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¹ **Observing the Local Emergence of the Southern**
² **Ocean Residual-Mean Circulation**

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Key Points.

- Horizontal buoyancy advection by mesoscale eddies at a Southern Ocean mooring site balances mean vertical advection on timescale of 100 days.
- The overturning circulation of the Southern Ocean emerges on this timescale as a residual between the eddy-induced and Eulerian-mean flows.
- The eddy-induced flow can be accurately parameterized with a Gent-McWilliams diffusivity of $\sim 2,000 \text{ m}^2 \text{ s}^{-1}$.

Abstract

4 The role of mesoscale turbulence in maintain-
 5 ing the mean buoyancy structure and overturn-
 6 ing circulation of the Southern Ocean is in-
 7 vestigated through a 2-year-long, single-mooring
 8 record of measurements in Drake Passage. The
 9 buoyancy budget of the area is successively
 10 assessed within the Eulerian and the Temporal-
 11 Residual-Mean frameworks. We find that a regime
 12 change occurs on timescales of 1 day to 100 days,
 13 characteristic of mesoscale dynamics, whereby
 14 the eddy-induced turbulent horizontal advec-

15 tion balances the vertical buoyancy advection
16 by the mean flow. We use these diagnostics
17 to reconstruct the region's overturning circu-
18 lation, which is found to entail an equatorward
19 downwelling of Antarctic Intermediate and Bot-
20 tom Waters and a poleward upwelling of Cir-
21 cumpolar Deep Water. The estimated eddy-
22 induced flow can be accurately parameterized
23 via the Gent-McWilliams closure by adopt-
24 ing a diffusivity of $\sim 2,000 \text{ m}^2 \text{ s}^{-1}$ with a mid-
25 depth increase to $2,500 \text{ m}^2 \text{ s}^{-1}$ at 2,100 m, im-
26 mediately underneath the maximum interior
27 stratification.

1. Introduction

28 The Southern Ocean plays a pivotal role in the global ocean circulation by connecting
29 the Indian, Pacific, and Atlantic basins [*Schmitz, 1996*], and the ocean's deep and surface
30 layers [*Lumpkin and Speer, 2007*]. The regional circulation is shaped by physical processes
31 acting on a wide range of spatio-temporal scales – from large-scale, persistent Ekman flows,
32 directly linked to surface wind forcing, to meso- and small-scale turbulent motions [*Speer*
33 *et al., 2000*]. Assessing the relative contributions of, and interplay between, this spectrum
34 of flows in determining the Southern Ocean limb of the Meridional Overturning Circulation
35 (MOC) remains an important open question, with implications for e.g., global ocean heat
36 uptake [*Liu et al., 2018*] and sequestration of carbon dioxide [*Landschützer et al., 2015*].
37 However, a complete assessment of these dynamics and their effects is outstanding due to
38 the difficulty of rigorously distinguishing between averaged quantities (produced through
39 inaccurate diagnostic) and mixed quantities (produced through actual mixing). Averaged
40 quantities can be likened to the development of a blurred picture in photography when
41 an overly long exposure time is used to capture a moving object. Hence, unlike mixed
42 quantities, they are the outcome of an inaccurate observation rather than a property
43 of the observed system. The erroneous interpretation of averaged quantities as mixed
44 quantities has been rationalized in the context of the time-mean ocean circulation through
45 the Temporal-Residual-Mean (TRM) framework put forward by *McDougall and McIntosh*
46 [2001].

47 Here, we acknowledge this issue, by examining and quantifying the way in which the
48 mean buoyancy structure and overturning circulation are established at a Southern Ocean

49 site, where a mooring was deployed for 2 years. This also allows us to estimate the eddy-
50 induced Gent-McWilliams diffusivities. To conclude, a discussion of the impact of our
51 observational results for our understanding and numerical model representation of the
52 Southern Ocean limb of the MOC is provided.

