The Scale of Submesoscale Baroclinic Instability Globally

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(Manuscript received 24 February 2020, in final form 7 July 2020)

ABSTRACT

The spatial scale of submesoscales is an important parameter for studies of submesoscale dynamics and multiscale interactions. The horizontal spatial scales of baroclinic, geostrophic-branch mixed layer instabilities (MLI) are investigated globally (without the equatorial or Arctic oceans) based on observations and simulations in the surface and bottom mixed layers away from significant topography. Three high-vertical-resolution boundary layer schemes driven with profiles from a MITgcm global submesoscale-permitting model improve robustness. The fastest-growing MLI wavelength decreases toward the poles. The zonal median surface MLI wavelength is 51–2.9 km when estimated from the observations and from 32, 25, and 27 km to 2.5, 1.2, and 1.1 km under the K -profile parameterization (KPP), Mellor–Yamada (MY), and κ - ϵ schemes, respectively. The surface MLI wavelength has a strong seasonality with a median value 1.6 times smaller in summer (10 km) than winter (16 km) globally from the observations. The median bottom MLI wavelengths estimated from simulations are 2.1, 1.4, and 0.41 km globally under the KPP, MY, and κ - ϵ schemes, respectively, with little seasonality. The estimated required ocean model grid spacings to resolve wintertime surface mixed layer eddies are 1.9 km (50% of regions resolved) and 0.92 km (90%) globally. To resolve summertime eddies or MLI seasonality requires grids finer than 1.3 km (50%) and 0.55 km (90%). To resolve bottom mixed layer eddies, grids finer than 257, 178, and 51 m (50%) and 107, 87, and 17 m (90%) are estimated under the KPP, MY, and κ - ϵ schemes.

1. Introduction

Ocean mixed layers (ML) are the vitally important channel for energy and material exchanges between the ocean interior and the atmosphere or ocean floor. For example, energy from surface winds is transferred downward to the ocean interior via the surface

mixed layer (SML), and much of it dissipates against the seafloor in the bottom mixed layer (BML). Alternatively, nutrients are extracted from sediments and suspended in the BML, then transported upward through the ocean interior and finally sustain life in the SML.

The recent recognition of submesoscale processes with spatial scales of $O(1\text{--}10)$ km has revolutionized modern oceanography (Thomas et al. 2008; McWilliams 2016). Submesoscale processes are ubiquitous in the ML over the World Ocean, as shown in observations (Pollard and Regier 1992; Rudnick and Ferrari 1999;

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Rudnick 2001; Munk et al. 2000; Callies et al. 2015; Buckingham et al. 2016; Ruan et al. 2017) and simulations (Capet et al. 2008; Fox-Kemper et al. 2011; Gula et al. 2015; Molemaker et al. 2015; Sasaki et al. 2017; Su et al. 2018; Zhang et al. 2020). Submesoscale processes in the SML have been intensively studied, and are demonstrated to make significant contributions to vertical heat and material transports, and impact ocean dynamics (Lévy et al. 2010; Fox-Kemper et al. 2011; Yang et al. 2017; Su et al. 2018; Buckingham et al. 2019) and ecosystem deeply (Taylor and Ferrari 2011; Lévy et al. 2012, 2018; Uchida et al. 2019). The typical horizontal buoyancy gradient in the SML and the potential energy barrier of the SML base favor growth of one kind of submesoscale baroclinic instability modes, called mixed layer instability (MLI; Boccaletti et al. 2007). Other instabilities are also specific to the SML, such as symmetric and Langmuir instabilities, but here the MLI will refer only to the baroclinic instability variant. Submesoscale processes generated via the surface mixed layer instability (SMLI) often dominate submesoscale variability in the SML, extracting energy from mean potential energy in SML fronts and leading to SML restratification (Boccaletti et al. 2007; Fox-Kemper et al. 2008; Hosegood et al. 2008; Fox-Kemper et al. 2011). Recent works exploring the bottom mixed layer instability (BMLI; Wenegrat et al. 2018; Callies 2018; Ruan and Callies 2020) reflect an interestingly different role affecting water masses, mean overturning, and interbasin exchanges. According to baroclinic instability theory (Boccaletti et al. 2007), the scale of the MLI mode is restricted by ML thickness, because submesoscale variability in higher stratification of the ocean interior is evanescent under larger resulting Richardson number. Submesoscale activity is greatly dependent on available potential energy—i.e., thermal wind shear of fronts—within the SML; similarly, BMLI depends on shear in the BML. Thus, the spatial scale of the SMLI mode varies seasonally, since the SML is thick and filled with fronts in winter but thin in summer. Hence, SMLI should have a relatively large scale in winter but small scale in summer. Activity of submesoscale processes is demonstrated to have a seasonal variability and is more active in winter, according to current model simulations and observations with resolutions around $O(1)$ km (Mensa et al. 2013; Callies et al. 2015; Rocha et al. 2016; Sasaki et al. 2017; Su et al. 2018; Dong et al. 2020). However, the small scale of SMLI in summer is potentially unresolved by many current models and observations (which are often on 1–2-km scale), and thus the summertime disappearance of SMLI may potentially stem from the seasonally changing submesoscale scale versus the analyzed scale.

In contrast, the role of submesoscale processes in the BML is still unclear, though limited studies have been focused on generation of submesoscale processes due to interaction between current and topography (Gula et al. 2015, 2016; Ruan and Callies, 2020) or exploiting dynamical similarity to SMLI (Wenegrat et al. 2018; Callies 2018). Inertial and symmetric instabilities triggered by topography in the coastal BML are capable of generating submesoscale processes, which is believed to enhance cross-shelf exchanges (Gula et al. 2015) and forward energy cascade for local dissipation (Gula et al. 2016). As a predominant mechanism for submesoscale processes in the SML, BMLI has been investigated by Wenegrat et al. (2018) theoretically. BMLI is found to occur in the BML causing vertical buoyancy flux, in analogy to SMLI. Hence, BMLI is expected to be a significant route for submesoscale generation, especially in the deep ocean.

