

Waves Explain Observed Similarity Scaling of Turbulence Dissipation in the Southern Ocean

ISABELLE GIDDY^{a,b,f,g}, DANIEL WHITT^b, ILKER FER^c, SEBASTIAAN SWART^{a,d}, ØYVIND BREIVIK^{c,e} AND SARAH-ANNE NICHOLSON^f

^a *University of Gothenburg, Gothenburg, Sweden*

^b *NASA Ames Research Center, Moffett Field, California*

^c *Geophysical Institute, University of Bergen, Bergen, Norway*

^d *Department of Oceanography, University of Cape Town, Rondebosch, South Africa*

^e *Norwegian Meteorological Institute, Oslo, Norway*

^f *Southern Ocean Carbon-Climate Observatory, Smart Places, CSIR, Cape Town, South Africa*

^g *California State University, Monterey Bay, Seaside, California*

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ABSTRACT: We investigate the turbulence kinetic energy (TKE) budget of the upper ocean and its response to wind and wave forcing in the Southern Ocean, using realistically forced large-eddy simulations (LESs) and in situ microstructure shear observations. The widely used law of the wall similarity scaling assumes that shear production of TKE is balanced by its dissipation. However, our findings reiterate that this assumption is violated under wave forcing: Dissipation is primarily balanced by local Stokes shear production, augmented by almost equal contributions from local Eulerian shear production and nonlocal convergence of TKE transport. Despite this, the canonical law of the wall scaling reasonably describes the vertical distribution of dissipation rates in the boundary layer in both realistically forced LES and observations, even though the underlying physical reasoning does not hold. We propose a modified scaling to account for the nonlocal component of the TKE budget that yields accurate predictions of the TKE budget as well as an interpretation accounting for wave effects. These insights have important implications for interpreting turbulence dissipation rate observations.

SIGNIFICANCE STATEMENT: This study provides and evaluates a physical model of turbulence in the near-surface layer of the Southern Ocean under strong winds and waves. This turbulence model quantifies how the vertical transport and dissipation of turbulence relate to the production of turbulence due to winds and waves. This model of turbulence is valuable because the Southern Ocean plays a key role in regulating climate through the uptake of heat and carbon from the atmosphere, which is mediated by small-scale turbulent motions that generally must be estimated without turbulence observations in global ocean and climate simulations. In addition, this model provides insight into dissipation rate observations, which are generally collected without observations of the corresponding turbulence production and transport.

KEYWORDS: Southern Ocean; Turbulence; Oceanic mixed layer; Surface layer; In situ oceanic observations; Large eddy simulations

1. Introduction

The Southern Ocean plays a disproportionately important role in the climate system due to its capacity to absorb carbon and heat (Williams et al. 2024; Gruber et al. 2019; Marshall et al. 2015). This uptake occurs through exchanges of momentum, heat, tracers, and gases in the ocean surface boundary layer (OSBL) that are regulated by small-scale turbulent motions. These turbulent motions must be parameterized in ocean models. Furthermore, because turbulence generation is concentrated in the surface layer—the uppermost 10% of the OSBL—it is essential to accurately represent the ocean surface layer's response to atmospheric and wave forcing.

Turbulence in the surface layer is often described using frameworks originally developed for the atmospheric boundary

layer, where surface stress drives local shear production near a rigid boundary. However, in contrast to the atmosphere which is treated as a no-slip boundary, the ocean surface is strongly influenced by both breaking and nonbreaking surface gravity waves that modify turbulence production and transport. Breaking waves enhance dissipation in the near-surface layer (~upper 1 m; Agrawal et al. 1992; Terray et al. 1996; Wang et al. 2018). Nonbreaking waves influence turbulence through Stokes drift. The vertical shear associated with Stokes drift decays exponentially with depth and acts as a source of turbulence production. Its interaction with the mean ocean current shear reorients vertical vorticity and generates counterrotating vortices aligned with the wind (Craig and Leibovich 1976; Leibovich 1983), known as Langmuir circulations. These circulations organize the flow into convergence zones and horizontal rolls, often visible as surface windrows (Smith 1992; Kukulka et al. 2009). Evidence from both observations and large-eddy simulation (LES) shows that compared to a wind-forced boundary layer with no waves, Langmuir circulations are associated with enhanced vertical velocities (e.g., Polton and Belcher 2007), turbulence (e.g., McWilliams et al. 1997; Grant and Belcher 2009; D'Asaro et al. 2014; Ferris et al. 2022;

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Corresponding author: Isabelle Giddy, isgiddy@gmail.com

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Cifuentes-Lorenzen et al. 2024), and turbulent transport (e.g., Gerbi et al. 2009; Kukulka et al. 2012; Chen et al. 2016; Chor et al. 2021; Jarosz et al. 2021).

Since appropriate oceanic measurements of simultaneous forcing and all the necessary terms of the turbulence kinetic energy budget are not attainable, LESs (Skylvingstad and Denbo 1995; Harcourt and D'Asaro 2008; Grant and Belcher 2009; Noh et al. 2011; Kukulka et al. 2012; Large et al. 2019a,b, 2021b; Skylvingstad et al. 2023) have been used to understand the role of Langmuir turbulence in shaping the OSBL structure and energetics. Realistically forced LESs more closely reproduce observed boundary layer properties when Langmuir effects are included (e.g., Kukulka et al. 2012; Fan et al. 2020; Skylvingstad et al. 2023; Pham et al. 2023). Further, the incorporation of Langmuir effects into OSBL parameterizations in ocean general circulation models (OGCMs) has reduced the shallow bias in mixed layer depth (MLD) (Li et al. 2019). However, substantial differences remain between parameterization schemes, particularly in the Southern Ocean during summer, where Langmuir turbulence is considered important (Belcher et al. 2012; Dong et al. 2024) due to strong winds and persistent wave fields with nearly unlimited fetch that drive intense upper-ocean mixing.

A widely used approach for diagnosing OSBL turbulence under neutral forcing is to compare observed dissipation rates (an observable variable) to predictions from the law of the wall. This classic framework assumes steady-state, horizontally homogeneous flow, where surface stress drives a local shear production that is entirely dissipated. Deviations from the scaling can reflect either a breakdown in the underlying assumptions behind the law of the wall or the presence of additional sources or sinks of turbulence kinetic energy (TKE). The law of the wall scaling was extended to include cases where buoyancy forcing has a nonnegligible effect on shear production by Monin–Obukhov similarity theory (MOST; Monin and Obukhov 1954) and seminal work by Lombardo and Gregg (1989) illustrated a scaling for dissipation under mixed wind and convective conditions. Several subsequent observation-based studies illustrate deviations from the law of the wall when surface gravity waves become important, specifically pointing toward enhanced or reduced magnitudes of dissipation (e.g., D'Asaro et al. 2014; Sutherland et al. 2014; Esters et al. 2018; Zheng et al. 2021). Accordingly, MOST has been modified to account for Stokes shear (e.g., McWilliams and Sullivan 2000; Smyth et al. 2002; Large et al. 2019a, 2021b). In particular, the assumption of shear production–dissipation balance when overturning cells are present, such as under strong convection or when there are Langmuir cells, does not hold. This was demonstrated for the atmospheric surface layer by Wyngaard and Coté (1971), where there is an imbalance between shear production and dissipation due to nonlocal turbulent transport when large convective cells are present. Under wave forcing, nonlocal transport has also been found to be important and was subsequently incorporated into the similarity scaling framework of Large et al. (1994) either via an additional nonlocal term (e.g., Chor et al. 2021) or through modifications to the structure function (e.g., Large et al. 2019b).

In the Southern Ocean—where effects of wave-driven turbulence might be expected to be large—enhanced dissipation relative to the law of the wall is not consistently observed. For example, in a campaign in the Drake passage, predictions were largely biased low compared to observations (Ferris et al. 2022); however, in two campaigns further to the east, predictions were equal or biased high (Nicholson et al. 2022; Giddy et al. 2023). This contrast is notable given that all campaigns took place during austral summer under substantial wave forcing. Interacting factors may explain these differences. Under conditions of wind-wave misalignment (e.g., Van Roekel et al. 2012) or when waves interact with fronts (e.g., Vrećica et al. 2022), Langmuir circulations weaken, and turbulence is reduced below expected levels. Additionally, under conditions of Stokes forcing, it has been shown that turbulent transport plays a nonnegligible role in the TKE budget throughout the boundary layer (e.g., Gerbi et al. 2009; Kukulka et al. 2012; Chor et al. 2021), violating the local balance of shear production and dissipation that is assumed in the law of the wall.

In this study, we use realistically forced LES combined with a suite of dissipation rate observations to show how the underlying assumptions of the law of the wall framework break down under wind-wave forcing. Building on this, we propose a modified similarity scaling that incorporates nonlocal TKE transport, offering a more physically grounded approach to the application of similarity scaling in wave-forced ocean boundary layers.

Section 2 summarizes the relevant similarity scaling theory; section 3 describes the data and methods. In section 4, we analyze similarity scaling relative to resolved turbulence in realistically forced LES and introduce a modified framework. Section 5 compares various scaling approaches to observations, and section 6 discusses the broader implications.

2. Theoretical framework

a. Turbulent kinetic energy

The velocity of a turbulent flow can be split into a mean flow and a fluctuating flow ($\mathbf{U} = \bar{\mathbf{u}} + \mathbf{u}'$). TKE is the kinetic energy per unit mass associated with the fluctuating flow. For a horizontally homogeneous flow, with no mean advection, and assuming a rigid lid where the effects of surface gravity waves are omitted,

$$\partial_t \text{TKE} = -\langle w' \mathbf{u}' \rangle \cdot \partial_z \mathbf{U} + \langle w' b' \rangle + \text{transport} - \varepsilon, \quad (1)$$

where $\partial_t \text{TKE}$ is the rate of change of TKE, $\langle \rangle$ denotes the time and/or horizontal space averaging, primed values are the turbulent fluctuations, and b is the buoyancy, with $b = -g(\rho/\rho_0)$, where ρ is the seawater density and ρ_0 is a reference density. We use a right-handed Cartesian coordinate system with z increasing upward. The first term on the right-hand side represents the production of TKE through the vertical shear of mean horizontal currents, $\mathbf{U}$, the Eulerian shear. The second term is the turbulent buoyancy flux, which is negative and a sink of TKE under stabilizing surface buoyancy forcing and positive and a source of TKE under destabilizing surface buoyancy forcing. The transport term represents divergence of TKE transport—hereinafter “TKE transport”—due to turbulent fluctuations of the vertical

velocity and pressure. The final term, ε , represents the rate of destruction of TKE due to viscosity (turbulence dissipation).

b. Similarity theory

Semiempirical MOST (Monin and Obukhov 1954; Wyngaard 2010; Foken 2006) provides a framework for describing planetary boundary layers under horizontally homogeneous and stationary conditions, assuming a rigid-lid boundary. It relates turbulent quantities to distance from the surface, $|z|$, using only the surface forcing parameters.

