1 **Synchronous intensification and warming of Antarctic Bottom Water**

2 **outflow from the Weddell Gyre**

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5 Michael P. Meredith[1], Arnold L. Gordon[2], Alberto C. Naveira Garabato[3], E. Povl

6 Abrahamsen[1], Bruce A. Huber[2], Loïc Jullion[3] and Hugh J. Venables[1]

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9 *[1] British Antarctic Survey, High Cross, Madingley Road, Cambridge, CB3 0ET, U.K.*

10 *[2] Lamont-Doherty Earth Observatory of Columbia University, 61 Route 9W,*

11 *Palisades, New York, 10964-8000, U.S.A.*

12 *[3] National Oceanography Centre Southampton, European Way, Southampton, SO14*

13 *7ZH, U.K.*

14

14 **Abstract**

15

16 Antarctic Bottom Water (AABW), the densest water in the global overturning

17 circulation, has warmed in recent decades, most notably in the Atlantic. Time series

18 recorded within the boundary currents immediately upstream and downstream of the

19 most significant outflow of AABW from the Weddell Sea indicate that raised outflow

20 temperatures are synchronous with stronger boundary current flows. These changes

21 occur rapidly in response to changes in wind forcing, suggesting that barotropic

22 dynamics and the response of the bottom Ekman layer are significant. The observed

23 synchronicity indicates that the previously-detected weakening of the export of the

24 colder forms of AABW from the Weddell Sea need not be associated with a reduction

25 in the total flux of AABW exported via this route. These points need careful

26 consideration when attributing the observed AABW warming in the Atlantic, and

27 when determining its contribution to global heat budgets and sea level rise.

28

Introduction

The Weddell Sea hosts the production of Weddell Sea Deep Water (WSDW), the main contributor to the cold Antarctic Bottom Water (AABW) that permeates much of the global ocean abyss [*Orsi et al.*, 1999]. While AABW in the low-latitude Atlantic is often taken as the ensemble of waters colder than 2 °C, WSDW itself has potential temperatures between ~0 and -0.7 °C, and forms directly through intense air-sea-ice interaction at the periphery of the Weddell Sea, as well as via upwelling of Weddell Sea Bottom Water [e.g. *Gordon et al.*, 2001; *Meredith et al.*, 2000]. It circulates cyclonically within the boundary currents of the Weddell gyre before being exported across and around the South Scotia Ridge (Figure 1), with Orkney Passage (OP) being the most direct throughflow for WSDW to enter the Atlantic overturning circulation [*Naveira Garabato et al.*, 2002; *Schodlok et al.*, 2002]. Around one quarter of all dense Antarctic-derived waters colder than 0 °C flow through OP, after which they flow north and east across the Scotia Sea [*Meredith et al.*, 2001; *Meredith et al.*, 2008] and westward in a boundary current toward Drake Passage. They do not, however, penetrate significantly into the Pacific [*Nowlin and Zenk*, 1988].

In recent decades, AABW has warmed significantly in many regions, most strongly in the Atlantic [e.g. *Meredith et al.*, 2008; *Zenk and Morozov*, 2007]. This warming has reached the North Atlantic, where it was interpreted as a slowing in the meridional overturning of Antarctic-derived waters [*Johnson et al.*, 2008]. The warming has a magnitude of potential significance for the global heat budget and calculations of sea level rise [*Purkey and Johnson*, 2010].

53   There have been a number of investigations of variability of deep and bottom waters

54   in the Weddell Sea [e.g. *Fahrbach et al.*, 2004; *Schröder et al.*, 2002], however the

55   absence of clearly-defined trends in WSDW properties has focused attention on

56   processes that might control its characteristics as it exits the Weddell Sea. Based on

57   hydrographic data from the Scotia Sea, *Meredith et al.* [2001; 2008] proposed that

58   changes in cyclonic wind-forcing over the Weddell gyre could impact on the

59   steepness of isopycnal surfaces in the northern Weddell Sea, and hence on the

60   temperature of the coldest water that crosses the South Scotia Ridge. *Jullion et al.*

