

Seasonality of submesoscale flows in the ocean surface boundary layer

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Key Points

- Submesoscale relative vorticity (RV) is estimated for approximately one year in the North-east Atlantic
- Variance and skewness of RV within the upper ocean are enhanced during winter
- Observations are consistent with seasonally varying submesoscale turbulence.

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A signature of submesoscale flows in the upper ocean is skewness in the distribution of relative vorticity. Expected to result for high Rossby-number flows, such skewness has implications for mixing, dissipation and stratification within the upper ocean. An array of moorings deployed in the North-east Atlantic for one year as part of the OSMOSIS experiment reveals that relative vorticity is positively skewed during winter even though the scale

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of the Rossby number is less than 0.5. Furthermore, this skewness is reduced to zero during spring and autumn. There is also evidence of modest seasonal variations in the gradient Rossby number. The proposed mechanism by which relative vorticity is skewed is that the ratio of lateral to vertical buoyancy gradients, as summarized by the inverse gradient Richardson number, restricts its range during winter but less so at other times of the year. These results support recent observations and model simulations suggesting the upper ocean is host to a seasonal cycle in submesoscale turbulence.

1. Introduction

The ocean surface boundary layer (OSBL) is an important component of Earth's climate system. It serves as the medium through which heat, momentum and gases (e.g., carbon dioxide) are exchanged between the ocean and atmosphere through a combination of physical, chemical and biological processes. Prediction of climate on seasonal, decadal and longer time scales therefore requires an accurate knowledge of the dynamics within this layer.

Wind, waves, precipitation and heat loss at the ocean surface have long been known to have a significant effect on vertical stratification within the OSBL and have successfully been modeled as one-dimensional processes (see *Large et al.* [1994] and references therein). More recently, however, a host of three-dimensional processes have been identified as influencing both stratification and dissipation in the upper ocean. These include (1) effects of Coriolis-Stokes forcing and Langmuir turbulence, (2) circulation due to frontogenesis/frontolysis and (3) forced and unforced instabilities [*Hoskins*, 1982; *Thomas*, 2005; *Polton et al.*, 2005; *Polton and Belcher*, 2007; *Boccaletti et al.*, 2007; *Thomas et al.*, 2008; *Fox-Kemper et al.*, 2008a; *Taylor and Ferrari*, 2009; *Belcher et al.*, 2012; *D'Asaro et al.*, 2011; *Hamlington et al.*, 2014; *Brüggemann et al.*, 2015]. The dynamical regime in which these processes reside is the oceanic submesoscale.

1.1. The oceanic submesoscale

The oceanic submesoscale is comprised of motions with horizontal scales of 0.1–10 km and evolving over time scales near the inertial period, $2\pi/f$. While generally characterized by $O(1)$ gradient Rossby and gradient Richardson numbers, $Ro = \zeta/f$ and

$Ri = N^2/|\partial_z \mathbf{u}_h|^2$, respectively [Thomas et al., 2008], implying large horizontal and vertical shears, this need not be the case. In these expressions, f is the vertical component of the Earth's rotation (i.e., the Coriolis parameter), $\zeta = (\nabla \times \mathbf{u}) \cdot \hat{\mathbf{k}}$ is the vertical component of relative vorticity, $\mathbf{u} = (u, v, w)$ is the total fluid velocity, $\mathbf{u}_h = (u, v)$ is the horizontal fluid velocity, $N^2 = \partial_z b$ is vertical stratification, $b = g(\rho_o - \rho)/\rho_o$ is buoyancy, g is gravity, ρ is density and ρ_o is a reference density.

1.2. Skewness in the distribution of relative vorticity

One of the hallmarks of submesoscale flows is skewness in the distribution of relative vorticity, ζ . This skewness is more common for submesoscale flows compared with motions at the mesoscale (30–200 km) since the Rossby number is considerably smaller at the mesoscale. Both Rudnick [2001] and Shcherbina et al. [2013] document positive skewness in winter from observed distributions of ζ estimated from shipboard velocity measurements, and similar skewness has been reported by Roulet and Klein [2010] and Qiu et al. [2014] in high-resolution ocean models. The simplest dynamical explanation for this observed asymmetry is that small perturbations grow exponentially when potential vorticity (PV) takes on sign opposite the Coriolis parameter, $f q < 0$, where

$$q = (f + \zeta)N^2 + \omega_h \cdot \nabla_h b \quad (1)$$

is the Ertel PV [Ertel, 1942; Hoskins, 1974]. Here, $(f + \zeta)$ is the vertical component of absolute vorticity, ω_h is the horizontal vorticity vector, N^2 is defined above and $\nabla_h b$ is the horizontal buoyancy gradient. Because relative vorticity is one factor determining q and since instabilities tend to drive $f q$ toward a positive state, these instabilities restrict the range of observed values of ζ . In particular, extreme negative values of ζ are rare in the

Northern Hemisphere and positive values of ζ are rare in the Southern Hemisphere. This results in a *positively* skewed probability density function (PDF) of $Ro = \zeta/f$. Note that increased variance in Ro resulting from vortex stretching will broaden the distribution of Ro and help reveal this skewness [Capet *et al.*, 2008].

1.3. Focus of this study

There are few open-ocean measurements of submesoscale Ro . Those of Rudnick [2001] are biased low owing to an approximation of ζ from along-track gradients alone, and Shcherbina *et al.* [2013] document Ro in a region influenced by the mean flow of the Gulf Stream. Furthermore, in most cases, existing observations are limited to wintertime conditions in which, based on modelling [Mensa *et al.*, 2013; Qiu *et al.*, 2014; Sasaki *et al.*, 2014; Brannigan *et al.*, 2015] and observational [Ostrovskii, 1995; Callies *et al.*, 2015] studies, we expect submesoscale flows to be most active.