The boundary layer of thickness h , taken as the actively mixing layer in contact with the surface, includes the viscous sublayer directly influenced by the surface ($\sim 0.01\text{--}0.03h$), the surface layer, and the bulk of the boundary layer. Across all observations and simulations analyzed in this study, we identify h as the first depth from the surface, where $\varepsilon \leq 10^{-8} \text{ m}^2 \text{ s}^{-3}$ (Brainerd and Gregg 1995). The surface layer is defined as the layer extending from the viscous sublayer to 10% of the height of the boundary layer ($0.1h$). Within this layer, the turbulent momentum and buoyancy fluxes are approximately constant. Surface forcing is characterized by the friction velocity, $u_* = \sqrt{\tau/\rho_0}$, where τ is the magnitude of the turbulent surface stress vector, and by the surface buoyancy flux, B_o , from which we define the buoyancy forcing scale $b^* = B_o/u_*$. The turbulent stress is related to the turbulent momentum flux via $\tau = -\rho_0 \langle w' \mathbf{u}' \rangle$. In MOST, the vertical shear and stratification profiles, characterized by the friction velocity and the surface buoyancy flux, respectively, become functions of a single dimensionless group—the stability parameter—which depends on z and the Monin–Obukhov length scale, L (Wyngaard 2010):

$$\zeta = \frac{z}{L} = \frac{\kappa z B_o}{u_*^3}, \quad (2)$$

with

$$L = \frac{u_*^3}{\kappa B_o}, \quad (3)$$

where $\kappa = 0.4$ is the universal von Kármán constant and L defines the depth at which the shear production of turbulence balances buoyancy production.

The nondimensional shear is given as

$$\Psi_m = \frac{\kappa z}{u_*} \partial_z U = \phi_m(\zeta), \quad (4)$$

determined experimentally in atmospheric boundary layers under unstable conditions as a function of the stability parameter (Businger et al. 1971; Höglström 1988) and fine-tuned to the Southern Ocean using LES (Large et al. 2019a):

$$\phi_m = (1 - 14\zeta)^{-1/3}. \quad (5)$$

Assuming all shear production is generated by mechanical forcing at the surface of the ocean and buoyancy is negligible such that $\Psi_m = \phi_m(\zeta) = 1$, the classic law of the wall scaling for the shear production of turbulence emerges where

$$-\langle w' \mathbf{u}' \rangle \cdot \partial_z \mathbf{U} = u_*^2 \left[\frac{u_*}{\kappa |z|} \right]. \quad (6)$$

Under certain conditions (e.g., wind-driven shear production), this scaling can be applied below the surface layer (i.e., beyond $0.1h$) but requires assuming a model (often linear) for how the turbulent stress decays to zero at the base of the boundary layer (Stull 1988; Watkins and Whitt 2020; Zippel et al. 2022; Zeiden et al. 2024) and that the stress aligns with the shear at all depths:

$$-\langle w' \mathbf{u}' \rangle = u_*^2 \left[1 - \frac{|z|}{h} \right], \quad (7)$$

giving

$$-\langle w' \mathbf{u}' \rangle \cdot \partial_z \mathbf{U} = u_*^2 \left[1 - \frac{|z|}{h} \right] \left[\frac{u_*}{\kappa |z|} \right]. \quad (8)$$

The ocean surface boundary layer differs from the atmospheric boundary layer in two ways. First, the coupling between the speed of the bulk fluid and the stress u_* differs since u_* is mainly defined by the wind in both cases. The ocean thus permits greater decoupling between the bulk current speed and the stress and the surface layer shear—the latter two of which similarity posits are tightly coupled and analogous in both atmospheric and ocean cases. Second, and the focus of this study, is the presence of surface gravity waves that introduce an additional source of turbulence production not captured by MOST. In particular, the interaction between wind and waves drives Langmuir turbulence, which requires modifying the similarity framework to include wave effects.

1) LANGMUIR TURBULENCE

Langmuir turbulence can be described through the Craik–Leibovich wave-averaged equations (Craik and Leibovich 1976), with Stokes drift being an imposed forcing (McWilliams et al. 1997; Belcher et al. 2012), which results in an additional term in Eq. (1)—Stokes shear production:

$$\partial_t \text{TKE} = -\langle w' \mathbf{u}' \rangle \cdot \partial_z \mathbf{U} - \langle w' \mathbf{u}' \rangle \cdot \partial_z \mathbf{U}_s + \langle w' b' \rangle + \text{transport} - \varepsilon, \quad (9)$$

where Stokes shear production is $-\langle w' \mathbf{u}' \rangle \cdot \partial_z \mathbf{U}_s$.

Large et al. (2019a) (hereafter L19) build on the MOST framework to develop a turbulent velocity scaling that includes the effects of Stokes drift and Langmuir turbulence, taking Stokes drift as an independent surface forcing. L19 introduce a second nondimensional function to Eq. (4), the Stokes similarity function, a function of the Stokes parameter, ξ , representing the proportion of Stokes production of TKE to the total production of TKE (the sum of shear production, Stokes shear production, and buoyancy production), such that

$$\Psi_m = \phi_m(\zeta) \chi_m(\xi), \quad (10)$$

with $\chi_m(\xi)$ being an empirically defined function fit to L19 LES output as

$$\chi_m = 1.05 + 2.43\xi + 1.69\xi^2, \quad (11)$$

where

$$\xi = \frac{\mu_s^3}{\mu_u^3 + \mu_s^3 + \mu_b^3}, \quad (12)$$

and $\mu_u^3 = P_u u_*^3$, $\mu_s^3 = P_s \text{La}^{-2} u_*^3$, and $\mu_b^3 = P_b w_*^3$ defining the cubes of three velocity scales for the surface forcing. The turbulent Langmuir number is defined as $\text{La} = \sqrt{u_* U_s}$ (McWilliams et al. 1997), where $\text{La} \leq 0.3$ indicates wave-driven turbulence conditions, and $w_*^3 = -B_o h$ defines the convective velocity scale. The empirically derived parameters, P_u and P_s , are based on turbulence and forcing in the L19 LES, with $P_b = 0.09$. This empiricism is needed as the surface forcing given by u_* , w_* , and La do not account for all the variability in TKE production in the surface layer. The parameters defined in L19 are specific to $0.1h$, but L19 shows that χ_m is depth invariant at least to $0.25h$ (their Fig. 6a).

Using the stability parameter ζ and the Stokes parameter ξ defined above, the Eulerian shear is scaled to account for both Stokes and surface buoyancy forcing. Thus, the law of the wall scaling for shear production given in Eq. (6) can be extended for unstable, wavy conditions:

$$-\langle \mathbf{w}'\mathbf{u}' \rangle \cdot \partial_z \mathbf{U} - \langle \mathbf{w}'\mathbf{u}' \rangle \cdot \partial_z \mathbf{U}_s = u_*^2 \left[1 - \frac{|z|}{h} \right] \left[\phi_m(\zeta) \chi_m(\xi) \frac{u_*}{\kappa|z|} + \partial_z \mathbf{U}_s \right]. \quad (13)$$

2) TURBULENCE DISSIPATION

Under the assumptions of stationarity, negligible turbulent transport, neutral buoyancy forcing, and the absence of surface gravity waves, the shear production of TKE is locally balanced by dissipation:

$$\varepsilon = -\langle \mathbf{w}'\mathbf{u}' \rangle \cdot \partial_z \mathbf{U}. \quad (14)$$

This relation forms the basis for approximating turbulence dissipation using the MOST framework {the law of the wall approximation under neutral buoyancy forcing—Eq. (6) and its extensions [Eqs. (7) and (8)]}.

However, under neutral conditions and in the presence of surface waves, two modifications need to be accounted for 1) a shear term associated with Stokes drift and 2) the vertical transport of TKE which becomes nonnegligible (Gerbi et al. 2009; Polton and Belcher 2007; Kukulka et al. 2012), as under strong destabilizing conditions (Wyngaard and Coté 1971). Although this transport term redistributes TKE vertically and integrates to zero over the full boundary layer, it modifies the local TKE balance and must be explicitly considered when predicting dissipation profiles.

Thus, for neutral buoyancy with both wind and wave forcing and no significant TKE tendency, $\partial/\partial t$ TKE, the local dissipation rate reflects contributions from Eulerian shear production, Stokes shear production, and the vertical transport of TKE:

$$\varepsilon = -\langle \mathbf{w}'\mathbf{u}' \rangle \cdot \partial_z \mathbf{U} - \langle \mathbf{w}'\mathbf{u}' \rangle \cdot \partial_z \mathbf{U}_s + \text{transport}. \quad (15)$$

Under conditions of pure wind forcing, it can be assumed that the stress and shear components of the shear production term are aligned. However, under the presence of waves, stress and shear are often misaligned (Van Roekel et al. 2012).

We define the magnitude of alignment between the stress and shear components as Θ , the normalized dot product of the stress and shear, which can be interpreted as the cosine of the angle between the stress and shear vectors such that a value of one indicates that the stress and shear are aligned:

$$\Theta_E = -\frac{\langle \mathbf{u}'\mathbf{w}' \rangle \cdot \partial_z \mathbf{U}}{|\langle \mathbf{u}'\mathbf{w}' \rangle| |\partial_z \mathbf{U}|}, \quad (16)$$

$$\Theta_S = -\frac{\langle \mathbf{u}'\mathbf{w}' \rangle \cdot \partial_z \mathbf{U}_s}{|\langle \mathbf{u}'\mathbf{w}' \rangle| |\partial_z \mathbf{U}_s|}. \quad (17)$$

With this, the Eulerian shear production can be represented as

$$-\langle \mathbf{u}'\mathbf{w}' \rangle \cdot \partial_z \mathbf{U} = \Theta_E |\langle \mathbf{u}'\mathbf{w}' \rangle| |\partial_z \mathbf{U}|, \quad (18)$$

and Stokes shear production can be represented as

$$-\langle \mathbf{u}'\mathbf{w}' \rangle \cdot \partial_z \mathbf{U}_s = \Theta_S |\langle \mathbf{u}'\mathbf{w}' \rangle| |\partial_z \mathbf{U}_s|. \quad (19)$$

L19 provides a framework for approximating Eulerian shear production under buoyancy and wave forcing but does not extend their results to directly predict turbulence dissipation rates and thus did not consider nonlocal transport in their framework. While the qualitative contribution of transport has been considered in a number of studies (e.g., Gerbi et al. 2009; Polton and Belcher 2007; Kukulka et al. 2012; Chen et al. 2016; Chor et al. 2021), its quantitative contribution to turbulence dissipation remains poorly constrained, particularly in observations.