2. Two distinct buoyancy budget regimes

53 To assess the contributions of motions of different scales to sustaining the extension of
54 the MOC across the Southern Ocean, we used *in situ* observations from a mooring located
55 in the Antarctic Circumpolar Current (ACC) at 56°S , $57^{\circ}50'\text{W}$ (Fig. 1a), leeward of the
56 Drake Passage [see Fig. 1a and *Brearley et al.*, 2013, for details of the exact location].
57 These 2-year long measurements were part of a 6-mooring cluster deployed under the aus-
58 pices of the Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean [DIMES,
59 *Naveira Garabato*, 2010; *Meredith*, 2011]. This study uses co-located horizontal velocity,
60 temperature, salinity and pressure data at 1200-, 1299-, 1853-, 1951-, 2049-, 2152-, 3400-,
61 3600-m depth (see Text S1 for further details). From these observations, vertical velocities
62 can be inferred, and the terms of a time-mean buoyancy budget can be diagnosed. The
63 method to compute vertical velocities from single-mooring measurements is described in
64 *Sévellec et al.* [2015] and summarized in Text S2; the mean statistical properties of the ob-
65 servational record are shown in Figs. 1b-g. Taking advantage of the equivalent-barotropic
66 nature of the ACC flow [*Killworth and Hughes*, 2002], we defined gradients in the along
67 and across directions of the time- and depth-mean flow (as indicated in Fig. 1a). This
68 enables us to eliminate contributions to the buoyancy budget from the time-mean flow in
69 the across direction (Fig. 1c), except for a small baroclinic component.

70 Since the measurements were obtained at broadly constant depth levels, it is natural
 71 to compute the buoyancy balance in an Eulerian framework with fixed depths. We start
 72 from the buoyancy conservation equation:

$$\partial_t b + u\partial_x b + v\partial_y b + w\partial_z b = 0, \quad (1)$$

73 where t is time; x , y , and z are the along, across, and vertical directions; b is the buoyancy;
 74 and u , v , and w are the along, across, and vertical velocities. By applying a Reynolds
 75 decomposition, set for a range of averaging periods (from 15 minutes to ~ 2 years), we
 76 separate mean quantities from fluctuations. Taking the overall long-time-mean of this
 77 decomposition allows us to build a mean buoyancy budget, which intrinsically depends on
 78 the averaging/filtering period (as described in Text S3). This can be expressed in terms
 79 of mean and turbulent advection as:

$$\text{Mean}_{\text{hor}}(\tau, z) + \text{Mean}_{\text{ver}}(\tau, z) \simeq \text{Turb}_{\text{hor}}(\tau, z) + \text{Turb}_{\text{ver}}(\tau, z), \quad (2)$$

80 where τ is the averaging period, Mean_{hor} and Mean_{ver} are the mean horizontal and vertical
 81 advection of buoyancy, and Turb_{hor} and Turb_{ver} are the turbulent horizontal and vertical
 82 advection of buoyancy, respectively. The mean buoyancy budget measures the relative
 83 contribution of the turbulent/fluctuation and mean terms in setting the long-time-mean
 84 equilibrium. It also shows how these contributions depend on the averaging/filtering
 85 period used in the Reynolds decomposition. Note that there is also a trend term in (??)
 86 compared to (2); however, it is negligible at all depths and for all averaging timescales by
 87 construction of the mean buoyancy budget (Text S3).

88 The resulting buoyancy advective terms are enhanced in the upper part of the measured
89 water column, between 1,000 m and 2,000 m (Fig. 2), where buoyancy gradients, and
90 notably vertical stratification, attain maximum values (Fig. 1). The turbulent vertical
91 advection (Fig. 2d) is negligible at all depths and for all averaging periods. This indicates
92 that variations in vertical velocity and vertical stratification do not correlate highly enough
93 to sustain a net turbulent advection, despite their substantial magnitudes compared to
94 their mean values (Figs. 1d and 1g). In contrast, the mean vertical advection induces a
95 buoyancy gain between 1,000 m and 2,000 m (Fig. 2b). This is consistent with a mean
96 downwelling of buoyant water: a downward mean vertical velocity through a positive
97 vertical stratification. The mean downwelling is robust to changes in averaging period,
98 confirming the steadiness of this buoyancy flux.

99 The mean horizontal advection induces a buoyancy loss between 1,000 m and 2,000 m
100 for averaging periods shorter than 1 day, but is insignificant for periods longer than
101 ~ 100 days (Fig. 2a). This dependency to averaging timescale suggests that this term
102 is not sustained by a steady horizontal flow. Finally, the turbulent horizontal advection
103 also induces a buoyancy loss between 1,000 m and 2,000 m (Fig. 2c) that emerges for
104 averaging periods longer than ~ 1 day, converging to a steady value for periods longer
105 than 100 days. This indicates that variations in horizontal velocity (Fig. 1b and 1c) and
106 in horizontal buoyancy gradients (Fig. 1e and 1f) are significantly correlated on periods
107 between 1 day and 100 days.