In this study, we extend the observational efforts of Rudnick [2001] and Shcherbina *et al.* [2013] by estimating $Ro = \zeta/f$ from measurements spanning an annual cycle. We conduct this analysis on observations from the Northeast Atlantic, a region characterized by weak mean kinetic energy and low-to-moderate mesoscale eddy kinetic energy (Supporting Material, Figure S1). We demonstrate that relative vorticity within the OSBL is positively skewed during winter even though a characteristic scale for $|Ro|$ is less than 0.5 at all times. Additionally, this skewness appears to vary seasonally, a result that has implications for stability. In summary, we (i) present rare observational evidence of submesoscale flows in the open ocean, (ii) provide a scale for $|Ro|$, and (iii) highlight seasonal changes in turbulence parameters with implications for stability.

2. Methods

2.1. Observations within the OSBL

The Ocean Surface Mixing, Ocean Submesoscale Interaction Study (OSMOSIS) is a research consortium formed to understand dynamics within the OSBL. An important component of the project is a year-long observational campaign conducted during September 2012–September 2013 in the North Atlantic (48.69°N, 16.19°W, Figure 1a) [Allen *et al.*, 2013]. Measurements include temperature, salinity and horizontal velocity from moorings (50–500 m) and hydrographic observations from two ocean gliders (0–1000 m). The present study focuses on mooring observations.

Two clusters of moorings were oriented around a central mooring. One cluster consisted of four moorings spaced approximately 1.5 km from this center mooring, while the other cluster consisted of four moorings spaced approximately 10 km from this same mooring (Figure 1b). We refer to regions spanned by these moorings as inner and outer mooring domains, respectively, and focus attention on inner moorings.

MicroCAT conductivity-temperature-depth (CTD) sensors and acoustic current meters (ACMs) were spaced approximately 30–100 m in the vertical, with increased numbers on the center mooring in order to resolve stratification and shear (Supporting Material, Figure S2). Sampling intervals of CTDs and ACMs were 5 and 10 minutes, respectively. Velocity, temperature and salinity at each mooring were interpolated to common depths and times, and smoothing was applied in the vertical and time (30 m and 2 hr) in order to minimize noise while retaining time and spatial scales of interest.

In an effort to prevent damage by severe weather throughout the year, OSMOSIS moorings resided below the ocean surface. This precluded use of a Global Positioning System

(GPS) and, owing to their considerable cost, acoustic positioning sensors were not used. For these reasons, knowledge of mooring position is somewhat limited. In the work that follows, we compute vorticity from neighboring moorings and, therefore, require accurate position information. To quantify the error introduced by position uncertainty, we modeled the distance between moorings as the sum of fixed distances and a stochastic parameter that represents perturbations from these constant distances. Fixed distances were estimated from anchor positions and perturbations were modeled as random variables with zero mean and standard deviations determined from the time-integration of horizontal currents (Appendix). A Monte Carlo approach was then used to quantify the range of uncertainty. Such an approach provides an upper bound on errors due to mooring motion since it effectively assumes instruments drift as unconstrained floats.

2.2. Vorticity from the inner moorings

We divided the inner mooring domain into four triangular regions in order to explore submesoscale characteristics described above. Relative vorticity was estimated from circulation following Stokes' theorem [*Batchelor*, 2000]:

$$\zeta = (\nabla \times \mathbf{u}) \cdot \hat{\mathbf{k}} = \frac{1}{A} \oint \mathbf{u}_h \cdot d\mathbf{l}. \quad (2)$$

Here, A is the area enclosed by the path, \mathbf{u}_h is the horizontal fluid velocity and $d\mathbf{l}$ is a differential pointing in the direction of the closed path. For the present case, our closed path is the area formed by a single triangle (cf. Figure 1b). We estimated the horizontal velocity, \mathbf{u}_h , as the average of velocities between two neighboring moorings and $|d\mathbf{l}|$ as the distance between these moorings. Alternatively, the vertical component of relative

vorticity can be estimated from velocity gradients, yielding numerically equivalent values [Bryden and Fofonoff, 1977].

From the above expression, $Ro = \zeta/f$ was computed ($f = 1.094 \times 10^{-4} \text{ s}^{-1}$) and the standard deviation and skewness of Ro were estimated as functions of depth and season.

The choice of depth and season are discussed below. We combined vorticity estimates from all four inner triangles in order to obtain robust statistics. We also computed seasonal profiles of stratification and gradient Richardson number from buoyancy and shear at the center mooring. Note, as Ri is approximately log-normal, averaging must be done in logged coordinates (before transforming it back to a characteristic value). Also, because the mean of Ro is approximately zero, the standard deviation of Ro provides a characteristic scale for the Rossby number.

We have included estimates of vertical stratification from ocean gliders [Thompson *et al.*, 2016] (see also *G. M. Damerell et al.*, submitted, 2016) to support those from moorings. These instruments sampled the water column at approximately 2–4-m vertical resolution throughout the entire year and, thus, provide a check on stratification from moorings.

2.3. Defining depths and seasons

Mooring measurements made within the OSBL, and where we expect submesoscales to be most active, do not extend for the entire observational record. This is most easily discerned from two-dimensional histograms of observation number as a function of time and normalized depth, $|z|/h$ (not shown). Here, $|z|$ is the observation depth and h is the mixed layer (ML) depth, defined as the depth at which potential density exceeds that observed at 20 m by 0.02 kg m^{-3} [de Boyer Montégut *et al.*, 2004]. We used these histograms

to define the following two seasons: winter (20 December–23 April) and autumn/spring (hereafter, non-winter; 11 October–19 December and 24 April–3 June). Additionally, to facilitate comparison across seasons, we have computed statistics in $|z|/h$ -coordinates.

