Diurnal variations in sea surface salinity (SSS) have been observed at a few selected locations with adequate in situ instrumentation. Such variations result primarily from imbalances between surface freshwater fluxes and vertical mixing of deeper water to the surface. New observations becoming available from satellite salinity remote sensing missions could help to constrain estimates of diurnal variations in air-sea exchange of freshwater, and provide insight into the processes governing diurnal variability of mixing processes in the upper ocean. Additionally, a better understanding of variation in near surface salinity is required to compare satellite measured SSS with in situ measurements at a few meters depth. The diurnal SSS variations should be reflected as differences between ascending and descending pass retrievals from the Aquarius and SMOS satellites; however, the diurnal signal may be masked by inadequacies of the geophysical corrections used in processing the satellite measurements. In this study, we quantify the expected range of diurnal SSS variations using a model developed for predicting diurnal sea surface temperature variations. We present estimates for the mean and variance of the global diurnal SSS cycle, contrasting it with the diurnal cycle of sea surface temperature. We find the SSS diurnal cycle can be significant throughout the tropics, with mean amplitudes of up to 0.1 psu in areas with heavy precipitation. Predicted maximum diurnal ranges approach 2 psu in select regions. Surface freshening in Aquarius salinity retrievals is shown to be larger for ascending than descending passes, consistent with the expectations from the model simulation.
The Aquarius/SAC-D [Lagerloef et al., 2008] and SMOS [Kerr et al., 2010] salinity remote sensing missions offer frequent global measurement of SSS, vastly increasing the coverage and frequency of salinity observations. Diurnal SSS variations should be detected as differences between ascending-pass and descending-pass SSS retrievals from the Aquarius/SAC-D or SMOS satellites. In the following, we discuss the sampling characteristics of Aquarius, but similar considerations apply to SMOS. The satellite travels in a sun-synchronous orbit, sampling the entire globe each week. Its descending pass (north to south) measures SSS in the morning hours (equator crossing at local time of 06:00 h), while its ascending pass (south to north) measures SSS in the evening hours (equator crossing at local time of 18:00 h). The climatological mean Aquarius ascending minus descending SSS difference is illustrated in Figure 1. While some features in the distribution of differences seem physically plausible, e.g., afternoon freshening along the intertropical convergence zones (ITCZ) of the Pacific and Atlantic basins, others features such as large values in midlatitude to high-latitude seem unlikely to be real. This is clearly a challenging remote sensing problem. Conversion of the brightness temperatures measured by the Aquarius L-band microwave radiometer to sea surface salinity requires accounting for a large number of factors and modeling their effects or masking out contaminated areas. These include radio frequency interference, land or sea ice contamination, direct or reflected solar radiation, or reflected background galactic radiation that can all impact the ascending and descending measurements differently, thereby projecting onto the estimated diurnal salinity variation [Meissner et al., 2014b]. While there are both first principles and empirical corrections employed for each of these, and the mapped ascending minus descending differences shown in Figure 1 exclude any values flagged as even “moderately” contaminated, it is still difficult to say with certainty how reliable these estimates are.

Motivated by both the physical science questions relating to near surface salinity stratification and the remote sensing issues described above, our objective in this study is to provide a first global estimate of the distribution of diurnal variations in SSS, and to provide some insight into the processes controlling this variability. We accomplish this by adapting a simple model developed for predicting diurnal variability in SST, as described in section 2. The character of the diurnal variation in SSS as predicted by the model, and its global and seasonal distribution are presented in section 3 along with comparison to in situ measurements. Section 4 discusses the application of these results to the mission of the Aquarius/SAC-D, and how the Aquarius satellite may be able to resolve the diurnal SSS cycle.

2. Methods

We use the Sea Surface Diurnal Cycling (SSDC) scheme described in Large and Caron (2015) (hereafter LC15) to simultaneously calculate the diurnal cycles of near surface temperature, salinity, and current shear, forced with atmospheric observations or reanalysis products as described in Table 1. For a full description of the model formulation, its validation for sea surface temperature, and application in a climate model, the reader is referred to LC15. We provide a brief summary here, with particular focus on the aspects relating to salinity. The structure of the model is similar in some respects to the model of Zeng and Beljaars [2005], in particular by assuming a prescribed depth of the diurnal layer, and a parameterized vertical distribution of the dependent variables within the layer. Also similar to the Zeng and Beljaars [2005] model, it is formulated to be computationally
efficient with the intention that it can be incorporated into global climate models at negligible cost. The primary difference from that model and its successors [e.g., Takaya et al., 2010] is that the shear at the base of the diurnal layer is explicitly taken into account when determining mixing with deeper waters.

