Surface Stress and Atmospheric Boundary Layer Response to Mesoscale SST Structure in Coupled Simulations of the Northern California Current System

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ABSTRACT

The influence of mesoscale sea surface temperature (SST) variations on wind stress and boundary layer winds is examined from coupled ocean–atmosphere numerical simulations and satellite observations of the northern California Current System. Model coupling coefficients relating the divergence and curl of wind stress and wind to downwind and crosswind SST gradients are generally smaller than observed values and vary by a factor of 2 depending on planetary boundary layer (PBL) scheme, with values larger for smoothed fields on the 0.25° observational grid than for unsmoothed fields on the 12-km model grid. Divergence coefficients are larger than curl coefficients on the 0.25° grid but not on the model grid, consistent with stronger scale dependence for the divergence response than for curl in a spatial cross-spectral analysis. Coupling coefficients for 10-m equivalent neutral stability winds are 30%–50% larger than those for 10-m wind, implying a correlated effect of surface-layer stability variations. Crosswind surface air temperature and SST gradients are more strongly coupled than downwind gradients, while the opposite is true for downwind and crosswind heat flux and SST gradients. Midlevel boundary layer wind coupling coefficients show a reversed response relative to the surface that is predicted by an analytical model; a predicted second reversal with height is not seen in the simulations. The relative values of coupling coefficients are consistent with previous results for the same PBL schemes in the Agulhas Return Current region, but their magnitudes are smaller, likely because of the effect of mean wind on perturbation heat fluxes.

1. Introduction

Offshore of a roughly 100-km-wide coastal strip in which orographic effects are important (Winant et al. 1988; Samelson 1992; Perlin et al. 2004), monthly mean mesoscale structure in wind stress over the California Current System (CCS; see appendix D for a list of acronyms) is dominated by the response to sea surface temperature (SST) and surface current variations associated with offshore extensions of cool coastal waters, which in turn originate from wind-driven coastal upwelling (Chelton et al. 2007; Seo et al. 2016; Renault et al. 2016). Although much progress has been made toward understanding the processes controlling SST–stress correlations throughout the global ocean (Chelton et al. 2001,

2004; Xie 2004; O'Neill et al. 2005, 2010; Samelson et al. 2006; Small et al. 2008; Schneider and Qiu 2015), many aspects of the coupling dynamics remain poorly understood, including the dependence of the response on specific characteristics of boundary layer turbulence, the vertical structure of the atmospheric response, and the factors that control regional variations in coupling strength.

One approach to the first of these questions is to consider the dependence of the simulated response on the particular planetary boundary layer (PBL) turbulence scheme used in a given atmospheric model simulation. Perlin et al. (2014) used this method to study SST–stress coupling in the Agulhas Return Current (ARC) region, and found that the strength of the simulated response was partially correlated with the amplitude of turbulent mixing represented by the various PBL schemes, where that amplitude was measured by the magnitude of the effective turbulent eddy viscosities and their variations. The same approach is taken here for the northern CCS region, using a suite of coupled ocean–atmosphere simulations that incorporate five of

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the PBL schemes considered by [Perlin et al. \(2014\)](#). This analysis examines further the dependence of the response on these PBL schemes, describes and interprets the corresponding boundary layer vertical structure, and compares the amplitude and characteristics of the responses for the CCS eastern boundary current and ARC open-ocean regimes. As such, it presents steps toward addressing all three of the questions listed above. Some of the simulations included surface-current coupling, but its inclusion is found to have no systematic effect on the SST coupling, and consequently the analysis of surface-current coupling processes and dynamics is not taken up here.

Two main hypotheses exist for the mechanisms that support the observed coupling: the vertical momentum mixing mechanism ([Wallace et al. 1989](#); [Hayes et al. 1989](#); [Samelson et al. 2006](#)) and the pressure gradient mechanism ([Lindzen and Nigam 1987](#)), both of which were originally proposed for tropical boundary layers. For SST–stress coupling in the CCS, scaling considerations suggest that the former will dominate the latter. An estimate of the maximum vertical-mean pressure gradient response $|\nabla p|$ from thermodynamic changes above an SST gradient $|\nabla T|$ is

$$|\nabla p| \approx \rho_a g \frac{h_e}{2} \theta^{-1} \frac{|\nabla T|}{2}, \quad (1)$$

where ρ_a is air density, g is gravity, and h_e and θ are representative boundary layer depth and potential temperature, respectively; in (1), the first factor of $1/2$ is geometric and the second is an upper bound for the thermal response, based on correlations of model SST and surface air temperature gradients ([section 3b](#)). The change in stress $\Delta\tau_e$ required to balance this pressure gradient in a momentum boundary layer of fixed thickness h and constant stress divergence is then

$$\Delta\tau_e \approx h \nabla p \approx \rho_a g \frac{h_e}{2} \theta^{-1} \frac{|\nabla T|}{2} h. \quad (2)$$

For the momentum (or momentum-deficit) mixing mechanism, an estimate of the change in stress $\Delta\tau_m$ required to balance a change Δh in h , for constant stress divergence $\tau/h = (\tau + \Delta\tau_m)/(h + \Delta h)$, is

$$\Delta\tau_m = \frac{\Delta h}{h} \tau. \quad (3)$$

The ratio of these two changes in stress is then

$$\left| \frac{\Delta\tau_e}{\Delta\tau_m} \right| \approx \rho_a g \theta^{-1} |\nabla T| \frac{h^2 h_e}{4 |\Delta h| \tau}, \quad (4)$$

suggesting that the momentum mixing mechanism will be favored for small h (and h_e) and large τ , while the

pressure gradient mechanism will be favored for large h (and h_e) and small τ . For $\rho_a = 1.2 \text{ kg m}^{-3}$, $g = 9.8 \text{ m s}^{-2}$, $\theta = 300 \text{ K}$, $|\nabla T| = 10^{-5} \text{ C m}^{-1}$, $h_e = h$, $\Delta h = 0.3h$, and $\tau = 0.1 \text{ N m}^{-2}$, the ratio (4) is less than one if $h < 600 \text{ m}$, and decreases quadratically with h . Thus, over the CCS, where boundary layer depths are typically of order 400 m or less, the vertical momentum mixing mechanism can be expected to dominate, and is the focus of the present analysis.

2. Methods

a. Coupled ocean–atmosphere simulations

Simulations were conducted using a coupled ocean–atmosphere numerical model configuration based on the Coupled Ocean Atmosphere Wave Sediment-Transport (COAWST; [Warner et al. 2010](#)) system and configured for the northern CCS region ([Figs. 1 and 2](#)). The Regional Ocean Modeling System (ROMS; [Shchepetkin and McWilliams 2005](#)) ocean model configuration had an approximately 2-km grid resolution and 40 terrain-following vertical levels, and the coupled, inner-nest configuration of the Weather Research and Forecasting (WRF) Model ([Skamarock et al. 2008](#)) regional atmospheric model had an approximately 12-km grid resolution with a surface-pressure-following vertical grid having 51 mass and horizontal velocity levels, 17 of them beneath a nominal height of 1500 m over the ocean, and the lowest at nominal 10-m height (see the appendixes). One additional simulation was done with a 58-level grid, consisting of the 51-level grid augmented by 7 additional mass and horizontal velocity levels below 200 m, and yielded results consistent with those for the original simulations, confirming that the standard 51-level grid provided adequate vertical resolution. Version 3.3 of the COAWST distribution was used, which contained WRF v3.9.1.1 and ROMS svn885. The 36-km WRF outer-nest, with a 130×138 horizontal grid, was in turn nested in output fields from the NCEP FNL operational analysis ([NCEP 2000](#)), and the ROMS ocean grid was nested in output fields from the HYCOM + NCODA Global $1/12^\circ$ Analysis (GLBa0.08/expt_90.8) product ([Bleck 2002](#); [Cummings 2005](#); [Fox et al. 2002](#)). Both the boundary and initial condition daily fields obtained from the HYCOM analysis were smoothed in time using a 5-day half-amplitude low-pass filter to remove aliased inertial oscillations that would otherwise arise.

Modifications to the standard COAWST distribution were made to allow bilinear interpolation of the atmospheric forcing fields onto the ocean model grid and to include surface current coupling in a subset of the simulations. The bilinear interpolation was introduced to

FIG. 1. Mean August 2009 (a) SST ($^{\circ}\text{C}$), (b) wind stress magnitude (N m^{-2}) and vectors (arrows, subsampled at every fifth grid row and column), (c) 10-m wind u_{10} (m s^{-1}), and (d) vector (arrows) and magnitude differences (shading) of 10-m equivalent neutral stability (ENS) wind u_{10N} and 10-m wind u_{10} (m s^{-1}) from an SST-coupled simulation with the WRF GBM planetary boundary layer scheme. The u_{10N} shown is that computed from stress using a constant z_0 . The region over which the coupling coefficients were computed is also indicated in each panel (black line).

avoid erroneous gridscale discontinuities in surface fluxes on the ocean model grid, which resulted from the nearest-neighbor interpolation in the standard COAWST distribution and caused spurious wind stress curl and divergence signals in the ocean component. The ocean model SST was passed to the atmospheric model grid using a nearest-neighbor interpolation without additional smoothing. The ocean model SST spectrum was found in these simulations to decay as inverse wave-number to the fourth power for wave-numbers greater than $1/24$ cpkm and to the eighth power for wave-numbers greater than $1/10$ cpkm, so errors or aliasing introduced by this subsampling approach should be small. The 7 March 2009 ocean field from an ocean-only

simulation initiated in October 2008 was used to initialize the coupled model, which was then integrated for the period 7 March 2009–15 October 2009. The coupled SST and flux fields were exchanged at each model time step. Numerous simulations of this period, with different atmospheric-model PBL schemes, and either with or without surface current coupling, were conducted. The primary analysis period was the four summer upwelling months June through September 2009.