Here, h has been smoothed using a 12-day window to eliminate intermittent changes in ML depth. With a few exceptions (e.g., early February and May), h tracks the depth of the pycnocline well (Figure 1c). We argue that this depth most appropriately characterizes submesoscale vortices since baroclinic instability (BCI) of the surface ocean is inherently tied to a local but non-intermittent index of ML depth [Stone, 1966; Boccaletti *et al.*, 2007; Fox-Kemper *et al.*, 2008b; Grooms, 2015].

3. Results

Year-long records of stratification, N/f , from ocean gliders and normalized relative vorticity, ζ/f , within the inner mooring domain are shown in Figures 1c and 1d. Overlaid on these are ML depth and contours of constant $|z|/h$. Note that Ro has values on the order of one but is generally smaller. Additionally, a number of short-duration (i.e., < 3 days) anomalies propagate through the inner mooring domain over the observational period, suggesting a sufficient number of samples for statistical estimates.

In Figure 2, we display statistics of stratification, vorticity and gradient Richardson number. We also highlight the depth of peak stratification, $|z|/h = 1.3$, and refer to locations above this depth as being *within the OSBL*.

Representative stratification profiles depict a strong seasonal cycle (Figure 2a). Non-winter values are as high as $N/f = 40$ while winter values fall below $N/f = 15$. This

pattern is reproduced by glider measurements. Higher non-winter magnitudes are present within the OSBL owing to the glider's ability to measure closer to the ocean surface.

Standard deviations of ζ/f also reveal seasonal differences (Figure 2b). Differences are pronounced within the OSBL but there is still evidence of seasonality immediately below the OSBL. Standard deviations of ζ/f within the OSBL are approximately 0.4 during winter and 0.3 during non-winter. These serve as scales for the Rossby number. Also noteworthy, the standard deviation of ζ/f decreases with increasing depth during both periods, supporting concepts of surface-intensified flow.

Similarly, there are seasonal differences observed in the skewness of ζ/f (Figure 2c). Skewness within the OSBL exceeds 1.0 during winter but is virtually non-existent during non-winter. The existence of positive skewness at depth below the OSBL during non-winter can be traced to anomalously strong cyclonic flow below the OSBL during late November and early December (cf. Figure 1d).

Finally, the logged-gradient Richardson number also varies seasonally (Figure 2d), with winter magnitudes close to 2 and non-winter magnitudes of approximately 6. Such seasonality has also been documented in a recent study involving ocean gliders [*Thompson et al.*, 2016].

4. Seasonality in vorticity skewness

The distribution of $Ro = \zeta/f$ is positively skewed within the OSBL during winter despite that a scale for the Rossby number is 0.5. What gives rise to vorticity skewness if, strictly-speaking, $|Ro|$ is not $O(1)$? To address this question, let us revisit the stability criterion. First, in the absence of lateral buoyancy gradients and assuming pos-

itive stratification, the condition for stable flow becomes, $f q > 0 \rightarrow f^2(1 + \text{Ro})N^2 > 0$, or $\text{Ro} > -1$. This is the well-known bound on Ro for centrifugally-stable flow. Second, assuming thermal wind balance and $w \ll u, v$, one can rewrite Eq. 1 as $q = (f + \zeta)N^2 + (-\partial_z v, \partial_z u) \cdot \nabla_h b \approx (f + \zeta)N^2 - |\nabla_h b|^2/f > 0$. In this case, the condition for stable flow simplifies to $\text{Ro} > -1 + \text{Ri}^{-1}$, where $\text{Ri} = f^2 N^2 / |\nabla_h b|^2$ is the balanced gradient Richardson number. Instability occurring for values less than $-1 + \text{Ri}^{-1}$ is referred to as symmetric instability [Hoskins, 1974; Colin de Verdière, 2012]. This expression states that the PDF of ζ is influenced, not only by its value relative to $-f$ but also by the ratio of lateral and vertical buoyancy gradients. This could explain why we observe vorticity skewness for only moderate $|\text{Ro}|$. Ignoring for the moment small seasonal changes in Ro , it also provides a plausible explanation for the observed seasonality in vorticity skewness, as illustrated in Figure 3.

5. Discussion

The ocean surface is subject to atmospheric forcing in the form of winds, precipitation and heat. Such forcing, particularly strong during winter, alters PV and can drive it toward zero [Thomas and Lee, 2005; Thomas, 2005]. In this case, q is not conserved and a host of instabilities can develop [Stone, 1966; Hoskins, 1974; Haine and Marshall, 1998; Taylor and Ferrari, 2009; Thomas et al., 2013; Lazar et al., 2013]. Note that BCI [Boccaletti et al., 2007] is not included here since it can occur for positive $f q$ [Stone, 1966].

The results presented above depict an active OSBL consistent with this interpretation. Decreased stratification in winter permits enhanced vertical motion. Through a combination of vortex stretching due to frontogenesis [Hoskins, 1982], stretching due to interfacial

waves at the base of the OSBL and vortex generation by baroclinic and shear instabilities [Stone, 1966; Munk *et al.*, 2000; Eldevik and Dysthe, 2002; Boccaletti *et al.*, 2007; Grooms, 2015], the standard deviation of $Ro = \zeta/f$ increases. Finally, as $|Ro|$ increases and Ri decreases, the lower bound associated with symmetric instability emerges.

Increased variance in Ro due to vortex stretching has the effect of revealing vorticity skewness even in the absence of instabilities [Capet *et al.*, 2008]. Thus, whether or not the marginally-stable state (i.e., $fq = 0$) is crossed, leading to the suite of instabilities referred to above, and to what degree wind forcing and buoyancy loss at the surface are relevant factors [Thomas and Lee, 2005; Thomas, 2005; Taylor and Ferrari, 2010; Thomas and Taylor, 2014; Hamlington *et al.*, 2014; Whitt and Thomas, 2015], are questions that we leave for a future study.

