1 Anisotropy of Eddy Variability in the Global Ocean

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12 Abstract

13 The anisotropy of eddy variability in the global ocean is examined in
14 geostrophic surface velocities derived from satellite observations and in the
15 horizontal velocities of a 1/12° global ocean model. Eddy anisotropy is of
16 oceanographic interest as it is through anisotropic velocity fluctuations that
17 the eddy and mean-flow fields interact dynamically. This study is timely be-
18 cause improved observational estimates of eddy anisotropy will soon be avail-
19 able with *Surface Water and Ocean Topography* (SWOT) altimetry data. We
20 find there to be good agreement between the characteristics and distributions
21 of eddy anisotropy from the present satellite observations and model ocean
22 surface. In the model, eddy anisotropy is found to have significant verti-
23 cal structure and is largest close to the ocean bottom, where the anisotropy
24 aligns with the underlying isobaths. The highly anisotropic bottom signal
25 is almost entirely contained in the barotropic variability. Upper-ocean vari-
26 ability is predominantly baroclinic and the alignment is less sensitive to the
27 underlying bathymetry. These findings offer guidance for introducing a pa-
28 rameterization of eddy feedbacks, based on the eddy kinetic energy and un-
29 derlying bathymetry, to operate on the barotropic flow and better account
30 for the effects of barotropic Reynolds stresses unresolved in coarse-resolution
31 ocean models.

32 *Keywords:* ocean eddies, barotropic/baroclinic variability,
33 bathymetry-induced variability, variance ellipses

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1. Introduction

Variability, particularly at the mesoscale (10-100 km), accounts for the majority of the ocean’s kinetic energy (e.g., *Ducet et al.*, 2000; *Wunsch*, 2007; *McWilliams*, 2008) and plays a fundamental role in setting the ocean state and circulation (e.g., *Hallberg & Gnanadesikan*, 2006; *Waterman et al.*, 2011; *Fox-Kemper et al.*, 2013), including air–sea CO₂ fluxes (*Dufour et al.*, 2013). The fluctuating, or eddy, component of the velocity field interacts with the time-mean flow only through the anisotropic or directional portion of the variability (*Hoskins et al.*, 1983; see also *Marshall et al.*, 2012, *Waterman & Hoskins*, 2013 and *Waterman & Lilly*, 2015), and as such, insight into the dynamical feedbacks between the eddy and mean-flow fields of the global oceans can be gained by quantifying and characterizing the magnitude and distribution of eddy anisotropy. Knowledge of eddy anisotropy is especially useful for improving parameterizations of unresolved variability in coarse-resolution ocean models (e.g., *Eden*, 2010), and defining criteria for *where* to prescribe these parameterizations (*Hallberg*, 2013).

Quantifying the anisotropy of ocean variability and understanding its mean-flow feedback requires an appropriate methodology. Typically this involves a Reynolds decomposition of ocean velocity and tracer fields, partitioning them into the time-mean and time-varying (eddy) components, and then identifying the conditions where these two components interact (e.g., *Marshall*, 1984; *Plumb*, 1986; *Gent & McWilliams*, 1996). A convenient framework are variance ellipses (see, for example, *Preisendorfer*, 1988) which provide a visual representation of eddy momentum fluxes. Variance ellipses characterize the horizontal velocity variances and covariance, thereby capturing the geometry of the fluctuating portion of the horizontal velocity and its associated horizontal momentum fluxes in terms of the eddy kinetic energy (which sets the ellipse amplitude), degree of anisotropy (which sets the ellipse eccentricity) and orientation (direction of the principle axis along which the anisotropy is oriented). Knowledge of the degree of anisotropy and its orientation are important because eddy-mean-flow interactions are sensitive to the magnitude of the anisotropic eddy kinetic energy and to the extent that the eddy anisotropy orientation departs from the direction of the mean-flow shear. Variance ellipses have previously been used in atmospheric research (e.g., *Hoskins et al.*, 1983), in the analysis of idealized ocean models (e.g.,

69 *Marshall et al.*, 2012; *Waterman & Hoskins*, 2013) and observational studies
70 (e.g., *Morrow et al.*, 1994; *Stanton & Morris*, 2004) to represent the char-
71 acteristics of the variability and the interaction between the eddying and
72 mean flows. Improved observational estimates for variance ellipses and eddy
73 anisotropy of the ocean surface will soon be available through *Surface Water*
74 *and Ocean Topography* (SWOT) altimetry data (e.g., *Durand et al.*, 2010; *Fu*
75 *et al.*, 2010), increasing the need to appreciate the representation of these
76 metrics in present observations and modelling efforts.

77 This paper examines the characteristics and global distributions of the
78 anisotropy of ocean eddy variability from satellite observations and a $1/12^\circ$
79 global ocean circulation model. It is found that eddy anisotropy is rich in
80 both horizontal and vertical structure. In the model, eddy anisotropy in-
81 creases towards the ocean bottom, where the highly anisotropic bottom vari-
82 ability is aligned with the underlying isobaths. Decomposing the horizontal
83 flow into its barotropic and baroclinic components reveals that the anisotropy
84 of the bottom variability is almost entirely contained in the barotropic flow,
85 while the relatively isotropic upper-ocean variability is predominantly baro-
86 clinic. This offers guidance for the development of a parameterization for the
87 effects of unresolved barotropic variability that is based on the shape of the
88 underlying bathymetry and operates only on the barotropic flow.

89 2. Methodology

90 This section briefly describes the theoretical basis of the variance ellipse
91 analysis from which the eddy anisotropy is derived; for a more complete
92 review refer to *Hoskins et al.* (1983), *Marshall et al.* (2012), *Waterman &*
93 *Lilly* (2015), and references therein.

94 The zonal (u) and meridional (v) velocities at a given location $\mathbf{x}(x, y, z)$
95 are Reynolds decomposed into their time-mean ($\bar{u}$, $\bar{v}$) and time-varying (u' ,
96 v') components:

$$97 \quad u(\mathbf{x}, t) = \bar{u}(\mathbf{x}) + u'(\mathbf{x}, t), \quad \text{and} \quad v(\mathbf{x}, t) = \bar{v}(\mathbf{x}) + v'(\mathbf{x}, t). \quad (1)$$

98 At location $\mathbf{x}$, the timeseries of the time-varying components of the horizontal
99 velocity define a variance ellipse in (u', v') -space. The amplitude, eccentricity
100 and orientation of this variance ellipse are defined in terms of the time-mean
101 velocity variances $\overline{u'^2}$ and $\overline{v'^2}$ and covariance $\overline{u'v'}$.