In the following, we adopt a terminology for the near surface salinity structure following the conventions in common use for near surface temperature [see Kawai and Wada, 2007 review]. The foundation temperature, \( T_f \), and salinity, \( S_f \), are defined as values at a depth below the direct influence of diurnal forcing. The subskin, or bulk temperature, \( T_{\text{bulk}} \), and salinity, \( S_{\text{bulk}} \), are defined as their foundation values with the addition of an incremental "warming," \( \Delta T_w \), or "salting," \( \Delta S \), which account for the diurnal variation within the upper several meters of the ocean.

\[
T_{\text{bulk}} = T_f + \Delta T_w \\
S_{\text{bulk}} = S_f + \Delta S 
\]  

(1)

The sea surface temperature is further modified by a cool-skin layer that differs from the bulk temperature by \( \Delta T_c \) where heat transfer is controlled by molecular diffusion.

\[
T_{\text{skin}} = T_{\text{bulk}} + \Delta T_c 
\]  

(2)

Note that while \( \Delta T_w \geq 0 \), and \( \Delta T_c \leq 0 \), \( \Delta S \) may take either sign depending on whether precipitation or evaporation dominates the surface forcing. In the following, for simplicity and consistency of notation, we will refer to the surface salinity difference from the foundation value \( \Delta S \) as "salting," even though in many cases it is actually a freshening. In contrast to the case for temperature, we do not consider a skin effect on salinity. As the Aquarius radiometer measures the salinity in the top centimeter of the ocean, and the skin effect generally only appears in the top 10–20 m, we expect the skin effect to only have a small effect on the remote sensing measurements of SSS [Yu, 2010; Zhang and Zhang, 2012].

The basis of the SSDC scheme is the coupled system of conservation equations for heat, salinity, and current speed within the diurnal layer:

\[
\frac{\partial}{\partial t} \int_{-d}^{0} T(z)dz = \left[ \frac{SW_0(1-A_d) + Q_w}{\rho c_p} \right] - K_c \frac{\partial T}{\partial z} \bigg|_{z=-d} \\
\frac{\partial}{\partial t} \int_{-d}^{0} S(z)dz = (E-P)S_{\text{bulk}} - K_s \frac{\partial S}{\partial z} \bigg|_{z=-d} \\
\frac{\partial}{\partial t} \int_{-d}^{0} V(z)dz = \frac{1}{\rho_w} \left| \tau \right| - 10 \frac{\partial V}{\partial z} \bigg|_{z=-d} 
\]  

(3)

The first term on the right hand side of each equation is the forcing of the diurnal layer by the atmosphere, while the second term is the flux through the base of the diurnal layer due to turbulent mixing. The diurnal layer depth \( d \) and the skin-layer thickness \( \delta \) are prescribed constants.

\( SW_0 \) is the net solar insolation at the sea surface. In the estimates presented in this paper, we use the daily average values from the International Satellite Cloud Climatology project (ISCCP-FD) [Zhang et al., 2004].
The diurnal dependence of solar insolation is prescribed by a latitude and time varying solar zenith angle function [Berger, 1978]. We have thus neglected the impact of changes in cloudiness during the daylight hours on the incident solar radiation. This was necessary because the available 6 hourly resolution of the ISCCP-FD product under resolves the daily variation of the solar insolation, and would make the amplitude of the diurnal warming a strong function of longitude, depending on how close local noon was to one of the four available UTC sample times. \( A_d \) is the fractional solar transmission at the base of the diurnal layer. In the present study, it is computed as the two-wavelength absorption for Jerlov water-type Ib [Paulson and Simpson, 1977]:

\[
A_d = 0.67 \exp \left(-\frac{d}{1m}\right) + 0.33 \exp \left(-\frac{d}{17m}\right) \tag{4}
\]

\( Q_n \) is the nonsolar surface heat flux, itself given by:

\[
Q_n = LW^i - LW^i - Q_{H} - Q_{E} \tag{5}
\]

The downwelling longwave radiation \( LW^i \) is provided at daily temporal resolution from the same data source as the shortwave flux. The upward longwave flux is computed using the model predicted skin temperature:

\[
LW^u = \sigma T_{s}^4 \tag{6}
\]

The turbulent latent and sensible heat fluxes \( Q_{E} \) and \( Q_{H} \), evaporation \( E \), and stress magnitude \( \mathbf{\tau} \) are computed from the model predicted surface bulk temperature \( T_{bulk} \) and the surface atmospheric state (wind speed, air temperature, and humidity) at 6 h temporal resolution from the Coordinated Ocean Reference Experiment (CORE) forcing data set [Large and Yeager, 2009] using the bulk formulae prescribed in the CORE protocol. The sensible heat flux associated with the temperature difference between rain water and the sea surface temperature is neglected in this study. To capture the potential diurnal forcing of precipitation, \( P \), on salinity we use the CMORPH satellite based precipitation estimates [Joyce et al., 2004] at 3 h resolution.