3. Description of large-eddy simulations and observational data

a. Study sites

The observations and simulations described in this study are from three regions of the Southern Ocean (Fig. 1a): 1) the subantarctic zone (SAZ, glider deployed at 0° , 42°S and LES initialized and forced with conditions at the Southern Ocean flux site, 46°S , 142°E ; Schulz et al. 2012), 2) the Antarctic polar frontal zone (APZ, glider deployed at 0° , 54°S), and 3) within the seasonal ice zone (SIZ, glider deployed at 0° , 60°S)—a region that is covered by sea ice in the winter and experiences open ocean conditions during summer. The boxes in Fig. 1 identify the regions, where the gliders were deployed, and the location of the Southern Ocean flux site (SOFS) mooring, whose oceanic and forcing conditions informed the LES. For scale, the gliders were piloted in bow-tie patterns of 20–40 km except for the SAZ glider which maintained a mooring sampling pattern with each surfacing approximately 4 h apart. The

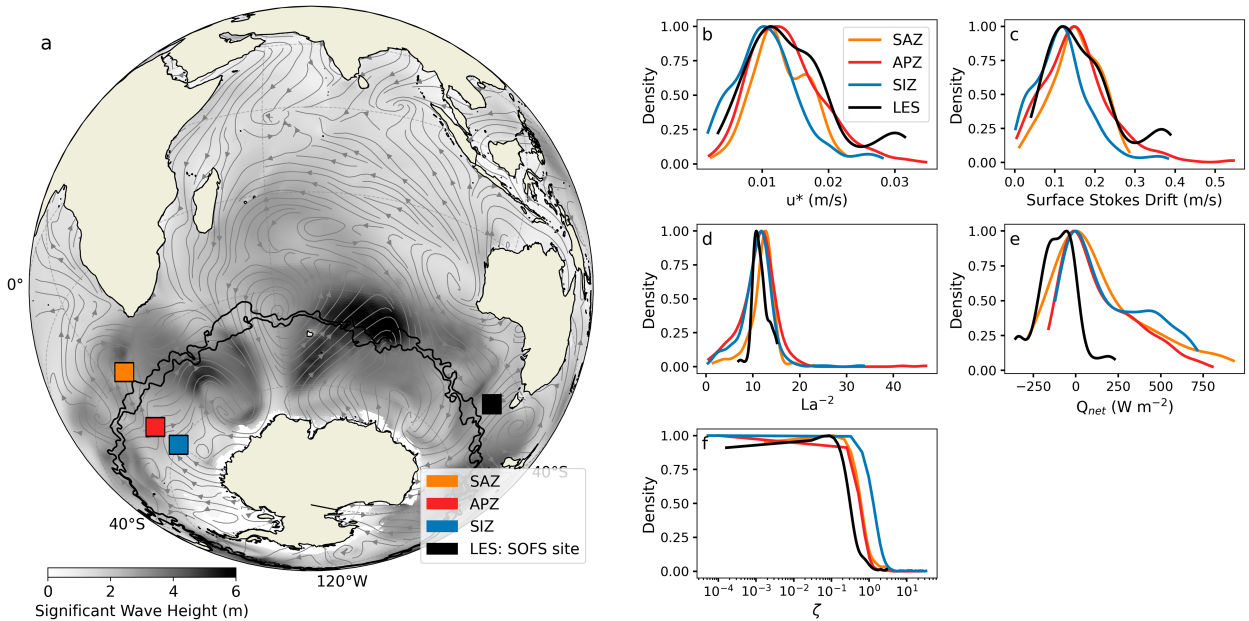


FIG. 1. (a) Locations of LESs (black: SOFS, 46°S, 142°E), and three glider MicroRider deployment sites (orange: SAZ, 42°S, 7°E; red: polar frontal zone, 0°, 64°S; blue: SIZ, 0°, 60°S). Gray color scale shows the mean ERA5 significant wave height for January 2019, with streamlines of ERA5 wind speed at 10 m for the same time period. Solid black contours indicate the mean location of the subantarctic front [northern contour; zero absolute dynamic topography (ADT) contour] and Antarctic polar front (southern contour; -0.3 ADT) during the same period (Sallée et al. 2008). (b)–(e) The distributions of the forcing conditions at each of the study sites: (b) friction velocity; (c) surface Stokes drift; (d) turbulent Langmuir number, $La^{-2} = U_s/u_*$; (e) net heat flux; and (f) stability parameter, ζ .

LES horizontal domain was $1250 \text{ m} \times 1250 \text{ m}$ and 300 m in the vertical.

These regions are hydrographically distinct in the water mass structure (Du Plessis et al. 2022); however, the surface forcing that they experience during the time of sampling is similar, dominated by strong wind and wave forcing with net buoyancy flux into the ocean on average (Figs. 1b–f). The LES runs were forced with stronger destabilizing buoyancy fluxes (net flux out of the ocean that occurred during austral autumn, April and June) compared to the glider observations (taken during austral summer, December–February; Fig. 1e). For the purpose of this study, we subset the data for destabilizing conditions in both observations and LES. Wind-wave conditions are typically fully developed (Fig. 1f, e.g., $La^{-2} \sim 11$; Belcher et al. 2012; L19), and buoyancy forcing is weak compared to shear ($|\zeta| = |z/L| \leq 1$, where z is taken at 10 m).

b. Large-eddy simulations

Realistically forced LES solutions of the ocean boundary layer representing conditions at the SOFS with and without the surface wave effects encapsulated in the wave-averaged Craik–Leibovich equations during April and June of 2010 are used in this study (Large et al. 2021a). This NCAR LES model is well documented (Moeng 1984; Sullivan et al. 1994; McWilliams et al. 1997). Modifications to the model to account for realistic surface wave effects are given by L19 which has been used to study wave forced Southern Ocean conditions by L19 and Large et al. (2019b, 2021b). The model solves the wave-averaged, incompressible, and Boussinesq Craik–Leibovich equation

set (Craik and Leibovich 1976; McWilliams et al. 1997). The computational domain spans a horizontal area of $1250 \text{ m} \times 1250 \text{ m}$ with a mesh grid of 1280×1280 cells. This configuration provides a fine horizontal resolution of approximately 1 m. The vertical grid configuration differs between the two simulation periods. For the April simulations, the domain reaches a vertical extent of 300 m; resolution starts at 0.375 m at the surface and stretches to 1.13 m at the base. The June simulations utilize a vertical extent of 500 m with a surface resolution of 0.376 m stretching to 2.02 m at the base. Both simulations use a stretching factor of $r = 1.0035$. In both cases, the vertical resolution is sufficient to resolve the rapid vertical decay of Stokes drift. The time steps are set to 0.9 s for the April simulations and 1.5 s for the June simulations. Further details are provided in section 3, Table 1 of L19. We use half-hourly profiles of horizontal and time-mean Eulerian flow and buoyancy with corresponding turbulent vertical fluxes, which include the correlations of vertical velocity w' with fluctuations, b' , u' , and v' , plus subgrid-scale (SGS) contributions (which are small except near the surface boundary). Profiles of Stokes drift are imposed as forcing. Variance of velocity fluctuations plus SGS give the TKE, whose dissipation ε is parameterized (i.e., throughout this manuscript, we refer to the dissipation of subgrid TKE as dissipation). The turbulent transport is determined as the residual term in the TKE budget, assuming steady-state conditions, by subtracting dissipation from the sum of the Eulerian shear production, Stokes shear production, and buoyancy production.

The Southern Ocean surface conditions based on SOFS in April and June are, on average, characterized by a buoyancy

flux negative out of the ocean, with equilibrium winds and waves ($La^{-2} = 11$). The forcing varies over 50 h in which the LESs are run, ranging between strong winds ($\sim 20 \text{ m s}^{-1}$) and quiescent conditions and similarly stronger convection ($\sim 200 \text{ W m}^{-2}$) to neutral buoyancy forcing, including one instance of stabilizing positive surface buoyancy forcing into the ocean (see Fig. A1). For a detailed breakdown of the LES into different regimes, see Table 2 of L19. We are interested in analyzing the turbulence that is relatively well developed and near-steady state, so we isolate those periods of the LES when the boundary layer is forced by relatively strong and consistent winds and near-wind-wave equilibrium and only modestly convective for the analysis. Of the 50 simulated hours in April, we exclude $t = 23\text{--}31 \text{ h}$ (when the winds are weak and rotating), and of the 41 simulated hours in June, we exclude the first 12 h and $t = 33\text{--}41 \text{ h}$ (when convection is dominant and winds are weak and rotating) (see Fig. A1 in appendix A—gray line coloring delineates the regimes that we exclude).

The LESs with Stokes forcing resolve coherent downward vertical velocities roughly aligned with the wind characteristic of Langmuir circulations—about 100 m wide, up to 800 m long, and separated by more than 300 m (see L19, their Fig. 4). Upwelling cells fill a broader area, characteristic of convective cells. Inertial oscillations are present in both Stokes and no Stokes cases, visible through the $\sim 15\text{-h}$ oscillations in the averaged mean velocities, characteristic of the inertial period at this latitude (see Fig. A1). Internal tides that are present in the real ocean are not simulated, and there are no mean lateral buoyancy gradients (fronts).

c. Microstructure shear observations and derivation of turbulence dissipation

Observations of turbulence dissipation are calculated from profiles of microstructure shear from three deployments of a Slocum glider during Southern Ocean Seasonal Cycle Experiment (SOSCEX)—Surface Turbulence and Ocean–Atmosphere Response to Momentum forcing (STORM) experiments (Swart et al. 2012; Nicholson et al. 2022). The experiments occurred in the SAZ (21 December 2017–17 January 2018; 42°S , 7°E), the APZ (19 December 2018–8 March 2019; 0° , 54°S), and the SIZ (17 December 2019–23 January 2020; 0° , 60°S) over consecutive years (2017–19) during austral summer, resulting in a total of 1220 profiles (see Fig. B1 in appendix B).

Microstructure data were collected using the self-contained turbulence instrument package MicroRider-1000LP (MR, Rockland Scientific International, Canada). The MR was attached to the top of a Slocum glider with turbulence sensors protruding at the front and was equipped with two airfoil velocity shear probes (SPM-38) oriented orthogonal to each other such that the first shear probe was sensitive to perturbations in the vertical, and the second shear probe was sensitive to perturbations in the horizontal direction relative to the direction of travel of the glider ($\partial w/\partial x$ and $\partial v/\partial x$, respectively). For processing the microstructure shear, additional sensors are required: a pressure transducer, a two-axis vibration sensor (a pair of piezo-accelerometers), and a high-accuracy dual-axis inclinometer (ADIS 16209, pitch and roll angles accurate to 0.18°). Shear microstructure measurements

were made to at least 500 m on both climbs and dives during the SAZ and the SIZ deployments, while during the APZ deployment, the MR was only powered on during climbs. The sampling frequency is 512 Hz on all turbulence channels (vibration, shear) and 64 Hz for the other channels (pitch, roll, and pressure).

The dissipation rate of TKE (dissipation hereinafter) is estimated by integrating the shear wavenumber spectrum, assuming isotropic turbulence:

$$\varepsilon_j = \frac{15}{2} \nu \left\langle \left(\frac{\partial u_j}{\partial x} \right)^2 \right\rangle \approx \frac{15}{2} \nu \int_{k_l}^{k_u} S'(k) dk, \quad (20)$$

where $\partial u_j/\partial x$ is the turbulent-scale shear component measured along the glider's along-path coordinate x and j identifies the shear probe number oriented orthogonal to measure the transverse ($\partial v/\partial x$) and vertical ($\partial w/\partial x$) components of the along-path shear. The kinematic viscosity of seawater, ν , is a function of the local water temperature. The shear wavenumber spectrum S' is integrated between k_l , set by the window length, and k_u , the minimum in a curve fit to the shear spectrum, that is unaffected by noise. The dissipation estimates are obtained following Fer et al. (2014) and the recommendations of the Scientific Committee on Oceanic Research (SCOR) Working Group on analyzing ocean turbulence observations to quantify mixing (ATOMIX; Lueck et al. 2024). Following the initial estimation of dissipation, a number of quality control steps were applied. Further details on the processing and quality control (QC) of the microstructure shear data are described in Giddy et al. (2025). Due to glider surface maneuvers, reliable estimates of dissipation rate start at 10 m during dives and below 3–5 m for climbs. Finally, the glider dataset is regridded to 1-m depth bins in the vertical. Each dive and climb is then treated as a single profile and allocated a time stamp that represents the average time during that dive.

d. Surface forcing

Similarity scaling of turbulence in the ocean surface boundary layer relies on knowledge of the surface forcing from which the structure of the boundary layer is approximated. Key turbulent fluxes are the momentum flux and the buoyancy flux. The third variable in the case of wave forcing is Stokes drift.

