On Sea Surface Salinity Skin Effect Induced by Evaporation and Implications for Remote Sensing of Ocean Salinity

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(Manuscript received 23 October 2008, in final form 4 August 2009)

ABSTRACT

The existence of a cool and salty sea surface skin under evaporation was first proposed by Saunders in 1967, but few efforts have since been made to perceive the salt component of the skin layer. With two salinity missions scheduled to launch in the coming years, this study attempted to revisit the Saunders concept and to utilize presently available air-sea forcing datasets to analyze, understand, and interpret the effect of the salty skin and its implication for remote sensing of ocean salinity.

Similar to surface cooling, the skin salinification would occur primarily at low and midlatitudes in regions that are characterized by low winds or high evaporation. On average, the skin is saltier than the interior water by 0.05–0.15 psu and cooler by 0.2–0.5°C. The cooler and saltier skin at the top is always statically unstable, and the tendency to overturn is controlled by cooling. Once the skin layer overturns, the time to reestablish the full increase of skin salinity was reported to be on the order of 15 min, which is approximately 90 times slower than that for skin temperature. Because the radiation received from a footprint is averaged over an area to give a single pixel value, the slow recovery by the salt diffusion process might cause a slight reduction in area-averaged skin salinity and thus obscure the salty skin effect on radiometer retrievals. In the presence of many geophysical error sources in remote sensing of ocean salinity, the salt enrichment at the surface skin does not appear to be a concern.

1. Introduction

Two salinity remote sensing satellite missions are expected to be launched in 2009–10. One mission is the Aquarius/Satelite de Aplicaciones Científicas-D (SAC-D) science mission, developed jointly by the National Aeronautics and Space Administration (NASA) and the Comisión Nacional de Actividades Espaciales (CONAE), the Argentine space agency (Lagerloef et al. 1995, 2008; Koblinsky et al. 2003; Le Vine et al. 2007). The other mission is the Soil Moisture and Ocean Salinity (SMOS) mission, which is a joint Earth observation mission between the European Space Agency (ESA), France, and Spain (Silvestrin et al. 2001). The two missions will provide the first ever global mapping of the complete oceanic sea surface salinity (SSS) field with an unprecedented resolution and coverage that are important for studies of the global water cycle, large-scale ocean circulation, and climate (e.g., Lukas and Lindstrom 1991; Webster 1994; Belkin et al. 1998; Delcroix et al. 2005; Gordon and Giulivi 2008; Riser et al. 2008). However, the SSS missions will not operate as a standalone; they will depend on in situ–sampled SSS to provide ground truth for satellite stability, instrument calibration, and data validation to achieve the desired accuracy of 0.2 practical salinity units (1 psu = 1 g kg⁻¹ salt concentration in seawater) or better for the final monthly mean products (Lagerloef et al. 2008).

Linking satellite SSS retrievals with in situ SSS measurements is challenged by the fact that the two types of SSS measurements are taken at different depths and can be different if vertical salinity gradients exist between the two measurement depths. The topmost salinity sampled by conventional in situ instruments (e.g., moored and drifting buoys, volunteer observing ships, and the automated Argo profiling buoy array) is at a depth of 1 m or more below the sea surface and is commonly referred to as bulk SSS. On the other hand, satellite SSS retrievals represent the salinity in a thin ocean surface layer that is penetrated by electromagnetic radiation. Aquarius/SAC-D and SMOS satellites retrieve SSS from microwave...
brightness temperature of seawater at L band (1.4 GHz; Klein and Swift 1977; Blume et al. 1978; Swift and McIntosh 1983), at which the microwave penetration depth (defined as the depth at which the incoming power density is reduced by a factor of $e^{-2}$) is about 1 cm for seawater at 20°C (Swift 1980). Though the penetration depth is much deeper than the sea surface haline boundary layer, which has a thickness on the order of 200 μm (Katsaros 1980), the exponential decay of the radiation energy after entering the ocean makes the sensor sensitive to the salinity change developed in the sea surface skin layer. Hence, questions arise as to what governs the SSS change in the surface skin layer, how different the skin SSS is from bulk SSS, and whether the skin SSS effect is large enough to affect the accuracy of 0.2 psu that is required for the final salinity product.

The skin–bulk difference in sea surface temperature (SST) has been studied extensively after a cool skin was first observed by Woodcock (1941) and Woodcock and Stommel (1947). The significance of correcting the cool skin for improving the absolute accuracy of satellite measurements of SST has further stimulated research into the physical processes that determine the near-surface thermal structure and the cause of the skin–bulk difference (e.g., Ewing and McAlister 1960; Hasse 1963; Saunders 1967; Liu and Businger 1975; Liu et al. 1979; Katsaros et al. 1977; Katsaros 1980; Paulson and Simpson 1981; Robinson et al. 1984; Schlüssel et al. 1990; Fairall et al. 1996; Wick et al. 1996; Katsaros and Soloviev 2004). The appearance of a cool skin layer is due to (i) the combined cooling effects of net longwave radiation, evaporation, and sensible heat flux from the sea surface to the atmosphere and (ii) the transport of heat by molecular conduction in the skin layer instead of turbulent mixing (Hasse 1971; Katsaros 1980; Schlüssel et al. 1990; Fairall et al. 1996). In the ocean interior, the heat transfer is carried out by turbulent eddies. When approaching the sea surface, the vertical component of turbulent motion is suppressed so that heat is transported primarily by molecular conduction. Because the molecular heat transfer is several orders of magnitude less efficient than that by turbulent mixing, a strong thermal gradient is therefore established across the molecular conduction layer. The stratification causes the surface skin of the ocean, which is in direct contact with the atmosphere, to be cooler than the subsurface water by 0.1°–0.5°C (Wick et al. 1996; Donlon et al. 1999).