108 The preceding Eulerian buoyancy budget exhibits a remarkable dependence on averag-
109 ing period. Whereas on short timescales the buoyancy balance is between the mean

110 horizontal advection and the mean vertical advection ($\text{Mean}_{\text{hor}} + \text{Mean}_{\text{ver}} \simeq 0$), on long
111 timescales it is the turbulent horizontal advection that balances the mean vertical advec-
112 tion ($\text{Mean}_{\text{ver}} \simeq \text{Turb}_{\text{hor}}$). The change in regime occurs between 1-100 days. This timescale
113 is characteristic of mesoscale dynamics, as may be shown e.g., by considering the 6-day
114 propagation timescale derived from the observed zonal propagation speed of mesoscale
115 eddies [2 cm s^{-1} , *Klocker and Marshall, 2014*] and the baroclinic Rossby deformation
116 radius [10 km, *Chelton et al., 1998*] characteristic of the ACC. Accordingly, variability
117 within this timescale will be referred to as mesoscale turbulence in the remainder of this
118 study [*Klocker and Abernathey, 2014*].

119 The permanent regime is reached for averaging periods of 100 days. This convergence
120 timescale was reported by *Sévellec et al. [2015]* in a previous investigation of regional
121 dynamics. This period is linked to the cumulative effect of mesoscale eddies propagating
122 over the mooring site, and defines the minimum time over which measurements of eddy
123 variables must be acquired to obtain robust statistics and a convergence of the buoyancy
124 budget. This is symptomatic of the central role of eddies in sustaining the turbulent
125 horizontal advection. On timescales of 100 days, a balance is established between mean
126 vertical advection leading to a buoyancy gain and turbulent horizontal advection inducing
127 a buoyancy loss, in the depth range between 1,000 m and 2,000 m (Fig 2). This is con-
128 sistent with the prevalent view of the Southern Ocean limb of the MOC equatorward of
129 the ACC's axis, where the mooring is located (i.e., equatorward of the maximum gradient
130 of dynamic ocean topography, Fig.1a). In this area, buoyant waters are expected to be
131 downwelled by the mean vertical circulation; this effect is balanced by the mesoscale ed-

132 dies inducing a sink of buoyancy at depth [*Toggweiler and Samuels*, 1995, 1998; *Marshall*
 133 *and Radko*, 2003, 2006]. If ergodicity of the ACC flow is assumed (i.e., if time averag-
 134 ing may be considered as equivalent to spatial averaging), our observational diagnostics
 135 indicate that mean downwelling at the ACC’s northern edge [which represents the equa-
 136 torward, downwelling branch of the ‘Deacon cell’, *Döös and Webb*, 1994] is compensated
 137 by horizontal mesoscale eddy turbulence.

3. The residual-mean circulation

138 Having documented the pivotal role of turbulent horizontal advection on timescales
 139 longer than 100 days in balancing persistent downwelling, we will test how the former is
 140 represented within the TRM framework [*McDougall and McIntosh*, 1996, 2001]. This will
 141 enable us to characterize the physical nature of turbulent horizontal advection. In par-
 142 ticular, we will assess the extent to which the turbulent buoyancy flux may be accurately
 143 parameterized in terms of an adiabatic advection, which has been regularly assumed to
 144 prevail over diabatic processes [*Gent and McWilliams*, 1990]. This parameterisation of-
 145 ten entails the addition of an eddy-induced velocity proportional to the mean isopycnal
 146 slope [*Gent and McWilliams*, 1990]. In assessing this parameterization approach, we will
 147 mainly concentrate on the permanent regime (i.e., the longest time average).

148 We apply the TRM framework and compute the quasi-Stokes velocities, as initially sug-
 149 gested by *McDougall and McIntosh* [1996] and elaborated by *McDougall and McIntosh*
 150 [2001], and summarized in Text S4. We obtain the eddy-induced velocities in the along and
 151 across directions of the time- and depth-mean flow (respectively referred to as along and
 152 across eddy-induced velocities henceforth), and in the vertical direction (Fig. 3a). The

153 along eddy-induced velocities are small compared to the mean along velocities (Fig. 3a1),
 154 indicating that the residual along transport is mainly determined by the intense mean
 155 flow along the path of the ACC. For the across direction, the residual velocities (com-
 156 puted as the sum of the mean and eddy-induced velocities) are determined primarily by
 157 the eddy-induced velocities, and exhibit a substantial baroclinic structure (Fig. 3a2).
 158 This suggests that the residual across transport is dominated by the integrated effect of
 159 turbulent processes. In turn, vertical eddy-induced velocities are of similar intensity to
 160 the vertical mean velocities (Fig. 3a3). This results in a significant compensation of the
 161 mean and eddy-induced vertical velocities at all depths, which leads to the occurrence of
 162 small residual vertical velocities (Fig. 3a3). Thus, the vertical transport of buoyancy is
 163 significantly weaker than suggested by the mean vertical velocities because of the oppos-
 164 ing contribution of eddy-induced flows. This local result is consistent with the common,
 165 general circulation model-based view of the residual-mean circulation in the Southern
 166 Ocean, in which the Eulerian-mean vertical flow is almost perfectly balanced by the eddy-
 167 induced flow [i.e., the vanishing of the Deacon cell in the residual-mean framework, *Döös*
 168 *and Webb, 1994; McIntosh and McDougall, 1996*]. Overall, along residual velocities are
 169 of the order of tens of centimeters per second, across residual velocities are of the order
 170 of centimeters per second, and vertical residual velocities are of the order of tenths of
 171 millimeters per second.