61   [2010] used a more comprehensive data set, and found a surprisingly short lag (just a

62   few months) between changes in wind stress curl over the Weddell gyre and

63   temperatures of WSDW exported through OP. If such processes were relevant on

64   decadal timescales, the observed strengthening of winds over the Southern Ocean

65   [e.g. *Thompson and Solomon*, 2002] could plausibly be responsible for the warming

66   of AABW in the Atlantic, due to the coldest classes of WSDW being progressively

67   restricted from crossing the South Scotia Ridge [*Meredith et al.*, 2008]. Here we

68   present observational evidence and dynamical arguments concerning local controls on

69   both the temperature and strength of the outflow of WSDW from the Weddell Sea,

70   and discuss the implications for large-scale water mass properties and ocean climate.

71

72   **Data sources**

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74   Data were obtained from two *in situ* instruments, one in the northern Weddell Sea,

75   and one in the southern Scotia Sea (Figure 1). The former was the uppermost

76   instrument on mooring M2 of the Consortium on the Ocean's Role in Climate: Abrupt

77   Climate Changes Studies (CORC-ARCHES) program [e.g. *Gordon et al.*, 2010]. M2

78   was deployed at 62°38'S, 43°15'W in water depth ~3059 m, with the uppermost

79   instrument (a Sea-Bird SBE37 Microcat temperature/conductivity recorder) located

80   2580 m below the surface. The latter was the Multi-Year Recording Tide Level

81   Equipment [MYRTLE; *Spencer and Foden*, 1996], a long-duration bottom pressure

82   recorder deployed on the seabed at 60°03'S, 47°10'W (~2350 m depth) between late

83   1999 and late 2003. The MYRTLE temperature sensor was not calibrated for absolute

84   temperature, so only anomalies are used; these are precise to 0.001 °C. We also use

85   wind stress data from the ERA-Interim reanalysis

86   (http://www.ecmwf.int/products/data/archive/descriptions/ei/index.html).

87

88   **Results**

89

90   MYRTLE and M2 temperature anomalies correlate well for much of their length with

91   a lag of 120 days (M2 leading MYRTLE; Figure 2a), specifically during 2002 and

92   2003 for which the correlation at this lag (0.6) is significant at the 95% level. (Note

93   that MYRTLE is scaled by 0.7 in Figure 2. M2 data were obtained from the

94   uppermost instrument on a long mooring, whilst MYRTLE data were obtained with a

95   bottom lander and hence came from closer to the core of the bottom-intensified

96   boundary current). For a path length between the two sites of ~500 km (Figure 1), this

97   120 day lag equates to a flow speed of ~4.8 cm s^{-1}, somewhat less than the 7.9 cm s^{-1}

98   quoted by *Nowlin and Zenk* [1988] for boundary current speeds at ~2000 m depth to

99   the west of MYRTLE. However, considering the full period of data overlap suggests

100   that ~4.8 cm s^{-1} may underestimate the true long-term mean. Specifically, during

101   2001, the lag of 120 days shows poor agreement between MYRTLE and M2 (Figure

102   2a), whereas 60 days yields significantly better agreement. (For this section of data, a

103   120 day lag gives a correlation of 0.2, compared with 0.55 for 60 days). This shorter

104   lag equates to a flow speed of ~9.6 cm s^{-1}. For comparison, long-term current meter

105   records at M2 give a mean speed of 8.3 cm s^{-1} [*Gordon et al.*, 2010], though no such

106   data are available from M2 for 2001 itself.

107

108   Significantly, temperature anomalies during the period of shorter-lag correlation

109   (2001) were higher than those during the period of longer-lag correlation (2002 and

110   2003) (Figure 2). The length of the transition between these two periods is hard to

111   quantify precisely, but cannot exceed a few months (Figure 2).

