

# Rapid sea-level rise along the Antarctic margins in response to increased glacial discharge

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**The Antarctic shelf seas are a climatically and ecologically important region, and are at present receiving increasing amounts of freshwater from the melting of the Antarctic Ice Sheet and its fringing ice shelves<sup>1,2</sup>, primarily around the Antarctic Peninsula and the Amudsen Sea. In response, the surface ocean salinity in this region has declined in past decades<sup>3-9</sup>. Here, we assess the effects of the freshwater input on regional sea level using satellite measurements of sea surface height (for months with no sea-ice cover) and a global ocean circulation model. We find that from 1992 to 2011, sea-level rise along the Antarctic coast is at least  $2 \pm 0.8 \text{ mm yr}^{-1}$  greater than the regional mean for the Southern Ocean south of  $50^\circ \text{ S}$ . On the basis of the model simulations, we conclude that this sea-level rise is almost entirely related to steric adjustment, rather than changes in local ocean mass, with a halosteric rise in the upper ocean and thermosteric contributions at depth. We estimate that an excess freshwater input of  $430 \pm 230 \text{ Gt yr}^{-1}$  is required to explain the observed sea-level rise. We conclude that accelerating discharge from the Antarctic Ice Sheet has had a pronounced and widespread impact on the adjacent subpolar seas over the past two decades.**

The Antarctic subpolar seas are a region of intense and complex interactions between the atmosphere, ocean and cryosphere, with an influence on Earth's climate that is greatly disproportionate to their area. Air-sea-ice interactions in these seas are central to the stability of the Antarctic Ice Sheet and global sea level<sup>1,2</sup>, the volume and extent of Antarctic sea ice<sup>10</sup>, the Earth's albedo<sup>11</sup>, and the generation of the Antarctic bottom water (AABW) that cools and ventilates much of the global ocean abyss<sup>12</sup>. The subpolar seas are at present experiencing a significant increase in freshwater discharge from the grounded Antarctic Ice Sheet<sup>1</sup> and its fringing ice shelves<sup>1,2,13</sup>. The current state of knowledge concerning the impact on the adjacent ocean of this rapid change in freshwater forcing is, however, extremely limited, consisting of a few suggestive, yet highly localized and temporally sparse time series of *in situ* hydrographic observations<sup>3-9</sup>. Here, we use multiple lines of evidence (satellite measurements of sea surface height (SSH), *in situ* hydrographic measurements, and results from ocean model simulations) to reveal the local response to the recent Antarctic freshwater imbalance.

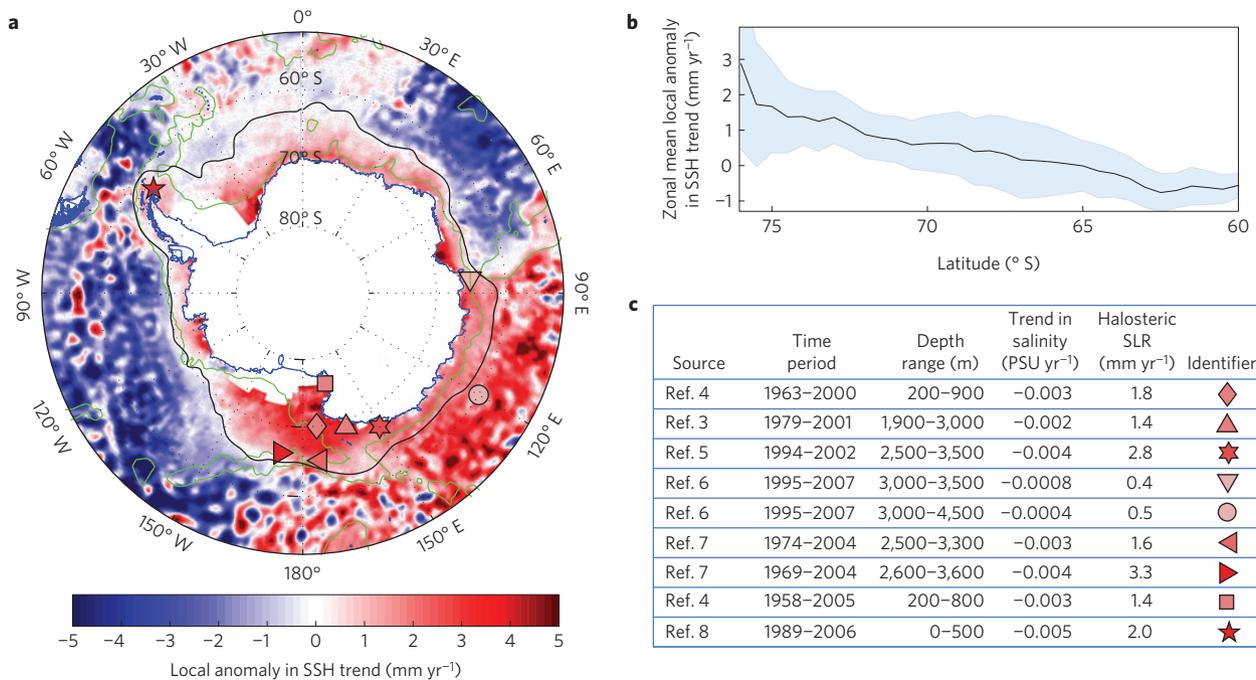
The grounded Antarctic Ice Sheet is at present losing mass overall through increased ice discharge, but gaining mass in places through enhanced snowfall<sup>1,14</sup>. With modest variability in evaporation and precipitation in subpolar waters<sup>15</sup>, the increased discharge is expected to freshen the nearby ocean. Such freshening should be attributable to an 'excess' freshwater discharge above a

baseline rate consistent with a steady ocean salinity. Discharge from the grounded ice sheet increased by  $150 \pm 50 \text{ Gt yr}^{-1}$  between 1992 and 2010 (ref. 14), implying an average of  $75 \pm 25 \text{ Gt yr}^{-1}$  excess discharge. An alternative estimate of the grounded ice contribution can be derived by assuming that the excess discharge is equal to net losses from West Antarctica and the Antarctic Peninsula, which sum to  $85 \pm 30 \text{ Gt yr}^{-1}$  between 1992 and 2011 (ref. 1). Further mass loss is occurring through the thinning of floating ice shelves. Although this mass loss is more uncertain than that of grounded ice, it may be estimated from satellite measurements and modelled surface accumulation, which indicate floating ice thinning of  $280 \pm 50 \text{ Gt yr}^{-1}$  between 2003 and 2008 (refs 16,17) and  $115 \pm 43 \text{ Gt yr}^{-1}$  for 1994–2008 (ref. 13). Finally, a series of large ice-shelf retreats has occurred along the Antarctic Peninsula that is not included in the previous figures, and which averages  $210 \pm 27 \text{ Gt yr}^{-1}$  between 1988 and 2008 (ref. 13). However, it is unclear how much of the freshwater from these breakups was injected into the ocean over the Antarctic subpolar seas, and how much removed to distance by icebergs. All the above estimates represent changes since the early 1990s, but *in situ* measurements suggest that the ocean was already freshening then, so these values constitute a lower bound for the actual excess discharge above a 'steady salinity' rate.

The excess freshwater flux to the Antarctic subpolar seas in the past two decades is estimated hereafter as the sum of mass losses from the thinning of grounded and floating ice,  $\sim 350 \pm 100 \text{ Gt yr}^{-1}$ . The bulk of this discharge is focussed around the Antarctic Peninsula and the Amudsen Sea. This excess freshwater input is anticipated to freshen the Antarctic subpolar seas, and to raise regional sea level through both steric (density-induced) and barostatic (mass-induced) effects. Consistent with this, the few available time series of *in situ* hydrographic measurements, collated in Fig. 1, suggest that Antarctic subpolar waters have undergone a marked freshening (by an order of 0.01 per decade) in recent decades<sup>3-9</sup>. An important limitation of these observations is their strong spatial bias to the Ross Sea, where tracer analyses suggest the implication of glacial meltwater in inducing the local freshening<sup>3</sup>. If the freshening is as widespread as suggested by the very sparse *in situ* measurements, and if the increase in Antarctic freshwater discharge is indeed the causal factor, we expect sea-level rise to be especially pronounced across the Antarctic subpolar seas, and to occur at a rate commensurate with the increase in freshwater input modulated by ocean dynamics.

To test this, we examine the evolution of sea level around Antarctica over the past two decades using satellite measurements of SSH. The primary data analysed are gridded maps of sea-level

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**Figure 1 | Regional anomaly in summer (January–April) sea-level trend, 1992–2011.** The anomaly is calculated relative to the full (barystatic and steric) global-mean rate of sea-level rise for summer months. **a**, Circumpolar view, showing the northern boundary of the sea-level anomaly in black. Markers indicate *in situ* estimates of interdecadal freshening, shaded by the magnitude of the corresponding halosteric sea-level rise. The information for each marker is given in the table in **c**. The 3,000 m isobath is shown in green. **b**, Zonal-mean regional sea-level rise. Shading highlights the 2σ zonal variation.

anomaly (MSLA) generated by AVISO (Archiving, Validation, and Interpretation of Satellite Oceanographic Data) for 1992–2011 (ref. 18; analysis methods and uncertainties are discussed in detail in Supplementary Methods). Satellite-derived measurements of sea level cannot be readily obtained in the presence of sea ice, so our analysis focuses on the largely ice-free summer months (January–April). Using data from these months, the linear trend in SSH was derived and the global-mean rate of sea-level rise for summer months (~3.2 mm yr<sup>-1</sup> between 1992 and 2012) subtracted to reveal the local anomaly (Fig. 1).

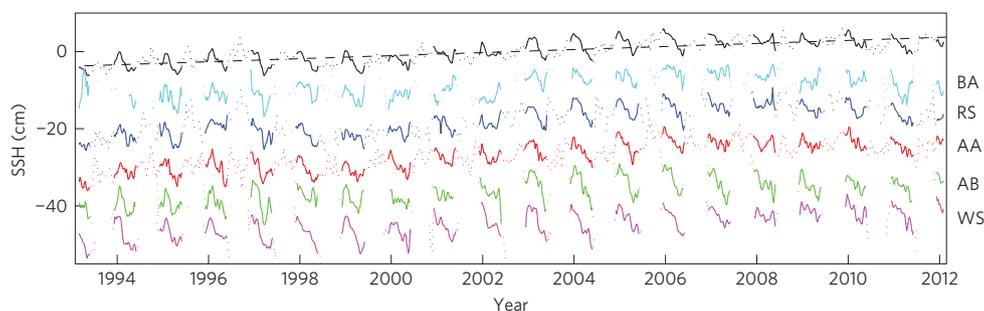
Over the mid-latitude Southern Ocean, the sea-level rise anomaly varies zonally, with alternating sign (Fig. 1a). This pattern arises from the superposition of sea-level impacts caused by various large-scale modes of atmospheric variability<sup>19</sup>. At high latitudes (south of ~62° S), however, our analysis reveals a circumpolar, topographically influenced signal of anomalously rapid sea-level rise that has not been observed previously (Fig. 1b), occurring at 1–5 mm yr<sup>-1</sup> above the global mean, with local peaks in the Ross Sea and Prydz Bay. The northern boundary of this rapid sea-level rise is identified here as the line where the SSH trend anomaly first changes sign or reaches a minimum with increasing distance from Antarctica (Fig. 1a). The mean sea-level rise south of this boundary is at least 2 ± 0.8 mm yr<sup>-1</sup> above the regional mean south of 50° S (1.2 ± 1.5 mm yr<sup>-1</sup> above the global-mean sea-level rise). Although the signal covers the broad Antarctic subpolar seas, the signal is most significant over the continental shelves, which are our primary focus.

