B., Gallée, H., 2011. Modelling
741 the seasonal variability of the Antarctic Slope Current. *Ocean Science* 7 (4),
742 445–532.

- 743 Mathiot, P., Jenkins, A., Harris, C., Madec, G., 2017. Explicit and
744 parametrised representation of under ice shelf seas in az* coordinate ocean
745 model NEMO 3.6. *Geoscientific Model Development* 10 (7), 2849–2874.
- 746 McMillan, M., Shepherd, A., Nienow, P., Leeson, A., 2011. Tide model ac-
747 curacy in the Amundsen Sea, Antarctica, from radar interferometry obser-
748 vations of ice shelf motion. *J. Geophys. Res.* 116 (C11).
- 749 McPhee, M. G., Morison, J. H., Nilsen, F., 2008. Revisiting heat and salt ex-
750 change at the ice-ocean interface: Ocean flux and modeling considerations.
751 *J. Geophys. Res.* 113 (C6).
- 752 Middleton, J. H., Foster, T. D., Foldvik, A., 1987. Diurnal shelf waves in the
753 southern Weddell Sea. *J. Phys. Oceanogr.* 17 (6), 784–791.
- 754 Millan, R., Rignot, E., Bernier, V., Morlighem, M., Dutrieux, P., 2017.
755 Bathymetry of the Amundsen Sea Embayment sector of West Antarctica
756 from Operation IceBridge gravity and other data. *Geophys. Res. Lett.*
757 44 (3), 1360–1368.
- 758 Mouginit, J., Rignot, E., Scheuchl, B., 2014. Sustained increase in ice dis-
759 charge from the Amundsen Sea Embayment, West Antarctica, from 1973
760 to 2013. *Geophys. Res. Lett.* 41 (5), 1576–1584.
- 761 Mueller, R. D., Hattermann, T., Howard, S. L., Padman, L., 2018. Tidal
762 influences on a future evolution of the Filchner–Ronne Ice Shelf cavity in
763 the Weddell Sea, Antarctica. *The Cryosphere* 12 (2), 453.
- 764 Mueller, R. D., Padman, L., Dinniman, M. S., Erofeeva, S. Y., Fricker, H. A.,
765 King, M. A., 2012. Impact of tide-topography interactions on basal melting
766 of Larsen C Ice Shelf, Antarctica. *J. Geophys. Res.* 117 (C5).
- 767 Müller, M., 2013. On the space-and time-dependence of barotropic-to-
768 baroclinic tidal energy conversion. *Ocean Model.* 72, 242–252.
- 769 Olbers, D., Eden, C., 2013. A global model for the diapycnal diffusivity
770 induced by internal gravity waves. *J. Phys. Oceanogr.* 43 (8), 1759–1779.
- 771 Padman, L., Erofeeva, S., Joughin, I., 2003. Tides of the Ross Sea and Ross
772 Ice Shelf cavity. *Antarctic Science* 15 (01), 31–40.

- 773 Padman, L., Fricker, H. A., Coleman, R., Howard, S., Erofeeva, L., 2002. A
774 new tide model for the Antarctic ice shelves and seas. *Annals of Glaciology*
775 34 (1), 247–254.
- 776 Padman, L., Howard, S. L., Orsi, A. H., Muench, R. D., 2009. Tides of the
777 northwestern Ross Sea and their impact on dense outflows of Antarctic
778 Bottom Water. *Deep Sea Research Part II: Topical Studies in Oceanogra-*
779 *phy* 56 (13), 818–834.
- 780 Padman, L., Kottmeier, C., 2000. High-frequency ice motion and divergence
781 in the Weddell Sea. *J. Geophys. Res.* 105 (C2), 3379–3400.
- 782 Padman, L., Siegfried, M. R., Fricker, H. A., 2018. Ocean Tide Influences on
783 the Antarctic and Greenland Ice Sheets. *Reviews of Geophysics* 56, 1–43.
- 784 Richet, O., Muller, C., Chomaz, J.-M., 2017. Impact of a Mean Current on
785 the Internal Tide Energy Dissipation at the Critical Latitude. *J. Phys.*
786 *Oceanogr.* 47 (6), 1457–1472.
- 787 Robertson, R., 2013. Tidally induced increases in melting of Amundsen Sea
788 ice shelves. *J. Geophys. Res.* 118 (6), 3138–3145.
- 789 Rousset, C., Vancoppenolle, M., Madec, G., Fichefet, T., Flavoni, S.,
790 Barthélemy, A., Benshila, R., Chanut, J., Lévy, C., Masson, S., et al.,
791 2015. The Louvain-La-Neuve sea ice model LIM3.6: global and regional
792 capabilities. *Geoscientific Model Development* 8 (10), 2991–3005.
- 793 Savage, A. C., Arbic, B. K., Alford, M. H., Ansong, J. K., Farrar, J. T.,
794 Menemenlis, D., O’Rourke, A. K., Richman, J. G., Shriver, J. F., Voet,
795 G., et al., 2017. Spectral decomposition of internal gravity wave sea surface
796 height in global models. *J. Geophys. Res.* 122 (10), 7803–7821.
- 797 Schaffer, J., Timmermann, R., Arndt, J. E., Kristensen, S. S., Mayer, C.,
798 Morlighem, M., Steinhage, D., 2016. A global high-resolution data set
799 of ice sheet topography, cavity geometry and ocean bathymetry. *Earth*
800 *System Science Data*.
- 801 Scheduikat, M., Olbers, D. J., 1990. A one-dimensional mixed layer model
802 beneath the Ross Ice Shelf with tidally induced vertical mixing. *Antarctic*
803 *Science* 2 (01), 29–42.