Yet, little is known about the near-surface haline structure. Evaporation and precipitation are the two forcing factors that can cause the skin effect to occur, but the two have opposite effects on sea surface water. Evaporation leads to the salt enrichment of the skin layer, whereas precipitation dilutes the skin water. Over the past several decades, there have been attempts to profiling the near-surface salinity structure (e.g., Woodcock 1941; Saunders 1967; Katsaros and Buettner 1969; Soloviev and Vershinsky 1982; Schlüssel et al. 1997), but few instruments had reached the top few centimeters of the oceans because salinity measurements could easily be contaminated by many possible error sources (e.g., sensor calibration accuracy; water sample contamination by rainwater, dirt, and air bubbles; sensor response times; etc.). At present, only rain-induced surface freshening is better observed (Katsaros and Buettner 1969; Ostapoff et al. 1973; Miller 1976; Price 1979; Wijesekera and Gregg 1996). These observations have shown that if winds are light so that wind-induced mixing is weak, a rainfall event can cause a fresh puddle to form at the shallow surface layer and enable a strong vertical salinity gradient to persist for some time (e.g., Soloviev and Lukas 1996; Wijesekera and Gregg 1996; Cronin and McPhaden 1998). By comparison, the salt enrichment of the skin layer under evaporation has never materialized, even though it has long been suggested (e.g., Woodcock 1941; Saunders 1967; Katsaros and Buettner 1969; Soloviev and Vershinsky 1982; Schlüssel et al. 1997). Saunders (1967) in his seminal study of the parameterization of the skin–bulk SST difference theorized that the cool skin layer should also be salty. The evaporation process, which is a key cooling mechanism, releases not only latent heat that cools the surface water but also water vapor that causes a salt enrichment in the skin layer water. However, because of the lack of clear observations, knowledge of the increase of the surface skin salinity by evaporation has not progressed much beyond the Saunders theoretical framework.

Evaporation occurs all the time, as long as air is unsaturated, whereas precipitation occurs sporadically in confined regions. Because of the perseverance of evaporation conditions, the global effect of evaporation on SSS in the molecular skin layer needs to be fully evaluated before likely implications of such effect for SSS remote sensing can be comprehended. Therefore, the objective is to utilize existing skin–bulk parameterization and available air–sea forcing datasets to analyze, understand, and interpret the increase of the skin salinity under evaporation, with the particular focus on its association with the cool skin.

The paper is organized as follows: Section 2 describes the model parameterization of the evaporation-induced skin effect on SST and its application to SSS. Section 3 presents the global simulation of the skin–bulk SSS and SST differences using newly developed air–sea flux datasets. Section 4 discusses the stability of the skin layer, the likelihood of the existence of strong salinity vertical
gradient in the presence of the cool skin, the impact of precipitation on the salty skin, and implications of such skin effect for SSS remote sensing. A summary is included in section 5.

2. Parameterization of the skin–bulk differences in evaporation conditions

a. Model

Much work has gone into investigating the difference between the skin and bulk SST and the predictability of such differences with the application for remotely sensed SST. As a result, two types of parameterizations are available for modeling the temperature difference across the skin layer. One type of parameterization treated the skin layer as a molecular boundary layer where the heat transfer is dominated by molecular conduction and the temperature difference is governed by the thickness of the molecular skin layer (Saunders 1967; Hasse 1971; Grassl 1976; Liu et al. 1979). The first of such models was developed by Saunders (1967), who derived the thickness of the skin layer by using approximations based on shear flows in rigid walls. The other type of parameterization was developed from a different background, assuming that the temperature difference is controlled by the process of surface renewal (Katsaros and Businger 1973; Liu and Businger 1975; Schlüssel et al. 1990; Soloviev and Schlüssel 1996). In this process, the surface water is episodically renewed by bulk water driven by turbulence elements acting on the surface, so that the key parameter governing the magnitude of the skin–bulk SST difference is the time interval between two successive surface-renewal events.

Wick et al. (1996) presented a review of these two types of parameterizations, showing that the Saunders and the surface-renewal-type models are very similar in form, despite the different derivations. They pointed out that the similarity between the two models is due to the similarity of the physical mechanisms taken by each approach: both models assumed that the heat transfer across the air–sea interface is governed by molecular diffusion, as long as there is no wave breaking or spray, and the air–sea interface is intact, with the only difference being whether the water in contact with the air should remain there or be renewed from below. Given the similarity between the two models, either one can serve as a starting point for estimating the salinity difference across the molecular skin layer in evaporation conditions. This study chose the Saunders (1967) parameterization to aid in the analysis.

Saunders (1967) deduced that the cool skin layer is also salty under evaporation conditions. By using dimensional analysis to estimate the skin layer thickness, he related the skin–bulk SST difference (denoted by \( \Delta T = T_{\text{skin}} - T_{\text{bulk}} \)) to the net heat flux \( Q \) and the wind-driven friction velocity, which can be written as

\[
-\Delta T = \lambda \left( \frac{\nu}{k} \right)^{1/2} \left( \frac{z}{\rho} \right)^{1/2},
\]

where \( \lambda \) is a to-be-determined empirical coefficient, \( \nu \) is the kinematic viscosity of seawater, \( k \) is the thermal conductivity of seawater, \( \tau \) is near-surface wind stress, and \( \rho \) is the density of seawater. The upward heat flux \( Q \) in the absence of insolation is given by \( Q = Q_{\text{LH}} + Q_{\text{SH}} + Q_{\text{NLW}} \), where \( Q_{\text{LH}} \) is latent heat flux, \( Q_{\text{SH}} \) is sensible heat flux, and \( Q_{\text{NLW}} \) is the net upward longwave radiation. Given that heat and salt transport in molecular sublayers beneath the ocean surface are governed by similar molecular dynamics (Soloviev and Lukas 2006), Saunders applied the same scaling arguments and obtained the parameterization for the salinity difference between the surface skin and interior (denoted by \( \Delta S = S_{\text{skin}} - S_{\text{bulk}} \)) under evaporation \( E \) at a given salinity \( S \):

\[
\frac{\Delta S}{S} = \lambda \left( \frac{\kappa_T}{\kappa_S} \right)^{2/3} \left( \frac{\rho c_p}{k} \right)^{1/2} \left( \frac{z}{\rho} \right)^{1/2},
\]

where \( \kappa_T \) is the thermal diffusivity, \( \kappa_S \) is the salinity diffusivity, and \( c_p \) is the specific heat of seawater. For simplicity of the analysis, \( -\Delta T \) is used, which is positive when the ocean surface skin is cooler than the subsurface bulk water. However, \( \Delta S \) is always positive because the skin layer is saltier than the subsurface water under evaporation.

b. Combining forced and free convection regimes through modification of \( \lambda \)