As a link between meso- and microscale, the spatial scale is an important parameter for MLI studies. The spatial scale provides a guidance for resolution requirements of simulations and observations to resolve MLI, further impacting research on submesoscales and multiscale dynamics. The MLI scale may vary spatially and temporally in both the SML and BML, since ML thickness varies in time and space. For further study of submesoscale processes, the MLI scale needs to be clarified globally, not only in the SML, but also BML. In this work, the spatial scales of MLI in the SML using observations and models and in the BML using models are investigated globally. The models used here are of two types—a global 3D submesoscale-permitting simulation and a variety of 1D boundary layer models. The latter provides high vertical resolution and a multiphysics ensemble for robustness. The paper is organized as follows: section 2 introduces the models and methods that used in the work, section 3 describes the global and seasonal patterns of the SMLI scale, section 4 investigates the global pattern of the BMLI scale, and the last section gives the discussion and conclusions.

2. Observations, models, and methods

a. Observations

Hydrographic profiles including temperature, salinity and pressure from the Argo program in Februaries and Augusts from 2010 to 2019 over the globe are used for the analysis of the SMLI scale. Each Argo float is equipped with sensors of conductivity, temperature, and pressure, which drifts freely at a parking depth (usually 1000–2000 m). It ascends to sea surface every 10 days and transmits the measured parameters during its

ascending to the satellites (<http://www.argo.ucsd.edu>). The data are provided by the NOAA National Centers for Environmental Information (https://www.nodc.noaa.gov/argo/basins_data.htm). To ensure calculation accuracy of ML thickness and stratification, the profiles are filtered with vertical grid spacings less than 2 m and the shallowest level above 5-m depth. SML thickness and stratification based on each profile are calculated and finally averaged into a global 4° grid field that is consistent with the following simulation.

b. Model description

Simulations of the SML and BML are accomplished by using the General Ocean Turbulence Model (GOTM; Umlauf et al. 2005). GOTM is a one-dimensional water column model that is focused on ocean turbulence related to vertical mixing. GOTM provides a number of turbulence closure schemes to parameterize turbulence fluxes, including the κ - ϵ two-equation model, the Mellor–Yamada second-order model (MY; Mellor and Yamada 1982), and the K -profile parameterization (KPP) scheme (Large et al. 1994), which are used as the multiphysics ensemble here. The source code is freely available online (<https://gotm.net>). It is a major advantage to be able to see a variety of different schemes in this work, because similarities and differences among them quantify both the understanding of the problem and the remaining uncertainty, as demonstrated in a recent comprehensive comparison (Li et al. 2019).

Initial and boundary conditions are provided by the Massachusetts Institute of Technology general circulation model (MITgcm) driven by the European Centre for Medium-Range Weather Forecasts (ECMWF) surface conditions. The MITgcm model simulation has a horizontal grid spacing of 1/48° with 90 vertical layers over the globe. External forcing includes 6-hourly atmospheric forcing from the 0.14° ECMWF and the full luni-solar tidal potential [although Arbic et al. (2018) note that the method of including tides is not standard]. The vertical mixing is the KPP scheme in boundary layers. The model is refined originally from a 1/6° global ocean state estimate generated by the Estimating the Circulation and Climate of the Ocean (ECCO) Phase II project, successively downscaling to 1/12° and then 1/24° simulations. The model is simulated from September 2011 to November 2012, and hourly model output is available at the ECCO data portal (<https://data.nas.nasa.gov/ecco/>). Air–sea fluxes including momentum and heat provided by the ECMWF have a spatial grid spacing of 0.25°, with a temporal interval of 6 h.

The MITgcm data are subsampled directly into fields with a spatial grid spacing of 4° globally and the air–sea fluxes are interpolated into the same resolution.

Then GOTM is integrated at each grid point where water depth exceeds 500 m to obtain full-depth tracer profiles including temperature and salinity, as well as turbulence statistics such as dissipation rate ϵ . The vertical grid spacing for GOTM is less than 3 m, and grids near the surface and bottom are squeezed to centimeter scale. Each GOTM simulation is conducted for a duration of one month in two different months, February and September, and only the last 15 days for each month are used for the analysis, considering strong seasonal variability of the SML.

Profiles from the MITgcm simulation have vertical grid spacings up to 1 m in the SML, and around 200 m near the bottom in the deep. To initialize the GOTM simulations, velocity and tracer profiles are extrapolated from the deepest wet layer to the seafloor, so as to provide the needed profiles. The velocity and tracers are processed differently. The tracers are extrapolated to the seafloor by assuming their gradients are constants and the same values as the deepest MITgcm layer. For the velocity (including tidal currents), a bottom Ekman layer depth is estimated based on an empirical expression (Caldwell et al. 1972; Armi and Millard 1976; Weatherly and Martin 1978),

$$h_{\text{Ekman}} = \frac{0.4u_*}{f}, \quad (1)$$

where $u_* = 0.03U_{\text{int}}$ is the friction velocity estimated from the interior velocity magnitude, U_{int} above the bottom Ekman layer, and f is the Coriolis parameter. The velocity is then set to decrease as depth increases by a natural exponential expression with zero velocity at the seafloor,

$$\mathbf{u} = \mathbf{u}_{\text{int}} \left\{ 1 - \exp \left[- \frac{(1+i)}{h_{\text{Ekman}}} z \right] \right\}. \quad (2)$$

Here, $\mathbf{u} = u + iv$ and $\mathbf{u}_{\text{int}} = u_{\text{int}} + iv_{\text{int}}$ are the complex forms of the Ekman layer velocity and interior velocity, and z is the distance above the seafloor. During the simulations, the temperature and salinity profiles are provided as the initial conditions, but velocity profiles are nudged to the MITgcm profiles with the bottom Ekman layer in a time scale of 1 h.

Sensitivity tests have been carried out, which indicate that a smaller time step (30 s, compared with 600 s that used here) and a finer vertical grid spacing (doubling grid points over the standard used) exert negligible impacts on the results. To evaluate the sensitivity of the Ekman layer processing of the velocity profiles, simulations without any processing on the velocity profiles are conducted, and it is found that the derived BMLI scales have similar patterns and are 5%–9% smaller.