Limitations and sources of error

Results of this study depend upon observations within a single year (September 2012–September 2013). Furthermore, owing to the minimum observable depth of moorings, the number of measurements made throughout the year is restricted to those in late Autumn through early Spring; that is, we are unable to comment on summertime statistics.

While intermittency presents a challenge, we can nevertheless estimate confidence levels on statistics. To do this, the effective sample number, n , was estimated as the total duration of time a given normalized depth was observed divided by $t_d = 3$ days. (The decorrelation time scale, t_d , of u and v is one day but we have opted for a more conservative value.) Standard errors of each statistic were then computed and used to determine the

significance of results. In all cases, the results given above remain unchanged (Supporting Material, Figure S3).

Another limitation pertains to the mooring array. Owing to the fixed spacing of moorings, we are unable to distinguish between (a) seasonally-varying amplitudes and (b) seasonally-varying *scales* of submesoscale phenomena. In particular, during winter, the ML deformation radius is approximately $\lambda_d = (N/f)h = 4$ km, while during summer months, it can be closer to $\lambda_d = 1-2$ km. In fact, this is a problem inherent in all studies thus far [Mensa *et al.*, 2013; Qiu *et al.*, 2014; Sasaki *et al.*, 2014; Brannigan *et al.*, 2015; Callies *et al.*, 2015] as their horizontal spacings only marginally resolve the seasonally-varying deformation radius.

Uncertainty has been introduced into this study as a result of mooring motion. Such motion gives rise to two types of errors. The first is additional variance inserted into ζ/f . The second is that observed velocities will be smaller than true velocities if moorings are advected with the flow. Consequently, vorticity (cf. Eq. 2) will be biased low and standard deviations of ζ/f reported here might be biased low. Note, however, that these two types of errors tend to compensate one another.

6. Conclusions

There is increasing evidence from modeling studies [Mensa *et al.*, 2013; Qiu *et al.*, 2014; Sasaki *et al.*, 2014; Brannigan *et al.*, 2015] and observational studies [Ostrovskii, 1995; Callies *et al.*, 2015; Thompson *et al.*, 2016] that submesoscale flows undergo a seasonal cycle. The observations presented here support this concept by depicting seasonal changes in gradient Rossby and Richardson numbers within the OSBL: (i) variances of Ro are

enhanced during winter months over non-winter months, (ii) Ro is positively skewed during winter months while approximately zero at other times of the year and (iii) we observe seasonal changes in Ri , consistent with seasonal changes in vorticity skewness. These observations have potential implications for mixing, stratification and dissipation within the OSBL with application for modelers seeking to represent submesoscale processes in climate-scale ocean models. Note, for example, that symmetric instability has yet to be parameterized in these models.

An important finding of this study is that submesoscale $|Ro|$ is approximately 0.3–0.45 in the Northeast Atlantic. These values are smaller than those observed near the Gulf Stream (e.g., ~ 0.7 –1.0) [*Shcherbina et al.*, 2013] and suggest a different balance of terms in the momentum equations. This may find wider application, for example, in other open-ocean regions where both mean and eddy kinetic energies are smaller.

Appendix A: Mooring motion

As mentioned in section 2.1, we have limited knowledge of the relative positions of moorings in time. Here, we quantify the error produced by such uncertainty using a Monte Carlo simulation.

A1. Mooring design

Considerable buoyancy was placed on mooring cables between 0- and 500-m depth so that the moorings stood as close to vertical as possible within the OSBL. This created an effective pivot point within the catenary at 600 m as evidenced by mooring design simulations (personal comm., P. Provost, 2014). In particular, the depth-integrated flow

determines the average positions of moorings but surface-intensified flows appear to perturb the positions of moorings with respect to one another.

A2. Coherence of motion at the inner moorings

Under the assumption that surface currents determine the relative position of moorings, we estimated the coherence and phase of \mathbf{u}_h between the center and each of the inner moorings (Supporting Material, Figures S4a and S4b). Horizontal velocities in the upper ocean are coherent for time scales greater than 12 hr (i.e., mean squared coherence, $|\gamma|^2 \geq 0.6$) with motions less than 6 hr being considerably incoherent (i.e., $|\gamma|^2 < 0.2$). This is also reflected in the phase, which becomes noisy for periods less than 6 hr but is well-defined for longer periods. While in part consisting of fluctuations from wind and waves, these shorter-period motions contain the signal of interest. Low-pass filtering the signal is therefore imprudent and we instead model relative mooring motion as a stochastic process.

A3. A stochastic model for mooring motion

The relative position of moorings is expressed as the sum of a fixed distance and a perturbation distance. Fixed distances were determined from estimated anchor positions, while perturbation distances, $\epsilon = (\epsilon_x, \epsilon_y)$, were modeled as Gaussian white noise processes with zero mean and non-zero variance, $\sigma^2 = (\sigma_x^2, \sigma_y^2)$, estimated from the time-integration of differential currents. The distance between center and neighboring moorings are assumed uncorrelated and motion is constant with depth at a given time.