The LES are forced with observations from the SOFS mooring, and the time-varying wave effects (2D wave spectra from which Stokes drift is computed) are generated from the wave prediction model, WAVEWATCH III (Tolman 2009), initialized with ECMWF wave spectra.

To interpret the microstructure shear observations, we use the fifth generation ECMWF atmospheric reanalysis (ERA5) (Hersbach et al. 2020, 2023) for surface buoyancy fluxes and wind stress and the ECMWF Wave Model (WAM) global wave model (ECMWF 2013) for Stokes drift in lieu of collocated in situ measurements of winds, waves, and buoyancy during the full time periods of the glider deployments. We choose to use the ECMWF WAM model when interpreting the dissipation observations for consistency with ERA5 surface forcing. Even though two different wave models are used in this study (WAVEWATCH III in the LES and ECMWF

WAM together with the dissipation observations), differences in Stokes drift profiles between WAVEWATCH III and ECMWF WAM are expected to be small (e.g., Swain et al. 2019).

The atmospheric component of ERA5 has a horizontal resolution of approximately $0.25^\circ \times 0.25^\circ$. The two-dimensional wave variance spectrum that describes how the mean sea surface elevation variance due to ocean waves is distributed as a function of frequency and propagation direction is given hourly at a resolution of approximately $0.5^\circ \times 0.5^\circ$.

Stokes drift was shown by Kenyon (1969), in the case of linear waves, to relate to the wave variance density spectrum $E(\omega, \theta)$. The circular frequency, $\omega = 2\pi f$, can be transformed to f , the linear frequency using a simple coordinate transform giving a description of Stokes drift as a function of linear wave frequency and direction (Breivik et al. 2014):

$$\mathbf{U}_s(z) = \frac{16\pi^3}{g} \int_0^{2\pi} \int_0^\infty f^3 \hat{\mathbf{k}} e^{2kz} F(f, \theta) df d\theta, \quad (21)$$

where θ is the measured clockwise from north and $\hat{\mathbf{k}}(f, \theta) \equiv \mathbf{k}/|\mathbf{k}|$ is the normalized direction vector of the wave component. The Stokes drift profile was computed from the full 2D spectrum following Eq. (21). A diagnostic f^{-5} tail was added to the spectrum beyond the cutoff frequency. This has a significant impact on the surface Stokes drift [up to 30%, see Breivik et al. (2016)]. The effect of high-frequency waves vanishes quickly with depth, creating a strong gradient in the Stokes drift profile in the upper part of the boundary layer.

In the case of the LES, wind stress magnitude, τ , is derived from wind speeds observed at the SOFS mooring using the COARE 3.0 bulk formulation (Fairall et al. 2003). For the wind stress applied to the observed dissipation, we use ECMWF, which is calculated by the ECMWF Integrated Forecasting System's turbulent diffusion and turbulent orographic form drag schemes (European Centre For Medium-Range Weather Forecasts 2023).

The net surface buoyancy flux into the ocean is given by

$$B_o = g \left[\frac{\alpha Q_{\text{net}}}{\rho_0 c_p} - \beta S(E - P) \right], \quad (22)$$

where $g = 9.81 \text{ m s}^{-2}$ is the gravitational acceleration; α is the thermal expansion coefficient; β is the haline contraction coefficient both computed from surface Absolute Salinity (S) and Conservative Temperature measured by the Slocum glider, using the Gibbs SeaWater Thermodynamic Equation of Seawater-2010 (TEOS-10) Toolbox (McDougall and Barker 2011); Q_{net} is the net heat flux into the ocean (shortwave + longwave + latent + sensible); c_p is the heat capacity of water = $3991.9 \text{ J K}^{-1} \text{ kg}^{-1}$; $\rho_0 = 1027 \text{ kg m}^{-3}$ is a reference seawater density; and E and P are the evaporation and precipitation rates, respectively.

As wind and wave variables are computed from a model, we acknowledge that the surface forcing as derived here from ERA5 may not be fully consistent with the observations. However, comparison with the collocated in situ wind observations from a waveglider from part of the deployment shows agreement with wind speed variability from ERA5, with

ERA5 tending to underestimate the mean by 10%–20%, although we note that the energy containing wind forcing is well resolved (see Fig. B2).

4. Similarity functions of turbulence kinetic energy in no Stokes and Stokes LES

A key term in the TKE equation [Eq. (1)] is the shear production term. Momentum at the ocean surface is transported to depth by turbulent eddies generated via shear production. This is the process that is parameterized in OGCMs as a turbulent viscosity term that is based on an understanding of shear production. Furthermore, turbulence generation by shear production, under neutral buoyancy forcing, is the primary source of energy that is ultimately dissipated and thus relates directly to observations of turbulence dissipation rate. In the following, we describe an example of shear production under wind-only and wind-wave-forced conditions in the realistically forced LES. We then statistically assess the shear production and dissipation of the LES runs compared to scaling parameterizations with the intention to understand the link between shear production and dissipation. Our interest is primarily in the surface layer (above $0.1h$), but we analyze depths down to $0.25h$.

a. Example of the vertical structure of shear production under wave forcing from an LES

Under neutral stratification and no buoyancy forcing, assuming constant momentum flux, and a rigid lid, the law of the wall scales shear production as a function of friction velocity and distance from the boundary [Eq. (6)]. Figure 2 decomposes the terms in the law of the wall scaling [Eq. (6)] and compares them to LES solutions of the Craik–Leibovich wave-averaged equations [Eq. (9)] for TKE under pure wind forcing and wind-wave forcing. First, consider pure wind forcing. Shear production (Fig. 2d) is the dot product of a vertical turbulent momentum flux (Fig. 2a) and a vertical shear (Fig. 2b). Friction velocity squared (u_*^2) is equivalent to the LES vertical turbulent momentum flux ($-\langle \mathbf{u}' w' \rangle$) at the ocean surface. In the LES, the vertical turbulent momentum flux generally decays to zero at the base of the boundary layer (as seen in the snapshot in Fig. 2a), while in the law of the wall, the parameterized momentum flux is constant with depth—a good approximation between the surface and $0.1h$. Vertical shear is represented as $u_*/(\kappa|z|)$ in law of the wall. Under pure wind forcing, the law of the wall accurately reproduces the LES shear down to $z \sim 0.8h$. Under pure wind forcing, Θ_E , shown in Fig. 2c, is close to 1 throughout the surface layer, indicating stress aligned with shear. The law of the wall and LES shear production are compared in Fig. 2d. Near the surface, the law of the wall is representative of shear production in the LES. The quality of the law of the wall fit to LES decays with depth. This is partly due to a failure in the assumption of constant momentum flux away from the boundary as seen in Fig. 2a. Using a linear decay in momentum flux $u_*^2[1 - (|z|/h)]$ [Eq. (7)] in place of a constant momentum flux, u_*^2 , significantly improves the representation of shear production below $0.1h$ for the pure wind forcing case. At the LES simulation latitude of 47°S , planetary rotation may significantly affect the turbulent stress

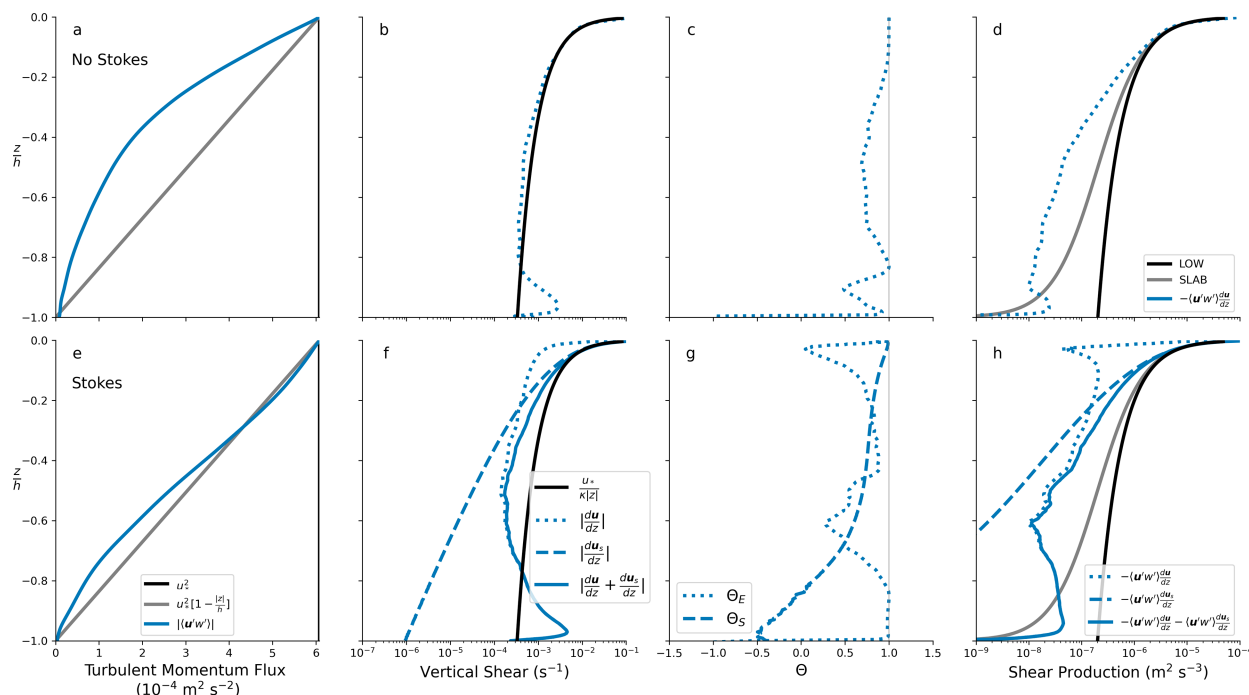


FIG. 2. Example of shear production in the surface boundary layer for (a)–(d) no Stokes and (e)–(h) Stokes forcing. The black lines represent terms in the law of the wall scaling. The gray lines replace the vertical turbulent momentum flux with a slab model. The blue lines are snapshots of data from the April LES runs of L19 at $t = 10$ h. (a),(e) Magnitude of the vertical turbulent momentum flux; (b),(f) magnitude of Eulerian vertical shear du/dz (blue dotted lines), and Stokes vertical shear u_s/dz (blue dashed lines), total shear (blue solid line), and wall scaling (black solid line); (c),(g) cosine of the angle between the stress and shear vectors; and (d),(h) shear production. In the no Stokes case, the boundary layer depth is 179 m. In the Stokes case, the boundary layer depth is 181 m.

within the boundary layer (e.g., Zikanov et al. 2003; Liu et al. 2018). Turbulent momentum stress is constrained by the Ekman spiral which results in a counterclockwise veering of momentum with depth in the Southern Hemisphere (not shown) and suppression of the stress. The stress profiles shown in the example (Figs. 2a and 2e) are qualitatively consistent with rotational effects in wind-only and wind-wave-affected boundary layers. The boundary layer depth is 179 and 181 m for no Stokes and Stokes cases, with an Ekman depth of 206 m (giving a scaled Ekman depth of -1.1). Under Stokes forcing, the effect of Coriolis on the stress profile tends to be reduced due to the increased mixing by Langmuir turbulence (Fig. 2e).