- 804 Schoof, C., 2007. Ice sheet grounding line dynamics: Steady states, stability,
805 and hysteresis. *J. Geophys. Res.* 112 (F3).
- 806 Schureman, P., 1958. Manual of harmonic analysis and prediction of tides.
807 No. 98. US Government Printing Office.
- 808 Simmons, H. L., Jayne, S. R., St-Laurent, L. C., Weaver, A. J., 2004. Tidally
809 driven mixing in a numerical model of the ocean general circulation. *Ocean*
810 *Modelling* 6 (3), 245–263.
- 811 Spence, P., Griffies, S. M., England, M. H., Hogg, A. M., Saenko, O. A.,
812 Jourdain, N. C., 2014. Rapid subsurface warming and circulation changes
813 of Antarctic coastal waters by poleward shifting winds. *Geophys. Res. Lett.*
814 41 (13), 4601–4610.
- 815 Stewart, A. L., Klocker, A., Menemenlis, D., 2018. Circum-Antarctic Shore-
816 ward Heat Transport Derived From an Eddy-and Tide-Resolving Simula-
817 tion. *Geophys. Res. Lett.*
- 818 Timmermann, R., Beckmann, A., Hellmer, H. H., 2002. Simulations of ice-
819 ocean dynamics in the Weddell Sea 1. Model configuration and validation.
820 *J. Geophys. Res.* 107 (C3).
- 821 Turner, J., Orr, A., Gudmundsson, G. H., Jenkins, A., Bingham, R. G.,
822 Hillenbrand, C.-D., Bracegirdle, T. J., 2017. Atmosphere-ocean-ice inter-
823 actions in the Amundsen Sea Embayment, West Antarctica. *Reviews of*
824 *Geophysics* 55 (1), 235–276.
- 825 Wang, Q., Danilov, S., Hellmer, H., Sidorenko, D., Schroeter, J., Jung, T.,
826 2013. Enhanced cross-shelf exchange by tides in the western Ross Sea.
827 *Geophys. Res. Lett.* 40 (21), 5735–5739.
- 828 Weertman, J., 1974. Stability of the junction of an ice sheet and an ice shelf.
829 *J. Glaciol.* 13, 3–11.
- 830 Zimmerman, J. T. F., 1979. On the Euler-Lagrange transformation and the
831 Stokes’ drift in the presence of oscillatory and residual currents. *Deep Sea*
832 *Research Part A. Oceanographic Research Papers* 26 (5), 505–520.
- 833

Simulation	Sea Ice	Atm. For.	Ice shelf melt	Lateral boundaries
BTP-07	-	-	zero	T_0, S_0 , tides (7 har.)
REF	LIM	CORE	3-equation	MOM025, no tides
TIDE-M2	LIM	CORE	3-equation	MOM025, tides (1 har.)
TIDE-04	LIM	CORE	3-equation	MOM025, tides (4 har.)
TIDE-18	LIM	CORE	3-equation	MOM025, tides (18 har.)
Utide-UNIF	LIM	CORE	3-equ. w. uniform u_{tide}^2 (1 value)	MOM025, no tides
Utide-PERISF	LIM	CORE	3-equ. w. uniform u_{tide}^2 (per ice shelf)	MOM025, no tides
Utide(x,y)	LIM	CORE	3-equ. w. $u_{\text{tide}}^2(x,y)$	MOM025, no tides
0.777Utide(x,y)	LIM	CORE	3-equ. w. $(0.777 \times u_{\text{tide}}(x,y))^2$	MOM025, no tides

Table 1: List of the nine 7-year AMU12.L75 simulations used in this paper. “3-equation” means that the square friction velocity involved in the three melt equations is directly proportional to the square TBL velocity, with no additional background RMS velocity unless specified. The “MOM025” lateral boundaries refer to 5-day temperature, salinity, velocity, sea-ice concentration and thickness prescribed from the 0.25° MOM simulation produced by Spence et al. (2014).

Ice Shelf	tidal residual circulation (mSv)	circulation w. 18 tidal harm. (mSv)	circulation w. no tide (mSv)
Getz	24	320	278
Dotson	5	247	241
Crosson	3	58	65
Thwaites	5	313	317
Pine Island	2	299	290
Cosgrove	1	48	41
Abbot	20	251	179

Table 2: Barotropic transport within each cavity, calculated as the maximum amplitude (maximum minus minimum) of the barotropic stream function under the ice-shelf. These numbers are calculated from averages over the 3rd year of the BTP-07 simulation (2nd column) and over the 7th year of the TIDE-18 (3rd column) and REF (4th column) simulations.

Ice Shelf	Melt with no tides	Melt with M_2	Melt with M_2, S_2, K_1, O_1	Melt with 18 harmonics	u_{tide} (RMS)	u_{tide} (harm.)
Getz	277.6	282.5 (+1.8%)	305.5 (+10%)	309.0 (+11%)	4.9	4.3
Dotson	25.0	26.4 (+5.6%)	31.7 (+27%)	32.6 (+30%)	6.1	3.1
Crosson	8.0	8.4 (+5.0%)	9.8 (+22%)	9.8 (+22%)	6.5	5.7
Thwaites	70.4	71.5 (+1.6%)	74.9 (+6.4%)	75.9 (+7.8%)	2.3	0.9
Pine Island	124.3	124.5 (+0.2%)	124.8 (+0.4%)	125.9 (+1.3%)	1.8	1.3
Cosgrove	20.2	21.2 (+4.9%)	25.4 (+26%)	27.0 (+34%)	4.9	3.3
Abbot	94.7	100.8 (+6.4%)	128.8 (+36%)	131.4 (+39%)	5.6	3.6

Table 3: Mean melt rates in the simulated ice shelf cavities (Gt.yr^{-1}). Increase relative to the first column is shown in brackets. The two last columns show the tidal TBL velocity in each individual cavity (in cm.s^{-1}) from the simulation with 18 tidal constituents, either deduced from a time-RMS difference between the simulation with tides and the one with no tides (RMS diff.) or from a harmonic analysis of the TBL velocity in the simulation with tides (Eq. 7).