An upper bound on the amplitude of mooring motion is estimated as follows. We decompose the observed velocities into mean and perturbation velocities:

$$u_i = \bar{u}_i + u'_i \tag{A1}$$

$$u_c = \bar{u}_c + u'_c, \quad (\text{A2})$$

where subscripts i and c denote inner and center moorings, respectively. Then

$$u_i - u_c = (\bar{u}_i - \bar{u}_c) + (u'_i - u'_c). \quad (\text{A3})$$

The first term is a secular signal while the second is a high-frequency signal representing the “flutter” of moorings. One can eliminate the former while retaining the latter by applying a high pass-filter to Eq. A3. We used a cut-off frequency of 48 hours, retaining all motions with periods smaller than this value. We next integrated this signal in time to obtain an upper bound on the zonal distance between the center and neighboring inner moorings:

$$\delta x = \int (u'_i - u'_c) dt. \quad (\text{A4})$$

It is an upper bound since the mooring cable will constrain the magnitude of relative mooring motion to be smaller than that inferred from Eq. A4. Finally, the variance, $\sigma_{ci_x}^2$, is obtained by averaging δx within a running window and estimating the deviation of δx from this mean:

$$\sigma_{ci_x}^2 = \frac{1}{T} \int_{t-T/2}^{t+T/2} (\delta x - \overline{\delta x})^2 d\tau, \quad (\text{A5})$$

where $T = 1$ month. A similar calculation for v_i and v_c yields δy and $\sigma_{ci_y}^2$. If repeated for all inner moorings, we have variances, $\sigma_{ci_x}^2$ and $\sigma_{ci_y}^2$, as a function of time and from which we defined the average magnitude of mooring motion, $\sigma = (\sigma_x, \sigma_y)$, as follows:

$\sigma_x = \frac{1}{m} \sum_{i=1}^m \sigma_{ci_x}$ and $\sigma_y = \frac{1}{m} \sum_{i=1}^m \sigma_{ci_y}$, where $m = 4$. Although variances in zonal and meridional directions are approximately equal, we retained their uniqueness in our

simulations. Supporting Material, Figure S4c illustrates the magnitude of relative mooring motion as a function of time.

A4. Quantifying uncertainty

The stochastic model presented above was used to estimate uncertainty in statistics. To accomplish this, we employed a Monte Carlo approach. We first expressed perturbation distances in terms of time series, $\epsilon(t)$. Time series of $\epsilon \equiv (\epsilon_x, \epsilon_y)$ were created by randomly sampling Gaussian distributions with zero means and standard deviations, σ , given above. We smoothed resulting time series using a 24-hr filter to prevent abrupt mooring motion. Next, the vector position of two moorings (cf. Eq. 2) was defined as $d\mathbf{l} = d\mathbf{x} + \epsilon(t)$, where $|d\mathbf{x}|$ denotes the fixed distance between the moorings.

Having obtained stochastic realizations of mooring position, ζ was calculated. For each time series, the standard deviation and skewness were estimated. Repeating this process 1000 times yielded distributions for each statistic as functions of depth and season. Shaded regions in Figure 2 illustrate the breadth of these distributions at the 90% level.

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References

Allen, J. T., A. Naveira-Garabato, A. Forryan and coauthors (2013), Cruise Report: RSS *Discovery* Cruise 381, 14 Sep. - 03 Oct. 2012: Ocean Surface Mixing, Ocean Submesoscale Interaction Study (OSMOSIS), National Oceanography Centre, Southampton, 1–196.

Batchelor, G. K. (2000): *An Introduction to Fluid Dynamics*. reprint, Cambridge University Press, Cambridge, 615 pp.

Belcher, S. E., A. L. M. Grant, K. E. Hanley, B. Fox-Kemper, L. Van Roekel, P. Sullivan, W. G. Large, A. Brown, A. Hines, D. Calvert, Daley, A. Rutgersson, H. Pettersson, J.-R. Bidlot, and P. A. E. M. Janssen, and J. A. Polton (2012), A global perspective on Langmuir turbulence in the ocean surface boundary layer, *Geophys. Res. Lett.*, *39*, L18605, doi:10.1029/2012GL052932.

Boccaletti, G., R. Ferrari, and B. Fox-Kemper (2007), Mixed layer instabilities and restratification, *J. Phys. Oceanogr.*, *37*, 2228–2250, doi:10.1175/JPO3101.1.

Brannigan, L., D. P. Marshall, A. C. Naveira Garabato, and A. J. G. Nurser (2015), The seasonal cycle of submesoscale flows, *Ocean Model.*, doi:10.1016/j.ocemod.2015.05.002.

Brüggemann, N., and C. Eden (2015), Routes to dissipation under different dynamical conditions, *J. Phys. Oceanogr.*, *45*, 2149–2168, doi:10.1175/JPO-D-14-0205.1.

Bryden, H. L., and N. Fofonoff (1977), Horizontal divergence and vorticity estimates from velocity and temperature measurements in the MODE Region, *J. Phys. Oceanogr.*, *7*, 329–337.

Callies, J., R. Ferrari, J. Klymak, and J. Gula (2015), Seasonality in submesoscale turbulence, *Nature Comm.*, *6*, 1–8, doi:10.1038/ncomms7862.

Capet, X., J. C. McWilliams, M. J. Molemaker, and A. F. Shchepetkin (2008), Mesoscale to submesoscale transition in the California Current System. Part III: Energy balance and flux, *J. Phys. Oceanogr.*, *38*, 2256–2269, doi:10.1175/2008JPO3810.1.

Colin de Verdière, A. (2012), The stability of short symmetric internal waves on sloping fronts: beyond the traditional approximation, *J. Phys. Oceanogr.*, *42*, 459–475, doi:10.1175/JPO-D-11-067.1.

de Boyer Montégut, C., G. Madec, A. S. Fischer, A. Lazar, and D. Iudicone (2004), Mixed layer depth over the global ocean: an examination of profile data and a profile-based climatology, *J. Geophys. Res.*, *109*, C12003, doi:10.1029/2004JC002378.

D’Asaro, E., C. Lee, L. Rainville, R. Harcourt, and L. Thomas (2011), Enhanced turbulence and energy dissipation at ocean fronts, *Science*, *332*, 318–322.

Eldevik, T. and K. .B. Dysthe (2002), Spiral eddies, *J. Phys. Oceanogr.*, *32*, 851–869, doi:10.1175/1520-0485(2002)032;0851:SEj;2.0.CO;2.