Now consider the Stokes forcing case (Figs. 2e–g). Turbulent momentum flux is well represented by the slab parameterization for momentum flux (similar to the pure wind case). Vertical shear, however, is no longer well represented by $u_*/(\kappa|z|)$. Vertical shear with Stokes forcing comprises Eulerian shear and Stokes shear with the majority of the shear near the surface attributed to Stokes shear—which arises from different (wave) physics than the $1/z$ form of the wind-driven Eulerian shear arising from law of the wall physics. A substantial angle difference is present between the stress and shear (Fig. 2g). Nevertheless, near the surface, the total shear production parameterized by law of the wall represents the shear production solved by the LES (Figs. 2d,h). This example shows that law of the wall provides a first approximation of shear production in the surface layer of the surface boundary layer in both pure wind and

Stokes forcing conditions. In other words, under realistic forcing conditions (such as the ones represented here), surface-layer shear production under wind forcing is not necessarily different in magnitude from surface-layer shear production under wind and wave forcing. In the next two sections, we show that this anecdotal example in Fig. 2 is in fact representative of the LES solutions across a broad range of forcing regimes.

Note that in both no Stokes and Stokes cases, shear production deviates from the slab parameterization below the surface layer. These deviations could be resulting from inadequately accounting for depth dependence such as with the structure functions of Large et al. (1994, 2019b), which are not accounted for here, as well as other processes that modify vertical shear and become relatively more important deeper in the boundary layer—such as shear related to inertial oscillations.

b. Shear production and dissipation dependence with depth

We evaluate the relationship between shear production and dissipation given by the realistically forced LES compared with the three shear production parameterizations described in section 2 [the classic wall scaling, Eq. (6); the slab wall scaling, Eq. (8); and L19, Eq. (13)] with no Stokes and Stokes cases in April and June (Figs. 3a,b). Recall that the wall scaling and the slab wall scaling applied here assume a local balance between shear production and dissipation [Eq. (14)]; however, the TKE budget under steady state also includes a turbulence

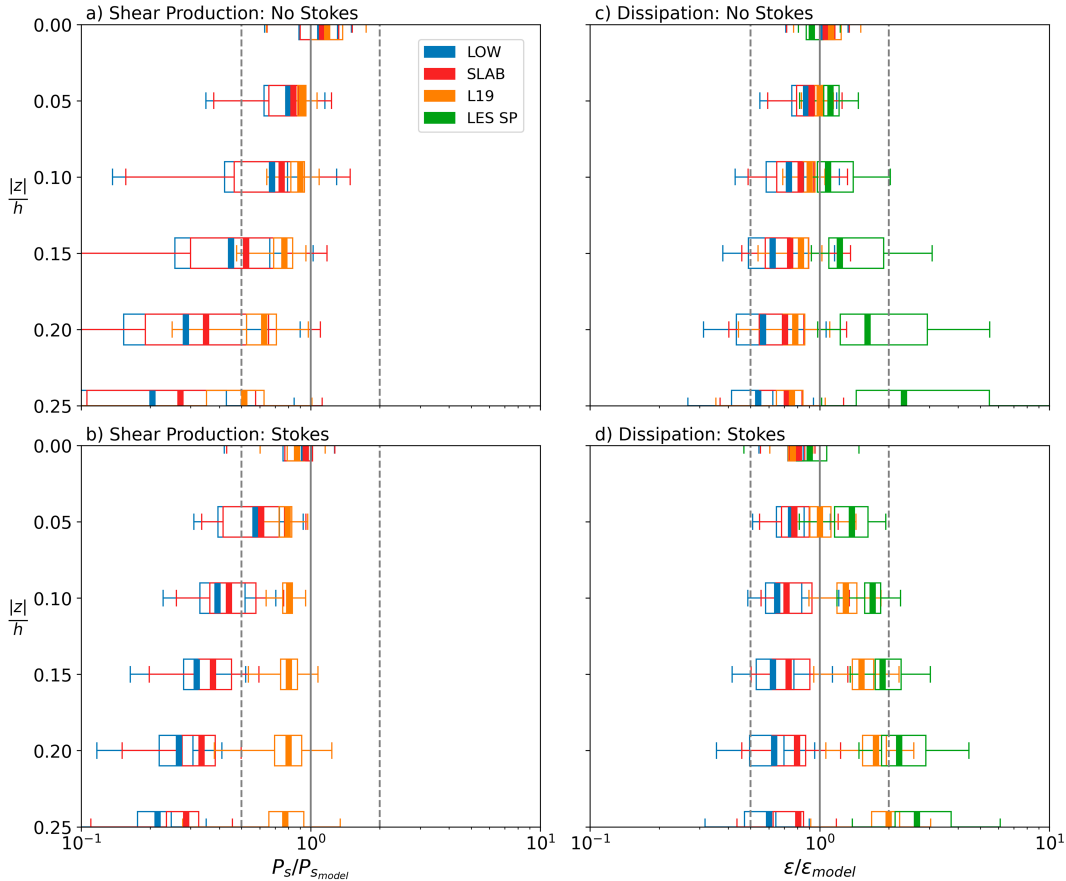


FIG. 3. Box-and-whisker plots of shear production from L19 (a) no Stokes and (b) Stokes LES runs, scaled by the P_S model and ϵ_{model} , where “model” refers to parameterization after law of the wall (blue) [Eq. (6)], slab wall scaling (red) [Eq. (8)], and L19 shear production parameterization (orange) [Eq. (13)]. (c) No Stokes and (d) Stokes turbulence dissipation scaled by the same parameterizations for shear production and, additionally, in green, LES dissipation scaled by LES shear production. Data are binned by nondimensional depth ($|z|/h$, where h is the depth of the boundary layer). The solid gray line identifies a 1:1 agreement, and the dashed gray lines show factor of 2 differences. Note the x axis is in log scale.

transport term and a buoyancy flux. Under wind-driven shear conditions, transport is assumed negligible, but the same assumption cannot be made under conditions where vertical velocities and thus turbulent transport become important (as in cases of strong convection or Langmuir circulations; Wyngaard and Côté 1971; Polton and Belcher 2007). We illustrate the role of turbulence transport in Fig. 3. The left panels in Fig. 3 show the ratio of LES shear production (LES SP) to parameterized shear production with depth. The right panels show the ratio of LES dissipation (LES ϵ) to parameterized dissipation with depth. A ratio of one indicates that the parameterization agrees with the simulation. L19 best represent the shear production in both no Stokes (Fig. 3a) and Stokes (Fig. 3b) forcing cases as it incorporates the effects of waves and buoyancy on Eulerian shear as well as a depth decaying turbulent momentum flux. In the pure wind-driven framework, dissipation is assumed to locally balance shear production. We test this assumption by comparing the same three shear production parameterizations with dissipation. In the no Stokes case, of all three shear production

parameterizations, L19 most closely match the LES (i.e., is nearest to 1:1) down to $\sim 0.15h$ (Fig. 3b). In the Stokes case, the L19 shear production parameterization (Fig. 3b, orange) increasingly underestimates the turbulence dissipation with depth, while the law of the wall and slab wall scaling better represent the dissipation rate (Fig. 3d).

In addition to turbulence transport under wave forcing, alignment of the stress and Stokes drift impacts shear production (e.g., Fig. 2g). In realistic ocean conditions, winds and waves are often misaligned, reducing the component of momentum flux aligned with Stokes shear and thus lowering shear production relative to idealized aligned cases (Van Roekel et al. 2012). The L19 parameterization accounts for this effect by explicitly incorporating wind-wave misalignment through an integrated Stokes parameter, further improving the agreement between the parameterized shear production with LES shear production (Fig. 3b).

Overall, the L19 shear production parameterization accurately models the LES shear production (Fig. 3b), so the

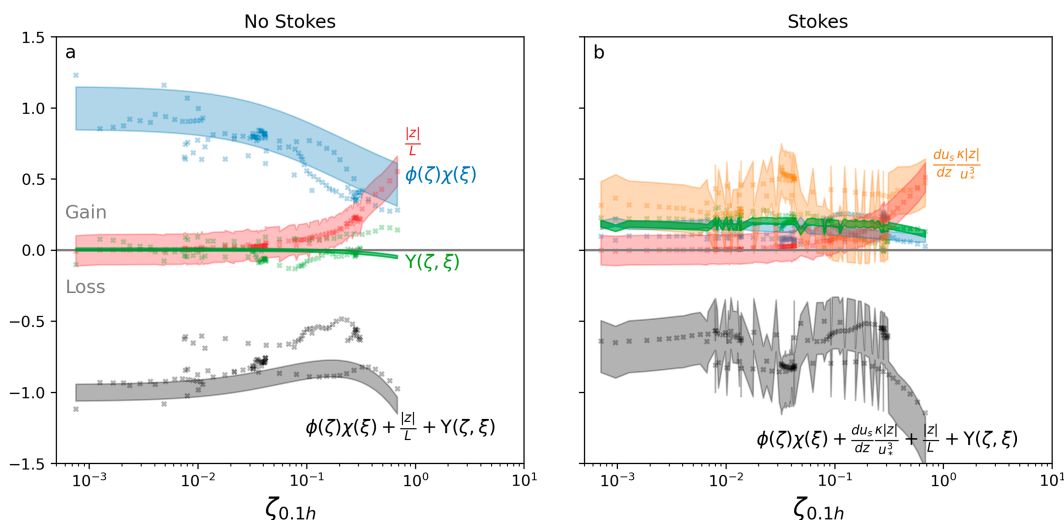


FIG. 4. Nondimensional TKE budget at $0.1h$ during the same time period as in Fig. 3. (a) No Stokes and (b) Stokes forcing. Positive values represent turbulence production, and negative values represent turbulence loss. Markers show data points from the LES. Shading shows the range of predictions from similarity functions (also annotated). Red: buoyancy production, blue: Eulerian shear production, orange: Stokes shear production, green: turbulence transport, and black: turbulence dissipation rate.

underestimation of dissipation by the L19 parameterization in Fig. 3d is not due to a failure to capture the shear production. Conversely, the wall and slab shear production parameterizations significantly overestimate the LES shear production even at 0.05 and $0.1h$ in Fig. 3b, so their relatively accurate estimate of dissipation is not due to their accurate modeling of shear production under Stokes forcing.