Γ_T	2.21×10^{-3}
C_d	1.00×10^{-3}
α	0.777
A_0	0.656
U_0	0.003 m.s^{-1}

Table 4: Values of the coefficients used in Eq. (9).

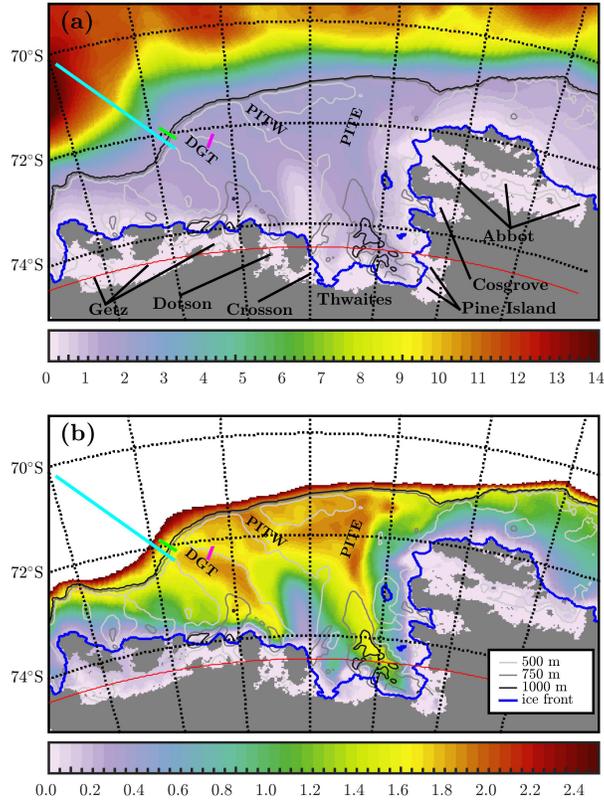


Figure 1: (a) Mean barotropic stream function (Sv) in the reference simulation (REF). Maxima indicate clockwise rotation and minima indicate anti-clockwise rotation. (b) same as (a) but with finer color range. Ice shelf names located within the domain are indicated, as well as DGT (Dotson-Getz Trough), PITW (Pine Island Trough - West) and PITE (Pine Island Trough - East). Are also shown: the M_2 critical latitude (red), the bathymetry on the continental shelf (grey contours), and the land ice terminus (blue). The grounded ice is in grey. The barotropic transports across the cyan, green and magenta sections are 11.7 Sv, 0.49 Sv and 0.13 Sv respectively.

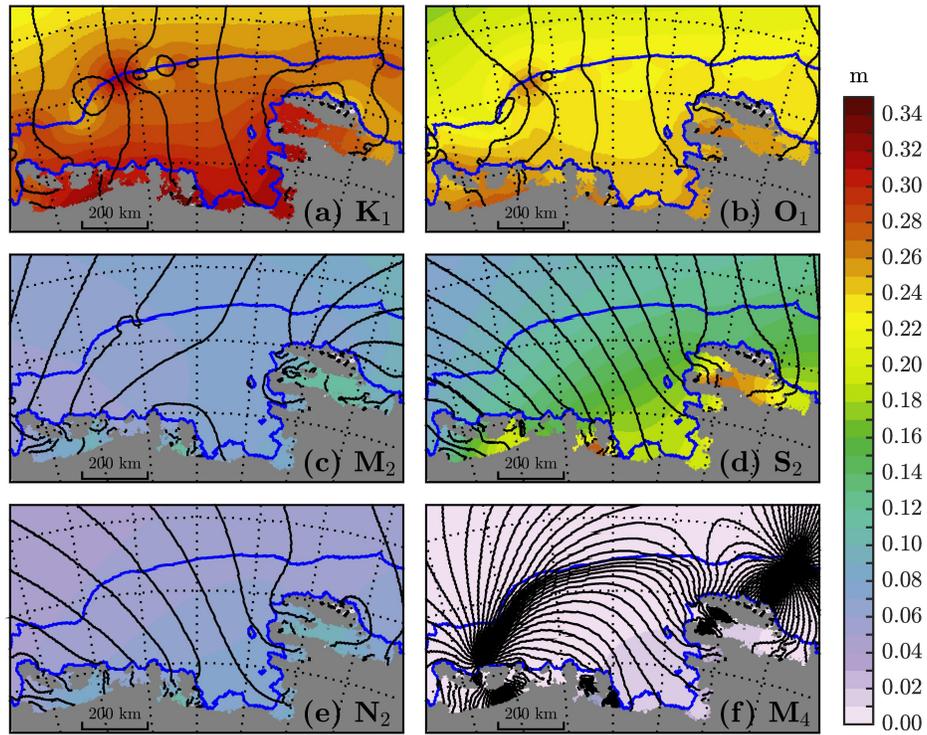


Figure 2: Amplitude (shaded) and phase (thin black, contour every 5°) of six SSH harmonics in the pseudo-barotropic experiment (harmonic analysis over the six first months of the third simulation year). Grounded ice is in gray, and the thick blue contours represent the ice-sheet margin and the 1500 m isobath (indicating the location of the continental shelf break). Dotted lines are latitude (every 2°) and longitude (every 5°).