102 The characteristics of the variance ellipse can be derived explicitly from
 103 the velocity covariance matrix Σ :

$$104 \quad \Sigma = \begin{bmatrix} \overline{u'^2} & \overline{u'v'} \\ \overline{u'v'} & \overline{v'^2} \end{bmatrix}, \quad (2)$$

105 which describes the time-mean horizontal eddy velocity variances and co-
 106 variance. It is convenient to separate this eddy covariance matrix into its
 107 isotropic and anisotropic components:

$$108 \quad \Sigma = K\mathbf{I} + \mathbf{A}, \quad (3)$$

109 where K is the eddy kinetic energy given by,

$$110 \quad K = \frac{(\overline{u'^2} + \overline{v'^2})}{2}, \quad (4)$$

111 $\mathbf{I}$ is the identity matrix, and $\mathbf{A}$ is the anisotropic component of the eddy
 112 covariance matrix,

$$113 \quad \mathbf{A} = \begin{bmatrix} M & N \\ N & -M \end{bmatrix}, \quad (5)$$

114 with

$$115 \quad M = \frac{(\overline{u'^2} - \overline{v'^2})}{2} \quad \text{and} \quad N = \overline{u'v'}. \quad (6)$$

116 M and N are composed of the eddy Reynolds stresses describing the eddy
 117 fluxes of horizontal momentum. These anisotropic elements of the eddy co-
 118 variance matrix describe the tendency for the variability to align along a
 119 preferred axis and combine to give the portion of the eddy kinetic energy
 120 associated with anisotropic variability L :

$$121 \quad L = \sqrt{M^2 + N^2}. \quad (7)$$

122 L is subject to the bound that it must always be less than or equal to the
 123 total eddy kinetic energy, such that,

$$124 \quad \frac{L}{K} \leq 1. \quad (8)$$

125 The ratio L/K is a measure of ellipse shape varying like the eccentricity
 126 between 0 and 1 and as such provides a useful measure of the degree of

127 so-called “eddy anisotropy”, the anisotropy associated with the eddy or fluc-
 128 tuating portion of the flow. As shown by *Waterman & Lilly* (2015), the
 129 covariance matrix (Eqn. 3) can be written in terms of the total eddy kinetic
 130 energy, the anisotropic eddy kinetic energy and an orientation Θ as,

$$131 \quad \Sigma = K\mathbf{I} + \mathbf{A} = K\mathbf{I} + L \begin{bmatrix} \cos 2\Theta & \sin 2\Theta \\ \sin 2\Theta & -\cos 2\Theta \end{bmatrix}. \quad (9)$$

132 Here Θ is the orientation of the principle axis of anisotropy (*Hoskins et al.*,
 133 1983; see also Fig. 2 of *Waterman & Hoskins*, 2013), hereinafter referred to
 134 as the “eddy orientation”. It is given by

$$135 \quad \Theta = \frac{1}{2} \tan^{-1} \left(\frac{N}{M} \right). \quad (10)$$

136 The orientation of eddy anisotropy relative to the orientation of the basic
 137 state ($\sim$ mean) shear defines the direction of the energy conversion between
 138 the mean and eddy fields. For eddies to extract energy from the horizontal
 139 shear of the mean current, the anisotropy must be orientated such that the
 140 variance ellipses “lean” against the mean shear; for ellipses that tilt with
 141 the mean flow shear, eddies transfer energy from the eddy field to the basic
 142 current (*Starr*, 1968; see also *Pedlosky*, 1987 Section 7.3, and Fig. 4 of
 143 *Marshall et al.*, 2012). It is clear from applying a coordinate rotation to
 144 Equation (9) that the first term $K\mathbf{I}$ is invariant to coordinate rotations, whilst
 145 the elements of the matrix comprising the second term will be modified by
 146 such a transformation; this indicates that L is associated with anisotropy or
 147 directionality of the velocity covariance.

148 Here, the eddy anisotropy of surface flows in the global ocean is examined
 149 in satellite observations and a $1/12^\circ$ global ocean model. For the case of the
 150 model output, the vertical structure of anisotropy and its relationship with
 151 the underlying model bathymetry is also considered. In addition to the full
 152 flow, the eddy anisotropy of the ocean model is computed for the barotropic
 153 and baroclinic variability separately; this is achieved by first decomposing
 154 the horizontal velocities into their barotropic (depth average) and baroclinic
 155 (deviation from the depth average) flows at each spatial location, and then
 156 repeating the above calculations (Eqns. 1–10).

157 The practical computation of these terms is simplified through ensemble
 158 averaging, that is,

$$159 \quad \overline{u'^2} = \overline{u^2} - \overline{u} \overline{u}, \quad \overline{v'^2} = \overline{v^2} - \overline{v} \overline{v}, \quad \overline{u'v'} = \overline{uv} - \overline{u} \overline{v}, \quad (11)$$

160 where $\bar{u}$, $\bar{v}$, $\overline{u^2}$, $\overline{v^2}$ and $\overline{uv}$ are ideally accumulated online, or alternatively
 161 from stored velocity fields representative of the temporal variability; here
 162 the latter is employed. Obviously, it is important to ensure the zonal and
 163 meridional velocities are collocated, which is not always the case for ocean
 164 model output.

165 3. Model and Observations

166 This study uses output from the ORCA12 (see description in *Treguier*
 167 *et al.*, 2014) configuration of NEMO (Nucleus for European Modelling of
 168 the Ocean; *Madec*, 2008), run by the DRAKKAR Group (*The DRAKKAR*
 169 *Group*, 2007). ORCA12 simulates the global ice-ocean circulation with a
 170 latitudinally-dependent horizontal resolution that is largest at the equator
 171 (9km, $1/12^\circ$) and decreases with the cosine of latitude towards the poles;
 172 the vertical resolution ranges from 5m at the surface to 250m at depth
 173 through 46 vertical levels. The model has been forced with the DRAKKAR
 174 Forcing Set #4 (DFS4; *The DRAKKAR Group*, 2007; *Brodeau et al.*, 2010)
 175 to produce a hindcast simulation for the 10-year period of January 2003 to
 176 December 2012, with the velocity fields output and stored as 5-day aver-
 177 ages. The timeseries of the zonal and meridional global velocity fields, and
 178 their barotropic and baroclinic components, are used to calculate the terms
 179 characterizing the eddy kinetic energy and its anisotropy (K , L , Θ).