Equations (1) and (2) are not valid when there are no winds (i.e., \( \tau = 0 \)). In fact, when wind speed approaches zero, the process that governs the transfer of heat and salt across the molecular skin layer undergoes a regime transition from forced (or shear-driven) convection to free convection (Katsaros 1980). Forced convection arises from shear flow generated by wind shear stress and is the regime from which the Saunders parameterization [Eqs. (1) and (2)] was established. Free convection is driven by buoyancy forces under conditions of upward heat flux and calm winds. The upward heat flux causes a cool skin that is denser than the underlying water; however, evaporation increases the salinity of the cool skin. If winds are light and mixing is weak, the cool and salty skin layer grows in depth by conduction, becomes gravitationally unstable, collapses, and sinks into the warmer and fresher interior water.
Fairall et al. (1996) attempted to incorporate a transition between the free and forced convection regimes through smoothly blending the expressions of the skin layer thickness of the two regimes. In their study, the skin layer thickness in the forced convection regime was the same as the one derived by Saunders, whereas the skin layer thickness in the free convection regime was derived from the Rayleigh number scaling. By blending the two forms of the skin layer thickness, a new expression for \( \lambda \) was derived as

\[
\lambda = 6 \left( 1 + \left[ \frac{16Q_b \rho \alpha c_p \nu^3}{\left( \frac{g}{\rho c_p} \right)^2 k^2} \right]^{3/4} \right)^{1/3},
\]

where \( Q_b \) is the virtual surface cooling that includes the buoyancy effects of salinity resulting from evaporation and is given by

\[
Q_b = Q + \frac{S\beta c_p}{\alpha L_e} Q_{LH},
\]

In Eqs. (3) and (4), \( \alpha \) denotes the thermal expansion coefficient for seawater, \( \beta \) is the salinity contraction coefficient, \( g \) is acceleration due to gravity, and \( L_e \) is the latent heat of vaporization of seawater. For most ocean regions, \( S\beta \) is relatively constant at about 0.026, whereas \( \alpha \) decreases with latitudes from about 3 \( \times \) 10^{-4} in the tropics to near zero in the polar region (Gill 1982). By inserting \( \lambda \) of Eq. (3) into Eq. (1), the Saunders parameterization of \(-\Delta T\) is then applicable to all wind conditions. At a lower wind speed, \( \lambda \) varies with wind speed. Once wind speed increases to more than 3 m s\(^{-1}\), \( \lambda \) approaches a constant value of 6. Robustness of the Saunders parameterization has been examined by many studies. For instance, Kent et al. (1996) evaluated the parameterization schemes of Saunders (1967), Hasse (1971), Schlüssel et al. (1990), and Soloviev and Schlüssel (1994) using SST data from a shipborne radiometer. They found that the Saunders model [Eq. (1)] had the best fit to the observed variations of the skin effect, particularly for wind speeds between 3 and 7 m s\(^{-1}\).

It is worth noting that the new \( \lambda \) formulation (3) was based on the skin layer thickness for molecular heat conduction. The diffusive skin layer, where the salinity skin effect is expected to occur, should be thinner than the conduction skin layer thickness by \((\kappa_s/\kappa_T)^{1/3}\) (Katsaros 1980). Therefore, the \( \lambda \) in formulation (3) should be scaled by \((\kappa_s/\kappa_T)^{1/3} \approx 0.17\) before inserting into Eq. (2) to compute \( \Delta S \) induced by molecular diffusion processes in the skin layer. By doing so, the scaled-down \( \lambda \) approaches a constant value near 1 as wind speed increases to 3 m s\(^{-1}\) or greater.

### 3. Global distribution of skin–bulk differences in temperature and salinity

#### a. Surface forcing data

To compute the skin–bulk differences of salinity (\( \Delta S \)) and temperature (\(-\Delta T\)) using Eqs. (1)–(4), air–sea forcing fields are needed as input. The required air–sea forcing parameters include global evaporation \( E \); near-surface wind stress \( \tau \); and air–sea heat flux components \( Q_{LH}, Q_{SH}, \) and \( Q_{NLW} \). All fluxes were provided by the Objectively Analyzed Air–Sea Fluxes (OAFlux) project at the Woods Hole Oceanographic Institution (Yu et al. 2008), except for \( Q_{NLW} \), which was obtained from the International Satellite Cloud Climatology Project (ISCCP; Zhang et al. 2004). The OAFlux products were computed from the Coupled Ocean–Atmosphere Response Experiment (COARE) bulk flux algorithm version 3.0 (Fairall et al. 2003) using surface meteorology (e.g., wind speed, sea surface and air temperature, and humidity) derived from an objective blending of satellite observations and numerical weather prediction output. The global daily products are gridded on 1° resolution. For the purpose of this study, which focuses on the characterization of global spatial and temporal distribution of the skin effects, daily climatology of the global forcing fields averaged over the 1988–2007 period was compiled and used.

The coefficients of seawater properties required by Eqs. (1)–(4) are given as follows (see also Paulson and Simpson 1981): the thermal diffusivity \( \kappa_T = 1.40 \times 10^{-7} \text{ m}^2 \text{s}^{-1} \), the salinity diffusivity \( \kappa_S = 0.74 \times 10^{-9} \text{ m}^2 \text{s}^{-1} \), the kinematic viscosity \( \nu = 1.14 \times 10^{-6} \text{ m}^2 \text{s}^{-1} \), the thermal conductivity \( k = 0.59 \text{ W m}^{-1} \text{K}^{-1} \), the specific heat of seawater \( c_p = 4.19 \times 10^3 \text{ J kg}^{-1} \text{K}^{-1} \), the density of seawater \( \rho = 1.025 \times 10^3 \text{ kg m}^{-3} \), and the acceleration of gravity \( g = 9.8 \text{ m s}^{-2} \). The latent heat of vaporization \( L_v = (2.501 - 0.00237 \times \text{SST}) \times 10^6 \text{ J kg}^{-1} \), where SST was taken from the OAFlux datasets. The mean sea surface salinity \( S \) in Eq. (2) was taken from the World Ocean Atlas 2005 (Antonov et al. 2006).