Ertel, H. (1942), Ein neuer hydrodynamischer Wirbelsatz, *Meteorol. Z.*, *59*, 271–281.

Fox-Kemper, B., G. Danabasoglu, R. Ferrari, and R. W. Hallberg (2008), Parameterizing submesoscale physics in global climate models, *CLIVAR Exchanges*, *13*, 3–5.

Fox-Kemper, B., R. Ferrari, and R. W. Hallberg (2008), Parameterization of mixed layer eddies. Part I: Theory and diagnosis, *J. Phys. Oceanogr.*, *38*, 1145–1165, doi:10.1175/2007JPO3792.1.

Grooms, I. (2015), Submesoscale baroclinic instability in the balance equations, *J. Fluid Mech.*, *762*, 256–272, doi:10.1017/jfm.2014.657.

Haine, T. W. N., and J. Marshall (1998), Gravitational, symmetric and baroclinic instability of the ocean mixed layer, *J. Phys. Oceanogr.*, *28*, 634–658.

Hamlington, P. E., L. P. Van Roekel, B. Fox-Kemper, K. Julien, and G. P. Chini (2014), Langmuir-submesoscale interactions: descriptive analysis of multiscale frontal spindown simulations, *J. Phys. Oceanogr.*, *44*, 2249–2272.

Hoskins, B. J. (1974), The role of potential vorticity in symmetric stability and instability, *Q.J.R. Meteorol. Soc.*, *100*, 480–482, doi:10.1002/qj.49710042520.

Hoskins, B. J. (1982), The mathematical theory of frontogenesis, *Ann. Rev. Fluid Mech.*, *14*, 131–151, doi:10.1146/annurev.fl.14.010182.001023.

Large, W. G., J. C. McWilliams, and S. C. Doney (1994), Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization, *Geophys. Res. Lett.*, *32*, 363–403, doi:10.1029/94RG01872.

Lazar, A., A. Stegner, and E. Heifetz (2013), Inertial instability of intense stratified anticyclones. Part 1. Generalized stability criterion, *J. Fluid Mech.*, *732*, 457–484, doi:10.1017/jfm.2013.412.

Mensa, J. A., Z. Garraffo, A. Griffa, T. M. Özgökmen, A. Haza, and M. Veneziani (2013), Seasonality of the submesoscale dynamics in the Gulf Stream region, *Ocean Dynamics*,

63, 923–941, doi:10.1007/s10236-013-0633-1.

Munk, W., L. Armi, K. Fischer and F. Zachariassen (2000), Spirals on the sea, *Proceedings of the Royal Society: mathematical, physical and engineering sciences*, 456, 1217–1280.

Ostrovskii, A. G. (1995), Signatures of stirring and mixing in the Japan Sea surface temperature patterns in autumn 1993 and spring 1994. *Geophys. Res. Lett.*, 22, 2357–2360, doi:10.1029/95GL02494.

Polton, J. A., D. M. Lewis, and S. E. Belcher (2005), The role of wave-induced Coriolis-Stokes forcing on the wind-driven mixed layer, *J. Phys. Oceanogr.*, 35, 444–457, doi:10.1175/JPO2701.1.

Polton, J. A., and S. E. Belcher (2007), Langmuir turbulence and deeply penetrating jets in an unstratified mixed layer, *J. Geophys. Res.*, 112, C09020, doi:10.1029/2007JC004205.

Qiu, B., S. Chen., P. Klein, H. Sasaki, and Y. Sasai (2014), Seasonal mesoscale and submesoscale eddy variability along the North Pacific Subtropical Countercurrent, *J. Phys. Oceanogr.*, 12, 3070–3098, doi:10.1175/JPO-D-14-0071.1.

Roulet, G., and P. Klein (2010), Cyclone-anticyclone asymmetry in geophysical turbulence, *Phys. Rev. Lett.*, 104, 218501 (1–4), doi:10.1103/PhysRevLett.104.218501.

Rudnick, D. L. (2001), On the skewness of vorticity in the upper ocean, *Geophys. Res. Lett.*, 28(10), 2045–2048, doi:10.1029/2000GL012265.

Sasaki, H., P. Klein, B. Qiu, and Y. Sasai (2014), Impact of oceanic-scale interactions on the seasonal modulation of ocean dynamics by the atmosphere, *Nature Comm.*, 5, 1–8, doi:10.1038/ncomms6636.

Shcherbina, A. Y., E. A. D'Asaro, C. M. Lee, J. M. Klymak, M. J. Molemaker, and J. C. McWilliams (2013), Statistics of vertical vorticity, divergence, and strain in a developed submesoscale turbulence field, *Geophys. Res. Lett.*, *40*, 4706–4711, doi:10.1002/grl.50919.

Stone, P. H. (1966), On non-geostrophic baroclinic instability, *J. Atmos., Sci.*, *23*, 390–400.

Taylor, J., and R. Ferrari (2009), On the equilibration of a symmetrically unstable front via a secondary shear instability, *J. Fluid Mech.*, *622*, 103–113, doi:10.1017/S0022112008005272.

Taylor, J., and R. Ferrari (2010), Buoyancy and wind-driven convection at mixed layer density fronts, *J. Phys. Oceanogr.*, *40*, 1222–1242, doi:10.1175/2010JPO4365.1.

Thomas, L. N. (2005), Destruction of potential vorticity by winds, *J. Phys. Oceanogr.*, *35*, 2457–2466, doi:10.1175/JPO2830.1.

Thomas, L. N., and C. M. Lee (2005), Intensification of ocean fronts by down-front winds, *J. Phys. Oceanogr.*, *35*, 1086–1102, doi:10.1175/JPO2737.1.