112

113   The depths of M2 and MYRTLE data were different, at ~2580 m and 2350 m

114   respectively. Recent multi-beam echo sounder data (not shown) give a sill depth of

115   ~3600 m for the narrowest part of OP, and depths ~2500 to ~3350m along a shallower

116   ridge immediately upstream. This indicates that WSDW passing the top of M2 can

117   cross the South Scotia Ridge relatively unimpeded, although it will sink and mix as it

118   does so. That the M2 and MYRTLE temperature anomalies show strong similarities

119   indicates vertical coherence in temperature, given this sinking and mixing and the

120   shallower location of MYRTLE compared with M2.

121

122   **Discussion**

123

124   Coincident with the period of shorter-lag correlation between MYRTLE and M2,

125   zonal wind stress along the South Scotia Ridge was anomalously strong during the

126   latter part of 2001 (Figure 2f, 2g, 3), with the end of this period matching closely the

127   transition to the longer-lag relationship between MYRTLE and M2 (Figure 2a, 2c).

128   This echoes the timescale of just a few months for the response of WSDW to winds

129   found previously [*Jullion et al.*, 2010]. Given the rapid and synchronous response of

130   WSDW temperature and speed to changes in forcing, we suggest that the key

131   mechanism involved is the response of the bottom Ekman layer to barotropic changes

132   in the strength of the deep boundary current that flows along the sloping topography

133   of the northern Weddell Sea.

134

135   As described by *Garrett et al.* [1993] and *Brink and Lentz* [2010] amongst others, any

136   along-slope geostrophic flow over sloping topography gives rise to a bottom Ekman

137   layer that displaces isopycnals away from their equilibrium depth and that is

138   ultimately arrested by buoyancy forces. The arrest time for a downwelling-favorable

139   along-slope geostrophic flow such as the deep boundary current in the northern

140   Weddell Sea (or, in other words, the time scale on which a change in the along-slope

141   geostrophic velocity is balanced by an adjustment of the bottom Ekman layer) is

142   estimated by $t_{arrest} = 0.5 \, c_d^{-1} \, N^{-1} \, S^{-3/2}$, where c_d is the frictional bottom drag

143   coefficient, N is the buoyancy frequency, and $S = N^2 f^{-2} \sin^2\theta$ is the Burger number

144   based on the bottom slope (tan θ) and the inertial frequency f [*Garrett et al.*, 1993].

145   Using values typical for the northern Weddell Sea boundary current ($c_d \approx 2.5 \times 10^{-3}$, N

146    $\approx 10^{-3}$ s^{-1}, $f \approx 10^{-4}$ s^{-1}, $\theta \approx 2°$, $S \approx 0.12$), we obtain $t_{arrest} \sim 54$ days, broadly consistent

147   with the baroclinic adjustment time scale suggested by the observations.

148

149   That the bottom Ekman velocity anomaly associated with the adjustment process is

150   sufficiently large to produce the observed change in WSDW temperature may be

151   shown using the relationship between the Ekman layer velocity v_{Ek} and the along-

152   slope geostrophic speed U_g, i.e. $v_{Ek} = 2.5 \, c_d^{1/2} \, U_g$, where the reduction in the Ekman

153   flow resulting from buoyancy effects has been neglected for simplicity [e.g., *Brink*

154   *and Lentz,* 2010]. An approximate halving of the boundary current speed from ~0.1 m

155   s^{-1} in the last 9 months of 2001 to ~0.05 m s^{-1} thereafter is suggested by observations,

156   implying a reduction in v_{Ek} of ~6 × 10^{-3} m s^{-1}. If such a reduced Ekman flow acted

157   adiabatically on a topographic slope $\theta \approx 2°$ over a time scale $t_{arrest} \sim 54$ days, the

158   implied uplift of isopycnals would be ~900 m, well in excess of the ~100 m suggested

159   by observations. In practice, this isopycnal displacement is a gross overestimate, due

160   to the assumption that the Ekman flow on the slope proceeds unimpeded by buoyancy

161   forces, and serves simply to quantitatively illustrate the plausibility of the proposed

162   mechanism of baroclinic adjustment.