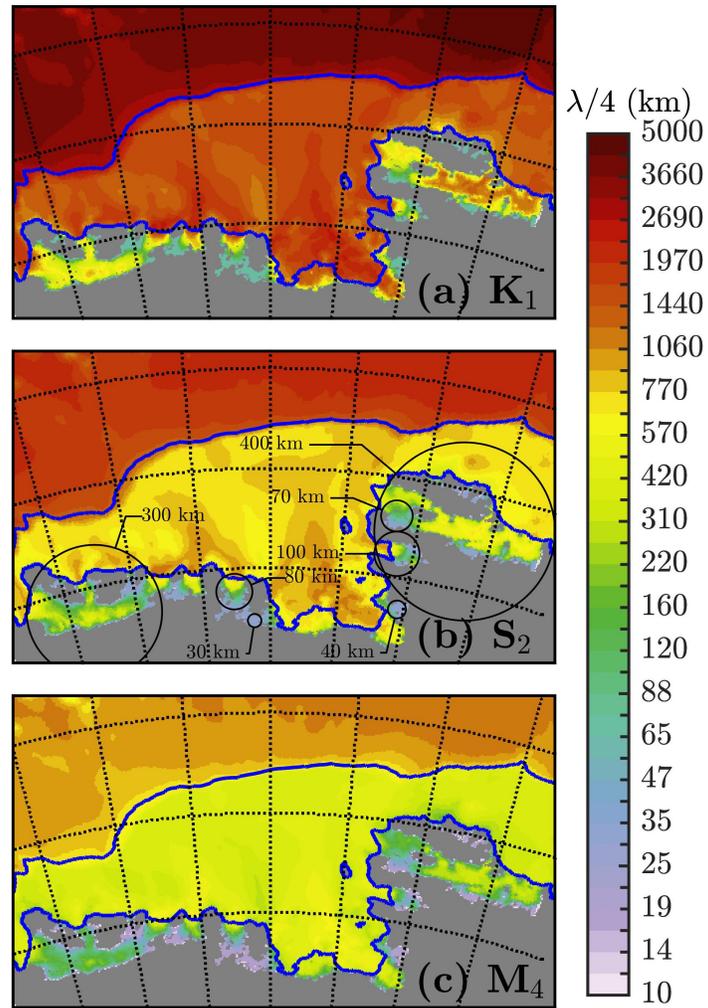


Figure 3: Quarter wavelength estimated under the shallow water approximation for (a) K_1 , (b) S_2 and (c) M_4 . The circles indicate typical ice-shelf and bay sizes on the map. The blue lines show the land ice terminus and the continental shelf break (1500 m isoline).

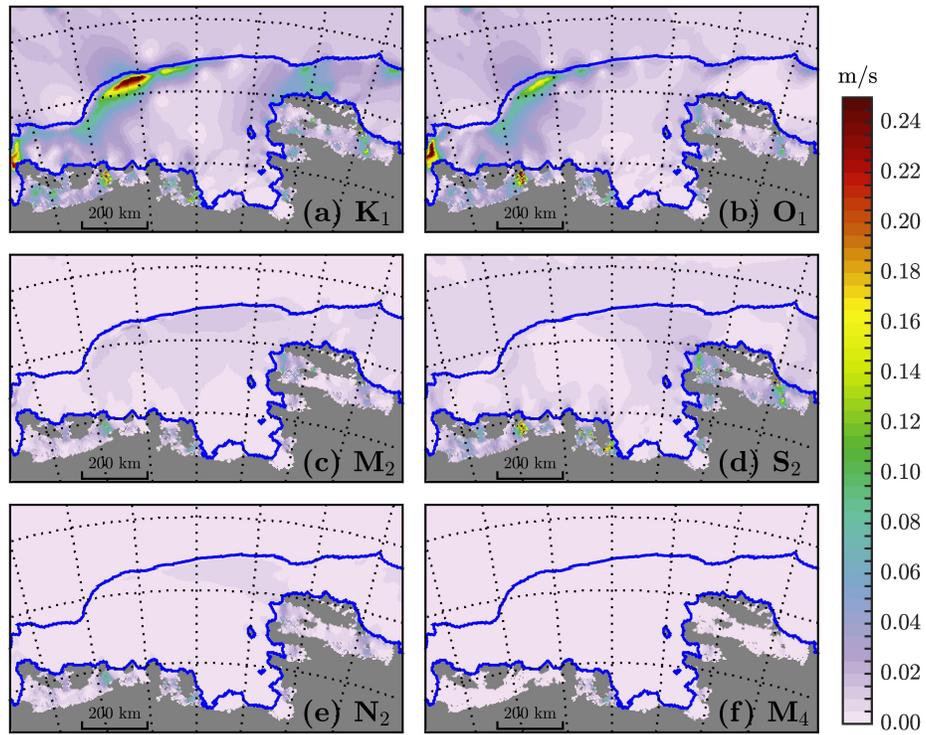


Figure 4: Major semi-axis of the tidal ellipse related to six barotropic-velocity harmonics (shaded) in the pseudo-barotropic experiment (harmonic analysis over the six first months of the third simulation year). Grounded ice is in gray, and the thick blue contours represent the ice-sheet margin and the 1500 m isobath (indicating the location of the continental shelf break). Dotted lines are latitude (every 2°) and longitude (every 5°).