Although its statistical significance appears modest, the above quantification is a highly conservative estimate of the regional sea-level rise anomaly induced by freshwater forcing. This is because the global-mean rate of sea-level rise contains a large thermosteric contribution from the low- and mid-latitude oceans that is unrelated to polar processes<sup>20,21</sup>. A more appropriate approach to isolating the local effect of Antarctic freshwater discharge would entail the subtraction of the global-mean rate of barystatic sea-level rise from the measured SSH trend (Supplementary Methods). This rate is

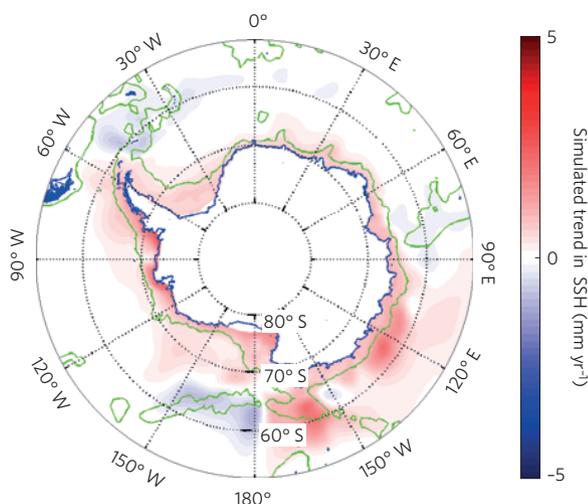
unlikely to exceed 1.5 mm yr<sup>-1</sup> (refs 16,20,21), resulting in a mean sea-level rise across the Antarctic subpolar seas of 2.8 ± 1.5 mm yr<sup>-1</sup> above the global barystatic mean. However, the global-mean rate of barystatic sea-level rise has substantial uncertainty, so hereafter we consider only the more conservative estimate relative to the global-mean rate of sea-level rise.

The temporal progression and regional distribution of SSH change across the Antarctic subpolar seas (Fig. 2) reveal several important features. For example, SSH displays a pronounced seasonal cycle that is most likely forced by seasonal fluctuations in upper-ocean temperature and salinity<sup>22</sup>. Although this seasonal cycle is larger than the interdecadal sea-level rise anomaly, it is distinct from the latter: the linear trend in SSH anomaly affects all stages of the seasonal cycle (Supplementary Methods).

To assess whether the enhanced rate of sea-level rise measured across the Antarctic subpolar seas is consistent with forcing by the recent acceleration in glacial discharge from Antarctica, we consider three distinct lines of evidence. First, we use a global ocean circulation model forced with realistic rates of Antarctic freshwater discharge to simulate the regional response to increased discharge (Supplementary Methods). All model experiments produce a striking, circumpolar, steric sea-level rise anomaly across the Antarctic subpolar seas that strongly resembles the altimetric observations, with a subpolar sea-average anomalous rise of 1–5 mm yr<sup>-1</sup> for a freshwater release of ~300 Gt yr<sup>-1</sup> (the approximate excess Antarctic freshwater discharge averaged over the past 20 years) centred in the Amundsen–Bellingshausen sector (Fig. 3). Remarkably, the modelled anomalous SSH signal is comprised of comparable halosteric and thermosteric contributions, with the former being focussed in the upper ocean and the latter at depth. Thus, the model suggests that the directly forced halosteric sea-level rise around Antarctica is amplified by a positive thermosteric feedback. The barystatic contribution of increased Antarctic freshwater discharge to the spatial distribution of the sea-level rise signal is shown to be negligible by the simulations.



**Figure 2 | Time series of sea-level anomaly in the Antarctic subpolar seas, 1992–2011.** Dotted lines show the full time series, and solid lines the ice-free summer month record. Black: circumpolar average south of the signal's boundary (trend =  $1.2 \text{ mm yr}^{-1}$ ); light blue: Bellingshausen and Amundsen seas (BA;  $135\text{--}60^\circ \text{ W}$ ; trend =  $0.2 \text{ mm yr}^{-1}$ ); dark blue: Ross Sea (RS;  $130^\circ \text{ E}\text{--}135^\circ \text{ W}$ ; trend =  $1.3 \text{ mm yr}^{-1}$ ); red: Australian-Antarctic basin (AA;  $50\text{--}130^\circ \text{ E}$ ; trend =  $1.9 \text{ mm yr}^{-1}$ ); green: Amery Basin (AB;  $10\text{--}50^\circ \text{ E}$ ; trend =  $1.0 \text{ mm yr}^{-1}$ ); pink: Weddell Sea (WS;  $60^\circ \text{ W}\text{--}10^\circ \text{ E}$ ; trend =  $0.5 \text{ mm yr}^{-1}$ ). The dashed black line indicates the linear trend fitted to the circumpolar average sea level.



**Figure 3 | Ocean model simulation of the regional anomaly in sea-level trend, 1992–2007.** This is generated by subtracting a control run with a 'background' Antarctic freshwater forcing from an experimental run perturbed by an excess Antarctic freshwater runoff of  $300 \text{ Gt yr}^{-1}$  (Supplementary Methods). The 3,000 m isobath is indicated in green.

Second, we quantitatively compare the altimetric results with recent observational estimates of steric sea-level rise around Antarctica. The altimetric rates of Antarctic coastal sea-level rise anomaly are found to be in broad agreement with (slightly exceeding) the halosteric sea-level rise contribution of  $\sim 0.5\text{--}3 \text{ mm yr}^{-1}$  implied by the available *in situ* measurements of interdecadal upper-ocean freshening around Antarctica<sup>3–9</sup> (Fig. 1 and Supplementary Methods), in line with model predictions of an important upper-ocean halosteric contribution to the anomalous SSH signal. Similarly, Southern Ocean deep and bottom waters have warmed significantly in the period of our study, inducing a thermosteric sea-level rise of  $\sim 1 \text{ mm yr}^{-1}$  (ref. 23) that is comparable to the signal discussed here. Although the spatial footprint of the deep thermosteric change extends well beyond the Antarctic subpolar seas<sup>23</sup>, in poor agreement with our observed signal and model results, the lack of spatial correspondence between thermosteric effects and regional sea-level trends may relate to other factors, such as changes in wind forcing or self-gravitation (Supplementary Methods). Thus, the existence of a significant contribution of deep-ocean thermosteric adjustment to the observed Antarctic coastal sea-level rise does not conflict with available observations.

Third, if it is assumed, on the basis of the preceding modelling and observational evidence, that the Antarctic coastal sea-level

rise signal is partitioned approximately equally between a directly forced halosteric component and a positive thermosteric feedback, the excess freshwater input required to explain the measured signal may be estimated. This involves multiplying half the linear trend in ocean volume inside the signal's boundary ( $11.6 \text{ km}^3 \text{ yr}^{-1}$ ;  $1.4 \text{ mm yr}^{-1}$ ) by a modified 'Munk multiplier' ( $36.7$ ; ref. 24; see Supplementary Methods), indicating a requirement of  $430 \pm 230 \text{ Gt yr}^{-1}$  of excess freshwater input above the nominal rate needed to maintain a steady ocean salinity. This agrees with the increase in Antarctic melt observed in the past two decades ( $\sim 350 \pm 100 \text{ Gt yr}^{-1}$ ), and lends support to our initial hypothesis that the recent imbalance in the Antarctic cryosphere is driving pronounced and widespread changes in the salinity of the high-latitude Southern Ocean.

Finally, we note that the observed anomalous SSH signal may also be influenced by several tertiary forcing mechanisms, which may account for at most  $\sim 10\%$  of the signal and are discussed comprehensively in Supplementary Methods. The most significant of these is the ocean's barystatic response to wind forcing. The gravitational effect of Antarctic ice-mass loss reduces relative sea-level rise in the Amundsen and Bellingshausen seas by  $\sim 1 \text{ mm yr}^{-1}$  (ref. 25). Other mechanisms, such as upper-ocean warming, precipitation-induced freshening or the ocean's barystatic response to the acceleration in Antarctic freshwater discharge, were found to be insignificant.

In summary, austral summer satellite altimetry measurements show a pronounced circumpolar rise in sea level across the Antarctic subpolar seas that significantly exceeds the global mean. The trend contains a significant halosteric contribution that originates in the increasing discharge of freshwater from Antarctica. Thermosteric sea-level rise from the observed warming of the deep Southern Ocean, which has itself been linked to the freshening of the shelf waters ventilating AABW (refs 23,26; see also Supplementary Methods), also contributes to the signal. Our findings therefore reveal that the accelerating discharge from the Antarctic ice sheet has had a pronounced and widespread impact on the adjacent subpolar seas over the past two decades, and indicate that a significant climatic perturbation to the cryospheric forcing of the Southern Ocean is under way. Given the key dependence of the Southern Ocean on freshwater forcing, this perturbation has major implications for the region's stratification, circulation and important biogeochemical and ecological processes<sup>26–29</sup>.