b. PBL and surface-layer schemes

Coupled simulations were conducted using five different WRF PBL schemes: YSU, MYJ, MYNN2, UW, and GBM, implemented, respectively, in the WRF Model by

FIG. 2. As in Fig. 1, but for an SST-coupled simulation with the WRF MYNN2 planetary boundary layer scheme.

setting the configuration parameter `bl_pbl_physics` = {1(YSU), 2(MYJ), 5(MYNN2), 9(UW), 12(GBM)}. These are the same five WRF schemes used by [Perlin et al. \(2014\)](#), and were chosen partly on that basis, in order to allow comparison between the CCS and ARC regions. [Perlin et al. \(2014\)](#) also conducted some additional simulations using a different atmospheric model, which is not considered here. The MYJ, MYNN2, UW, and GBM schemes are all based on local turbulent closures, with eddy viscosities and diffusivities proportional to a local measure of turbulent kinetic energy and a mixing length. The YSU scheme is based on a nonlocal closure, in which the eddy viscosities and diffusivities are forced to have a fixed vertical profile structure, relative to a diagnosed planetary boundary layer height and turbulent intensity. A more detailed summary of the characteristics of each of these schemes, with supporting references, may be found in [Perlin et al. \(2014\)](#) and is not

repeated here. The GBM PBL scheme was not generally available for WRF prior to v3.5, when the [Perlin et al. \(2014\)](#) implementation, developed for WRF v3.3, was incorporated into the WRF distribution.

In all simulations except one, the revised MM5 Monin–Obukhov surface layer and flux scheme (WRF configuration parameter `sf_sfclay_physics` = 1) was used, which is a standard WRF option for the YSU, MYNN2, UW, and GBM PBL schemes. One SST-coupled MYJ simulation was conducted using the MYJ Monin–Obukhov (Janjić Eta) surface layer and flux scheme (`sf_sfclay_physics` = 2), which is the standard WRF option for the MYJ PBL scheme. The MYJ PBL code was then modified following [Perlin et al. \(2014\)](#) to allow use of the revised MM5 surface-layer scheme (`sf_sfclay_physics` = 1), and this combination of MYJ PBL and revised MM5 surface layer was used for the second SST-coupled and the surface-current-coupled

MYJ simulations. Some elements of the YSU and MYNN2 PBL schemes and the MM5 surface-layer scheme were modified between WRF v3.3 used by Perlin et al. (2014) and WRF v3.9.1.1 used here, but these changes related primarily to terrain and cloud effects, and evidently did not have a major effect on the representation of SST-stress coupling processes. A subset of simulations repeated using a version of WRF without those changes gave essentially equivalent results to those reported here, and, as described below in section 5, the relative intensity of the coupling represented by the various schemes found here was largely consistent with that reported by Perlin et al. (2014).

The CCS simulations differ from those of Perlin et al. (2014) in that they include a coupled ocean model component, whereas the Perlin et al. (2014) study was based on atmosphere-only simulations. Thus, the SST field is different in each of the CCS simulations, whereas it was imposed and therefore identical in all of the Perlin et al. (2014) ARC simulations. The inclusion of the coupled ocean component and associated SST field evolution provides a more consistent dynamical context for the diagnosis of SST-stress coupling but does not appear to affect strongly the basic coupling processes.

c. Model-based coupling coefficients

The surface stress and wind responses to the upwelling-driven mesoscale SST variability can be quantified in terms of coupling coefficients that measure the stress and wind response to SST fluctuations (Chelton et al. 2007; O'Neill et al. 2012; Perlin et al. 2014). The model SST-wind and SST-stress coupling coefficients were computed for monthly mean fields. Because of the presence of the coastal boundaries and the general spatial inhomogeneity of the eastern boundary current regime, spatial high-pass filtering of the model fields following Perlin et al. (2014) or other similar studies was not practical. Instead, following Chelton et al. (2004, 2007) and O'Neill et al. (2005), the coupling coefficients were computed only for the divergence and curl of wind and wind stress, for which the derivative operations effectively produced a high-pass-filtering effect.

The coupling coefficients were computed from 3-hourly WRF Model output fields on the 12-km inner-nest grid. Spatial derivatives were computed by standard centered differences on that grid. Surface (10-m) winds and stress (friction velocity) output fields were located at the temperature points and used directly for the derivative computations. The horizontal components of the three-dimensional wind field were interpolated from their respective staggered C-grid cell-edge points to the temperature (and pressure) points at cell centers prior to the derivative computations. In addition, for comparison

with the scatterometer-based wind coupling coefficients, 10-m equivalent neutral-stability (ENS) wind fields were computed from the stress output fields, as described below (section 2e).

Based on the 10-m wind direction, 3-hourly downwind and crosswind SST gradients were computed. Monthly means of the 3-hourly values for these and the other, undifferentiated fields at the temperature points were computed for each month during June through September 2009. The horizontal divergence and the vertical component of the curl were computed for the mean surface stress, 10-m and 10-m ENS winds, and the three-dimensional wind field. Points outside the range 36° to 48° N, west of 128° W, or within 100 km (9 grid points) of the coastline, as well as those over land, were removed prior to computing the derivatives (Figs. 1 and 2). The resulting subdomain contained the dominant mesoscale SST structure and avoided the coastal region where orographic effects dominate the mesoscale stress structure. For the three-dimensional wind field, the horizontal derivative operations were computed on fixed model coordinate surfaces, which over the ocean were nearly equivalent to fixed pressure surfaces, with departures from the fixed pressure surfaces induced only by changes in the surface pressure and decreasing essentially linearly with height. Seasonal anomalies of all of the derivative fields, including the downwind and crosswind SST gradients, were computed as the departure of each monthly mean from the June through September (JJAS) seasonal mean.

Two different methods were then used to compute the coupling coefficients. First, binned averages were computed for the wind and stress derivative quantities as a function of the downwind and crosswind SST gradients, and the coupling coefficients were computed as linear regression slopes for the corresponding means over the SST gradient range between $\pm 10^{-5} \text{C m}^{-1}$ [$\pm 1^{\circ}\text{C} (100 \text{ km})^{-1}$]. Second, linear regressions were performed simultaneously on the entire set of monthly anomaly points, without bin-averaging. Differences in the model coefficients computed on the 12-km model grid by these two different regression methods were small (see section 3b). The second regression method was also applied to model fields that were first interpolated to the $0.25^{\circ} \times 0.25^{\circ}$ QuikSCAT/SST grid, using a loess smoother (Cleveland and Devlin 1988) with a 48-km half-power-point. As discussed below, these two gridded versions gave slightly different results for the model coupling coefficients, but with consistent relative magnitudes of the different coefficients for the various PBL schemes.

Because of the monthly mean time averaging and the single summer season of simulation, June through September 2009, the distributions of binned values for

FIG. 3. Means ($\circ$) and means $\pm$ standard deviations (vertical lines) of binned distributions of monthly means of seasonal anomalies of (a),(c) divergence of stress (10^{-7} N m^{-3}) vs downwind SST gradient ($10^{-5} \text{ }^{\circ}\text{C m}^{-1}$) and (b),(d) curl of stress (10^{-7} N m^{-3}) vs crosswind SST gradient ($10^{-5} \text{ }^{\circ}\text{C m}^{-1}$) computed on the model grid from examples of SST-coupled simulations using the (a),(b) GBM and (c),(d) MYNN2 planetary boundary layer schemes.

each simulation primarily represent spatial rather than temporal variability. Standard deviations were computed from the variances of the four monthly means for each bin, and represent the month-to-month variability of the bin means; the standard deviations from spatial variability of the point-wise monthly means within each mean are larger. For brevity, the coupling coefficients for the divergence response to downwind SST gradients and the curl (vorticity) response to crosswind gradients are denoted in the text by, respectively, $\alpha_d(X)$ and $\alpha_c(X)$, where $X = \{\tau, U, U10N\}$ indicates, respectively, stress, 10-m wind, or 10-m ENS wind.

d. Scatterometer-based coupling coefficients

SST-wind and SST-stress coupling coefficients were computed for the time period of the model simulations using wind stress fields derived from the QuikSCAT satellite scatterometer surface winds and SST from the daily NCEI OI v2 AVHRR-only analysis (National Centers for Environmental Information 2016; Reynolds et al. 2007), which is constructed on the same uniform 0.25° latitude–longitude grid as the QuikSCAT gridded wind fields. Only the second method, linear regression to all monthly anomaly points, was used to compute the observed coupling coefficients. The QuikSCAT surface wind observations used in this analysis were obtained from, and processed by, JPL/PO.DAAC in Level-2B format, version 4.0 (SeaPAC 2018). These fields were interpolated onto the 0.25° grid using a two-dimensional loess smoother weighted by a tricubic function with a radial half-power point of 80 km, corresponding to a wavelength resolution of about 48-km. Spatial derivatives were computed within each gridded swath using centered finite differences, which is often referred to as “in-swath.” Time-mean derivative wind fields were computed from the gridded swaths of instantaneous derivatives.

e. Stress and ENS winds

The wind stress was computed (see appendix B) from the QuikSCAT winds using both a constant roughness length $z_0 = 0.152 \times 10^{-3}$ m, which corresponds to a constant 10-m neutral drag coefficient $C_{DN} = 1.3 \times 10^{-3}$, and the COARE 3.0 10-m neutral drag coefficient (Fairall et al. 2003), which is a more accurate representation of the empirical physical relation between 10-m wind speed and stress under neutral stability conditions. The QuikSCAT scatterometer measures backscatter

from the ocean surface, and is more directly a measure of surface conditions—and therefore stress—than of 10-m winds, so the implied relation between stress and 10-m winds incorporated into the calibration method affects the encoding of stress information in the QuikSCAT 10-m ENS wind product. The original QuikSCAT geophysical model function incorporated the Large and Pond (1981) neutral drag coefficient. For the most recent buoy-wind calibration of the QuikSCAT geophysical model function, the constant- z_0 neutral drag coefficient was used to correct the buoy winds to 10-m height, evidently without any stability corrections (F. Wentz, personal communication). The current geophysical model function was calibrated instead to WindSat winds (Ricciardulli and Wentz 2015), but the latter were apparently calibrated to the same buoy winds, which effectively retains the original constant- z_0 dependence.