180 In addition to the ORCA12 output, this study uses 10 years of geostrophic
 181 velocities derived from merged satellite altimetries; these altimetry products
 182 were produced by Ssalto/*Duacs* and distributed by *AVISO*, with support
 183 from *CNES* (<http://www.aviso.oceanobs.com/duacs/>). The velocities are pro-
 184 vided as daily-averages on a regular $1/4^\circ$ grid and cover the same period as
 185 the model simulations (2003–2012). Considering the ~ 10 day repeat orbital
 186 cycle and objective remapping of along-track observations to the gridded
 187 product, the global $1/4^\circ$ altimetry record is expected to only be able to re-
 188 solve a portion of the mesoscale variability, corresponding to eddies having
 189 an e-folding scale of greater than 30–45 km (see, for example, *Chelton et al.*,
 190 2011; *Hallberg*, 2013). Nevertheless it is useful to look at the eddy anisotropy
 191 in this data set globally as it contains valuable information about the sur-
 192 face geostrophic variability at the scales that are resolved (e.g., *Ducet & Le*
 193 *Traon*, 2001; *Huang et al.*, 2007; *Scott et al.*, 2008), and it offers guidance
 194 for understanding future estimates of eddy anisotropy derived from improved
 195 observational systems, such as the upcoming SWOT altimetry data (e.g., *Du-*

196 *rand et al.*, 2010; *Fu et al.*, 2010). It is important to note that the different
 197 spatial and temporal resolutions of the model and altimetry data do not al-
 198 low for a direct and fair comparison of calculated variabilities, and that the
 199 ambiguity of the altimetry data resolution (especially in regards to the scales
 200 of variability resolved) complicates the process of selectively-sampling the
 201 model output so as to provide a fair comparison for the altimetry product.

202 4. Results

203 This section presents the anisotropy of ocean variability calculated from
 204 the AVISO product and ORCA12 model output. The section is divided into
 205 four parts; the focus of the first is on the anisotropy of surface velocities,
 206 the second on the vertical structure of eddy anisotropy in the model velocity
 207 fields, the third on the eddy anisotropy of the barotropic and baroclinic
 208 velocities in the modelled Southern Ocean, and the fourth on the model eddy
 209 anisotropy in the vicinity of the Macquarie Ridge and Campbell Plateau.
 210 The focus on the Southern Ocean, and in particular the vicinity of Macquarie
 211 Ridge and Campbell Plateau, serve as illustrative examples of the distinction
 212 between eddy anisotropy of the barotropic and baroclinic velocity fields in
 213 eddy-rich regions where bathymetry is known to have a strong influence on
 214 the flow.

215 4.1. Surface Eddy Anisotropy

216 The AVISO product was used to characterize the global distribution of
 217 eddy anisotropy by calculating K , L and their ratio (Fig. 1). The distri-
 218 bution of L (Fig. 1(b)) shows many similarities with the distribution of K
 219 (Fig. 1(a)), being largest in and surrounding regions of large flow speeds.
 220 However relative to K , the spatial distribution of L appears more localized
 221 to the large flow speed regions and is also richer in mesoscale structure. The
 222 eddy anisotropy L/K tends to be large where currents are strongly influ-
 223 enced by bathymetry or by equatorial dynamics, and exhibits complicated
 224 mesoscale structure (Fig. 1(c)). Whilst the distribution of L/K is maximal
 225 along coastlines and shelfbreaks, and near the equator, the central basin re-
 226 gions tend to be more isotropic. The global average value of surface eddy
 227 anisotropy is 0.25, implying that approximately 25% of surface eddy kinetic
 228 energy is associated with anisotropic variability (Tab. 1).

229 The eddy anisotropy of the ORCA12 model surface fields generally com-
 230 pares well with that derived from the satellite observations in both magnitude

231 and spatial distributions, a consequence of the fact that both the eddy ki-
 232 netic energy (Figs. 1(a) & 2(a)) and its anisotropic component (Figs. 1(b)
 233 & 2(b)) in the model and satellite observations are similar in magnitude and
 234 distribution. However, regions where the ORCA12 anisotropy is consistently
 235 greater than that of the AVISO-derived fields include the coastlines and equa-
 236 torial latitudes, a consequence of the anisotropic eddy kinetic energy L being
 237 significantly larger in the model fields (Figs. 1(b) & 2(b)). Globally, the
 238 average eddy anisotropy in the model surface is 0.29, implying that 29% of
 239 ORCA12 surface eddy kinetic energy is associated with anisotropic variabil-
 240 ity (Tab. 1); this is slightly more anisotropic than the AVISO-derived fields
 241 and is reflected in the global distributions of L/K (Figs. 1(c) & 2(c)).

242 *4.2. Vertical Structure of Eddy Anisotropy*

243 The ORCA12 model output provides an opportunity to investigate the
 244 vertical structure of ocean variability and, in particular, eddy anisotropy.
 245 An initial step is to compare the anisotropy of surface variability with that
 246 of the bottom; for this, the horizontal velocities of the second-deepest non-
 247 bathymetry grid cell are used to calculate the bottom covariance fields, so
 248 as to avoid complications with viscous boundary conditions and bathymetry
 249 representation schemes. In contrast to the surface, the bottom variability is
 250 predominantly anisotropic, as reflected in the similarities between the bottom
 251 K and bottom L fields (Figs. 3(a) & 3(b)). Although the global average of
 252 bottom eddy kinetic energy is almost 2 orders of magnitude smaller than
 253 the surface, the global average value of eddy anisotropy is 0.65, implying
 254 that 65% of all bottom eddy kinetic energy is associated with directional
 255 or anisotropic variability (Tab. 1). The distribution of the bottom eddy
 256 anisotropy appears to be influenced by the underlying bathymetry; regions of
 257 significant bathymetric features (e.g., mid-ocean ridges, continental shelves,
 258 island chains) tend to be highly anisotropic, whilst the relatively featureless
 259 plains and basins (e.g., Aleutian, Admundsen, Argentine, Bellingshausen,
 260 Ceylon, Enderby, Weddell) are more isotropic (Fig. 3(c)).

261 Vertical sections of eddy anisotropy L/K highlight this transition be-
 262 tween the anisotropic bottom and relatively isotropic surface. The bottom
 263 intensification of eddy anisotropy is evident in the zonal average of L/K (Fig.
 264 4(a)), and with the exception of the equatorial regions, the eddy anisotropy
 265 generally increases monotonically with depth.