Thomas, L. N., and J. Taylor (2014), Damping of inertial motions by parametric subharmonic instability in baroclinic currents, *J. Fluid Mech.*, *743*, 280–294, doi:10.1017/jfm.2014.29.

Thomas, L. N., A. Tandon and A. Mahadevan (2008), Submesoscale processes and dynamics, *Ocean Modeling in an Eddy Regime*, Geophysical Monograph Series 177, doi:10.1029/177GM04.

Thomas, L., J. Taylor, R. Ferrari and T. M. Joyce (2013), Symmetric instability in the Gulf Stream, *Deep-Sea Res. II*, *91*, 96–110, doi:10.1016/j.dsr2.2013.02.025.

Thompson, A. F., A. Lazar, C. Buckingham, A. C. Naveira Garabato, G. M. Damarell, and K. J. Heywood (2016), Open-ocean submesoscale motions: a full seasonal cycle of mixed layer instabilities from gliders, *J. Phys. Oceanogr.*, in press.

Whitt, D. B., and L. N. Thomas (2005), Resonant generation and energetics of wind-forced near-inertial motions in a geostrophic flow, *J. Phys. Oceanogr.*, *45*, 181–208, doi:10.1175/JPO-D-14-0168.1.

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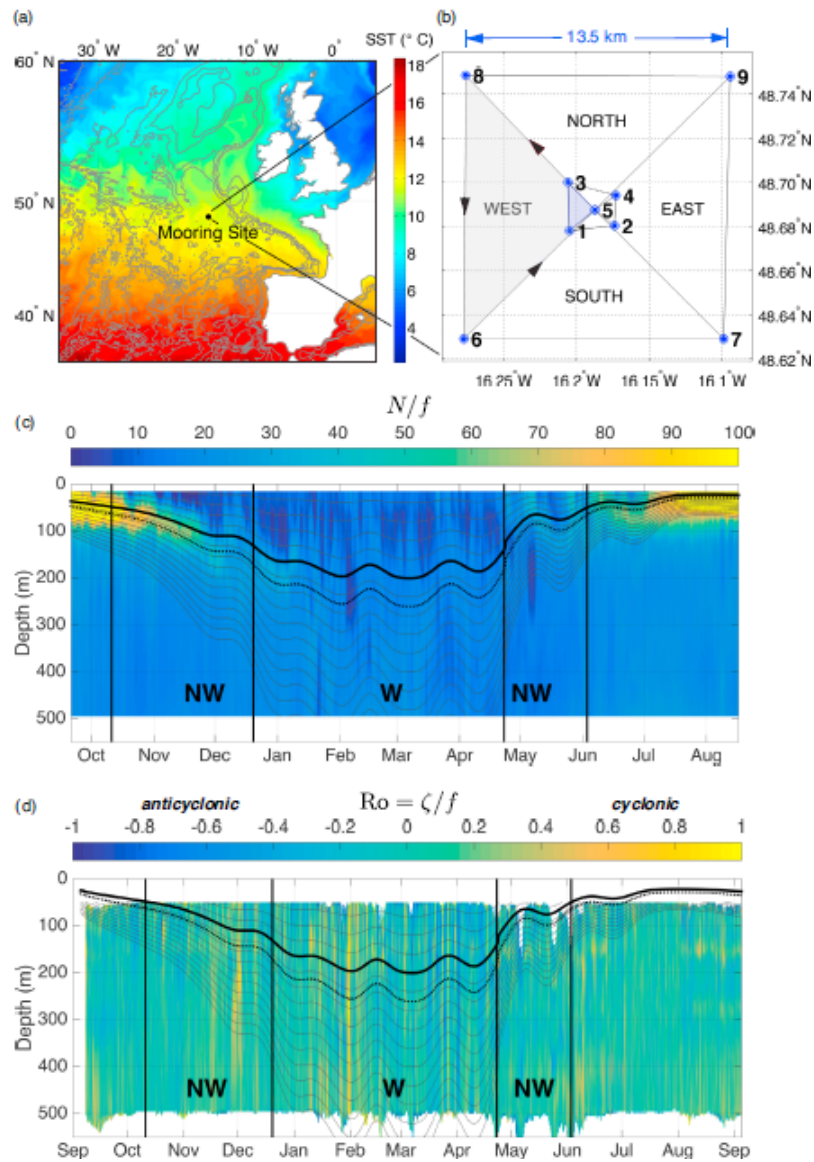


Figure 1. (a) The location of the OSMOSIS mooring site overlaid on sea surface temperature (SST) from a $1/12^\circ$ NEMO ocean model. Bathymetric contours are shown every 1000 m between 0 and 6000 m, and moorings were located in waters approximately 4800 m in depth. (b) Locations of inner (1,2,3,4), outer (6,7,8,9) and center (5) moorings. The inner mooring domain was divided into four triangle-shaped regions and vorticity was estimated within these regions. Black arrows denote the sense of circulation associated with positive vorticity. (c) Vertical stratification, N/f , as observed by gliders (September 2012 – August 2013). (d) Relative vorticity, $Ro = \zeta/f$, estimated within the inner mooring domain (September 2012 – September 2013). In both (c) and (d), winter (W) and non-winter (NW) seasons are labeled. For reference, we also display ML depth, h (thick black), $1.3h$ (dotted black) and lines of constant normalized depth, $|z|/h$, (thin gray). The latter range between 0 and 3.0 with spacing of 0.2.