Equations (1)–(4) were computed on a daily basis using daily climatological datasets averaged over the 20-yr period from 1988 to 2007. Monthly mean fields were then constructed and serve as the base for the analysis of salinity/temperature skin effects in the following sections. To facilitate the analysis, the averaged 20-yr mean forcing fields of upward heat flux \( Q \), evaporation \( E \), and surface wind speed at the height of 10 m \( w_{10m} \) in February and August are shown in Fig. 1. For simplicity, \( w_{10m} \) is plotted and relates to wind stress by \( \tau^2 = c_d \rho a (w_{10m})^2 \), where \( c_d \) is the drag coefficient computed from the COARE 3.0 bulk flux algorithm (Fairall et al. 2003) and \( \rho a \) is the air density given by 1.2 kg m^{-3}.
Similarity between the mean pattern of $Q$ and that of $E$ is evident. In both fields, maximum values are located in regions associated with such western boundary currents (WBCs) as the Gulf Stream of the North Atlantic, the Kuroshio and Kuroshio Extension of the North Pacific, and the Agulhas Current off the South African coast. The similarity between $Q$ and $E$ is due to the fact that the two air–sea fluxes are not independent: the latent heat released from the ocean surface during the evaporation process is the component that dominates the total heat loss from the ocean. At mid- and low latitudes, $Q_{LH}$ is about an order of magnitude larger than $Q_{SH}$ (Yu and Weller 2007) and about 3–4 times larger than $Q_{NLW}$.

Theoretically, each water molecule that becomes water vapor takes a parcel of heat with it; hence, the strength of evaporation is proportional to the amount of latent heat released from the sea surface. Methodologically, the OAFlux $E$ is estimated from $Q_{LH}$ using the following relation: $E = Q_{LH}/(\rho c_p)$ (Yu 2007). Because evaporation not only cools the sea surface but also enriches the salt concentration in the skin layer, some similarity in global spatial distribution of $\Delta T$ and $\Delta S$ is to be expected.

The $\lambda$ values based on Eqs. (3) and (4) are primarily wind speed dependent and less sensitive to heat flux. To help understand the relationship, the global monthly mean distribution of $\lambda$ is shown in Fig. 2 for every other month starting with February. A constant value of 6 appears over most of the ocean basins where prevailing winds are strong (Fig. 1c). Low $\lambda$ values are identified in three major regions, including (i) the horse latitudes (i.e., the east–west belt between 30° and 35° both north and south of the equator), (ii) the tropical Indian and the western Pacific warm water pools, and (iii) the intertropical convergence

**FIG. 1.** Forcing fields of (a) upward heat flux $Q$, (b) evaporation $E$, and (c) surface wind speed at height of 10 m ($w_{10m}$) for (left) February and (right) August. The monthly mean fields were constructed from OAFlux daily time series from 1988 to 2007.
zone (ITCZ) in the eastern Pacific and Atlantic Oceans at about 10°N. These regions of low \( \lambda \) values have one feature in common: winds are predominantly light. The \( \lambda \) values show strong sensitivity to the seasonally varying winds. For instance, \( \lambda \) in the tropical Indian Ocean is weakest in April (and May, which is not shown) and October (and November, which is not shown), which are the monsoon transition periods that are characterized by weaker lower-level winds. These \( \lambda \) daily fields are inserted directly to Eq. (1) to compute \( -\Delta T \), whereas they are scaled by \( (\kappa_S/\kappa_T)^{1/3} \approx 0.17 \) when applied to Eq. (2) to compute \( \Delta S \).

b. Monthly distribution of \( -\Delta T \)

The global monthly distribution of the skin–bulk temperature differences \( (-\Delta T) \) are displayed for every other month starting with February (Fig. 3). The magnitude of the skin cooling on monthly mean basis ranges between 0.2° and 0.5°C. Weak values of \( -\Delta T \) are found at high latitudes, where high winds are prevalent, whereas larger values of \( -\Delta T \) dominate the low and midlatitudes. The most significant skin cooling is shown in three main regions: (i) the horse latitude belts at 25°–35° north and south of the equator; (ii) the WBC regions of the Northern Hemisphere; and (iii) the warm water pools in the northern Indian Ocean, the western equatorial Pacific, and the coastal region off Mexico.

The factors affecting \( -\Delta T \) have been well documented in the literature (e.g., Saunders 1967; Grassl 1976; Katsaros 1977, 1978; Paulson and Simpson 1981; Kent et al. 1996; Wick et al. 1996; Donlon et al. 1999). Wind speeds are a key factor in determining whether the

FIG. 2. Global distribution of monthly mean \( \lambda \) for (a) February, (b) April, (c) June, (d) August, (e) October, and (f) December. The \( \lambda \) values were computed on daily basis using Eq. (3).
turbulent transports of heat/salt are in the regime of free convection driven by buoyancy, forced convection driven by wind shear stress, or forced convection driven by surface wave breaking. In the case of free convection, the calm winds (no wind or very weak winds) produce very weak turbulent mixing so that buoyancy forces govern the turbulent transports (Katsaros et al. 1977). When the net upward heat flux increases, the surface conduction layer becomes thinner and the cooling of the skin layer temperature intensifies, which leads to significant departure of the skin layer temperature from the interior temperature (Saunders 1967; Katsaros 1977). For winds with moderate magnitude, the turbulent transports are driven by shear stress primarily and by buoyancy forces secondarily. In this case, wind speed affects \( -\Delta T \) by two parallel mechanisms, turbulent mixing and the net upward heat flux, and the two processes have opposite effects. Increased wind speed enhances turbulent mixing and increases latent and sensible heat fluxes and thus the net upward heat flux. Although strong turbulent mixing reduces the magnitude of \( -\Delta T \), large net heat flux increases the surface cooling and causes larger \( -\Delta T \). Hence, the net effect of wind speed on \( -\Delta T \) will have to depend on the relative significance of the two processes (Wick et al. 1996). For very strong winds, the situation is completely changed. Microscale wave breaking in the form of capillary waves and rollers or short gravity waves (Eifler and Donlon 2001) breaks down the ocean skin layer, and no significant temperature gradients exist at the air–sea interface. Disappearance of the thermal skin

![Global distribution of skin–bulk temperature differences (\(-\Delta T\)) for (a) February, (b) April, (c) June, (d) August, (e) October, and (f) December. The monthly mean fields were averaged from daily fields.](image-url)
layer at wind speeds greater than 6 m s$^{-1}$ has been reported (e.g., Konda et al. 1994; Donlon et al. 1999).