266 The anisotropic bottom signal tends to be localized to regions of slop-
 267 ing bathymetry whilst in regions of flat bathymetry, the bottom variability

268 tends to be more isotropic. This is evident in the relative intensification of
 269 the bottom anisotropy when the calculation of the zonal average of L/K
 270 is restricted to regions above sloping bathymetry (i.e. $\nabla H \neq 0$, where H
 271 is the ocean model depth; Fig. 4(b)), compared to when it is restricted to
 272 regions of locally flat model bathymetry (i.e. $\nabla H = 0$; Fig. 4(c)). This
 273 differing influence of flat or sloping bathymetry is readily observed in various
 274 basin transects of L/K , such as a circumpolar transect at 55°S (Fig. 5(a)).
 275 In regions of sloping bathymetry, the anisotropic bottom signal penetrates
 276 upwards ~ 1000 m into the water column, occasionally reaching the ocean
 277 surface. In addition to the bottom intensification, the eddy anisotropy is
 278 notably enhanced downstream of major obstacles across the Antarctic Cir-
 279 cumpolar Current (ACC), such as the Drake Passage, Kerguelen Plateau and
 280 Macquarie Ridge.

281 Vertical transects of the eddy orientation, Θ , reveal that in spite of the
 282 significant vertical structure in eddy anisotropy, the eddy orientation re-
 283 mains relatively unchanged with depth (Fig. 5(b)). Note that while the
 284 cardinal direction of the eddy orientation is of little consequence (rather, the
 285 eddy orientation direction relative to that of the mean shear is important for
 286 diagnosing eddy-mean-flow interactions; see *Waterman & Lilly*, 2015), the
 287 observation that the orientation of the eddy anisotropy maintains vertical co-
 288 herence suggests a significant portion of the anisotropic signal is barotropic.

289 Given the apparent link between eddy anisotropy and sloping bathymetry,
 290 it is useful to consider the alignment of the eddy anisotropy relative to the
 291 direction of the underlying isobath (Fig. 6). This is an angle between 0
 292 and $\pi/2$, where 0 implies the eddy anisotropy is aligned with the isobath,
 293 and $\pi/2$ implies they are perpendicular. At the surface, there is a slight
 294 tendency for the variability to align itself with the underlying isobath (Fig.
 295 6(a)); approximately 15% of the anisotropic surface eddy kinetic energy is
 296 aligned within 10° of the underlying isobath. This effect is amplified at
 297 depth: bottom variability has an increased tendency to align with the local
 298 isobath (Fig. 6(b)); over 20% of the anisotropic bottom eddy kinetic energy
 299 is aligned within 10° of the underlying isobath.

300 4.3. Barotropic vs. Baroclinic Eddy Anisotropy

301 The barotropic tendency of eddy anisotropy warrants further investiga-
 302 tion. Given the prevalence of Southern Ocean eddies, and their crucial role
 303 in the ACC and its vertical structure, the region south of $\sim 30^\circ$ S serves

304 as an illustrative example of the distinction between eddy anisotropy of the
 305 barotropic and baroclinic flows (e.g., *Rintoul*, 2010).

306 Separating the horizontal flow into its barotropic (depth-averaged) and
 307 baroclinic (deviation from the depth-average) components shows that over
 308 60% of the total eddy kinetic energy in the ORCA12 Southern Ocean is con-
 309 tained in the barotropic flow (Tab. 2). The distribution of barotropic eddy
 310 anisotropy is highly variable and rich in mesoscale (10-100 km) structure; the
 311 barotropic variability is anisotropic on continental shelves, shelfbreaks, and
 312 plateaus, and its distribution resembles that of $|\nabla(f/H)|$ (where f is the
 313 Coriolis parameter; compare Fig. 7(c) with 7(g)).

314 Baroclinic variability in the ORCA12 Southern Ocean contains less than
 315 40% of the total eddy kinetic energy (Tab. 2). At the surface, the distri-
 316 bution of eddy anisotropy in the baroclinic flow is virtually unchanged from
 317 that of the full flow case (compare Fig. 7(b) with 7(d)), and the anisotropy is
 318 strongest where $|\nabla(f/H)|$ is large (Fig. 7(g)). The Southern Ocean average
 319 of baroclinic bottom K is similar to the full flow case, whilst the baroclinic
 320 bottom L , and thus the eddy anisotropy L/K , almost halves (Tab. 2). This
 321 shift towards a more isotropic bottom baroclinic eddy field is most obvious in
 322 the distribution of baroclinic bottom L/K (compare Fig. 7(e) with 7(f)); for
 323 both the baroclinic and full flow cases, the bottom eddy anisotropy is weakest
 324 where $|\nabla(f/H)|$ is small (Fig. 7(g)). Indeed, the vertical penetration of the
 325 strong bottom eddy anisotropy above sloping bathymetry that is evident in
 326 the full flow (Fig. 5(a)) is not present in the baroclinic flow (Fig. 8). Addi-
 327 tionally, with the absence of the bottom intensification, the enhancement of
 328 eddy anisotropy downstream of major obstacles is more apparent.

329 A dominant feature of the circumpolar transect of baroclinic L/K is an
 330 intermediate-depth maximum in eddy anisotropy. This is an artifact of the
 331 baroclinic decomposition; at depths where the current velocity in a given
 332 direction is equal to the barotropic velocity (for example, $u = u_{BT}$ or $v =$
 333 v_{BT}), the baroclinic velocity becomes zero, leading to an anisotropizing of
 334 the variability in the perpendicular direction. This feature also suggests
 335 that for ORCA12 the majority of the baroclinic variability at this latitude is
 336 contained in the first baroclinic mode, such that $u = u_{BT}$ or $v = v_{BT}$ at only
 337 one depth in the water column.

338 4.4. Eddy Anisotropy at the Macquarie Ridge and Campbell Plateau

339 Given the rich spatial structure of eddy anisotropy, and its apparent re-
 340 lationship with bathymetry, it is useful to consider the eddy anisotropy in a

341 localized region with significant bathymetry that is known to influence the
 342 flow. One such region is the Macquarie Ridge; this ridge presents a large
 343 bathymetric obstacle for the ACC, which crosses the ridge primarily through
 344 a few narrow gaps. A major gap in the ridge is north of Macquarie Island
 345 at (53.3°S, 159.4°E) where a flow equivalent to almost one-third of the total
 346 ACC transport crosses the ridge as a predominantly barotropic and steady
 347 current; it is one of the rare places in the ocean where the mean kinetic
 348 energy is observed to exceed the eddy kinetic energy (*Rintoul et al.*, 2014).
 349 Eastward of the Macquarie Ridge, the ACC encounters Campbell Plateau
 350 where, rather than flowing over the Plateau, the ACC tends to flow along
 351 the slope as an energetic and highly variable current (*Stanton & Morris*,
 352 2004).