163

164   Other potential mechanisms do not seem able to explain our rapid synchronous

165   changes in WSDW outflow temperature and strength. For example, at typical rates of

166   O(10 m y^{-1}), local forcing by anomalous surface Ekman pumping / suction would take

167   several years to displace isopycnals vertically by ~100 m, the amount required to

168   change the temperature of the coldest WSDW exported by the ~0.04°C observed [e.g.

169   *Gordon et al.,* 2001]. Further, changes in the formation properties of WSDW such as

170   have been inferred [e.g. *Fahrbach et al.*, 2004] are unlikely to have the rapid

171   timescale that characterize the shifts in both temperature and flow speed that we have

172   observed, nor are they likely to have such a short-period relationship to changes in

173   local atmospheric forcing.

174

175   We hypothesized previously that large-scale cyclonic wind forcing over the extent of

176   the Weddell gyre may control the northward export of WSDW across the South

177   Scotia Ridge [*Meredith et al.*, 2001; 2008]. Our new observations do not preclude

178   this, however it is important to note that classical gyre-wide baroclinic adjustment at

179   high latitudes is characterized by time scales of decades [e.g., *Anderson and Gill,*

180   1975], which would result in a significantly longer response timescale than that

181   observed here. We thus suggest that any comparatively slow changes in outflow due

182   to changes in large-scale cyclonic forcing will have more rapid (and potentially

183   larger) changes superposed on them due to barotropic and Ekman processes near the

184   outflow itself.

185

186   **Conclusions**

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188   AABW in the Atlantic is warming rapidly, and determining the causes of this is

189   important if the implications for global heat budgets and sea level rise are to be

190   correctly quantified and understood. If, as has been suggested previously, the cause of

191   this warming were a reduction in the northward export of the colder classes of

192   WSDW (as opposed to the same volume being exported with higher temperatures)

193   such calculations will need to account for this to avoid regional overestimates of heat

194   content and thermal expansion. Our findings support the notion of the colder classes

195   of WSDW being progressively restrained from leaving from the Weddell Sea,

196   however the synchronous nature of the temperature and export speed of WSDW in

197   response to changing winds adds significant further complexity. Specifically, it now

198   seems possible that a wind-induced reduction in the export of the colder classes of

199   AABW into the Atlantic overturning circulation can be associated with an increase in

200   the export of the warmer classes. The balance between these factors in controlling the

201   overall export flux is presently unknown.

202

203   Similar to the AABW in the North Atlantic [*Johnson et al.*, 2008], bottom waters in

204   the North Pacific have also warmed recently [*Fukasawa et al.*, 2003]. The rapid

205   timescale of this warming (order of four decades) was likely accomplished by a fast

206   teleconnection facilitated by internal Rossby and Kelvin waves, which have the action

207   of slowing the deep currents that transport AABW northwards. If true also in the

208   Atlantic, with the perturbations that generated the wave teleconnection being changes

209   in the WSDW outflow of the type described here, the superposition of such fast wave-

210   mediated signals upon slower advective changes would generate significant

211   complexity, and make attribution of measured deep ocean climate change even more

212   challenging. This is compounded by the fact that the outflow processes outlined here

213   are not represented in current climate-scale ocean models.

214

215   The changes in WSDW observed here were caused by changes in winds that project

216   strongly onto the pattern of the Southern Annular Mode (SAM), consistent with

217   *Jullion et al.* [2010]. The SAM has moved to a higher-index state in recent decades,

218   associated with anthropogenic forcing from ozone depletion and/or greenhouse gases,

219   resulting in stronger winds over the Southern Ocean [*Marshall*, 2003; *Thompson and*

220   *Solomon*, 2002]. To the extent that the mechanisms described here are relevant on

221   decadal timescales, this carries the implication that anthropogenic processes may be

222   significant in the abyssal warming along the length of the Atlantic. If the winds over

223   the Southern Ocean continue to strengthen in future years and decades (either via

224   there being a stronger mean, or more frequent events such as that seen in 2001), this

225   could result in warmer classes of WSDW being supplied more efficiently to the lower

226   limb of the global overturning circulation. To detect and correctly attribute such

227   changes, a system capable of monitoring both properties and fluxes is required.