The role of wind speed and the net upward heat flux in determining the magnitude of $-\Delta T$ is clearly characterized in Fig. 3. The thermal skin effect is insignificant at high latitudes poleward of 40° north and south of the equator, where winds ($U_{10} > 6$ m s$^{-1}$) are strong. In mid and low latitudes, the dominant mechanism for larger $-\Delta T$ changes with the season and the region. During the boreal fall and winter, the WBC regions boast the strongest net upward heat flux on the global scales. The sufficiently large $-\Delta T$ is caused more by intensive heat loss at the surface of the WBCs and less by the turbulent mixing generated by shear stress. On the other hand, the calm winds over the horse latitudes create a favorable condition for buoyancy-driven convection to occur, resulting in larger $-\Delta T$ in direct response to the net upward heat flux. The free convection regime appears to have also played a key role in maintaining a significant thermal gradient across the thin skin layer over the warm pools of the tropical Indian Ocean, the tropical western Pacific, and the coastal region off Mexico.

c. Monthly distribution of $\Delta S$

The global monthly mean fields for the skin–bulk salinity differences ($\Delta S$) are shown in Fig. 4 for the same six months as in Fig. 3. The monthly mean $\Delta S$ has a magnitude between 0.05 and 0.15 psu and a pattern similar to that of $-\Delta T$. Like the latter, values of $\Delta S$ are small and insignificant at latitudes poleward of 40° north and south of the equator but are sufficiently large in mid- and low latitudes, particularly over the horse latitudes.
between 25° and 35° north and south of the equator; the WBCs of the Northern Hemisphere; and the warm water pools of the tropical Indian Ocean, the western Pacific Ocean, and the coastal waters off Mexico. Over the tropical Indian Ocean, sufficient $\Delta S$ (>0.08 psu) appears mostly in a narrow band centered on the equator and the feature persists throughout the year. Only in April (and May, which is not shown) is the skin salinity effect seen over the basin. A year-round presence of $\Delta S$ is also shown in the western equatorial Pacific, albeit the center of maximum $\Delta S$ migrates slightly over the north and south of the equator with seasons. Prevailing winds over these warm tropical regions are all weak.

The similarity in the global distribution of $\Delta S$ and $-\Delta T$ implies similar mechanisms at work. In addition, the parameterizations for the two skin effects were constructed in a similar fashion [Eqs. (1) and (2)], except that $-\Delta T$ is forced by $Q$ but $\Delta S$ is forced by $E$. As pointed out in section 3a, the latent heat $Q_{\text{LH}}$ released from the ocean surface during the evaporation process is the dominant component of the total heat loss $Q$ from the ocean, and $Q_{\text{LH}}$ and $E$ are exchangeable following the relation $E = Q_{\text{LH}}/(\rho L_v)$ (Yu 2007). The close association between $E$ and $Q$ (Fig. 1) links together the cooling and salinification effects of evaporation on the skin layer. Hence, following the analysis of $-\Delta T$ in the previous section, the factors controlling $\Delta S$ can be summarized as follows: At high latitudes, strong winds break down the skin layer and cause both $\Delta S$ and $-\Delta T$ to disappear. Over the WBC regions, the enhanced evaporation during boreal fall/winter not only cools the skin layer but also causes skin salinity to increase and a larger $\Delta S$ to occur. The calm wind conditions, which occur in the tropical Indo-Pacific warm water pools and along the northern and southern horse latitudes, promote the development of the buoyancy-driven free convection.

The plot of variations of the zonally averaged $-\Delta T$ and $\Delta S$ with seasons (Fig. 5) recapitulates the low and mid-latitudes (40°S–40°N) as the regions of focus when looking for the sea surface skin effects on SST and SSS. Within the regions, the skin effects are most pronounced along the northern (southern) horse latitudes (i.e., 25°–35°) during September–March (March–September) with the maximum occurring in October (June), during which $-\Delta T$ peaks at 0.5°C and $\Delta S$ peaks at 0.1 psu. Considerable skin effects are also noted along the equator from March to November, reflecting primarily the persistence of the skin effect over the western Pacific warm pool.

d. Annual-mean patterns and standard deviations

The annual-mean fields of $-\Delta T$ and $\Delta S$ and the respective standard deviations (STD) based on the 12 climatological months are shown in Fig. 6. Not only do the two annual-mean patterns resemble each other, but the two STD patterns resemble each other as well. Larger values of STD indicate stronger seasonal variations. The regions of larger $-\Delta T/\Delta S$ values also have larger seasonal variability. This is further demonstrated in the plot of the zonally averaged annual-mean $-\Delta T \pm 1\sigma_{\Delta T}$ and $\Delta S \pm 1\sigma_{\Delta S}$ (Fig. 7), where $\sigma_{\Delta T}$ and $\sigma_{\Delta S}$ denote the STD for $-\Delta T$ and $\Delta S$, respectively. The dominance of the thermal and saline skin effects at low and mid-latitudes within 40°S–40°N is clearly seen. In these regions, the averaged mean value of $\Delta S$ is 0.075 psu with a STD of $\pm 0.01$ psu seasonally, whereas the averaged mean value of $-\Delta T$ is 0.30°C with a STD of $\pm 0.08°C$.

4. Discussions

a. Stability of the surface skin layer

Figures 3–7 provided the first-order estimation of the possible occurrence of the cooling and salinification effects of the sea surface skin over the global oceans and suggested the low and midlatitudes, particularly the regions that are characterized by weak winds and high evaporation, as the key focus for such effects. However,
the decrease in surface temperature and increase in surface salinity are surely going to increase the surface density, leading to the destabilization of the upper-ocean stratification and convective mixing of surface waters. Because the cooler and saltier skin on top of the warmer and less salty interior water is always statically unstable and tends to sink, this leads to the question as to what drives the instability: the thermal stratification $\Delta T$ or the haline stratification $\Delta S$.