The model assumes a parameterized functional form of the temperature, salinity, and current speed profiles within the diurnal layer above the prescribed depth \( d \):

\[
T(z) = T_{bulk} - \left(\frac{z + \delta}{-d + \delta}\right)^p \Delta T_W \tag{7}
\]

\[
S(z) = S_{bulk} - \left(\frac{z}{-d}\right)^p \Delta S \tag{7}
\]

\[
V(z) = V_{bulk} - \left(\frac{z}{-d}\right)^p \Delta V \tag{7}
\]

with \( \Delta T_W \) and \( \Delta S \) defined above, and \( \Delta V \) the analogous shear across the diurnal layer. LC15 find empirically that an exponent \( p = \frac{1}{5} \) and \( d = 3m \) provides a good simulation of the global distribution of the diurnal range of sea surface temperature and these values are retained in the present study. Schematic diagrams of the temperature and salinity profiles are shown in Figure 2. It should be emphasized that the constraints used by LC15 in optimizing the parameters \( p \) and \( d \) were based on satellite and in situ sea surface temperature observations, not high-resolution profile data. We expect these empirically determined values to provide reasonable values of the surface response to diurnal forcing, and implicitly the gradients at the base of the diurnal layer such that the mixing with deeper waters is appropriate, but we do not assert that they can be used to reliably predict the detailed vertical structure within the diurnal layer. Since sea surface salinity is governed by the same mixing processes as temperature, we proceed on the assumption that the results generated for surface salinity will be similarly accurate. Part of the analysis below will be directed at testing this assumption.

Using the expressions (7), the budget equations (3) reduce to a set of ordinary differential equations:

\[
\frac{d}{dt} \Delta T_W = H_d \frac{(p+1)}{pd} - K_d \Delta T_W \frac{(p+1)}{d^2} \tag{8}
\]

\[
\frac{d}{dt} \Delta S = -F_d \frac{(p+1)}{pd} - K_d \Delta S \frac{(p+1)}{d^2} \tag{8}
\]

\[
\frac{d}{dt} \Delta V = u_d \frac{(p+1)}{pd} - v_d \Delta V \frac{(p+1)}{d^2} \tag{8}
\]
where $H_d$ and $F_d$ are the net heat and freshwater forcing. These equations are integrated using an implicit time-stepping scheme with a 30 min time step on a spatial grid with resolution of 0.25° in latitude and longitude. All forcing data described above are linearly interpolated to the model time and grid positions. The integration described below is for the period 2007–2009 when all of the forcing data sets are available, and is the same as the experiment referred to as DD* in LC15.

It remains to describe how the turbulent mixing is determined. Each day is split up into four regimes with different near surface mixing characteristics: a nocturnal deep well-mixed regime, a deep-stable regime, a shallow-stable regime, and a deepening-convective regime (Figure 3). At midnight, it is assumed that evening convection has mixed the upper ocean creating a nocturnal boundary layer much deeper than $d$, and Regime I is designated. The diurnal component of salting, warming, and shear are set to zero. When the net surface buoyancy flux first becomes positive (stabilizing), Regime II is designated, which then persists as long as the Monin-Obukov length is large compared to the diurnal layer depth $d$. However, LC15 found that Regime II typically only persists for a short time, so as a simplification, the model is run without Regime II, and Regime I transitions directly to Regime III. In Regime III, the fluxes across the base of the diurnal layer are governed by shear driven mixing, and the diffusivity $K_d$ and viscosity $t_d$ in the prognostic equations (8) are given by

\[
K_d = K_0 + \nu_0 v(R_{ld}) \\
\nu_d = \nu_0 + \nu_0 v(R_{ld})
\]

where $Pr$ is the Prandtl number and $R_{ld}$ is the gradient Richardson number at depth $d$, which can be evaluated from the assumed profile shapes (7) as

\[
R_{ld} = \frac{g (\alpha \Delta T_W - \beta \Delta S) d}{\Delta V^2}
\]

The details of the functional form of $v(R_{ld})$ and the choices for the other parameters are described in LC15. Regime III ends when the net surface buoyancy flux becomes negative, and the ocean boundary layer becomes convective, resulting in Regime IV. In this regime, the diffusivity and viscosity become proportional to the convective velocity scale, following the K-profile parameterization [Large et al., 1994].

The SSDC scheme has been tested through comparison with the global diurnal SST differences measured by the AMSR-E satellite (LC15). We additionally compare the model results with observations from the Global Tropical Moored Buoy Array in the following section. To illustrate the functioning of the model, and the regime transitions, we consider the solution at two representative buoy locations for a sample model day. A location with heavy rainfall, 5°N 147°E, is shown in Figures 3a–3d. Typical of the deep tropics, the
maximum precipitation occurs in the early morning hours around dawn. Despite the net freshwater input, the buoyancy flux remains negative until after sunrise, so the SSDC remains in the deep convective Regime I, preventing a fresh layer from developing. During this period, the surface fluxes would force the foundation temperature and salinity directly, but in this experiment we consider the diurnal layer values in isolation. After sunrise, as the surface layer stabilizes due to positive buoyancy input, freshwater from precipitation begins to accumulate on the surface, causing rapid local freshening. At the same time, as the diurnal layer becomes more stratified, momentum is trapped near the surface, increasing the vertical shear. As a result, $K_d$ and hence vertical mixing of heat and salinity across the base of the diurnal layer increase.

While the net freshwater flux remains positive (freshening) until around local noon, the SSS begins to increase around 0930h. As precipitation decreases and $K_d$ increases through the late afternoon, SSS returns to its nighttime value. Note that at this location, while there is a significant variation in surface salinity through the day, it would be missed by the Aquarius observations because shear driven mixing has completely removed the near surface stratification before the time of the ascending overpass.

