CHAPTER 10

Surface Heat Fluxes and Wind Remote Sensing

Mark A. Bourassa and Paul J. Hughes

Dept. of Earth, Ocean and Atmospheric Science and Center for Ocean-Atmosphere Prediction Studies, Florida State University, Tallahassee, FL, USA

The exchange of heat and momentum through the air-sea surface are critical aspects of ocean forcing and ocean modeling. Over most of the global oceans, there are few in situ observations that can be used to estimate these fluxes. This chapter provides background on the calculation and application of air-sea fluxes, as well as the use of remote sensing to calculate these fluxes. Wind variability makes a large contribution to variability in surface fluxes, and the remote sensing of winds is relatively mature compared to the air sea differences in temperature and humidity, which are the other key variables. Therefore, the remote sensing of wind is presented in greater detail. These details enable the reader to understand how the improper use of satellite winds can result in regional and seasonal biases in fluxes, and how to calculate fluxes in a manner that removes these biases. Examples are given of high-resolution applications of fluxes, which are used to indicate the strengths and weakness of satellite-based calculations of ocean surface fluxes.

Introduction

Satellite remote sensing works by receiving an electromagnetic signal that is dependent on the geophysical variable that is being measured. For example, ocean color is determined through measuring wavelengths of reflected light in the visible band; temperature is determined from measurements of emitted radiation, and wind speed can be determined through a wide variety of approaches including emissions at multiple bands or the fraction of radar energy that interacts with the surface and returns to the satellite. The geophysical variable determined in this manner is called a retrieval (e.g., wind speed retrievals). Passive remote sensing usually measures how much energy (over a band of wavelengths) is emitted, reflected, or scattered from an object (e.g., the ocean surface). Active satellites emit a signal and measure how much of this emitted signal is reflected or scattered from an object and returned to the satellite or a different satellite. For remote sensing to be effective, the geophysical variable that is retrieved must modify the measured electromagnetic signal. It is easy to imagine how wind modifies the ocean surface (e.g., water waves), making the remote sensing of wind an obvious candidate for observation from space. A better understanding of the geophysical processes that modify the surface can be used to develop better retrievals and to more carefully define the variable being measured.

Bourassa, M.A., and P.J. Hughes, 2018: Surface heat fluxes and wind remote sensing. In "*New Frontiers in Operational Oceanography*", E. Chassignet, A. Pascual, J. Tintoré, and J. Verron, Eds., GODAE OceanView, 245-270, doi:10.17125/gov2018.ch10.

There is a long history of efforts to use satellite observations to determine surface turbulent and radiative fluxes. Surface fluxes are commonly thought of as the transfer of something (e.g., momentum, heat, moisture, a gas, or particulates) from the ocean to the atmosphere or vice versa. For example, the latent heat flux is related to evaporation at the ocean surface, but it is also the vertical transport of this energy (stored in water vapor) in the atmospheric boundary-layer. Stress is the vertical transport of momentum at the air/sea interface, but there must also be an identical transport of momentum in the atmospheric and oceanic boundary-layers, unless that momentum is released or transported away by waves in the process of crossing the air/sea interface. Air-sea fluxes transfer momentum, heat, or material through the ocean surface and the layers near the surface, in the process modifying the surface and materials near the surface (e.g., modifying the motions, temperatures near the surface). A variety of approaches can be used to measure information at slightly different depths, and hence the changes caused by fluxes can be used to measure fluxes. Consequently, we realize that air/sea fluxes modify the air/sea interface and could plausibly be measured from satellite. Understanding the geophysical processes that modify the surface and thus modify the electromagnetic characteristics of the surface can be useful in developing more accurate retrievals and retrievals of new variables.

The above considerations are extremely useful for the developers of remote sensing retrievals, but are not relevant to most users of the retrieved geophysical data. That said, because the above considerations have been used to improve retrievals and quantify error characteristics, the descriptions of what is retrieved is often very carefully defined and often is not quite consistent with the assumptions of a casual user of the data. In some cases, the consequences of the errors associated with ignoring these definitions are small compared to other assumptions made in the application; however, in many cases they are not small and can result in a very misleading result. For example, satellite winds are 'equivalent neutral winds' (defined below, which differ slightly from traditional winds. If this slight difference is ignored, the average impact on mid-latitude fluxes is approximately a 10 Wm^{-2} bias, in addition to further regional and seasonal biases. It is very useful to know enough about the physical interpretation of retrievals to be able to assess if the assumptions made in using them will seriously alter the interpretation resulting from using the satellite data.

An introductory discussion of how air-sea coupling processes cause responses in the ocean and atmosphere provides relatively simple examples that show the complexity of coupling in the real world. These example illustrate problems that can be examined with uncoupled and coupled models. Satellite data can be used to assess the realism of the modeled near surface conditions, but a new generation of much finer resolution observations would be much more useful in determining the accuracy of models.

This chapter is designed to provide novices to remote sensing a very general idea of how remote sensing works, with a focus on surface wind and stress. It includes the basic physics and parameterizations of air/sea fluxes and how these fluxes modify the ocean surface in ways that can be used to measure air/sea fluxes. The goals are that readers will be able to better understand satellite remote sensing, gain an appreciation of the physics that impacts air/sea fluxes and satellite retrievals, and learn the types of questions that should be asked when applying any satellite retrieval.

Basics of Remote Sensing of Winds

Remote sensing relies on measuring electromagnetic radiation that is either emitted by the surface or emitted off the surface and interacts with the surface. If the emission or interaction of the electromagnetic radiation changes enough as a function of wind speed, then the observations can be used to determine wind speed. In addition, if the emissions or interactions are non-isotropic (directionally-dependent), then the observations can be used to determine the wind direction. The measurement of wind direction requires observations of the electromagnetic signal from several different directions. One further point is that if the signal being measured is non-isotropic, then the measurement of speed will be more accurate if the direction is determined. In other words, not knowing or measuring the direction can result in a greater uncertainty in the measurement of speed. These observations can be made more accurate by better accounting for how the electromagnetic radiation interacts with the atmosphere or by using wavelengths that are not heavily altered or modified by the atmosphere. Measurements of several wavelengths or polarizations can be used to learn a lot more about the atmosphere and have been shown to be effective in retrieving the near surface air temperature and humidity (Jackson et al., 2006, 2009; Jackson and Wick, 2010; Roberts et al., 2010; Bourassa et al., 2010; Smith et al., 2012), which are very important for determining surface fluxes of heat and moisture. The complexities of interaction with the atmosphere will not be discussed in this chapter, other than as warnings about when the atmospheric conditions cause trouble with the accuracy of remotely sensed variables.

Active remote sensing of surface winds

Radar is an example of active remote sensing. An electromagnetic signal, often microwaves, is generated and aimed at the surface. The signal that is reflected or scattered from the surface is measured and the ratio of the input to the output signal provides information about the surface. Radars use wavelengths that respond to the shape or roughness of the ocean surface. The term roughness is used for electromagnetic interactions and aerodynamic interactions; these do not mean the same thing. Electromagnetic waves interact with water waves of similar wavelengths to the electromagnetic wavelengths. For example, a radar wavelength of 6 cm moving roughly parallel to the surface would interact with surface waves of around 6 cm in wavelength. If the electromagnetic waves approach the surface at a greater angle, then they interact with shorter wavelengths. For example, scatterometers (described below) tend to operate at Ku-band interacting with longer ultragravity waves or at C-band interacting with very short gravity waves. These wavelengths are desired because these waves respond very quickly to changes in the surface winds. However, shorter radar wavelengths are more adversely impacted by the atmosphere and rain, making longer wavelengths, such as those used to measure ocean salinity, preferred for observations in rainy conditions.

Scatterometers and synthetic aperture radars

Scatterometers and synthetic aperture radars (SARs) are active sensors that use short microwaves (specifically Ka- to C-band) to measure the surface roughness. This roughness is a function of wind stress, which is closely related to wind speed (see the section on air-sea fluxes for details). The backscatter at these wavelengths has a directional dependence that allows direction to be retrieved if there are observations from multiple directions or the direction can be assumed from features in the data. Scatterometers retrieve direction by observing the surface from multiple directions as the satellite moves over the surface. This approach requires that the satellite move relative to the ocean surface. Therefore, we will never see a scatterometer in geosynchronous orbit. Similarly, SARs move relative to the surface and take observations from slightly different locations to effectively increase the size of the antenna and allow for much finer resolution observations. This resolution is used to identify bands of higher winds from rolls or other features to estimate the wind direction. SAR directions are less accurate than scatterometer directions and can, in both cases, have substantial biases in areas because of incorrect assumptions.

It is clear that scatterometer and SAR observations respond to stress rather than wind (Bourassa et al., 2010); however, surface wind datasets are produced. This is in part because there are far more wind observations than stress observations with which to calibrate the satellites, and in part because it is believed that operational weather centers will be able to more easily assimilate wind than surface stress.

The strengths of scatterometer and SAR observations are that they are vector observations (i.e., speed and direction), which is much more useful than scalar winds (i.e., speed only) for many ocean and atmospheric applications. Vector observations allow for the calculation of curl and divergence, and have much more dynamical impact during data assimilation in weather models. The vector observations also usually come from active systems, which can penetrate cloud cover and rain far better than passive observations. However, in both cases rain can cause serious complications in retrieving accurate surface winds. In the case of scatterometers, those that use Ku-band (e.g., QuikSCAT) are more sensitive to rain than C-band instruments (e.g., ASCAT), which have longer wavelengths relative to rain drop sizes. The problems are greatest when there are light winds and heavy rain, and negligible for most applications when there are high winds and light rain. Perhaps the greatest weakness of any single wind sensing satellite is the temporal sampling. There is a global average of two overpasses per day from a QuikSCAT-like instrument (1800 km swath width), with about 1.5 observations per day in the equator and four or more (in two clusters) near the poles where adjacent swaths overlap. An example of the coverage for one day is shown below. This limited coverage means that it is highly desirable to use data from a constellation of intercalibrated instruments. While intercalibration is very good for commonly occurring wind speeds (4–17 ms^{-1}), the two popular calibrations (from KNMI and Remote Sensing Systems) diverge outside this range (Verspeek et al., 2012; Chakraborty et al., 2013; Wentz et al., 2017; Holbach and Bourassa, 2017).