353 In the model, the mean surface flow is fastest across the Macquarie Ridge
 354 at the (53.3°S, 159.4°E) gap, and remains at current speeds of $> 0.2 \text{ m s}^{-1}$ as
 355 it flows along the southeastern edge of Campbell Plateau (Fig. 9(a)). This
 356 mean flow pattern is largely maintained with depth; the flow through the gap
 357 and that rounding Campbell Plateau are connected by the $> 0.1 \text{ m s}^{-1}$ con-
 358 tour to a depth of $\sim 2000 \text{ m}$ (Figs. 9(b)–(d)). The surface variability in the
 359 region is mostly isotropic except in shallow waters, against coastlines, and
 360 above the eastern and western flanks of the Macquarie Island ridge section
 361 and the steeper edges of Campbell Plateau (Fig. 9(a)). The variability be-
 362 comes more anisotropic with depth, especially against the sloping bathymetry
 363 (Figs. 9(b)–(d)). In agreement with observations (e.g., *Rintoul et al.*, 2014),
 364 the flow through the (53.3°S, 159.4°E) gap is predominantly barotropic with
 365 the current speed remaining over 0.1 m s^{-1} beyond 3000 m depth. The eddy
 366 kinetic energy K (not shown here) is a local minimum in the flow through
 367 the gap. Any variability within the gap is isotropic at all depths except
 368 against the ridge flanks. Immediately upstream to the north of the gap, the
 369 variability is highly anisotropic below depths of 1000 m, indicative of eddy-
 370 mean-flow interaction at depth (Fig. 9(c)–(d)). After crossing the Macquarie
 371 Ridge, the current flows along the southeastern edge of the Campbell Plateau
 372 where the bathymetry is steepest. The variability is highly anisotropic along
 373 the Plateau edge at all depths including the surface; this anisotropic feature
 374 is evident in the satellite observations (Fig. 7(a)) and previous studies (e.g.,
 375 *Stanton & Morris*, 2004), and is presumably related to the presence of the
 376 bathymetric slope.

377 5. Discussion

378 There is good agreement between the satellite and model surface distribu-
 379 tions of K , L , and L/K . Both exhibit large values of K in and surrounding
 380 the regions of large flow speed, especially in the western boundary currents,
 381 equatorial currents and the Antarctic Circumpolar Current (ACC). The dis-
 382 tributions of the anisotropic eddy kinetic energy L exhibit more mesoscale
 383 structure than their respective distributions of K ; L appears more local-
 384 ized and confined to the regions of large flow speeds. The distributions of
 385 L/K reveal that the largest surface eddy anisotropy occurs along the coast-
 386 lines and in the vicinity of the equator. The strong eddy anisotropy along
 387 the coastlines can be understood by considering the constraint the coastline
 388 imposes on the onshore flow and its fluctuations: the onshore flow and fluc-
 389 tuations are smaller than the alongshore flow and fluctuations. At and near
 390 the equator, the eddy orientation (not shown here) suggests that the strong
 391 eddy anisotropy is due to a combination of equatorial Kelvin waves elongat-
 392 ing fluctuations zonally and tropical instability waves elongating fluctuations
 393 meridionally.

394 The surface variability fields derived from the satellite altimetry are gen-
 395 erally more isotropic than those of the model. This is likely due to the
 396 reduced spatial resolution of the satellite observations, and perhaps in part
 397 to the objective remapping procedure used to generate a daily $1/4^\circ$ global
 398 product from along-track altimetry measurements. Irrespective of the cause
 399 of this apparent isotropic bias relative to the model, it does suggest that any
 400 anisotropic signals detected by the satellite observations are indeed robust
 401 features of the circulation. Therefore, the common anisotropic signals found
 402 in the AVISO and model surface fields (such as enhanced anisotropy on the
 403 Campbell and Kerguelen Plateaus, Macquarie Ridge, Drake Passage, etc.)
 404 are likely to result from common causes, and as such we expect that the
 405 model provides the means to diagnose such causes.

406 Ocean variability tends to be anisotropic in the vicinity of major bathy-
 407 metric features, such as coastlines, shelfbreaks, seamounts or plateaus, and
 408 this relationship between eddy anisotropy and f/H is worthy of further dis-
 409 cussion. Previous studies have drawn attention to the effect of f/H on
 410 shaping the eddy anisotropy (e.g., *Morrow et al.*, 1994; *Stanton & Morris*,
 411 2004; *Scott et al.*, 2008), and generally conclude that, with the exception of
 412 a few specific regions, the alignment of surface variability is not sensitive to
 413 the direction of underlying isobaths. This is consistent with the analysis of

414 surface variability presented here, especially in regions of weakly barotropic
 415 flows (compare Figs. 7(b) with 7(g)). However, the alignment of sub-surface
 416 variability is found to be strongly influenced by bathymetry; this link be-
 417 tween anisotropic eddy kinetic energy and bathymetry is most apparent in
 418 bottom fluctuations. Although the bottom eddy kinetic energy is about two
 419 orders of magnitude smaller than that at the surface, where the bathymetry
 420 is sloping the bottom variability is almost entirely anisotropic and aligned
 421 with the underlying isobaths. Interestingly, the regions of largest bottom K
 422 coincide with regions of isotropic variability and flat bathymetry unable to
 423 support topographic wave activity, indicative of f -plane turbulence. These
 424 findings are consistent with known dynamics of variability near bathymetry,
 425 such as bottom-trapped Rossby waves (*Rhines*, 1970) and the topostrophic
 426 processes encompassed in the Neptune Effect (*Holloway*, 2008), and can
 427 offer guidance for shaping parameterizations for the unresolved variability
 428 based on the underlying bathymetric slope. Such a parameterization scheme
 429 for bottom variability is especially warranted considering the current ocean
 430 modelling practice for z -coordinate models to reduce vertical resolution with
 431 depth.

432 Further insight into the vertical structure of ocean variability is gained
 433 by partitioning the flow into its barotropic and baroclinic components. The
 434 isobath-aligned anisotropic bottom variability is contained in the barotropic
 435 flow. This makes good physical sense; given that the barotropic flow tends
 436 to follow contours of constant f/H , it should come as no surprise that the
 437 barotropic variability follows suit (compare Figs. 7(c) with 7(g)). On the
 438 other hand, the baroclinic flow is less constrained by f/H contours, and
 439 this is reflected in the finding that bathymetric slope appears to have little
 440 influence on the baroclinic variability (compare Figs. 7(d) with 7(g)).

441 The fundamental difference in the nature of the barotropic and baroclinic
 442 variability suggests there may be benefit in distinct parameterizations for
 443 unresolved barotropic and unresolved baroclinic variability. One approach
 444 could take advantage of the fact that, given the separation of time- and
 445 length-scales between barotropic and baroclinic dynamics, many ocean mod-
 446 els solve these flows separately with a fast two-dimensional barotropic solver
 447 sub-cycled within the slower three-dimensional baroclinic solution (e.g., *Kill-*
 448 *worth et al.*, 1991; *Griffies et al.*, 2001). This separation presents an opportu-
 449 nity to parameterize the unresolved barotropic variability independent of the
 450 baroclinic variability and in a way that reflects its tendency to be anisotropic
 451 and aligned with the underlying isobath. Presumably, a parameterization for

unresolved barotropic variability would be more active in models that over-partition energy to the barotropic and low baroclinic modes, and be relatively idle in ocean models with greater vertical resolution.