---

**Figure 3.** Simulated diurnal warming $\Delta T_{gw}$ (a,e), freshwater flux P-E (b,f), diffusivity at the base of the diurnal layer $K_d$ (c,g), and diurnal surface salinity change $\Delta S$ (d,h) from a single day at two locations: 5°N 147°E (left) and 2°S 140°W (right). Regimes defining the model mixing parameterization as described in the text are noted as vertical lines.
positive flux due to solar heating resulting in higher values than the corresponding flux due to rain, and the negative buoyancy flux due to evaporation opposes the diurnal cycle. The regression coefficient with 95% confidence interval is 0.92.

The amplitude of the composite day diurnal SSS and SST cycles in both the simulation and buoy observations is quite linear. The relationship between the diurnal cycle plotted against the TAO and RAMA buoy measurements. The regression coefficient tends to be an order of magnitude larger than the mean. This variability is driven by the day-to-day variability in precipitation. In Figure 4, we show the model predictions for the amplitude of the diurnal cycle. There are very few observations of diurnal variations in SSS; however, 15 moorings in the Global Tropical Moored Buoy Array measure near-surface salinity every half hour during the 2007–2009 time period (Table 2).

Table 2. Composite Mean and Standard Deviation (in Parentheses) of Amplitude of Diurnal Cycle of Sea Surface Temperature and Salinity Computed From Tropical Buoy Array Observations and for the Sea Surface Diurnal Cycling Model Prediction Interpolated to the Buoy Locations*  

<table>
<thead>
<tr>
<th>Label</th>
<th>Location</th>
<th>Buoy Observations</th>
<th>SSDC Model Prediction</th>
<th>Correl. Buoy and CMORPH Precip.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>SSS (psu)</td>
<td>SST (°C)</td>
<td>SSS (psu)</td>
</tr>
<tr>
<td>1</td>
<td>0°N 147°E</td>
<td>0.009 (0.229)</td>
<td>0.429 (0.390)</td>
<td>0.028 (0.168)</td>
</tr>
<tr>
<td>2</td>
<td>0°N 156°E</td>
<td>0.011 (0.192)</td>
<td>0.453 (0.341)</td>
<td>0.012 (0.125)</td>
</tr>
<tr>
<td>3</td>
<td>2°N 137°E</td>
<td>0.015 (0.247)</td>
<td>0.456 (0.375)</td>
<td>0.026 (0.137)</td>
</tr>
<tr>
<td>4</td>
<td>2°N 147°E</td>
<td>0.011 (0.273)</td>
<td>0.387 (0.323)</td>
<td>0.025 (0.162)</td>
</tr>
<tr>
<td>5</td>
<td>2°N 156°E</td>
<td>0.007 (0.255)</td>
<td>0.364 (0.329)</td>
<td>0.017 (0.133)</td>
</tr>
<tr>
<td>6</td>
<td>2°S 140°W</td>
<td>0.007 (0.045)</td>
<td>0.307 (0.236)</td>
<td>0.007 (0.016)</td>
</tr>
<tr>
<td>7</td>
<td>2°S 156°E</td>
<td>0.011 (0.220)</td>
<td>0.434 (0.371)</td>
<td>0.017 (0.156)</td>
</tr>
<tr>
<td>8</td>
<td>5°N 137°E</td>
<td>0.014 (0.236)</td>
<td>0.376 (0.376)</td>
<td>0.023 (0.144)</td>
</tr>
<tr>
<td>9</td>
<td>5°N 147°E</td>
<td>0.013 (0.225)</td>
<td>0.308 (0.366)</td>
<td>0.035 (0.193)</td>
</tr>
<tr>
<td>10</td>
<td>5°N 156°E</td>
<td>0.009 (0.244)</td>
<td>0.269 (0.288)</td>
<td>0.029 (0.154)</td>
</tr>
<tr>
<td>11</td>
<td>5°S 95°E</td>
<td>0.006 (0.216)</td>
<td><strong>missing</strong></td>
<td>0.014 (0.104)</td>
</tr>
<tr>
<td>12</td>
<td>5°S 156°E</td>
<td>0.009 (0.306)</td>
<td>0.358 (0.364)</td>
<td>0.029 (0.192)</td>
</tr>
<tr>
<td>13</td>
<td>8°N 137°E</td>
<td>0.005 (0.215)</td>
<td>0.302 (0.305)</td>
<td>0.017 (0.142)</td>
</tr>
<tr>
<td>14</td>
<td>8°N 156°E</td>
<td>0.007 (0.145)</td>
<td>0.232 (0.266)</td>
<td>0.013 (0.122)</td>
</tr>
<tr>
<td>15</td>
<td>1.5°S 90°E</td>
<td>0.008 (0.207)</td>
<td><strong>missing</strong></td>
<td>0.018 (0.120)</td>
</tr>
</tbody>
</table>

*Labels in column one are used in Figures 4 and 5. Last column list correlation between CMORPH precipitation interpolated to the buoy location and the precipitation reported for the buoy rain gauge.