Sensitivity of the density to changes in temperature and salinity can be analyzed by using the thermal expansion coefficient $\alpha$ and the haline contraction coefficient $\beta$ of seawater, where $\alpha = -(1/\rho)(\partial \rho / \partial T)_{P,S}$ and $\beta = -(1/\rho)(\partial \rho / \partial S)_{P,T}$. The magnitude of $\alpha$ has a strong dependence on the temperature changes: there is a factor of 4 difference between 2.5°C and 30°C (781 $\times$ 10$^{-7}$ to 3413 $\times$ 10$^{-7}$ K$^{-1}$). However, the magnitude of $\beta$ varies only slightly over the range of water masses found in the ocean (8010 $\times$ 10$^{-7}$ to 7490 $\times$ 10$^{-7}$ psu$^{-1}$; Gill 1982). The large contrast between the thermal and saline responses to changes in water properties indicates that in the tropical warm-water province, the surface density flux into the tropical oceans is controlled primarily by changes of temperature (Schmitt et al. 1989).

To examine the relative importance of the cooling versus salinification to the surface density, the density flux ratio (Schmitt 1981; Webster 1994) given by

$$R_p = \frac{-\alpha \Delta T}{\beta \Delta S}$$

is computed (Fig. 8) using the monthly mean values of $\Delta T$ and $\Delta S$ estimated from Eqs. (1) and (2). If $R_p$ is greater than 1, the temperature stratification is a dominant component of the density profile; otherwise, if $R_p$ is less than 1, the salinity stratification is a dominant component. Figure 8 shows that, at the low and mid-latitudes where major skin effects are identified, the density ratio $R_p$ is between 1.5 and 2. In these regions, the surface buoyancy and hence the convective mixing are controlled primarily by the surface cooling and secondarily by the salinity increase. At high latitudes, $R_p$ is less than 1. But because the regional $\Delta T$ and $\Delta S$ are so small to be negligible, the skin layer effects on surface buoyancy are expected to be unimportant.

![Figure 6](image_url)

**Fig. 6.** Annual-mean fields of (a) $\Delta T$ and (b) $\Delta S$ and the monthly standard deviations of (c) $\Delta T$ and (d) $\Delta S$. 
The pattern of the zonally averaged monthly mean $R_r$ values (Fig. 9) is very different from that of $2D_T$ or $D_S$ (Fig. 5). For example, the latter two have pronounced seasonal maxima at the northern and southern horse latitudes, but these localized seasonal intensifications do not appear in $R_r$. It appears that no matter how $2D_T$ and $D_S$ changes with the season, the dominance of the thermal effect on $R_r$ is unchanged, and this keeps $R_r$ in a range of 1.4–1.8 for the regions within $40^\circ S$–$40^\circ N$.

**b. Likelihood of the existence of strong salinity stratification**

It is clear from the previous discussion that the temperature stratification is the leading factor for the onset of the convection (Figs. 8 and 9). This raises three questions. First, what is the role of the saline stratification in the thermally driven instability? Second, when the skin layer is destroyed by convection, how long does it take to reestablish $-\Delta T$ and $\Delta S$? Finally, can a strong salinity gradient exist? Answers to the three questions are important because they determine the impact of the skin layer on remote sensing of ocean salinity.

The work of Katsaros (1969) and a series of subsequent laboratory experiments by her and collaborators (Katsaros 1973, 1976, 1977, 1978, 1980; Katsaros and Buettner 1969; Katsaros and Businger 1973; Katsaros and Liu 1974; Katsaros et al. 1977; Liu et al. 1979; DeCosmo et al. 1996) laid a solid theoretical foundation for understanding the molecular skin layer. From these studies, some interesting insights can be gained into the respective roles of thermal and saline effects in the stability of the skin layer and the time constants (i.e., the time intervals between two successive renewal events) for heat conduction and salt diffusion to reestablish the surface thermal and saline gradients after destruction. In the surface-renewal theory, the molecular skin layer undergoes a cyclic growth and destruction (Katsaros and Businger 1973; Liu and Businger 1975). The surface skin is continually renewed as new parcels of water are brought to the surface by the effects of turbulent mixing. These parcels, once at the surface, are assumed to equilibrate instantaneously. Therefore, the properties of molecular skin layer depend on how fast the surface is renewed (i.e., the time constants).

Katsaros (1969) made visualization of the convection in a saltwater tank and calculated the time constants and the stability limiting values of $-\Delta T$ and $\Delta S$ for a critical Rayleigh number of 600 (see Table 3.4–1 in Katsaros 1969). She came to the conclusion that a strong salinity gradient does not really materialize because of the vast difference in time constants for restoring heat and salt diffusion layers that have different thickness. The difference in layer thickness results in a time constant of $O(15 \text{ min})$ for salt diffusion but a time constant of $O(10 \text{ s})$ for heat conduction. Note that the exact time depends on the Rayleigh number, which determines the skin layer thickness, and also on the molecular coefficients that are used in computation (K. B. Katsaros 2009, personal communication).

Other laboratory experiments (e.g., Ewing and McAlister 1960) showed that the skin layer can re-establish itself within 9–12 s after it is destroyed by a breaking wave. More recent infrared camera measurements have demonstrated that the skin layer has the ability to restore itself in as little as 1 s (Jessup et al. 1997). The range of the thermal skin layer recovery times is due to the dependence of the recovery process on the net heat flux and intensity of the background turbulence (Grassl 1976; Paulson and Simpson 1981; Katsaros 1980; Robinson et al. 1984; Schlüssel et al. 1990, 1997; Wick et al. 1996; Kent et al. 1996; Donlon and Robinson 1997; Zappa et al. 1998). The fast restoration for heat conduction suggests that the thermal

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**Fig. 7. Zonally averaged annual-mean fields of (a) $-\Delta T \pm 1\sigma$ and (b) $\Delta S \pm 1\sigma$, where $\sigma$ is the standardized deviation and defines the upper and lower bounds of the gray shaded areas.**
skin layer is almost always present; indeed, the existence of a cool surface skin both at day and at night has been well observed by satellite radiometers.

By contrast, the long restoration time \( \sim O(15 \text{ min}) \) for salt diffusion, which is about 90 times longer than that for heat conduction, imposes a limitation for a strong salinity stratification to exist. The surface cooling increases the density, and the cooling effect alone could be sufficient enough to cause the Rayleigh number to increase to the critical value, at which convective mixing sets in but the salinity gradient is not yet at its maximum intensity. The latter affects the magnitude of \( \Delta S \). In particular, when averaging \( \Delta S \) over an area, there would be a reduction in the average value of \( \Delta S \). Given that the radiation received from a footprint is averaged over an area to give a single pixel value, the reduction in the averaged \( \Delta S \) would make the skin effect less effective.

