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Highlights:

- 1) Simple coupled sea ice – ocean model has been developed to simulate the seasonal cycle of sea ice formation.
- 2) Salt flux associated with the wind-forced ice production causes haline convection affecting the characteristics of the entire water column.
- 3) The comparison between model-derived polynya extents and MODIS IST images was performed.
- 4) High resolution wind forcing is necessary to capture in more detail coastal sea ice processes, such as coastal polynyas, ice drift and ice compression against coastline features.

1 **Modelling sea ice formation in the Terra Nova Bay polynya**

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16 **Abstract**

17 Antarctic sea ice is constantly exported from the shore by strong near surface winds that open leads
18 and large polynyas in the pack ice. The latter, known as wind-driven polynyas, are responsible for
19 significant water mass modification due to the high salt flux into the ocean associated with
20 enhanced ice growth. In this article, we focus on the wind-driven Terra Nova Bay (TNB) polynya,
21 in the western Ross Sea. Brine rejected during sea ice formation processes that occur in the TNB
22 polynya densifies the water column leading to the formation of the most characteristic water mass
23 of the Ross Sea, the High Salinity Shelf Water (HSSW). This water mass, in turn, takes part in the
24 formation of Antarctic Bottom Water (AABW), the densest water mass of the world ocean, which
25 plays a major role in the global meridional overturning circulation, thus affecting the global climate
26 system. A simple coupled sea ice – ocean model has been developed to simulate the seasonal cycle
27 of sea ice formation and export within a polynya. The sea ice model accounts for both thermal and
28 mechanical ice processes. The oceanic circulation is described by a one-and-a-half layer, reduced
29 gravity model. The domain resolution is $1\text{ km} \times 1\text{ km}$, which is sufficient to represent the salient
30 features of the coastline geometry, notably the Drygalski Ice Tongue. The model is forced by a
31 combination of Era Interim reanalysis and in-situ data from automatic weather stations, and also by
32 a climatological oceanic dataset developed from in situ hydrographic observations. The sensitivity
33 of the polynya to the atmospheric forcing is well reproduced by the model when atmospheric in situ
34 measurements are combined with reanalysis data. Merging the two datasets allows us to capture in
35 detail the strength and the spatial distribution of the katabatic winds that often drive the opening of
36 the polynya. The model resolves fairly accurately the sea ice drift and sea ice production rates in the
37 TNB polynya, leading to realistic polynya extent estimates. The model-derived polynya extent has
38 been validated by comparing the modelled sea ice concentration against MODIS high resolution
39 satellite images, confirming that the model is able to reproduce reasonably well the TNB polynya
40 evolution in terms of both shape and extent.

1. Introduction

Observations and models have clearly shown that changes in atmospheric forcing and ocean circulation affect the Antarctic sea ice extent (Jacobs and Comiso, 1997; Liu et al., 2004; Lefebvre et al., 2005; Zhang 2007; Turner et al., 2009; Liu and Curry, 2010). The pronounced pattern of increasing ice cover in the Ross Sea region, found to be the highest contributor to sea ice expansion amongst the five Southern Ocean sectors in the 1979 - 2010 period with a positive trend of $13700 \pm 1500 \text{ km}^2 \text{ yr}^{-1}$, has been ascribed to changes in atmospheric circulation (Parkinson and Cavalieri, 2012). Enhanced northward winds have changed sea ice drift and export offshore affecting the dynamics of the local oceanography. These changes impact on the occurrence of wind driven polynyas along the Antarctic coastal margin, modifying the production of dense water masses through sea ice growth (Holland and Kwok, 2012). Variation in size or extent of polynyas are believed to be suitable indicators of climatic change (Morales Maqueda et al., 2004).

The wind-driven Terra Nova Bay (TNB) polynya, located in the western sector of the Ross Sea, plays a major role in shaping the sea ice and ocean dynamics of this region (Kurtz and Bromwich, 1985; Bromwich, 1989). The polynya opening results principally from the synergy of meteorological, oceanographic, and physical geography features of this region (Fig.1). Especially during winter, the TNB polynya is frequently forced by cold and strong katabatic downslope flows that push sea ice away from the coast. Their action prevents sea ice from consolidating as a thick pack and, at the same time, facilitates its continuous formation by leaving the relatively warm open water exposed to the cold atmosphere. Also, the presence and the orientation of the Drygalski Ice Tongue is essential for the polynya maintenance, since this barrier blocks the incoming sea ice from the south and controls, through its length, the polynya extent (Frezzotti and Mabin, 1994). Due to the constant formation and offshore drift of new ice, the TNB polynya contributes significantly to the sea ice mass budget of the whole area, producing approximately 10% of the sea ice formed annually in the Ross Sea (Kurtz and Bromwich, 1985; Van Woert, 1999b). Associated with the wind-forced ice production is a salt flux that causes haline convection that affects the characteristics

67 of the entire water column in the TNB polynya and the thermohaline structure of the whole Ross
68 Sea (Kurtz and Bromwich, 1985; Trumbore et al., 1991). The TNB polynya is considered to be by
69 far the largest producer of High Salinity Shelf Water (HSSW) (Kurtz and Bromwich, 1983, 1985;
70 Jacobs et al., 1985; Van Woert, 1999a, b; Budillon and Spezie, 2000; Budillon et al., 2003; Fusco et
71 al., 2009), a water mass that plays a crucial role in the formation of Antarctic Bottom Water
72 (AABW) (Kurtz and Bromwich, 1985; Jacobs and Comiso, 1989; Van Woert, 1999a), thus
73 contributing to deep ocean ventilation and the global thermohaline circulation (Jacobs et al., 1985;
74 Orsi et al., 1999; Jacobs 2004).

Fig. 1: Overview map of TNB (Western Ross Sea) showing the major geographical features of this region and its surroundings (Kurtz and Bromwich, 1983).

77 The main goal of this study is to investigate the sea ice behaviour in the Terra Nova Bay polynya in
78 response to external forcing and to estimate the associated sea ice and HSSW production. To this
79 purpose, a coupled sea ice – ocean model was developed and applied to the TNP polynya area. The
80 model simulates the seasonal cycle of sea ice formation in the TNB polynya, accounting for both
81 sea ice dynamic and thermodynamic processes. Dynamics does not produce ice directly, but causes
82 the ice to drift in and out of the area and leads to ice deformation in form of rafting or ridging due to
83 convergence. Thermodynamics processes are responsible for local ice growth or melt and heat
84 transfer at the ice-air and ice-ocean interfaces (Rothrock, 1979). Both ice dynamics and
85 thermodynamics alter the local mean thickness (ice volume per unit area) and result in the exchange
86 of mass, momentum and energy with the atmosphere and the ocean (Flato, 2003).

87 A further goal of this work is to estimate the variation of the TNB polynya extent, i.e. the size of the
88 area of low ice concentration, in response to the forcing. Computing the polynya extent is difficult
89 given the limitations of models and remote sensing tools as regards both their accuracy and their
90 ability to resolve polynya variability in space and time. Polynya extent estimates are not trivial to
91 derive since local ice thickness and ice production rates are often unknown. Papers focusing on the
92 variability of sea ice and open water in the TNB polynya exist in literature, mainly concerning the
93 wintertime season. Authors have investigated the TNB polynya extent either through one
94 dimensional models forced by in situ and reanalysis data (Van Woert, 1999a, 1999b; Fusco et al.,
95 2002; Petrelli et al., 2008) or through satellite observations (Kern, 2007; Ciappa et al., 2012).

96 The polynya extent in this paper is derived from modelled sea ice concentration (Markus et al.,
97 1998; Kwok et al., 2007; Tamura et al., 2008) and validated by comparison with polynya extents
98 estimates from MODIS satellite images (Key et al., 1994; Key et al., 1997).

99 The paper is organized as follows. Section 2 provides a description of the coupled sea ice – ocean
100 model and the main formulations adopted to resolve sea ice dynamics and thermodynamics. Section
101 3 presents experiments on model sensitivity to variations in specific physical processes and
102 parameterizations in order to better tune the model to the peculiarities of the TNB polynya region.

103 In particular, the sea ice and polynya response to wind forcing variations is studied. Section 4 shows
104 the results of a one year simulation of sea ice formation and polynya extent in the TNB polynya
105 region. Section 5 focuses on the comparison between the numerical TNB polynya extent estimates
106 and those derived from high resolution MODIS images. Finally, a discussion of results and a few
107 concluding remarks are presented in Section 6.

108 **2. Description of the model**

109 **2.1 Model equations**

110 The coupled sea ice – ocean model presented here provides an intermediate complexity formulation
111 of the TNB polynya dynamics. In contrast by polynya flux models describing the evolution of a
112 polynya in terms of the polynya edge contour, our model predicts sea ice concentrations over a
113 regular spatial grid (Willmott et al., 2007). The model has a relatively high resolution ($1\text{ km} \times 1$
114 km) in order to capture the complexity of the coastline geometry and the meteorological patterns of
115 the region. Both dynamic and thermodynamic sea ice and ocean processes are incorporated in the
116 model. An accurate representation of the main sea ice processes, often overlooked in numerical
117 simulations of the polar regions (Russell et al., 2006; Maksym et al., 2012), and a realistic
118 representation of sea ice dynamics, are crucial for the accurate description of the interactions of thin
119 ice and polynyas with the atmospheric and oceanic circulation (Stössel et al, 1990). The ocean is
120 represented by a one-and-a-half layer, reduced gravity, Boussinesq ocean model. The stratification
121 is simplified using a description in which the active layer (effectively, the upper ocean mixed layer)
122 moves above a lower stagnant (motionless) layer of infinite depth. The formulation of the ocean
123 model is inspired in Morales Maqueda et al. (1999) and Biggs and Willmott (2001), with the only
124 major departure that an eddy bolus transport term is added to the advection of scalars. The
125 equations are as follows.