The 2°S 140°W location shown in Figures 3e–h experiences no precipitation on this day. The only freshwater flux comes from a small amount of evaporation throughout the day. In this case, once Regime III begins and the surface layer stabilizes due to warming, the evaporation causes an increase in surface salinity. This increased salinity persists until late afternoon, decreasing slightly at the end of Regime III, and more sharply following the transition to Regime IV, at which point increased mixing causes the surface salinity to approach its nighttime value. The salinity increase at this location has a significantly smaller magnitude than the salinity decrease shown at 5°N 147°E, since the freshwater flux due to evaporation is much smaller than the corresponding flux due to rain, and the negative buoyancy flux due to evaporation opposes the positive flux due to solar heating resulting in higher values of $K_s$ during the morning hours.

### 3. Results

There are very few observations of diurnal variations in SSS; however, 15 moorings in the Global Tropical Moored Buoy Array measure near-surface salinity every half hour during the 2007–2009 time period (Table 2). Thirteen of these moorings are from the TAO/TRITON array and two are from the RAMA array. To estimate typical diurnal cycles for each location, we create composite average days. Finding the difference between the maximum and minimum salinity over the composite day gives the composite diurnal salinity cycle amplitude. We additionally calculate composite diurnal temperature and precipitation cycles.

The amplitude of the composite day diurnal SSS and SST cycles in both the simulation and buoy observations, along with their standard deviations are provided in Table 2. For SST, the mean amplitude of the diurnal cycle is comparable to the standard deviation of the amplitude. However, for SSS the standard deviation tends to be an order of magnitude larger than the mean. This variability is driven by the day-to-day variability in precipitation. In Figure 4, we show the model predictions for the amplitude of the diurnal SST cycle plotted against the TAO and RAMA buoy measurements. The relationship between the model predictions and the observations is quite linear. The regression coefficient with 95% confidence interval is 0.92 ± 0.29, and with an $R^2$ value of 83.2%. This is plotted alongside a one-to-one line. We do not expect the model to fit the one-to-one line, because the moorings measure temperature at 1.5 m depth (except for the mooring at 2°S 140°W, which measures at 1 m depth), while the model predicts a sea surface temperature. Additionally, the presence of the mooring itself may affect the measured temperature—Farrar et al. (2007) found that a buoy created a cool wake downstream, even under low-wind conditions. The observed composite warming maximum always occurs at 16:00 h, while the model predicts the maximum a little earlier in the day, at 14:00 h or 14:30 h. This is consistent with observations that the sea surface warming peak propagates downwards during the afternoon, as the warmed surface mixes with deeper water.
The relationship between the modeled and observed values for the amplitude of the diurnal salinity cycle is shown in Figure 5. The regression coefficient with 95% confidence interval is $1.37 \pm 1.39$, with an $R^2$ value of 32.7%. As for SST, the TAO/RAMA moorings measure salinity at 1.5 m depth (1 m depth for 2°S 140°W). Again, the SSDC predicted amplitude is systematically larger than the buoy measurements. However, for SSS both the relative offset between the model predicted surface values and the buoy subsurface values, and the dispersion about the best-fit line are larger than in the case of temperature such that the regression coefficient interval does not exclude zero and the variance explained is much lower. A number of factors may be contributing to the apparent differences between the simulation skill of surface temperature and salinity. First, the diurnal forcing of precipitation and evaporation only occurs directly at the surface. While a portion of the solar radiation easily penetrates to 1.5 m depth, the freshwater flux accumulated during the day must be transported to depth by diffusive processes. Thus, we might expect stronger salinity than temperature stratification in the diurnal layer, and hence a larger difference between the buoy observations and the true surface values. Additionally, the model only considers one-dimensional vertical processes in predicting the diurnal cycles, and does not account for advection. With no feedback between SSS and surface freshwater fluxes, SSS anomalies may persist longer than warming anomalies leading to a larger error associated with advection in salinity than in temperature. Perhaps most importantly, the accuracy of the gridded satellite precipitation product and how representative it is at the buoy location is likely poorer than are the heat flux components used in the SST prediction, leading to larger discrepancies for SSS than for SST. An indication of this given by the correlations between the CMORPH precipitation rates interpolated to each buoy location and the buoy rain gauge precipitation rate (Table 2). While values differ between locations, generally the CMORPH data set explains less than half the variability of the rain gauge observed precipitation. While this may be due to biases in the CMORPH values, the difference in scale represented by the two measurements likely also plays a role. The satellite-
based product provides quarter-degree average values, while the mooring measures values at a specific location.

In areas that are precipitation-dominated, the diurnal SSS cycle typically shows a freshening in the late morning and early afternoon hours and a salinification during the late afternoon and at night, when shear driven turbulence and convection dominate and low salinity surface water mixes with saltier deeper water. The results shown here are consistent with the analysis of Cronin and McPhaden [1999] for a single mooring. In evaporation-dominated areas, the opposite occurs—evaporation during the day results in a salinity maximum in the afternoon, which freshens overnight. The buoy observations show more variability across locations in when the peak occurs than the model predicts.