However, a nonnegligible difference in ML thickness results from using different turbulence closures. To illustrate this uncertainty, simulations with three different turbulence closures are conducted and presented, including the KPP, MY, and κ - ϵ schemes. Note that using 1D mixing models uncoupled to the 3D flow limits effects of important feedbacks, but the initialization from a 3D model reduces this concern. Furthermore the use of the 1D models allows for a significantly finer vertical grid spacing, which is required to have precise boundary layer estimation, than is possible in the MITgcm simulations.

c. Determination of mixed layer instability scale

According to quasigeostrophic (QG) instability theory, one specific horizontal scale of MLI has the fastest growth rate, and the scale is tightly related to the local internal Rossby deformation radius,

$$R = \frac{NH}{f}. \quad (3)$$

Here, N , H , and f are the buoyancy frequency, vertical scale of the perturbation, and Coriolis parameter, respectively. For mesoscale instabilities, the QG theory predicts the wavelength of the most unstable perturbation to be somewhat larger than the local deformation radius as (Eady 1949)

$$L \approx 4R. \quad (4)$$

However, the baroclinic instability wavelength of MLI differs from the QG scale due to ageostrophic effects which shift the instability to a longer scale at small Richardson number (Stone 1966; Nakamura 1988; Boccaletti et al. 2007),

$$L_{\text{ML}} \approx 4\sqrt{1 + \text{Ri}^{-1}}R. \quad (5)$$

Here, $\text{Ri} = N_{\text{ML}}^2 H_{\text{ML}}^2 / \Delta U^2$ is the bulk Richardson number, measuring a relative magnitude between stratification N_{ML} and vertical shear of horizontal velocity in the ML (ΔU is the change in horizontal velocity across the ML depth H_{ML}). As a result, the scale of the fastest growing mode in the ML falls in a range of (Eldevik and Dysthe 2002; Özgökmen et al. 2011)

$$4R \leq L_{\text{ML}} \leq 6R. \quad (6)$$

In this work, the upper bound is chosen for the estimation (corresponding to $\text{Ri} \approx 0.8$), namely,

$$L_{\text{ML}} = 6R = 6 \frac{N_{\text{ML}} H_{\text{ML}}}{f}, \quad (7)$$

where $N_{\text{ML}} = \int_0^{H_{\text{ML}}} N(z) dz / H_{\text{ML}}$, which is the depth-averaged buoyancy frequency in the ML consistent with the WKB instability theory (Chelton et al. 1998), H_{ML} is the ML thickness. Given that an estimation of Ri cannot be conducted based on the observations, a constant Ri is applied here for consistency of the scale calculation. The ML in the real ocean tends to stay near a neutral state with $\text{Ri} = 1$ and zero potential vorticity (PV) due to the restratification processes of symmetric instability (Bachman et al. 2017) and geostrophic adjustment (Tandon and Garrett 1994; 1995), although MLI, solar, and wind-driven restratification by upfront winds can lead to much larger Ri than 1 (Fox-Kemper et al. 2008; Thomas et al. 2008; Haney et al. 2012). However, the mixing schemes implemented in the 1D models do not represent most of these restratification processes. As a result, there is an important systematic bias in using 1D models to estimate the stratification in the ML required to estimate the scale of the MLI (the analysis in the following will show that the 1D models tend to underestimate the stratification in the SML), potentially leading to Ri much less than unity and smaller than reality. Following the works of Eldevik and Dysthe (2002) and Özgökmen et al. (2011), a constant $\text{Ri} \approx 0.8$ is chosen in this work.

Based on H_{ML} and N_{ML} in the ML determined from the ARGO profiles and GOTM simulations using the method above, the spatial scales of MLI in both the SML and BML over the globe can be obtained. In this work, equatorial regions (equatorward of 10°) are excluded because the small f values preclude geostrophy.

d. Determination of mixed layer thickness

Numerous works have been undertaken for developing calculation methods of ML thickness (e.g., de Boyer Montégut et al. 2004; Holte and Talley 2009; Chu and Fan 2011; Lozovatsky and Shapovalov 2012; Huang et al. 2018; Banyte et al. 2018; Huang et al. 2019). Considering different features between the SML and BML, different methods are applied to determine ML thickness here, which are classified into two major types, threshold (Thrd) and relative variance (Rvar) methods (details in the appendix).

Each method has advantages and disadvantages. But here the best method for ML thickness calculation is determined by a comparison of the MLI scales to the results from a QG instability analysis. Temporally averaged density and velocity profiles in February at a site in the Pacific basin are chosen for the analysis (Fig. 1). An obvious SML with thickness around 200 m and a BML with thickness less than 100 m can be found from the potential density profile (Fig. 1a). Given the density

FIG. 1. Temporally averaged (a) potential density (reference seawater pressure: 0 dbar) and (b) velocity profiles in February in the Pacific basin (40.16°N, 178.76°E). The inserted panel in (a) shows the density profile near the bottom. The profiles are obtained from the GOTM simulations with the κ - ϵ scheme.

and velocity vertical shear profiles, the QG instability analysis predicts growth rates of the MLI versus wavenumbers (Fig. 2). Two peaks are observed from the growth rate distribution with growth time less than one day (Fig. 2a). The first peak is around 8.8 km with a vertical structure trapped in the SML, indicating that it is SMLI (Fig. 2b). The wavenumber of the other peak is much larger with a wavelength of 56 m, and its vertical structure is trapped in the BML (Fig. 2c), so it is BMLI.

Parameters H_{ML} and L_{ML} calculated from different ML thickness determination methods are shown on top of the SMLI and BMLI growth rates and profiles. In the SML, the ML thickness H_{SML} from different methods varies from about 153 to 207 m, with maximum difference exceeding 50 m (red and yellow dots in Fig. 2b). Correspondingly, the SMLI wavelength, L_{SML} is in a magnitude of $O(10)$ km, and the maximum difference

exceeding 10 km (red and yellow lines in Fig. 2a). Generally, the dissipation Thrd and Rvar methods are both close to the peak from the QG analysis. The thickness of the BML, H_{BML} shows a big difference ranging from 85 to 920 m (blue and green dots in Fig. 2c). Hence, the BMLI wavelength, L_{BML} is up to 10 km for the temperature Thrd method with a potential temperature difference of 0.005°C, but as low as 51 m for the dissipation Thrd method (blue and green lines in Fig. 2a). Evaluating the SMLI and BMLI wavelengths versus the QG instability analysis, the Rvar method is the best for the MLI scale determination. As the dissipation rate is not available from observations globally or simulations using the KPP scheme, the dissipation Thrd method is not used here, although it is similarly accurate to the Rvar method. The scatterplot in Fig. 3 gives a statistical relation indicating that L_{ML} from the Rvar method is about 1.2 times larger than the scale from the QG

FIG. 2. (a) Growth rate vs wavenumber (black curve) and vertical structures associated with the most unstable modes in the (b) SML (peak at 1.14×10^{-4} cycles per meter) and (c) BML (peak at 1.77×10^{-2} cycles per meter). The lines and dots with different colors denotes L_{ML} and H_{ML} that derive from different methods.

analysis. The scales are derived based on the GOTM simulation results with the κ - ϵ scheme. As MLI occurs at relatively low Ri , the QG instability analysis used here underpredicts the fastest growing wavelength, as mentioned in the preceding section. As a result, the slight tendency of the Rvar method to overpredict the scale when compared to the QG scale is acceptable. Reproducing this analysis with the other turbulence schemes yields similar results.