228   Accordingly, we will deploy shortly a comprehensive mooring array in the narrowest

229   part of OP. Sustaining this for several decades will be difficult, but is required to meet

230   the challenges outlined here.

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244   **References**

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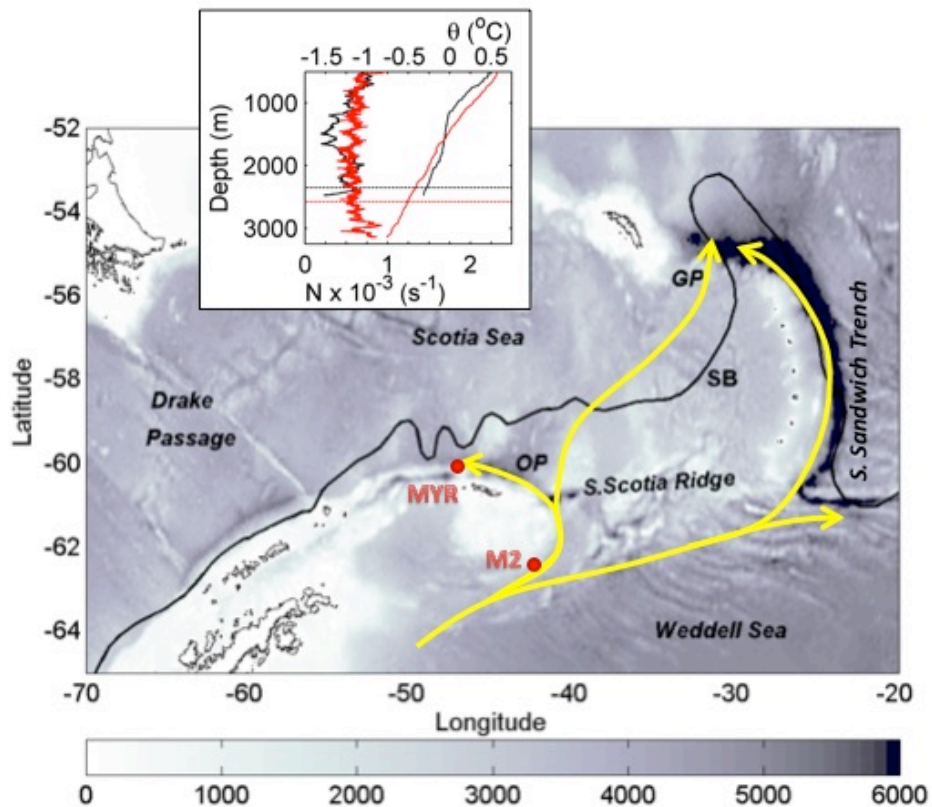
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Figure 1. Locations of M2 and MYRTLE in the northern Weddell Sea and southern

Scotia Sea (red dots). Yellow arrows depict the primary flow paths of WSDW through

the region: Orkney Passage (OP) on the South Scotia Ridge and Georgia Passage

(GP) at the northeastern Scotia Sea are key throughflows. The Southern Boundary

(SB) of the Antarctic Circumpolar Current is marked. Background shading is depth

(m). Inset shows typical potential temperature and buoyancy frequency profiles from

the location of M2 (red) and close to the site of MYRTLE (black). The depths of the

MYRTLE and M2 temperature sensors used here are marked.