172 Using these eddy-velocities we can express the buoyancy balance within the TRM frame-
 173 work as: $\text{Trnd} + \text{Hor} + \text{Ver} = 0$, where Trnd is the trend of modified buoyancy ($\hat{b} = \bar{b} + \tilde{b}$, the
 174 three terms denoting the modified buoyancy, mean buoyancy, and the rescaled buoy-

175 ancy variance), $\text{Hor}=\hat{u}\partial_x\hat{b}+\hat{v}\partial_y\hat{b}$, and $\text{Ver}=\hat{w}\partial_z\hat{b}$ (where $\hat{u}=\bar{u}+\tilde{u}$, $\hat{v}=\bar{v}+\tilde{v}$, and $\hat{w}=\bar{w}+\tilde{w}$
 176 are residual, mean, and eddy-induced along, across, and vertical velocities). Closer ex-
 177 amination of this buoyancy balance reveals that the trend of modified buoyancy is small
 178 compared to the horizontal and vertical advection of modified buoyancy, which tend to
 179 mutually compensate (Fig. 3b1). Vertical and horizontal fluxes of buoyancy are de-
 180 composed into four terms, in accordance with the TRM framework of *McDougall and*
 181 *McIntosh* [2001] and as summarized in Text S4. These terms correspond to: mean ad-
 182 vection of mean buoyancy (MAM); turbulent advection of mean buoyancy (TAM); mean
 183 advection of re-scaled buoyancy variance (MAV); and turbulent advection of re-scaled
 184 buoyancy variance (TAV); such that $\text{MAM}_{\text{ver}}=\bar{w}\partial_z\bar{b}$, $\text{TAM}_{\text{ver}}=\tilde{w}\partial_z\bar{b}$, $\text{MAV}_{\text{ver}}=\bar{w}\partial_z\tilde{b}$, and
 185 $\text{TAV}_{\text{ver}}=\tilde{w}\partial_z\tilde{b}$ (and equivalently for the horizontal terms).

186 From this diagnostic, we find that the horizontal buoyancy flux is primarily determined
 187 by the mean and turbulent advection of mean buoyancy and by the mean advection of
 188 re-scaled buoyancy variance, with a negligible contribution of the turbulent advection
 189 of re-scaled buoyancy variance (Fig. 3b2). In contrast, the vertical buoyancy flux is
 190 dominated by the contribution of the mean and turbulent advection of mean buoyancy,
 191 with negligible influence from the mean and turbulent advection of re-scaled buoyancy
 192 variance (Fig. 3b3). Note that the vertical advection terms are one order of magnitude
 193 larger than their horizontal advection counterparts (Fig. 3b2 vs Fig. 3b3). To summarize,
 194 the TRM framework indicates that the mean vertical advection of mean buoyancy and the
 195 turbulent vertical advection of mean buoyancy dominate the overall buoyancy balance,
 196 and act to compensate one another.

197 Now that the eddy-induced velocities and associated buoyancy fluxes have been derived
 198 and described, we can compute the eddy-induced velocity coefficients required in the *Gent*
 199 *and McWilliams* [1990] parameterisation. The mathematical description of this procedure
 200 is summarized in Text S4.