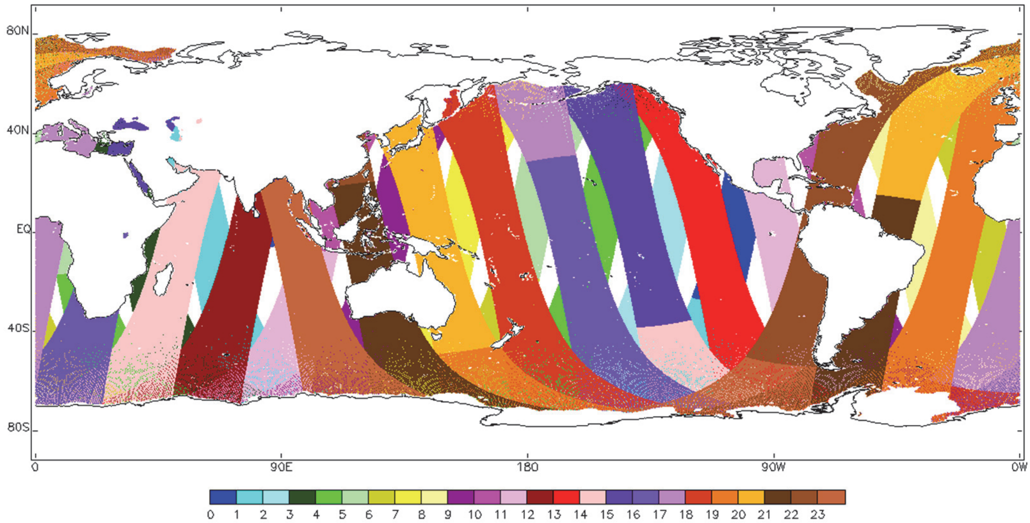

Figure 10.1. Example of daily coverage from QuikSCAT, where the color indicates the hour of the overpass. Any satellite in low earth orbit takes roughly 100 minutes to circle the Earth. The swath width and orbit determine the temporal sampling.

GPS reflectometry

Surface winds can also be estimated from the reflection of signals from the Global Positioning System (GPS). This approach is called GPS reflectometry (Zavorotny and Voronovich, 2000) and it requires a separate satellite to measure the signal emitted from the GPS satellite. The footprint from this technique is currently rather large, making it difficult to observe extreme features, which usually are much smaller than the footprint (e.g., fonts, Tropical Cyclone eyewalls, and near-shore features). NASA has recently developed the CYGNSS observing system (Ruf et al., 2013), which is a set of eight satellites flown in formation to improve the coverage. This observing system is focused on the tropics, providing improved sampling in time. However, the pattern in space is not well suited for spatial derivatives (e.g., calculation of curl and divergence, in addition to gradients of speed). The calibration of the CYGNSS system is an ongoing effort. It is expected to provide scalar wind speeds only, with much more sampling needed to provide vector observations, which require looks at the same surface location from difference azimuthal angles (angles relative to true north). One advantage of this system is that the GPS wavelengths are very long and interact very weakly with rain. Therefore, this system is expected to be useful for activities (operational and research) involving tropical cyclones.

Passive remote sensing (radiometry) of surface winds

Passive systems for observing surface winds measure the brightness temperature of the ocean surface. This is the temperature that the ocean would have if the observed radiance came from a black body. The ocean is not a black body, therefore this temperature is less than the ocean's surface temperature (Petty, 2006). This radiance is a function of temperature and surface wind speed (technically stress – see the section on air-sea fluxes) to understand this distinction) with greater radiance coming from whitecaps (Uhlhorn and Black, 2003; Petty, 2006; Paget et al., 2015). As wind stress increases there is greater breaking and a larger fraction of the ocean surface is covered by whitecaps. Combined with an estimate of the ocean temperature, the excess emissivity (above that expected from the temperature and a flat water surface) can be determined (Uhlhorn and Black, 2003; Petty, 2006). This temperature estimate is usually made from the same satellite used to estimate wind speed. There is a wind direction dependence in these observations, which is usually not considered. However, the WindSat mission uses polaromitry (Wentz, 1983; Gordon and Wang, 1994; Yueh et al., 1995; Wentz, 1997; Smith, 1998; Gordon and Voss, 1999; Rose et al., 2002; Gaiser et al., 2004; Anguelova and Webster, 2006; Anguelova and Gaiser, 2012, 2013) to measure this directional dependence and determine wind direction. In general, radiometers can accurately measure surface wind speeds >3 ms^{-1}; however, those designed to measure hurricane winds are typically not effective for wind speeds <8 ms^{-1}. WindSat can usefully determine directions for speeds >8 ms^{-1}. Radiometry is much more sensitive to rain than scatterometry and SAR, and usually does not provide wind data coincident with rain, whereas scatterometry (particularly C-band) and SAR can work in rain rates up to a wind speed dependent threshold (Draper and Long, 2004).

Air/Sea Fluxes

Air/sea fluxes refer to the rate of transfer, per unit area, through the air/sea interface and the ocean and atmosphere near this interface. Physicists refer to a transfer that includes the units of ‘per unit area’ as flux densities, but meteorologists and oceanographers drop the word ‘density’ and simply say ‘fluxes.’ Similar vertical rates of turbulent transfer exist in the atmospheric boundary-layer. Stress in the near-surface ocean is slightly reduced because some stress is lost to surface water waves. Heat fluxes are radiative fluxes and turbulent heat fluxes (sensible and latent heat). Radiative fluxes are similar in the lower atmosphere, but change rapidly with depth in the ocean. Radiative fluxes refer to the rate of transfer of energy per unit area in the form of electromagnet radiation, rather than the much more scary things that people normally associate with the word radiation. Fluxes are associated with motion and the net transport of something like heat and momentum. Most people think of conduction (heat transferred through the motion of atoms and molecules) as a good method for transporting energy, but that is an extremely poor assumption for the atmosphere. For example, Styrofoam coolers work very well not because of the foam, but because air is trapped in the Styrofoam and air is a terrible conductor. Similarly, Brownian motion is a very slow way of transporting energy compared to turbulent transport and radiative transport. Turbulence, where it exists, is much more efficient than conduction and Brownian motion.

In simple and somewhat recursive terms, turbulence transport is a transport due to turbulent motion. This motion comes about from vertical shear in horizontal motion and from unstable stratification. In contrast, stable stratification inhibits turbulent motion. Turbulent motion can be thought of as due to local changes of the speed of air or water associated with friction on a boundary and buoyancy. For example, any broad surface that has friction will have slower wind speeds near this surface and larger wind speeds away from the surface. These wind speeds increase logarithmically with height in the atmospheric boundary-layer. Increases in mechanical mixing tend to reduce the impacts of buoyancy, and increased buoyancy (either upward or downward) tends to reduce vertical shear except very near a boundary (e.g., the ocean surface). Turbulent fluxes (described in detail below) include stress (the flux of momentum), sensible heat, latent heat, and mass fluxes such as moisture or gas fluxes.

Turbulent fluxes are proportional to the covariance of vertical perturbations in vertical winds and perturbations in the quantity being transported. The example in Fig. 10.2 shows how a vertical perturbation of an air parcel is associated with a vertical motion (*w*'), and that when that air parcel moves up (positive *w*') it is slower than the surrounding air (a negative perturbation in momentum, *u*'). If the air parcel moves down it has negative *w*' and a positive *u*'. Except for the very rare cases when the current is moving faster than the air above, a wind profile will always have this shape. Consequently, the covariance of u' and w' is always negative. Hence the stress is proportional to $-\overline{u'w'}$.

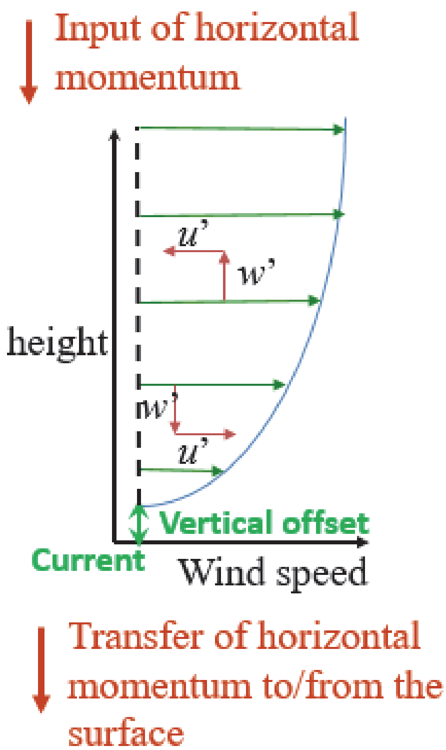

Figure 10.2. Log-wind profile with vertical perturbations in the wind (w') and the corresponding perturbation in horizontal velocity, which is proportional to momentum.

Stress

If the boundary layer has neutral stratification, meaning that force related to buoyancy is not contributing to turbulence, then stress is a shear stress, which is due to a change in velocity with height (Fig. 10.2). Near the surface, in the part of the atmospheric boundary layer known as the log layer (because the variables have a logarithmic function of height above the surface), all fluxes are independent of height within that layer. Very near the surface, part of the momentum flux can change to a horizontal pressure on the waves but the net momentum flux remains constant.

The top of the ocean's mixed layer feels this same stress as a 'surface stress,' reduced by any net rate of loss of momentum into surface waves. If the waves are in balance with the wind (i.e., the time and space averaged distributions of wave characteristics are not changing thus waves are not growing higher to reach equilibrium with wind that is stronger than needed for equilibrium, and waves are becoming shorter and longer because the wind has reduced in speed), the stress experienced by the ocean is a few percent less than the stress in the atmosphere because some of the momentum from the waves maintains the current. Stress is particularly important for oceanography because it can be closely related to wind-driven currents and vertical motions in the upper ocean (Ekman transport; Knauss, 2005) as well as horizontal motions in the deep ocean (Sverdrup transport; Sverdrup, 1947). Stress is also related to the generation of eddy kinetic energy, which is the dot product of surface stress and surface current.