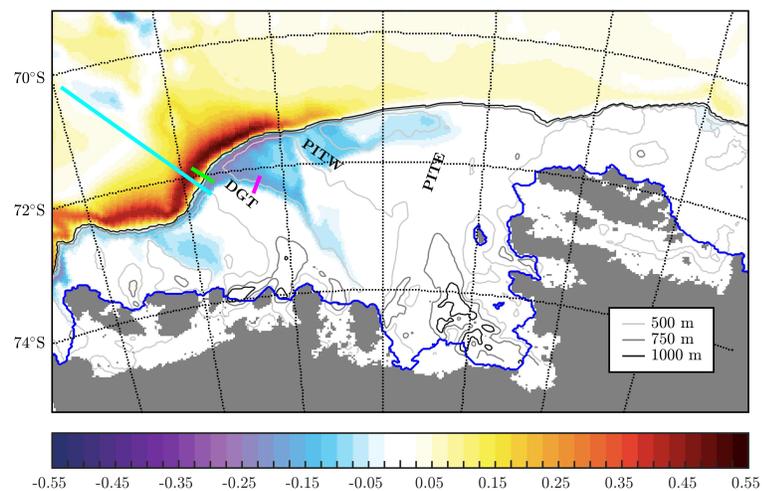


Figure 5: Residual tidal circulation, calculated as the annual mean barotropic stream function (Sv) in the BTP-07 simulation. Maxima indicate clockwise rotation and minima indicate anti-clockwise rotation. The grey contours show the bathymetry on the continental shelf, and the blue contour indicates the land ice temrinus. The grounded ice is in grey. The barotropic transports across the cyan, green and magenta sections are 0.12 Sv, 0.70 Sv and 0.17 Sv respectively.

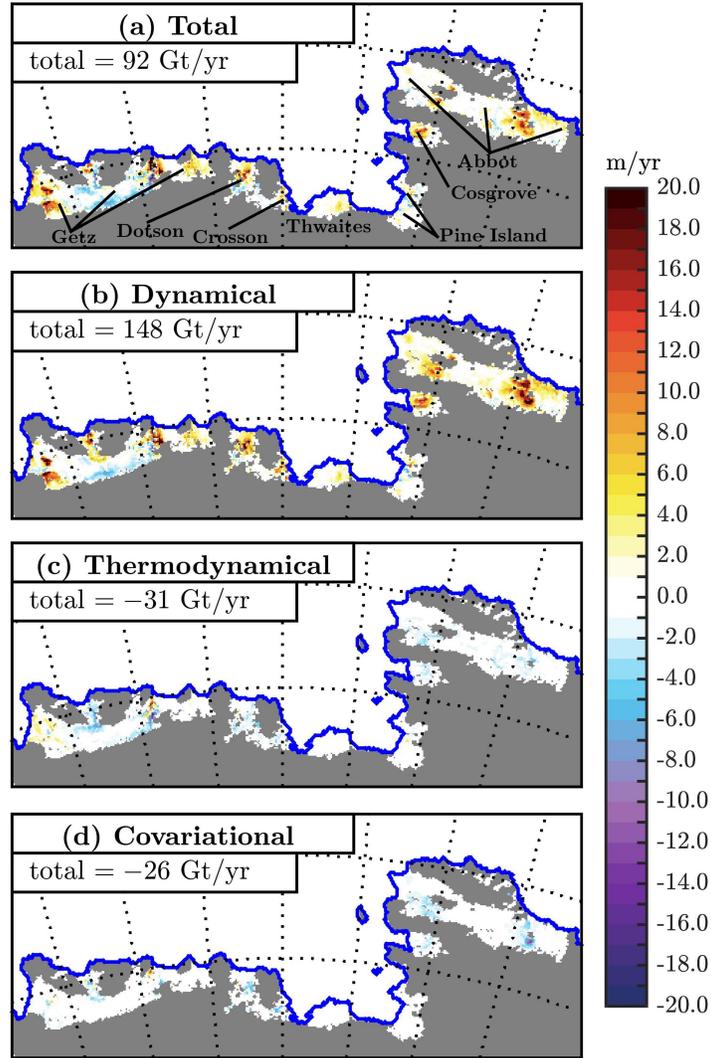


Figure 6: (a) Change in melt rate explained by the presence of 18 tidal harmonics, i.e. Δm in Eq. (5). (b-d) Terms of the decomposition of Eq. (5). The numbers in the upper left corners give the total melt anomaly over the domain.

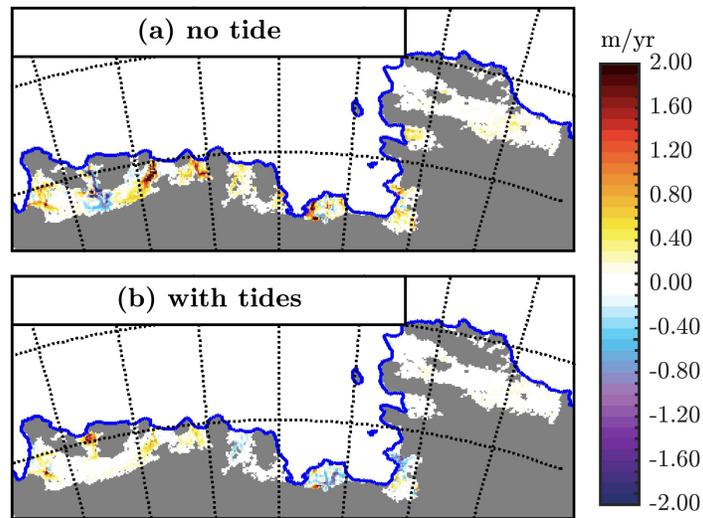


Figure 7: Changes in annual mean melt rates due to a 1°S shift of latitudes (i.e. bathymetry and ice drafts are shifted by 1°N) in simulations with (a) no tides, (b) 18 tidal harmonics. The blue contour indicates the land ice terminus and the grounded ice is in grey. Note that the amplitude of the color bar is ten times smaller than in Fig. 6.