Katsaros (1978) illustrated that salt affects how quickly the density of the uppermost layer increases enough to become unstable. Because of this, salinity stratification is regarded as a limiting factor for the growth of the skin cooling (Katsaros 1969). Laboratory experiments by Hasse (1963) showed a cutoff at 1°C for \( -\Delta T \), even for a strong evaporation rate. The satellite-sensed \( -\Delta T \) has a mean value of 0.3°C with root-mean-square variability of up to 0.4°C (Donlon et al. 1999), because the instantaneous value depends on heating–cooling and surface wind conditions (Wick et al. 1996; Donlon and Robinson 1997). For these \( -\Delta T \) values that are commonly observed, the cool skin would remain stable if referring to Table 3.4–1 in Katsaros (1969). Surface cooling

![FIG. 8. Density flux ratio](image-url)
c. Possible effects of precipitation

Precipitation has effects opposite than evaporation; it makes the surface layer fresher, reduces surface density, and helps to stabilize the surface layer (e.g., Katsaros and Buettner 1969; Katsaros 1976; Wijesekera and Gregg 1996; Schlüssel et al. 1997; Cronin and McPhaden 1999). Precipitation also differs from evaporation in intensity, duration, and spatial extension. Precipitation occurs sporadically in localized areas and its duration varies, whereas evaporation takes place at all times and over all regions, as long as the air is not saturated. Wijesekera and Gregg (1996) observed that during extreme episodic rainfall events in the tropical western Pacific warm pool, precipitation can exceed evaporation by more than 1 m and the surface layer in the top 2–3 m can be freshened by several psu. When this occurs, the salty skin layer, which has a maximum magnitude not exceeding 0.15 psu (Fig. 4), would be wiped out easily. A question thus arises as to how the global distribution of $\Delta S$ could be affected by precipitation.

Precipitation is pronounced in low latitudes, especially within the ITCZ, the warm pools, and the monsoon regions. As shown in Figs. 4–6, $\Delta S$ with sufficient magnitude tends to occur in the horse latitudes at 25°–35° north and south of the equator, the WBCs of the Northern Hemisphere, and the warm water pools. It is apparent that some of the key regions of precipitation are also the regions favorable for larger $\Delta S$. To see how much the two sets of regions overlap on the global scale, the monthly mean distribution of $\Delta S$ is superimposed onto the total number of days that rainfall would normally occur on a monthly basis (Fig. 10). The precipitation data are the NASA Tropical Rainfall Measuring Mission (TRMM) global daily merged analysis (Huffman et al. 2007) from 1998 to 2007. The number of precipitation days was calculated for each calendar month and then averaged over the TRMM 10-yr period. The calculation was based on the TRMM domain from 40°S to 40°N. Given that the salinity skin effect is negligible for latitudes poleward of 40° north and south of the equator, the TRMM domain will not affect the impact assessment of the precipitation. It should be noted that, on a monthly basis, the intensity of precipitation is generally proportional to the total number of days of precipitation. To facilitate the discussion, Fig. 11 plots the global annual-mean fields of precipitation, evaporation, and their differences, which shows that precipitation exceeds evaporation over most tropical oceans and at high latitudes, whereas evaporation exceeds precipitation in the extratropical regions.

Precipitation is frequent year-round along the ITCZ and over the warm waters of the equatorial eastern Indian and the western Pacific Oceans, with the maximum occurrence (more than 25 days per month) during June–October (Fig. 10). The precipitation in these regions is also intense (Fig. 11); its mean magnitude is about 3 times larger than that of evaporation (Adler et al. 2003; Yu 2007). Under extreme episodic rainfall events, the difference in magnitude could be even larger (Wijesekera and Gregg 1996). The predominance of precipitation over evaporation along the ITCZ and the Indo-Pacific warm pools implies that a salty skin layer would not be a long-lasting feature, because it would be washed away whenever it rains. On the other hand, the salty skin could exist along the horse latitudes between 25° and 35° north and south of the equator, where evaporation forcing is predominant and precipitation is infrequent and light (Figs. 10, 11). The salty skin layer may also be featured in the western and northern tropical Indian Ocean during boreal winter and spring, the time that enhanced evaporation is featured.

d. Implication for remote sensing of ocean salinity

The sensitivity of sea surface $T_B$ to SSS is near the maximum at L band (1.413 GHz). However, even at this optimal frequency band, the sensitivity of $T_B$ to SSS is low: 0.5 K psu$^{-1}$ for SST of 20°C and down to 0.25 K psu$^{-1}$ for SST of 0°C, both at nadir incidence. This has presented a great technical challenge for SSS remote sensing, because the magnitude of the radiometric sensitivity to SSS is comparable to the effects induced by many geophysical parameters, such as sea
surface roughness, SST, the Faraday rotation (i.e., the change in direction of polarization of microwave emissions as the radiation passes through the ionosphere), solar radiation, and atmospheric gases (Yueh et al. 2001; Koblinsky et al. 2003; Le Vine et al. 2007). Likely error contributions from these geophysical parameters to SSS retrievals are listed in Table 1 of Koblinsky et al. (2003), which shows that surface roughness (e.g., waves), with error size estimated to be about 0.27 psu, is the largest error source. The magnitude of the errors resulting from other parameters ranges between 0.01 and 0.13 psu, dominated mostly by the effects of Faraday rotation and solar radiation.

The design of Aquarius has taken into account the correction of geophysical effects (Le Vine et al. 2007). Three efforts are particularly noted. The foremost important correction is the sea roughness induced by winds. Given that the radar backscatter of sea surface is sensitive to sea roughness but nearly insensitive to SST and SSS, Aquarius will fly a scatterometer to provide independent information of surface roughness to make the needed correction. The scatterometer at 1.26 GHz will operate at nearly the same frequency as the radiometer, share the same antenna feed, and look at the same pixel with approximately the same footprint (Wilson et al. 2001). The second potential source of error
is Faraday rotation, which can be corrected by using polarimetric measurements (Yueh et al. 2001; Le Vine and Abraham 2002). Aquarius will include a polarimetric channel and will use the measured third Stokes parameter and an algorithm suggested by Yueh (2000) to retrieve the angle of polarization rotation. Finally, Aquarius will avoid reflection of solar radiation from the ocean surface into the main beam of the antenna by flying in a 0600–1800 LT equatorial-crossing, sun-synchronous orbit with the antenna beams pointing toward the nighttime side of the orbit.