Figure 6 shows modeled diurnal variation of surface salinity and temperature and observed values at 1.5 m depth for composite days from four different locations chosen to characterize different forcing regimes and levels of model veracity. The buoy observations are normalized by subtracting the midnight values of temperature and salinity from each day’s measurements prior to averaging, to minimize the impact of longer time-scale variation. 5°N 147°E is an area that is precipitation-dominated, and both the model and the observations show low salinity in the afternoon and a higher salinity overnight. The model predicts a much stronger salinity decline during the day. This is partly attributable to the stronger precipitation estimate in the CMORPH data than observed by the rain gauge on the buoy. Note also that the scale of the composite mean diurnal salinity variation is an order of magnitude smaller than shown in Figure 3 for an individual
day at the same location due to the much stronger than average precipitation on that day. The model-predicted surface warming is also stronger than the observed variation at 1.5 m depth at this location for the reasons described above.

2°S 140°W is an evaporative area. Here, the afternoon salinity is higher than the overnight salinity and the diurnal range predicted by the model is in close agreement with the buoy observations, despite the difference in depth of the measurements. Note that the diurnal salting is plotted on a different scale at this location, as the SSS diurnal cycle caused by day-time precipitation is much larger than the evaporation-driven diurnal cycle.

2°S 156°E is a location where the model predicts a precipitation-dominated region with daytime freshening, while observations show that this location is fresher at night. Finally, buoy observations for 5°S 156°E show a salinity cycle with two peaks. This suggests that there is some stratification occurring during the night. However, the model only predicts one of these peaks, as it does not allow any stratification to form overnight. Also note that this location has one of the highest standard deviations (Table 2) of any of the locations for both the model and observations.

It is also apparent that for all of these locations, observed near-surface temperatures continue to cool overnight reaching a minimum around dawn, while the model does not allow variations between midnight and the time when warming begins. The SSDC scheme does not account for this cooling because the deep convective mixing in Regime I drives the foundation temperature directly and the diurnal layer disappears.

Even in areas with heavy precipitation, rainfall is quite episodic and highly spatially variable. The average diurnal cycle is shaped by days on which there is heavy rain, as the signal due to evaporation is so small. Figure 7 displays a model time series of salting and freshwater flux from the location 5°N 147°E. While the composite average amplitude is less than 0.04 psu, there are many days when the diurnal change exceeds 0.5 psu, and occasional days when it exceeds 1 psu. The correlation coefficient between diurnal SSS change and precipitation is −0.54. While the strongest freshening events coincide with heavy rain, not all heavy rain events result in strong freshening due to other factors such as strong wind that drives mixing of the surface layer. Also shown is the standard deviation of salting over position for each day, calculated from 16 quarter-degree grid cells surrounding the buoy location. This approximate (100 km)² area is representative of the footprint of the Aquarius radiometer. Thus, the subfootprint variation each day is comparable to the day-to-day variation at the buoy location. This result is a further indication of the challenge of comparing insitu and remote sensing observations as discussed above in the context of precipitation.

Figure 8 displays histograms of amplitude and timing of temperature and salinity diurnal variation from the model predictions and observations at 5°N 147°E and 2°S 140°W. In both locations, the observations show a broader distribution of diurnal salinity amplitude than the model predictions. This is possibly due to the advection of salinity anomalies, a process not captured by the model. However, the model’s most extreme values for the daily amplitude (not visible in Figure 8) tend to be larger than the values observed. This may be partially because the observed data is measured below the sea surface, dampening extreme effects. The observations also suggest that the timing of the extrema of diurnal salinity varies slightly more than the model predicts, possibly also due to advection. Table 3 gives the percent of days on which the amplitude of the diurnal salinity change exceeds the given threshold in both observations and SSDC simulation. While the evaporative location is unlikely to have many days on which a diurnal cycle could be observed by the Aquarius radiometer, both the observations and SSDC simulation show a significant number of days at the 5°N 147°E location for which Aquarius could potentially detect the diurnal change.

The global distribution of the predicted mean diurnal SSS amplitude and its standard error are shown in Figure 9 for the composite constructed from all days in 2007–2009, as well as for days with or without precipitation. Days with precipitation show a much stronger diurnal SSS cycle than days without. The salting due to evaporation only creates a very small difference between morning and evening SSS values, while precipitation events create a diurnal SSS signal that is stronger by a factor of 10. This is consistent with the observations at individual buoy locations shown above. The average amplitude on days without precipitation tracks regions of high evaporation, including the northern and southern subtropical oceans. However, while western boundary current regions experience high evaporation, these regions do not have a correspondingly strong diurnal SSS cycle. This may be due to other effects that limit the stabilization of the surface layer, such as high wind conditions. The average amplitude over days with precipitation is strongly
correlated with global precipitation patterns. The diurnal SSS cycle is strongest along the ICTZ and SPCZ, as well as along the western boundary currents. Due to the large response precipitation evokes relative to evaporation, the average amplitude of the diurnal SSS cycle over all days is quite similar to the average over days with precipitation. The standard error is also much larger for days with precipitation. However, the biggest contribution to the standard error across all days is the differences between precipitating and nonprecipitating days.

Figure 7. Time series at 2°N 147°E of (a) modeled salinity diurnal cycle from, (b) net freshwater flux, and (c) standard deviation of salinity over the 16 grid point (100 km × 100 km) neighborhood surrounding the buoy location.
We additionally calculate the largest diurnal SSS amplitudes expected in a year, shown in Figure 10. These values reach 2 psu in regions with heavy rainfall. Values of this magnitude would be easily detected by the Aquarius radiometer, provided they are not masked by contamination by radio frequency interference, background radiation, or other factors mentioned in the introduction.