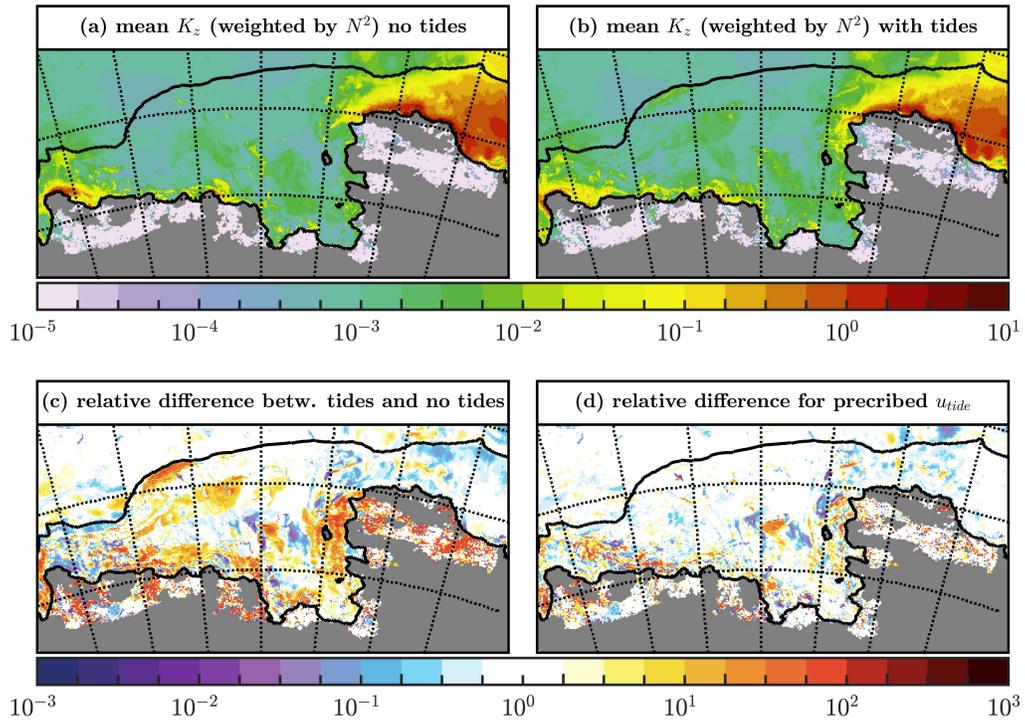


Figure 8: Stratification-weighted depth-mean vertical diffusivity (K_z , in $\text{m}^2 \cdot \text{s}^{-1}$) in: (a) REF and (b) TIDE-18. (c) relative difference (i.e. difference divided by half sum) between (a) and (b). (d) Same as (c) but with prescribed tidal TBL velocity ($0.777U_{\text{tide}}(x,y)$) instead of explicit tides. The land ice terminus and the 1500 m isobath are in black, the grounded ice sheet is in gray.

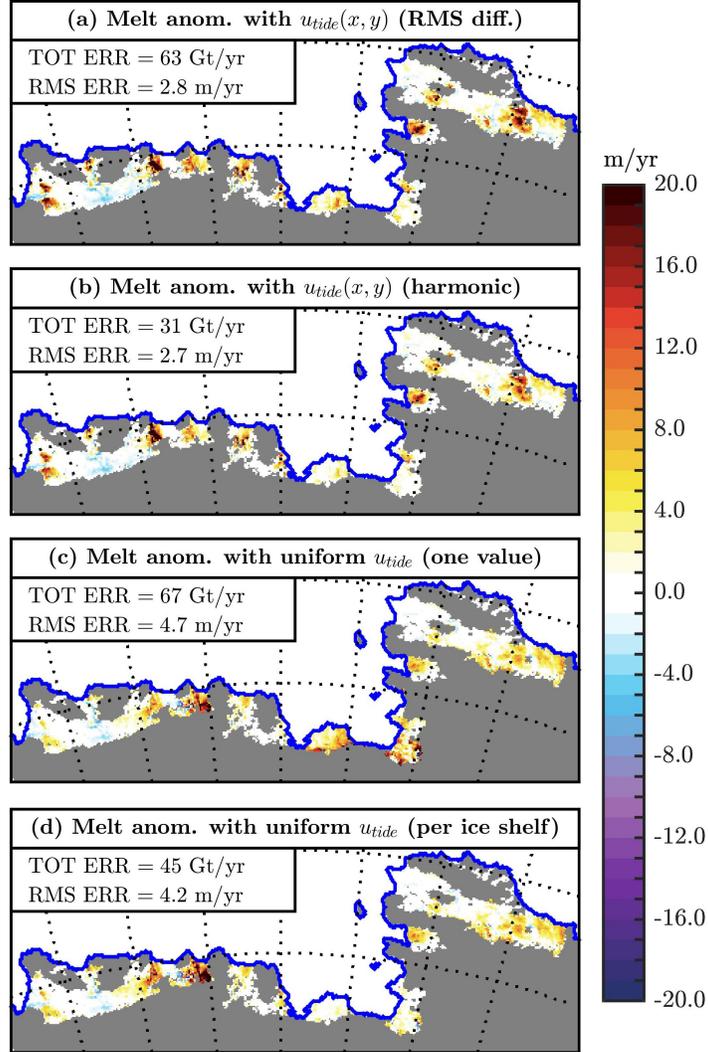


Figure 9: Change in melt rate due to the use of prescribed tidal TBL velocity in the melt equations, with (a) spatially-varying tidal velocity from the RMS difference between TIDE-18 and REF, (b) spatially-varying tidal velocity from a harmonic analysis in TIDE-18, (c) uniform tidal velocity all over the domain, calculated from the RMS of (b) under all the ice shelves, and (d) uniform tidal velocity calculated from the RMS of (b) under each individual ice shelf. The total error and RMS error with respect to the simulation with explicit tides (TIDE-18) are indicated for each panel.