Compared to the geophysical errors in SSS remote sensing, the effect of the molecularly diffusive skin layer in evaporation conditions does not appear sufficiently large to give rise to serious concern when validating satellite SSS retrievals with in situ–sampled SSS measurements. The salt enrichment with a mean magnitude of 0.075 psu and STD of ±0.01 psu is identified (Figs. 6 and 7) in regions of weak winds (e.g., the northern and southern horse latitudes and the tropical warm pools) or large evaporation (e.g., the WBC in the Northern Hemisphere), but the salty skin is always associated with cooling, which is unstable and tends to sink. Contrary to the recovery time of \( \sim O(10 \text{ s}) \) for \( \Delta T \), the long restoration time \( \sim O(15 \text{ min}) \) needed for a full development of \( \Delta S \) after the destruction of the skin layer would reduce the chance of the existence of strong salinity stratification. As the radiation received from a footprint is averaged over an area to give a single pixel value, the long restoration time would cause a reduction in area-averaged \( \Delta S \), making the skin effect less effective. It appears that there is little point in seeking to implement a correction for the salinity skin effect under evaporation; any improvement that might be made could be easily obscured by geophysical effects. Whether the salinity skin effect could become a comparable source of error once the performance of SSS measurement sensors meets the prescribed accuracy is not yet known.

5. Summary and conclusions

The existence of a cool and salty sea surface skin under evaporation conditions was first proposed by Saunders in 1967, but few efforts have since been made to perceive the salt component of the skin layer. With two salinity missions scheduled to launch in the coming years, this study attempted to revisit the Saunders concept and to utilize presently available air–sea forcing datasets to analyze, understand, and interpret the effect of the salty skin and its implication for remote sensing of ocean salinity.

This study found that the evaporation-induced salification would occur primarily at low and mid-latitudes in regions characterized by low winds or high evaporation, and it found that the salty layer is always accompanied by a cooling. On average, the skin layer is saltier than the interior water by 0.05–0.15 psu and is cooler by 0.2°–0.5°C. The cooler and saltier skin on top of the warmer and less salty interior water is always statically
unstable and tends to overturn. Calculation of the density flux ratio \( R_T = -\alpha \Delta T/\beta \Delta S \) suggested that the surface buoyancy is controlled primarily by cooling and secondarily by the salinity increase.

The three issues, the role of the salinity in thermally driven instability, the restoration times needed to re-establish \(-\Delta T\) and \(\Delta S\) at a full scale after the destruction of the skin layer, and the likelihood of the existence of strong salinity stratification, were reviewed with reference to previous studies, particularly to the work of Katsaros (1969) and a series of subsequent studies by her and collaborators. Surface cooling and salt enrichment both contribute positively to the increase of the density of the surface water. Given that the instability threshold can be reached with a lesser degree of cooling in the presence of a higher salt concentration, a sensible inference appears to be that salinity may be the factor for limiting strong cooling to develop (Katsaros 1969). The time constants needed to restore a full increase of \(-\Delta T\) and \(\Delta S\) after the destruction of skin layer were calculated by Katsaros (1969, Table 3.4–1). Because of the difference in layer thickness for heat conduction and salt diffusion, Katsaros (1969) estimated that it takes \(\sim O(10 \text{ s})\) to establish \(-\Delta T\) but \(\sim O(15 \text{ min})\) to reset \(\Delta S\): the latter is 90 times longer than the former. Because the radiation received from a footprint is averaged over an area to give a single pixel value, it is deduced that the slow recovery process by salt diffusion would cause a reduction in the value of \(\Delta S\) when averaged over an area and obscure the salty skin effect on radiometer retrievals. In the presence of many geophysical error sources in SSS remote sensing, the salt enrichment of the sea surface skin under evaporation conditions does not appear to be a concern. Therefore, there is little point to correct the skin effect under evaporation; any improvement that might be made could be easily overshadowed by the geophysical effects. Whether the salinity skin effect could become a comparable source of error, once the performance of SSS measurement sensors meets the prescribed accuracy, is not yet known.

This study also examined the possible effect of precipitation on the evaporation-induced salt enrichment. Some of the regions of sufficient \(\Delta S\), such as the warm water pools of the equatorial eastern Indian Ocean, western Pacific Ocean, and off the coast of Mexico, are found to overlap with the regions of frequent and intense precipitation. The magnitude of the evaporation-induced \(\Delta S\) appears much weaker than the surface freshening of several psu in the case of heavy rainfall events. It appears that the salty skin layer may not be a persistent feature in regions where precipitation dominates evaporation, whereas the skin effect may occur in regions such as the western tropical Indian Ocean and the horse latitudes in the sub tropics where evaporation dominates precipitation.

It should be noted that the heat and salt transfers at the molecularly diffusive skin layer are affected by many variables, including not only wave action, wind speed, temperature of the atmosphere and seas but also conditions of surface water (e.g., the phytoplankton-generated surfactants). Surfactants can reduce evaporative heat transport at free surfaces and thus complicate the processes governing \(-\Delta T\) and \(\Delta S\). The effect of surfactants is expected to be most pronounced under low wind speed conditions. It should also be noted that the salt enrichment at the skin layer is just one mechanism that could cause the skin–bulk SSS differences. Other factors, such as the diurnal cycle and precipitation, could be a source of the deviation between satellite and in situ SSS, with work ongoing.

Acknowledgments. This study was supported by the Remote Sensing Science for Carbon and Climate program sponsored by the National Aeronautics and Space Administration under Grant NNX07AF97G. The author was grateful to Dr. Kristina Katsaros for providing thoughtful and constructive review of the manuscript, for suggesting that the difference in time constants for heat and salt diffusion may limit the formation of strong salinity stratification, and for forwarding a scanned copy of the relevant chapter of her Ph.D dissertation. The comments from an anonymous reviewer were also appreciated. Support from the NOAA/Office of Climate Observation (OCO) and NASA Ocean Vector Wind Science Team (OVWST) in developing air–sea heat fluxes and surface wind stress used in this study are gratefully acknowledged. The flux datasets are available from the OAFlux project Web site (available online at http://oaflux.whoi.edu/).

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