201 We find that the eddy-induced velocity coefficients for the across direction (relative to
 202 the time- and depth-mean flow, as above) range from $500 \text{ m}^2 \text{ s}^{-1}$ to almost $2,500 \text{ m}^2 \text{ s}^{-1}$
 203 (Fig. 4a), and vary widely in the vertical. These values including the *Gent and*
 204 *McWilliams* [1990] closure [e.g., $2,000 \text{ m}^2 \text{ s}^{-1}$ for NEMO in its $2^\circ \times 2^\circ$ configuration, *Madec*
 205 *and Imbard*, 1996; *Madec*, 2008], or those diagnosed in regional eddy-resolving models of
 206 e.g., the North Atlantic [$500 \text{ m}^2 \text{ s}^{-1}$ and $2,000 \text{ m}^2 \text{ s}^{-1}$, *Eden et al.*, 2007] or the California
 207 Current System [from $300 \text{ m}^2 \text{ s}^{-1}$ to $750 \text{ m}^2 \text{ s}^{-1}$, *Colas et al.*, 2013]. Following an abrupt
 208 increase from $500 \text{ m}^2 \text{ s}^{-1}$ at 3,200 m to $1,500 \text{ m}^2 \text{ s}^{-1}$ at 3,000 m depth, the across coeffi-
 209 cient increases linearly up to $2,000 \text{ m}^2 \text{ s}^{-1}$ at 2,300 m. Directly underneath the maximum
 210 deep stratification, the coefficient reaches its peak to values up to $2,500 \text{ m}^2 \text{ s}^{-1}$ at 2,100 m
 211 (Fig. 1g vs Fig. 4a). At shallower levels, the coefficient remains almost constant at
 212 $1,800 \text{ m}^2 \text{ s}^{-1}$ all the way to the upper ocean. Notably, we find that the closure converges
 213 when the time-averaging period is increased (Fig. 4a).

4. Discussion and Conclusions

214 Our analysis of a mooring in the ACC reveals that a regime shift in the buoyancy bal-
 215 ance occurs between 1-100 days, associated with the emergence of mesoscale dynamics.
 216 A timescale of 100 days characterises the convergence of flow statistics, and corresponds
 217 to the minimum time required for enough mesoscale eddy features to propagate past the

218 mooring. On this and longer timescales, an Eulerian buoyancy budget shows that the
 219 mean vertical advection of buoyancy is balanced by turbulent horizontal advection. The
 220 latter arises from velocities and buoyancy gradients evolving on timescales ranging from
 221 1 day to 100 days, characteristic of the mesoscale eddy field. The $O(1 \text{ day})$ timescale
 222 over which the regime change starts can be used to objectively distinguish between the
 223 eddying regime and the non-eddying regime in ocean models. Indeed, using an estimate
 224 of the speed of first-baroclinic, non-dispersive internal gravity waves ($NH/\pi=0.55 \text{ m s}^{-1}$,
 225 where $N=1.7\times 10^{-3} \text{ s}^{-1}$ and $H=1,000 \text{ m}$ following Fig. 1g) and the longest timescale for
 226 which the turbulent horizontal advection (i.e., the use of a turbulent closure) is negligible
 227 ($1/f \simeq 0.1 \text{ day}$, Fig. 2c), we obtain $0.55 \times 8640 = 5,000 \text{ m}$ or $\sim 1/13^\circ$ at the mooring latitude,
 228 which broadly corresponds to the first-baroclinic Rossby deformation radius. This value is
 229 significantly smaller than typical horizontal resolutions adopted by global climate models
 230 ($\sim 1/4^\circ$ at best) but is close to those used in recent global ocean models ($\sim 1/12^\circ$ at best,
 231 except for a few specific studies). This suggests that to accurately represent the buoyancy
 232 balance and residual circulation of the Southern Ocean, models require a horizontal reso-
 233 lution of a few kilometres or a robust turbulent closure for the effects of mesoscale eddy
 234 flows.

235 To validate this type of closure against observations, we quantified the eddy-induced
 236 circulation through the quasi-Stokes velocities of the TRM framework [*McDougall and*
 237 *McIntosh, 2001*]. We find that the mean horizontal velocities are only weakly modified
 238 by the eddy-induced velocities along the direction of the time- and depth-mean flow,
 239 whereas eddy-induced velocities dominate for the across direction. This leads to a signifi-

240 cant residual-mean horizontal transport across the direction of the time- and depth-mean
241 flow (of the order of centimeters per second). In contrast, the mean vertical velocities are
242 strongly compensated by the eddy-induced velocities. This leads to a weak residual-mean
243 vertical transport. The TRM view of the buoyancy balance differs from the Eulerian
244 view. In the former (which distinguishes averaged quantities from mixed quantities, un-
245 like the Eulerian framework), the buoyancy balance is primarily established between the
246 mean vertical advection of mean buoyancy and the turbulent vertical advection of mean
247 buoyancy.