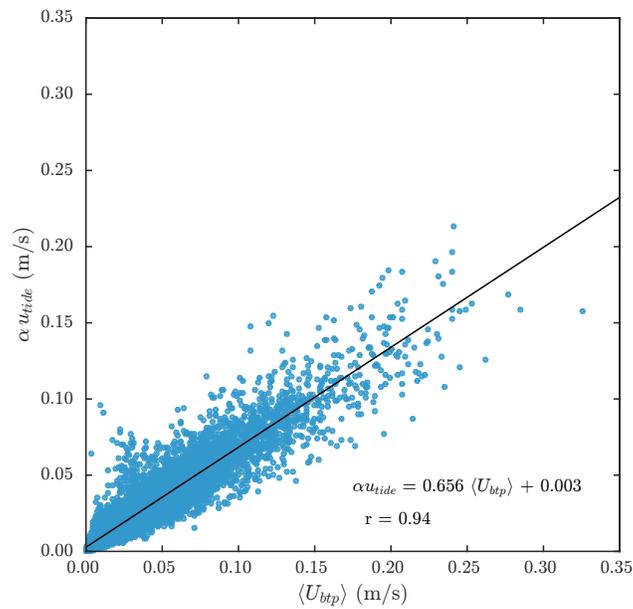


Figure 10: Harmonic tidal TBL velocity (from TIDE-18) versus the root-sum-square barotropic velocity from the harmonic analysis of the BTP-07 simulation as defined in eq. (9). Also indicated are the correlation coefficient (r) and regression coefficients of the least-squares linear fit.

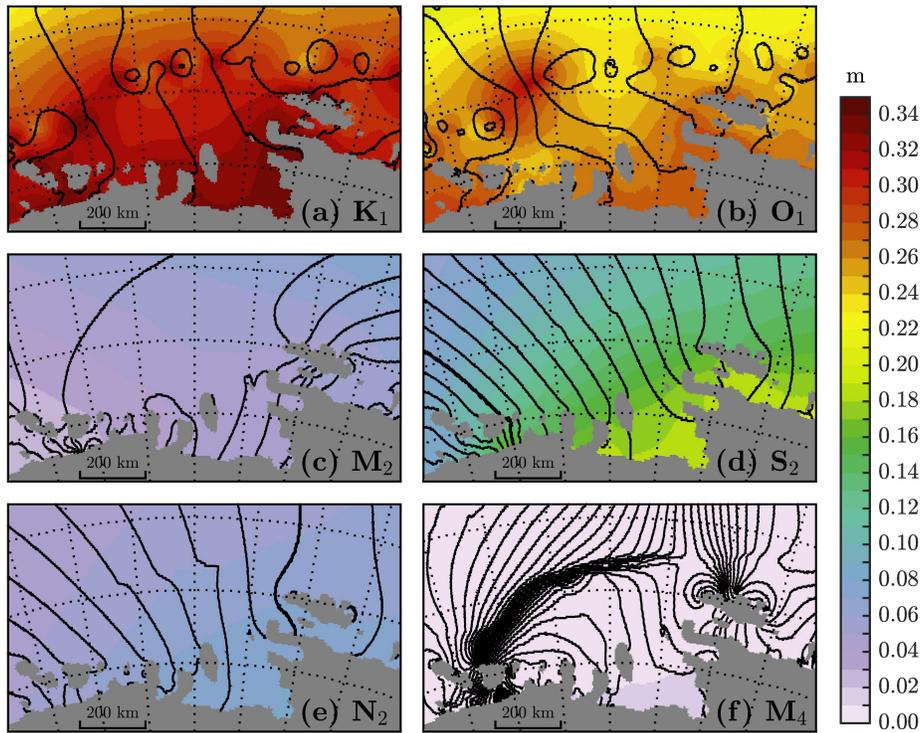


Figure S1: Amplitude (shaded) and phase (thin black, contour every 5°) of six SSH harmonics in FES2012. Grounded ice is in gray and dotted lines are latitude (every 2°) and longitude (every 5°).

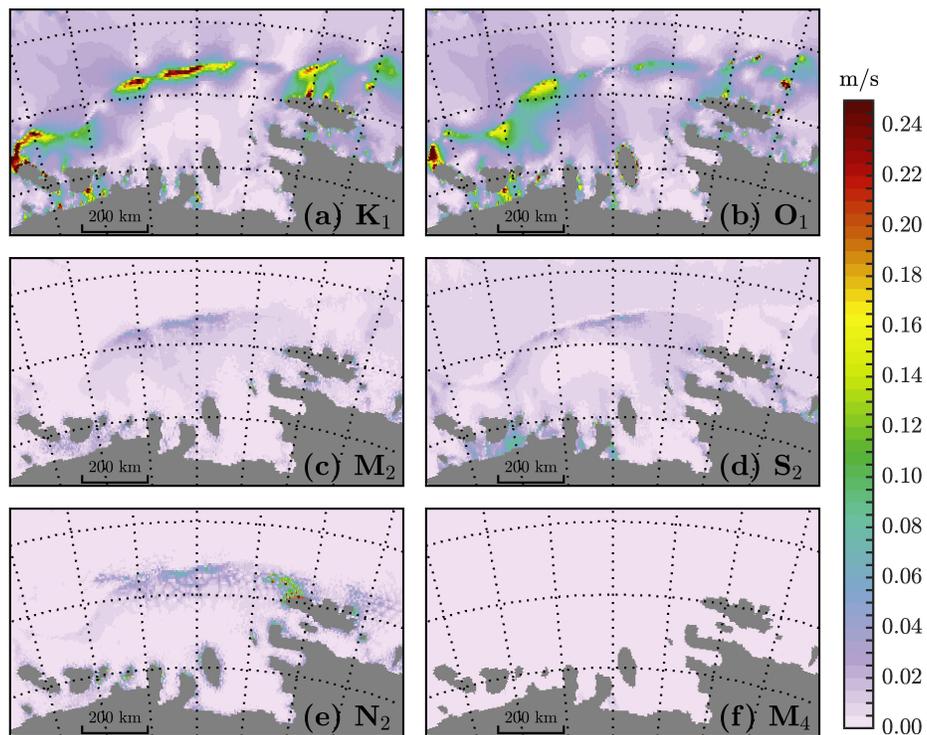


Figure S2: Major semi-axis of the tidal ellipse related to six barotropic-velocity harmonics (shaded) in FES2012. Grounded ice is in gray, and dotted lines are latitude (every 2°) and longitude (every 5°).