This realistically forced LES corroborates previous studies showing that shear production does not locally balance dissipation under Stokes forcing (Fig. 3d), highlighting the role of turbulent transport, even in the surface layer.

c. Dependence of transport on the stability parameter

Since dissipation, a measurement of the rate of conversion of turbulence to heat, is the primary measure of turbulent mixing in the ocean, it is useful to understand the relationship between dissipation and the TKE budget. Wyngaard and Coté (1971) find an imbalance between measured terms of the TKE budget under very unstable conditions (strong convective cells) in the atmosphere that they attribute to turbulent transport. Similarly here, but instead for the case of Langmuir cells, we assess the LES terms of the TKE budget. We use this to build a quantitative model of the transport of TKE by considering the nondimensional terms of the TKE budget and their associated similarity functions as a function of the stability parameter, ζ , at $0.1h$ for no Stokes and Stokes cases (Fig. 4). Under neutral conditions ($|\zeta| \leq 10^{-1}$) in the no Stokes case (Fig. 4a), shear production balances dissipation, while buoyancy production and turbulence transport are negligible. For the same conditions under Stokes forcing (Fig. 4b), Eulerian shear production is reduced to $\sim 20\%$ of its no Stokes value, with the largest contribution to shear production derived from Stokes shear (orange in Fig. 4b). The transport term is no longer negligible,

with an almost equivalent contribution to the balance of turbulence dissipation as that of Eulerian shear production.

Thus, under realistic neutral wind-wave-forced conditions, dissipation is balanced by shear production and transport following the form of Eq. (15) (Fig. 4b), similarly to that observed in the atmosphere under strong convection (Wyngaard and Coté 1971) and as discussed in a number of studies focused on Langmuir circulation (Polton and Belcher 2007; Gerbi et al. 2009; Grant and Belcher 2009; Jarosz et al. 2021).

L19 and Large et al. (2019b, 2021b) suggest that in the Southern Ocean under Stokes forcing, a similarity theory approach can be reasonably applied to model turbulent velocity scales—essentially treating Stokes drift as a surface forcing. We extend this reasoning to turbulent transport by fitting a simple multiple linear regression to nondimensional LES TKE transport [scaled by $u_*^3/(k|z|)$] to derive the empirical function, $Y_m(\xi, \zeta)$, that is a function of the stability parameter and the Stokes parameter, shown in Fig. 5:

$$T_m(\xi, \zeta) \approx Y_m(\xi, \zeta) = 0.00475 + 0.25\xi - 0.08\zeta. \quad (23)$$

This gives an empirically determined universal quantitative functional representation of dimensionless transport. The function defined by Eq. (23) explains 65% of the variance in transport at the base of the surface layer in the LES which spans a wide parameter space (see Fig. A1). Nevertheless, further LES studies are needed in different parameter spaces (e.g., stable/strongly convective) to more extensively test Eq. (23) and refine our understanding.

Finally, the nondimensional law of the wall scaling for dissipation is extended to unstable, wavy conditions:

$$\varepsilon \frac{\kappa|z|}{u_*^3} = \phi_m(\xi)\chi_m(\xi) + \frac{dU_s}{dz} \frac{\kappa|z|}{u_*^3} + \frac{z}{L} + Y_m(\xi, \zeta). \quad (24)$$

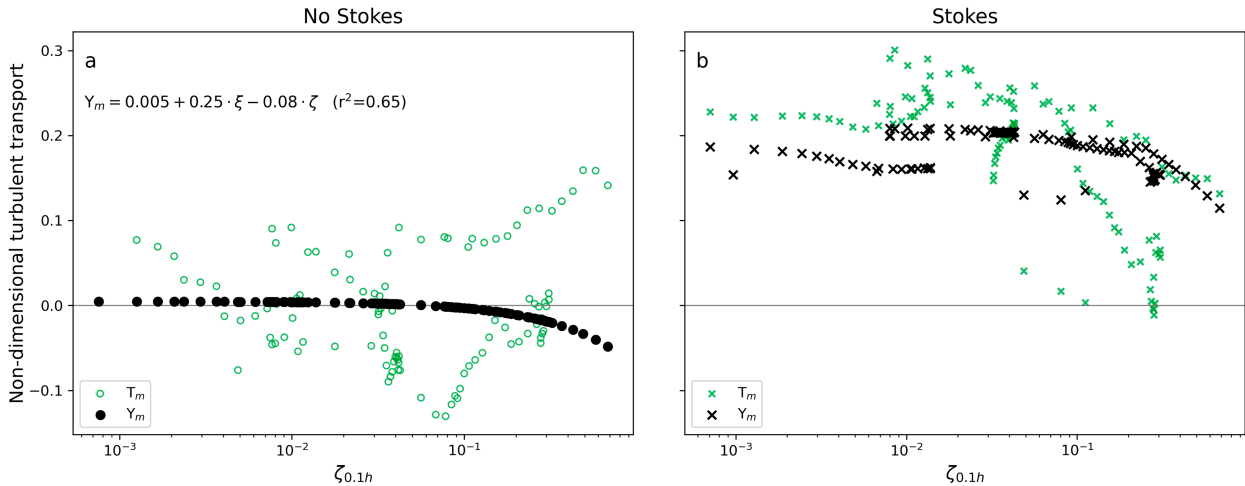


FIG. 5. Nondimensional similarity functional fit, $Y_m(\xi, \zeta)$ [Eq. (23)], of the convergence of turbulence transport of TKE in terms of the Stokes parameter ξ and the stability parameter ζ , plotted along ζ at $0.1h$, derived from LES solutions during the same time periods as in Fig. 3. (a) No Stokes and (b) Stokes. The green markers indicate the dimensionless convergence of TKE due to turbulent transport T_m , and the black markers show Y_m . The fit from the combination of Stokes and no Stokes data has an r^2 of 0.65.

With Eq. (24), we have now built on L19 to include turbulent transport giving a scaling for dissipation that is representative of surface boundary layer with wave contributions under weak or neutral buoyancy forcing. This provides a physically grounded interpretation of dissipation observations in regimes where Langmuir turbulence is significant.

5. Scaling observations of dissipation in the Southern Ocean

The four (wall scaling, slab wall scaling, L19, and L19 + transport) semiempirical similarity functions for dissipation in the surface layer are now compared to observations of dissipation in the SAZ, APZ, and SIZ during destabilizing buoyancy conditions. Both the observed and similarity-derived profiles are vertically averaged from 5 to 15 m. This depth average ranges between 0.1 and $0.5h$ depending on the depth of h (Fig. B1). Averaging from 5 m is a necessity due to limitations in the data as noted in section 3b (limited data in upper 5–10 m). A lower bound of 15 m is chosen as a compromise, providing sufficient data and reasonable results. The 5–15-m depth range is also easily observable and thus a useful metric for microstructure observations in the ocean. Following the vertical averaging, a profilewise rolling mean of four profiles (and about 16 h) is applied, thereby averaging over the inertial periods in these regions.

First, consider predictions of observed dissipation, assuming wind-driven shear production only. Law of the wall [Eq. (6)] explains 75%–85% of the variance in all regions, with a tendency to overestimate dissipation rates (Figs. 6a–c). This overestimation is most exaggerated in the SAZ and during weaker forcing (lower values of dissipation rate). During the observational campaign, the SAZ experienced the shallowest boundary layers, such that an average of 5–15 m is well below $0.1h$, invalidating the assumption of constant momentum flux. Adjusting for a linearly decaying momentum flux [Eq. (8)] improves the variance

explained and removes the bias averaged across all regions (Figs. 6d–f).

Now, consider predictions of observed dissipation accounting for wind- and wave-driven shear production. L19 [Eq. (13)] explains ~80% of the variance across all regions, where shear production is now largely attributed to Stokes drift as opposed to friction velocity, but the parameterization tends to underestimate observed dissipation rates (Figs. 6g–i). Adjusting L19 for the redistribution of turbulence via the transport of TKE [Eq. (24)] reduces this bias by ~20% (Figs. 6j–l).

An outstanding observation is that the law of the wall slab scaling and the modified L19 + transport slab scaling can explain the variability and magnitude of observed dissipation rates equally well even though they represent different physics. The wall scaling (and the wall slab scaling) represents a shear resulting from wind stress, while the modified L19 + transport slab scaling represents a shear from Stokes drift, a shear from wind stress, a buoyancy flux (and the interactions of Stokes drift and buoyancy on the Eulerian shear), and turbulent transport term. With Eq. (24), we have, therefore, provided a new quantitative parameterization of the TKE budget that includes all the terms of the TKE budget and reproduces both realistically forced LES over a broad parameter space (Fig. 4) and observations in the Southern Ocean (Fig. 6).

6. Discussion and conclusions

Through the interrogation of realistically forced LES across a wide parameter space of wind and wave forcing, we have developed a similarity scaling for the dissipation of TKE [Eq. (24)] that is a more physically consistent representation of dissipation under the wave-forced conditions that dominate in the Southern Ocean. Specifically this scaling, based on similarity theory, builds on the scaling of L19 to include all four terms of the TKE budget: Eulerian shear production, Stokes shear production, Buoyancy production, and turbulent transport.