Computing the wind stress from the QuikSCAT winds using the 10-m neutral drag coefficient from the COARE 3.0 formulation results in QuikSCAT-derived coupling coefficients that are approximately 30% larger than those obtained using the constant- z_0 formulation. This difference presumably arises from the steeper slope of the COARE 3.0 stress versus 10-m wind speed relation for wind speeds greater than roughly 7 m s^{-1} (see appendix B, Fig. B1). In the absence of a more direct comparison of QuikSCAT stresses to measured stresses, it is difficult to determine which of these approaches is more accurate. For consistency and comparison, model 10-m ENS winds have also been computed from the model stress using both of these formulations. Thus, two sets of stress coupling coefficients are presented for the QuikSCAT observations, and two sets of 10-m ENS wind coupling coefficients are presented for the model simulations. The neutral drag coefficient for the WRF Model `sf_sfclay_physics = 1` surface layer scheme used in all but one of the simulations is very similar to the COARE 3.0 coefficient, while the Large and Pond (1981) coefficient is also similar to COARE 3.0 but somewhat smaller (see appendix B, Fig. B1).

3. Results

a. Model SST and stress fields

As expected for the coastal upwelling regime of the northern California Current System, the monthly mean

←

The linear regression lines (thick solid lines) to the means computed using the first regression method and the corresponding observed scatterometer results (dashed lines, slopes 3.16 and 1.76 for COARE3.0 and 2.49 and 1.41 for constant z_0 , respectively) computed using the second regression method are also shown (see section 2c).

FIG. 4. As in Fig. 3, but for means ($\circ$) and means $\pm$ standard deviations (vertical lines) of binned distributions of monthly means of seasonal anomalies of (a),(c) divergence of winds (10^{-5} s^{-1}) vs downwind SST gradient (10^{-5} C m^{-1}) and (b),(d) curl of winds (10^{-5} s^{-1}) vs crosswind SST gradient (10^{-5} C m^{-1}), for (a),(b) 10-m and (c),(d) constant- z_0 10-m ENS winds computed on the model grid from the simulation using the GBM planetary boundary layer scheme. The corresponding observed results for the 10-m ENS scatterometer winds (dashed lines, slopes 0.96 and 0.56, respectively) computed using the second regression method (see section 2c) are also shown.

SST fields during JJAS have distinct minima in the coastal zone, with coastal surface waters often 10°C colder than those offshore (Figs. 1 and 2). Tongues of this cold, coastal upwelled water may extend 250 km or more offshore, creating large offshore mesoscale SST gradients beneath the predominantly northerly summertime surface winds. The interaction of these SST gradients with the overlying atmospheric boundary

layer produces large mesoscale variations in surface stress and 10-m winds. The associated variations in lower boundary layer stability and the shallow depth of the boundary layer result in systematic differences of order $\pm 1 \text{ m s}^{-1}$ in model 10-m wind and 10-m ENS wind, with fluctuations in the magnitude of these winds visually correlated with mesoscale SST variability (Figs. 1 and 2). The difference between the magnitudes of the

FIG. 5. As in Fig. 4, but for the simulation using the MYNN2 planetary boundary layer scheme.

10-m ENS wind and the 10-m wind, $|\mathbf{u}_{10N}| - |\mathbf{u}_{10}|$, is itself also visually correlated with the mesoscale SST variability, suggesting that the lower boundary layer is more strongly stratified over relatively cool SST and more weakly stratified over relatively warm SST (Figs. 1 and 2).

b. CCS model and scatterometer coupling coefficients

Binned averages of the monthly mean seasonal anomalies of the divergence and curl of stress as functions of, respectively, the downwind and crosswind SST gradient components show a linear dependence of the

stress derivatives on the SST gradients over the SST gradient range between $\pm 10^{-5} \text{C m}^{-1}$ (Fig. 3), qualitatively consistent with the results of Chelton et al. (2007) for the CCS and with Perlin et al. (2014) and many other previous studies for other regions. This linearity motivates the computation of the linear regression fits to the bin means and the interpretation of the corresponding regression slopes as coupling coefficients.

The qualitative structure of the binned averages is similar for all simulations, but the values of the resulting coupling coefficients depend upon the planetary boundary layer scheme used in the respective simulation. Two

FIG. 6. As in Fig. 3, but for means ($\circ$) of binned distributions of monthly means of seasonal anomalies of (a),(c) downwind SAT gradient vs downwind SST gradient and (b),(d) crosswind SAT gradient vs crosswind SST gradient, for the (a),(b) GBM and (c),(d) MYNN2 examples. The linear regression lines (black lines, slopes as indicated in label) to the means computed using the first regression method are shown.

examples are described in detail. A simulation with the GBM planetary boundary layer scheme, for example, had linear regression slopes $\alpha_d(\tau) = 1.53$ (stress divergence versus downwind SST gradient) and $\alpha_c(\tau) = 1.55$ (stress curl versus crosswind SST gradient), for stress and SST derivatives in units of 10^{-7} N m^{-3} and 10^{-5} K m^{-1} , respectively (or, equivalently, for slope units of $10^{-2} \text{ N m}^{-2} \text{ K}^{-1}$), where both were computed by the first method (see section 2c). The corresponding values for a simulation conducted with the MYNN2 scheme were 30%–40% larger, $\alpha_d(\tau) = 2.18$ and $\alpha_c(\tau) = 1.98$ (Fig. 3).

For these examples, the corresponding linear regression slopes computed by the second method on the model grid were $\alpha_d(\tau) = \alpha_c(\tau) = 1.50$ for GBM, and $\alpha_d(\tau) = \alpha_c(\tau) = 2.24$ for MYNN2. The corresponding observed values were $\alpha_d(\tau) = 3.16$ and $\alpha_c(\tau) = 1.76$ using the COARE 3.0 drag coefficient, or $\alpha_d(\tau) = 2.49$ and $\alpha_c(\tau) = 1.41$ using constant z_0 , where both sets were computed by the second method on the QuikSCAT/SST grid. Thus, for these examples, the model coupling coefficients computed on the model grid captured only 50%–90% of the stress divergence coupling, but at least 85% of the stress curl

FIG. 7. Summary of coupling coefficients computed using the second regression method from QuikSCAT observations (lines; stress results for both COARE 3.0 and constant- z_0) and using the first regression method on the $12 \text{ km} \times 12 \text{ km}$ model grid from the suite of simulations (small and large $\times$ and $\circ$), as a function of WRF bl_pbl_physics boundary layer scheme; simulations were conducted for bl_pbl_physics={1 (YSU), 2 (MYJ), 5 (MYNN2), 9 (UW), 12 (GBM)}. (a) Divergence of stress vs downwind SST gradient [$\alpha_d(\tau)$], (b) curl of stress vs crosswind SST gradient [$\alpha_c(\tau)$], (c) divergence of constant- z_0 10-m ENS (large $\times$) and 10-m (small $\times$) wind vs downwind SST gradient [$\alpha_d(U_{10N})$, $\alpha_d(U)$], and (d) curl of constant- z_0 10-m ENS (large $\circ$) and 10-m (small $\circ$) wind vs crosswind SST gradient [$\alpha_c(U_{10N})$, $\alpha_c(U)$].

coupling, with both MYNN2 model values for stress curl exceeding both observed values.

The 10-m ENS winds are, locally, related uniquely to the stress through the neutral drag coefficient, and so it might be expected that the corresponding coupling coefficients will have values relative to the observed neutral-wind coefficients that are similar to the relative values for stress. This expectation is only partly met, with the difference attributable to the combination of the nonlinearity in the neutral drag coefficient relation and the averaging procedure used to compute the mean response (O'Neill et al. 2012). For the two example solutions, the constant- z_0 neutral-wind divergence coupling for the GBM scheme is again much weaker than the observed divergence coupling, with $\alpha_d(U_{10N}) = 0.69$ versus the observed 0.96, respectively, but the neutral-wind curl coupling for GBM is similar to the observed value, with coefficient $\alpha_c(U_{10N}) = 0.62$ versus

the observed 0.56, respectively (Fig. 4). For the MYNN2 example, $\alpha_d(U_{10N}) = 0.91$, close to the observed 0.96, and $\alpha_c(U_{10N}) = 0.77$, 40% larger than the observed 0.56 (Fig. 5). The standard deviations of the monthly means are again systematically larger for curl than for divergence.