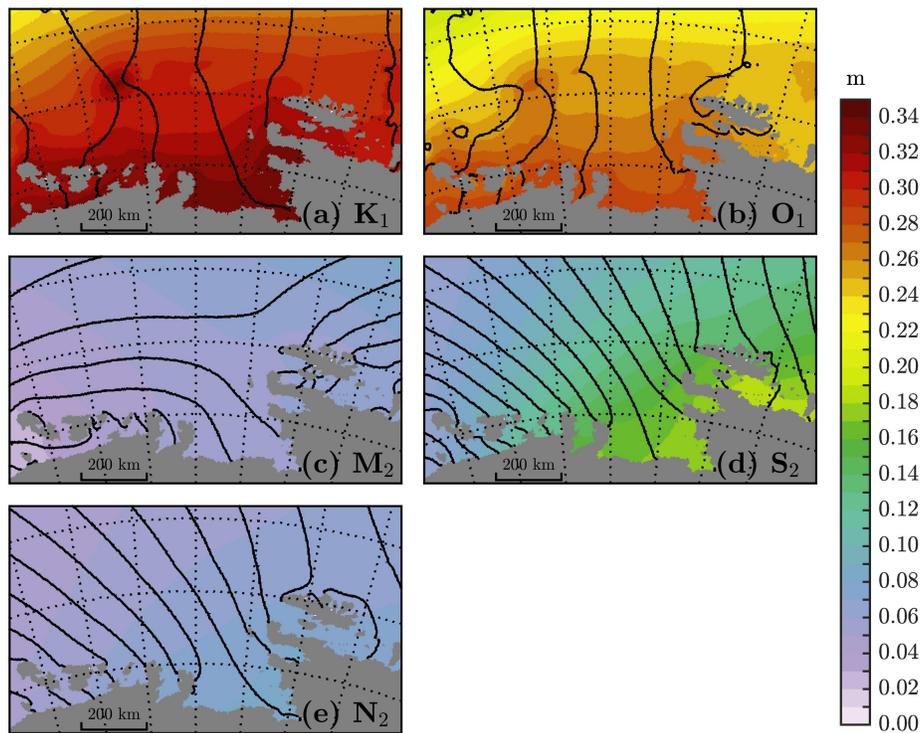


Figure S3: Amplitude (shaded) and phase (thin black, contour every 5°) of six SSH harmonics in CATS2008. Grounded ice is in gray and dotted lines are latitude (every 2°) and longitude (every 5°).

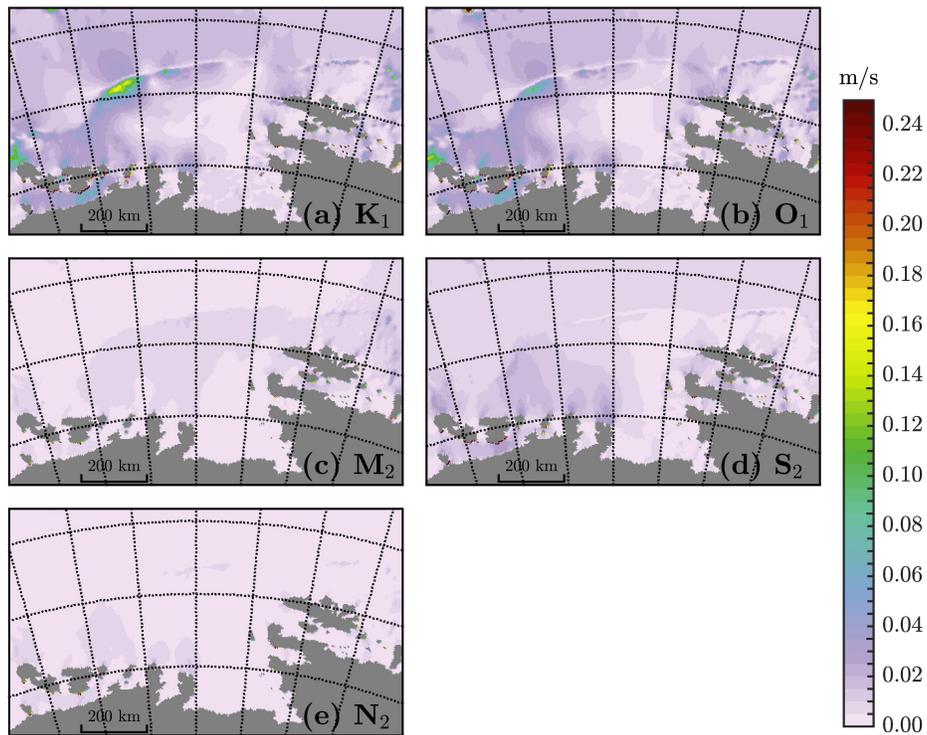


Figure S4: Major semi-axis of the tidal ellipse related to six barotropic-velocity harmonics (shaded) in CATS2008. Grounded ice is in gray, and dotted lines are latitude (every 2°) and longitude (every 5°).