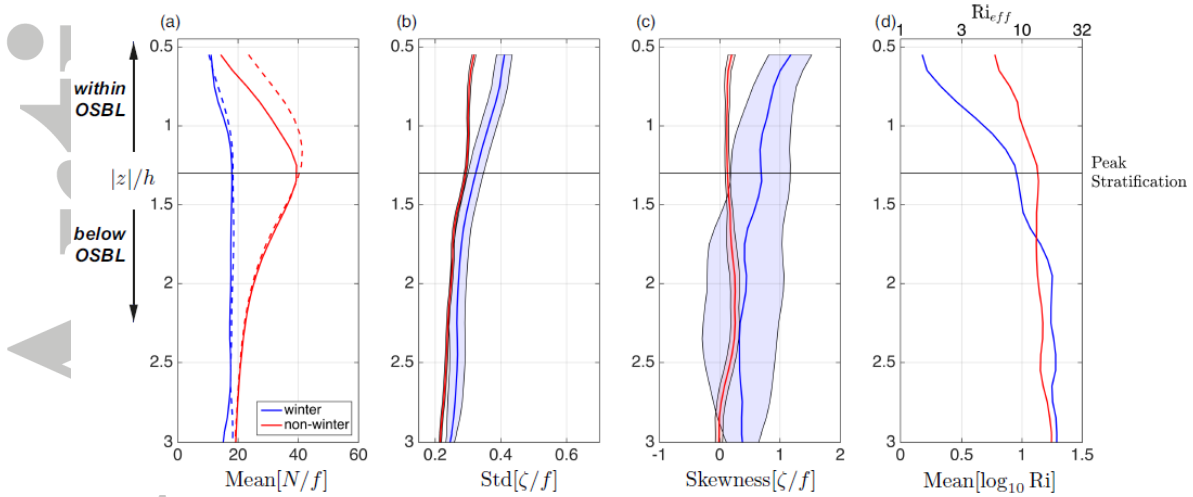


Figure 2. Submesoscale statistics as a function of normalized depth and season: winter (blue) and non-winter (red). Left to right: (a) mean vertical stratification estimated from moorings (solid) and ocean gliders (dashed), (b) standard deviation of relative vorticity, ζ/f , (c) skewness of ζ/f and (d) mean of the logged-gradient Richardson number, Ri . The gradient Richardson number was estimated as $Ri = N^2/|\partial_z \mathbf{u}|^2$. Observation depths have been normalized by a smoothed ML depth, h .

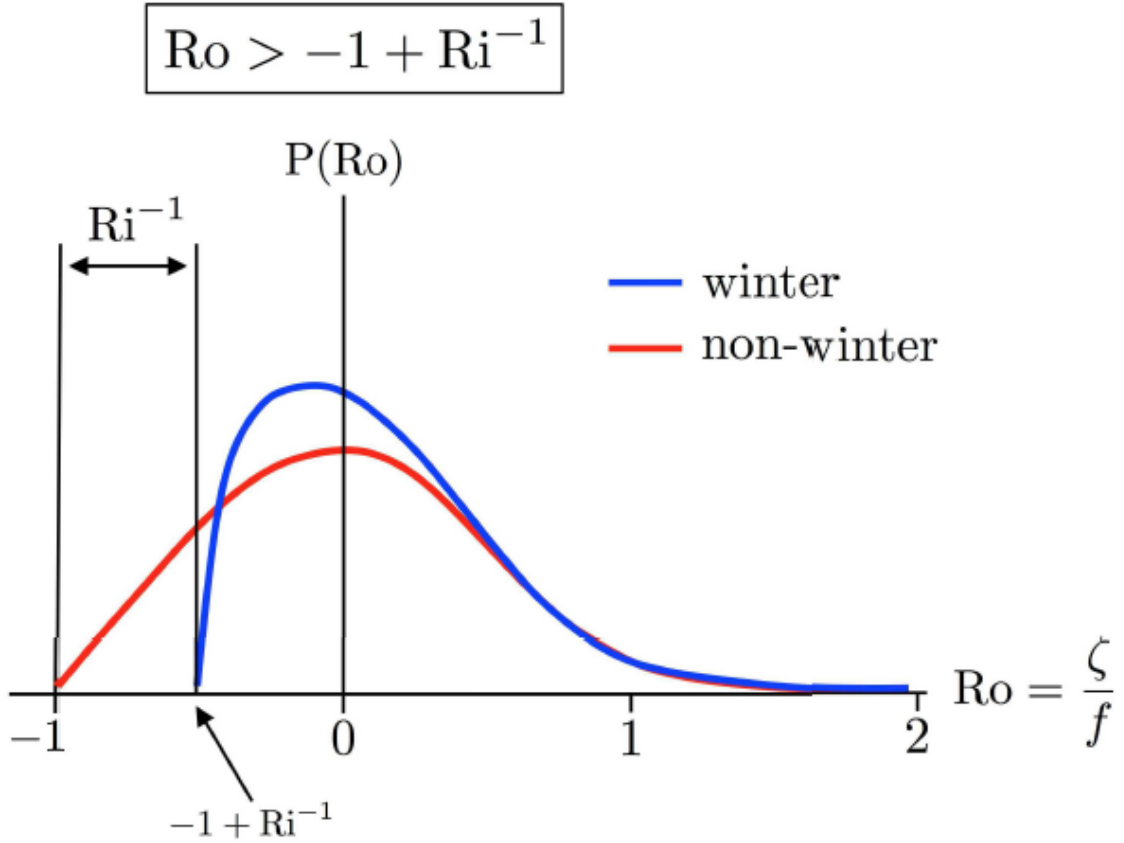


Figure 3. Illustration of how the inverse gradient Richardson number, $Ri^{-1} = |\nabla b|^2 / (f^2 N^2)$, modifies the distribution of relative vorticity, $P(Ro)$. In winter, Ri^{-1} is large and can result in skewness for low-to-moderate Ro . At other times of the year, Ri^{-1} is small and has minimal effect on vorticity. The graphic displays hypothetical PDFs of Ro and is strictly valid for geostrophic flow and finite stratification.