Stress (τ) can be defined in terms of perturbations as $-\rho\,\overline{u'w'}$, where ρ is the density of the fluid. This is a great definition when measuring fluxes, but is not very useful for modeling stress in atmospheric and ocean models because tenth of a second changes in winds must be resolved to determine stress this way. In such models it is more useful to define stress in terms of friction velocity ($\mathbf{u}_*$, Eq. 1; where bold text indicates a vector, and non-bold indicates a scalar), which is closely related to stress (Eq. 1) and a scaling parameter in the log-profile for wind speed (Eq. 2).

$$\tau = \rho\,|\mathbf{u}_*|\mathbf{u}_* \tag{1}$$

$$\overline{\mathbf{u}}(z) - \overline{\mathbf{u}}_{\mathrm{sfc}} = \frac{\mathbf{u}_*}{k_v}\left[\log\left(\frac{z-d}{z_o}+1\right) - \varphi_M\left(z, z_o, L\right)\right] \tag{2}$$

In Eq. 2, $\overline{\mathbf{u}}$ is the mean horizontal wind speed at a height z above the displacement height d. $\overline{\mathbf{u}}_{\mathrm{sfc}}$ is the mean ocean surface current, κ is von Kármán's constant, z_o is the momentum roughness length, φ_M is an atmospheric stability term (Stull 1988) for momentum, L is the Moniv-Obukhov scale length (Monin and Obukhov, 1954; Stull, 1988), and $\overline{\mathbf{u}}$ - $\overline{\mathbf{u}}_{\mathrm{sfc}}$ is parallel $\boldsymbol{u}_*$ and stress. The displacement height (d) is a vertical offset in the log-profile, and it is the height at which the left side of the equation extrapolates to zero. As implied by the word 'extrapolate,' the log profile does not apply near the displacement height. For example, over crops the displacement height is roughly 79% the height of crops. Waves have a height-to-spacing ratio that is much less than crops, suggesting that d is much less than 70% of wave height. In fact, d is almost always assumed to be zero over water. The impact of considering displacement height has been tested in a wind/wave tank (Bourassa et al., 1999), where the stress estimated using a log profile technique was compared to a stress determined from eddy covariance. The impact of displacement height on open ocean fluxes is small when a displacement height of 80% of the wind driven significant wave height was found to optimize the fit to observed fluxes (Bourassa, 2006) for wind waves. A student project later revealed that a displacement height of zero works well for swell. The Monin-Obukhov scale length is a related to atmospheric stratification, and is also dependent on the shear stress. The friction velocity can be obtained from Eq. **Error! Bookmark not defined.**, given L and a relation between roughness length and friction velocity, such as Charnock's relation (Stull, 1988). Typically, friction velocity is calculated in the atmosphere, so the density is that of air rather than water. Using the density of water rather than that of air results in roughly a factor of 800 in stress.

Solving for friction velocity and stress as a function of wind speed is a slow iterative process. Therefore, many ocean and atmosphere models define stress in terms of a drag coefficient (C_D).

$$\tau = \rho\, C_D\, |\bar{\mathbf{u}} - \bar{\mathbf{u}}_{sfc}|(\bar{\mathbf{u}} - \bar{\mathbf{u}}_{sfc}) \tag{3}$$

The drag coefficient can be parameterized in terms of z_o and φ, assuming the displacement height is negligible, or the impact of displacement height can be included empirically while fitting the drag coefficient to observations, analogous to what is done with the COARE flux model (Fairall et al., 2003).

$$C_D(z) = \left[\frac{1}{k_v}\left[\ln\left(\frac{z-d}{z_o}+1\right) - \varphi_M\left(z, z_o, L\right)\right]\right]^{-2} \tag{4}$$

This combination of Eqs. (2) and (3) clearly shows that specifying a drag coefficient is analogous to specifying a roughness length. The roughness length is a combination of roughness lengths for multiple types of surfaces (Smith, 1988; Bourassa et al., 1999; Zheng et al., 2013): a smooth surface (Nikuradse, 1933; Kondo, 1975), capillary waves (Wu, 1968, 1994; Bourassa et al., 1999; Zheng et al., 2013) and gravity waves (Wu, 1980; Smith, 1988; Dobson et al., 1994; Taylor and Yellend, 1999; Drennan et al., 2005; Bourassa, 2006). The fitting parameters in the gravity wave roughness length parameterizations are highly sensitive to the surface current and displacement height (Bourassa, 2006). In all cases, the parameterizations for stress are tuned to observations. If stability is considered (and it should be), then the parameterizations become complicated. If sea state is known, it is almost always a good idea to use a flux parameterization that considers sea state. Some models greatly speed up the time it takes to calculate fluxes by using polynomial curve fits to stability dependent parameterizations (Kara, 2000).

Current topics in stress parameterization include the dependency on sea state and the atmospheric response to winds flowing over sea surface temperature (SST) gradients. Another open question is the importance of the atmospheric response to changes in the ocean. There is no question that the atmosphere responds to changes in the ocean, but the outstanding questions are *how much does it respond?* and *what are the key physical processes?* Similar questions are asked about the response of the ocean to changes in the atmosphere. These are important questions because they could have large impacts on the importance of two-way coupling between the ocean and atmosphere, which could add a great deal of computer processing requirements relative to the common practice of assuming that these feedbacks are small.

The ocean's mixed layer has a very thin log layer compared to the atmosphere. Ekman transport cause ocean responses to changes in the surface stress. Changes in the thermal forcing (heat budget) also cause an ocean response. If the winds are light and there is no wave breaking, the current profile is explained in terms of Stokes drift. However, if breaking occurs there should be more mixing at the surface and a log layer should thicken as the breaking increases. Breaking greatly modifies the transfer of energy and some gasses (Gulev et al., 2010; and references therein).

Sensible heat

The sensible heat flux (*H*) is the rate at which thermal energy (associated with heating, but without a phase change) is transferred from the ocean to the atmosphere. In the tropics, the latent heat flux is typically an order of magnitude greater than the sensible heat flux; however, in the polar regions the *H* can dominate. The quantity used to describe the storage of thermal and potential energy is the potential temperature. The sensible heat flux can be measured as proportional to the covariance of vertical velocity and potential temperature (Θ). Hence the sensible heat flux is proportional to $\overline{\theta' w'}$.

Sensible heat flux (*H*) can be defined in terms of perturbations as $-\rho\, C_p\, \overline{\theta' w'}$, where C_p is the heat capacity of air. In such models it is more useful to define sensible heat in terms of the magnitude of friction velocity (u_*, Eq. 1), an analogous term in the log-temperature profile (θ_*, Eq. 6).

$$H = \rho\, C_p\, |\mathbf{u}_*|\theta_* \tag{5}$$

$$\overline{\theta}_{\mathrm{sfc}} - \overline{\theta}(z) = \frac{\theta_*}{k_v}\left[\ln\left(\frac{z}{z_{o\theta}} + 1\right) - \varphi_\theta\left(z, z_{o\theta}, L\right)\right] \tag{6}$$

In Eq. 6, the displacement height for potential temperature is believed to be zero, however, it could easily be argued that evaporation from sea spray would change the temperature profile and make this term non-zero. A non-zero value for the displacement height for momentum (Eq. 2) suggests that a non-zero displacement height for potential temperature is quite plausible. Such considerations would be difficult to separate from adjustments due to stability. The stability term for temperature and moisture (φ_θ) is different than the stability term for momentum. Since *H* is an atmospheric flux, stability is calculated based on the atmospheric profile, rather than the oceanographic profile. The sensible heat flux should be non-zero in the near surface ocean, provided the net surface heat flux is non-zero (which implies that there is a temperature difference between the ocean and atmosphere) and wave breaking is causing turbulent motion. Turbulence will be damped out rapidly with increasing depth, meaning that other forms of energy transport will be more important away from the surface. Consequently, sensible heat flux almost always refers to the atmospheric side of the interface, with the understanding that the net heat flux is maintained on the ocean side of the interface.

Many models define *H* in terms of a transfer coefficient (C_H).

$$H = \rho\ C_p C_H\, |\overline{\mathbf{u}} - \overline{\mathbf{u}}_{sfc}|(\bar{\theta}_{sfc} - \bar{\theta}) \tag{7}$$

The merging of Eqs. (5) and (7) shows that C_H is equal to the square root of the drag coefficient times and similarly structured term with velocities replaced by potential temperature. The sensible heat flux should be sensitive to changes in drag, but it is rarely parameterized as such. Another issue with both sensible and latent heat fluxes is that organized eddies, such as atmospheric rolls, arguably act to increase fluxes beyond the expectations of Eq. (7). Similarly, at low enough wind speeds (how low is debatable) these parameterizations (Eqs., 1, 3, 5, and 7) underestimate fluxes because they assume that the vast majority of horizontal transport is due to the mean flow rather than eddies. Both of these issues have been attempted to be accounted for by slightly increasing the wind shear (Fairall et al., 1996). However, if this approach is used, the wind, temperature, and humidity cannot be correctly adjusted to different heights. For example, a temperature measured at a height of 4 m cannot be correctly adjusted to 10 m. If this type of fix is not used, then something else must be done to increase fluxes at low wind speeds. It has also been argued that roughness and turbulence from capillary waves partially account for this underestimation at low wind speeds (Bourassa et al., 1999; Zheng et al., 2013). Both approaches have a similar impact on the fluxes, but the inclusion of roughness from capillary waves does not cause problems with height adjustments. Such parameterizations work very well with the temperature and moisture roughness length parameterizations developed by Clayson et al. (1996).

Latent heat and evaporation

The evaporative moisture flux is the rate, per unit area, at which moisture is transferred from the ocean to the air. The latent heat flux (E) is related to the moisture flux (L); it is the rate (per unit area) at which energy associated with the phase change of water is transferred from the ocean to the atmosphere or the rate at which this energy is vertically transported within the atmosphere. The latent heat flux can be measured as proportional to the covariance of vertical velocity and specific humidity (q): $\overline{q'w'}$.