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## References

1. Shepherd, A. *et al.* A reconciled estimate of ice-sheet mass balance. *Science* **338**, 1183–1189 (2012).

2. Shepherd, A., Wingham, D. & Rignot, E. Warm ocean is eroding West Antarctic Ice Sheet. *Geophys. Res. Lett.* **31**, L23402 (2004).
3. Jacobs, S. S., Giulivi, C. F. & Mele, P. A. Freshening of the Ross Sea during the late 20th century. *Science* **297**, 386–389 (2002).
4. Jacobs, S. S. & Giulivi, C. F. Large multidecadal salinity trends near the Pacific–Antarctic continental margin. *J. Clim.* **23**, 4508–4524 (2010).
5. Aoki, S., Rintoul, S. R., Ushio, S., Watanabe, S. & Bindoff, N. L. Freshening of the Adélie Land bottom water near 140° E. *Geophys. Res. Lett.* **32**, L23601 (2005).
6. Johnson, G. C., Purkey, S. G. & Bullister, J. L. Warming and freshening in the abyssal Southeastern Indian Ocean\*. *J. Clim.* **21**, 5351–5363 (2008).
7. Ozaki, H., Obata, H., Naganobu, M. & Gamo, T. Long-term bottom water warming in the north Ross Sea. *J. Oceanogr.* **65**, 235–244 (2009).
8. Hellmer, H. H., Huhn, O., Gomis, D. & Timmermann, R. On the freshening of the northwestern Weddell Sea continental shelf. *Ocean Sci.* **7**, 305–316 (2011).
9. Williams, G. D. *et al.* Antarctic bottom water from the Adélie and George V Land coast, East Antarctica (140–149° E). *J. Geophys. Res.* **115**, C04027 (2010).
10. Parkinson, C. L. & Cavalieri, D. J. Antarctic sea ice variability and trends, 1979–2010. *Cryosphere* **6**, 871–880 (2012).
11. Meehl, G. A. & Washington, W. M. CO<sub>2</sub> climate sensitivity and snow-sea-ice albedo parameterization in an atmospheric GCM coupled to a mixed-layer ocean model. *Climatic Change* **16**, 283–306 (1990).
12. Orsi, A. H., Johnson, G. C. & Bullister, J. L. Circulation, mixing, and production of Antarctic bottom water. *Prog. Oceanogr.* **43**, 55–109 (1999).
13. Shepherd, A. *et al.* Recent loss of floating ice and the consequent sea level contribution. *Geophys. Res. Lett.* **37**, L13503 (2010).
14. Rignot, E., Velicogna, I., van den Broeke, M. R., Monaghan, A. & Lenaerts, J. T. M. Acceleration of the contribution of the Greenland and Antarctic ice sheets to sea level rise. *Geophys. Res. Lett.* **38**, L05503 (2011).
15. Bromwich, D. H., Nicolas, J. P. & Monaghan, A. J. An assessment of precipitation changes over Antarctica and the Southern Ocean since 1989 in contemporary global reanalyses. *J. Clim.* **24**, 4189–4209 (2011).
16. Depoorter, M. A. *et al.* Calving fluxes and basal melt rates of Antarctic ice shelves. *Nature* **502**, 89–92 (2013).
17. Rignot, E., Jacobs, S., Mouginot, J. & Scheuch, B. Ice-shelf melting around Antarctica. *Science* **341**, 266–270 (2013).
18. Le Traon, P. Y., Nadal, F. & Ducet, N. An improved mapping method of multisatellite altimeter data. *J. Atmos. Ocean. Technol.* **15**, 522–534 (1998).
19. Spence, P., Fyfe, J. C., Montenegro, A. & Weaver, A. J. Southern Ocean response to strengthening winds in an eddy-permitting global climate model. *J. Clim.* **23**, 5332–5343 (2010).
20. Leuliette, E. W. & Miller, L. Closing the sea level rise budget with altimetry, Argo, and GRACE. *Geophys. Res. Lett.* **36**, L04608 (2009).
21. Gregory, J. M. *et al.* Twentieth-century global-mean sea level rise: Is the whole greater than the sum of the parts? *J. Clim.* **26**, 4476–4499 (2013).
22. Meredith, M. P., Renfrew, I. A., Clarke, A., King, J. C. & Brandon, M. A. Impact of the 1997/98 ENSO on upper ocean characteristics in Marguerite Bay, western Antarctic Peninsula. *J. Geophys. Res.* **109**, C09013 (2004).
23. Purkey, S. G. & Johnson, G. C. Antarctic bottom water warming and freshening: Contributions to sea level rise, ocean freshwater budgets, and global heat gain. *J. Clim.* **26**, 6105–6122 (2013).
24. Munk, W. Ocean freshening, sea level rising. *Science* **300**, 2041–2043 (2003).
25. Riva, R. E. M., Bamber, J. L., Lavallée, D. A. & Wouters, B. Sea-level fingerprint of continental water and ice mass change from GRACE. *Geophys. Res. Lett.* **37**, L19605 (2010).
26. Purkey, S. G. & Johnson, G. C. Global contraction of Antarctic bottom water between the 1980s and 2000s. *J. Clim.* **25**, 5830–5844 (2012).
27. Dierssen, H. M., Smith, R. C. & Vernet, M. Glacial meltwater dynamics in coastal waters west of the Antarctic Peninsula. *Proc. Natl Acad. Sci. USA* **99**, 1790–1795 (2002).
28. Boyd, P. W. & Ellwood, M. J. The biogeochemical cycle of iron in the ocean. *Nature Geosci.* **3**, 675–682 (2010).
29. Stammerjohn, S. E., Martinson, D. G., Smith, R. C. & Ianuzzi, R. A. Sea ice in the western Antarctic Peninsula region: Spatio-temporal variability from ecological and climate change perspectives. *Deep-Sea Res. II* **55**, 2041–2058 (2008).

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### Author contributions

C.D.R. conducted the data analysis, with regular input from A.C.N.G., A.J.G.N., C.W.H., M.P.M. and P.R.H. All these authors contributed to the writing of the manuscript. A.C.C. and D.J.W. assisted with the modelling component of the work.

### Additional information

The altimetry data used in the study is available via the AVISO website ([www.aviso.oceanobs.com/en/](http://www.aviso.oceanobs.com/en/)) as well as the MyOcean website ([www.myocean.eu.org/](http://www.myocean.eu.org/)). The reanalysis wind data is available via the ECMWF website ([www.ecmwf.int/](http://www.ecmwf.int/)). Supplementary information is available in the online version of the paper. Reprints and permissions information is available online at [www.nature.com/reprints](http://www.nature.com/reprints). Correspondence and requests for materials should be addressed to C.D.R.

### Competing financial interests

The authors declare no competing financial interests.

# Rapid sea level rise along the Antarctic margins in response to increased glacial discharge

## 6 Uncertainty in SSH measurements and trends

7

8 The measurement of SSH from satellite altimeters suffers from multiple sources of  
9 uncertainty. These are reviewed extensively elsewhere<sup>30</sup> and are therefore only  
10 summarised here. One of the largest sources of uncertainty in the altimetric data sets of  
11 sea level anomaly is that arising from the frame of reference used to correct for the  
12 satellite's orbit error<sup>30-32</sup>. Preceding estimates of this orbit error for the TOPEX / Jason  
13 satellites are as high as 1.5 mm yr<sup>-1</sup> in the z-direction (31), although this error is  
14 thought to be smaller for more recent data processing. Here we assume an upper limit  
15 of 1 mm yr<sup>-1</sup>. The orbit correction is known to be symmetric along the z-axis. For a z-  
16 translation the error is proportional to the sine of the latitude; therefore, an error of  
17 magnitude 1 mm yr<sup>-1</sup> gives a value of 0.9 mm yr<sup>-1</sup> at 65 °S, and a difference of 0.09  
18 mm yr<sup>-1</sup> between 65°S and 75°S. From this small difference, it is clearly not possible to  
19 produce the narrow band of increased sea level observed (Fig. 1). The lower orbit  
20 heights of ERS and Envisat result in larger orbit errors, particularly in the y-direction,  
21 but the dominant role of TOPEX / Jason observations in the AVISO gridded data  
22 essentially eliminates this error, which has also greatly reduced in recent solutions.  
23 Additionally, the altimetric data north of 62°S have been re-processed with an  
24 improved frame of reference that reduces orbit errors<sup>33</sup>. The sea level anomaly data in  
25 the Antarctic subpolar seas north of 62°S are well correlated (p > 99%) with this  
26 improved data set, and our sea level rise estimates in that region are unchanged.

27

28 Following orbit error, the main sources of uncertainty in SSH trend are the wet  
29 troposphere correction, and the biases applied to link together SSH records from  
30 different altimeters<sup>30</sup>. The wet troposphere error is a function of atmospheric water  
31 content and therefore latitude; it can be as high as 2 mm yr<sup>-1</sup> in the tropics but is  
32 negligible at high latitudes<sup>30</sup>. Further, the biases applied to link together SSH records  
33 from different altimeters<sup>30</sup> contribute an uncertainty for the reference TOPEX / Jason  
34 missions of about 0.15 mm yr<sup>-1</sup>. Other sources of uncertainty, such as instrumental,  
35 meteorological, and tide-related factors contribute less than 0.1 mm yr<sup>-1</sup> (ref. 30).

36

37 Since the geographical variability of some potentially significant sources of uncertainty  
38 (particularly those associated with the satellite orbit calculation and the bias between  
39 different satellite missions) has not been accurately characterized for the AVISO data  
40 set (see e.g., discussion in ref. 30-34), here we adopt a multi-line approach to further  
41 demonstrate the robustness of the signal of anomalously rapid sea level rise in the  
42 Antarctic subpolar seas at the core of this study.

43

44 First, the relative significance (assuming negligible systematic error) of the linear trend  
45 in SSH for each data point in the gridded AVISO data set is estimated with a Patterson  
46 t-test accounting for auto-covariance in the sea level record (Fig. S1). This indicates  
47 that, in general, the Antarctic coastal sea level rise signal is significantly different from  
48 zero with 95% confidence, with some exceptions in areas of weak anomalous sea level  
49 rise in the Amundsen Sea, eastern Ross Sea and Weddell gyre.

51 Second, we show that the Antarctic coastal sea level rise signal is present in the  
52 measurements of individual satellite missions (Fig. S2). To do this, we consider the  
53 along-track SSH data recorded by the ERS-1, ERS-2 and Envisat altimeters, as  
54 obtained from the Centre National d'Etudes Spatiales (CNES)<sup>18</sup>. These data sets have  
55 had the aforementioned corrections applied, but have not been cross-calibrated in the  
56 same way as the merged data set. We select regions of significant anomalous sea level  
57 rise (identified on the basis of Fig. S1) and bin-average single-track measurements for  
58 every intersection of the satellite track with each specified area, typically every 10  
59 days. The resulting SSH time series are illustrated by Fig. S2, which shows the single-  
60 mission sea level records in the western Ross Sea, representative of that in other  
61 regions inspected. The presence of an anomalous sea level rise signal that is both  
62 consistent between different satellites and highly coherent with the gridded AVISO  
63 data set indicates that the Antarctic coastal sea level rise signal does not arise from  
64 uncertainty in the bias between different missions.

65  
66 Finally, we emphasize that the Antarctic coastal sea level rise signal identified in this  
67 study has a spatial footprint that is both distinct from those of any known sources of  
68 uncertainty in altimetric measurements and consistent with expectations from  
69 numerical simulations of the present deglaciation of Antarctica<sup>35-36</sup>. This provides  
70 further endorsement for the inference that the signal is of physical origin. Following  
71 these arguments, the combination of dominant error terms suggests an overall regional  
72 average error of 1-1.5 mm yr<sup>-1</sup>, and an error of 0.2-0.6 mm yr<sup>-1</sup> for the difference  
73 between 65°S and 75°S. In the main text we use the upper bounds of these values.

#### 74 75 76 **Impact of glacial isostatic adjustment**

77  
78 Despite correcting for a number of measurement errors, the sea level anomaly products  
79 do not intrinsically correct for changes in the geoid, of which the dominant component  
80 is glacial isostatic adjustment (GIA). This correction is estimated using output from a  
81 GIA model<sup>37</sup> (Fig. S3) and subtracted from the MSLA fields before analysis. In  
82 addition, an estimate of the more recent changes in the geoid resulting from e.g., mass  
83 loss from the Antarctic Peninsula, is made. The recent geoid trend is estimated by  
84 subtracting the GIA correction from the total GRACE geoid correction<sup>38</sup> (Fig. S3).  
85 This provides an estimate of recent changes to the geoid over the GRACE period  
86 (2003-2011). As the altimetric record is initiated in 1992, the recent geoid correction  
87 cannot be easily compared and thus directly subtracted from the altimetric data. The  
88 recent geoid correction shows a strong negative anomaly with a maximum of ~ -2 mm  
89 yr<sup>-1</sup> in the Amundsen - Bellingshausen region, and a weaker positive anomaly of ~ 0.5  
90 mm yr<sup>-1</sup> between the Weddell Sea and the Amery region. An Antarctic subpolar sea-  
91 mean of this correction yields -0.2 mm yr<sup>-1</sup>, which is subtracted from the circumpolar-  
92 average subpolar sea level anomaly and accounted for in its error budget. A direct  
93 subtraction of this correction would increase the SSH trend anomaly in the Amundsen  
94 - Bellingshausen region, and decrease it in East Antarctica, thereby enhancing the  
95 agreement between observational and model results.