The coupling coefficients for model 10-m wind were systematically only two-thirds those for the corresponding 10-m ENS wind. The model 10-m wind coefficients for the GBM example were $\alpha_d(U) = 0.45$ and $\alpha_c(U) = 0.39$ compared to model 10-m ENS wind values $\alpha_d(U_{10N}) = 0.69$ and $\alpha_c(U_{10N}) = 0.62$ (Fig. 4). For the MYNN2 example, the corresponding values were $\alpha_d(U) = 0.59$ and $\alpha_c(U) = 0.48$, relative to $\alpha_d(U_{10N}) = 0.91$ and $\alpha_c(U_{10N}) = 0.77$ (Fig. 5). Relative to the standard deviations of the distributions of the corresponding monthly means, these slopes are nonetheless still significantly different from zero, with the curl again

FIG. 8. As in Fig. 7, but for the model coupling coefficients computed using the second regression method on the $12 \text{ km} \times 12 \text{ km}$ model grid. Results are shown for both constant- z_0 and COARE 3.0 model 10-m ENS winds.

having standard deviations systematically larger than the divergence.

In general, stability corrections to 10-m wind speeds under midlatitude conditions are only 10%–20% of the 10-m ENS wind speed (O'Neill et al. 2012; Small et al. 2008). The large, systematic differences between the 10-m wind and 10-m ENS wind coupling coefficients therefore must be attributed to a stability effect that is correlated with the SST gradients. The sign of the boundary layer stability correction is such that, for a given 10-m wind speed, the stress—and therefore the 10-m ENS wind—is relatively larger under unstable conditions and relatively smaller under stable conditions, where the latter corresponds to surface air temperature (SAT) relatively cooler than SST and the former to SAT warmer than SST. Therefore, if SAT is uniform, then this correction will tend to increase the ENS wind coupling coefficients relative to the 10-m wind coupling coefficients. This effect will be reduced as the local SAT approaches equilibrium with the local SST.

A measure of the mean approach to this equilibration can be derived by computing a coupling coefficient

between SAT and SST, using the same methodology as for the stress and wind coupling coefficients. By analogy to the latter, it is natural to consider the coupling between the downwind SAT gradient and the downwind SST gradient, and between the crosswind SAT gradient and the crosswind SST gradient. For both the GBM and the MYNN2 SST-coupled simulations, the corresponding coupling coefficients are 0.33 for the downwind components and 0.40 for the crosswind components (Fig. 6). The locally equilibrated state would have coefficients of 1.00 for both components, so both of these coefficients indicate that local SAT is relatively far from equilibrium with local SAT, for the mesoscale gradients of interest here. Stronger fluctuations in surface buoyancy flux and stability will occur when local SAT and SST are farther from equilibrium (Chelton et al. 2001), and it is these fluctuations in stability that control the depth of momentum mixing and drive the stress and wind response both at the surface and throughout the boundary layer (Samelson et al. 2006). The along-SST-isotherm winds that occur in the crosswind case would be parallel also to the SAT isotherms in the equilibrated

FIG. 9. As in Fig. 7, but for the model coupling coefficients computed on the $0.25^\circ \times 0.25^\circ$ QuikSCAT/SST grid. Results are shown for both constant- z_0 and COARE 3.0 model 10-m ENS winds.

case, and thus do not act to drive the boundary layer away from the local SAT–SST equilibrium, consistent with the larger crosswind SAT–SST coupling coefficient.

For the five WRF PBL schemes considered, the relative magnitudes of the coupling coefficients vary by a factor of two, with a systematic dependence on PBL scheme (Fig. 7). Ordered from weakest to strongest response, the PBL schemes are MYJ, YSU, GBM, UW, and MYNN2. For the divergence of stress, none of the PBL schemes give coupling coefficients as large as observed, despite the factor of two variation in model values. For the stress curl and constant- z_0 10-m ENS wind coefficients, the observed values all fall within the ranges spanned by the schemes. The best matches to the observations, for this comparison based on model coefficients using the first linear regression method and the constant- z_0 neutral stability winds, are found for the MYNN2 scheme for both stress divergence and 10-m ENS wind divergence, and for the GBM scheme for both stress curl and 10-m ENS wind curl (Fig. 7). Thus no single scheme is optimal for all of the coefficients, but those with the largest responses might arguably be favored over the others.

Notably, there is no systematic difference between the magnitudes of the divergence and the curl coupling coefficients, for these values computed on the model grid by either the first or the second method (Figs. 7 and 8). This result contrasts with the systematically larger divergence coefficients, relative to curl, found in a number of previous studies (Chelton et al. 2001, 2004, 2007; O'Neill et al. 2005) and for the CCS QuikSCAT observations themselves (Figs. 7–9). In contrast to the coefficients computed on the model grid, a systematic difference is found for the coefficients computed by the second method for the spatially smoothed fields on the 0.25° QuikSCAT/SST grid (Fig. 9). This difference can apparently be attributed to the different scale dependencies of the divergence and curl responses. Meridional wave-number spectral squared coherences and transfer functions for the model divergence of stress versus downwind SST gradient show a roughly 20% amplification in the 50–200-km wavelength range relative to the same quantities for curl versus crosswind SST gradient (Figs. 10a,b). A similar spatial dependence is found for the 10-m wind and ENS wind quantities. This suggests that the systematically larger

FIG. 10. Ensemble mean spectral (a) coherence squared, (b) transfer function amplitude, and (c) transfer function phase vs meridional wavenumber for wind stress divergence response to downwind SST gradient (thick solid black lines) and wind stress curl response to crosswind SST gradient (thick dashed black), from the suite of CCS model simulations. The corresponding quantities for the individual ensemble members are also shown (thin solid and dashed lines; colors as in legend, for: I—YSU; J—GBM; K—UW; L—MYJ; M—MYNN2).

divergence coefficients relative to those for curl arise from the smoothing and interpolation from the 12-km model grid to the coarser $0.25^\circ \times 0.25^\circ$ QuikSCAT/SST grid, which removes relatively more of the coherent curl than divergence response.

The corresponding transfer function phase spectra also indicate different spatial equilibration scales. For $\alpha_d(\tau)$ and $\alpha_c(\tau)$, the phase angle ϕ is approximately zero

for wavelengths L_0 near 400 and 300 km, respectively, and increases in magnitude linearly with increasing wavenumber, to maxima near $L_1 = 60$ km wavelength of roughly 30° for divergence and 20° for stress (Fig. 10c). Taking $L_{eq} = (2\pi)^{-1} \Delta\phi / (1/L_1 - 1/L_0)$ for ϕ in radians gives equilibration scales L_{eq} of roughly 6 km for the wind stress divergence response to downwind SST gradient and 4 km for wind stress curl response to crosswind SST gradient. Although these are smaller than the atmospheric model grid spacing, they are consistent with adjustment scales for the atmospheric boundary layer in high-resolution numerical studies of the response to abrupt surface temperature changes (e.g., Skyllingstad et al. 2005). It is likely that the 90° difference in relative SST gradient and wind directions is the critical factor causing this difference in equilibration scales, with heat advection tending to drive the system more directly away from equilibration in the downwind than the crosswind SST gradient case.

There is also a dependence of the SAT gradient versus SST gradient coupling on PBL scheme (Fig. 11a). Provided that the same surface-layer scheme is used in each case, the associated ordering of schemes is consistent with, although less systematic than, that for the stress and wind coupling coefficients, with the MYJ scheme giving the weakest response, the YSU scheme giving the next weakest, and the GBM, UW, and MYNN2 schemes giving essentially equal and slightly larger responses. When the WRF-standard MYJ surface-layer scheme is used for MYJ, the MYJ SAT-SST gradient coupling is instead the strongest of all. Surprisingly, this dependence of the MYJ response on the surface-layer scheme does not carry over the surface heat-flux gradient coupling coefficient (Fig. 11b), and consequently it seems likely that the different SAT-SST gradient coupling may primarily reflect the different algorithm by which the surface air temperature is related to the surface fluxes in the MYJ surface-layer scheme. For all schemes, the crosswind SAT gradient is more strongly coupled to the crosswind SST gradient than is the downwind SAT gradient to the downwind SST gradient, with the opposite true for the heat-flux gradient coupling. This difference is consistent with the shorter equilibration scales for the wind stress curl versus crosswind SST gradient relative to the wind stress divergence versus downwind SST gradient, with the latter being locally farther from equilibrium (neutral) stability than the former: the greater degree of SAT equilibration to local SST for the crosswind gradients, with surface wind along SST isotherms, relative to the downwind gradients, with surface wind across SST isotherms, may lead to a weaker effect of stability from SAT-SST differences and a corresponding reduction in the stability-controlled vertical mixing of horizontal momentum (Chelton et al. 2001).

FIG. 11. Crosswind and downwind (left panel) SAT and (right) surface heat flux gradient coupling coefficients for downwind gradients vs downwind SST gradient ($\times$) and crosswind gradients vs crosswind SST gradient ($\circ$), computed using the first regression method on the $12\text{ km} \times 12\text{ km}$ model grid from the suite of simulations, as a function of WRF `bl_pbl_physics` boundary layer scheme; simulations were conducted for `bl_pbl_physics` = {1 (YSU), 2 (MYJ), 5 (MYNN2), 9 (UW), 12 (GBM)}. In both panels, the left-most pairs of MYJ coefficients were computed with the WRF-standard MYJ Monin–Obukhov (Janjic Eta) surface layer scheme (`sf_sfclay_physics` = 2), while all of the others including the two additional MYJ coefficient pairs were computed with the revised MM5 Monin–Obukhov scheme (`sf_sfclay_physics` = 1). The coefficients for downwind SAT gradient vs crosswind SST gradient and for crosswind SAT gradient vs downwind SST gradient were all effectively zero.