The latent heat flux can be defined in terms of perturbations as $-\rho L_v \overline{q'w'}$, where L_v is the latent heat of vaporization for water, and C_E is the transfer coefficient analogous to the drag coefficient. In such models it is more useful to define latent heat flux (Eq. 8) in terms of friction velocity (u_*, Eq. 1), and an analogous term in the log-humidity profile (Eq. 9):

$$Q = \rho L_v |\mathbf{u}_*| q_* \tag{8}$$

$$\bar{q}_{\mathrm{sfc}} - \bar{q}(z) = \frac{q_*}{k_v}\left[\ln\left(\frac{z}{z_{oq}}+1\right) - \varphi_\theta\left(z, z_{oq}, L\right)\right] \tag{9}$$

The stability term for temperature and moisture (φ_θ) is identical to the stability term for temperature. There is no latent heat flux in the ocean, so the density is that of the air. Many models define E in terms of a transfer coefficient (C_E).

$$E = \rho L_v C_E |\bar{\mathbf{u}} - \bar{\mathbf{u}}_{sfc}|(\bar{q}_{sfc} - \bar{q}) \tag{10}$$

The merging of Eqs. (8) and (10) shows that C_E is equal to the square root of the drag coefficient times and similarly structured term with velocities replaced by potential temperature. Evaporation (E) is the flux of moisture (not including precipitation).

It is almost identical to Q except that the water vapor is not multiplied by L_v.

$$L = E/\, L_v = \rho\, C_E\, |\bar{\mathbf{u}} - \bar{\mathbf{u}}_{sfc}|(\bar{q}_{sfc} - \bar{q}) \tag{11}$$

Radiative fluxes

Radiative fluxes are normally subdivided as either solar or terrestrial, and further subdivided as upward or downward. Upward and downward simply refers to the direction of propagation. Solar usually refers to near infrared and more energetic frequencies (i.e., larger frequencies and shorter wavelengths). These are energies that are rarely generated in a natural terrestrial environment. Therefore, upward terrestrial radiation is from reflected (or scattered) downward solar radiation rather than emitted from something in the terrestrial environment. For example, a red shirt looks red because it is relatively efficient in reflecting red light. The shirt does not have the temperature of a hot burner required to emit red light! Terrestrial radiation is generated by any body at terrestrial temperatures, and the rate of output of terrestrial radiation is closely tied to the temperature of the emitter.

Fig. 10.3 shows one of the two key concepts in both the diel (daily) cycle of light and seasonal changes in light. The other key concept is that the length of daytime changes as a function of season and latitude. For the diel cycle, the rotation of the earth causes the shift in the angle $\boldsymbol{\theta}$**.** For the seasonal cycle, the change in tilt is caused by the tilt of the Earth relative to the sun and the location of the Earth in its orbit about the sun. High latitude summers can have a daily input of solar radiation that is similar to tropical input because of the long length of a summer day.

Fig. 10.3 illustrates the importance of the angle of the surface relative to the incoming sunlight. The amount of incoming solar radiation depends on the distance from the Sun (Earth's orbit is not very elliptical, so this doesn't change much), the angle of the surface relative to the sun (which is a function of latitude, season and the time of day), and the amount of absorbing, reflecting and scattering materials in the atmosphere. The solar radiation that makes it through the atmosphere without interaction is call direct solar radiation, and reaches the surface as indicated in Fig. 10.3. Indirect light was either scattered (e.g., the blue light from the sky) or reflected from a cloud. The amount of solar radiation directly reaching the ground is easy to calculate if the mass and type of atmospheric absorbers are known, and if the number and type of scatters are known. The indirect solar radiation is much harder to model.

The absorption and emission of terrestrial radiation can be an effective method of transferring heat. In forest fires, enough heat can be transferred through radiation to start fires in neighboring trees untouched by flame. Temperatures in the lower atmosphere and ocean are much less extreme (excepting volcanoes), nevertheless radiative transfer can be an effective mechanism for transferring heat. This mechanism should be much more effective than diffusion in the ocean, and is perhaps more important than conduction in areas with weak thermal gradients.

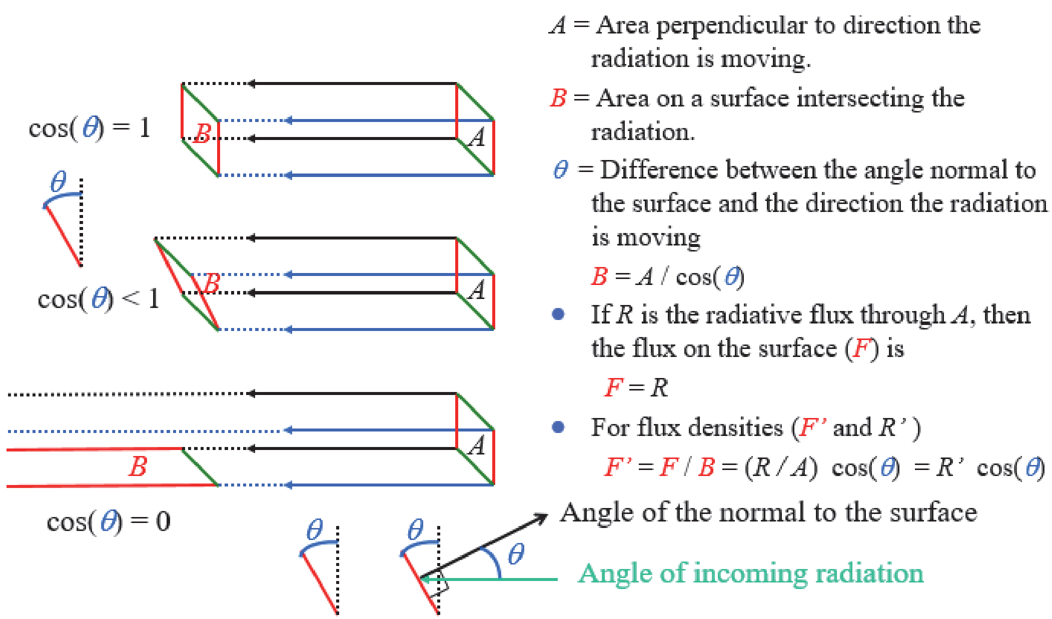

Figure 10.3. The flux of solar energy passing through the area A, which is perpendicular to light emitted from the sun, is constant in all three examples. This light strikes an ocean surface B that is tilted relative to A. In the top example, the tilt of A is zero relative to B, and the flux through A is equal to the flux through B. For the middle and bottom cases, the tilt is not zero, causing the energy passing through B to be spread over more area. The flux density through B is less than the flux density through A, even though the same amount of energy is passing through both areas. In the bottom case, the area of B is infinite, making the flux density through B equal to zero.

Terrestrial radiation is a function of the temperature and the emissivity (the fraction of energy emitted relative to a perfect emitter). The amount emitted is related to the surface area or mass of emitters. The atmosphere is a good absorber of most emitted terrestrial radiation, and the ocean is an excellent absorber. Emitted terrestrial radiation is usually absorbed and reemitted in all directions, contributing to the redistribution of heat in the earth system. The atmosphere also emits terrestrial radiation, and this contributes to heating the ocean's surface. Since the atmosphere is usually colder than the ocean, the terrestrial radiation emitted from the ocean tends to be larger than the input radiation. Cloud cover will increase the downward terrestrial radiation and block some of the solar radiation. In polar winter environments, most of the input radiation is terrestrial radiation emitted from clouds in the lower 200 m (Bourassa et al., 2013).

In the atmosphere, emissions at some frequencies can travel through the cloud-free atmosphere, and are very useful for remote sensing of the surface. Microwaves can travel through the atmosphere including clouds making them very effective for satellite observations of the Earth's surface. Longer wavelength microwaves can see through rain, whereas the shorter end of the microwave spectrum is largely blocked by rain. Therefore, microwaves are very effective for satellite observations of surface processes in the sense that they are rarely blocked by the atmosphere, with longer wavelengths being more effective. Unfortunately, the resolution of satellite observations is a function of wavelength, with finer resolution being much easier (i.e., less costly) to achieve with shorter wavelengths. This tradeoff between being able to see through rain and resolve smaller ocean features is a concern in the development of satellite missions.

Net fluxes

The net heat flux is critical to long-term energy budgets and to the diurnal cycle. The atmospheric net heat flux is equal to the net radiative fluxes (solar and terrestrial, downward minus upward) plus the sensible heat flux and the latent heat flux. The dominant mechanisms for transporting thermal energy in the ocean are different than those in the atmosphere, but in conditions of equilibrium the fluxes in the atmosphere must equal the fluxes in the ocean. While the whole integrated Earth system is in near-equilibrium when averaged over decades, equilibrium is very rare on regional and local scales. For example, more solar radiation is absorbed in the tropical oceans than at higher latitudes, resulting in a net transfer of energy from the tropics to the poles. The change in radiative forcing due to the seasonal cycle, and changes due to the diurnal cycle also change radiative forcing and cause non-equilibrium on regional, seasonal and sub-daily scales. In fact, modeling the Earth System assuming radiative equilibrium results in unrealistically extreme summers. Weather plays a large role in reducing the equator to pole imbalance of energy, but changes in weather cause large local imbalances of the energy budget.

Consider the diurnal cycle in ocean surface temperature, because this temperature is important to terrestrial radiative fluxes and turbulent heat fluxes. The diurnal cycle of solar radiative energy causes a large change in the solar radiative flux, most of which is absorbed in the top meter of the ocean. At night, the surface water temperature deceases, increasing the density of the water, and causing warmer sub-surface water to replace the surface water. This process greatly reduces the nighttime change in surface temperature. The daytime warming is more complicated. The stronger solar radiation heats the surface water, making it less dense, and inhibits downward mixing. At wind speeds above three or four m s^{-1}, the cooling due to the latent heat flux is sufficient to prevent the build-up of enough energy near the ocean surface, hence prevent increases in temperature of more than a few tenths of a degree Celcius (C°). At wind speeds <3 ms^{-1}, the latent heat flux and ocean mixing are too weak to prevent much larger increases in temperature (Weihs and Bourassa, 2016), which can easily exceed 0.6 C°. This heating is eventually disrupted by increased convection and by the duration of strong solar heating. This diurnal heating process is estimated to increase the neat heat flux by 10Wm-2 in the tropics and sub-tropics (Weihs and Bourassa, 2016). Modeling and explaining this change in temperature is a subject of ongoing research, and is clearly an important consideration in climate modeling but is a much less important consideration for short-term ocean and atmosphere forecasts.