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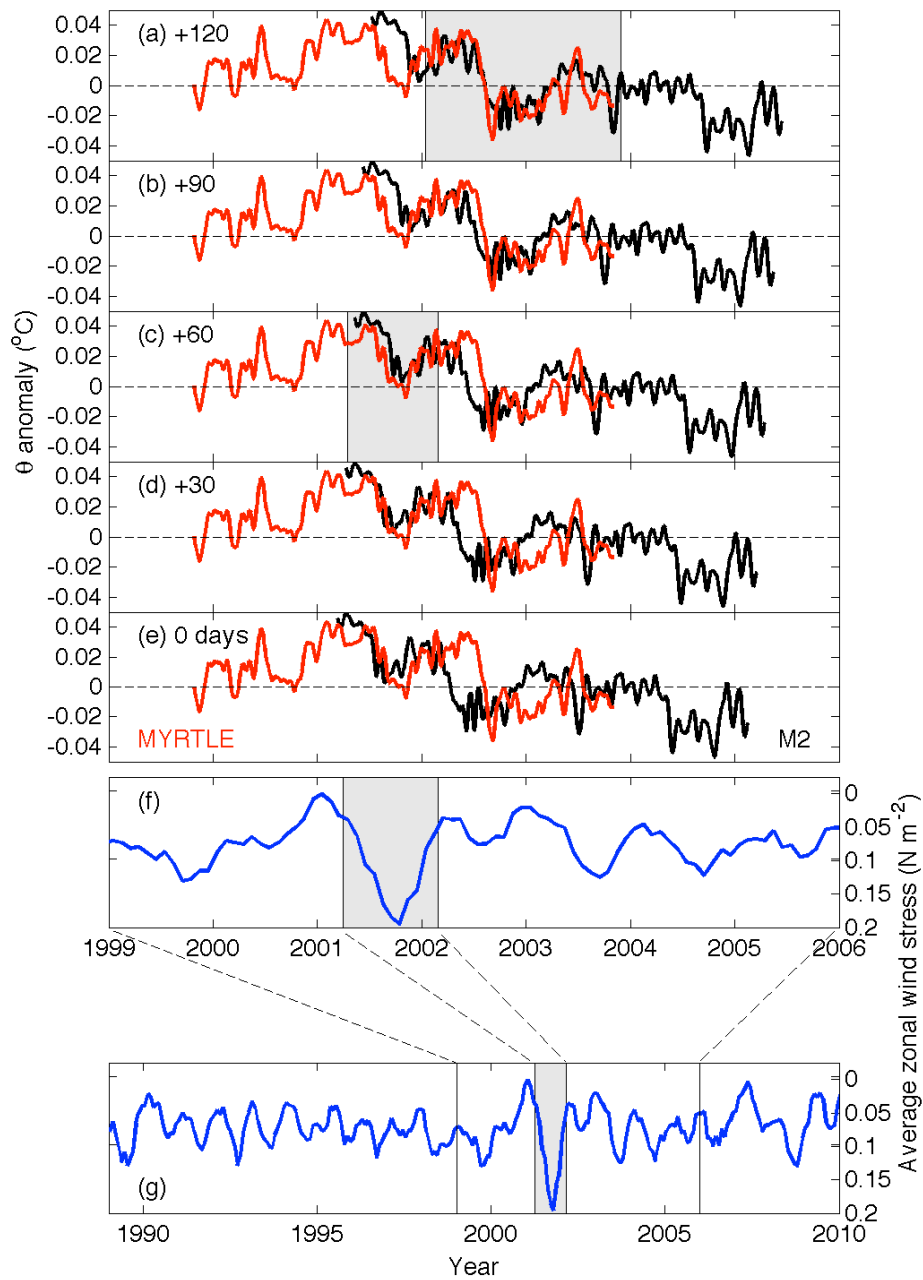
**Figure 2(a-e).** Temperature anomaly at M2 (black) and MYRTLE (red), for lags of 0

to 120 days. MYRTLE and M2 correlate well at 120 days lag, except for the first ~9

months, during which a 60 day lag yields the highest correlation (shaded boxes in (a)

and (c) highlight these periods). **(f)** Zonal wind stress averaged over the northern

Weddell Sea (area of averaging marked in Figure 3a). Curve is a five-point running

average of monthly-mean values, plotted inverted. Shaded box is as per Figure 2c,

314   during which period wind forcing reached a record-length extreme. **(g)** as for (f), but

315   for the full length of ERA-Interim.