248 Finally, we considered a parameterization of the eddy-induced circulation as a purely
249 advective process. Following *Gent and McWilliams* [1990], the quasi-Stokes velocity is
250 set proportional to the mean isopycnal slope [computed for the modified buoyancy, *Mc-*
251 *Dougall and McIntosh*, 2001]. This leads to across eddy-induced velocity coefficients of
252 $\sim 2,000 \text{ m}^2 \text{ s}^{-1}$, but varying by up to a factor 5 over the 2,000-m depth range examined.
253 Our analysis also shows an enhancement of the coefficient up to $2,500 \text{ m}^2 \text{ s}^{-1}$ at a depth
254 of 2,100 m immediately below the maximum of deep stratification.

255 All in all, our analysis suggests the existence of a residual-mean circulation that re-
256 distributes buoyancy vertically and in the across direction. Mapping the residual-mean
257 flow onto the temperature-salinity relation measured by the moored instrumentation, we
258 can diagnose the motion of the three major water masses present at the mooring loca-
259 tion: Antarctic Intermediate Water (AAIW), upper and lower Circumpolar Deep Water
260 (CDW), and Antarctic Bottom Water (AABW). These three water masses are derived
261 from three distinct water types (temperature-salinity properties at their origin/formation):

262 AAIW defined in the range of 3-7°C and 34.3-34.5 psu [*Carter et al.*, 2009], North Atlantic
263 Deep Water defined in the range of 3-4°C and 34.9-35 psu [*Defant*, 1961], and AABW
264 defined in the range of -0.9-1.7°C and 34.64-34.72 psu by [*Emery and Meincke*, 1986].
265 Following the density classes set by these three water types, we find that AAIW (light-
266 est layer) flows downward and equatorward, CDW (high-salinity layer) flows upward and
267 poleward, and AABW (densest layer) flows downward and equatorward (Fig. 4b). In
268 this way, the classical description of the large-scale overturning circulation of the South-
269 ern Ocean [*Schmitz*, 1996; *Speer et al.*, 2000], derived largely from analyses of basin- or
270 global-scale water mass property distributions and general circulation models, is seen to
271 emerge from local measurements of the buoyancy balance (Fig. 4c).

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Figure 3. (a1) Along, (a2) across, and (a3) vertical velocities for the (red dashed line) time-mean, (blue dashed line) eddy-induced and (black solid line) residual components for the longest averaging timescale ($\tau = t_2 - t_1 \simeq 2$ years). Eddy-induced velocity is computed through the quasi-Stokes velocity of the Temporal-Residual-Mean framework, and the residual velocity is the sum of the time-mean and eddy-induced velocities. The results are shown on a uniformly spaced 100-meter vertical grid (crosses) and are connected by a cubic spline interpolation (line). (b1-3) Buoyancy budget within the Temporal-Residual-Mean framework as a function of depth for the longest averaging timescale ($\tau = t_2 - t_1 \simeq 2$ years). (b1) Total buoyancy budget between the trend (TRND, black line), horizontal advection (HOR, red line), and vertical advection (VER, blue line). (b2 and b3) Horizontal and vertical buoyancy advection balance between the mean advection of mean buoyancy (MAM, black lines), the turbulent advection of mean buoyancy (TAM, red lines), the mean advection of rescaled buoyancy variance (MAV, blue lines), and the turbulent advection of rescaled buoyancy variance (TAV, purple lines). The results are shown on a uniformly spaced 100-meter vertical grid (crosses) and are connected by a cubic spline interpolation (line).