$$\begin{aligned}
126 \quad & h_1 \left(\frac{\partial U_1}{\partial t} + \frac{\partial(U_1 U_1)}{\partial x} + \frac{\partial(V_1 U_1)}{\partial y} \right) - f h_1 V_1 = -\frac{\partial}{\partial x} \left(\frac{1}{2} g \frac{\rho_r - \rho_1}{\rho_r} h_1^2 \right) + \frac{1}{\rho_r} (\tau_s^x - \tau_b^x) + \frac{1}{\rho_r} \left(\frac{\partial}{\partial x} \left(A_H h_1 \frac{\partial U_1}{\partial x} \right) + \right. \\
127 \quad & \left. \frac{\partial}{\partial y} \left(A_H h_1 \frac{\partial U_1}{\partial y} \right) \right), \tag{1}
\end{aligned}$$

$$\begin{aligned}
128 \quad & h_1 \left(\frac{\partial V_1}{\partial t} + \frac{\partial(U_1 V_1)}{\partial x} + \frac{\partial(V_1 V_1)}{\partial y} \right) + f h_1 U_1 = -\frac{\partial}{\partial y} \left(\frac{1}{2} g \frac{\rho_r - \rho_1}{\rho_r} h_1^2 \right) + \frac{1}{\rho_r} (\tau_s^y - \tau_b^y) + \frac{1}{\rho_r} \left(\frac{\partial}{\partial x} \left(A_H h_1 \frac{\partial V_1}{\partial x} \right) + \right. \\
129 \quad & \left. \frac{\partial}{\partial y} \left(A_H h_1 \frac{\partial V_1}{\partial y} \right) \right), \tag{2}
\end{aligned}$$

$$130 \quad \frac{\partial h_1}{\partial t} = - \left\{ \frac{\partial[h_1(U_1 + U_e)]}{\partial x} + \frac{\partial[h_1(V_1 + V_e)]}{\partial y} \right\} - E + P + M_s + M_i - G_i + w_b, \tag{3}$$

$$131 \quad \frac{\partial(h_1 T_1)}{\partial t} = - \left\{ \frac{\partial[h_1 T_1(U_1 + V_e)]}{\partial x} + \frac{\partial[h_1 T_1(V_1 + V_e)]}{\partial y} \right\} + \frac{Q_1 + Q_i}{\rho_r c_p} + w_b T_b + h_1 \frac{T_c - T_1}{T_r}, \tag{4}$$

$$132 \quad \frac{\partial(h_1 S_1)}{\partial t} = - \left\{ \frac{\partial[h_1 S_1(U_1 + U_e)]}{\partial x} + \frac{\partial[h_1 S_1(V_1 + V_e)]}{\partial y} \right\} + (M_i - G_i) S_i + w_b S_b + h_1 \frac{S_c - S_1}{T_r}. \tag{5}$$

133

134 In the equations above, h_1 is the depth of the ocean active layer, U_1 and V_1 are the x and y
135 components of the ocean current in the active layer, U_1 , f is the Coriolis parameter, g is the
136 acceleration of gravity, ρ_r is a constant and uniform reference density representative on average of
137 the densities encountered in the region below the active layer, ρ_1 is the density of the active layer,
138 τ_s^x and τ_s^y are the components of the horizontal wind stress at the top of the active layer and,
139 similarly, τ_b^x and τ_b^y are the components of the vertical shear stress at the bottom of the active layer,
140 $U_e = -\kappa_e h_1^{-1} \partial h_1 / \partial x$ and $V_e = -\kappa_e h_1^{-1} \partial h_1 / \partial y$ are the components of the eddy bolus velocity
141 (Gent and McWilliams, 1990), with κ_e a constant and uniform thickness diffusivity, E is
142 evaporation, P is precipitation, M_s and M_i are the volume fluxes associated with snow and ice melt,
143 respectively, G_i is ice growth (exception made of snow ice formation, which is described below),
144 w_b represents the vertical volume flux at the base of the active layer caused by entrainment, T_1 is
145 the temperature of the active layer, and Q_1 is the net surface heat flux into the layer, incorporating
146 contributions from the ice free and ice covered areas and also including the latent heat loss required
147 to melt the snow that falls over leads. Specifically,

$$Q_1 = (1 - A)(Q_{sw} + Q_{lw} + Q_s + Q_e - P L_f \rho_s) - A Q_c \quad (6)$$

where A is the fractional oceanic area covered by ice, or ice concentration, Q_{sw} , Q_{lw} , Q_s and Q_e are the shortwave, longwave, sensible and latent heat fluxes in the open ocean, the four quantities positive if they flow into the ocean and calculated according to Budillon et al. (2000), L_f is the latent heat of fusion of ice, ρ_s is snow density, and Q_c is the conductive heat flux through ice, which we take as positive if directed upwards and assume to be identical to the heat flux from the ocean into the ice at the base of the ice cover. Small enthalpy changes associated with the mass fluxes E , P , M_s , M_i and G_i are all ignored here since we neglect the thermal inertia of both snow and ice. Further in (4), Q_i is a latent heat flux associated with the formation or melting of ice in the active oceanic layer (see below) and T_b is the temperature associated with the volume flux w_b across the base of the active layer. The last term on the right hand side of (4) represents a relaxation to observations, T_c , with a time scale T_r . In (5), S_1 is the salinity of the active layer and S_i and S_b are the salinity of sea ice, which we take as constant and equal to 4, and the salinity of water entrained at the base of the active layer, respectively. Relaxation to salinity observations, S_c , is also included. The temperature T_1 is approximately maintained at the freezing point, T_f , which is calculated according to Fofonoff and Millard (1983), for as long as ice is present or if ice forms for the first time in the season in the open ocean. The term Q_i in (4) ensures this, and it has the form

$$Q_i = \rho_r c_p \{ H(A) + [1 - H(A)] H(T_f - T_1) \} h_1 \frac{T_f - T_1}{T_i}. \quad (7)$$

In (7), H is the Heaviside function, A is the ice concentration, and T_i is a restoring time scale which we have chosen to be equal to the model time step. The entrainment velocity, w_b , depends on the air-sea stress and buoyancy fluxes that control the strength of turbulence in the mixed layer and is parameterized according to Lemke (1987) as

$$w_b = \frac{2 u^{*3} e^{-h_1/h_w} + h_1 B [1 + H(B) (e^{-h_1/h_c} - 1)]}{g \frac{\rho_r - \rho_1}{\rho_r} h_1} \quad (8)$$

171 where $\rho_r u^{*2} = \sqrt{\tau_s^{x^2} + \tau_s^{y^2}}$, $h_w = 7 \text{ m}$, B is the surface buoyancy flux, and $h_c = 50 \text{ m}$. The
 172 temperature and salinity of entrained water, T_b and S_b , are determined using mooring based
 173 observations (see Section 2.3) linearly interpolated onto the bottom of the active layer.

174 The surface stress term is formulated as in Mellor and Kantha (1989) and is a linear combination of
 175 the shear stress at the surface of the ice covered ocean and the wind stress acting on the open ocean,
 176 weighted by the fractional area of sea ice and leads, respectively, namely, $\tau_s^x = (1 - A)\tau_{sw}^x + A\tau_{si}^x$,
 177 where τ_{sw}^x is the x component of the surface stress over the ocean and τ_{si}^x is the sea ice counterpart.
 178 An analogous equation holds for τ_s^y . In full, the ice-ocean stress components are

$$179 \quad (\tau_{si}^x, \tau_{si}^y) = \rho_i C_i \sqrt{(U_1 - U_i)^2 + (V_1 - V_i)^2} (U_i - U_1, V_i - V_1), \quad (9)$$

180 where $C_i = 5 \times 10^{-3}$ is the ice-ocean drag coefficient, and U_i and V_i are the components of the ice
 181 velocity. The stress at the base of the active layer is calculated according to the parameterization of
 182 Pacanowski and Philander (1981):

$$183 \quad (\tau_b^x, \tau_b^y) = \left(\nu_b + \frac{\nu_0}{(1 + \alpha Ri)^n} \right) \frac{1}{h_1} (U_1, V_1), \quad (10)$$

184 where $\nu_b = 1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$, $\nu_0 = 1 \times 10^{-1} \text{ m}^2 \text{ s}^{-1}$, $\alpha = 0.5$, $n = 2$

185 and Ri is a Richardson number:

$$186 \quad Ri = \frac{g \frac{\rho_r - \rho_1}{\rho_r} h_1}{u_1^2 + v_1^2}. \quad (11)$$

187 The horizontal friction terms in (1) and (2) are included to make it possible to impose no-slip
 188 boundary conditions at coastlines. The viscosity, A_H , has a value of $2 \times 10^2 \text{ m}^2 \text{ s}^{-1}$.

189 Sea ice behaves as a floating, zero layer system (i.e., without thermal inertia), as proposed by
 190 Semtner (1976). Sea ice interacts thermodynamically and dynamically with the atmosphere and the
 191 underlying mixed layer of the ocean. The coupling of sea ice with the surface ocean layer allows us
 192 to simulate the seasonal cycle of sea ice formation and export within the polynya. The model
 193 requires atmospheric and ocean forcing as inputs that are applied as surface and bottom boundary
 194 conditions. The atmospheric forcing is given by air temperature, surface pressure, humidity, cloud

210 the thickening of sea ice when the snow layer is depressed below the sea surface by its own weight,
 211 is parameterized. The conservation equations for the sea ice and snow state variables are as follows.

$$212 \quad \frac{\partial A}{\partial t} = - \left\{ \frac{\partial(Au_i)}{\partial x} + \frac{\partial(Av_i)}{\partial y} \right\} + S_A, \quad (12)$$

$$213 \quad \frac{\partial(Ah_i)}{\partial t} = - \left\{ \frac{\partial(Ah_i u_i)}{\partial x} + \frac{\partial(Ah_i v_i)}{\partial y} \right\} + \dot{V}_i + \dot{V}_{i0} + \dot{V}_{si}, \quad (13)$$

$$214 \quad \frac{\partial(Ah_s)}{\partial t} = - \left\{ \frac{\partial(Ah_s u_i)}{\partial x} + \frac{\partial(Ah_s v_i)}{\partial y} \right\} + \dot{V}_{s0} - \frac{\rho_i}{\rho_s} \dot{V}_{si}, \quad (14)$$

215 where u_i and v_i are the components of the sea ice velocity. The term S_A is a placeholder for all non-
 216 advective processes that lead to changes in ice concentration. It is formulated in the following way.

$$217 \quad S_A = (1 - A) \frac{Q_i / (L_f \rho_i)}{d}, \quad (15)$$

218 where, L_f is the latent heat of formation of ice, and, if $Q_i > 0$ (i.e., there is frazil ice formation in
 219 the water column), d is a collection thickness of frazil ice which is either a constant or a function of
 220 wind speed as in Winsor and Bjork (2000), and we denote it by H , while, if $Q_i < 0$, $d = h_i$. The
 221 terms $\dot{V}_i = Q_i / (L_f \rho_i)$, $\dot{V}_{i0}$ and $\dot{V}_{si}$ in (13) account for ice growth/decay as a result of the heat budget
 222 in the active oceanic layer, for sea ice melting at the ice-atmosphere interface when the ice heat
 223 balance is such that the surface is at the melting point, and for snow ice formation, respectively. We
 224 note that the term G_i in (3) is simply equal to $\rho_i / \rho_r \dot{V}_i$. Snow ice formation occurs when the weight
 225 of snow depresses the snow-ice interface below sea level. In such cases, we transform the
 226 submerged snow into ice, leading to a contribution to the ice growth rate that obeys the formula

$$227 \quad \dot{V}_{si} = \frac{A}{T_{si}} H \left(\frac{\rho_s h_s - (\rho_r - \rho_i) h_i}{\rho_r} \right) \frac{\rho_s}{\rho_i} \frac{\rho_s h_s - (\rho_r - \rho_i) h_i}{\rho_r}, \quad (16)$$