#### 96 97 98 **Impact of temporal aliasing of SSH measurements**

99  
100 Our calculation of the linear trend in summertime SSH over the 1992-2011 period

101 incorporates all data points between January and April not covered by sea ice.  
102 Summertime sea ice cover across the Antarctic subpolar seas is, however, highly  
103 variable in both space (i.e. the regularity of sea ice cover varies with location across  
104 the Antarctic subpolar seas) and time (i.e. at any given location, there may be  
105 substantial intraseasonal and interannual variability in sea ice cover), so that any  
106 estimate of interdecadal SSH change in the region may be affected by aliasing. Indeed,  
107 there are significant regional trends in Antarctic summertime sea ice extent during this  
108 period<sup>3</sup>.

109  
110 In order to show that our results are robust to this aliasing issue, the time series of  
111 circumpolar-mean SSH anomaly south of the oceanic boundary of the Antarctic coastal  
112 sea level rise signal (Fig. 2) is compared with a time series of SSH anomaly averaged  
113 over the ~5% of that region that is ice-free over the entire altimetric record (Fig. S4).  
114 The two time series are in good agreement, and exhibit a high correlation coefficient of  
115 0.8. Further, the ice-free time series confirms the existence of a strong seasonal cycle  
116 in Antarctic coastal sea level, discussed in the main text.

117

118

### 119 **Global-mean sea level rise**

120

121 In this work, the rate of sea level rise in the Antarctic subpolar seas is calculated by  
122 subtracting the rate of global-mean *total* sea level rise,  $\sim 3.2 \text{ mm yr}^{-1}$ , from the linear  
123 trend in SSH. The rate of global-mean *total* sea level rise is computed as an area-  
124 weighted average of the AVISO-gridded sea level anomaly data over austral summer  
125 months only. Global-mean sea level rise results from the combined effect of barystatic  
126 and steric contributions. Barystatic changes in sea level are rapidly communicated  
127 away from their source and so are considered global<sup>36</sup>. Steric changes are less  
128 effectively communicated and remain more regionally confined over the two-decade  
129 time scale relevant to our study. It is unlikely that the global-mean rate of barystatic  
130 sea level rise accounts for more than half of the trend in total sea level (ref. 20-21),  
131 requiring a steric contribution of  $\sim 1.5 \text{ mm yr}^{-1}$  to balance the budget. While in the  
132 tropics the majority of this steric signal is readily accounted for by upper-ocean  
133 warming, the thermal expansion coefficient for seawater decreases in polar regions and  
134 salinity dominates steric sea level changes there. Thus, in seeking to quantify polar  
135 steric anomalies, subtracting the global-mean rate of barystatic (rather than total) sea  
136 level rise is more appropriate (Fig. S5). However, since the global-mean rate of  
137 barystatic sea level rise suffers from substantial uncertainty, in this study we discuss  
138 only the more conservative estimate relative to the global-mean rate of sea level rise.

139

140

### 141 **Estimation of halosteric SSH change from *in situ* observations**

142

143 Our assessment of the halosteric contribution to the observed sea level rise signal  
144 involves the estimation of steric SSH change from several *in situ* observations of  
145 interdecadal ocean freshening available within the region of the signal (Fig. 1). The  
146 estimation of halosteric change in sea level,  $h_s$ , relies on the use of a linear  
147 approximation to the equation of state<sup>39</sup>,

148

$$149 \quad h_s = -\beta \Delta S h_0 \quad (1)$$

150

151 where  $\beta = 7.6 \times 10^{-4}$  is the haline contraction coefficient,  $\Delta S$  is the change in salinity  
152 observed in situ, and  $h_0$  (m) is the depth over which the change in salinity has occurred.  
153

154

155

### Ocean circulation model experiments

156

157 The NEMO (Nucleus for European Modelling of the Ocean; ref. 40) model was used  
158 to assess our physical interpretation of the observed Antarctic coastal sea level rise  
159 signal. Our model configuration has a  $1^\circ$  resolution tri-polar grid (ORCA1). NEMO is  
160 a z-level Boussinesq ocean model (OPA) coupled to a dynamic-thermodynamic sea ice  
161 model (LIM2), and uses a linear free surface. Precipitation and evaporation affect the  
162 model via volume input through the ocean surface, and therefore influence sea surface  
163 height both through volume input and steric forcing. The model has 75 vertical levels,  
164 is forced by CLIVAR/WGOMD Coordinated Ocean-sea ice Reference Experiments  
165 (COREII) atmospheric reanalysis data, and utilises the Gent-McWilliams eddy  
166 parameterisation<sup>41</sup>.

167

168 The model was qualitatively validated by comparison to a Southern Ocean climatology  
169 based on *in situ* measurements, the CSIRO Atlas of Regional Seas (CARS; ref 42;  
170 Figs. S6-S9). Here we compare the NEMO annual-mean fields for 1992 with the  
171 CARS mean fields, which are derived from the last 50 years of measurements. The  
172 data and model agree surprisingly well, most notably in the zonal-mean diagnostics  
173 (Fig. S8-S9). Fig. S7 contrasts the bottom distributions of temperature and salinity in  
174 the model with those in the CARS climatology, and illustrates the model's satisfactory  
175 degree of realism, particularly as regards the salinity and density fields, though bottom  
176 temperatures are too cold in the Pacific sector. Further information on and extensive  
177 validation of the model are provided in ref. 43.

178

179 Here we consider two types of model runs: a standard control run, and several  
180 perturbation runs. The control run is the last 15 years (1992–2007) of the last of four  
181 cycles of the CORE2 forcing dataset from 1948–2007, so has been spun up for 225  
182 years<sup>43</sup>. A time-invariant runoff from Antarctica of  $0.073 \text{ Sv}$ , or  $2200 \text{ Gt yr}^{-1}$  (ref. 41),  
183 is assumed. The perturbation runs are identical to the control run, except for an  
184 additional surface freshwater flux anomaly applied on a  $8^\circ$  by  $2^\circ$  area centred on the  
185 Amundsen – Bellingshausen continental shelves. A total of three perturbation  
186 experiments are conducted, with surface freshwater flux anomalies of  $\sim 300$  (the  
187 measured approximate excess Antarctic freshwater discharge averaged over the last 20  
188 years, see main text),  $550$  and  $900 \text{ Gt yr}^{-1}$ . The SSH and steric anomalies resulting  
189 from the anomalous freshwater forcing in these runs were evaluated by subtracting the  
190 relevant physical fields in the control run from those in the perturbation runs. The  
191 linear trends in SSH, and the halosteric and thermosteric components of sea level  
192 change, were then calculated from the pertinent anomaly fields.

193

194 The key results for the  $300 \text{ Gt yr}^{-1}$  perturbation run are illustrated in Figures 3 and S6.  
195 The simulated linear trend in SSH anomaly agrees well with observations, both in  
196 magnitude and in spatial distribution. This distribution is also consistent with those in  
197 the perturbation experiments with more vigorous freshwater forcings, with the  
198 magnitude of the SSH response scaling approximately linearly with the amplitude of  
199 the freshwater forcing anomaly, echoing the findings of ref. 35–36. In all runs, the  
200 halosteric and thermosteric contributions each account for approximately 50% of the

201 local SSH trend anomaly (i.e. the barystatic response is spatially uniform and therefore  
 202 does not contribute significantly to the local anomaly in the Antarctic subpolar seas). A  
 203 vertical decomposition of the steric constituents reveals that the halosteric change  
 204 occurs primarily in the upper ocean (approximately half of the vertically integrated  
 205 halosteric change is accounted for by the uppermost 300 m), whereas the bulk of the  
 206 themosteric signal occurs in the deep ocean. This result is broadly consistent with the  
 207 analyses of *in situ* observations described in the main text.

208  
 209 While the general endorsement of our physical interpretation of the observed Antarctic  
 210 coastal sea level rise signal by the model experiments is reassuring, two significant  
 211 caveats must be noted. First, there are resolution-related limitations to the model's  
 212 representation of the narrow boundary current surrounding Antarctica, which is  
 213 thought to play a significant role in mediating the transmission to the deep ocean of  
 214 freshwater anomalies on the Antarctic continental shelves. Second, the simulated deep-  
 215 ocean themosteric changes are likely to result from variations in the formation and  
 216 circulation of Antarctic Bottom Water, the production of which NEMO (like most  
 217 ocean circulation models) fails to represent in a realistic fashion.

## 218 219 220 **The Munk multiplier**

221  
 222 Here we use first-principle arguments, assuming a linear equation of state  
 223 approximation, to estimate the amount of freshwater required for a given steric SSH  
 224 signal. This topic is discussed in depth by ref. 24.

225  
 226 The definition of steric height anomaly in pressure coordinates is the integral between  
 227 atmospheric pressure ( $ap$ ) and bottom pressure ( $bp$ ), of the specific volume anomaly,  
 228  $v_f - v_\rho$ :

$$229 \Delta h_s = \int_{ap}^{bp} \frac{v_f - v_\rho}{g} dp, \quad (2)$$

230  
 231 where  $v_s = \frac{1}{\rho}$  is the specific volume of seawater,  $v_f$  is the specific volume of fresh  
 232 water, and  $g$  is gravity.

233  
 234 The mass anomaly associated with a freshwater release in the Antarctic subpolar seas  
 235 is assumed to be locally negligible as it is rapidly communicated to the global ocean,  
 236 establishing a global barystatic equilibrium within  $\sim 14$  days<sup>36</sup>. Thus, as there is  
 237 negligible change in bottom pressure following a freshwater release, equation (2) then  
 238 simplifies to an integral over the freshwater layer (from  $ap$ , to the base of the  
 239 freshwater layer,  $lp$ ).

240  
 241 Further, we note that

$$242 \int_{ap}^{lp} dp = lp - ap = \rho_f g \delta h, \quad (3)$$

243  
 244 and as  $\frac{v_f - v_s}{g}$  is constant (2) becomes

245  
246  
247

248  $\Delta h_s = \rho_f g \delta h \frac{v_f - v_s}{g}$ . (4)

249

250 After some re-arrangement, we find

251

252  $\Delta h_s = \frac{\delta h(\rho_s - \rho_f)}{\rho_s}$ , (5)

253

254 and

255

256  $\Delta h_s = (1 - \frac{\rho_f}{\rho_s})\delta h$ . (6)

257

258 Following equation (6), the change in steric height of the water column is then

259

260  $\Delta h_s \approx \delta h \beta s_s$ . (7)

261

262 Therefore, the amount of freshwater required to produce a given change in steric height  
263 in the Antarctic subpolar seas is

264

265  $\delta h \approx \Delta h_s \frac{1}{\beta} s_s = \Delta h_s \cdot 37.6$ . (8)

266

267 This equation states that for a given freshwater discharge, the resultant steric change in  
268 SSH is 37.6 times smaller than the height of freshwater added.