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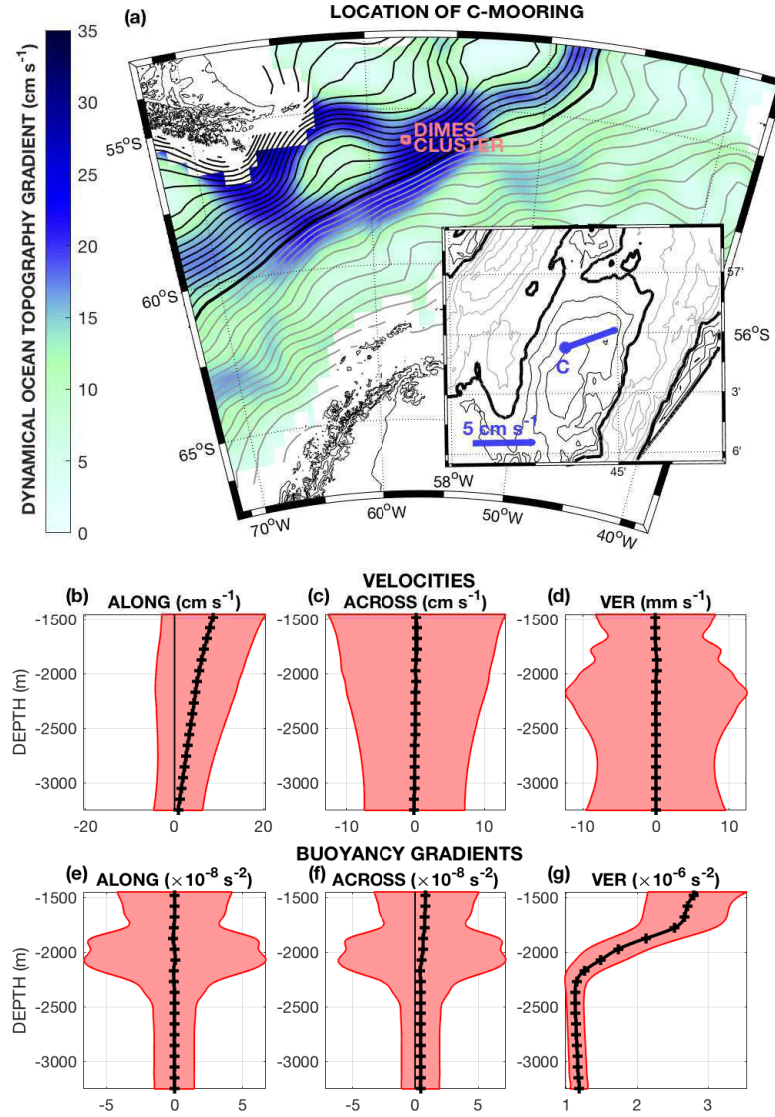


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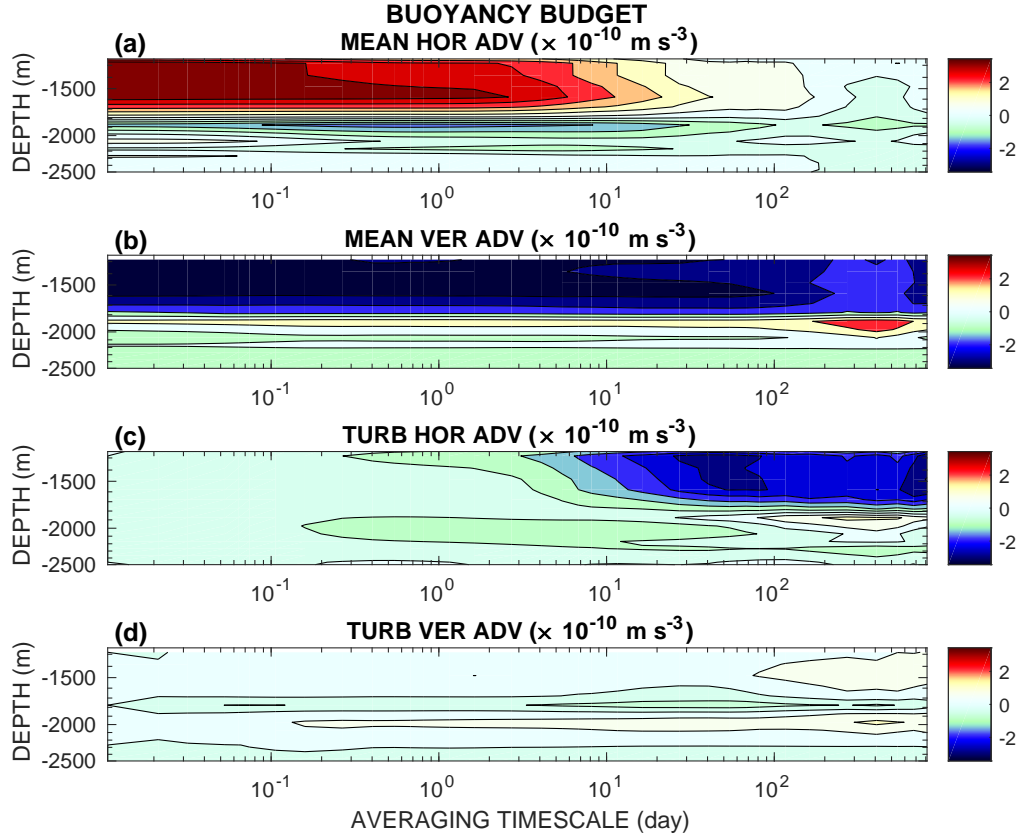