228 where T_{si} is a time scale that we take to be equal to the time step in the model. The term $\dot{V}_{s0}$ in (14)
 229 accounts for both the accumulation of precipitating snow on top of sea ice and the subsequent
 230 surface snow melt during the thaw. As stated above, we make use of an approach for the sea ice and
 231 snow thermodynamics that is commonly termed zero layer approximation. This means that the
 232 vertical temperature profiles in the snow and the ice are linear and fully determined by the
 233 temperatures at their respective top and bottom interfaces. At the ice-ocean interface, the

234 temperature is always taken to be T_f , and we further assume that the oceanic heat flux into the ice
 235 matches the conductive heat flux through the ice, Q_c , so that there is never a flux divergence at the
 236 bottom of the ice cover. At the ice-snow interface, the conductive heat flux is also assumed to be
 237 continuous. The conductive heat flux is given by the following expression (e.g., Fichefet and
 238 Morales Maqueda, 1997):

$$239 \quad Q_c = \frac{\kappa_i}{h_i + \frac{\kappa_i}{\kappa_s} h_s} (T_f - T_0), \quad (17)$$

240 where κ_i and κ_s are the heat conductivities of ice and snow, respectively, and T_0 is the surface
 241 temperature. Finally, at the surface of the ice or snow cover a balance is postulated between surface
 242 and conductive heat fluxes, namely,

$$243 \quad Q_0 = Q_{swi} + Q_{lwi} + Q_{si} + Q_{ei} + Q_c = 0, \quad (18)$$

244 where Q_{swi} , Q_{lwi} , Q_{si} and Q_{ei} are the shortwave, longwave, sensible and latent heat fluxes at the
 245 surface of the snow or ice layer. The balance $Q_0 = 0$ in (18) can be guaranteed as long as T_0
 246 remains below the freezing point. However, if (18) requires that the surface temperature be above
 247 freezing then melting will ensure. During melting, T_0 will remain at the freezing point for
 248 freshwater (if snow is present) or for ice (if there is no snow) and the excess heat $Q_0 > 0$ will be
 249 used to melt snow, $M_s = Q_0 / (L_f \rho_s)$, or ice $M_i = Q_0 / (L_f \rho_i)$.

250 Sea ice drift, $\mathbf{U}_i$, with components U_i and V_i , is computed by postulating a balance of momentum
 251 between the Coriolis force, wind and ocean stresses and the ice internal force resulting from the
 252 interaction between floes during ice deformation. The momentum equations are:

$$253 \quad m \frac{\partial U_i}{\partial t} - f m V_i = A(\tau_w^x - \tau_o^x) + F^x, \quad (19)$$

$$254 \quad m \frac{\partial V_i}{\partial t} + f m U_i = A(\tau_w^y - \tau_o^y) + F^y, \quad (20)$$

255 where m is the mass of snow plus ice per unit area, τ_w^x and τ_w^y are the components of the wind stress
 256 acting on the ice, τ_o^x and τ_o^y are the components of the ice-ocean stress at the base of the ice, and
 257 F^x and F^y are the components of the ice internal stress force. Note that advection of sea ice

258 momentum is ignored and that the atmosphere and ocean stresses term includes the ice
259 concentration as a multiplicative factor to be consistent with the theory of free drift in regions of
260 low ice concentration according to Connolley et al. (2004). The internal ice forces are resolved
261 using the elastic-viscous-plastic rheology by Hunke and Dukowicz (1997). The internal ice pressure
262 is formulated as a function of sea ice thickness, h_i , and concentration, A , as in Hibler (1979) (see
263 further discussion in section 3.1).

264 **2.2 Model domain and set up**

265 The model domain consists of a wide region of the western Ross Sea including an extended area
266 along the coast of Victoria Land south of the Drygalski Ice Tongue and the northern region of the
267 Wood Bay (Fig. 3). It is $154 \text{ km} \times 488 \text{ km}$, extending approximately from 74°S to 78°S in latitude
268 and from 162°E to 168°E in longitude. A spatially uniform horizontal resolution of 1 km is used to
269 study the small scale behaviour of sea ice in TNB. This resolution is considered to be sufficient in
270 representing the salient features of the coastline geometry, such as the Drygalski Ice Tongue.
271 Hence, the horizontal grid is a rectangle of width X and length Y subdivided in square grid cells
272 resulting in a grid of $154 \times 488 = 75152$ grid points. An Arakawa B-grid is used for the spatial
273 discretization. A land mask is specified in the center of the cells with 0 representing land and 1
274 oceanic cells. A corresponding mask is defined for all corner quantities such as the wind speed, sea
275 ice velocity and stress components.

Fig. 3: The model domain showing the Drygalski Ice Tongue (DIT) and the two preferential paths of the katabatic flows, the Priestley and the Reeves Glaciers. The Italian base, Mario Zucchelli Station (MZS), and the location of the automatic weather stations, Rita (AWS-R: 74.72°S, 164.03°E) and Manuela (AWS-M: 74.95°S, 163.69°E) are also indicated.

276

277

278 The advection of scalars is discretized with a first order upstream scheme. The solutions of the
 279 momentum equations (1), (2), (19) and (20) and the advective and thermodynamic processes in eqs.
 280 (3), (4), (5), (12), (13) and (14) are computed using two different time steps: a small one for the
 281 momentum (Δt) and a larger one for the advection (ΔT_a). All the input parameters such as constants
 282 and coefficients are shown in Table 1. Some values of the input parameters, referred to as “x” in
 283 Table 1, are let to vary in the sensitivity experiments. A one month spin-up length (repeated twice)
 284 is used for each experiment. Sea ice concentration and thickness and ocean fields are initialized at
 285 the beginning of each integration with a prescribed value or with a restart from a previous

286 integration. The initial values are zero for both ice concentration and thickness. The initial
287 temperature and salinity of the ocean are from monthly climatological observations (described in
288 section 2.3) and the initial active layer depth is determined from the former using the density
289 threshold criterion of 0.125 kg m^{-3} relative to near surface densities (Monterey and Levitus, 1997).
290 Open lateral boundary conditions ensuring a minimum of signal reflections at the boundary have
291 been used so that advective flows leaving the domain are allowed to freely exit the domain using an
292 upstream formulation, while flows into the domain use a simple sponge boundary condition that
293 relaxes the variables to their climatological external values (Martinsen and Engedahl, 1987). The
294 main physical parameters of atmosphere, sea ice and ocean used in the model are showed in Table
295 2. Figure 4 is the flow diagram of the coupled sea ice-ocean model, showing the basic steps in
296 computing the diagnostic variables of the model.

297

298	Parameter	Symbol	Value
	X domain	X	154000 m
	Y domain	Y	488000 m
299	T domain	T	x days
	Time step for momentum	Δt	1.2 s
	Time step for advection	Δt_a	600 s
300	Elastic timescale (EVP ice rheology)	Δt_e	180 s
	Air drag coefficient	C_{da}	x
	Ocean drag coefficient	C_{do}	x
301	Ice strength parameter	P^*	x N/m^2
	Ice concentration parameter	C	20
	Creep limit	c	$5 \times 10^{-11} \text{ 1/s}$
302	Eccentricity of the elliptical yield curve	e	2
	Demarcation ice collection thickness in leads	H	x m

303

304 **Table 1:** Input parameters of the model. The “x” stands for a varying value assigned to that parameter in the sensitivity experiments.

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Parameter	Symbol	Value
Ocean horizontal viscosity	A_H	$2 \times 10^2 \text{ m}^2 \text{ s}^{-1}$
Ocean eddy thickness diffusivity	κ_e	$2 \times 10^2 \text{ m}^2 \text{ s}^{-1}$
Thermal conductivity of sea ice	κ_i	2.2 W/m/K
Thermal conductivity of snow	κ_s	0.3 W/m/K
Emissivity of atmosphere	ε_a	0.95
Emissivity of ocean	ε_o	0.985
Albedo of ocean	α_o	0.07
Albedo of ice	α_i	0.07-0.7
Albedo of snow	α_{sn}	0.85
Latent heat of fusion of ice	L_{fi}	$3.34 \times 10^5 \text{ J/kg}$
Latent heat of vaporization of water	L_e	$2.5 \times 10^6 \text{ J/kg}$
Latent heat of fusion of snow	L_{fsn}	$3.34 \times 10^5 \text{ J/kg}$
Latent heat of sublimation of snow	L_{ssn}	$2.834 \times 10^6 \text{ J/kg}$
Specific heat capacity of ocean	c_{pa}	3985 J/kg/K
Specific heat capacity of air	c_{pa}	1004 J/kg/K
Density of air	ρ_a	1.3 Kg/m ³
Density of ice	ρ_i	900 Kg/m ³
Density of snow	ρ_s	330 Kg/m ³
Density of ocean	ρ_o	1024 Kg/m ³
Melting point of freshwater ice	t_{fus}	0°C
Salinity of sea ice	s_i	4
Exchange coeff. for sensible heat (leads/ice)	C_H	1.75×10^{-3}
Exchange coeff. for latent heat over leads	C_E	1.75×10^{-3}
Exchange coeff. for latent heat over ice	C_E	1×10^{-3}
Stefan-Boltzmann constant	K	$5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$
Minimum vertical viscosity	v_{min}	$1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$
Scale depth of mechanical dissipation	h_w	7 m
Scale depth of convective dissipation	h_c	50 m

Table 2: Physical parameters of atmosphere, sea ice and ocean.

Fig. 4: Diagram flow of the coupled sea ice - ocean model.