269

270

### 271 **Tertiary mechanisms of sea level rise**

272

273 In addition to the mechanisms contributing significantly to Antarctic coastal sea level  
274 rise (halosteric adjustment to an acceleration in freshwater discharge from Antarctica,  
275 and thermosteric response to the warming of the deep Southern Ocean), discussed in  
276 the main text, we have assessed the likely importance of other candidate mechanisms  
277 contributing to our observed signal.

278

#### 279 *Thermosteric adjustment to upper-ocean warming*

280

281 A possible forcing of anomalous sea level rise in the Antarctic subpolar seas is a  
282 regional increase in the mean temperature of the ocean. Satellite measurements of sea  
283 surface temperature do not suggest coherent circumpolar warming of the upper-ocean  
284 waters of the subpolar seas over the last two decades<sup>44</sup>. In addition, *in situ* temperature  
285 measurements in the Ross Sea, a region of anomalously rapid sea level rise (Fig. 1a),  
286 show a negligible thermosteric sea level rise between 1958 and 2008 for upper-ocean  
287 waters (200-800 m; ref. 45). It is therefore likely that surface warming does not  
288 contribute significantly to the sea level trends examined here.

289

#### 290 *Halosteric adjustment to changes in precipitation and sea ice volume*

291

292 There is currently no evidence to suggest significant contributions to halosteric sea  
293 level rise in the Antarctic subpolar seas from non-glacial sources. A widespread  
294 increase in Antarctic precipitation is a common result of climate models simulating the  
295 atmospheric response to changes in global climatic forcing over the late 20<sup>th</sup> and the

296 21<sup>st</sup> centuries<sup>46-47</sup>, yet atmospheric reanalyses have thus far proven too uncertain to  
297 detect whether this predicted precipitation increase is presently underway<sup>46</sup>. Similarly,  
298 the possibility that a significant reduction in sea ice volume may have contributed to  
299 halosteric sea level rise cannot be definitively excluded due to the scarcity of sea ice  
300 thickness measurements, but it seems highly unlikely given satellite observations that  
301 show Antarctic sea ice area increasing slightly over our study period<sup>3</sup>. In fact, a  
302 modelling study that assimilated these sea ice concentration data indicated an increase  
303 in sea ice volume<sup>48</sup>.

304

#### 305 *Barystatic response to variable wind forcing*

306

307 Another possible cause of the observed sea level rise anomaly is a barystatic  
308 adjustment to perturbed wind forcing. Prevalent westerly winds along the northern  
309 boundary of the Antarctic subpolar seas drive a northward Ekman transport that  
310 exports water from the region, establishing a mean sea level slope upwards to the  
311 north<sup>49</sup>. Any reduction in the intensity of the northward Ekman transport would thus  
312 cause a relative increase in sea level close to Antarctica by allowing this slope to relax.

313

314 The SSH impact of variability in wind forcing was investigated using ERA-Interim  
315 reanalysis data<sup>50</sup>, following comparison to other reanalysis products<sup>51</sup>, as well as  
316 verification of this product's winds and their trends against sea-ice drift  
317 measurements<sup>52</sup> and *in situ* observations<sup>53</sup>. A time series of monthly values of the mean  
318 SSH anomaly in the Antarctic subpolar seas shows a significant correlation with a  
319 record of cross-boundary northward Ekman transport (Figs. S10-S11;  $r^2 = 0.5$ ), with a  
320 linear gradient of  $-(3.8 \pm 0.6) \times 10^{-4} \text{ mm (Gt yr}^{-1})^{-1}$ . The linear trend in annual-mean  
321 northward Ekman transport out of the region is  $-480 \pm 140 \text{ Gt yr}^{-2}$  (i.e. a decrease in  
322 export) between 1992 and 2011, implying that a barystatic adjustment to wind forcing  
323 contributed  $0.2 \pm 0.1 \text{ mm yr}^{-1}$  of sea level rise in the Antarctic subpolar seas relative to  
324 the rest of the ocean, a minor fraction (20% at most) of the observed signal. Repeating  
325 this analysis at a circumpolar contour following the 3000-m isobath (Fig. 1a) yielded  
326 very similar results.

327

328 The frequency dependence of the relationship between Antarctic coastal sea level and  
329 Ekman transport can be assessed by temporally averaging both time series with a  
330 moving window of variable width (Fig. S12a). This exercise demonstrates that the  
331 transfer function between the Ekman transport away from the Antarctic subpolar seas  
332 and the regional SSH anomaly is essentially constant at  $\sim 3.8 \times 10^{-4} \text{ mm (Gt yr}^{-1})^{-1}$  for  
333 periods of one year to longer than a decade. The values of the transfer function are  
334 calculated simply as the linear fit between Ekman transport and Antarctic coastal sea  
335 level for each temporal resolution.

336

337 This method of assessment of the relationship between Antarctic coastal sea level and  
338 the cross-boundary Ekman transport suffers from both a lack of resolution of sub-  
339 annual time scales and the temporal aliasing issues outlined above. To assess the  
340 robustness of our basic result, we repeat the analysis with the mean SSH anomaly in  
341 the  $\sim 5\%$  of the Antarctic subpolar seas that is permanently ice-free (Fig. S12b),  
342 permitting a consideration of sub-annual timescales. The resulting transfer function  
343 shows relatively modest variability, ranging from  $1 \times 10^{-4}$  to  $3 \times 10^{-4} \text{ mm (Gt yr}^{-1})^{-1}$  for  
344 periods of 2 months to longer than a decade, where the different value of the transfer  
345 function simply reflects its relevance to a different area. Thus, we conclude that our

346 estimate of the transfer function in Figure S12a is likely to be robust and representative  
347 of the sensitivity of the mean SSH anomaly in the Antarctic subpolar seas to changes  
348 in wind forcing.

349  
350

## 351 **Bibliography**

352  
353

354 30 Ablain, M., Cazenave, A., Valladeau, G. & Guinehut, S. A new assessment of  
355 the error budget of global mean sea level rate estimated by satellite altimetry  
356 over 1993-2008. *Ocean Sci.* **5**, 193-201,

357

358 31 Beckley, B.D., Lemoine, F.G., Luthcke, S.B., Ray, R.D., Zelensky N.P., A  
359 reassessment of global and regional mean sea level trends from TOPEX and  
360 Jason-1 altimetry based on revised reference frame and orbits. *Geophys. Res.*  
361 *Lett.* **34**, L14608

362

363 32 Prandi, P., Ablain, M., Cazenave, A. & Picot, N. Sea level variability in the  
364 Arctic Ocean observed by satellite altimetry. *Ocean Sci.* **9**, 2375-2401,

365

366 33 Beckley, B. D. Zelensky, N. P. Holmes, S. A., Lemoine, F. G. Ray, R. D.  
367 Mitchum, G. T. Desai S. D. & Brown S. T. Assessment of the Jason-2  
368 Extension to the TOPEX/Poseidon, Jason-1 Sea-Surface Height Time Series for  
369 Global Mean Sea Level Monitoring. *Marine Geodesy* **33**,

370

371 34 Prandi, P., Ablain, M., Cazenave, A. & Picot, N. A. New Estimation of Mean  
372 Sea Level in the Arctic Ocean from Satellite Altimetry. *Mar. Geodesy* **35**, 61-  
373 81,

374

375 35 Stammer, D. Response of the global ocean to Greenland and Antarctic ice  
376 melting. *J. Geophys. Res.* **113**, C06022

377

378 36 Lorbacher, K., Marsland, S. J., Church, J. A., Griffies, S. M. & Stammer, D.  
379 Rapid barotropic sea level rise from ice sheet melting. *J. Geophys. Res.* **117**,  
380 C06003

381

382 37 Tamisiea, M. E.. 2011 Ongoing glacial isostatic contributions to observations  
383 of sea level change. *Geophysical Journal International*, **186** (3), 1036-1044,

384

385 38 Chambers, D.P., Wahr, J., Tamisiea, M.E., and Nerem R.S. Ocean mass from  
386 GRACE and glacial isotactic adjustment. *J. Geophys. Res.: Solid Earth* **115**,  
387 B11, B11415

388

389 39 Wunsch, C., Ponte, R. M. & Heimbach, P. Decadal trends in sea level patterns:  
390 1993-2004. *J. Clim.* **20**, 5889-5911,

391

392 40 Madec, G. NEMO ocean engine. *Note du Pole de modélisation, Institut Pierre-*  
393 *Simon Laplace (IPSL)* **27** (2008).

394

395 41 Danabasoglu, G., et al., North Atlantic Simulations in Coordinated Ocean-ice

- 396 Reference Experiments phase II (CORE-II). Part I: Mean States, Ocean  
397 Modelling, 73 (2014) 76-107,  
398
- 399 42 Ridgway K.R., Dunn, J.R. and Wilkin, J.L. Ocean interpolation by four-  
400 dimensional least squares; Application to the waters around Australia. *Journal*  
401 *of Atmospheric and Oceanic Technology* **19**, 9, 1357-1375, (2002)  
402
- 403 43 Barrier, N., Treguier A., Cassou C., Deshayes J. Impact of the winter North-  
404 Atlantic weather regimes on subtropical sea-surface height variability *Climate*  
405 *Dynamics* **41**, 1159-1171.  
406
- 407 44 Reynolds, R. W., Smith, T. M. Liu, C. Chelton, D. B. Casey, K. S. & Schlax,  
408 M. G. Daily high-resolution blended analyses for sea surface temperature. *J.*  
409 *Clim.* **20**, 5473-5496,).  
410
- 411 45 Jacobs, Stanley S., Claudia F. Giulivi, Large Multidecadal Salinity Trends near  
412 the Pacific–Antarctic Continental Margin. *J. Clim.* **23**, 4508–4524,  
413
- 414 46 Masson-Delmotte, V. *et al.* Past and future polar amplification of climate  
415 change: climate model intercomparisons and ice-core constraints. *Clim. Dyn.*  
416 **26**, 513-529,  
417
- 418 47 Genthon, C., Krinner, G. & Castebrunet, H. Antarctic precipitation and climate-  
419 change predictions: horizontal resolution and margin vs plateau issues. *Ann.*  
420 *Glaciol.* **50**, 55-60,  
421
- 422 48 Massonnet, F. *et al.* A model reconstruction of the Antarctic sea ice thickness  
423 and volume changes over 1980-2008 using data assimilation. *Ocean Model.* **64**,  
424 67-75,  
425
- 426 49 Rintoul, S. R., Hughes, C. W. & Olbers, D. The Antarctic Circumpolar Current  
427 system. In *Ocean Circulation and Climate*, Academic Press, San Diego, pp.  
428 271–302 (2001).  
429
- 430 50 Dee, D. *et al.* The ERA-Interim reanalysis: Configuration and performance of  
431 the data assimilation system. *Quart. J. Roy. Met. Soc.* **137**, 553-597,  
432
- 433 51 Bromwich, D. H., Nicolas, J. P. & Monaghan, A. J. An assessment of  
434 precipitation changes over Antarctica and the Southern Ocean since 1989 in  
435 contemporary global reanalyses. *J. Clim.* **24**, 4189-4209,  
436
- 437 52 Holland, P. R. & Kwok, R. Wind-driven trends in Antarctic sea ice drift.  
438 *Nature Geosci.* **5**, 872-875,  
439
- 440 53 Bracegirdle, T. J. & Marshall, G. J. The reliability of Antarctic tropospheric  
441 pressure and temperature in the latest global reanalyses. *J. Clim.* **25**, 7138-  
442 7146,  
443  
444  
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446 **Figure Legends**

447

448 **Figure S1 | Significance of Antarctic subpolar sea linear trend in SSH anomaly.**

449 Green shading indicates the area in which the anomalous linear trend in SSH (Fig. 1) is  
450 significantly different from zero with 95% confidence, determined using the Patterson  
451 t-test accounting for auto-covariance under the assumption of negligible systematic  
452 error.