In conclusion, the SMLI and BMLI scales can be accurately estimated from Eq. (7) together with the Rvar method to determine ML thickness. An added advantage of the Rvar method is that it can be used with the Argo observations for the SML as well as the

GOTM model results, so that model biases can be assessed.

3. The scale of surface mixed layer instability

The SML thickness, H_{SML} in winter (February in the Northern Hemisphere and August in the Southern Hemisphere, the same below) and summer (August in the Northern Hemisphere and February in the Southern Hemisphere, the same below) determined using the Rvar method shows a remarkable seasonality (Fig. 4). H_{SML} derived from the ARGO profiles in winter is generally deep, especially at mid- and high latitudes with thickness generally exceeding 100 m (Fig. 4a). In contrast, H_{SML} in

FIG. 3. Scatterplot of L_{ML} vs the scales derived from the QG analysis L_{QG} . The red and blue dots denote scales in the SML and BML, respectively. The green line denotes the linear regression curve, $L_{ML} = 1.2 L_{QG}$. The scales are calculated based on the GOTM simulation results with the κ - ϵ scheme.

summer is relatively shallow (less than 50 m), aside from a deep ML belt of 50–100 m along the Antarctic Circumpolar Current (ACC) (Fig. 4b). These seasonal and spatial features are well-known and have been widely discussed based on different observations and calculation methods (e.g., Monterey and Levitus 1997; Kara et al. 2003; de Boyer Montégut et al. 2004). The seasonal variability and spatial pattern are captured by the GOTM simulations with different turbulence closures (Figs. 4c–h). The buoyancy frequency in the SML, N_{SML} is generally weak but varies over two orders of magnitude (2×10^{-4} – $4 \times 10^{-2} \text{ s}^{-1}$) (Fig. 5). Opposing the ML thickness, N_{SML} is weak where the ML is deep but strong where the ML is shallow, with an antiphase seasonality (Figs. 5a,b). Similarly, the GOTM simulations reproduce the buoyancy frequency features reasonably under each different turbulence closure (Figs. 5c–h). Quantitatively, there are persistent biases in the estimation of both H_{SML} and N_{SML} between the observations and each GOTM simulation, but these constituent biases will not be discussed, since biases of the SML scale which is the primary target of the investigation will be discussed below.

Based on H_{SML} and N_{SML} , the spatial scale of SML, L_{SML} is estimated according to Eq. (7). In the spatial distribution derived from the observations, L_{SML} is tightly dependent on latitude, largest scales near the equator and smaller near the poles, with the summertime

pole smallest of all (Figs. 6a,b). Parameter L_{SML} varies from a median scale of 51 km at low latitudes to 2.9 km at high latitudes from the observations. Comparing different seasons indicates a strong seasonal variability in the MLI scale. Parameter L_{SML} decreases from winter to summer, which is easily observed at mid- and high latitudes. The median L_{SML} value globally is 16 km in winter and decreases to 10 km in summer. Considering uneven area coverage of the grid cells at different latitudes, the median and other percentiles in the paper are determined based on area covered. Both H_{SML} and N_{SML} contribute to L_{SML} , as Eq. (7) shows. However, the seasonality of L_{SML} is mainly dominated by SML thickness, since L_{SML} variation patterns are roughly consistent with those of H_{SML} .

All of the simulations with different turbulent closures show similar spatial patterns and seasonal variability of H_{SML} , with decrease of L_{SML} from winter to summer mainly occurring at mid- and high latitudes (Figs. 6c–h). The simulated L_{SML} decreases with latitude as well, with median values range from 32 to 2.5 km under the KPP scheme, from 25 to 1.2 km under the MY scheme, from 27 to 1.1 km under the κ - ϵ scheme. Nevertheless, differences can still be found between the observations and simulations. The most obvious difference is that L_{SML} is smaller in the simulations than the observations close to the Antarctica, an inconsistency whose pattern stems more from biases in N_{SML} than H_{SML} .

The zonal median L_{SML} shows a more quantitative comparison, since the scale tightly depends on latitude (dots in Fig. 7). The corresponding shaded intervals show the 10th and 90th percentile bounds of the scales. The observed median L_{SML} is comparable in winter between two hemispheres (blue dots in Fig. 7a). However, an asymmetry of L_{SML} exists between two hemispheres in summer. Parameter L_{SML} in summer from the observations in the Southern Hemisphere is generally large and its variation slope is relatively flat compared with that in the Northern Hemisphere (blue dots in Fig. 7b). The mean difference of the median L_{SML} between two hemispheres is 1.4 km. The asymmetry of the zonal median L_{SML} is quantified by a skew parameter, defined as

$$\text{Skew} = \frac{\sum [L_{SML}(\text{Lat} - \mu)^3]}{\left\{ \sum [L_{SML}(\text{Lat} - \mu)^2] \right\}^{3/2}} \sqrt{\sum L_{SML}}. \quad (8)$$

Here, Lat is the latitude with a range from -65° to 65° (corresponding to 65°S – 65°N), and $\mu = \sum \text{Lat} L_{SML} / \sum L_{SML}$. The Skew parameter definition resembles the metric of normalized skewness from probability theory. A positive (negative) value of Skew indicates a larger scale in

FIG. 4. Global distributions of the SML thickness H_{SML} (m) derived from the (a),(b) observations, and simulations with the (c),(d) KPP, (e),(f) MY, and (g),(h) κ - ϵ schemes. Results in (left) winter (February in the Northern Hemisphere and August in the Southern Hemisphere) and (right) summer (August in the Northern Hemisphere and February in the Southern Hemisphere).

the Northern (Southern) Hemisphere than the Southern (Northern) Hemisphere. The calculated Skew based on observational L_{SML} is 0.056 in winter, implying a slightly larger scale in the Northern Hemisphere. Skew becomes negative with a value of -1.3 in summer, consistent with the asymmetry observed in Fig. 7b. This asymmetry is mainly caused by a deeper ML in summer in the Southern Hemisphere, especially in the ACC.