330 2.3 Forcing Fields

331 The ocean forcing consists of climatological oceanographic profiles of ocean temperature and
332 salinity developed through the analysis of available in situ temperature and salinity datasets. These
333 datasets consist of hydrographic mooring and CTD profile data collected from February 1995 to
334 January 2008 within the CLIMA (Climatic Long-term Interaction for the Mass-balance in
335 Antarctica) project of the Italian National Research Antarctic Program (PNRA). The two
336 climatological datasets include idealized monthly temperature and salinity values, spatially uniform
337 in the model domain and varying vertically down to 800 meters in depth. In detail, 8 depth levels (0
338 m, -30 m, -50 m, -100 m, -150 m, -300 m, -500 m, -800 m) are chosen for the computation of
339 climatological temperature and salinity profiles (see Section 2.3) that are subsequently used to
340 calculate, by linear interpolation, the temperature, T_b , and salinity, S_b , at the base of the active layer
341 as well as the restoring temperature, T_c , and salinity, S_c , in the active layer. As main atmospheric
342 forcing, the Era-Interim reanalysis from the European Centre for Medium-Range Weather Forecasts
343 (ECMWF), has been prescribed. The data extracted from the global domain provide surface six-
344 hourly parameters at a 0.5×0.5 degree horizontal resolution covering the model domain with $16 \times$
345 11 grid points, in latitude $\times$ longitude. Specifically, the input data consist of the 10 meter eastward
346 and northward wind components (m/s), the 2 meter temperature (K), the downward surface solar
347 radiation (Wm^{-2} s), the surface (1000 mb level) pressure (Pa), the relative humidity (%), the total
348 cloud cover (0-1) and the total precipitation accumulation (m of water). The oceanographic and
349 atmospheric data have been spatially and temporally interpolated over the whole model domain.
350 Meteorological observations from Automatic Weather Stations (AWSs) have also been employed
351 to force the model.

352

353

354 2.3.1 Atmospheric field setting

355

356 The resolution of the local winds is a crucial factor in estimating sea ice and HSSW production,
357 especially in a small coastal polynya like the TNB. In particular during winter, sea ice production in
358 TNB is largely determined by katabatic winds which are the main control of the TNB polynya size
359 (Petrelli et al., 2008; Gallé, 1997). Petrelli et al. (2008) showed that an insufficient resolution of the
360 katabatic winds leads to an underestimation of sea ice winter production of up to 50%, which results
361 in an underestimation of the formation rate of HSSW and consequently of AABW. On the other
362 hand, low resolution winds reanalyses can also result in higher ice and AABW production because
363 of an overestimated offshore component due to the coarse resolution orography (Stössel et al.,
364 2011).

365 In spite of their relatively high resolution, the ECMWF reanalysis have been found to underestimate
366 the wind speeds in several studies (Cullather et al., 1997; Fusco et al., 2002; Petrelli et al., 2008),
367 providing therefore an improper representation of the wind fields along and offshore TNB. Also
368 Mathiot et al. (2010) investigated the effect of the katabatic winds on sea ice and shelf water
369 properties by correcting the ECMWF reanalyses winds with results from the MAR regional
370 atmospheric model. To remedy this problem, we have applied a wind correction to coastal and
371 offshore model grid points value combining the Era Interim data with in-situ atmospheric data from
372 Automatic Weather Stations (AWSs), which show a significantly increased skill over ECMWF
373 atmospheric variables (Petrelli et al., 2008).

374 A merging function has been designed so that the correction factor for each grid point value varies
375 with the distance from the weather station. Era Interim and AWS data are merged resulting in the
376 effective wind vector defined as

$$377 V_{eff} = V_{AWS} e^{-\frac{r}{R}} + V_{Era} \left(1 - e^{-\frac{r}{R}}\right), \quad (21)$$

378 where V_{AWS} and V_{Era} are the wind vectors from AWS and ERA-Interim, respectively, r is the
379 distance from the AWS and R is an e-folding length scale.

380 In particular, atmospheric data from two AWSs have been used. In a first phase of the sensitivity
381 tests, only the Rita AWS (-74.72° S, 164.03° E), installed within the Meteo-Climatological
382 Observatory of the PNRA in close proximity to the Italian base “Mario Zucchelli”, downstream of
383 the Priestley Glacier (Fig. 3), has been considered.

384 Subsequently, the Manuela AWS (-74.95° S, 163.69° E), installed as part of the AWS project of
385 the University of Wisconsin-Madison Antarctic Meteorology Program on Inexpressible Island, has
386 been also included. The AWS Manuela lies downstream of the Reeves Glacier (Fig. 3), which
387 represents one of the main routes for the katabatic flows from the interior of Antarctica.

388 The Rita and Manuela datasets consist, respectively, of one hourly and ten minute intervals data
389 including air temperature ($^{\circ}$ C), wind speed (m/s) and direction ($^{\circ}$ N), surface pressure (hPa), and
390 relative humidity (%). The merging function (21) has been applied also to the air temperature and
391 relative humidity data.

392 Three experiments (Table 3) were carried out in order to examine the influence of the parameter R
393 (see eq. 21) on the wind fields and, consequently, on sea ice and ocean currents in the TNB area.

394 The model was run with the rheology parameter $P^* = 27500 \times 10^4$ N/m², as in Hibler and Walsh
395 (1982), which is the most widely used value for the ice strength parameter. A demarcation ice
396 collection thickness, $H = 0.1$ m and a relaxation time of $T_r = 7$ days to climatological oceanic data
397 were used. The time interval of the atmospheric input was set to 6 hours, while that of the output
398 fields is such that the model gives a daily output for each computed variable.

399 The control experiment, referred to as CASE 1, uses $R=25$ km, the second experiment, CASE 2,
400 uses a larger $R=50$ km, while CASE 3 differs from the control experiment in the absence of the
401 merging between reanalyses and AWS data. The results of the simulations (Fig. 5a to c) show that
402 an increasing of the parameter R (CASE 2) leads, as expected, to larger wind and ice drift and ocean
403 velocities. The wind velocities have maximum values of 19.66 m/s in CASE 1 and 20.65 m/s in
404 CASE 2, while in CASE 3, where the merging function is switched off, they reach a maximum

405 value of only 7.27 m/s. Also sea ice and ocean current velocities, which reach values of 0.34 m/s
406 and 0.06 m/s respectively in CASE 2, are significantly reduced in CASE 3.

407

Experiment	R (km)
CASE 1	25
CASE 2	50
CASE 3	no merging

410 **Table3:** Case studies on the influence of the parameter R on wind fields.

Fig. 5: Maps of wind speed (a), modelled ice drift velocity (b) and modelled ocean current (c) overlaid by the corresponding wind speed vectors, ice drift velocity vectors and ocean current vectors on 8th July 2000 for CASE 1 to CASE 3.

3. Sensitivity experiments

An improper choice of the parameters which describe sea ice evolution results often in unrealistic simulations leading to inaccurate results. Several sensitivity experiments were performed to define the best set of parameters controlling TNB sea ice dynamics and thermodynamics in response to wind forcing. Two key parameters have been found to control the wind driven polynyas: the rheological ice strength parameter P^* and the demarcation ice collection thickness H , also named the lead-closing parameter. The rate at which the leads close under freezing conditions is inversely proportional to the value of H . Both parameters have a strong effect on polynya size and sea ice extent and volume estimates (Hibler, 1979; Stössel et al., 1990; Stössel, 1992).

Finally a sensitivity analysis was carried out turning attention to the air-ice and ice-ocean drag coefficients which control the stresses on the sea ice cover. The choice of these parameters depends on the study area and especially on the wind forcing time and spatial resolution, therefore the model was opportunely tuned and optimized in this regard.

3.1 Sensitivity to ice strength parameter

The ice strength parameter P^* is a key element in sea ice rheology that relates sea ice strength (P) to its concentration, A , and mean thickness, Ah_i . It was first introduced by Hibler (1979) in the constitutive equation for sea ice strength as

$$P = P^* Ah_i \exp [-C(1 - A)] \quad (22)$$

where P^* and C are empirical values. The ice strength exhibits a strong dependence on sea ice concentration and especially on the amount of thin ice. For a large amount of thin ice, the ice strength decreases significantly and most of thin ice is deformed (Hibler, 1979; Willmott et al., 2007; Feltham, 2008). Sea ice also offers less resistance to compression when Ah_i and P^* are low, and tends to pile up more easily because of enhanced mechanical ridging and rafting. Therefore, P^* is a critical parameter controlling sea ice drift behaviour in wind driven polynyas and represents the

main tuning parameter to achieve a realistic sea ice drift pattern (Owens and Lemke, 1990; Stössel et al., 1990; Steele et al., 1997).

The strength of the ice internal forces depends on the state of deformation of the sea ice cover, which will, in turn be partly controlled by the wind stress field. Based on this observation, a sensitivity experiment was performed to investigate the impact on the ice dynamics of varying P^* . This experiment, referred to as CASE 4, is equal to CASE 1 shown in section 2.3.1 except for $P^* = 5000 \text{ N/m}^2$ (Hibler, 1979). The wind field is the same as CASE 1 (see Fig. 5a) and the sea ice drift and ocean fields on 8th July 2000 for CASE 4 are shown in Fig. 6. These results are very similar to those of CASE 1 for both sea ice and ocean outputs.

The reduced ice strength does not affect significantly the ridging of sea ice or the sea ice drift in convergent regions, altering relatively little the ice concentration and thickness distribution (not shown). This indicates the polynya area is not highly sensitive to P^* in the determination of its opening/closure for this set of forcing.

Fig. 6: Modelled ice drift velocity (left) and modelled ocean current (right) overlaid by the corresponding ice drift velocity vectors and ocean current vectors on 8th July 2000 for CASE 4.

451 This is probably due to the fact that this parameter has a major influence only in areas of thick ice
452 but not so much in regions covered by thin and broken ice cover (Kreysner et al., 2000).

453

454 **3.2 Sensitivity to demarcation ice collection thickness**

455 While the dynamic behaviour of the consolidated ice is greatly determined by the ice rheology, the
456 interior of the polynya is affected by the new ice thickness parameterization. The new ice thickness
457 is controlled by the demarcation ice collection thickness parameter H in eq. (15) that is expressed as
458 a transition value between thin ice (open water) and thick ice (Hibler, 1979). It is as crucial an
459 element in sea ice models as the ice thickness collection depth in polynya flux models (Tear et al.,
460 2003; Willmott et al., 2007) since it represents the thickness at which newly-formed ice in the
461 polynya is transferred into thicker sea ice. Thereby, it affects sea ice thermodynamics, lowering heat
462 loss through thin ice inside the polynya and determining primarily the mean thickness and sea ice
463 concentration of newly formed ice (Hibler, 1979; Olason and Harms, 2010).

464 H has been often defined as a constant in the literature, with typical values in the range 0.1-0.5 m
465 (Hibler, 1979; Pease, 1987; Ou, 1988; Darby et al. 1994, 1995). However, wind speed is an
466 important controlling factor of the collection thickness of new ice (Mellor and Kantha, 1989;
467 Winsor and Björk, 2000; Olason and Harms, 2010).