453

454 **Figure S2 | Linear trend in SSH anomaly in the western Ross Sea for individual**

455 **satellite missions.** The global-mean rate of sea level rise is not removed for simplicity,  
456 and the bin-averaging box is indicated in the inset. ERS-1 (black), ERS-2 (dark blue)  
457 and Envisat (green) records are shown alongside the Antarctic subpolar sea mean SSH  
458 anomaly, including the rate of global-mean sea level rise (light blue). The average  
459 linear trend in the western Ross Sea box for the gridded AVISO data set is  $\sim 6 \text{ mm yr}^{-1}$ .  
460 The uncertainties for the single-mission trends are estimated using a bootstrap method  
461 accounting for the standard deviation of SSH within the box for each time step.

462

463 **Figure S3 | GIA corrections for the AVISO-MSLA altimetry data.** Left: the

464 correction made to altimetry data associated with the Tamisiea (2011) GIA model.  
465 Right: the (highly uncertain) correction for recent ice mass loss, predominantly from  
466 the Antarctic Peninsula, computed from Tamisiea (2011) GIA correction and  
467 Chambers (2010) GRACE-derived geoid data.

468

469 **Figure S4 | Time series of Antarctic subpolar sea SSH anomaly showing**

470 **continuously sea ice-free record.** The circumpolar mean of the SSH anomaly south of  
471 the oceanic boundary of the Antarctic coastal sea level rise signal (Fig. 1) is indicated  
472 in red. Data gaps show times of widespread sea ice cover. The mean of the SSH  
473 anomaly in a small subset of the Antarctic subpolar seas that is permanently sea ice-  
474 free is indicated in black. Both data sets have had the global-mean rate of sea level rise  
475 subtracted.

476

477 **Figure S5 | Regional anomaly in summer (January to April) linear sea level trend,**

478 **1992-2011, relative to the global barystatic rate of sea level rise.** As in Figure 1a,  
479 the black line demarks the northern boundary of the Antarctic coastal sea level rise  
480 anomaly. Markers indicate the location of *in situ* estimates of interdecadal freshening,  
481 shaded by the magnitude of the corresponding halosteric sea level rise. The reference  
482 for and information synthesised by each marker are given in the table in Figure 1c. The  
483 3000-m isobath is shown in green.

484

485 **Figure S6 | A decomposition of the NEMO-simulated linear trend in Antarctic**

486 **subpolar sea steric height anomaly.** a. Upper-ocean (0-800 m) trend in halosteric  
487 height. b. Deep ( $> 800 \text{ m}$ ) trend in halosteric height. c. Upper-ocean (0-800 m) trend in  
488 thermosteric height. d. Deep ( $> 800 \text{ m}$ ) trend in thermosteric height. The green  
489 contours show the 3000-m isobath.

490

491 **Figure S7 | Comparison of bottom temperature and salinity in the NEMO model**

492 **and the CARS Southern Ocean climatology.** a. and c. show bottom salinity and  
493 temperature distributions from the CARS Southern Ocean climatology. b. and d.  
494 indicate bottom salinity and temperature distributions estimated from the NEMO  
495 model.

496

497 **Figure S8 | Comparison between zonal-mean sections of salinity in the CARS**  
498 **climatology and the NEMO model.** The zonal-mean salinity distributions for CARS  
499 (upper) and NEMO (lower), with  $\sigma$ -4 density contours.

500

501 **Figure S9 | Comparison between zonal-mean sections of temperature in the CARS**  
502 **climatology and the NEMO model.** The zonal-mean temperature distributions for  
503 CARS (upper) and NEMO (lower), with  $\sigma$ -4 density contours.

504

505 **Figure S10 | Time series of Ekman transport into the Antarctic subpolar seas.** Full  
506 lines show the monthly and yearly-averaged Ekman transport into the Antarctic  
507 subpolar seas, with the dashed line indicating the linear fit to the yearly averaged data.

508

509 **Figure S11 | Relationship between the Ekman transport into the Antarctic**  
510 **subpolar seas and regional SSH anomaly.** The circles indicate monthly averaged  
511 values of the two variables (AASS: Antarctic subpolar seas). The solid line shows the  
512 linear fit to the circles, with the  $2\sigma$  uncertainty denoted by the dashed lines. The  
513 rectangle has sides of length defined by uncertainties in the trends in both variables,  
514 and indicates the area of Ekman transport - sea level space that the solid line would  
515 have to pass through in order for the observed trend in regional SSH anomaly to be  
516 explained by wind forcing.

517

518 **Figure S12 | Time scale dependence of the transfer function between changes in**  
519 **the Ekman transport across the northern boundary of the Antarctic subpolar seas**  
520 **(Fig. 1) and the regional SSH anomaly.** The upper panel illustrates results using SSH  
521 anomaly measurements in the entire Antarctic subpolar seas, and the lower panel  
522 shows results derived from the SSH anomaly record in the subset of the Antarctic  
523 subpolar seas that is permanently sea ice-free. The transfer function was estimated by  
524 averaging the time series of Ekman transport and SSH anomaly in temporal bins of  
525 variable length (indicated by the horizontal axis in both panels) and calculating the  
526 linear gradient of the resulting Ekman transport versus Antarctic subpolar seas SSH  
527 anomaly distribution.

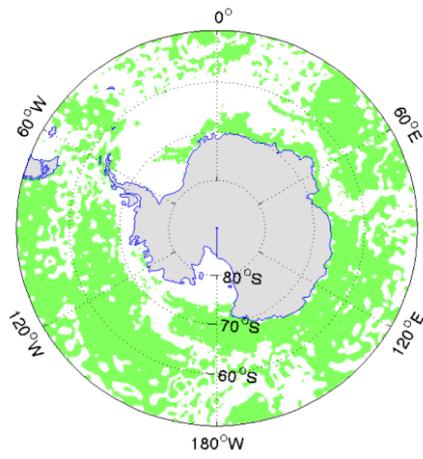
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529 **Figure S13 | Barystatic relative sea level rise resulting from glacial melt between**  
530 **2003 and 2009.** Derived from GRACE gravitometry (Riva et al., 2010).

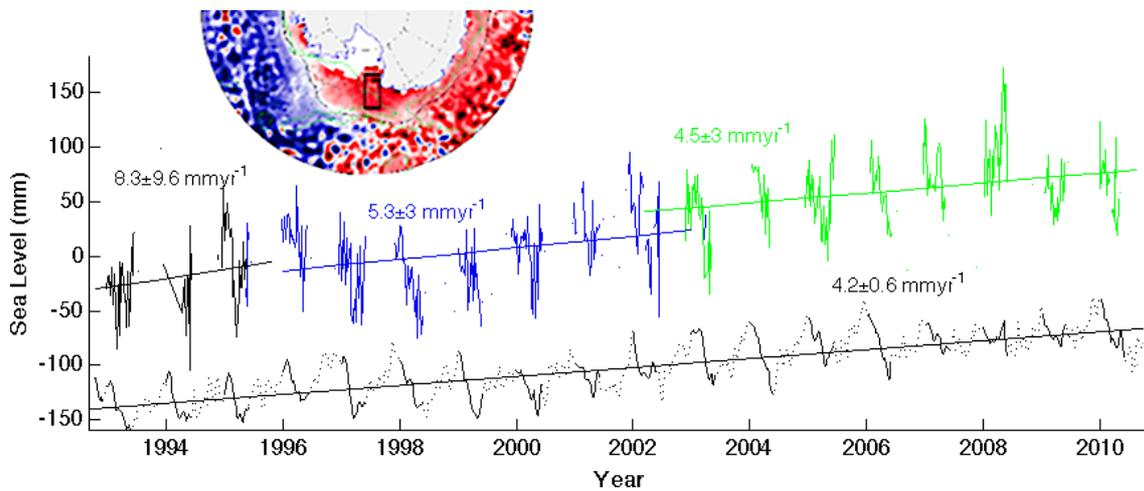
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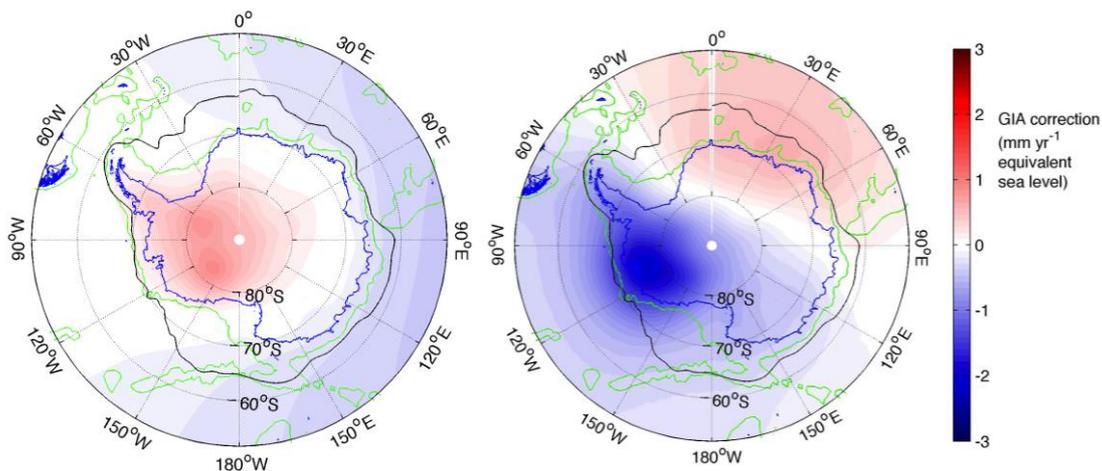


**Figure S1 | Significance of Antarctic subpolar sea linear trend in SSH anomaly.** Green shading indicates the area in which the anomalous linear trend in SSH (Fig. 1) is significantly different from zero with 95% confidence, determined using the Patterson t-test accounting for auto-covariance under the assumption of negligible systematic error.