The timing of the maximum salting amplitude is somewhat later in the day than the maximum diurnal warming, and shows greater spatial variation (Figure 11). This is true whether or not precipitation plays a role, although days with precipitation tend to have maximums that occur earlier in the day. Around 2 pm, the surface layer of water begins cooling, causing the peak warming to occur, however, evaporation and precipitation events may continue later into the afternoon. This allows the salinity peak to sometimes persist until sundown, when convective mixing obscures the changes in salinity at the surface. However, we find that in most locations the peak occurs before sunset, around 2 or 3 pm when shear driven mixing becomes large enough to offset the surface forcing.

The diurnal SSS variation exhibits a seasonal cycle (Figure 12). The amplitude of the diurnal SSS variation tracks the seasonal migration in precipitation, with largest values collocated with the ITCZ and SPCZ. In the seasonal maps, local maxima in diurnal salting become apparent over the midlatitude western boundary currents and their extensions during the summer seasons. The amplitude is expected to be larger in summer when wind speeds and hence vertical mixing are weaker.
4. Discussion

We now return to the interpretation of Figure 1 with the guidance obtained from the SSDC model results presented in section 3. The mean Aquarius ascending minus descending SSS difference of \(\Delta 0.1 \text{ psu}\) under the ITCZ in the eastern Pacific and in the Atlantic are roughly consistent with the amplitudes shown in Figure 9. The eastern Pacific ITCZ signal is stronger in boreal summer (not shown), consistent with the SSDC predictions in Figure 12. Much of the western Pacific warm pool where the SSDC scheme predicts the largest diurnal SSS amplitudes are masked by various quality flags in the Aquarius level 2 data, but the remaining portions of this region do not show large ascending minus descending differences. However, as discussed above, the peak of the diurnal salinity cycle generally occurs around 2 or 3 pm, significantly before the Aquarius ascending overpass. The mean SSDC values of the 6 A.M. to 6 P.M. local time difference

<table>
<thead>
<tr>
<th>Data Set</th>
<th>% Over .2 psu</th>
<th>% Over .5 psu</th>
<th>% Over 1 psu</th>
<th>% Over 2 psu</th>
</tr>
</thead>
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<tr>
<td>2S 140W observed</td>
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<td>0</td>
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<tr>
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<td>6.7908</td>
<td>1.862</td>
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<tr>
<td>5N 147E modeled</td>
<td>11.335</td>
<td>2.468</td>
<td>0.45704</td>
<td>0.091408</td>
</tr>
</tbody>
</table>

Figure 9. Mean amplitude of the diurnal salting (a–c) and standard error of the mean (d–f), divided into days without precipitation occurred (top), days with precipitation (middle), and all days (bottom).
in SSS are about half the size of the mean values of the complete diurnal amplitude. Therefore, the small Aquarius measured diurnal differences in this area cannot be rejected.

On the other hand, the larger positive ascending minus descending differences along the equator in the central Pacific, and in the central subtropical gyres are more difficult to reconcile with the model predictions. As discussed above and shown in Figure 9, net evaporative forcing is not expected to produce as large of a diurnal response as does precipitation, but the Aquarius amplitudes in Figure 1 are more symmetric between negative and positive values. Larger values poleward of 30° latitude seem unlikely to be physical because the surface forcing conditions are not conducive to the formation of a diurnal layer.

In view of the highly episodic nature of diurnal salinity variability, it is perhaps more useful to examine the surface salinity variations as a function of forcing regime rather than geographical position. Meissner et al. [2014c] examined signatures of rain-induced surface freshening in the tropics in Aquarius observations, finding robust relationships between the Aquarius and colocated subsurface salinity difference and simultaneous rain rate or rain rate within the previous hour. They show that the difference between the Aquarius SSS estimate and the Aquarius ancillary salinity provided by HYCOM is a good proxy for near-surface salinity stratification in comparison to estimates derived from colocated Argo buoy or TAO/TRITON measurements. We have repeated this analysis but considered ascending and descending pass observations separately. As shown in Figure 13, the near surface salinity freshening is a decreasing function of simultaneous rain rate, consistent with the results of Meissner et al. [2014c]. For a given rain rate, the freshening is weaker for higher wind speeds.

At low wind speeds (< 3 m/s), the ascending pass surface freshening is slightly stronger than the descending pass surface freshening for any given rain rate. This can be understood in view of the SSDC simulation results described above. At the time of the ascending pass, 18:00 h, there is more likely to be a remnant shallow diurnal warm layer created during daylight hours, while at the time of the descending pass, 06:00 h, a deeper nocturnal boundary layer is more likely to exist. Therefore, under the same precipitation and wind
Figure 11. Time of the maximum of salting (a and b) and warming (c and d), separated into days with precipitation (top) and days without precipitation (bottom).