468 A number of experiments, outlined in Table 4, were performed with different values of H . The
469 control experiment (CASE 5) was run with $H=0.2$ m, that is considered more appropriate than the
470 $H=0.5$ m proposed by Hibler (1979), in simulating the behaviour of thin ice inside the polynya
471 (Olason and Harms, 2010). In the second (CASE 6) and third experiment (CASE 7), sea ice
472 concentration and thickness are simulated using a constant $H=0.3$ m and $H=0.4$ m respectively,
473 while in the fourth experiment (CASE 8) a varying H has been used. Specifically, the collection
474 depth parameterization of Winsor & Björk (2000) is employed, namely,

$$475 \quad H = \frac{a + V \cdot b}{c}, \quad (23)$$

476 where V is the surface wind speed (m/s) and the constants are $a = 1$ m, $b = 0.1$ s and $c = 15$. This
 477 means that H varies in the range 0.1-0.3 m in the presence of wind speeds between 5-35 m/s.
 478 In all these experiments, R is fixed to 50 km, except in CASE 9, in which the merging function is
 479 not applied. The value of 27500 N/m^2 for P^* and of 30 days for the relaxation time to oceanic
 480 forcing were used. Figure 7 shows the results of sea ice concentration (a) and thickness (b)
 481 simulation on 8th July 2000 for CASE 5–to–CASE 9. Note that a lower ice demarcation thickness
 482 gives higher ice concentration values and lower ice thickness values due to lower heat losses
 483 through leads.

484

Experiment	H (m)
CASE 5	0.2
CASE 6	0.3
CASE 7	0.4
CASE 8	f (V)
CASE 9	0.2 no merging

487 **Table 4:** Sensitivity tests of sea ice evolution with respect to H .
 488
 489

Fig. 7: Modelled sea ice concentration (a) and thickness (b) on 8th July 2000 for CASE 5 to CASE 9.

490

491 The sea ice distribution in CASE 8 is similar to that of CASE 5, suggesting that the dependence of
 492 the ice collection thickness on the wind velocities provides plausible values for H . This is well
 493 supported by the estimates of daily sea ice production (km^3/day) in TNB region on July 2000 (Fig.
 494 8). Cumulative sea ice production (km^3) for the whole of July 2000 is also showed in Table 5. Note
 495 that the TNB region is identified by the area of the domain that extends within the ranges 1-120 km
 496 in X (longitude) and 310-425 km in Y (latitude) as shown in Fig. 8. Sea ice production rate
 497 computed using a non-spatially uniform H (CASE 8) depicted by the solid line with square markers
 498 shows a trend very similar to that of CASE 5, except for a few days when wind speeds in the

499 polynya were particularly large. As it can easily be observed, CASE 9, where wind forcing is given
 500 by the winds reanalysis alone, underestimates considerably sea ice production rate compared to
 501 CASE 5 and, indeed, all the others. These results suggest reasonable agreement between the wind
 502 forcing and the simulated sea ice dynamics in the TNB.
 503 In the next section we will consider the sensitivity of the TNB polynya to the wind forcing and
 504 wind stress (Stössel, 1992; Stössel et al., 2011) pointing out the importance of high temporal
 505 resolution of wind data.

506

Experiment	Sea ice production (km ³) in July 2000
CASE 5	10.08
CASE 6	11.09
CASE 7	12.12
CASE 8	9.79
CASE 9	6.83

510 **Table 5:** Sea ice production in July 2000 for the experiments CASE 5 to CASE 9.

511 **Fig. 8:** Daily ice production (left) on July 2000 for CASE 5 to CASE 9 within a smaller area of the domain, defined as TNB region (right), extending
 512 approximately from 310 km to 425 km in Y and bordered by X = 120 km.

513 3.3 Sensitivity to air-ice and ice-ocean drag coefficients

514 Along with P^* , the atmospheric and oceanic drag coefficients have been identified as crucial
515 parameters for sea ice drift. Several sensitivity experiments were performed to obtain the optimal
516 set of drag coefficients that would allow us to run the model under more realistic conditions.

517 We also focused on the regime of the katabatic winds and its impact on sea ice evolution in the
518 TNB and on the polynya size. The latter, in fact, is very responsive to variations in the freezing
519 rates in the bay as a result of a weakening of the katabatic flows or a change in their direction
520 (Bromwich and Kurtz, 1984; Priestley, 1914). The duration of the katabatic wind events has a
521 greater contribution than the intensity and frequency of the katabatic flows in determining the
522 polynya extent (Ciappa et al., 2012; Rusciano et al., 2013). Rusciano et al. (2013) found most
523 frequent katabatic events take place during the winter season and last on average from one to three
524 hours. That means that long time intervals (daily/six hourly) atmospheric input probably
525 misrepresent the real and local atmospheric fields in a given temporal period. On the other hand, a
526 single source of AWS data fails to properly reproduce the geometry of the coastal wind regime
527 resulting from the drainage of the interior katabatic airflows through the different confluence
528 pathways (Petrelli et al., 2008). In view of these considerations, in the next experiments, the time
529 resolution was increased so that the model is able to capture any katabatic events. Furthermore, a
530 second dataset from AWS Manuela (see section 2.3.1) was taken into account to enlarge the area of
531 influence of the katabatic flows. Unfortunately, no other weather station is available in the
532 southernmost region of the bay and near to the Drygalski Ice Tongue. In addition, the merging
533 function was modified and the range of influence of the AWS data on the reanalysis data was let to
534 assume an elliptic shape rather than a circumference as follows:

$$535 \quad V_{eff} = V_{AWS} e^{-\sqrt{\frac{x^2}{R_1^2} + \frac{y^2}{R_2^2}}} + V_{Era} \left(1 - e^{-\sqrt{\frac{x^2}{R_1^2} + \frac{y^2}{R_2^2}}} \right), \quad (24)$$

536 where x and y are the components of the position vector of a particular point in the domain with
 537 respect to the AWS, and $R_1 = 50$ km and $R_2 = 20$ km are e-folding length scales in the x and y
 538 directions. Table 6 summarises the experiments performed to explore the impact of varying the C_{da}
 539 and C_{do} drag coefficients, increasing and/or decreasing the one with respect to the other, on sea ice
 540 drift and polynya dynamics. Substantially, an increasing of C_{da} and/or at the same time a decreasing
 541 of C_{do} allows sea ice to move faster and vice versa. We have made use of a double sub-index to
 542 identify easily the wind and ocean drag coefficients used in a particular experiment, e.g. the
 543 experiment denoted by E_{ab} uses $C_{da} = a \times 10^{-3}$ and $C_{do} = b \times 10^{-3}$.
 544 The first experiment (E_{15}) is the control simulation of one winter month of the year 2005 for which
 545 the model has been configured with constant and more commonly used values for the drag
 546 coefficients, $C_{da} = 1 \times 10^{-3}$ and $C_{do} = 5 \times 10^{-3}$. In the next experiments, the values of the two drag
 547 coefficients were allowed to vary individually or simultaneously with respect to those of the control
 548 run. Specifically, in the second experiment (E_{35}) C_{da} varies and C_{do} is the same as in the control
 549 run, in the third experiment (E_{11}), only C_{do} varies, while, in the fourth (E_{31}) and in the fifth (E_{34})
 550 experiments, both parameters vary together. The sixth experiment (E_r), which is in more detail
 551 described afterwards, was carried out using non constant values for the drag coefficients. All the
 552 experiments are forced with atmospheric forcing from the AWS Manuela at ten minutes resolution,
 553 combined with hourly data from the AWS Rita. The resulting values are averaged with the six
 554 hourly ERA-Interim data so as to adjust the background atmospheric fields, especially the winds. In
 555 addition, the output time of the variables simulated by the model were set equal to 3 hours since, as
 556 explained above, this value would appear to be a good compromise to capture the effects of
 557 katabatic winds.

558 As the sensitivity experiments described in previous sections, a significant dependence of the sea
 559 ice simulation on the wind forcing can be inferred from the results of the modelled output fields.
 560 The sea ice distribution appears to be very sensitive to the pattern of the wind stress which varies
 561 considerably depending on the surface winds. Fig. 9a shows the wind speeds and the wind stress

vector fields for the E_{15} , E_{35} , E_{11} , E_{31} and E_{34} . The wind field is the same for all the experiments since they have been forced with the same wind configuration, which has maximum wind speed values of up to 23 m/s and a mean value of 9 m/s. The wind stress, depending on the drag parameters, exhibits average values of 0.16, 0.41, 0.13, 0.27 and 0.40 N/m^2 in E_{15} , E_{35} , E_{11} , E_{31} , E_{34} , respectively. The largest values have been found, as expected, in E_{35} , E_{31} and E_{34} with maxima of 1.48, 1.34 and 1.47 N/m^2 versus much smaller maxima in the CTRL run (E_{15}) and in E_{11} of approximately 0.54 N/m^2 .

Experiment	C_{da}	C_{do}
E_{15} CTRL	1×10^{-3}	5×10^{-3}
E_{35}	3×10^{-3}	5×10^{-3}
E_{11}	1×10^{-3}	1×10^{-3}
E_{31}	3×10^{-3}	1×10^{-3}
E_{34}	3×10^{-3}	4×10^{-3}
E_r	1×10^{-3} $V \leq 10$ m/s 3×10^{-3} $V \geq 20$ m/s	$1.3 \times C_{da}$

Table 6: Sensitivity tests with respect to the air-ice and ice-ocean drag coefficients. The double sub-index identifies the wind and ocean drag coefficients used in each experiment.

Fig. 9: Wind speeds (a), sea ice drift velocities (b) and ocean currents (c) with the superimposed wind stress, ice drift and ocean current vectors, respectively on 30th July 2005 for E_{15} , E_{35} , E_{11} , E_{31} and E_{34} .

581 The bigger wind stress in E_{35} , E_{31} and E_{34} leads to maximum ice drift speeds (Fig. 8b) of 0.65, 1.37
582 and 0.70 m/s respectively, and also to larger ocean currents (e. 8c). A maximum ice drift of 0.81
583 m/s, comparable to that from E_{34} , result from E_{11} , where the two coefficients C_{da} and C_{do} differ the
584 least from each other. Smaller values, as expected, result from E_{15} with a maximum of 0.37 m/s and
585 a mean of 0.12 m/s. Sea ice concentration and thickness charts displayed in Fig. 10a and b reveal
586 that the sea ice distribution in E_{35} , E_{11} and E_{34} show a good comparison, from a qualitative point of
587 view, with MODIS scenes represented in Ciappa et al. (2012). In these experiments the gap between
588 C_{da} and C_{do} is small. In contrast, when C_{do} is much smaller than C_{da} , the ice drift becomes
589 unrealistic and too strong also in regions out of the range of the coastal winds or, in the opposite
590 case, really insignificant along shore. These results supports the importance of the C_{da}/C_{do} ratio
591 considered to be the most basic dynamics parameter determining the mean drift speed (McPhee,
592 1980; Lepparanta, 1981; Stössel, 1992; Geiger et al., 1998; Harder and Fisher, 1999).