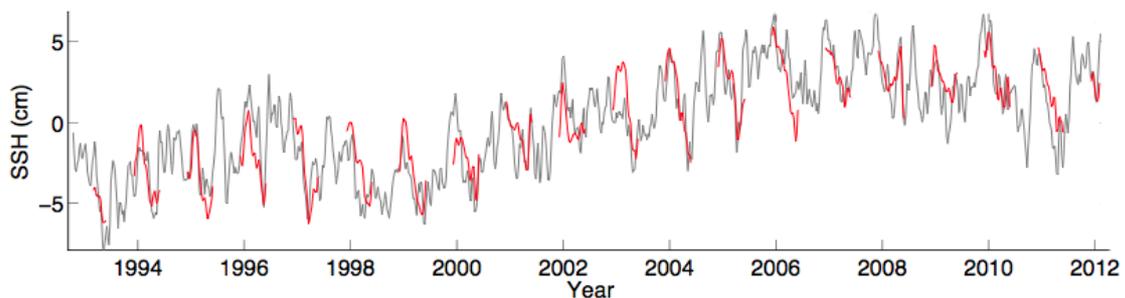
**Figure S2 | Linear trend in SSH anomaly in the western Ross Sea for individual satellite missions.** The global-mean rate of sea level rise is not removed for simplicity, and the bin-averaging box is indicated in the inset. ERS-1 (black), ERS-2 (dark blue) and Envisat (green) records are shown alongside the Antarctic subpolar sea mean SSH anomaly, including the rate of global-mean sea level rise (light blue). The average linear trend in the western Ross Sea box for the gridded AVISO data set is  $\sim 6 \text{ mm yr}^{-1}$ . The uncertainties for the single-mission trends are estimated using a bootstrap method accounting for the standard deviation of SSH within the box for each time step.

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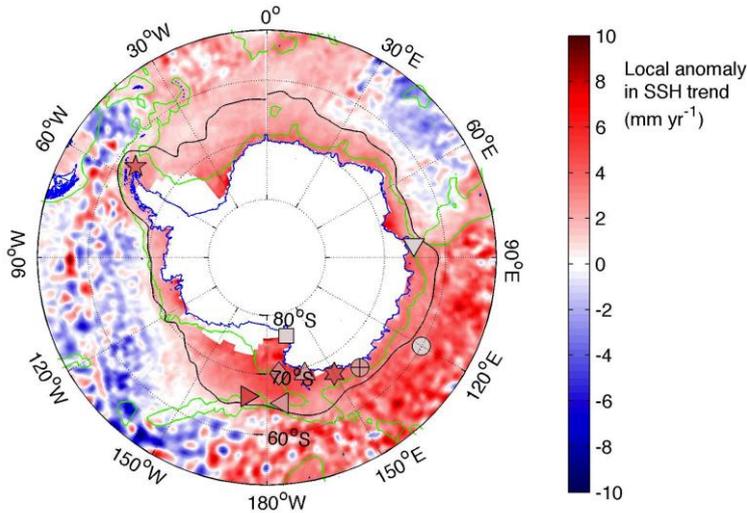


**Figure S3 | GIA corrections for the AVISO-MSLA altimetry data.** Left: the correction made to altimetry data associated with the Tamisiea (2011) GIA model. Right: the (highly uncertain) correction for recent ice mass loss, predominantly from the Antarctic Peninsula, computed from Tamisiea (2011) GIA correction and Chambers (2010) GRACE-derived geoid data.

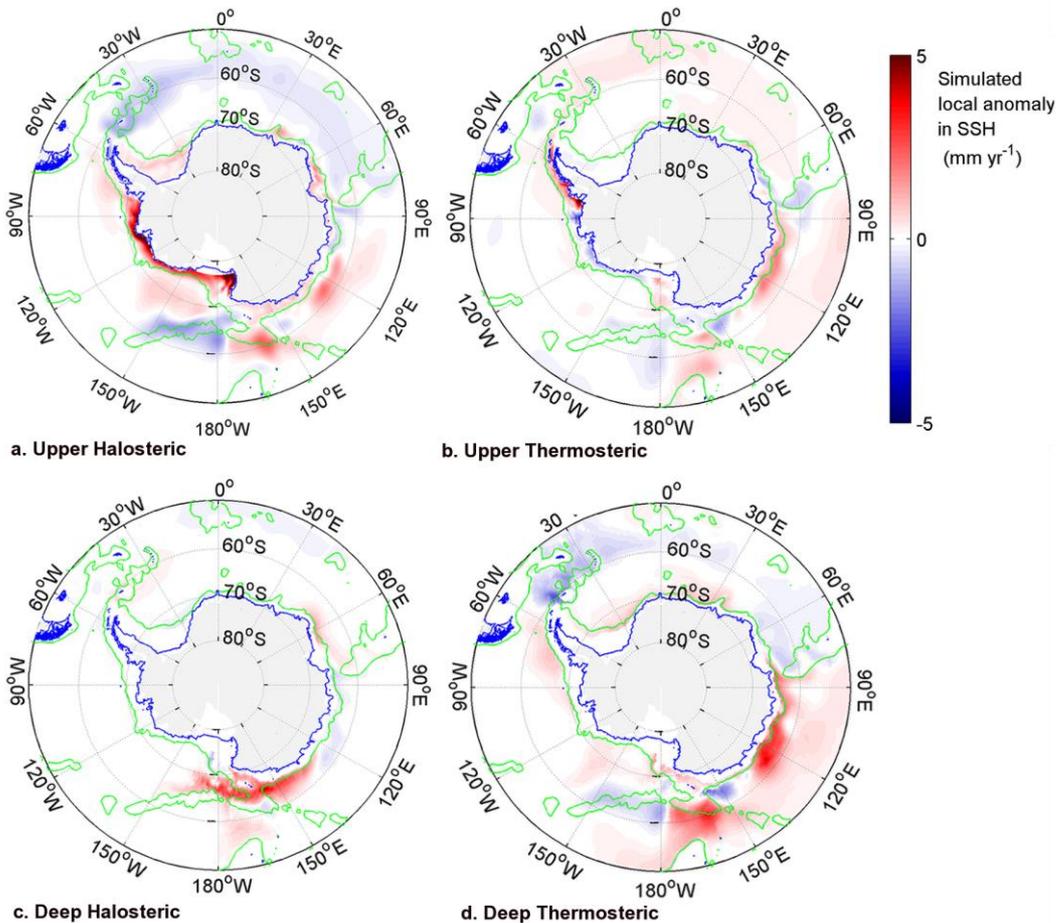
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**Figure S4 | Time series of Antarctic subpolar sea SSH anomaly showing continuously sea ice-free record.** The circumpolar mean of the SSH anomaly south of the oceanic boundary of the Antarctic coastal sea level rise signal (Fig. 1) is indicated in red. Data gaps show times of widespread sea ice cover. The mean of the SSH anomaly in a small subset of the Antarctic subpolar seas that is permanently sea ice-free is indicated in black. Both data sets have had the global-mean rate of sea level rise subtracted.

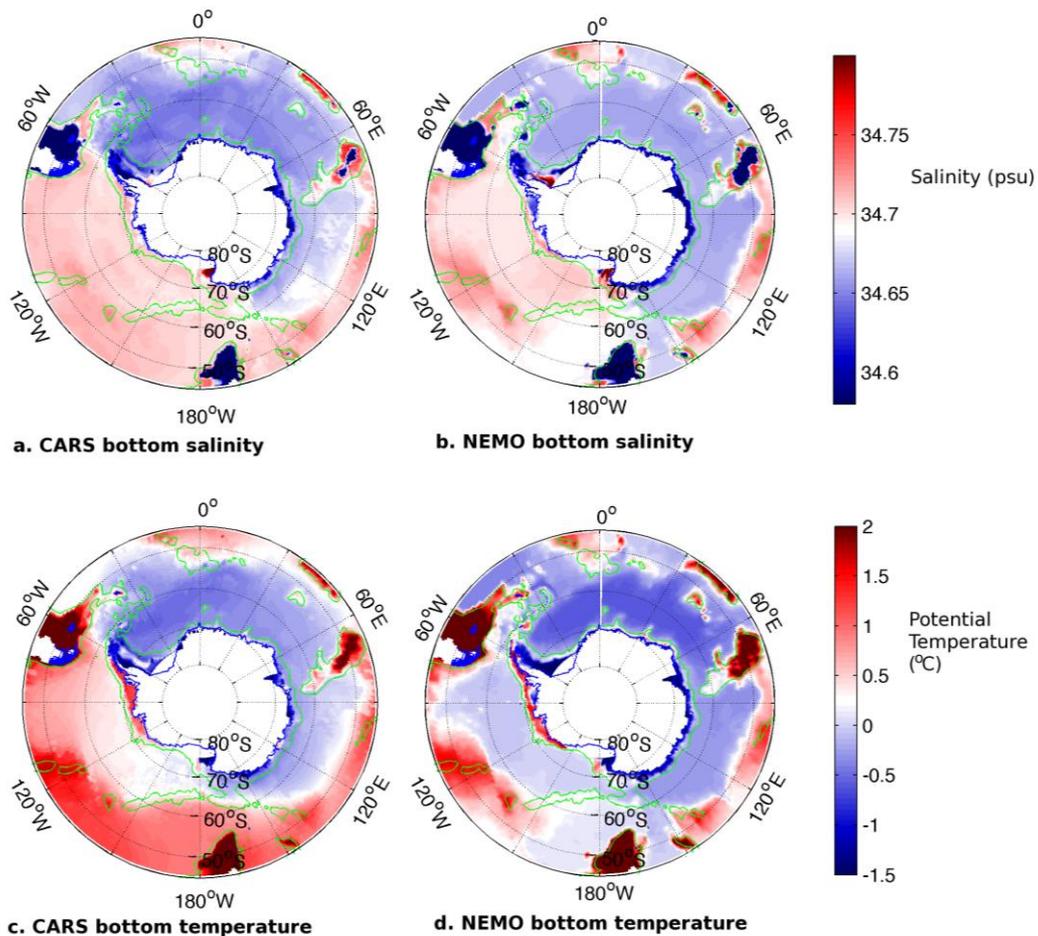


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 619 **Figure S5 | Regional anomaly in summer (January to April) linear sea level trend,**  
 620 **1992-2011, relative to the global barystatic rate of sea level rise.** As in Figure 1a,  
 621 the black line demarks the northern boundary of the Antarctic coastal sea level rise  
 622 anomaly. Markers indicate the location of *in situ* estimates of interdecadal freshening,  
 623 shaded by the magnitude of the corresponding halosteric sea level rise. The reference  
 624 for and information synthesised by each marker are given in the table in Figure 1c. The  
 625 3000-m isobath is shown in green.  
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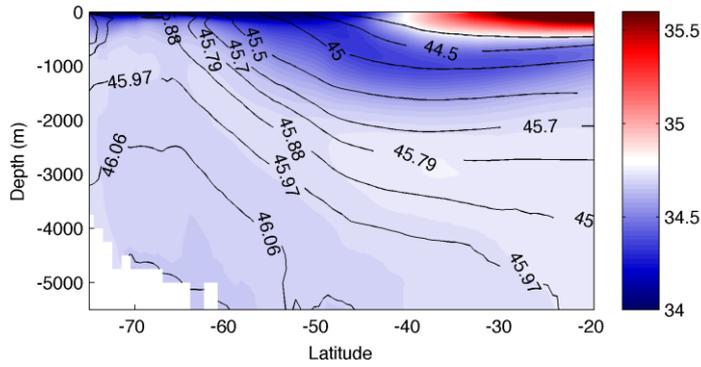