Efforts to capture the effect of eddies in ocean models have focused on improving the representation of lateral eddy structures through increased horizontal resolution (e.g., *Hallberg & Gnanadesikan*, 2006), whilst relatively less attention has gone to improving the representation of the vertical structure of the eddy field. The vertical structure of ocean variability presented here highlights the need for increases of the vertical resolution to track with those of the horizontal resolution. Indeed, atmospheric studies have identified substantial benefits to increasing both the horizontal and vertical resolution consistently (*Roeckner et al.*, 2006). It may be the case that ocean models with $1/12^\circ$ horizontal resolution require > 46 vertical levels to resolve the dynamics of the finer lateral structures. For instance, it is obvious that there is little physical benefit to laterally resolving the high-order baroclinic Rossby deformation radii when the vertical resolution only permits low-order baroclinic and barotropic mode dynamics, and vice-versa.

The present study has focused on the horizontal eddy anisotropy to highlight the structure and distribution of the eddy anisotropy globally and as a function of depth. A topic of ongoing work is to incorporate eddy buoyancy flux terms into the analysis. To do this from the perspective of eddy geometry involves the characterization of a 3D variance ellipsoid in momentum-buoyancy eddy flux space in place of the 2D variance ellipse employed here. It is expected that these eddy buoyancy flux terms will also have distinctive vertical structure, and interact with the eddy momentum field primarily through the baroclinic eddies. Additional utility of using the eddy geometry framework to characterize and understand eddy fluxes potentially lies in understanding the time dependence of the eddy geometry, such as seasonality, or in the case of barotropic variability, diurnal/tidal cycles.

6. Conclusion

Quantifying the anisotropy of eddy variability is important as eddy feedbacks on the larger scale flow result solely from the anisotropic components of the eddy variability. Here, the anisotropy of ocean variability has been examined in satellite observations and a $1/12^\circ$ global ocean model. It is found that the anisotropy of eddy variability is largest near the equator, along coastlines and close to bathymetry. Eddy anisotropy is found to have a dis-

Table 1: Global averages of K ($\text{m}^2 \text{s}^{-2}$), L ($\text{m}^2 \text{s}^{-2}$), and L/K for surface geostrophic velocities derived from the AVISO product and for the surface, bottom and depth-averaged ORCA12 model output.

Term	AVISO	ORCA12		
		Surface	Bottom	All
K	2.34e-2	2.42e-2	5.57e-4	2.17e-3
L	6.37e-3	8.81e-3	3.12e-4	6.90e-4
L/K	0.245	0.294	0.644	0.418

488 tinctive vertical structure and become increasingly anisotropic with depth,
 489 especially in the vicinity of sloping bathymetry. The highly anisotropic bot-
 490 tom variability tends to be aligned with the underlying isobaths, which is
 491 useful knowledge given that eddy-mean-flow interactions are sensitive to the
 492 orientation of the eddy anisotropy, and in particular the extent to which
 493 the eddy anisotropy orientation departs from the direction of the mean-flow
 494 shear. Partitioning the eddy variability into the barotropic and baroclinic
 495 components reveals that the anisotropic bottom variability is predominantly
 496 barotropic, whilst the upper-ocean variability is relatively baroclinic. Given
 497 the distinct natures of barotropic and baroclinic variability, the findings sug-
 498 gest that parameterizations for unresolved variability in ocean models may be
 499 improved by the development of a parameterization for unresolved barotropic
 500 Reynolds stresses that is based on the shape of the underlying bathymetry
 501 and operates only on the barotropic flow.

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Table 2: Southern Ocean (south of 30°S) averages of K ($\text{m}^2 \text{s}^{-2}$), L ($\text{m}^2 \text{s}^{-2}$), and L/K for the surface, bottom and depth-averaged ORCA12 model output. These terms are calculated for the Southern Ocean using (i) the full flow, (ii) only the barotropic flow, and (iii) only the baroclinic flow of the ORCA12 model output.

Term	AVISO	ORCA12 Southern Ocean			ORCA12 Barotropic	ORCA12 Baroclinic		
		Surface	Bottom	All		Surface	Bottom	All
K	1.53e-2	1.61e-2	8.14e-4	2.82e-3	1.72e-3	9.01e-3	8.40e-4	1.09e-3
L	3.13e-3	3.46e-3	3.65e-4	6.86e-4	4.28e-4	1.84e-3	2.27e-4	2.41e-4
L/K	0.218	0.222	0.588	0.361	0.277	0.205	0.314	0.262

Figure 1: Distributions of (a) K ($\text{m}^2 \text{s}^{-2}$), (b) L ($\text{m}^2 \text{s}^{-2}$), and (c) L/K derived from the 1/4° AVISO product of surface geostrophic velocities for the period of 2003–2012. The 0.22 m s^{-1} time-mean speed contour is shown in grey.

Figure 2: Surface distributions of (a) K ($\text{m}^2 \text{s}^{-2}$), (b) L ($\text{m}^2 \text{s}^{-2}$), and (c) L/K derived from the surface velocities of the 1/12° ORCA12 ocean model for the period of 2003–2012. The 0.22 m s^{-1} time-mean speed contour is shown in grey.

Figure 3: Bottom distributions of (a) K ($\text{m}^2 \text{s}^{-2}$), (b) L ($\text{m}^2 \text{s}^{-2}$), and (c) L/K derived from the horizontal velocities of the second-deepest grid cell of the 1/12° ORCA12 ocean model for the period of 2003–2012. The 0.1 m s^{-1} time-mean speed contour is shown in grey.

Figure 4: Zonal averages of L/K south of $\sim 30^\circ \text{N}$ derived from the ORCA12 horizontal velocities for (a) all regions, (b) regions over sloping model bathymetry ($\nabla H \neq 0$, where H is the local model ocean depth), and (c) regions over flat model bathymetry ($\nabla H = 0$). The 0.02 m s^{-1} zonal average of the time-mean zonal velocity is contoured (solid eastward, dashed westward).

Figure 5: Circumpolar transects of (a) L/K and (b) Θ derived from the ORCA12 horizontal velocities at 55°S . The 0.1 m s^{-1} time-mean speed contour is shown in black. Locally flat model bathymetry ($\nabla H = 0$) is indicated in white. Note that for Θ , north (red) is equivalent to south (blue), however for zonal mean-flow it is useful to identify regions where the eddy anisotropy orientation transitions through the north-south direction, hence the non-periodic colorbar.