593

Fig. 10: Modelled sea ice concentration (a) and thickness (b) on 30th July 2005 for E_{15} , E_{35} , E_{11} , E_{31} and E_{34} .

594

595

596 Furthermore, unlike the strength parameter P^* which has a strong impact mainly on thick and more
597 compact areas of the pack ice, the C_{da}/C_{do} ratio influences the ice drift in all regions during all
598 seasons (Kreysner et al., 2000). The dependence of the air-ice drag coefficient on the wind speed
599 has been also investigated by several authors (Large and Pond, 1981; Overland, 1985; Lynch et al.,
600 1997). Accordingly, in the last experiment (E_r) C_{da} was allowed to vary linearly from 1×10^{-3} for
601 wind speeds below 10 m/s, to 3×10^{-3} for wind speeds above 20 m/s. Then, C_{do} is allowed to
602 depend linearly on the C_{da} through a constant factor of 1.3 (McPhee, 1980; Lepparanta, 1981;
603 Stössel, 1992).

604 Figure 11 shows wind and ice velocities with the superimposed wind stress and ice velocity vector
605 fields, ice concentration and thickness maps for E_r on the 30th July 2005 at 24:00. Mean and
606 maximum values of the wind stress are very similar to those resulting from E_{34} . The results of E_r
607 provide the best simulations of the sea ice dynamics of TNB.
608

Fig. 11: Wind speeds and modelled sea ice drift velocities with the superimposed wind and ice drift vectors field (top) and sea ice concentration and thickness distribution (bottom) on 30th July 2005 for E_r .

4. One year numerical simulation and results

One year simulation of the TNB sea ice evolution has been carried out to investigate the polynya behaviour in response to the local katabatic flows. The main results of the 2005 simulation are discussed. The modelled polynya behaviour follows the characteristic dynamics of sea ice and ocean circulation in TNB. During the summer season, approximately from November to March, the bay is mostly ice free. It starts to be covered by sea ice in late March, when the low atmospheric and oceanic temperatures let the sea surface freeze. The evolution of the polynya is strongly controlled by the action of katabatic winds which allow TNB to be almost never completely ice covered in winter. Katabatic winds are very intense between April and October (Rusciano et al. 2013), and within this period several cycles of opening/closure of the polynya occurred.

Model-derived polynya extents in TNB region, defined in the section 3.2, have been computed for 2005. The polynya area is usually defined as the sum of the surfaces of open water and thin sea ice and therefore is restricted to the oceanic region within which the ice concentration is smaller than a given threshold (Willmott et al. 2007). This threshold is rather arbitrary, varying commonly from 0.5 to 0.7 (Parmiggiani 2006; Kern et al., 2007). An ice concentration threshold of 0.7 has been used here to estimate the TNB polynya extent (Fig. 12).

Fig. 12: Model-derived polynya extent in the TNB region from March to October 2005.

627

628 An increase of the polynya size is associated with the occurrence of katabatic events (not shown).

629 The peak extent in midwinter occurred in July with a maximum value of 2962 km², followed by
 630 other two large extents of the polynya in August and September of 2868 km² and 2674 km²
 631 respectively (Table 7).

632 Polynya mean extents vary approximately from just over 500 km² up to almost 900 km², except in
 633 March/April, when sea ice formation processes start, and in October, which represents the end of
 634 the wintertime and the beginning of sea ice melting processes. The computed polynya extents are in
 635 good agreement with the wintertime values estimated by Petrelli et al. (2008) and with those
 636 recently published by Ciappa et al. (2012) who computed a mean annual open water of around 900
 637 km² in the period 2005-2010 and 600 km² in 2006 using MODIS thermal infrared data. In any case,
 638 the computation of the polynya extent is not trivial since it depends on the accuracy and the
 639 limitations of the models and the remote sensing tools, as well as on their capability to resolve in
 640 time and in space the processes involved in the polynya variability. In addition, the local coastal
 641 winds have a strong but not exclusive impact on the polynya size which is caused by the interaction
 642 between katabatic forcing and synoptic weather conditions on longer timescales. The major effect

643 of the katabatic winds on short timescales is the local recirculation of sea ice in TNB and its
644 redistribution within the polynya area (Petrelli et al., 2008). The recirculation forced by these local
645 winds enhances the ice production maintaining high ice production rates in open water and thin ice
646 regions.

Winter months	Maximum Polynya extent (km ²)	Mean Polynya extent (km ²)
March	7946	5574
April	1806	1174
May	2688	871.2
June	2205	557.2
July	2962	532.5
August	2868	766.9
September	2674	875.6
October	5637	2304

651 **Table 7:** Monthly maximum and mean polynya extent of the TNB polynya from March to October 2005.
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654 Sea ice production in the TNB region (Table 8) has been also computed by model sea ice fields
655 outputs. The ice production rate, depends primarily on the presence of open water and on the
656 surface wind speeds, therefore following the same trend as the TNB polynya extent. The spatial
657 maximum sea ice production daily rate over TNB area exhibits a maximum of 0.70 km³/day on 30th
658 July that is equivalent to 48.08 cm/day. These estimates are comparable to those of Petrelli et al.
659 (2008), who simulated the TNB polynya using a coupled atmosphere–sea ice model. She found in
660 her high resolution experiment an ice production maximum daily rate of 26.4 cm/day during winter.
661 Our results are quite consistent, even if slightly smaller, with ice production estimates obtained by
662 Fusco et al. (2002) by applying a one-dimensional flux model to the TNB polynya. She computed
663 for August 1993 and 1994 a maximum value of ice production of 85 cm/day and 72 cm/day
664 respectively. Our daily ice productions result in a cumulative ice production value of 39.29 m over
665 2005 versus her yearly ice production of 81.7 m and 68.8 m for 1993 and 1994. However, these
666 large values in Fusco et al. (2002) were obtained using AWS forcing, whereas the ice production
667 was found to be significantly underestimated when computed by using the ECMWF data only. The
668 spatially cumulative daily ice production is also showed in Fig. 13. The highest peaks of ice
669 production occur in May, June and July with maxima of 0.61, 0.54 and 0.70 km³ respectively. The

670 cumulative ice production, that is the sea ice volume produced in the whole year 2005, is 57.91
671 km³. This value is consistent with the estimation by Tamura et al. (2008) based on satellite data in
672 combination with ERA-40 reanalysis data, which shows for the TNB polynya a mean annual
673 cumulative sea ice production of 59.2 ± 10 km³. In particular, the ice volume created in the months
674 of June and July amounts overall to 16.37 km³, which is in good agreement with the value of 16.4
675 km³ computed by Petrelli et al. (2008) in her winter experiment. The brine rejection, associated with
676 the new ice production, and the HSSW production are also calculated. The brine rejection (kg/day)
677 is parameterized as $P_S = \rho_i P_i (S_1 - S_i) \times 10^{-3}$ (see Markus et al., 1998; Van Woert, 1999a) where
678 P_i is the ice production rate. The HSSW production (m³/day) is computed following Van Woert
679 (1999a) as $P_{HSSW} = P_S / \rho_{HSSW} (S_{HSSW} - S_{LSSW}) \times 10^{-3}$ where ρ_{HSSW} is the density of HSSW
680 (1030.45 kg/m³), S_{HSSW} is the salinity of HSSW (34.8) and S_{LSSW} is the salinity of Low Salinity
681 Shelf Water or Warm Core Water (34.5) (Jacobs et al., 1985).
682 The salt and HSSW (Fig. 14) production are larger in wintertime, when the ice production is higher.
683 Their cumulative values in the year 2005 within the TNB polynya are 1.7×10^{12} kg and 0.5×10^{13} m³
684 respectively. These values are in good agreement with those of Fusco et al. (2002), Fusco et al.
685 (2009) and Van Woert (1999a). Fusco et al. (2002), for example, estimated a salt production of
686 about 4.6×10^{12} kg and a HSSW production of 1.5×10^{13} m³ in the years 1993-94.

Winter months	Maximum daily rates of sea ice production (km ³ /day)	Mean daily rates of sea ice production (km ³ /day)	Monthly cumulative sea ice (km ³)
March	0.42	0.16	4.99
April	0.40	0.26	7.86
May	0.61	0.30	9.25
June	0.54	0.25	7.52
July	0.70	0.22	6.98
August	0.58	0.30	9.39
September	0.44	0.24	7.34
October	0.39	0.14	4.29

692 **Table 8:** Daily sea ice production rates from spatially cumulated ice production in TNB polynya region from March to October 2005.

Fig. 13: Spatially cumulated daily rate of sea ice production in the TNB region from March to October 2005.

Fig. 14: Daily salt production (top) and HSSW production (bottom) in the TNB region from March to October 2005.

698 **5. Model comparison with MODIS data**

699 In situ measurements are particularly poor in remote or hardly accessible areas during the Antarctic
700 winter, therefore satellite observations represent a useful tool in tuning sea ice-ocean models (Linch
701 et al., 1997). Satellite images in combination with numerical weather prediction model data and in
702 situ data from Automatic Weather Stations provide a good database to study polynya-atmosphere
703 interactions in TNB area (Gallée, 1997; Ciappa et al., 2012).

704 Measurements of ice thickness and total ice volume in Terra Nova Bay do not exist. However, the
705 model can be ground-truthed, at least in part, by comparing the polynya shape and extent to satellite
706 images, which we do in the following. The NASA's MODIS (Moderate Resolution Imaging
707 Spectroradiometer) sensor provides high temporal and spatial resolution measurements of Earth's
708 land, ocean and atmospheric processes in several spectral bands and swath. The MODIS/Aqua
709 Level 1B 1km Calibrated Radiances at 1 km resolution have been used to retrieve the ice surface
710 temperature (IST) in the TNB region and subsequently to derive the polynya extent. Radiance data
711 from MODIS channels 31 and 32 are converted to brightness temperatures (Kelvins) through the
712 inversion of the Planck's law equation (Key et al., 1994). For ice/snow surface temperature (IST)
713 computation the equation based on the technique of Key et al. (1997), originally developed for the
714 Advanced Very High Resolution Radiometer (AVHRR), is used.