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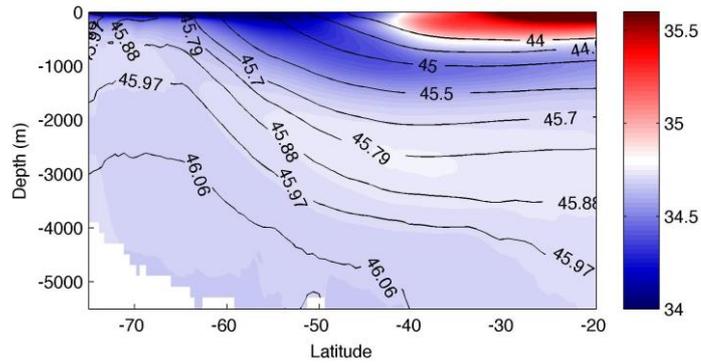
630 **Figure S6 | A decomposition of the NEMO-simulated linear trend in Antarctic**  
 631 **subpolar sea steric height anomaly.** a. Upper-ocean (0-800 m) trend in halosteric  
 632 height. b. Deep (> 800 m) trend in halosteric height. c. Upper-ocean (0-800 m) trend in  
 633 thermosteric height. d. Deep (> 800 m) trend in thermosteric height. The green  
 634 contours show the 3000-m isobath.  
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638 **Figure S7 | Comparison of bottom temperature and salinity in the NEMO model**  
 639 **and the CARS Southern Ocean climatology.** a. and c. show bottom salinity and  
 640 temperature distributions from the CARS Southern Ocean climatology. b. and d.  
 641 indicate bottom salinity and temperature distributions estimated from the NEMO  
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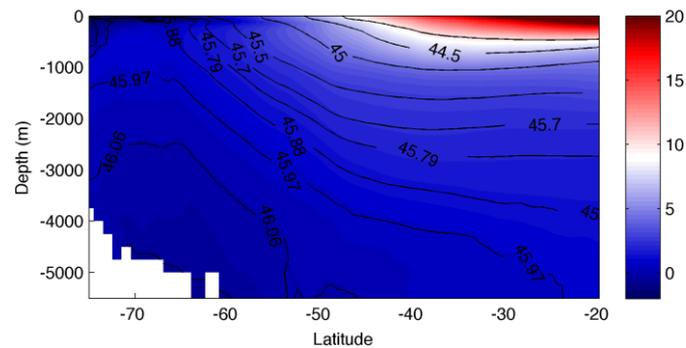
655 **Figure S8 | Comparison between zonal-mean sections of salinity in the CARS**  
 656 **climatology and the NEMO model.** The zonal-mean salinity distributions for CARS  
 657 (upper) and NEMO (lower), with  $\sigma$ -4 density contours.

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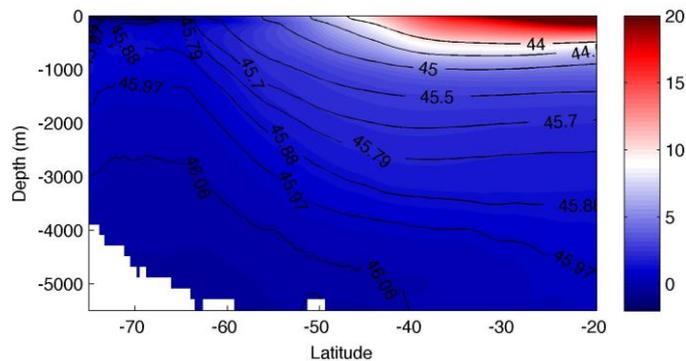
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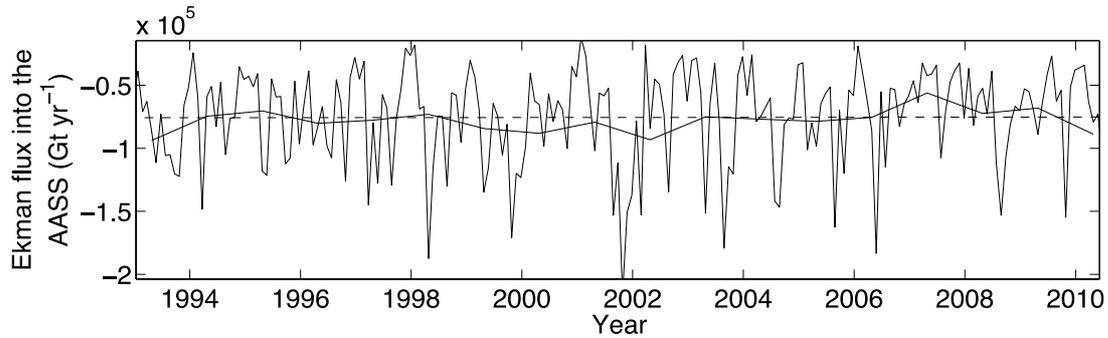
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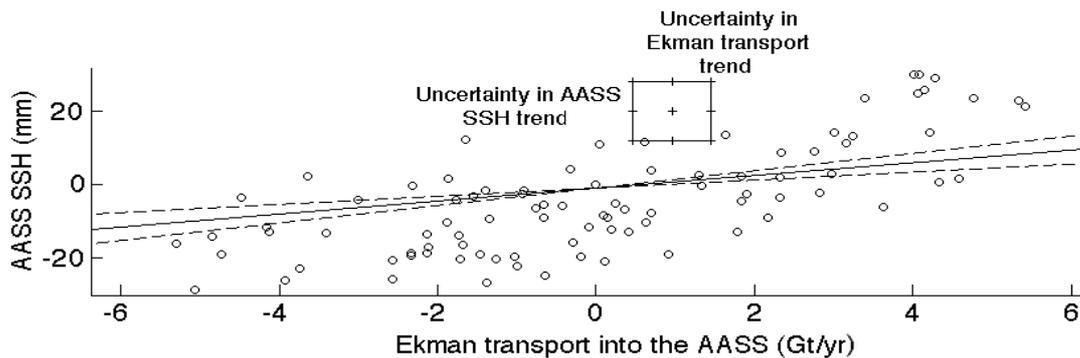
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664 **Figure S9 | Comparison between zonal-mean sections of temperature in the CARS**  
 665 **climatology and the NEMO model.** The zonal-mean temperature distributions for

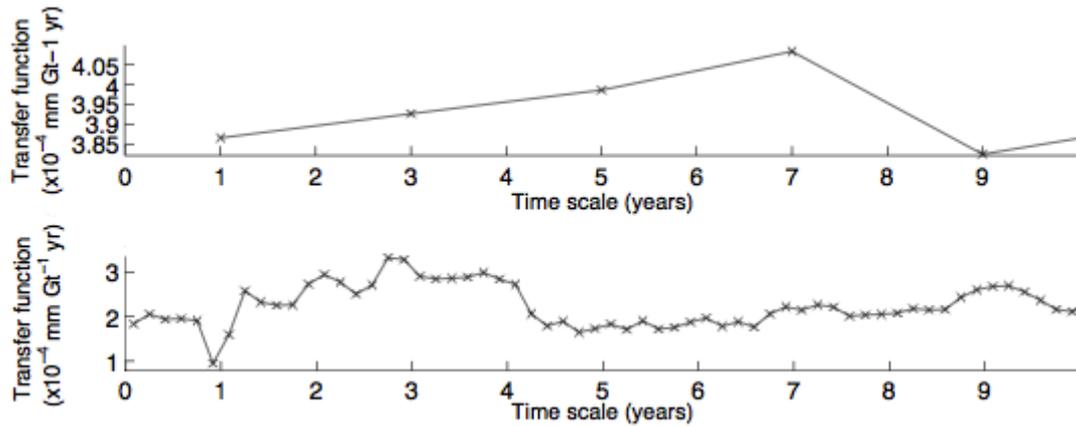
666 CARS (upper) and NEMO (lower), with  $\sigma$ -4 density contours.  
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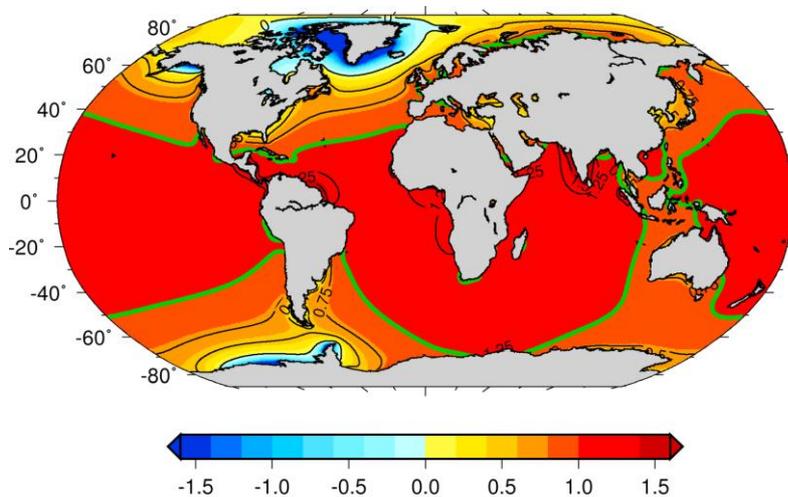
671 **Figure S10 | Time series of Ekman transport into the Antarctic subpolar seas.** Full  
 672 lines show the monthly and yearly averaged Ekman transport into the Antarctic  
 673 subpolar seas, with the dashed line indicating the linear fit to the yearly averaged data.  
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 679 **Figure S11 | Relationship between the Ekman transport into the Antarctic**  
 680 **subpolar seas and regional SSH anomaly.** The circles indicate monthly averaged  
 681 values of the two variables (AASS: Antarctic subpolar seas). The solid line shows the  
 682 linear fit to the circles, with the  $2\sigma$  uncertainty denoted by the dashed lines. The  
 683 rectangle has sides of length defined by uncertainties in the trends in both variables,  
 684 and indicates the area of Ekman transport - sea level space that the solid line would  
 685 have to pass through in order for the observed trend in regional SSH anomaly to be  
 686 explained by wind forcing.  
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 693 **Figure S12 | Time scale dependence of the transfer function between changes in**  
 694 **the Ekman transport across the northern boundary of the Antarctic subpolar seas**  
 695 **(Fig. 1) and the regional SSH anomaly.** The upper panel illustrates results using SSH  
 696 anomaly measurements in the entire Antarctic subpolar seas, and the lower panel  
 697 shows results derived from the SSH anomaly record in the subset of the Antarctic  
 698 subpolar seas that is permanently sea ice-free. The transfer function was estimated by  
 699 averaging the time series of Ekman transport and SSH anomaly in temporal bins of  
 700 variable length (indicated by the horizontal axis in both panels) and calculating the  
 701 linear gradient of the resulting Ekman transport versus Antarctic subpolar seas SSH  
 702 anomaly distribution.



724 **Figure S13 | Barystatic relative sea level rise resulting from glacial melt between**  
 725 **2003 and 2009.** Derived from GRACE gravitometry (Riva et al., 2010).