Figure 6: Histograms of the angle between Θ and the local isobath for the (a) surface and (b) bottom of ORCA12 anisotropic variability in regions of sloping model bathymetry ($\nabla H \neq 0$). The dashed lines represent uniform distributions (i.e. what would be expected if there were no connection between Θ and the local isobath). The local maxima in bottom eddy orientation at $\pi/4$ is related to Ekman boundary layer dynamics; the second-lowest grid cell is occasionally entirely contained within the lower Ekman layer.

Figure 7: Surface distributions of Southern Ocean L/K derived from (a) AVISO surface geostrophic velocities, (b) ORCA12 surface layer velocities, (c) the ORCA12 barotropic velocities, and (d) the ORCA12 baroclinic velocities; the 0.22 m s^{-1} time-mean speed contour is shown in grey. The distributions of Southern Ocean L/K derived from (e) bottom ORCA12 velocities, and (f) the bottom ORCA12 baroclinic velocities; the 0.1 m s^{-1} time-mean speed contour is shown in grey. The (g) distribution of ORCA12 Southern Ocean $|\nabla(f/H)|$ ($\text{m}^{-2}\text{s}^{-1}$).

Figure 8: Circumpolar transect of L/K at 55°S derived from the baroclinic velocities of ORCA12. The 0.1 m s^{-1} time-mean speed contour is shown in black. Locally flat model bathymetry ($\nabla H = 0$) is indicated in white. The intermediate-depth maxima in L/K relates to the depth where the current speed is equal to the barotropic speed, so that the baroclinic flow is zero.

Figure 9: Distributions of L/K in the vicinity of Macquarie Ridge and Campbell Plateau at (a) the surface, (b) 1000 m, (c) 2000 m and (d) 3000 m depth. The 1000, 2000 and 3000 m model isobaths are contoured in magenta. The 0.2 m s^{-1} time-mean speed contour is contoured in black for the surface (a); the 0.1 m s^{-1} time-mean speed contour is contoured for other depths (b–d).

514 Bibliography

- 515 Brodeau, L., B. Barnier, A.-M. Treguier, T. Penduff, and S. Gulev (2010),
 516 An ERA40-based atmospheric forcing for global ocean circulation models.
 517 *Ocean Modelling*, *31*, 88–104. DOI: 10.1016/J.OCEMOD.2009.10.005.
- 518 Chelton, D. B., M. G. Schlax, and R. M. Samelson (2011), Global observa-
 519 tions of nonlinear mesoscale eddies. *Progress in Oceanography*, *91*, 167–
 520 216. DOI: 10.1016/J.POCEAN.2011.01.002.
- 521 Ducet, N., P.-Y. Le Traon, and G. Reverdin (2000), Global high-
 522 resolution mapping of ocean circulation from TOPEX/Poseidon and ERS-
 523 1 and -2. *Journal of Geophysical Research*, *105*, 19477–19498. DOI:
 524 10.1029/2000JC900063.
- 525 Ducet, N., and P.-Y. Le Traon (2001), A comparison of surface eddy ki-
 526 netic energy and Reynolds stresses in the Gulf Stream and the Kuroshio
 527 Current Systems from merged TOPEX/Poseidon and ERS-1/2 altimet-
 528 ric data. *Journal of Geophysical Research*, *106*, 16603–16622. DOI:
 529 10.1029/2000JC000205.
- 530 Dufour, C. O., J. Le Sommer, M. Gehlen, J. C. Orr, J.-M. Molines, J. Simeon,
 531 and B. Barnier (2013), Eddy compensation and controls of the enhanced
 532 sea-to-air CO₂ flux during positive phases of the Southern Annular Mode.
 533 *Global Biogeochemical Cycles*, *27*, 950–961. DOI: 10.1002/GBC.20090.
- 534 Durand, M., L.-L. Fu, D. P. Lettenmaier, D. E. Alsdorf, E. Rodrigues, and D.
 535 Esteban-Fernandez (2010), The Surface Water and Ocean Topography mis-
 536 sion: Observing terrestrial surface water and oceanic submesoscale eddies.
 537 *Proceedings of IEEE*, *98*, 766–779. DOI: 10.1109/JPROC.2010.2043031.
- 538 Eden, C. (2010), Parameterising meso-scale eddy momentum fluxes based on
 539 potential vorticity mixing and a gauge term. *Ocean Modelling*, *32*, 58–71.
 540 DOI: 10.1016/J.OCEMOD.2009.10.008.
- 541 Fox-Kemper, B., R. Lumpkin, and F. O. Bryan (2013), Lateral transport
 542 in the ocean interior. *Ocean Circulation and Climate: A 21st Century*
 543 *Perspective*, International Geophysical Series, *103*, Elsevier, 185–210. DOI:
 544 10.1016/B978-0-12-391851-2.00008-8.

- 545 Fu, L.-L., D. B. Chelton, P.-Y. Le Traon, and R. Morrow (2010),
 546 Eddy dynamics from satellite altimetry. *Oceanography*, *23*, 14–25. DOI:
 547 10.5670/OCEANOG.2010.02.
- 548 Gent, P. R. and J. C. McWilliams (1990), Isopycnal mixing in ocean circu-
 549 lation models. *Journal of Physical Oceanography*, *20*, 150–155.
- 550 Gent, P. R. and J. C. McWilliams (1996), Eliassen-Palm Fluxes and the mo-
 551 mentum equation in non-eddy-resolving ocean circulation models. *Journal*
 552 *of Physical Oceanography*, *26*, 2539–2546.
- 553 Griffies, S. M., R. Pacanowski, M. Schmidt, and V. Balaji (2001), Tracer
 554 conservation with an explicit free surface method for z-coordinate ocean
 555 models. *Monthly Weather Review*, *129*, 1081–1098.
- 556 Hallberg, R. and A. Gnanadesikan (2006), The role of eddies in determin-
 557 ing the structure and response of the wind-driven southern hemisphere
 558 overturning: results from the Modelling Eddies in the Southern Ocean
 559 (MESO) Project. *Journal of Physical Oceanography*, *36*, 2232–2252. DOI:
 560 10.1175/JPO2980.1.
- 561 Hallberg, R. (2013), Using a resolution function to regulate parameteriza-
 562 tions of oceanic mesoscale eddy effects. *Ocean Modelling*, *72*, 92–103. DOI:
 563 10.1016/J.OCEMOD.2013.08.007.
- 564 Holloway, G. (2008), Observing global ocean topostrophy. *Journal of Geo-*
 565 *physical Research*, *113*, C07054. DOI: 10.1029/2007JC004635.
- 566 Hoskins, B. J., I. N. James, and G. H. White (1983), The shape, propagation
 567 and mean-flow interaction of large-scale weather systems. *Journal of the*
 568 *Atmospheric Sciences*, *40*, 1595–1612.
- 569 Huang, H.-P., A. Kaplan, E. N. Curchitser, and N. A. Maximenko (2007),
 570 The degree of anisotropy for mid-ocean currents from satellite observa-
 571 tions and an eddy-permitting model simulation. *Journal of Geophysical*
 572 *Research*, *112*, C09005, DOI: 10.1029/2007JC004105.
- 573 Killworth, P. D., D. Stainforth, D. J. Webb, and S. M. Paterson (1991), The
 574 development of a free-surface Bryan-Cox-Semtner ocean model. *Journal of*
 575 *Physical Oceanography*, *21*, 1333–1348.