715 In order to investigate the dependence of the opening/closing cycles of the polynya on the wind
716 forcing, a few significant periods in the wintertime of 2005 characterized by strong katabatic events
717 have been identified. For each period, sea ice concentration charts from ice fields model outputs
718 have been produced. The polynya edge is identified by the first contour line characterized by an ice
719 concentration threshold of 0.7. These maps have been compared with MODIS IST images obtained
720 following the aforementioned procedure for the same period. Figures 15 and 17 show the wind
721 speed from both Rita and Manuela AWSs during two katabatic events observed in May and July

722 (1th - 5th May and 28th - 31th July respectively). The evolution of the polynya extent detected by
723 MODIS can be seen in Figs. 16 and 18 where the modelled sea ice concentration for the same days
724 is also showed. Sea ice concentration maps at the temporal steps closer to those of satellite scenes
725 have been chosen to match at best model and MODIS products.

726 The model reproduces, reasonably well, sea ice concentration, depth and velocity as seen from the
727 comparison with MODIS images. The drift of sea ice responds to wind forcing which shows a
728 predominant West-West North West direction. Stronger winds are responsible for sea ice advection
729 offshore, opening the polynya, and contributing to increasing its extent, while weaker winds just
730 hamper the closure keeping the polynya opened. According to Pease (1987), a seaward wind
731 component exceeding 10m/s is sufficient to maintain a polynya in coastal zones. Our results are in
732 agreement with the suggested threshold.

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Fig. 15: Wind speed from Rita (top) and Manuela (bottom) AWSs on 1th-5th May 2005.

Fig. 16: IST MODIS scenes (top) and the modelled sea ice concentration and thickness maps (bottom) displaying the polynya evolution on 1th - 5th May 2005.

Fig. 17: Wind speed from Rita (top) and Manuela (bottom) AWSs on 28th - 31th July 2005.

Fig. 18: IST MODIS scenes (top) and the modelled sea ice concentration and thickness maps (bottom) displaying the polynya evolution on 28th - 31th July 2005.

741

742 The small polynya observed at the beginning of the 1st of May (Fig. 16) increases its extent on the
743 2nd day of the month upon an increase of the wind speed measured by the AWS Rita, exhibiting a
744 value well over 20 m/s and reaching a peak of almost 50 m/s. Wind speed from the AWS Manuela
745 with an average value of 20 m/s contributes to enlarge the polynya eastward. The polynya size
746 keeps on increasing at the beginning of 3rd of May until the wind speed drops sharply, below 10 m/s
747 for AWS Rita, and the polynya starts closing on 4th and 5th of May. The discrepancies between the
748 spatial distribution of sea ice in model simulations and IST MODIS scenes are thought to be partly
749 due to the iceberg B-15A drifting in front of the TNB approximately in April-May 2005. The
750 presence of this iceberg blocks the drift of sea ice offshore forcing the ice to accumulate in its
751 proximity. In fact, in IST MODIS scenes the edge of the polynya is located more toward the coast
752 and southward reducing thus the northern portion of the whole polynya extent. The simulation of

753 sea ice distribution in July 2005 (Fig. 18) shows a higher degree of similarity with that observed in
754 satellite images, probably because the advection of sea ice is less affected by the iceberg moving
755 out of the bay. On 28th July the polynya is almost totally closed because the wind speeds are near to
756 zero. After an enhancement of the wind forcing, the polynya starts opening at the beginning of the
757 29th of July and expands rapidly seaward. The largest opening of the polynya occurs on 31th July
758 2005 in response to the stronger wind speeds values recorded previously by AWS stations, near to
759 50 m/s for Rita and 40 m/s for Manuela. Some discrepancies between the simulated polynya and
760 that observed in MODIS scenes may probably due to the gaps (missing data) in the AWSs wind
761 datasets.

762 The TNB polynya extents have been also derived from both MODIS IST scenes and sea ice
763 concentration maps on 28th -31th July. The aforementioned sea ice concentration threshold of 0.7
764 has been used for the modelled ice. A varying threshold for IST proposed by Ishikawa et al. (1996)
765 and Zwally et al. (1983) that discriminates open water and thin ice from thick ice or land fast ice has
766 been employed for satellite maps. Setting sea ice concentration to 0.7, our IST threshold is given by
767 $T_{th} = 0.3T_f + 0.7T_{ice}$ where T_f is the temperature of the open water at the freezing point and T_{ice} is
768 the temperature of sea ice around the polynya. Both temperature values are extracted from the IST
769 scenes after they have been visually inspected one by one. In particular, T_f is given by the warmest
770 IST found within the polynya and T_{ice} is estimated as the average of the IST values found around
771 the open water.

772 Polynya extents from the 28th July to the 31th July are showed in Table 9. Polynya extent on the 31th
773 July represents the largest opening of the whole of 2005, as also found in Ciappa et al. (2012). The
774 model-derived polynya extents mostly agree with those computed from MODIS IST images,
775 revealing the same temporal trend in polynya increasing during the observed katabatic event. The
776 polynya extent values are less comparable to the MODIS based extents retrieved by Ciappa et al.
777 (2012) showing a polynya extent of approximately 7615 km² on 31 July at 04:25 versus the
778 corresponding MODIS-derived and model-derived polynya extents of 3393 km² and 2831 km²,

779 respectively. That is due to the wider domain considered in his estimates, including all the open
 780 water fraction occurred north of TNB (Wood Bay) and south of the Drygalski Ice Tongue.

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TNB polynya event in July 2005	Model-derived polynya extent (km ²)	MODIS-derived polynya extent (km ²)
28 th 12:00	12	40
29 th 06:00	0	25
29 th 15:00	389	391
30 th 06:00	1858	1936
30 th 12:00	2148	2385
31 th 03:00	2831	3393

786 **Table 9:** TNB polynya extents from model sea ice concentration outputs and from MODIS IST from 28th to 31th July 2005.

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789 **6. Summary and concluding remarks**

790 This work focuses on the investigation of sea ice formation in the TNB polynya in response to wind
 791 forcing. Because of the lack of direct observations related to sea ice fields, models provide valuable
 792 insight into the mean state of the ice cover (Flato, 2003) together with satellite observations which
 793 indeed fail often in availability and spatial resolution. A coupled sea ice-ocean model that simulates
 794 the seasonal cycle of sea ice formation and export within the polynya was presented. The model is
 795 applied to the TNB area, including also the nearby regions north and south of the bay in order to
 796 characterize at the best seasonal sea ice variability and polynya behaviour. The horizontal resolution
 797 is of 1 km, which is sufficient to represent the salient features of the coastline geometry, notably the
 798 Drygalski Ice Tongue. The model has been forced by a combination of Era Interim reanalysis by
 799 ECMWF and in-situ data from Rita and Manuela AWS, and also by in situ oceanic data.

800 The modelled sea ice fields have proved to be very sensitive to the atmospheric forcing. The sea ice
 801 evolution has been found to be shaped by different parameters involved in the dynamics of sea ice
 802 which in turn affects the thermal processes that occur in the ice cover. Several sensitivity
 803 experiments have been performed in order to optimize and set up a few main parameterizations and

804 coefficients, thus improving the model outputs. The use of a demarcation ice collection thickness
805 (H) varying with the wind speed used by Winsor & Björk (2000) seems to be the best choice,
806 amongst the ones considered here, for simulating sea ice fields and thermodynamic heat losses
807 through thin ice inside the polynya. In contrast, the rheology parameter P^* has not been found to
808 affect significantly the drift of sea ice in this region, resulting in almost unchanged outputs of sea
809 ice concentration and thickness distribution irrespective of the value used for P^* . The importance of
810 the air drag coefficient, one of the most important factors in modelling ice motion, has been also
811 stressed. First the responses of the model to constant values of the air-ice (C_{da}) and ice-ocean (C_{do})
812 drag coefficients and subsequently to the C_{da}/C_{do} ratio have been investigated, the latter being the
813 most basic parameter of sea ice dynamics in determining the mean sea ice drift speed (Geiger et al.,
814 1998; Harder and Fisher, 1999). A C_{da} varying with wind speed has been adopted, while C_{do} is
815 forced to depend linearly on C_{da} through a constant factor. Also a wind enhancement function has
816 been developed in order to try to improve the prediction of sea ice fields. However, its application
817 was unsuccessful, causing unrealistically high large values of the wind stress.

818 A simulation of sea ice formation in TNB has been performed for the entire year 2005 to investigate
819 the response of the polynya dynamics to wind forcing. Unsurprisingly, the largest openings of the
820 polynya match the stronger katabatic winds which have been found in wintertime, mainly from
821 April to October. The largest polynya opening occurs in July, with an extent of 2962 km², while the
822 polynya extent over the wintertime 2005 ranges between approximately 500 km² and 900 km². Sea
823 ice production and the associated brine and HSSW productions have also been computed, exhibiting
824 values cumulated over 2005 of 57.91 km³, 1.7×10^{12} kg and 0.5×10^{13} m³, respectively. These results
825 are in good agreement with those reported by Fusco et al. (2002, 2009) who estimated a salt
826 production of about 4.6×10^{12} kg and a HSSW production of 1.5×10^{13} m³ for the period 1993-1994.

827 In order to support and validate the model outputs, a comparison with sea ice conditions detected by
828 satellite images has been thought essential. Satellite images from the MODIS sensor have been
829 chosen for this purpose since they reach a high spatial resolution of 1 km, the same as that of the

830 model. In order to explore the strong relationship between the wind field and the TNB polynya
831 extent, some wintertime periods including significant katabatic events have been selected. For these
832 periods the MODIS IST scenes have been compared with the modelled sea ice concentration maps.
833 The TNB polynya area seems to be reproduced reasonably well by the model in terms of both shape
834 and distribution of sea ice. However, differences in sea ice distribution respect to that observed in
835 the MODIS IST scenes are visually detectable in some regions. These differences are most
836 prominent in the areas located along the coast characterized by the variable shelf-ice borders and
837 the presence of land fast ice. In particular, two areas, namely, the region south of Drygalski Ice
838 Tongue and the region north of the TNB (Wood Bay) appear almost recurrently ice free in the
839 modelled sea ice maps.

840 The TNB polynya extent has been also derived from MODIS IST scenes and the corresponding
841 model sea ice maps from the 28th to 31th July 2005. The application of an ice state dependent
842 threshold for IST in MODIS images let us to validate the polynya extent with higher reliability. The
843 model-derived polynya extents are very similar to those computed from MODIS IST images.