Because of the mixed ordering of the schemes, the SAT–SST gradient coupling coefficients do not, through this argument, explain the systematic variations in SST–stress coupling strength between schemes. However,

they do provide modest evidence for the systematic influence across all schemes of the SAT–SST gradient coupling on the difference between 10-m wind and 10-m ENS wind coupling coefficients. If the conditions were perfectly

FIG. 12. Ratio of model 10-m wind to constant- z_0 10-m ENS wind coupling coefficients vs SAT–SST gradient coupling coefficients for (a) wind divergence and downwind SAT gradient vs downwind SST gradient, and (b) wind curl and crosswind SAT gradient vs crosswind SST gradient, for each of the model simulations.

FIG. 13. Ensemble mean CCS model coupling coefficients for (a) wind divergence response to downwind SST gradient, (b) wind curl response to crosswind SST gradient, (c) wind divergence response to crosswind SST gradient, and (d) wind curl response to downwind SST gradient, vs corresponding SST gradients (10^{-5}C m^{-1}) and pressure (hPa). A mean profile over all corresponding SST gradient bins has been removed in each case. Wind divergence and curl units are 10^{-5} s^{-1} . The maximum values near the surface are of order $\pm 0.5 \times 10^{-5}\text{ s}^{-1}$, as can be inferred from the 10-m wind coupling coefficients in Fig. 7.

neutral, the ratio of the latter coefficients would be unity. In the present regime, however, the 10-m ENS wind coefficients are all systematically larger than the 10-m wind coefficients, consistent with the surface layer being unstable in the mean. When all schemes and simulations are considered jointly, there is a recognizable trend for stronger SAT–SST gradient coupling – consistent with more rapid adjustment of boundary layer temperature to SST and thus weaker overall boundary layer stability – to give coupling coefficients for which the ratio of the 10-m wind to 10-m ENS wind coefficients are relatively closer to unity (Fig. 12).

4. Low-level atmospheric response

a. Coupling coefficient profiles

From the QuikSCAT and SST observations, only the surface stress and 10-m ENS wind coupling can be assessed. For the simulations, however, coupling

coefficients can be computed for winds above the surface. The resulting wind–SST coupling coefficients for winds on constant pressure levels show systematic indications of a reversal in sign above a surface boundary layer of approximately 100 m (10 hPa below surface pressure). For the GBM solution discussed above (e.g., Fig. 4), for example, the regression slopes from the binned averages for wind divergence and curl at 400 m height (966 mb) are -0.025 and -0.080 . Relative to the standard deviations of the distributions of monthly means, these slopes are not significantly different from zero, but the tendency for linear anticorrelation is systematic and consistent across the bins.

Ensemble mean profiles of the mean wind divergence and curl response to downwind and crosswind SST gradients, respectively, for the complete set of coupled CCS model simulations show both the near-surface and the reversed midboundary layer responses (Fig. 13). Similar structures, with more variability, are evident in

FIG. 14. Ensemble mean CCS model coupling coefficients for (a) downwind θ gradient response to downwind SST gradient, (b) crosswind θ gradient response to crosswind SST gradient, (c) downwind θ gradient response to crosswind SST gradient, and (d) crosswind θ gradient response to downwind SST gradient, vs corresponding SST gradients (10^{-5}C m^{-1}) and pressure (hPa). A mean profile over all corresponding SST gradient bins has been removed in each case.

all of the individual simulation means. The reversed midboundary layer response is evident in both the curl response to the crosswind SST gradient and the divergence response to the downwind SST gradient. Perhaps surprisingly, a similar structure is found for the complementary pair, the divergence response to crosswind SST gradient and the curl response to downwind SST gradient, with maximum amplitudes approximately half those of the standard pair (Fig. 13). The simulations appear to show an additional deep curl response, above the first reversal, but examination of spatial maps of this curl response and the SST gradients suggests that it is a spurious correlation, arising from large-scale structures in the curl and SST gradient fields that have no immediate physical connection through the boundary layer. No evidence is thus found in these profiles for an interior, free-atmosphere response to the SST-induced modulations of wind stress and boundary layer winds. Vertical velocities induced by the near-surface divergence may either be too small to detect in these

means, or they may disperse through radiation of gravity waves (Kilpatrick et al. 2014).

Ensemble mean profiles of the coupling coefficients for downwind and crosswind potential temperature gradients at fixed model levels versus downwind and crosswind SST gradients show that the coupling is weak, with coefficients generally no larger than 0.1 in all cases and at all levels (Fig. 14). The temperature responses are confined below roughly 100 m, while weak responses above are likely related to geostrophic contributions to the surface wind. These results are consistent with the hypothesis that vertical momentum mixing effects dominate pressure gradient effects in the SST-induced wind response. Both the crosswind and the downwind near-surface potential temperature gradients show a stronger response to crosswind SST gradients than to downwind SST gradients. The correlation of crosswind near-surface potential temperature gradient with crosswind SST gradient can be understood as a indicating partial equilibration

FIG. 15. Profiles vs height z (m) of (a) eddy viscosity $A_V(z)$ ($\text{m}^2 \text{s}^{-1}$ for $z_1 = 1.15 z_1^m$, $z_2 = 1.05 z_2^m$, representing the warm-SST case with deeper turbulence (solid blue line), and for $z_1 = 0.85 z_1^m$, $z_2 = 0.95 z_2^m$, representing the cool-SST case with shallower turbulence (dashed green), for the log-Ekman model (A7) with $(z_1^m, z_2^m, z_3^m) = (50, 150, 700)$ m; (b) kinematic upward stress ($\text{m}^2 \text{s}^{-2}$) components $\tau^x = -A_V du/dz$ (thin solid) and $\tau^y = -A_V dv/dz$ (dashed) and magnitude $|(\tau^x, \tau^y)|$ (thick solid) from the solutions for the warm-SST (blue) and cool-SST (green) eddy viscosity profiles; (c) velocity components (m s^{-1}) u (thin solid) and v (dashed) and magnitudes $|(u, v)|$ (thick solid) from the solutions for the warm-SST (blue) and cool-SST (green) eddy viscosity profiles; (d) warm-SST minus cool-SST velocity differences (m s^{-1}) $u_+ - u_-$ (solid) and $v_+ - v_-$ (dashed) from the profiles in (c).

of the near-surface potential temperature with the local SST, for winds parallel to SST isotherms, while that of downwind near-surface potential temperature gradient with crosswind SST gradient is difficult to explain through a local argument, given that the SST field is locally uniform downwind. The latter may arise as a remnant of the mean large-scale structure, with alongshore near-surface winds over cool upwelled water resulting in near-surface atmospheric cooling in predominantly southward flow (Figs. 1c and 2c) over predominantly westward SST gradients (Figs. 1a and 2a).

b. Log-Ekman boundary layer model: Formulation

To analyze the dynamical context of the structure revealed in the coupling coefficient cross sections, an idealized, analytical model of the atmospheric boundary

layer is constructed, and variations in the specified eddy coefficient profile are inferred by approximately matching the profiles of the SST-correlated lower atmosphere responses from the simulations. The analytical model is based on the Ellison (1956) Hankel-function solution of the Ekman-layer equations with log-layer eddy viscosity $A_V = \kappa u_*(z + z_0)$, where A_V is the eddy viscosity, κ is a von Kármán constant, u_* is the friction velocity derived from surface stress, z_0 is the roughness length, and z is height above the mean sea surface. The model equations are given in the appendixes. This approach is complementary to that of Perlin et al. (2014), who diagnosed the numerical model eddy diffusivity profiles and found correlations between their variations and variations in the intensity of SST-wind and SST-stress coupling. The Perlin et al. (2014) approach addresses the model

parameter variations more directly, but the present approach has the advantage of providing an accessible, closed-form representation of the associated PBL physics and response that facilitates dynamical interpretation. The log-Ekman model vertical structure is similar to that of the SST-dependent generalized Ekman model considered by Schneider and Qiu (2015).

Consider the difference between two solutions of the log-Ekman boundary layer model (C3) that are taken to represent conditions over, respectively, warm-SST and cool-SST perturbations. The warm and cool SST cases are represented by increasing or decreasing the depth penetration and amplitude of the specified eddy viscosity, respectively (Fig. 15). These variations are consistent with the general argument (Wallace et al. 1989; Hayes et al. 1989; Samelson et al. 2006) that the SST-stress correlations arise from stability-controlled variations in the depth of vertical momentum mixing, that is, the depth range over which the momentum lost to the ocean through surface drag is extracted from the atmospheric boundary layer. To maintain the horizontal momentum balance near equilibrium, the stress vertical divergence must remain essentially unchanged as this depth of momentum mixing changes. Consequently, when under relatively unstable conditions the momentum deficit is spread over a deeper momentum boundary layer, the surface stress must increase, while when under relatively stable conditions the momentum deficit is spread over a shallower momentum boundary layer, the surface stress must decrease (Samelson et al. 2006).

The different profiles of the eddy viscosity A_v will in general result in the 10-m wind in the two cases having slightly different directions (O'Neill et al. 2010). To produce a model prediction of the wind response to down-wind and crosswind SST perturbations, which can be compared with similar responses in the observations or the coupled numerical circulation model, the wind direction for the pair of solutions is defined here as the direction of the mean 10-m wind for the two profiles. Let the x axis be taken along the direction of this mean 10-m wind, so that the corresponding wind velocity component is u , with y and v denoting the perpendicular axis and velocity component. Because the solutions are computed for horizontally homogeneous conditions, the implied orientation of the SST gradient ∇T relative to the mean 10-m wind may be specified arbitrarily. If this SST gradient is taken to be in the x -direction, then the velocity differences may be interpreted as responses to downwind SST gradients, while if it is taken instead to be in the negative y direction, the velocity differences may be interpreted as responses to crosswind SST gradients.