- 576 Madec, G. (2008), NEMO Ocean Engine. *Note du Pôle de Modélisation de*
577 *l’Institut Pierre-Simon Laplace*, 27.
- 578 Marshall, J. C. (1984), Eddy-mean-flow interaction in a barotropic ocean
579 model. *Quarterly Journal of the Royal Meteorological Society*, 110, 573–
580 590.
- 581 Marshall, D. P., J. R. Maddison, and P. S. Berloff (2012), A framework for pa-
582 rameterizing eddy potential vorticity fluxes. *Journal of Physical Oceanog-*
583 *raphy*, 42, 539–557. DOI: 10.1175/JPO-D-11-048.1.
- 584 McWilliams, J. C. (2008), The nature and consequences of oceanic eddies.
585 *Geophysical Monograph Series*, 177, 5–15. DOI: 10.1029/177GM03.
- 586 Morrow, R., R. Coleman, J. Church, and D. Chelton (1994), Surface eddy
587 momentum flux and velocity variances in the Southern Ocean from Geosat
588 altimetry. *Journal of Physical Oceanography*, 24, 2050–2071.
- 589 Pedlosky, J. (1987), Geophysical Fluid Dynamics, Second Edition. *Springer-*
590 *Verlag*.
- 591 Plumb, R. A. (1986), Three-dimensional propagation of transient quasi-
592 geostrophic eddies and its relationship with the eddy forcing of the time-
593 mean flow. *Journal of the Atmospheric Sciences*, 43, 1657–1678.
- 594 Preisendorfer, R. W. (1988), Principal component analysis in meteorology
595 and oceanography. *Elsevier*, 425 pp.
- 596 Rhines, P. (1970), Edge-, bottom-, and Rossby waves in a rotating stratified
597 fluid. *Geophysical Fluid Dynamics*, 1, 273–302.
- 598 Rintoul, S. R. (2010), Antarctic Circumpolar Current. *Ocean Currents: A*
599 *Derivative of the Encyclopedia of Ocean Sciences*, 196.
- 600 Rintoul, S. R., S. Sokolov, M. J. M. Williams, B. Peña Molino, M. Rosenberg,
601 and N. L. Bindoff (2014), Antarctic Circumpolar Current transport and
602 barotropic transition at Macquarie Ridge. *Geophysical Research Letters*,
603 41, 7254–7261. DOI: 10.1002/2014GL061880.
- 604 Roeckner, E., R. Brokopf, M. Esch, M. Giorgetta, S. Hagemann, L. Ko-
605 rnblueh, E. Manzini, U. Schlese, and U. Schulzweida (2006), Sensitiv-
606 ity of simulated climate to horizontal and vertical resolution in the

607 ECHAM5 atmosphere model. *Journal of Climate*, *19*, 3771–3791. DOI:
608 10.1175/JCLI3824.1.

609 Scott, R. B., B. K. Arbic, C. L. Holland, A. Sen, and B. Qiu (2008), Zonal
610 versus meridional velocity variance in satellite observations and realistic
611 and idealized ocean circulation models. *Ocean Modelling*, *23*, 102–112.
612 DOI: 10.1016/J.OCEMOD.2008.04.009.

613 Stanton, B. R. and M. Y. Morris (2004), Direct velocity measurements
614 in the Subantarctic Front and over Campbell Plateau, southeast of
615 New Zealand. *Journal of Geophysical Research*, *109*, C01028. DOI:
616 10.1029/2002JC001339.

617 Starr, V. P. (1968), Physics of Negative Viscosity Phenomena. *McGraw-Hill*,
618 New York, 256pp.

619 The DRAKKAR Group (2007), Eddy-permitting ocean circulation hindcasts
620 of past decades. *CLIVAR Exchanges*, *12*, 3.

621 Treguier, A.-M., J. Deshayes, J. Le Sommer, C. Lique, G. Madec, T. Penduff,
622 J.-M. Molines, B. Barnier, R. Bourdalle-Badie, and C. Talandier (2014),
623 Meridional transport of salt in the global ocean from an eddy-resolving
624 model. *Ocean Science*, *10*. DOI: 10.5194/OS-10-243-2014.

625 Waterman, S., N. G. Hogg, and S. R. Jayne (2011), Eddy-mean flow interac-
626 tions in the Kuroshio Extension region. *Journal of Physical Oceanography*,
627 *41*, 1182–1208. DOI: 10.1175/2010JPO4564.1.

628 Waterman, S. and B. J. Hoskins (2013), Eddy shape, orientation, propaga-
629 tion, and mean flow feedback in western boundary current jets. *Journal of*
630 *Physical Oceanography*, *43*, 1666–1690. DOI: 10.1175/JPO-D-12-0152.1.

631 Waterman, S. and J. M. Lilly (2015), A geometric decomposition of eddy
632 feedbacks in barotropic systems. *Journal of Physical Oceanography*, *45*,
633 1009–1024. DOI: 10.1175/JPO-D-14-0177.1.

634 Wunsch, C. (2007), The past and future ocean circulation from a contempo-
635 rary perspective. *Ocean Circulation: Mechanisms and Impacts, Geophys-*
636 *ical Monograph Series*, *173*, American Geophysical Union, 53–74. DOI:
637 10.1029/173GM06.

Figure 1a

Figure 1b

Figure 1c

Figure 2a

Figure 2b

Figure 2c

Figure 3a

Figure 3b

Figure 3c

Figure 4a

Figure 4b

Figure 4c

Figure 4z

Figure 5a

Figure 5b

Figure 6a

Figure 6b

Figure 7a

Figure 7b

Figure 7c

Figure 7d

Figur 7e

Figure 7f

Figure 7g

Figure 7z1

Figure 7z2

Figure 8

Figure 9

Figure 9z