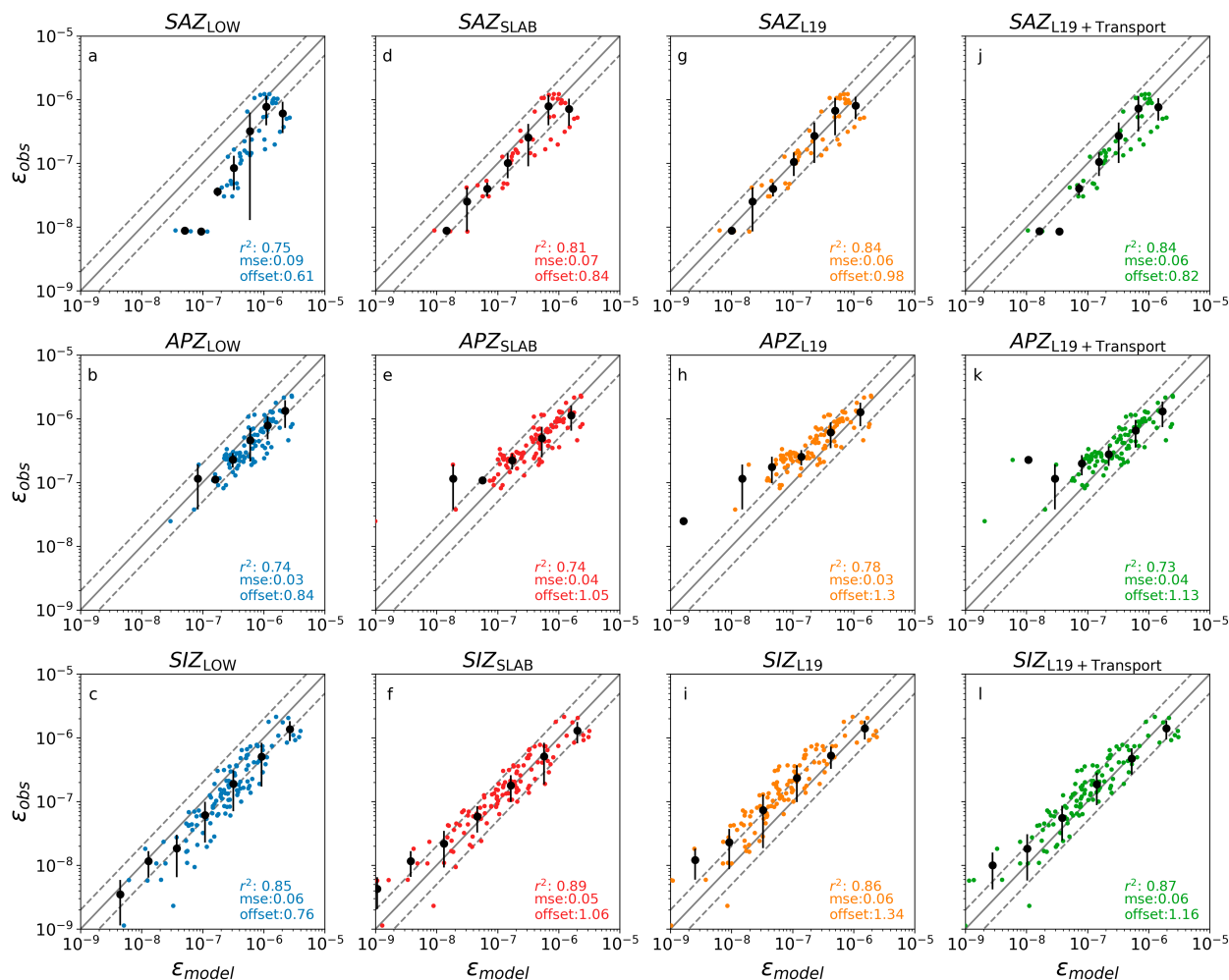


FIG. 6. Models for turbulence dissipation compared to observations of turbulence dissipation in three Southern Ocean locations during destabilizing buoyancy conditions. (a),(d),(g),(j) The SAZ, (b),(e),(h),(k) the APZ, and (c),(f),(i),(l) the SIZ. (a)–(c) Law of the wall [Eq. (6)], (d)–(f) slab [Eq. (8)], (g)–(i) L19 [Eq. (13)], and (j)–(l) L19 plus transport [Eq. (24)], with transport defined in this manuscript [Eq. (23)]. The solid gray line indicates a 1:1 agreement. The dashed gray lines show a factor of 2 offset from the 1:1 line. Black markers indicate binned averages, with the error bars showing the interquartile range. The r^2 value, mean-square error, and bias from the mean of the regressions are indicated in the legends of each panel.

We find that turbulent transport can be represented in a non-dimensional functional form [Eq. (23)] with a dependence on buoyancy flux and Stokes drift and incorporated within a similarity scaling framework [Eq. (24)]. This scaling is representative to both realistically forced LES, where the full TKE budget is explicitly simulated (Large et al. 2021a), and a suite of observations in the Southern Ocean (Giddy et al. 2025), where dissipation rates are measured directly. We expect that it will be applicable to cases at least within the parameter space tested here (weak buoyancy forcing, wind-wave forcing mostly in equilibrium). Further work is needed to assess if the parameterizations used and developed here are applicable in other parts of parameter space.

Idealized LES studies that incorporate the surface wave effects encapsulated the wave-averaged Craik–Leibovich (CL) equations and compare dissipation profiles with wall scaling

have shown that dissipation is enhanced relative to wall-scaling predictions (e.g., McWilliams et al. 1997; Grant and Belcher 2009). Realistically forced LESs that solve for the CL vortex force (include wave effects) more closely match dissipation observations (e.g., Skillingstad et al. 2023; Pham et al. 2023), but their results have not been directly compared to wall scaling—a common metric used to analyze observations of dissipation. Exceptions (reduced dissipation to that expected) have been noted that arise when winds and waves are misaligned (e.g., Van Roekel et al. 2012) and when swell is important (e.g., Fan et al. 2020). In this study, we compare realistically forced LES with and without Stokes forcing and observed dissipation profiles with law of the wall. Law of the wall is found to be an accurate predictor of surface-layer dissipation in the Southern Ocean (Figs. 3c,d and 6a–c). This is consistent with a number of prior observational studies (e.g., Gerbi et al. 2009;

Sutherland et al. 2013; Esters et al. 2018; Ferris et al. 2022; Miller et al. 2023; Zeiden et al. 2024). The slab wall scaling improves the fit as many of our observations are taken below the surface layer. It is intriguing that the physics-aligned scaling we present [Eq. (24)] does not perform statistically better against our observations than the slab wall scaling that does not account for wave physics. In part, the similarity across parameterizations may be reflecting the approximately linear relation between Stokes drift and friction velocity (e.g., Esters et al. 2018; Gargett 2022). This result may appear to suggest that surface gravity waves and Stokes drift are not participating in the TKE budget and specifically the dissipation rate in the surface layer. Zippel et al. (2022) follow this hypothesis with data from the North Atlantic, splitting the boundary layer into a breaking wave layer (upper few meters) in which turbulence is enhanced from breaking waves and the layer below down to the base of the boundary layer, which is explained by the slab wall scaling. However, since our scaling, which includes transport and Stokes shear production, performs equally well as the slab scaling, we cannot corroborate their results with the same physical argument. While in the regimes sampled in this study and, for example, that of Zippel et al. (2022), the wall scaling well represents dissipation, we have shown here with realistically forced LES that the physics behind the wall scaling is not representative of the actual physics at play. As such, we cannot be certain that the law of the wall will hold in all regime spaces. And indeed, deviations from the wall scaling compared to dissipation measurements when waves are present are often observed, supporting our interpretation (D'Asaro et al. 2014; Sutherland et al. 2016; Jarosz et al. 2021; Zheng et al. 2021).

In all parameterizations and across all the regions observed, there is a remaining variance of $\sim 20\%$. Part of this variability may reflect the natural intermittency of turbulence (e.g., Moun et al. 1995). Remaining variance in dissipation may be attributed to limitations in using reanalysis wind and modeled waves, inadequate representation of stress-shear misalignment, unaccounted for sources of shear production, or imperfect closure of the TKE budget. In addition to Langmuir turbulence, other boundary layer processes can modify both the Langmuir turbulence directly as well as acting as additional sources or sinks of turbulence dissipation. The transport of TKE is nonnegligible for the local TKE balance under cases of strong convection as shown in the atmosphere (e.g., Wyngaard and Coté 1971). Time-varying wind (e.g., Watkins and Whitt 2020) and wave (e.g., Wagner et al. 2021) forcing can excite the growth of near-inertial waves which act as a source of shear, and the rate of growth of the waves modulates Langmuir turbulence (Wagner et al. 2021). In the simulated dissipation of the LES presented here, near-inertial resonance likely contributes to enhanced shear at the base of the boundary layer in both wind- and wind-wave-forced cases, illustrated in Figs. 2b and 2f. The effects of differences in the rate of growth of waves have not been explored. The presence of fine-scale lateral fronts will also modify turbulence. Submesoscale mixed layer eddies are associated with overturning cells with strong vertical velocities (Fox-Kemper et al. 2008). Winds aligned with fronts provide energy to maintain the geostrophic shear and, thus, turbulence, associated with those fronts (Thomas et al. 2016). In this study, lateral processes are

not resolved in the simulation, and while they may be present in the observations, they do not appear to play a dominant role—at least in the upper half of the boundary layer since the wind forcing is much greater than the strength of the lateral gradients [$O(10^{-8})$]. The small contribution of submesoscale instabilities to the TKE budget was shown in observations by Buckingham et al. (2019); however, they may become important in other regimes (Dong et al. 2024). Since the parameterizations tested here perform similarly across the three regions of our observations (Fig. 6), we speculate that turbulence related to submesoscale fronts (D'Asaro et al. 2011; Thomas et al. 2016; Yu et al. 2019; Dong et al. 2024) and the effects of time-varying processes like inertial oscillations which would be expected to vary with latitude (Large and Crawford 1995; Dohan and Davis 2011; Watkins and Whitt 2020) are secondary sources of shear production in the surface layer and likely only play a substantial role in the outer boundary layer—at deeper depths than the focus of this analysis but important to consider if closing the full boundary layer budget with this framework. A focus on these processes, especially their role in the TKE budget in the outer boundary layer in the Southern Ocean, is left to future work.

A weakness of the observational data is that we largely miss the upper 5 m, where Stokes shear production is expected to be its strongest as Stokes drift follows an exponential decay. Because of the limited near-surface observations, we also did not consider the role of breaking waves (e.g., Agrawal et al. 1992; Zippel et al. 2022), although we do not expect the results here to be sensitive to breaking waves because their impact is generally confined to one significant wave height (approximately 8 m at its maximum in the conditions of these observations). In addition, lacking collocated observations of forcing variables, we relied on model and reanalysis products in this study, which could add further uncertainty. Nevertheless, the results are consistent with the proposed physics based on realistically forced LES of the Southern Ocean, giving confidence to the conclusions drawn here. The conclusions of this analysis apply to the upper-ocean surface boundary layer, under moderate-to-strong wind and wave forcing, weak stratification, and weak destabilizing buoyancy forcing. Cases of strongly stabilizing buoyancy, weak and variable winds, rapidly changing winds, and strong convection were not considered. LES over a broader range of conditions representative of the real ocean together with observations would enable further generalization of these conclusions. Future observational studies could reduce uncertainty through collocated observations of surface forcing, direct observation of friction velocity (e.g., with a sonic anemometer), and measurements of Stokes and Eulerian shear in conjunction with turbulence dissipation.

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Data availability statement. Large-eddy simulations were made available by Large et al. (2021a) at <https://doi.org/10.5065/87n8-9r86>. The in situ data glider and microstructure data are

available via Zenodo (Giddy et al. 2025) at <https://doi.org/10.5281/zenodo.14981745>. Fifth generation ECMWF atmospheric reanalysis (ERA5) (Hersbach et al. 2023) was downloaded from the Copernicus Climate Change Service. The Antarctic Circumpolar Current (ACC)—fronts product used in Fig. 1, <https://doi.org/10.24400/527896/a01-2023.004>—was produced by CTOH and distributed by AVISO+ (<https://www.aviso.altimetry.fr/>) with support from CNES.

APPENDIX A

LES Forcing and Surface Current Response

Figure A1 describes the surface forcing imposed on the LES for April and June cases. Note the predominantly convective conditions in Fig. A1c. Figure A1d illustrates the Eulerian surface currents under the no Stokes cases, while Fig. A1e illustrates the Eulerian surface currents under the Stokes cases. Oscillations spanning 15 h (about the inertial period at this latitude, 42°S) indicate the presence of inertial oscillations in all cases. These are most prominent in April (solid lines). Under the Stokes forcing cases, the mean Eulerian current is reduced in magnitude, but the oscillations remain.