Measurements of Stress and Example Practical Applications

Stress can be measured by buoys and research vessels simply by measuring horizontal and vertical winds frequently enough (roughly 10 Hz) to accurately determine the stress from the covariance of vertical and horizontal winds. However, because the spatial sampling is very poor for fluxes determined this way, these observations are not used to force ocean models. These observations are well suited for determining transfer coefficients (C_D, C_H and C_E) and roughness length parameterizations. The resulting flux models can then be used to determine fluxes from more typical ocean observations of wind speed, temperature, humidity and pressure which are much more commonly measured. Pressure is used to determine the air density and some humidity characteristics. The importance of wind speed, surface current, temperature and humidity are clear from Eqs. 3, 7 and 10. These equations are also dependent on the density of air, which is a function of pressure, temperature, and humidity. These transfer coefficients are also used in atmospheric models, which lack the 10 Hz temporal resolution needed to calculate fluxes through covariances, but do model the above list of flux-related variables. However, in situ based flux-related observations have insufficient space–time sampling to resolve fluxes needed to determine mesoscale fluxes without biases, and hence to determine basin-scale and global fluxes without biases without biases. For short-term weather forecasts, these biases are not a serious problem. For forecasts on monthly scales and longer, these biases contribute to model errors that must be someone compensated. Satellite observations and data assimilative numerical weather predictions can be used to reduce these biases, but will not be sufficient solutions unless the sufficiently small scales can be resolved. 'How small?' is a key question that is yet to be resolved, however, it appears that from the perspective of surface fluxes and coupled air-sea processes that a 2 or 3 km-scale must be resolved (Shi, 2017). A constellation of satellites in addition to in situ observations would be required to meet temporal sampling requirements (at least hourly sampling) over the global oceans. The daily coverage by a satellite (e.g., Fig. 1) is vast, particularly when it is realized that winds are separated by 12.5 km within each 1800 km-wide swath. However, a single satellite in low Earth orbit takes roughly 100 minutes to orbit the globe and does not come close to providing global coverage in that orbit. This explains the need for a constellation of satellites. The in situ observations are needed to test satellite calibration and to examine processes (including the subsurface) that are not well observed from satellite. In situ platforms are capable of making many more types of observations than would be available from satellites focused on surface fluxes and related variables.

Determining stress from satellite observations

Surface stress is particularly easy to determine from satellites because satellite winds are not quite winds but are equivalent neutral winds (Eq. 12; Kara et al. 2008). Such winds, when squared and multiplied by a neutral drag coefficient and an air density, are transformed into a stability-dependent stress (Eq. 13; Bourassa et al., 2010). This roughness is a function of the stress in a manner that is consistent with boundary layer stratification (Kara et al., 2008). Equivalent neutral winds were chosen as the calibration standard for satellite winds because (1) the satellites respond to ocean surface roughness rather than the wind at a 10-m height, (2) there are much more wind observations that stress observations to use for calibration, and (3) atmospheric models were already capable of assimilating surface winds.

$$\overline{\mathbf{u}}_{10EN} = \overline{\mathbf{u}}_{\text{sfc}} + \frac{\mathbf{u}_*}{k_v}\log\left(\frac{10}{z_o}+1\right) \tag{12}$$

$$\tau = \rho\, C_{D10N}\, |\overline{\mathbf{u}}_{10EN}|\, \overline{\mathbf{u}}_{10EN} \tag{13}$$

Eq. 12 uses a friction velocity that is consistent with a stability-dependent stress. The great advantage of this approach is that it uses a neutral drag coefficient (one that is determined for neutral buoyancy, which means the boundary layer stability term in Eq. 4 is set to zero). This approach removes the need to know the temperature and humidity, which are often not available from observations and are relatively poorly modeled. The above neutral drag coefficient (C_{D10N}) calibrated to a height of 10 m.

While the flux-related variables listed above must be known to calibrate the satellite, it has long been argued that they do not need to be known to use the calibrated satellite products. However, satellite retrievals of stress have been shown to depend on air density (Bourassa et al., 2010), which was considered to be a minor error when the calibration approach was developed. Density-related errors are smaller than the accuracy requirements in the original mission planning. A stress-equivalent wind product, that accounts for this density dependence, is now produced by KNMI (De Kloe et al., 2017). It is clear that satellite winds are relative to the moving surface (Kelly et al., 2001; Cornillion and Park, 2001; Plagge et al., 2012), dependent on density (Bourassa et al., 2010), and dependent on boundary layer stability (personal communication, Jim Edson), all in the manner we expect for stress. Therefore, it is apparent that remotely-sensed winds are more directly sensitive to wind stress than wind at an arbitrary height. While this finding creates some difficulties for determining surface winds, it is excellent for determining surface stress from satellites.

The concept of a neutral equivalent wind works well because surface characteristics are relatively sensitive to wind speed, making an equivalent neutral wind similar to a wind speed, with most differences <0.5 ms^{-1}. It has proven much more challenging to provide similarly good calibrations for heat fluxes, with several groups publishing remarkable improvements in the last decade. Very recently calibrations for air/sea differences in temperature and humidity have become quite good, with random errors about 1.5 times as large as buoys, and about 25 km sampling from satellites (Jackson and Wick, 2010; Roberts et al., 2010; Bourassa et al., 2010; Smith et al., 2012). These techniques use satellite observations of air-sea temperature difference and air-sea humidity differences, applied in Eqs. 7 and 10.

Heat fluxes determined from equivalent neutral winds

Since satellite winds are referred to as ‘winds’ rather than equivalent neutral winds, there is an unsurprising tendency for satellite winds to be used in bulk formulas (Eqs. 7 and 10) to calculate sensible and latent heat fluxes. There are stability-dependent biases between equivalent neutral winds and winds, resulting in biased fluxes of sensible and latent heat. For mid-latitudes, this bias in heat flux is roughly 10 Wm^{-2} in mid-latitude fluxes, which can very safely be ignored for weather forecasts of several days or less. Such biases become important on scales of 10 to 14 days On seasonal scales, regional errors of this magnitude could alter the layering of ocean circulation. Bulk formulas can be revised to account for this error in interpretation, but that is rarely done. Fluxes using satellite winds should be calculated using Eqs. 5, 6, 8, 9, and 12. However, the resulting equations are computationally intensive, which makes them awkward to use for ocean and atmospheric modeling. Ideally, equations similar to 7 and 10 would be developed, using forms of C_H and C_E that account for equivalent neutral winds like Eq. 13.

Datasets of satellite-derived near surface potential temperature and humidity

At this time, there are several relatively high-quality datasets of satellite-derived potential temperature and humidity, which can be used to determine turbulent heat fluxes as described in above These are the Seaflux dataset (Roberts et al., 2010), which is available from Woods Hole Oceanographic Institution, the NOAA Earth System Research Laboratory (ESRL) products (Jackson et al., 2010), and the U.S. Naval Research Laboratory NFLUX products, which are not yet available to the public. The Ocean Heat Flux product has been produced by Abderrahim Bentamy at IFREMER. More information about these products is given in Table 10.1. These turbulent fluxes can be combined with radiative fluxes from Clouds and the Earth's Radiant Energy System (CERES) or Surface Radiation Budget (SRB) to determine net heat fluxes.

Product	Reference	Website
Seaflux	Roberts et al. (2010)	http://seaflux.org/
NOAA/ESRL	Jackson et al. (2009)	https://mdc.coaps.fsu.edu/data/noaa-cires-multi-satellite-air-temperature-and-humidity
NFLUX	May et al. (2017)	Not yet publicly available
IFREMER	None at this time	ftp://o1ef56:DeJd6uNv@eftp.ifremer.fr/oceanheatflux/data/third-party/fluxes/ifremerflux_v4
SRB		https://gewex-srb.larc.nasa.gov
CERES		https://ceres.larc.nasa.gov

Table 10.1. Modern satellite-derived near-surface temperature and humidity datasets.

Atmospheric response to ocean variability

It is well known that stress depends on sea state and that air/sea temperature differences are important for determining fluxes. Sea state and surface winds also respond to ocean surface currents and gradients of atmospheric stratification and SST. These concepts become more interesting and presumably important when examining air/sea interaction around strong ocean currents and eddies. For example, ocean surface currents (e.g., tides) moving in the same direction as the wind reduce wind shear, which reduces stress, which increases atmospheric stability departures from neutral conditions. Most ocean conditions have unstable stratification, thus currents moving in the wind direction tend to make stratification more unstable, increasing stress and more effectively stronger winds down from the top of the boundary layer. This series of adjustments represents a negative feedback. Stable stratification would lead to positive feedback in a manner that reduces fluxes and might eventually trigger a negative feedback through reduced wind speeds as described earlier in relation to diurnal changes in SST.

Coupling coefficients between stress and SST gradients

Winds and surface stress has been shown (detailed below) to be modified by gradients in SST. These changes imply changes in surface turbulent fluxes, as well. These changes in winds and stress have been identified in satellite data on scales <1000 km. Coupling coefficients are the regression coefficient between an SST gradient and a characteristic of wind or stress such as speed, a vector component, divergence and curl. If the focus is on surface fluxes, then links to wind and stress are clearly important. Changes in the divergence influence vertical motion in the atmosphere, and changes in curl influence vertical motion in the ocean.

There are several hypotheses concerning the mechanisms that fundamentally drive the surface wind and wind stress response to strong SST fronts: 1) stability-dependent adjustment of turbulent mixing of momentum from the upper atmosphere to the surface (e.g., Sweet et al., 1981; Hayes et al., 1989; Wallace et al., 1989; Wai and Stage, 1989; Liu et al., 2000; Hashizume et al., 2001, 2002; Tokinaga et al., 2005; Spall, 2007); 2) generation of hydrostatic sea level pressure gradients through changes in atmospheric baroclinicity across the SST front (e.g., Lindzen and Nigam, 1987; Wai and Stage, 1989; Hashizume et al., 2001; Small et al., 2003, 2005; Song et al., 2006; Spall et al., 2007; O'Neill et al., 2010b), and 3) a rapid change in turbulent mixing that results in an unbalanced Coriolis force in the vicinity of the SST front (Spall et al., 2007). A great deal of controversy surrounds these hypotheses. Most of these suggested mechanisms are related to surface fluxes of heat, and the winds and SSTs used in studying these relationships are often measured from satellites.