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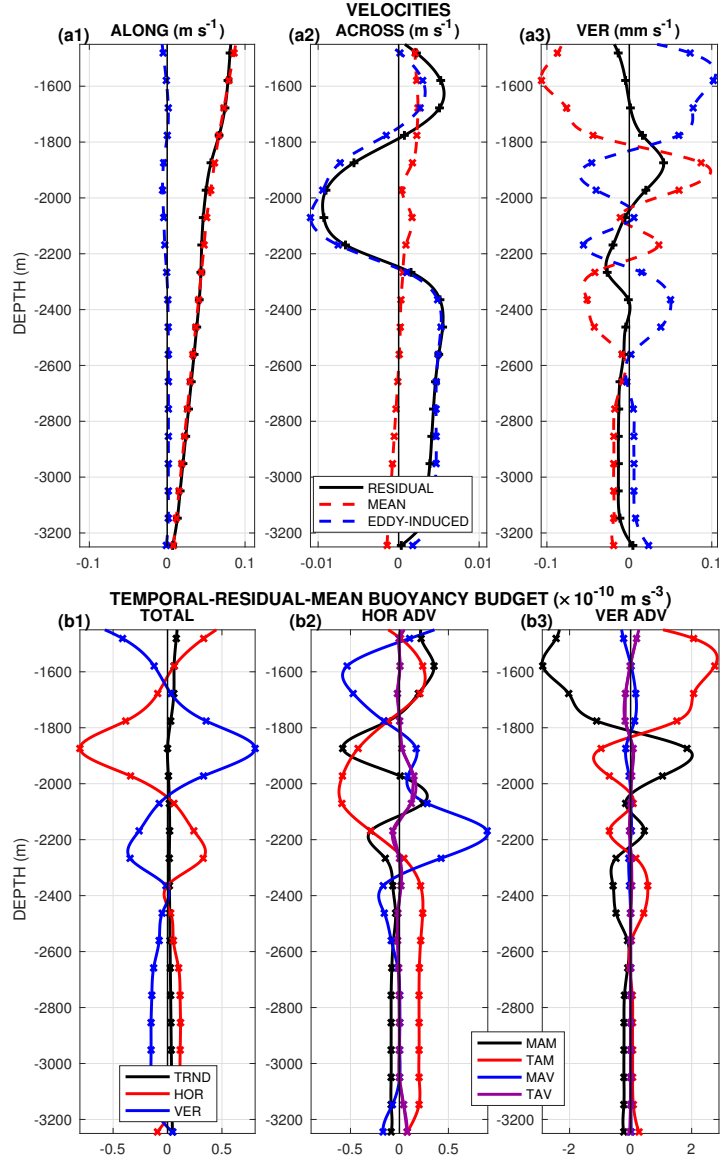


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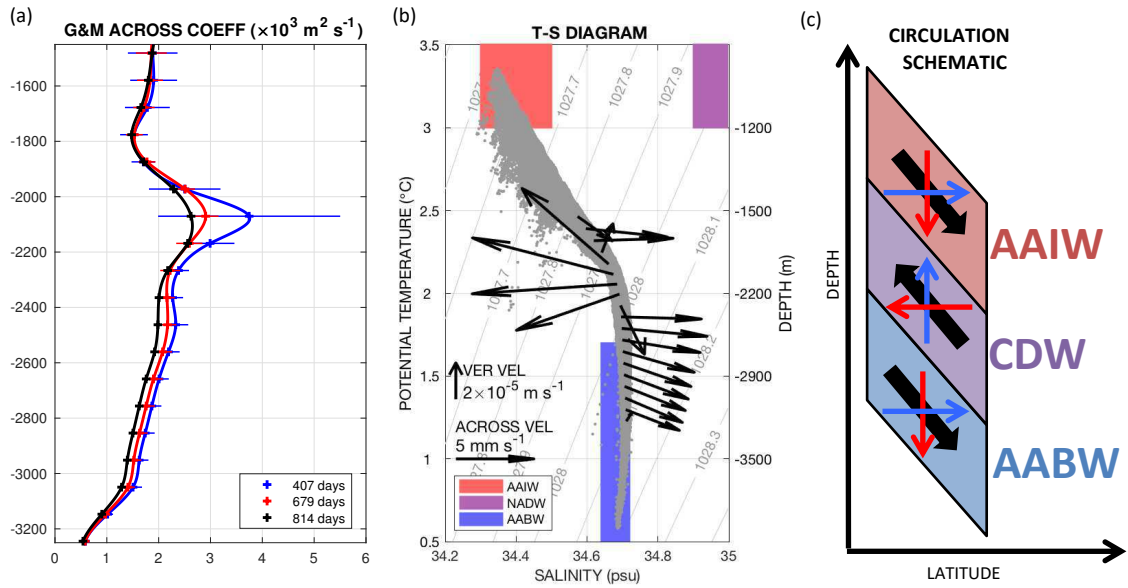


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