844 Finally, despite the discrepancies in both sea ice distribution in some regions and polynya extents,
845 the model performs well in reproducing sea ice evolution. These discrepancies will be investigated
846 more extensively in the future through either an improvement of the model to capture land-fast ice
847 or, more simply the use of a more accurate land mask including fast ice. The remote sensing
848 detection of the polynya area and its extent is obviously affected by fog, clouds or other
849 atmospheric disturbance that often compromise the quality of the used satellite images. At any rate,
850 modelling the opening and closing polynya events is a difficult task especially if the size of polynya
851 is relatively small, as is the case in Terra Nova Bay (Pease, 1987; Lynch et al., 1997; Petrelli et al.,
852 2008). Our results have further highlighted the sensitivity of sea ice simulations to wind forcing,
853 which is the major aspect stressed in numerous modelling works on Southern Ocean. Accurate sea
854 ice simulations in terms of sea ice distribution and thickness can be achieved, provided that the
855 model is forced with realistic winds and surface boundary conditions, in particular ocean

temperatures, as found by Stössel et al. (2011). High resolution wind forcing is necessary to capture in more detail coastal sea ice processes, such as coastal polynyas, ice drift and ice compression against coastline features.

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1032 **Figure and table captions**

1033 **Fig. 1:** Overview map of TNB (Western Ross Sea) showing the major geographical features of this
1034 region and its surroundings (Kurtz and Bromwich, 1983).

1035 **Fig. 2:** Schematic view of the main variables of the coupled sea ice – ocean model. The radiative
1036 and turbulent heat fluxes are separately calculated over the ice free (leads) and ice covered areas.

1037 **Fig. 3:** The model domain showing the Drygalski Ice Tongue (DIT) and the two preferential paths
1038 of the katabatic flows, the Priestley and the Reeves Glaciers. The Italian base, Mario Zucchelli
1039 Station (MZS), and the location of the automatic weather stations, Rita (AWS-R: 74.72°S,
1040 164.03°E) and Manuela (AWS-M: 74.95°S, 163.69°E) are also indicated.

1041 **Fig. 4:** Diagram flow of the coupled sea ice - ocean model.

1042 **Fig. 5:** Maps of wind speed (a), modelled ice drift velocity (b) and modelled ocean current (c)
1043 overlaid by the corresponding wind speed vectors, ice drift velocity vectors and ocean current
1044 vectors on 8th July 2000 for CASE 1 to CASE 3.

1045 **Fig. 6:** Modelled ice drift velocity (left) and modelled ocean current (right) overlaid by the
1046 corresponding ice drift velocity vectors and ocean current vectors on 8th July 2000 for CASE 4.

1047 **Fig. 7:** Modelled sea ice concentration (a) and thickness (b) on 8th July 2000 for CASE 5 to CASE
1048 9. The color version of this figure is available online.

1049 **Fig. 8:** Daily ice production (left) on July 2000 for CASE 5 to CASE 9 within a smaller area of the
1050 domain, defined as TNB region (right), extending approximately from 310 km to 425 km in Y and
1051 bordered by X = 120 km.

1052 **Fig. 9:** Wind speeds (a), sea ice drift velocities (b) and ocean currents (c) with the superimposed
1053 wind stress, ice drift and ocean current vectors, respectively on 30th July 2005 for E₁₅, E₃₅, E₁₁, E₃₁
1054 and E₃₄.

1055 **Fig. 10:** Modelled sea ice concentration (a) and thickness (b) on 30th July 2005 for E₁₅, E₃₅, E₁₁, E₃₁
1056 and E₃₄. The color version of this figure is available online.

1057 **Fig. 11:** Wind speeds and modelled sea ice drift velocities with the superimposed wind and ice drift
1058 vectors field (top) and sea ice concentration and thickness distribution (bottom) on 30th July 2005
1059 for E_r.

1060 **Fig. 12:** Model-derived polynya extent in the TNB region from March to October 2005.

1061 **Fig. 13:** Spatially cumulated daily rate of sea ice production in the TNB region from March to
1062 October 2005.

1063 **Fig. 14:** Daily salt production (top) and HSSW production (bottom) in the TNB region from March
1064 to October 2005.

1065 **Fig. 15:** Wind speed from Rita (top) and Manuela (bottom) AWSs on 1th - 5th May 2005.

1066 **Fig. 16:** IST MODIS scenes (top) and the modelled sea ice concentration and thickness maps
1067 (bottom) displaying the polynya evolution on 1th - 5th May 2005. The color version of this figure is
1068 available online.

1069 **Fig. 17:** Wind speed from Rita (top) and Manuela (bottom) AWSs on 28th - 31th July 2005.

1070 **Fig. 18:** IST MODIS scenes (top) and the modelled sea ice concentration and thickness maps
1071 (bottom) displaying the polynya evolution on 28th - 31th July 2005. The color version of this figure
1072 is available online.

1073 **Table 1:** Input parameters of the model. The “x” stands for a varying value assigned to that
1074 parameter in the sensitivity experiments.

1075 **Table 2:** Physical parameters of atmosphere, sea ice and ocean.

1076 **Table3:** Case studies on the influence of the parameter *R* on wind fields.

1077 **Table 4:** Sensitivity tests of sea ice evolution with respect to H .

1078 **Table 5:** Sea ice production in July 2000 for the experiments CASE 5 to CASE 9.

1079 **Table 6:** Sensitivity tests with respect to the air-ice and ice-ocean drag coefficients. The double
1080 sub-index identifies the wind and ocean drag coefficients used in each experiment.

1081 **Table 7:** Monthly maximum and mean polynya extent of the TNB polynya from March to October
1082 2005.

1083 **Table 8:** Daily sea ice production rates from spatially cumulated ice production in TNB polynya
1084 region from March to October 2005.

1085 **Table 9:** TNB polynya extents from model sea ice concentration outputs and from MODIS IST
1086 from 28th to 31th July 2005.

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CASE 4

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Parameter	Symbol	Value
X domain	X	154000 m
Y domain	Y	488000 m
T domain	T	x days
Time step for momentum	Δt	1.2 s
Time step for advection	Δt_a	600 s
Elastic timescale (EVP ice rheology)	Δt_e	180 s
Air drag coefficient	C_{da}	x
Ocean drag coefficient	C_{do}	x
Ice strength parameter	P^*	x N/m ²
Ice concentration parameter	C	20
Creep limit	c	5×10 ⁻¹¹ 1/s
Eccentricity of the elliptical yield curve	e	2
Demarcation ice collection thickness in leads	H	x m

Parameter	Symbol	Value
Ocean horizontal viscosity	A_H	$2 \times 10^2 \text{ m}^2 \text{ s}^{-1}$
Ocean eddy thickness diffusivity	κ_e	$2 \times 10^2 \text{ m}^2 \text{ s}^{-1}$
Thermal conductivity of sea ice	κ_i	2.2 W/m/K
Thermal conductivity of snow	κ_s	0.3 W/m/K
Emissivity of atmosphere	ϵ_a	0.95
Emissivity of ocean	ϵ_o	0.985
Albedo of ocean	α_o	0.07
Albedo of ice	α_i	0.07-0.7
Albedo of snow	α_{sn}	0.85
Latent heat of fusion of ice	L_{fi}	$3.34 \times 10^5 \text{ J/kg}$
Latent heat of vaporization of water	L_e	$2.5 \times 10^6 \text{ J/kg}$
Latent heat of fusion of snow	L_{fsn}	$3.34 \times 10^5 \text{ J/kg}$
Latent heat of sublimation of snow	L_{ssn}	$2.834 \times 10^6 \text{ J/kg}$
Specific heat capacity of ocean	c_{pa}	3985 J/kg/K
Specific heat capacity of air	c_{pa}	1004 J/kg/K
Density of air	ρ_a	1.3 Kg/m ³
Density of ice	ρ_i	900 Kg/m ³
Density of snow	ρ_s	330 Kg/m ³
Density of ocean	ρ_o	1024 Kg/m ³
Melting point of freshwater ice	t_{fus}	0°C
Salinity of sea ice	s_i	4
Exchange coeff. for sensible heat (leads/ice)	c_H	1.75×10^{-3}
Exchange coeff. for latent heat over leads	c_E	1.75×10^{-3}
Exchange coeff. for latent heat over ice	c_E	1×10^{-3}
Stefan-Boltzmann constant	K	$5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$
Minimum vertical viscosity	ν_{min}	$1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$
Scale depth of mechanical dissipation	h_w	7 m
Scale depth of convective dissipation	h_c	50 m

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Experiment	<i>R</i> (km)
CASE 1	25
CASE 2	50
CASE 3	no merging

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Experiment	<i>H</i> (m)
CASE 5	0.2
CASE 6	0.3
CASE 7	0.4
CASE 8	f (V)
CASE 9	0.2 no merging

Table(s)
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Experiment	Sea ice production (km ³) in July 2000
CASE 5	10.08
CASE 6	11.09
CASE 7	12.12
CASE 8	9.79
CASE 9	6.83

Table(s)
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Experiment	C_{da}	C_{do}
E ₁₅ <u>CTRL</u>	1×10^{-3}	5×10^{-3}
E ₃₅	3×10^{-3}	5×10^{-3}
E ₁₁	1×10^{-3}	1×10^{-3}
E ₃₁	3×10^{-3}	1×10^{-3}
E ₃₄	3×10^{-3}	4×10^{-3}
E _t	1×10^{-3} $V \leq 10 \text{ m/s}$ 3×10^{-3} $V \geq 20 \text{ m/s}$	$1.3 \times C_{da}$

Winter months	Maximum Polynya extent (km ²)	Mean Polynya extent (km ²)
March	7946	5574
April	1806	1174
May	2688	871.2
June	2205	557.2
July	2962	532.5
August	2868	766.9
September	2674	875.6
October	5637	2304

Winter months	Maximum daily rates of sea ice production (km ³ /day)	Mean daily rates of sea ice production (km ³ /day)	Monthly cumulative sea ice (km ³)
March	0.42	0.16	4.99
April	0.40	0.26	7.86
May	0.61	0.30	9.25
June	0.54	0.25	7.52
July	0.70	0.22	6.98
August	0.58	0.30	9.39
September	0.44	0.24	7.34
October	0.39	0.14	4.29

TNB polynya event in July 2005	Model-derived polynya extent (km ²)	MODIS-derived polynya extent (km ²)
28 th 12:00	12	40
29 th 06:00	0	25
29 th 15:00	389	391
30 th 06:00	1858	1936
30 th 12:00	2148	2385
31 th 03:00	2831	3393