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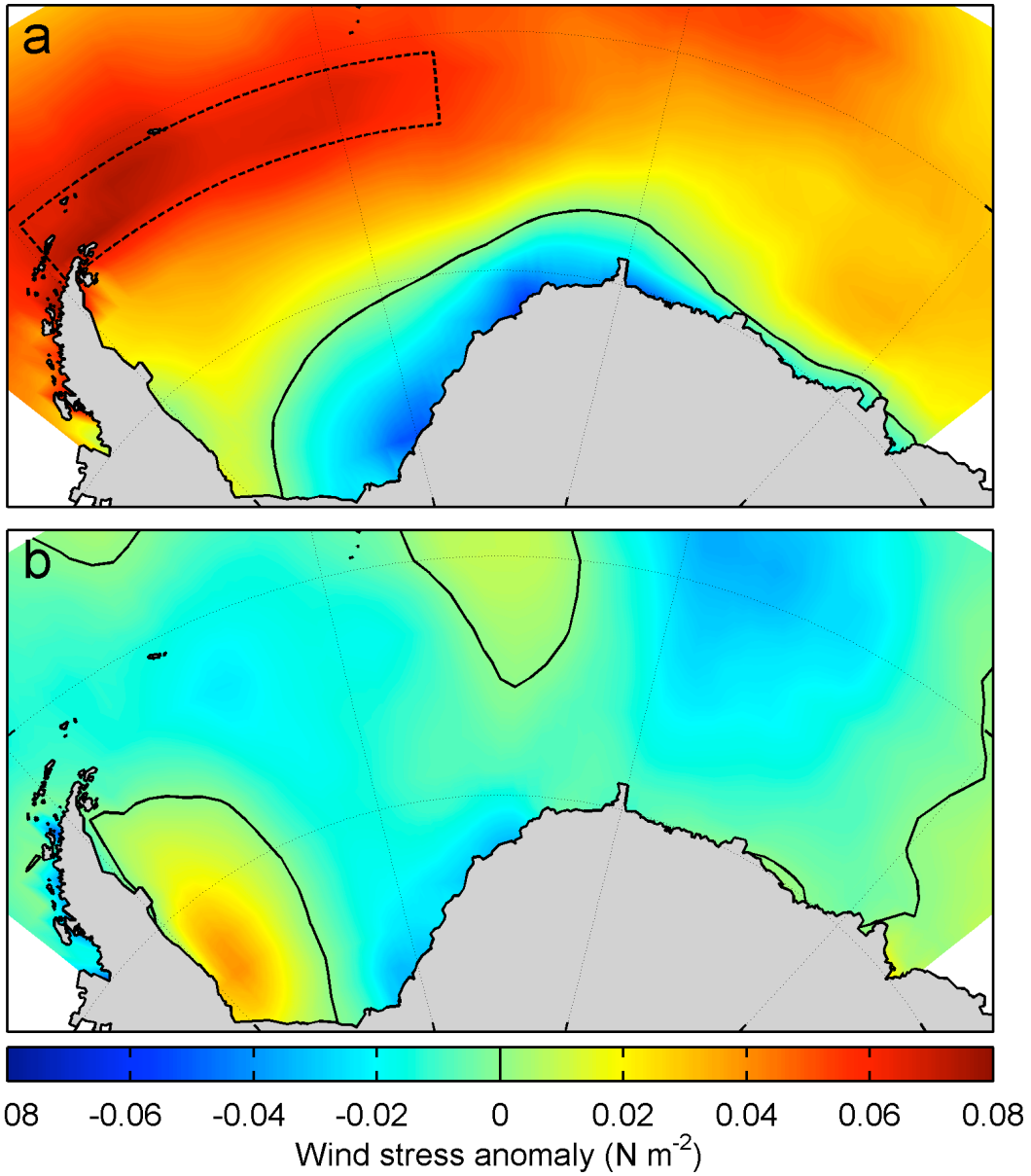
Figure 3 (a) Zonal and **(b)** meridional wind stress anomalies respectively, for the last 9 months of 2001. The zero N m^{-2} contour is plotted. Note the strong anomaly in eastward wind stress at the northern edge of the Weddell Sea during this period.