Compared with the observations, L_{SML} derived from the simulations is generally small regardless of the turbulence closures or months. The bias is large (as much as 26 km between the observations and the MY scheme) as the latitude gets close to the tropical regions and becomes relatively small (less than 2 km) at high latitudes (a result of increasing f versus latitude). A further analysis of both H_{SML} and N_{SML} indicates that the relatively smaller simulated L_{SML} is mainly a result of the

relatively weaker simulated N_{SML} . The bias may be due to differences in averaging, since L_{SML} is a 10-yr average for the observations but a 15-day average for the simulations. Another possible reason for the weaker N_{SML} may be a missing of restratification processes, such as submesoscale processes (Thomas et al. 2008; Fox-Kemper et al. 2011; Bachman et al. 2017) in the simulations. Overall, the simulation with the KPP scheme tends to most accurately match the observed SMLI scale. See Li et al. (2019) for a more detailed comparison of boundary layer schemes.

Despite the quantitative bias between the observations and simulations, all different turbulence closures highlight the asymmetry of L_{SML} as well. The calculated Skew is 0.84 in winter and -1.4 in summer for the KPP, 2.9 in winter and -0.59 in summer for the MY, and 1.1 in winter and -1.2 in summer for the κ - ϵ (recall, the

FIG. 5. Global distributions of the depth-averaged buoyancy frequency in the SML $N_{\text{SML}} (\times 10^{-3} \text{ s}^{-1})$ derived from the (a),(b) observations, and simulations with the (c),(d) KPP, (e),(f) MY, and (g),(h) $\kappa-\epsilon$ schemes. Results in (left) winter (February in the Northern Hemisphere and August in the Southern Hemisphere) and (right) summer (August in the Northern Hemisphere and February in the Southern Hemisphere).

observational Skew is 0.056 in winter and -1.3 in summer). This seasonal asymmetry of L_{SML} emphasizes the temporal variability of the SMLI scale, which is notably different from a seasonally unchanging length scale associated with the full-depth mesoscale instability mode (Chelton et al. 1998). The asymmetry implies that the seasonal variability of L_{SML} is more remarkable in the Northern Hemisphere than the Southern. To measure the seasonal variability of L_{SML} , a ratio of L_{SML} between summer and winter is calculated as

$$R_t = \frac{L_{\text{SML}}^{\text{sum}}}{L_{\text{SML}}^{\text{win}}} \times 100\%. \quad (9)$$

R_t describes the ratio of L_{SML} in summer relative to the value in winter. Here, an averaged R_t of the median, 10th percentile, and 90th percentile L_{SML} values

is shown (Fig. 8). According to R_t calculated from the observations, R_t varies along latitude, ranging from 116% at Southern Hemisphere high latitudes to 44% at Northern Hemisphere high latitudes (blue dots in Fig. 8). Generally, R_t is large with an average of 77% in the Southern Hemisphere, but decreases to 56% in the Northern Hemisphere. Thus, the MLI scale is more seasonal in the Northern Hemisphere. In the Southern Hemisphere, the median L_{SML} values at high latitudes (poleward of 60°S) are 6.2 km in winter and 4.1 km in summer, but 15 and 11 km at midlatitudes (equatorward of 60°S). In contrast, the median L_{SML} values are 6.0 and 3.3 km at high latitudes (poleward of 60°N), and 19 and 11 km at midlatitudes (equatorward of 60°N) in the Northern Hemisphere. The stronger seasonal variability of the MLI scale is obtained from the simulations as well, since a larger R_t in the North Hemisphere from the

FIG. 6. Global distributions of the spatial scale of the SMLI L_{SML} (km) derived from the (a),(b) observations, and simulations with the (c),(d) KPP, (e),(f) MY, and (g),(h) κ - ϵ schemes. Results in (left) winter (February in the Northern Hemisphere and August in the Southern Hemisphere) and (right) summer (August in the Northern Hemisphere and February in the Southern Hemisphere).

simulations can be observed similarly (red, green and yellow dots in Fig. 8).

4. The scale of bottom mixed layer instability

Unlike in the MITgcm, which has a coarse vertical grid spacing at depth, and according to Fig. 1, the GOTM simulations are capable of reproducing the ocean BML as well as the SML. Meanwhile, the QG instability analysis in Fig. 2 predicts MLI occurring with a vertical mode trapped in the BML (Wenegrat et al. 2018). In this section, the BMLI scale will be estimated and analyzed based on the GOTM simulation results.

The BML thickness, H_{BML} is derived from the simulation results with different turbulence closures (Fig. 9). Parameter H_{BML} varies remarkably with turbulence

closures, and unfortunately the Argo program does not measure the BML as it does in the SML and other observations are too sparse (Huang et al. 2019). Thus, for the time being one must remain content with the simulated range. The KPP scheme simulates the largest H_{BML} with a global median value of 273 m. The MY scheme gives a median H_{BML} of 268 m, and the κ - ϵ scheme has the smallest median H_{BML} of 102 m. Seasonally, the median H_{BML} in winter is slightly larger than that in summer (Table 1). The slight difference indicates an inconspicuous seasonal variability of the BML, which is true for all the simulations with different turbulence closures. Meanwhile, similar patterns of H_{BML} are observed between the different simulations. The most obvious similarity is located at the ACC region, where all of the three schemes show a thick BML compared with other regions, presumably due

FIG. 7. The zonal median L_{SML} in (a) winter and (b) summer. The blue, red, green, and yellow dots denote the results from the observations, and simulations with the KPP, MY, and κ - ϵ schemes, respectively. The shaded intervals denotes the corresponding bounds of the 10th and 90th percentile L_{SML} values zonally. The y axis is in a natural logarithmic coordinate.

to large shear of the ACC. Moreover, a relatively thick BML in the North Atlantic and a thin BML in the south-east Pacific are consistent. Inconsistency is found in the North Pacific basin. Both the KPP and MY schemes show a generally thick BML in the North Pacific basin (median H_{BML} of 250 m), while the κ - ϵ scheme produces a thin H_{BML} (median H_{BML} of 85 m).