Let $(u, v)_w$ and $(u, v)_c$ represent the velocities over the relatively warm and cool SST perturbations, respectively,

FIG. 16. Log-Ekman model of coupling coefficient profiles vs SST gradient and height (m) for (a) wind divergence response to downwind SST gradient, or wind curl response to crosswind SST gradient, and (b) wind divergence response to crosswind SST gradient, or the negative of wind curl response to downwind SST gradient, for the profiles of eddy viscosity $A_v(z)$ shown in Fig. 15. For wind divergence and curl units of 10^{-5} s^{-1} and SST gradient units of 10^{-5} C m^{-1} as in Fig. 13, the wind speed difference profiles in Fig. 15 would be found over a 100-km distance for a mean SST gradient of 10^{-5} C m^{-1} [$1^\circ \text{C (100 km)}^{-1}$].

which are separated by arbitrary distances Δx or Δy , depending on the SST gradient orientation. For the case of a downwind SST gradient ($dT/dx \neq 0$ and $dT/dy = 0$), there are no implied gradients of any properties or variables in y , so the divergence $\partial u/\partial x + \partial v/\partial y = du/dx$ and the curl $\partial v/\partial x - \partial u/\partial y = dv/dx$. The implied wind divergence and curl responses can then be obtained from the warm-cool differences as $(u_w - u_c)/\Delta x$ and $(v_w - v_c)/\Delta x$, respectively. For the case of a crosswind SST gradient ($dT/dx = 0$ and $-dT/dy \neq 0$), there are no implied gradients of any properties or variables in x , and by a similar argument the implied divergence and curl responses can be obtained as $(v_c - v_w)/\Delta y$ and $-(u_c - u_w)/\Delta y$, respectively. The corresponding warm-cool SST perturbation differences

FIG. 17. As in Fig. 15, but for z_1 held constant and z_2 and z_3 altered by $\pm 30\%$.

are $(dT/dx)\Delta x$ for the downwind SST gradient and $-(dT/dy)\Delta y$ for the crosswind SST gradient. For given values of the SST gradient and the distances Δx and Δy , the amplitudes of the corresponding divergence and curl responses are determined by the values of the vector velocity difference $(u, v)_w - (u, v)_c$.

It follows that the wind divergence response to downwind SST gradients is identical in this model to the wind curl response to crosswind SST gradients. Because of the change in wind direction between the profiles, there is also a wind curl response to downwind SST gradients, and this is identical in the model to the wind divergence response to crosswind SST gradients, except for a reversal of sign that is determined by the right-handed definition of crosswind direction. Thus, the model produces two predictions: one for both the wind divergence response to downwind SST gradients and the wind curl response to crosswind SST gradients, and a second for both the wind curl response to downwind SST gradients and the (negative of the) wind divergence response to crosswind SST gradients. This model therefore

does not and cannot directly explain the apparent difference in the scale dependence of the observed and simulated divergence-downwind coupling relative to curl-crosswind coupling.

c. Log-Ekman boundary layer model: Results

An illustrative pair of solutions of the log-Ekman model (C3) can be obtained by increasing or decreasing the parameters z_1 and z_2 by, respectively, 15% and 5% relative to their undisturbed values $z_1 = 50 \text{ m}$, $z_2 = 100 \text{ m}$, with $z_3 = 700 \text{ m}$ held constant and an upper-level geostrophic wind of 10 m s^{-1} , corresponding to a 10-m wind of approximately 7 m s^{-1} (Fig. 15). The relative increase or decrease of the height parameters is consistent with, respectively, deeper momentum mixing under unstable (relatively warm SST) or shallower momentum mixing under stable (relatively cool SST) conditions. The upward stress difference in the 10-m wind direction is negative and nearly constant below 500 m, while the orthogonal stress difference is generally small. For this change in model height parameters, the magnitude of

the velocity difference is roughly $\pm 0.2 \text{ m s}^{-1}$, which is less than 5% of the 10 m s^{-1} geostrophic wind outside the boundary layer.

Profiles of effective wind divergence and curl responses versus SST gradient can be constructed from this pair of solutions by linearly interpolating between opposite-signed profiles obtained from the profile differences (Fig. 16). For example, the differences $u_w - u_c$ and $u_c - u_w$ represent, respectively, the divergence response to positive and negative downwind SST gradients, or, alternatively and again respectively, the curl response to positive and negative crosswind SST gradients. With these endpoints representing mean SST gradients of $\pm 10^{-5} \text{ } ^\circ\text{C m}^{-1}$, the corresponding wind speed differences (Fig. 15) would be found over distances $\Delta x = \Delta y = 100 \text{ km}$. The intermediate, interpolated values then represent the responses for intermediate SST gradients, with an assumed linear dependence on SST gradient. The resulting profiles for this pair of solutions reproduce much of the structure of the cross sections of ensemble-mean wind divergence and curl responses from the numerical simulations (Fig. 13). The velocity component difference $u_+ - u_-$ in the direction of the mean 10-m wind is positive in a surface layer roughly 100 m deep, negative between 100 and 500 m heights, and again positive between 500 and 800 m heights, and is essentially zero above 800 m. This difference represents the log-Ekman model prediction of both the wind divergence response to downwind SST gradient and the wind curl response to crosswind SST gradient. The orthogonal component difference $v_+ - v_-$, which represents either the wind divergence response to crosswind SST gradient or the negative of the wind curl response to downwind SST gradient, has a qualitatively similar pattern, but shifted upward in phase, so that it is positive in a surface layer roughly 200 m deep, negative between 200 and 700 m heights, positive again between 700 and 800 m heights, and is essentially zero above 800 m.

The log-Ekman model thus reproduces both the near-surface and the reversed midboundary layer responses found in the simulations. However, the third, uppermost lobe of response predicted by the log-Ekman model, which has the same sign as the near-surface response, is not seen in any of the mean coupled CCS model cross sections. Also, the near-surface response is not as strong, for the parameters chosen, as is observed; the parameters could have been chosen to match the near-surface response magnitude, but then would more dramatically overestimate the midlevel (as well as upper-level) boundary layer response. Both of these differences suggest that the linear dynamics described by the log-Ekman model are progressively poorer or more incomplete descriptions of the dynamics with height above the surface.

FIG. 18. As in Fig. 16, but for profiles of eddy viscosity in Fig. 17.

The structure of the profiles of effective coupling coefficients derived from the log-Ekman model is sensitive to the choice of perturbations of the eddy-viscosity profile height parameters z_j . If the lower height parameter z_1 is held constant, but the middle and upper parameters z_2 and z_3 are altered by $\pm 30\%$, there is essentially no surface signature of the wind divergence response to downwind SST gradient or the wind curl response to crosswind SST gradient, with the dominant structure in the resulting coupling-coefficients confined to upper boundary layer levels near 500 m (Figs. 17 and 18). If instead all of the height parameters are held constant, but the von Kármán coefficient κ in the surface layer is increased by 5% for unstable conditions and decreased by 5% for stable conditions, the sign of the near-surface wind response reverses, so that it is opposite to the observed response, and the depth of that near-surface response increases to encompass most of the boundary layer (Figs. 19 and 20). In this case, the stability correction is large enough that the stress still increases for unstable conditions, despite the decrease in near-surface winds, and still decreases for

FIG. 19. As in Fig. 15, but for z_1 , z_2 , z_3 held constant and the effective von Kármán coefficient modified by a $\pm 5\%$ surface-layer stability correction.

stable conditions, despite the increase in near-surface winds.

5. Discussion

The dependence of the CCS model coupling coefficients on PBL scheme may be directly compared to that obtained by Perlin et al. (2014) for the Agulhas Return Current region, for the corresponding schemes. The dependence on scheme is generally consistent in the two cases, with the same relative ordering of schemes based on coupling strength prevailing in both, for both stress and 10-m ENS winds (Fig. 21; see also Figs. 7–9 and Perlin et al. 2014, their Table 3). The variations in stress coupling coefficients between schemes for the ARC region are roughly twice as large as for the CCS region, while the variations in 10-m ENS wind coupling coefficients are only modestly larger for the ARC than for the CCS. Given the systematic differences in the model coupling coefficients for the ARC and the CCS, it is surprising that the observed, QuikSCAT-derived

coefficients are apparently similar (Fig. 21). The following discussion provides a plausible explanation for the differences in the model coefficients, but does not address this conflicting similarity of the observed coefficients, which is most apparent for the observed divergence coefficients.

There are two primary differences between the boundary layer structure and dynamics in the ARC and CCS regimes. The ARC boundary layer has a depth of order 1 km, more typical of open-ocean conditions and more than twice as deep as the 400-m CCS boundary layer, which is more typical of eastern boundary current regimes (Lima et al. 2018; Bridger et al. 1993). The deeper ARC boundary layer is apparent, for example, in profiles of PBL model eddy viscosity diagnosed by Perlin et al. (2014) [their Fig. 11], for which large values penetrate 2–3 times deeper than can be inferred for the CCS either from the typical 400-m boundary layer depth or from the log-Ekman viscosity profiles (Fig. 15). Mean low-level wind speeds were also at least twice as large in the ARC simulations of Perlin et al. (2014) as in the CCS

FIG. 20. As in Fig. 16, but for profiles of eddy viscosity in Fig. 19.

simulations, with mean 10-m winds typically $11\text{--}13\text{ m s}^{-1}$ for the ARC (e.g., [Perlin et al. 2014](#), their Fig. 9) compared with $5\text{--}7\text{ m s}^{-1}$ for the CCS simulations (Figs. 1 and 2).