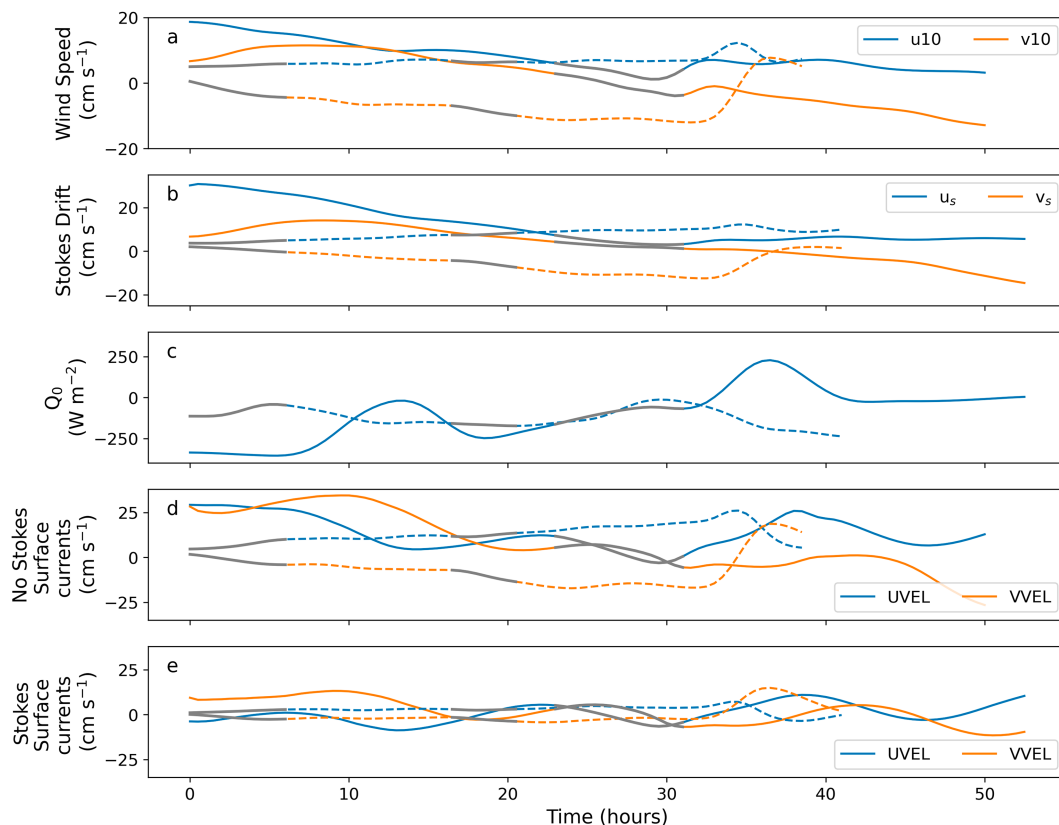


FIG. A1. Surface forcing in April (solid lines) and June (dashed lines) LES: (a) wind speed, (b) Stokes (c) surface heat flux, Q_0 , and the Eulerian surface current response: (d) no Stokes surface currents and (e) Stokes surface currents. Gray lines mark the regimes which are not included in this analysis.

APPENDIX B

Observations in the Southern Ocean

a. Dissipation

Figure B1 details the dissipation sections from the subantarctic zone, the Antarctic polar frontal zone, and the seasonal ice zone, computed from microstructure shear profiles as described in section 3c, together with collocated surface forcing from the ERA5 reanalysis.

b. In situ observations of wind speed compared to ERA5 reanalysis

During two of the Slocum glider deployments, part of the deployments were coupled to surface wavegliders which were

recording wind speed from an Airmar sensor mounted at a height of 1 m. Because the deployment was not coupled for the full missions, we use ERA5 reanalysis in this study; however, here we show that using ERA5 reasonably captures the variance and magnitude of winds (which impose stress on the surface ocean and drive waves and mixing). The regressions of ERA5 with the in situ wind observations in Figs. B2b and B2c show that 80% of the variance observed is captured with ERA5, with ERA5 being biased low, especially at stronger wind speeds. Spectra of wind speeds from ERA5 and the Airmar observations are computed using a multitaper spectral method with a smoothing of $P = 16$, the time-bandwidth product. The resulting power density spectra (Figs. B2c and B2f) indicate that the energy containing scales (lower frequencies) of the winds are well represented in ERA5.

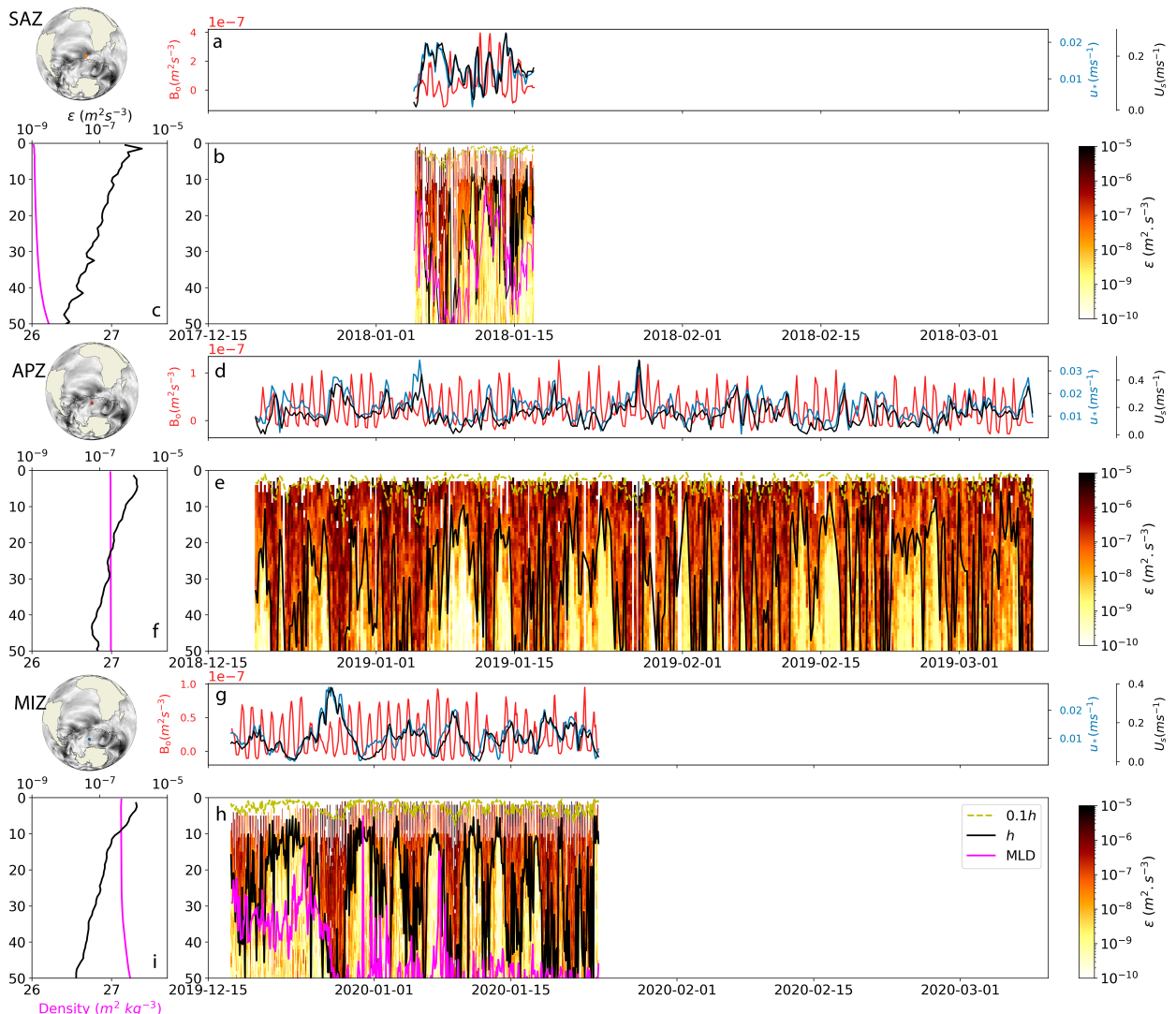


FIG. B1. Observed turbulence dissipation rate and their atmospheric forcing. The SAZ, 2017–18: (a) forcing, (b) dissipation sections, and (c) mean profiles of density (magenta) and dissipation (black); the APZ, 2018–19: (d) forcing, (e) dissipation section, and (f) mean profiles of density (magenta) and dissipation (black). The SAZ, 2019–20: (g) forcing, (h) dissipation section, and (i) mean profiles of density (magenta) and dissipation (black). The $0.1h$ (yellow dashed), h (solid black), and the MLD (solid magenta) are overlain in (b), (e), and (h). Each deployment occurred in a different year, but time series are aligned from December–March. The maps identify the locations of the deployments.

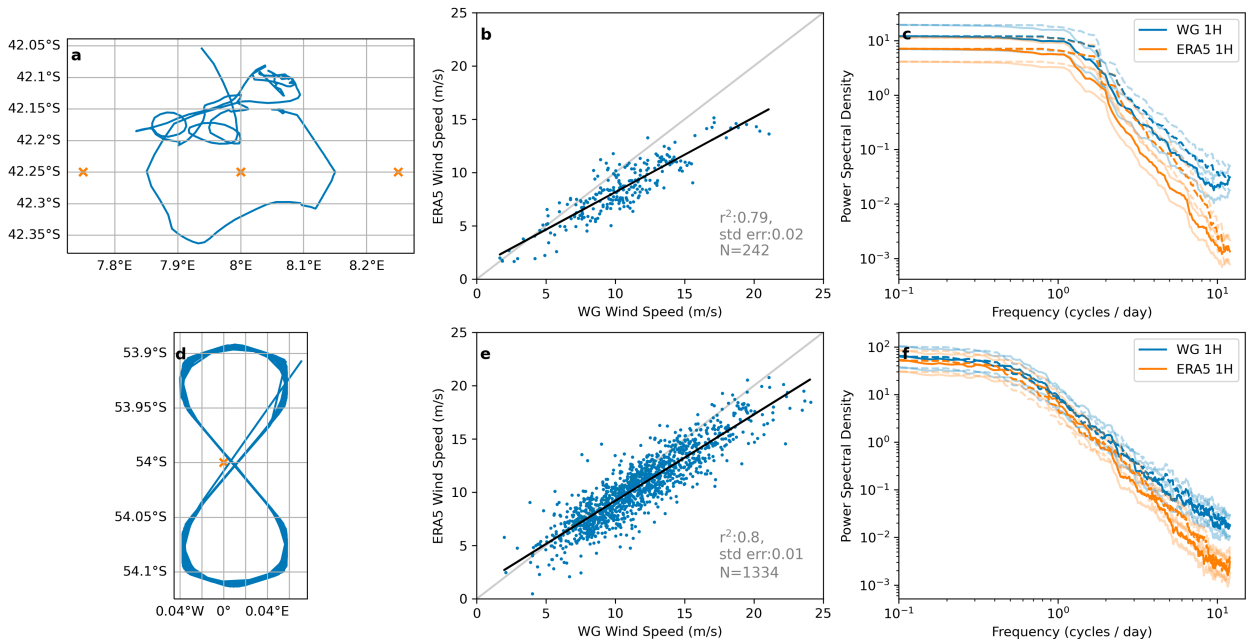


FIG. B2. Comparison of in situ wind observations from times when a waveglider was coupled with the Slocum during the deployments in (a)–(c) the SAZ and (d)–(f) the APZ. Waveglider observations are interpolated to hourly from 10-min samples. (a), (d) Trajectories of the wavegliders in blue and ERA5 in orange; (b), (e) regressions of waveglider observed wind speeds at 1 m adjusted to 10 m against collocated ERA5 reanalysis 10-m wind speeds, and the gray line indicates a 1:1 agreement; (c), (f) rotary spectra of wind speeds: waveglider observations are in blue, ERA5 reanalysis is in orange, solid lines are the negative spectra, and dashed lines are the positive spectra. The 95% confidence intervals are shown.

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