It is difficult to obtain a global map of BML distribution based on observations, due to scarce observations in the deep basin especially near the ocean bottom. Based on hydrographic data, the work by Huang et al. (2019) shows a global BML thickness distribution with a spatial grid spacing of 10° . The median H_{BML} over the globe from their estimation is 47 m, much smaller than

FIG. 8. The average of R_t between summer and winter from the median, 10th percentile, and 90th percentile L_{SML} values. The blue, red, green, and yellow dots denote the results from the observations, and simulations with the KPP, MY, and κ - ϵ schemes, respectively. The shaded intervals show the minimum and maximum R_t of the median, 10th percentile, and 90th percentile L_{SML} values.

FIG. 9. Global distributions of the BML thickness H_{BML} (m) derived from the simulations with the (a),(b) KPP, (c),(d) MY, (e),(f) κ - ϵ schemes. Results in (left) winter (February in the Northern Hemisphere and August in the Southern Hemisphere) and (right) summer (August in the Northern Hemisphere and February in the Southern Hemisphere).

the values obtained here with simulations. Overall, the κ - ϵ scheme has a median H_{BML} closest to these observations. However, the observed H_{BML} has a thick BML in the North Pacific basin, similar to the simulations with the KPP and MY schemes. And the thick BML in the ACC region from the simulations is not observed. The inconsistencies may result from many reasons, especially observation accuracy and sampling, and limitations of turbulence closures. Nevertheless, the BMLI scale from simulations will be analyzed.

The depth averaged buoyancy frequency in the BML, N_{BML} varies two orders of magnitude (from 10^{-4} to 10^{-3} s^{-1}), one order smaller than N_{SML} (Fig. 10). Unlike the SML, the distribution of N_{BML} is correlated to not only H_{BML} , but also water depth. N_{BML} is generally weak at regions with deep water depth or thin BML. Hence, N_{BML} is stronger for simulations with the KPP and MY schemes compared to the one with the κ - ϵ scheme, which tends toward thinner BML. Moreover, N_{BML} is enhanced slightly from February to August, consistent with the slightly thickening H_{BML} .

The spatial scale of BMLI, L_{BML} estimated based on H_{BML} and N_{BML} , is much smaller than L_{SML} , ranging from hundreds of meters to up to 10 km (Fig. 11).

Latitude affects L_{BML} , again with polar scales smaller than equatorial ones. Nevertheless, a relatively large L_{BML} exists in the ACC, in addition to low latitudes, due to a thicker and more stratified BML in that region associated with stresses from relatively strong bottom currents. L_{BML} derived from the KPP and MY schemes tends to be larger than that from the κ - ϵ scheme, due to the thicker simulated H_{BML} and larger simulated N_{BML} . Globally, the median L_{BML} is 2.0 km for the KPP scheme, 1.4 km for the MY scheme, and 0.41 km for the κ - ϵ scheme. The latitude dependence of the L_{BML} variability is characterized by collecting statistics of L_{BML} zonally over all data in both months (Fig. 12). The zonal median L_{BML} derived from the KPP scheme is the largest, with a value of 3.7 km at low latitudes and

TABLE 1. Median H_{BML} derived from different turbulence schemes.

Scheme	H_{BML} (m)	
	Winter	Summer
KPP	276	270
MY	271	268
κ - ϵ	104	101

FIG. 10. Global distributions of the depth-averaged buoyancy frequency $N_{\text{BML}} (\times 10^{-3} \text{ s}^{-1})$ in the BML derived from the simulations with the (a),(b) KPP, (c),(d) MY, and (e),(f) κ - ϵ schemes. Results in (left) winter (February in the Northern Hemisphere and August in the Southern Hemisphere) and (right) summer (August in the Northern Hemisphere and February in the Southern Hemisphere).

0.93 km at high latitudes. The MY scheme gives a L_{BML} close to the KPP scheme with median scales from 3.8 to 0.72 km. The median L_{BML} from the κ - ϵ scheme is roughly smaller than 1 km from 0.97 to 0.13 km. Despite the quantitative differences, the three schemes depict similar patterns, including the peaks at the ACC region and high latitudes in the Northern Hemisphere.

In this work, regions with water depth deeper than 500 m over the globe are investigated, and effects of the topographic slope are not considered on the instability scale. Over these depths, slope angles are estimated to be smaller than 0.1 over 95% of the World Ocean (Costello et al. 2010). Theoretical analysis demonstrates that the features of BMLI with high wavenumbers are robust to the slope angle (Wenegrat et al. 2018), and BML isopycnal slopes will tend to be steep due to turbulent mixing, meaning that the key controlling ratio of isopycnal slopes to the topographic slope will remain large (Wang and Stewart 2018). The topographic slope thus is likely to have negligible impacts on the estimate of L_{BML} . Moreover, the estimation is mainly implemented over relatively flat seafloor. Regions with rough topography are not considered in this work, and these regions

usually have a thick BML due to strong mixing associated with internal waves (e.g., Nikurashin and Ferrari 2010).

5. Discussion and conclusions

a. Discussion

Seasonal activity of submesoscale MLI in the SML revealed by observations and model simulations has been demonstrated to be a result of ML thickness variation and energy cascade (Mensa et al. 2013; Callies et al. 2015; Dong et al. 2020). However, our analysis reveals that the SMLI scale has a remarkable seasonal variability as well, which can be decreased by more than 60% from winter to summer (Fig. 8). Hence, SMLI activity detected relies heavily on spatial grid spacings of models and observations. As observations show (Callies et al. 2015), SML submesoscales tend to be weaker in summertime, which could be a result of stronger solar restratification in competition with submesoscale restratification, or more rapid wind-driven restratification by upfront winds in shallower Ekman layers, or other potential mechanisms. What is clear from this study, however, is that modeling studies must have grid spacings

FIG. 11. Global distributions of the BMLI scale L_{BML} (km) derived from the (a),(b) KPP, (c),(d) MY, and (e),(f) κ - ϵ simulations. Results in (left) winter (February in the Northern Hemisphere and August in the Southern Hemisphere) and (right) summer (August in the Northern Hemisphere and February in the Southern Hemisphere).

sufficiently fine as to resolve the smallest potential submesoscales of the year before seasonality can be examined.

For context, the evolution of horizontal grid spacings of selected high-resolution global ocean models is shown in Fig. 13 (Following Fox-Kemper et al. 2014). The finest grid spacings reported by the IPCC reports are denoted by gray dots in the corresponding year of publication.