While much larger eddy viscosity values and variances, and consequently much stronger vertical mixing, are associated with the deeper ARC boundary layers, relative to those for the CCS, this in itself does not seem to provide an explanation or hypothesis for the systematically larger coupling coefficients found for the ARC. The ARC boundary layer depths are large enough, by the scaling argument given above (section 1), that pressure gradient effects might be expected to supplement the vertical mixing effect, but [Perlin et al. \(2014\)](#) found that the coupling coefficient variations were well explained by variations in the vertical mixing, so that pressure gradient effects were likely secondary. The same scaling argument, and more detailed arguments by [Samelson et al. \(2006\)](#), imply that the amplitude of SST-induced stress variations is proportional to the fractional change in the vertical extent of momentum mixing, not the mean depth. It is not evident why a boundary layer with larger mean depth

should experience larger effective fractional changes in depth in response to a given SST gradient than a boundary layer with smaller mean depth; the opposite might even be anticipated, if the absolute change in depth is assumed to be controlled by the SST gradient.

On the other hand, the larger wind speeds in the ARC region do suggest a plausible explanation for the regional difference in coupling coefficients. The underlying physical mechanism that is believed to control the SST–stress coupling is the effect of stability modulations on the vertical penetration of surface-generated turbulence and the associated vertical mixing of momentum. These stability modulations presumably take the form either of mean changes in the local stability of the boundary layer that are driven by mean local heat fluxes, or of correlated variations in the instantaneous heat fluxes carried by the same turbulent eddies that carry the momentum fluxes. In the latter case, relatively large upward heat fluxes should drive deeper penetration of momentum fluxes, while relatively small upward heat fluxes should leave momentum fluxes trapped closer to the surface. In the former case, these same effects are felt indirectly, through control of the local mean stability by local heat flux anomalies. In either case, the stability effects are mediated by the local sensible heat flux anomalies arising from the SST gradients. The amplitude of these heat fluxes is known empirically to be proportional to the local mean wind speed. Consequently these fluxes thus can be anticipated to be substantially larger for a given air–sea temperature difference in the ARC region than in the CCS region. It follows that the corresponding depth penetration of momentum fluxes will have a larger variance in the ARC region than in the CCS region. This effect appears to be sufficiently large that the fractional change in momentum mixing depth and the resulting surface stress response to SST variations are larger for the ARC than the CCS, despite the deeper ARC mean boundary layer depth.

Evidence for a reversal, relative to the surface, of the response at midlevels in the boundary layer was found in the ARC simulations by [Perlin et al. \(2014\)](#), generally consistent with that found here in the CCS simulations. Consistent with the deeper boundary layer in the ARC simulations, the reversal tended to occur near 200-m height ([Perlin et al. 2014](#), their Figs. 9c and 10), rather than near 100-m height for the CCS (Fig. 13). This similarity is consistent also with the present explanation of this reversal structure as a consequence of SST-induced changes in vertical mixing of momentum, as described by the log-Ekman model, rather than by appeal to the pressure gradient mechanism.

FIG. 21. Scatterplots of coupling coefficients computed on the $0.25^\circ \times 0.25^\circ$ QuikSCAT grid for the CCS vs those reported by [Perlin et al. \(2014\)](#) for the ARC, for the corresponding WRF PBL schemes. (a) Divergence of stress vs downwind SST gradient ($\times$) and curl of stress vs crosswind SST gradient ($\circ$). (b) Divergence of 10-m ENS wind vs downwind SST gradient ($\times$) and curl of 10-m ENS wind vs crosswind SST gradient ($\circ$). The observed coupling coefficients computed from QuikSCAT observations for the two regions are also shown in each panel (divergence: small solid dot; curl: large solid dot). For (a) QuikSCAT stress and (b) model 10-m ENS winds, CCS results are shown for both constant- z_0 and COARE-3.0 neutral drag coefficients, with the constant- z_0 formulation giving relatively smaller stress and larger 10-m ENS wind coupling coefficients than the COARE-3.0 formulation. Units are (a) $10^{-2} \text{ N m}^{-2} \text{ } ^\circ\text{C}^{-1}$ and (b) $\text{m s}^{-1} \text{ } ^\circ\text{C}^{-1}$.

6. Summary

The coupling of surface stress and winds to local SST on the ocean mesoscale is a robust effect found throughout the world's oceans. However, quantifying this coupling and assessing the accuracy and adequacy of the representation of the relevant physics in ocean-atmosphere models remains a challenge. The present results suggest that the intensity of the observed coupling for the CCS is only partly reproduced by the suite of PBL schemes considered here, with none of the schemes matching the observed intensity of the stress-divergence coupling, and no single scheme emerging as optimal for matching all four of the primary coupling measures: divergence and curl of stress and 10-m ENS winds versus, respectively, downwind and crosswind SST gradients. A systematic reversal of the surface response at midlevels in the boundary layer is found in the simulations; in situ observations of boundary layer structure would seem necessary to determine whether these reversals indeed occur in the low-level atmosphere.

Analysis of the coefficients derived from the model simulations suggests that differences in the amplitudes of divergence coefficients relative to curl may arise from scale dependence of the processes, with the curl response extending to smaller spatial scales and thus increasing less rapidly with smoothing scale, while the

divergence response peaks at larger scales and increases more rapidly with smoothing scale. This evidence for scale dependence of the coupling complicates the analysis of and comparison with observations, and indicates that careful attention to the spatial smoothing and resolution of wind, wind stress, and SST fields used in calculations of both observed and model coupling coefficients is required. A further complication is that the relation between the 10-m ENS QuikSCAT wind and the corresponding geophysical stress field is not completely determined, leaving uncertainties as to whether the stress field should be computed using the most recent and accurate formulation for the neutral drag coefficient, or the formulation that was used, albeit indirectly, in the calibration of the QuikSCAT geophysical model function. In view of these complications, it is difficult to use the observed coupling coefficients as a fixed, absolute metric against which to measure the relative performance of the model PBL schemes. This finding apparently contrasts with that of [Renault et al. \(2016a\)](#), who argue that, in their U.S. West Coast simulations, the MYNN2 scheme gave “more realistic” wind features than the YSU or UW schemes.

The comparison with the log-Ekman model is consistent with the hypothesis that variations in vertical mixing of momentum – the depth over which the

momentum injected into the ocean is extracted from the atmosphere – arising from changes in stability, and not from a thermodynamically induced pressure gradient in the boundary layer, control the coupling of wind and stress to SST. This in turn implies that, in accordance with the results of [Perlin et al. \(2014\)](#), the differences in coupling coefficients obtained for the different PBL schemes arise primarily from differences in the intensity of the vertical momentum mixing that is parameterized by those schemes. The smaller amplitude of the SST-induced coupling found in the simulations for the CCS relative to the ARC likely arises indirectly from the weaker mean surface winds in the CCS, which reduce the amplitude of sensible heat flux anomalies for given SAT–SST variability, in turn reducing the stability fluctuations that drive the coupled response. However, the observed coefficients do not appear to show a difference in amplitude between the ARC and the CCS, leaving uncertainty as to whether this interpretation of the difference in model results is relevant to that comparison.

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APPENDIX A

WRF Vertical Grid and Physics Options

The vertical grid for the atmospheric component of the coupled numerical circulation model simulations consisted of 51 “terrain-following” pressure-based mass and horizontal velocity levels with 17 levels below a nominal height of 1500 m, lowest level at nominal 10-m height above the sea surface, and a resolution of 20 m below 100 m height above the sea surface ([Fig. A1](#)). One additional simulation was done with a 58-level grid, consisting of the 51-level grid augmented by 7 additional mass and horizontal velocity levels below 200 m, lowest level at nominal 5-m height above the sea surface, and a resolution of 10 m below 100 m height above the sea surface ([Fig. A1b](#)). The results from this additional, fine-resolution simulation, which used the GBM PBL scheme, were consistent with the three analyzed GBM PBL simulations using the original vertical grid, with no indication of systematic bias. For all simulations, the remaining WRF physics parameters were: $mp_physics = 10$, $ra_lw_physics = 4$, $ra_sw_physics = 4$,

$sf_surface_physics = 2$, $cu_physics = 1$, $sf_lake_physics = 0$, $sf_urban_physics = 0$, $sheu_physics = 0$.