For the momentum-mixing mechanism, the modification of the surface wind and wind stress is attributed (see references above) to SST-induced changes in the stratification of the marine atmospheric boundary-layer. Over warm water, the stability of the marine atmospheric boundary-layer is reduced and the buoyancy-driven turbulent mixing is increased, which enhances the downward mixing of momentum from aloft to the surface. This decreases the vertical wind shear in the boundary layer and increases the surface winds. The converse is true over cool water, where the vertical turbulent mixing is suppressed by the stronger static stability, resulting in greater vertical wind shear and lighter surface winds.

The sea level pressure gradient mechanism attributes the modification of the surface wind field to the SST-induced changes in sea level pressure that develop across SST fronts (see references above). The resulting perturbation pressure gradient force tends to accelerate the surface wind toward locally warmer water. Therefore, the surface wind can be enhanced or reduced depending on its directional alignment with the SST-induced sea level pressure gradient vector. For example, Song et al. (2006) showed that the perturbation pressure gradient force enhanced the surface wind speed in conditions where the mean surface flow traversed the Gulf Stream from the cool to the warm side, and reduced the surface wind speed in the converse. Small et al. (2003) found similar results over tropical instability waves in the eastern equatorial Pacific, where the perturbation pressure gradient force was responsible for strengthening the trade winds over warm SST anomalies and weakening the trade winds over colder SST anomalies.

Satellite-based observations show that the wind stress divergence and curl fields are linearly related to the downwind and crosswind component of the SST gradient, respectively (see the review by O'Neill). The perturbation in wind stress divergence or convergence is locally maximized where the wind stress vector is oriented parallel to the SST gradient vector (across isotherms). Conversely, the perturbation in wind stress curl is locally largest where the wind stress vector is aligned perpendicular to the SST gradient vector (along isotherms). This linear relationship between surface wind stress and SST perturbations is observed remotely (to varying degrees) over the Gulf Stream (e.g., O'Neill, 2012), the Kuroshio Extension (e.g., O'Neill et al., 2010a, 2012b; Maloney and Chelton, 2006), the Agulhas Return Current (e.g., O'Neill et al., 2005, 2010a, 2010b, 2012; Maloney and Chelton, 2006), the Brazil-Malvinas Confluence in the South Atlantic (e.g., O'Neill et al., 2010a, 2012), the California Current System (e.g., Chelton et al., 2007), and the eastern equatorial Pacific (e.g., Xie et al., 1998; Liu et al., 2000; Chelton et al., 2007). O'Neill (2012) confirmed the linear dependence using moored buoys in the Gulf Stream and eastern equatorial Pacific, where the wind speed, 10 m equivalent neutral wind speed, and surface wind stress were found to respond linearly to SST differences.

The strength of the linear relationship or coupling between the perturbations in surface wind and wind stress, and SST is shown to exhibit large geographical variability (see aforementioned references). The prominence of each proposed mechanism or force balance is, in part, dependent on the local oceanic and atmospheric conditions. Some of the factors that may contribute to the spatial variability are: the strength and spatial extent of the SST front; the strength and directional steadiness of the prevailing wind; the marine atmospheric boundary-layer stratification; the state of equilibrium between marine atmospheric boundary-layer and underlying SST, capping inversions, latitude, and proximity to land (Hashizume et al., 2001; Chelton et al., 2007; Spall et al., 2007; O'Neill et al., 2010b). Therefore, knowledge of the relative dominance of each mechanism is important and should be investigated over an extensive range of oceanic and atmospheric conditions.

Responses to fluxes

This section provides a subset of near-surface responses to surface fluxes on timescales that are typically of less than one day, and describes how changes to the near surface ocean and atmosphere in turn cause changes in the fluxes. Take away points are that the ocean and atmosphere both respond to surface forcing on short timescales (as well as on timescales greater than one day, which are well described elsewhere). These changes alter surface fluxes, causing further changes, or countering changes to the near-surface ocean and atmosphere. This coupling is stronger near surface features with strong gradients, such as some eddies, ocean fronts, and strong currents, which tend to occur on scales <50 km. These are time- and space scales that are difficult to observe and model in a coupled system: observing, modeling, and understanding coupled processes on these scales will be a hot topic for the foreseeable future, and both air-sea fluxes and remote sensing will be important in understanding these processes.

Subsections address responses at the surface, in the ocean, and in the atmosphere. This topic is of interest for many applications, such as diurnal variability, satellite sampling of diurnal cycles, waves and currents changes with changing winds, thermodynamical coupling of the ocean and atmosphere, and air-sea coupling around ocean fronts. For example, if an ocean model is forced with atmospheric (e.g., heat, momentum, or CO_2) fluxes, it accumulates the influence of biases, often resulting in very unrealistic ocean characteristics such as surface temperature, salinity, current speed, and acidity. Thus, errors on sub-daily timescales are temporally integrated to cause longer term errors, which can lead to very unrealistic ocean and atmospheric conditions and weather in models. Of course, when errors get too large additional processes come into play to reduce biases. For example, excessive ocean surface temperatures are reduced to fluxes of terrestrial radiation, latent heat and sensible heat. Similarly, currents that are too strong are reduced by friction with surrounding slower moving water.

Models that are one-way forced (where an atmosphere is modeled with a specified SST, or where an ocean is forced with surface fluxes from an atmospheric model) cannot modify the surface forcing to reduce the impacts of biases. The impacts of these biases can be greatly mitigated by calculating surface fluxes based on atmospheric variables (wind, air temperature, humidity, and surface pressure) coupled with ocean variables (SST, currents, and sea state) based on Eqs. 4, 7, and 10. As biases in these flux-related variables occur, there are associated changes in surface fluxes (momentum, heat, and others) that reduce the biased fluxes that are causing the problems. This solution to reduce biases in fluxes might seem odd because the atmospheric data used to force an ocean model does not change in response to the ocean model. However, the changes in surface fluxes imply changes in the near surface atmosphere consistent with Eqs. 2, 6, and 9. This approach to reducing biases does not remove biases in the atmospheric variables (e.g., excessive humidity or winds) and it does result in errors in the ocean model (e.g., excess temperatures and currents), but it does reduce biases in fluxes and thereby reduced biases in the ocean model.

Full two-way coupling could be thought of as a further extension of calculating fluxes to reduce biases in forcing. Two-way coupling is more complicated because it allows the atmosphere to respond to the ocean including atmosphere-related changes to the ocean, and the ocean to respond to the atmosphere including ocean-related changes to the atmosphere. Currently there is considerable interest in evaluating the need for two-way coupling. If there is a substantial small-scale and short-term atmospheric response to ocean coupling, then ocean forcing will benefit from two way coupling. However, two-way coupling and forcing from atmospheric variables is sensitive to the physics included in the model. For example, including surface currents (without considering sea state and atmospheric response) results in a roughly 25% drop in the production of ocean eddy kinetic energy (Renault et al., 2016a, 2016b), which is clearly unrealistic. Preliminary results (Shi, 2017) suggest that there is a quite substantial atmospheric response over western boundary currents, and that this response compensates to cause a net impact of increased stresses. Considerably more work needs to be done to investigate what physics and resolutions are required to provide sufficiently accurate coupling.

Surface responses

One of the most well-known responses to fluxes is the changes in sea surface temperature and wind vectors associate with the diurnal cycle. The changes in wind vectors are best known in land-sea breezes. Currently, these are difficult to observe from satellite because most satellites cannot measure closer than roughly the width of a footprint from the coast. However, weather radars show the complicated interaction between surface temperature gradients, the horizontal pressure gradients and winds they induce, and atmospheric convection and outflow. In this example it is quite clear that the atmospheric response is substantial.

Large diurnal changes in SST are associated with low wind speeds, little mixing of the near-surface ocean, and daytime heating. There can be very large SST changes associated with strong advection of cold air, but these are not daily repeating changes. The low wind speeds are essential because they reduce the vertical transport of heat downward in the ocean and upward through latent heat. Consequently, the absorbed solar energy raises the water temperature, with greater absorption near the surface and hence greater heating near the surface. The resulting vertical gradient in density is stably stratified, which further reduces vertical transport in the ocean. Thus far, the process has a positive feedback, with strengthens the diurnal cycle (Weihs and Bourassa, 2014). The negative feedback is presumably due to increased surface winds associated with atmospheric convection and possibly associated with temperature gradients (discussed earlier). If there was no atmospheric feedback, the magnitude of the diurnal heating would be constrained only by the mean heat flux over a diurnal cycle.

Ocean responses

The case of diurnal heating is a simplified example of a thermodynamical ocean response to air sea coupling. In most cases or on large scales, the energy transfer in the ocean is dominated by vertical heat fluxes. Strong warm and cold currents are obvious exceptions, but the currents are not strong over much of the ocean, and when averaged over a large enough scale (several hundred kilometers), the impact currents are averaged down. However, around strong currents and eddies the horizontal transport of energy can be important. On small scales (mesoscale) there can be large gradients in surface stress, causing relatively strong vertical motions and vertical transport of energy. Surface-layer vertical motion is usually due to Ekman pumping, which is proportional to the curl of the surface stress. This makes resolution of the data and the complexity of the surface stress model more important. Surface stress parameterizations that are commonly applied consider the important of wind speed and (in more modern parameterizations) air-sea temperature gradients. Considering currents and sea state influences causes greater gradients and curls of stress, which makes vertical motion locally important. The importance of spatial scale and complexity of the flux parameterizations are topics of ongoing research. Satellite data are much easier to utilize in the sense that they account for additional physics; however, the sampling from satellite is sufficiently sparse to present additional challenges, and greater resolution of vector stresses is highly desired.