The only exception is the latest dot, which indicates the ground-breaking effective horizontal grid spacing of the ECCO MITgcm LLC4320 simulation. Up until this simulation, the evolution of the horizontal grid spacings is roughly consistent with the estimated scaling of faster computing by Moore's Law that predicts the model resolution doubling every six years (Moore 1965), although it is important to note the limited number of years simulated

FIG. 12. The zonal median L_{BML} . The red, green, and yellow dots denote the results from the simulations with the KPP, MY, and κ - ϵ schemes. The shaded intervals denotes the corresponding bounds of the 10th and 90th percentile L_{SML} values zonally.

FIG. 13. Estimate of horizontal grid spacings of the IPCC ocean models. The gray dots denote the finest grid spacings reported by the IPCC reports by year of publication, except the latest one from the ECCO MITgcm LLC4320 simulation by the publication year of Rocha et al. (2016). The black line denotes the estimate predicted by Moore's Law, while the shaded regions denote the grid spacing intervals resolving 50% and 90% SML eddies globally based on the observations and BML eddies based on simulations. This figure is reproduced based on Fig. 1 from Fox-Kemper et al. (2014).

in the MITgcm run versus the centennial-scale IPCC runs. On this same figure, required grid spacings to resolve submesoscale eddies in the ML under different circumstances are estimated. Considering that Eq. (7) estimates the MLI sinusoidal wavelength as L_{ML} , a required grid spacing resolving ML eddies should be near $L_{ML}/8$, since at least two grid cells per eddy radius (i.e., the Nyquist sampling rate for an eddy radius set to 1/4 the wavelength) are necessary to resolve eddies. Shaded regions in Fig. 13 denote the grid spacing intervals resolving ML eddies in 50% and 90% of the analysis regions globally. According to the estimated L_{SML} from the observations, the required grid spacings to resolve winter and summer 50% SML eddies globally are 1.9 and 1.3 km, respectively (upper bounds of the blue and light blue intervals). The grids decrease to 0.92 km in winter and 0.55 km in summer to resolve SML eddies in 90% of the regions (lower bounds of the blue and light blue intervals). Regionally speaking, the required grid spacings at midlatitudes are 2.0 km (50% resolved) and 1.0 km (90% resolved) in winter, and 1.4 km (50% resolved) and 0.66 km (90% resolved) in summer. At high latitudes, the grid spacings become smaller, which are 0.78 km (50% resolved) and 0.47 km (90% resolved), and 0.50 km (50% resolved) and 0.37 km (90% resolved) in summer.

Up to now, most of the regional or global models used for submesoscale simulation have horizontal grid spacings around 1–2 km. Taking the LLC4320 model shown in Fig. 13 as an example, the model has an effective grid spacing of 2.1 km globally (poleward of 10°). The grid spacing allows the model to resolve about 46% of the SML eddies in winter based on the estimates in this work, mainly concentrating at midlatitudes. But in summer, the model can only resolve about 26% of the SML eddies globally, mainly in the ACC (Su et al. 2018). Hence, claims about the submesoscale seasonality need to be carefully weighed.

In the BML, the required grid spacings to resolve BML eddies are much smaller. The required grid spacings to resolve submesoscale eddies in BML are estimated to be as low as 257 m (50% resolved) and 107 m (90% resolved) under the KPP scheme, 178 m (50% resolved) and 87 m (90% resolved) under the MY scheme, and 51 m (50% resolved) and 17 m (90% resolved) under the κ - ϵ scheme. These grid spacings beyond the reach of current ocean models for submesoscale simulation (Fig. 13), especially when considering that increased vertical grid spacing as in the GOTM simulations here is likely also required. The estimated BML scale varies among turbulence closures, indicating that grid spacings in submesoscale-permitting models will differ based on what boundary layer schemes are used; present direct observations are insufficient for a global perspective.

According to Moore's Law, which tracks the IPCC high-resolution runs well, the timeline when the MLI are routinely resolved in an IPCC-class simulation is estimated. It will take about 25 years until coupled IPCC models completely resolve SML eddies including summertime, which will be achieved by approximately 2045. But for BML eddies, at least 40 years are needed (the KPP and MY schemes), i.e., around 2060. The wait is even longer for the boundary layers simulated with the κ - ϵ scheme: about 55 years. Thus, subgrid parameterization for ML eddies and process studies at higher grid spacing will remain important for decades.

b. Conclusions

The scale of baroclinic instability in the SML and BML, including spatial and seasonal variabilities, is investigated over the global ocean in this work, based on observations and simulations. The analysis indicates that SMLI and BMLI differ significantly regionally and seasonally due to their different forcings.

As captured by both the observations and simulations, the spatial distribution of the SMLI scale is tightly dependent on latitude, smaller toward the poles and smallest

at the summer pole. The median L_{SML} from the observations varies from as large as 51 km at low latitudes to 2.9 km at high latitudes. Due to the seasonality of ML thickness and buoyancy frequency in the ML, the SMLI scale is larger in winter than summer. The median L_{SML} value globally is 16 km in winter and decreases to 10 km in summer. The seasonal variation is more obvious in the Northern Hemisphere, and the SMLI scale in summer is 1.8 times smaller than that in winter. As a comparison, the scale in summer is 1.3 times smaller in the Southern Hemisphere. The seasonality of SMLI scale leads to a symmetry of L_{SML} between two hemispheres. The skewness of L_{SML} demonstrates that the SMLI scale is slightly larger in the Northern Hemisphere in winter with a Skew of 0.056, but reverses in summer with a Skew of -1.3 . Based on the estimated L_{SML} , the required model grid spacings to resolve ML eddies globally are 1.9 km (50% resolved) and 0.92 km (90% resolved) in winter, and 1.3 km (50% resolved) and 0.55 km (90% resolved) in summer.

The BML thicknesses derived from different turbulence schemes differ quantitatively with globally median H_{BML} of 273 m for the KPP, 268 m for the MY and 102 m for the κ - ϵ . Nevertheless, the three turbulence schemes depict similar patterns of L_{BML} with slight seasonality. And BMLI has a scale much smaller than SMLI with a magnitude of $O(1)$ km due to weak stratification and small stresses near the bottom. The latitude dependence of L_{BML} is still observed with a decreasing trend with latitude increasing, but relatively large L_{SML} exists in the ACC region. The required grid spacings to resolve submesoscale eddies in the BML is estimated as 257 m (50% resolved) and 107 m (90% resolved) under the KPP scheme, 178 m (50% resolved) and 87 m (90% resolved) under the MY scheme, and 51 m (50% resolved) and 17 m (90% resolved) under the κ - ϵ scheme. It will take 40–55 years for an IPCC-class coupled ocean model to resolve them.