APPENDIX B

Neutral Drag Coefficient

The COARE 3.0 neutral drag coefficient $C_{3.0}$ ([Fairall et al. 2003](#)) was computed using a polynomial approximation ([O’Neill et al. 2012](#)):

$$C_{3.0} = \frac{a}{U_{10N}} + b + cU_{10N}, \quad (B1)$$

where U_{10N} is the 10-m neutral stability wind speed and

$$(a, b, c) = (4.5, 5.7, 0.7) \times 10^{-2}. \quad (B2)$$

The constant roughness length $z_0 = 0.152 \times 10^{-3}$ m corresponds to a neutral drag coefficient $C_{z_0} = \kappa^2/\ln^2(z_{10}/z_0) = 1.3 \times 10^{-3}$, where $\kappa = 0.4$ is the von Kármán constant and $z_{10} = 10$ m. The roughness length z_0^{WRF} for the WRF $sf_sfclay_physics = 1$ scheme, which was used in all of the simulations except one MYJ case, is given by

$$z_0^{WRF} = \min\{z_0^*, 2.85 \times 10^{-3} \text{ m}\}, \quad (B3)$$

where

$$z_0^* = 0.185 \frac{u_*^2}{g} + 0.11 \times \frac{1.5 \times 10^{-5}}{u_*}, \quad (B4)$$

$u_* = (\tau/\rho_a)^{1/2}$ is the friction velocity, g is the acceleration of gravity, and ρ_a is the surface air density, taken here as $\rho_a = 1.22 \text{ kg m}^{-3}$; the corresponding neutral drag coefficient is $C_{WRF} = \kappa^2/\ln^2(z_{10}/z_0^{WRF})$. The stress τ is then related to the 10-m ENS wind speed by $\tau = \rho_a C_{DN} U_{10N}^2$, where $C_{DN} = \{C_{3.0}, C_{z_0}, C_{WRF}\}$ is the neutral drag coefficient for each case ([Fig. B1](#)). Note that $C_{WRF} \approx C_{3.0}$, so that under neutral conditions, the 10-m ENS wind derived for the coupling analysis from the model stress and the $C_{3.0}$ coefficient ([B1](#)) will be close to the model 10-m wind, but the full WRF drag coefficient includes stability corrections that will modify the relation between the model stress and the model 10-m wind when the stability is not neutral.

APPENDIX C

Log-Ekman Boundary Layer Model

A continuous, piecewise-linear eddy viscosity profile is constructed with the log-layer form near the sea surface ($0 < z < z_1$), constant viscosity A_{V1} in the midlevel

FIG. A1. Vertical model grid: mass and horizontal velocity levels for the primary suite of simulations (•) and for the test simulation with 7 additional levels below 200 m (□). (a) All levels (km), with the 1.5 km height indicated (dashed line). (b) Levels (m) below 1.5 km, with the 200-m height indicated (dashed line).

boundary layer above the surface layer ($z_1 \leq z < z_2$), a modified log-layer form with viscosity decreasing upward in the upper boundary layer ($z_2 \leq z < z_3$), and constant, small viscosity $A_{V3} \ll A_{V1}$ above ($z > z_3$):

$$A_V(z) = \begin{cases} \kappa u_*(z + z_0), & 0 < z < z_1 \\ A_{V1}, & z_1 \leq z < z_2 \\ A_{V1} - \kappa_2 u_*(z - z_2), & z_2 \leq z < z_3 \\ A_{V3}, & z \geq z_3 \end{cases} \quad (\text{C1})$$

where

$$A_{V1} = \kappa u_*(z_1 + z_0), \quad \kappa_2 = \frac{A_{V1} - A_{V3}}{u_*(z_3 - z_2)}. \quad (\text{C2})$$

The generalized Ekman-layer equation for the departure $\mathbf{V}'$ of the wind $\mathbf{V}$ from the geostrophic wind $\mathbf{V}_G$ is then

$$f\mathbf{k} \times \mathbf{V}' = \frac{d}{dz} \left[A_V(z) \frac{d\mathbf{V}'}{dz} \right], \quad (\text{C3})$$

where f is the constant Coriolis parameter, $\mathbf{k}$ is a vertical unit vector, and the boundary conditions are

$$\mathbf{V}' = -\mathbf{V}_G \quad \text{at} \quad z = 0, \quad \mathbf{V}' \rightarrow 0 \quad \text{as} \quad z \rightarrow \infty. \quad (\text{C4})$$

With the piecewise-linear profile (C1) for A_V , (C3) may be solved in terms of Hankel or modified Bessel

functions in the intervals $0 < z < z_1$ and $z_2 \leq z < z_3$ (Ellison 1956), and in terms of exponential functions—as for the classical Ekman layer—in the interval $z_1 \leq z < z_2$ and for $z \geq z_3$. The natural matching conditions are that velocity $\mathbf{V}'$ and (kinematic) stress $A_V d\mathbf{V}'/dz$ are continuous at z_1 , z_2 , and z_3 , which supplement the boundary conditions (C4) and are sufficient to close the problem. The fundamental velocity and time scales u_*

FIG. B1. Stress (N m^{-2}) vs 10-m wind (m s^{-1}) for the COARE 3.0 (thick), constant- z_0 (thin), WRF v3.9.1.1 sf_sfclay_physics = 1 (dashed), and Large and Pond (1981) (dotted) neutral drag coefficients.

and f^{-1} may to express the equations in dimensionless form, with dimensionless height $\zeta = z/D$, where $D = u_*/f$.

The effects of SST on the stability and turbulence of the boundary layer may be modeled by changing the parameters κ , z_1 , z_2 , and z_3 that determine the eddy viscosity profile (C1). Deeper penetration of turbulence, corresponding to relatively unstable conditions over warmer water, are modeled by increasing one or more of the heights z_1 , z_2 , and z_3 . Shallower penetration, and relatively stable conditions over cooler water, are modeled by decreasing one or more of these heights. Changes in the von Kármán constant κ correspond to stability corrections in the surface log layer, with larger κ representing relatively unstable and smaller κ relatively stable conditions. For small perturbations of the parameters relative to mean values, the differences between pairs of wind profiles obtained for perturbations of opposing signs give an effective amplitude of the response to SST fluctuations that can be compared with the model results.

For a given fluid, the roughness length and the geostrophic velocity are typically related to each other and to the stress at the surface, and consequently these three parameters are generally not independent. For the atmosphere, the relation is encoded in the neutral drag coefficient C_{DN} , which relates the 10-m neutral wind $U_{10N} = U(z = z_{10} = 10 \text{ m})$ to the kinematic stress:

$$u_*^2 = C_{DN} U_{10N}^2. \quad (\text{C5})$$

Under the further assumption that the wind profile is logarithmic from the surface through 10 m, that is, that $U(z)u_* = (1/\kappa)\ln[(z + z_0)/z_0]$, C_{DN} may be expressed in terms of the roughness length:

$$C_{DN} = \frac{\kappa^2}{\ln^2(z_{10}/z_0)} = \frac{\kappa^2}{\ln^2(\zeta_{10}/\zeta_0)}, \quad (\text{C6})$$

where z_0 and $\zeta_0 = z_0/D$ are the dimensional and dimensionless roughness lengths, respectively, $\zeta_{10} = z_{10}/D$, and it has been assumed that $z_0 \ll z_{10}$. The boundary conditions on the dimensionless velocity perturbation $\hat{\mathbf{V}}' = \mathbf{V}'/u_*$ are supplemented by the consistency condition:

$$\left| \kappa \zeta_0 \frac{d\hat{\mathbf{V}}'}{d\zeta} \right| = 1 \quad \text{at} \quad \zeta = 0, \quad (\text{C7})$$

which follows from the scaling of the unknown kinematic stress by the unknown friction velocity u_* .

For the calculations here, it is assumed that the roughness length is given by the Charnock relation for

the atmospheric roughness length over the sea surface:

$$z_0 = \frac{a_{\text{Ch}} u_*^2}{g}, \quad (\text{C8})$$

where $a_{\text{Ch}} \approx 0.02$ is a dimensionless constant and $g = 9.81 \text{ m s}^{-2}$ is the acceleration of gravity. The dimensionless roughness length ζ_0 is then a function of u_* :

$$\zeta_0 = \frac{a_{\text{Ch}} f}{g} u_*. \quad (\text{C9})$$

Consequently, the unknown friction velocity u_* can be obtained from ζ_0 :

$$u_* = \frac{g}{a_{\text{Ch}} f} \zeta_0. \quad (\text{C10})$$

For the standard computation of bulk fluxes from known 10-m winds, the Charnock relation would be used to solve iteratively for the roughness length z_0 and the friction velocity u_* . For the model, the dimensionless ζ_0 and the direction of the geostrophic velocity are given, and the equations are solved when the dimensionless geostrophic velocity magnitude $|\hat{\mathbf{V}}_G|$ is found that gives a dimensionless surface stress of unit magnitude. The corresponding dimensional geostrophic and 10-m velocities are then, respectively,

$$\{\mathbf{V}_G, \mathbf{V}_{10}\} = \frac{g \zeta_0}{a_{\text{Ch}} f} \{\hat{\mathbf{V}}_G, \hat{\mathbf{V}}(\zeta_{10})\}, \quad (\text{C11})$$

where the dimensionless 10-m height is

$$\zeta_{10} = \frac{a_{\text{Ch}} f^2}{g \zeta_0} z_{10}. \quad (\text{C12})$$

APPENDIX D

Acronyms

ARC	Agulhas Return Current
CCS	California Current System
CW	Crosswind gradient
DW	Downwind gradient
ENS	Equivalent neutral stability
GBM	WRF bl_pbl_physics=12 PBL scheme (Grenier–Bretherton–McCaa)
MYJ	WRF bl_pbl_physics=2 PBL scheme (Mellor–Yamada–Janjić)
MYNN2	WRF bl_pbl_physics=5 PBL scheme (Mellor–Yamada–Nakanishi–Niino-2)

PBL	(Atmospheric) planetary boundary layer
SAT	Surface (2-m height) air temperature
UW	WRF bl_pbl_physics=9 PBL scheme (University of Washington)
WRF	Weather Research and Forecasting mesoscale atmospheric model (Skamarock et al. 2008)
YSU	WRF bl_pbl_physics=1 PBL scheme (Yonsei University)

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