Atmospheric responses

As stated above, it is often assumed that there is no atmospheric response to the ocean; however, it is also pointed out that small-scale ocean variability imparts considerable forcing on the atmosphere. Since the atmosphere has relatively little inertia and thermal inertial compared to the ocean, it seems likely that atmospheric responses should occur. Such reasoning is strongly supported by the complex interaction of atmospheric convection and outflow associated with land-sea breezes. Clearly changes in surface stress associated with currents, waves, wave-current interaction, and horizontal temperature gradients (in the ocean and the atmosphere) should impact the lower atmospheric boundary layer. It has been argued that the Gulf Stream impacts the strength of cyclogeneses. This concept is strongly supported by models of atmospheric cyclogenesis; however, the limited time spent over the Gulf Stream greatly complicates analyses of cyclone evolution. Preliminary results indicated that observed coupling between SST gradients and stress can be used to test the sensitivity of coupled models to stress parameterizations (Shi, 2017). On smaller scales, it is much clearer how changes in atmospheric circulation modify air-sea fluxes and the ocean response. Such questions seem to be important for weather forecasts around western boundary currents (and downwind of them), to address biases in seasonal forecasting, to reduce biases in climate models, and, of more immediate interest, to improve operational ocean models.

Acknowledgements

This work was funded, in part, by the Jet Propulsion Laboratory, National Aeronautics and Space Administration under contract #1419699 and NNX15AD45G S02. This work was also funded, in part, by the Northern Gulf Institute (grant number 18-NGI3-42), National Oceanic and Atmospheric Administration, U.S. Department of Commerce under award number NA11OAR432019 and NA16OAR4320199.

References

Anguelova, M.D., and P.W. Gaiser, 2013: Microwave emissivity of sea foam layers with vertically inhomogeneous dielectric properties, *Remote Sensing of Environment*, **139**, 81-96, doi:10.1016/j.rse.2013.07.017.

Anguelova, M.D., and P.W. Gaiser, 2012: Dielectric and radiative properties of sea foam at microwave frequencies: Conceptual understanding of foam emissivity, *Remote Sensing*, **4**, 1162-1189, doi:10.3390/rs4051162.

Anguelova, M.D. and F. Webster, 2006: Whitecap coverage from satellite measurements: A first step toward modeling the variability of oceanic whitecaps, *J. Geophys. Res*., **111**, C03017, doi:10.1029/2005JC003158.

Bourassa, M.A., E. Rodriguez, and R. Gaston, 2010: NASA's Ocean Vector Winds Science Team workshops. *Bull. Amer. Meteor. Soc*., **91**, doi:10.1175/2010BAMS2880.1

Bourassa, M.A., 2006: Satellite-based observations of surface turbulent stress during severe weather, Atmosphere - Ocean Interactions, Vol. 2., ed., W. Perrie, Southampton, UK, Wessex Institute of Technology Press, 35 – 52 pp.

Bourassa, M.A., E. Rodriguez, and R. Gaston, 2010: NASA's Ocean Vector Winds Science Team workshops. *Bull. Amer. Meteor. Soc*., **91**, doi:10.1175/2010BAMS2880.1

Bourassa, M.A., D.G. Vincent, W.L. Wood, 1999: A flux parameterization including the effects of capillary waves and sea state. *J. Atmos. Sci*., **56**, 1123-1139.

Bourassa, M.A., S.T. Gille, D.L. Jackson, B.J. Roberts, and G.A. Wick, 2010: Ocean winds and turbulent air-sea fluxes inferred from remote sensing. *Oceanography*, **23**, 36-51.

Chakraborty, A., R. Kumar and A. Stoffelen, 2013: Validation of ocean surface winds from the OCEANSAT-2 scatterometer using triple collocation. *Remote Sensing Letters*, **4**(1), 84–93, doi:10.1080/2150704X.2012.693967.

Chelton, D.B., M. G. Schlax, and R.M. Samelson, 2007: Summertime coupling between sea surface temperature and wind stress in the California current system. *J. Phys. Oceanogr*., **37**, 495–517.

Clayson, C.A., C. W. Fairall, and J.A. Curry, 1996: Evaluation of turbulent fluxes at the ocean surface using surface renewal theory. *J. Geophys. Res*., **101**, 28503-28513.

Cornillon, P., Park, K.A., 2001: Warm core ring velocities inferred from NSCAT. *Geophys. Res. Letters*, **28**, 575–578, doi:10.1029/2000GL011487.

De Kloe, A. Stoffelen and A. Verhoef, 2017: Improved Use of Scatterometer Measurements by Using Stress-Equivalent Reference Winds. JSTARS, 10, doi:10.1109/JSTARS.2017.2685242

Dobson, F.W., S.D. Smith, and R.J. Anderson, 1994: Measuring the relationship between wind stress and sea state in the open ocean in the presence of swell. *Atmos.-Ocean*, **32**, 327-256.

Draper, D.W., Long, DG (2004), Evaluating the effect of rain on SeaWinds scatterometer measurements. *J. Geophys. Res.-Oceans*, **109**(C02005), doi:10.1029/2002JC001741.

Drennan, W.M., P.K. Taylor, and M.J. Yelland, 2005: Parameterizing the sea surface roughness. *J. Phys. Oceanogr.,* **35***,* 835-848.

Fairall, C.W., E.F. Bradley, D.P. Rogers, J.B. Edson, and G.S. Young, 1996: Bulk parameterizations of air-sea fluxes for Topical Ocean-Global Atmosphere Coupled-Ocean Atmosphere Response Experiment. *J. Geophys. Res*., **101**, 3747-3764.

Fairall, C., E.Bradley, J. Hare, A. Grachev, and J. Edson, 2003: Bulk parameterization of air–sea fluxes: updates and verification for the CORE algorithm. *J. Clim*, **16**, 571–591.

Gasier, P.W., and Coauthors, 2004: The WindSat spaceborne polarimetric microwave radiometer: Sensor description and early orbit performance. *IEEE Trans. Geosci. Remote Sens.*, **42**, 2347-2361.

Gordon, H.R., and M. Wang, 1994: Retrieval of water-leaving radiance and aerosol optical thickness over the oceans with SeaWiFS: A preliminary algorithm, *Appl. Opt.*, **33**, 443–452.

Gulev, K.S., S.A. Josey, M. Bourassa, L.-A. Breivik, M.F. Cronin, C. Fairall, S. Gille, E.C. Kent, C.M. Lee, M.J. McPhaden, P.M.S. Monterio, U. Schuster, S.R. Smith, K.E. Trenberth, D. Wallace, and S.D. Woodruff, 2010: Surface energy, CO_2 fluxes and sea ice. *Proceedings of the OceanObs'09: Sustained Ocean Observations and Information for Society Conference (Vol. 1)*, Venice, Italy, eds. J. Hall, D.E. Harrison and D. Stammer, ESA Publication WPP-306. doi:10.5270/OceanObs09.pp.19

Hansen, J., and Coauthors, 2005: Earth's energy imbalance: Confirmation and implications. *Science,* **308,** doi:10.1126/science.1110252.

Hashizume, H., S.-P. Xie, W. T. Liu, and K. Takeuchi, 2001: Local and remote atmospheric response to tropical instability waves: a global view from space. *J. Geophys. Res.*, **106**, 10173–10185.

Hashizume, H., S.-P. Xie, M. Fujiwara, M. Shiotani, T. Watanabe, Y. Tanimoto, W. T. Liu, and K. Takeuchi, 2002: Direct Observations of atmospheric boundary layer response to SST variations associated with tropical instability waves over the eastern Equatorial Pacific. *J. Climate*, **15**, 3379–3393.

Hayes, S. P., M. J. McPhaden, and J. M. Wallace, 1989: The influence of sea-surface temperature on surface wind in the Eastern Equatorial Pacific: weekly to monthly variability. *J. Climate*, **2**, 1500–1506.

Holbach, H.M. and M.A. Bourassa, 2017: Platform and across-swath comparison of vorticity spectra from QuikSCAT, ASCAT-A, OSCAT, and ASCAT-B Scatterometers. *IEEE Journal of Selected Topics in Applied Earth Observations and Remote Sensing*, **10**(5), 2205–2213, doi:10.1109/JSTARS.2016.2642583.

Jackson, D.L., and G.A. Wick, 2010: Near-surface air temperature retrieval derived from AMSU-A and sea surface temperature observations. *J. Atmos. Oceanic Technol.*, **27**, 1769-1776.

Jackson, D.L., G.A. Wick, and J.J. Bates, 2006: Near-surface retrieval of air temperature and specific humidity using multisensor microwave satellite observations, *J. Geophys. Res*., **111**, D10306, doi:10.1029/2005JD006431.

Jackson, D.L., G.A. Wick, and F.R. Robertson, 2009: Improved multi-sensor approach to satellite-retrieved near-surface specific humidity observations. *J. Geophys. Res*., 114, D16303, doi:10.1029/2008JD011341.

Kara, A.B., A.J. Wallcraft, and M.A. Bourassa, 2008: Air-Sea Stability Effects on the 10m Winds Over the Global Ocean: Evaluations of Air-Sea Flux Algorithms. *J. Geophys. Res*., **113**, C04009, doi:10.1029/2007JC004324.

Kara, B., Kara A.B., Hurlburt H.E., Rochford P.A., 2000: Efficient and accurate bulk parameterizations of air–sea fluxes for use in general circulation models. *J. Atmos. Ocean. Technol.*, **17,** 1421-1438.

Kelly, K.A., S. Dickinson, M.J. McPhaden, and G.C. Johnson, 2001: Ocean currents evident in satellite wind data. *Geophys. Res. Letters*, **28**, 2469–2472, doi:10.1029/2000GL012610.

Knauss, J.A., 2005: *Introduction to Physical Oceanography*, Waveland Press. Second Edition. ISBN 978-1-57766-429-1

Kondo, J., 1975: Air-sea bulk transfer coefficients in diabatic conditions. *Bound.-Layer Meteor*., **9**, 91-112.

Lindzen, R.S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the tropics. *J. Atmos. Sci.*, **44**, 2418–2436.

Liu, W.T., and W. Tang, 1996: Equivalent neutral wind. JPL Publ., 96–17, 8 pp.