Some recent studies have elected to define the submesoscale as being smaller than 50 km, so as to label the scales that upcoming satellite missions will be able to sample as “submesoscale.” This study suggests that such a definition is misleading in two ways: this scale is too large over most of Earth, and strong variations in H_{ML} , N_{ML} , and especially f with latitude, as well as season and region, make such a definition by a single dimensional estimate less useful than a dynamical scale estimate. Mesoscale eddy scales also vary spatially (Chelton et al. 1998), but depth and stratification of the interior ocean are only modestly seasonal. The extreme seasonal and regional variability of the SML makes the variation in submesoscale eddy scales much larger than those of mesoscale eddies.

Finally, the baroclinic MLI is not the only generation mechanism for submesoscale structures. Large Eddy Simulations have shown that symmetric instability (Bachman et al. 2017), frontogenesis and filamentogenesis (McWilliams 2016), Stokes forces and submesoscale straining (Suzuki et al. 2016), or turbulent thermal wind (Sullivan and McWilliams 2018) can lead to structures that are smaller than MLI, but are still hydrostatic and affected by rotation and stratification at leading order: i.e., these too are dynamical denizens of the submesoscale. These submesoscale structures may also tend to have seasonal variability in the SML, since their generation is highly dependent on air–sea fluxes (momentum, buoyancy), horizontal buoyancy gradient and flow strain. Symmetric instability occurs when the PV has an opposite sign to the local planetary vorticity, indicating that it should be more active in winter due to relatively stronger Ekman buoyancy flux and sea surface cooling (Dong et al. 2020, manuscript submitted to *J. Phys. Oceanogr.*). Meanwhile, relatively stronger wind in winter may also favor submesoscales due to enhanced Stokes force and wind stirring in the SML. Given that geostrophic strain is usually stronger in summer, the activity of submesoscales generated by the frontogenesis may be determined by the activity of submesoscale fronts and strain related to MLI.

Acknowledgments. The authors acknowledge support from the National Key Research Program of China (2017YFA0604100). JHD partially acknowledges support from the National Natural Science Foundation of China (91958205, 41806025) and the China Scholarship Council. BFK is partially supported by NSF OCE-1350795 and ONR N00014-17-1-2963. CMD appreciates the support from National Key Research and Development Program of China (2016YFA0601803), the National Natural Science Foundation of China (41476022, 41490643), the National Programme on Global Change and Air-Sea Interaction (GASI-03-IPOVAI-05), and the foundation of China Ocean Mineral Resources R & D Association (DY135-E2-2-02, DY135-E2-3-01). Shafer Smith kindly provided the quasigeostrophic instability code. The authors also acknowledge the data support from Dimitris Menemenlis’s group from the Jet Propulsion Laboratory. The LLC4320 data are available on ECCO Data Portal (<https://data.nas.nasa.gov/ecco/data.php>), and can be downloaded conveniently using the xmitgcm python package (<https://xmitgcm.readthedocs.io/en/latest/index.html>). The ARGO data are available for download from NOAA National Centers for Environmental Information (https://www.nodc.noaa.gov/argo/basins_data.htm). The data for reproducing figures in this work are available at <https://doi.org/10.26300/s501-ew23>.

TABLE A1. Methods for ML thickness calculation. The potential temperature θ and potential density σ are calculated with zero reference pressure for SML thickness determination and a reference pressure corresponding to local water depth for BML determination. The subscripts of 10 and H denote 10-m depth and local water depth.

Method		SML	BML
Thrd	Temperature	$\theta = \theta_{10} \pm 0.2^\circ\text{C}$	$\theta = \theta_H \pm 0.005^\circ\text{C}$ $\theta = \theta_H \pm 0.001^\circ\text{C}$
	Density	$\sigma = \sigma_{10} + 0.03 \text{ kg m}^{-3}$ $\sigma = \sigma_{10} + \Delta\sigma, \Delta\sigma = \sigma(\theta - 0.2^\circ\text{C}) - \sigma(\theta)$	$\gamma_z = 10^{-5} \text{ kg m}^{-4}$
	Dissipation	$\varepsilon = 10^{-11} \text{ W}$	$\varepsilon = 10^{-11} \text{ W}$
Rvar	Density	$\chi \rightarrow \min$	$\chi \rightarrow \min$

APPENDIX

Methods for Mixed Layer Thickness Calculation

Both the SML and BML thicknesses are calculated using different methods, and these methods are clarified into two major types, threshold and relative variance methods (Table A1). For consistency between different methods, the starting depth is 10 m for SML thickness calculation and is equal to local water depth for BML thickness calculation. The potential temperature θ and potential density σ are calculated with zero reference pressure for SML thickness determination and a reference pressure corresponding to local water depth for BML thickness determination.

- 1) Threshold method (Thrd): the threshold method is applied on different variables, mainly including temperature and density. For the SML, ML thickness is determined with a fixed temperature threshold of 0.2°C , a fixed density threshold of 0.03 kg m^{-3} , a variable density threshold equivalent to a 0.2°C temperature decrease (de Boyer Montégut et al. 2004, 2007). For the BML, ML thickness is determined with a fixed temperature threshold of 0.005° and 0.001°C (Hogg et al. 1982; Lozovsky and Shapovalov 2012; Huang et al. 2018), and a fixed vertical neutral density gradient threshold of $1 \times 10^{-5} \text{ kg m}^{-4}$ (Banyte et al. 2018). Moreover, an energy dissipation threshold of $1 \times 10^{-11} \text{ W}$ is applied for ML thickness determination as well. This method determines a thickness named mixing layer thickness, little different from ML thickness.
- 2) Relative variance method (Rvar): this method determines ML thickness based on a relative variance, χ obtained from potential density profiles (Huang et al. 2018). For each density profile, the relative variance χ is defined as a ratio of the root-mean-square relative to the absolute difference from a reference depth level. The depth at the minimum value of the relative variance is used to identify the boundary of the ML. Details of the algorithm can be referred to

Huang et al. (2018). Both the SML and BML are calculated using this method.

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