Figure 12. Seasonal variation of the amplitude of the diurnal salting signal averaged over 2007–2009. (a) Boreal winter. (b) Boreal summer.
conditions, we expect, on average, weaker mixing and stronger freshening at the time of the ascending pass. At slightly higher winds speeds (3 – 9 m/s), there is no significant difference between ascending and descending pass surface freshening for rain rates below about 2 mm/h, while at the highest winds speeds (> 9 m/s) the difference only becomes apparent for very high rain rates. These changes with wind speed are also consistent with the physical processes represented in the SSDC simulations. At higher wind speeds, additional positive buoyancy flux from stronger precipitation is required to suppress the shear-driven vertical mixing before diurnal surface freshening can begin. The highest rain rates are quite rare events, so while the descending pass data show consistently weaker freshening, the results become less statistically robust in the high rain rate limit. It is clear from the results shown in section 3 that the surface freshening is a function of the history of both precipitation and mixing leading up to the time of observation, not just the precipitation and wind speed at that moment. However, segregating the data in this manner does seem to provide a physically meaningful way of interpreting the Aquarius ascending minus descending SSS difference.

As a final application of the SSDC results to the interpretation of Aquarius observations, we evaluate one aspect of the geophysical corrections used in producing the surface salinity estimates. Version 2 of the Aquarius level 2 data removed reflected galactic radiation by using a geometric optics model that represents the ocean surface as an ensemble of reflecting facets [Wentz and LeVine, 2012; LeVine and Abraham, 2004; Meissner et al., 2014a]. This algorithm was developed prior to data collection, and successfully removed about 90% of the reflected galactic radiation. However, the remaining error reaches values of 1.0 psu. This remaining error is thought to be a result of inaccuracies in modeling the ocean surface due to effects such as wind and small waves, as well as other processes that are very difficult to accurately model. Version 3 of the level 2 data incorporates an empirical correction intended to remove the remaining retrieval errors apparent in the ascending-descending differences [Meissner et al., 2014a, 2014b]. An underlying assumption of this empirical correction method is that there are no zonal mean diurnal salinity variations when averaged over a 7 day window. Using the diurnal salinity model, we are able to predict the size of this 7 day zonal mean (Figure 14). While we find physical signals in this zonal mean, they are very small compared to the 0.2 psu uncertainty in the Aquarius measurements. The signal is largest close to the

Figure 13. Surface freshening, defined as the difference between the Aquarius v3.0 level 2, bias adjusted SSS and the HYCOM ancillary SSS versus colocated, simultaneous rain rate from the 8 km/30 minute resolution CMORPH product. The data are segregated by winds speed range (obtained from the Aquarius scatterometer) and ascending and descending passes as shown in the key in the lower left. Thin lines bounding each curve are ± one standard error of the mean. Averages are constructed using data from the beginning of the Aquarius mission (August 2011) through the end of 2013, and within the tropical band equatorward of 15° latitude, excluding values with any quality flag set to either moderate or severe.
equator, and is larger in boreal autumn than in boreal spring. The version 3 correction is unlikely to remove a significant portion of the physical diurnal signal, and thereby facilitate use of Aquarius SSS observations to investigate the associated physical processes.

5. Conclusions

Through the use of a diurnal warming model, predictions are made for the amplitude and variance of a global diurnal SSS cycle. This cycle is driven by freshwater flux in the form of both evaporation and precipitation, with evaporation (precipitation) causing an increase (decrease) in afternoon SSS as solar warming stratifies the surface layer. Comparison with TAO/TRITON and RAMA buoys suggests the model is quite successful in capturing diurnal warming, and while some issues remain in the predictions for diurnal salting there are areas in which the model captures essential features of this cycle.

The strength of the diurnal SSS cycle is correlated with precipitation, which produces a much larger response than evaporation, and in areas with heavy precipitation predicted values frequently are large enough to be observed by the Aquarius satellite, as well as other salinity remote sensing missions. We find broad consistency between the currently available Aquarius ascending minus descending pass salinity differences and the expected diurnal variations in much of the tropics. On the other hand, the differences in the subtropics and higher latitudes appear larger than would be expected from the physical processes represented by the diurnal cycle model. By analyzing the ascending minus descending pass differences as a function of rain rate and wind speed, a consistent picture emerges that Aquarius is able to detect diurnal salinity freshening under conditions favorable to the formation of a diurnal layer. The existence of this cycle and its dependence on precipitation allows for the possibility of exploiting diurnal salinity variations as a proxy for local precipitation as our understanding of the physical processes improves.

The SSDC model, by design, is limited in the physics it represents. This is motivated by the need for computational efficiency in its originally intended application, but perhaps more importantly by the limited observations of the detailed near surface vertical structure of salinity and the processes controlling it. Recent developments in instrumentation such as described in Asher et al. [2014] and Anderson and Riser [2014], observational campaigns such as the recently completed and forthcoming Salinity Processes in the Upper Ocean experiments (http://spurs.jpl.nasa.gov/), along with high-resolution detailed processes models are required to guide further refinements of global models capable predicting the diurnal response of the upper ocean.
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References
Meissner, T., F. Wentz, J. Scott, and K. Hilburn (2014c), Upper ocean salinity stratification and rain freshening in the tropics observed from Aquarius, in Proceedings of IGARSS 2014, Quebec City, Canada. [Available at http://www.rems.com/about/profiles/thomas-meissner/]