Liu, W.T., X. Xie, P.S. Polito, S.-P. Xie, and H. Hashizume, 2000: Atmospheric manifestation of tropical instability wave observed by QuikSCAT and Tropical Rain Measuring Mission. *Geophys. Res. Lett.*, **27**, 2545–2548.

Maloney, E.D., and D.B. Chelton, 2006: An assessment of the sea surface temperature influence on surface wind stress in numerical weather prediction and climate models. *J. Climate*, **19**, 2743–2762.

May, J.C., C. Rowley, and C.N. Barron, 2017: NFLUX Satellite-Based Surface Radiative Heat Fluxes. Part I: Swath-Level Products. *J. Applied Met. Clim*., **56**, 1025 – 1041, doi:10.1175/JAMC-D-16-0282.1

Monin, A.S. and A.M. Obukhov, 1954: Basic laws of turbulent mixing in the surface layer of the atmosphere. Tr. Akad. Nauk. SSSR Geophiz. Inst., **24** (151): 163–187.

Nikuradse, J., 1933: *Stromungsgesetze in rauben Rohren*. V. D. I. Forschungsheft 361, 22 pp.

O'Neill, L.W., 2012: Wind speed and stability effects on coupling between surface wind stress and SST observed from buoys and satellite. *J. Climate*, **25**, 1544–1569.

O'Neill, L.W., D.B. Chelton, and S.K. Esbensen, 2005: High-resolution satellite measurements of the atmospheric boundary layer response of SST variations along the Agulhas return current. *J. Climate*, **18**, 2706–2723.

O'Neill, L. W., D. B. Chelton, and S. K. Esbensen, 2010a: The effects of SST-indeced surface wind speed and direction gradients on midlatitude surface vorticity and divergence. *J. Climate*, **23**, 225–281.

O'Neill, L.W., D. B. Chelton, and S. K. Esbensen, 2010a: The effects of SST-indeced surface wind speed and direction gradients on midlatitude surface vorticity and divergence. *J. Climate*, **23**, 225–281.

O'Neill, L.W., S.K. Esbensen, N. Thum, R.M. Samelson, and D.B. Chelton, 2010b: Dynamical analysis of the boundary layer and surface wind responses to mesoscale SST perturbations. *J. Climate*, **23**, 559–581, doi:10.1175/2009JCLI2662.1.

Paget, A., M.A. Bourassa, and M.D. Anguelova, 2015: Comparing in situ and satellite-based observations of oceanic whitecaps. *J. Geophys. Res*., **120**, 2826-2843, doi:10.1002/2014JC010328.

Petty, G. W., 2006: Thermal emission. A first course in atmospheric radiation. Second Edition, Sundog Publishing, 113–154.

Plagge, A.M., D. Vandemark and B. Chapron, 2012: Examining the impact of surface currents on satellite scatterometer and altimeter ocean winds. *J. Atmos. Oceanic Technol.*, **29**, 1776–1793, doi:10.1175/JTECH-D-12-00017.1.

Renault, L., M. J. Molemaker, J. Gula, S. Masson, and J. C. McWilliams, 2016a: Control and Stabilization of the Gulf Stream by Oceanic Current Interaction with the Atmosphere, *J. Phys. Oceanogr.*, **46**(11), 3439–3453, doi:10.1175/JPO-D-16-0115.1.

Renault, L., M.J. Molemaker, J.C. McWilliams, A.F. Shchepetkin, F. Lemarié, D. Chelton, S. Illig, and A. Hall, 2016b: Modulation of wind-work by oceanic current interaction with the atmosphere, *J. Clim.*, 0–52, doi:10.1175/JPO-D-15-0232.1.

Roberts, B., C.A. Clayson, F.R. Robertson, and D.L. Jackson, 2010: Predicting near-surface characteristics from SSM/I using neural networks with a first guess approach. *J. Geophys. Res.*, **115**, D19113, doi:10.1029/2009JD013099.

Rose, L., W. Asher, S. Reising, P. Gaiser, K. St Germain, D. Dowgiallo, K. Horgan, G. Farquharson, and E. Knapp, 2002: Radiometric measurements of the microwave emissivity of foam, *IEEE Trans. Geosci. Remote Sens.,* **40**(12), 2619–2625, doi:10.1109/TGRS.2002.807006.

Ruf, C., M. Unwin, J. Dickinson, R. Rose, D. Rose, M. Vincent, A. Lyons, "CYGNSS: Enabling the Future of Hurricane Prediction", *Geoscience and Remote Sensing Magazine IEEE*, vol. 1, pp. 52-67, 2013.

Small, R.J., S.-P. Xie, Y. Wang, S.K. Esbensen, and D. Vickers, 2005: Numerical simulation of boundary layer structure and cross-equatorial flow in the Eastern Pacific. *J. Atmos. Sci.*, **62**, 1812–1830.

Small, R.J., S.P. deSzoeke, S.P. Xie, L. O'Neill, H. Seo, Q. Song, P. Cornillon, M. Spall, and S. Minobe, 2008: Air–sea interaction over ocean fronts and eddies. *Dyn. Atmos. Oceans*, **45**, 274–319, doi:10.1016/j.dynatmoce.2008.01.001.

Shi, Q., 2017: Coupling ocean currents and waves with wind stress over the Gulf Stream. Ph.D Dissertation, Florida State University

Smith, P., 1998: The emissivity of sea foam at 19 and 37 GHz. *IEEE Trans. Geosci. Remote Sens., 26*, 541-547.

Smith, S. D., 1988: Coefficients for sea surface wind stress, heat flux, and wind profiles as a function of wind speed and temperature. *J. Geophys. Res*., **93**, 15467-15472.

Smith, S.R., M.A. Bourassa, and D.L. Jackson, 2012: Supporting Satellite Research with Data Collected by Vessels. *Sea Tech*., June 2012, 21 – 24.

Stull, R. 1988: An introduction to boundary layer meteorology. Netherlands: Springer. ISBN 978-94-009-3027-8.

Song, Q., P. Cornillon, and T. Hara, 2006: Surface wind response to oceanic fronts. *J. Geophys. Res.*, **111**, C12006, doi:10.1029/2006JC003680.

Spall, M.A., 2007: Midlatitude wind stress–sea surface temperature coupling in the vicinity of oceanic fronts. *J. Climate*, **20**, 3785–3801, doi:10.1175/JCL14234.1.

Sweet, W., R. Fett, J. Kerling, and P. La Violette, 1981: Air–sea interaction effects in the lower troposphere across the North Wall of the Gulf Stream. *Mon. Wea. Rev.*, **109**, 1042–1052.

Sverdrup, H.U., 1947: Wind-driven currents in a baroclinic ocean; with application to the equatorial currents of the Eastern Pacific. *Proc. Natl. Acad. Sci. U.S.A.*, **33**(11): 318–26.

Taylor, P.K., and M.J. Yelland, 2001: The dependence of sea surface roughness onthe height and steepness of the waves. *J. Phys. Oceanogr.,* **31,** 572–590.

Tokinaga, H., Y. Tanimoto, and S.-P. Xie, 2005: SST-induced surface wind variations over the Brazil-Malvinas confluence: satellite and in situ observations. *J. Climate*, **18**, 3470–3482.

Ulaby, F.T. and D.G. Long, 2014: Microwave Radar and Radiometric Remote Sensing, *University of Michigan Press*, Ann Arbor, 984 pp., ISBN: 978-0-472-11935-6

Uhlhorn, E.W., and P.G. Black, 2003: Verification of remotely sensed sea surface winds in hurricanes. *J. Atmos. Oceanic Technol.*, **20**, 99–116, doi:10.1175/1520-0426(2003)020<0099:VORSSS>2.0.CO;2.

Wai, M. M.-K., and S.A. Stage, 1989: Dynamical analyses of marine atmospheric boundary layer structure near the Gulf Stream oceanic front. *Q. J. R. Meteorol. Soc.*, **115**, 29–44.

Wallace, J.M., T.P. Mitchell, and C. Deser, 1989: The influence of sea-surface temperature on surface wind in the Eastern Equatorial Pacific: seasonal and interannual variability. *J. Climate*, **2**, 1492–1499.

Weihs, R., and M.A. Bourassa, 2014: Modeled diurnally varying sea surface temperatures and their influence on surface heat fluxes*. J. Geophys. Res*., **119***,* 4101–4123*,* doi:10.1002/2013JC009489*.*

Wentz, F. (1983), A model function for ocean microwave brightness temperature, *J. Geophys Res., 88*, 1892-1908.

Wentz, F. (1997), A well-calibrated ocean algorithm for special sensor microwave.imager, *J. Geophys. Res., 102*, 8703-8718.

Wu, J., 1968: Laboratory studies of wind-wave interactions. J. Fluid Mech., 34, 91-111.

Wu, J., 1980: Wind-stress coefficients over sea surface near neutral conditions--A revisit. *J. Phys. Oceanogr*., 10, 727-740.

Wu, J., 1994: The sea surface is aerodynamically rough even under light winds, *Bound..-Layer Meteor*., **69**, 149-158.

Xie, S.P., Ishiwatari, M., Hashizume, H. and Takeuchi, K., 1998: Coupled ocean-atmospheric waves on the equatorial front. *Geophys. Res. Letters*, **25**(20), 3863-3866.

Yueh, S.H., W.J. Wilson, F.K. Li, S.V. Nghiem, W.B. Ricketts, 1995: Polarimetric measurements of sea surface brightness temperatures using an aircraft K-band radiometer. *IEEE Trans. Geosci. Remote Sens.*, **33,** 1, 85–92.

Zavorotny, V.U. and A.G. Voronovich, "Scattering of GPS signals from the ocean with wind remote sensing application", *IEEE Trans. Geosci. Remote Sensing*, vol. 38, pp. 951-964, 2000.

Zheng, Y., M.A. Bourassa, and P.J. Hughes, 2013: Influences of sea surface temperature gradients and surface roughness changes on the motion of surface oil: a simple idealized study. *J. Appl. Meteor. Clim*., **52**, 1561 –1575, doi:10.1175/JAMC-D-12-0211.1