Evaluation of turbulent fluxes at the ocean surface using surface renewal theory

C. A. Clayson
Department of Earth and Atmospheric Science, Purdue University, West Lafayette, Indiana

C. W. Fairall
NOAA Environmental Technology Laboratory, Boulder, Colorado

J. A. Curry
Program in Atmospheric and Oceanic Sciences, University of Colorado, Boulder

Abstract. An internally consistent model is presented that can be used to determine the ocean surface fluxes of heat, moisture, and momentum, given bulk sea surface temperature and atmospheric temperature, humidity, and wind measured at a single level within the atmospheric surface layer. This model is based upon surface renewal theory as described by Brutsaert [1975a]. Liu et al. [1979] (hereinafter referred to as LKB) made partial use of this theory, and further improvements to the LKB parameterization have been made by Fairall et al. [1996a]. The present model includes the following improvements relative to the LKB and Fairall et al. bulk flux models: incorporation of a new time-scale parameterization for surface renewal, inclusion of capillary waves in the surface roughness model, derivation of the surface roughness scales of water vapor and heat based solely upon surface renewal theory, and incorporation of a new surface skin temperature model. The model is validated using shipborne observations of surface fluxes and surface meteorology that were obtained in the central Pacific Ocean, the western tropical Pacific, the subtropical Pacific, and the midlatitude North Atlantic. Comparisons of model results with covariance fluxes of latent heat show biases of less than 3% for all locations, with little dependence of error on wind speed; similar results are obtained for sensible heat and momentum flux. An assessment is given of the advantages of the present scheme over the LKB and Fairall et al. schemes. The model results are interpreted in the context of the physical processes involved in determining the surface roughness length and the surface renewal timescale.

1. Introduction

Determination of the ocean surface fluxes of heat, fresh water, and momentum is critical for understanding and modeling air-sea interactions. The surface wind stress is a major driving force of surface waves, and the heat and fresh water fluxes determine the buoyancy flux into the upper ocean. An oft-quoted number is that an accuracy of 10 W m\(^{-2}\) in the net surface heat flux into the ocean is required to accurately force an ocean model [e.g., Webster and Lukas, 1992]. In regions where horizontal processes are inefficient in transporting heat and where shallow mixed layers are commonly found, such as the western tropical Pacific, vertical processes drive changes in sea surface temperature and upper ocean heat content, which are very sensitive to the local surface heat flux. Seager et al. [1988] and Gent [1991] have shown that a change of 10 W m\(^{-2}\) in the net surface heat flux of the tropical western Pacific can change the modeled sea surface temperature by 1°C over a 1-year period. Boundary layer modeling, wave modeling, and remote sensing all depend upon the surface momentum flux.

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Significant differences in the response of a coupled ocean-atmosphere model based on varying determinations of surface turbulent fluxes were reported by Geernaert [1987].

The direct measurement of the turbulence fluxes by eddy correlation techniques and inertial dissipation methods requires measurements of high-frequency atmospheric turbulence properties. Such measurements are scarce and sensitive to low-frequency platform motions induced by the ocean. Bulk aerodynamical methods are therefore an attractive alternative to determining the turbulent surface fluxes. Bulk aerodynamical methods employ standard meteorological observations of surface air temperature, humidity, and wind speed, and sea surface temperature, humidity, and current speed, all of which can be routinely measured by ships, buoys, and satellites. The fluxes can then be computed from the meteorological variables if the bulk transfer coefficients are known. However, significant uncertainties still remain in the determination of the bulk transfer coefficients (for a review see Blanc [1987] and Geernaert [1990]).

Previous research into the value and dependence of the drag coefficient has yielded results which differ by as much as a factor of 3 [for reviews see Wu [1969, 1980], Smith [1970], Garratt [1977], Blanc [1985], Geernaert et al. [1986], and Geernaert [1990]]. In a review of 10 different drag coefficient schemes, Blanc [1985] determined differences in calculated
wind stresses of up to 60%. Much of the difference between the various parameterizations of drag coefficient has been due to differing stability regimes between the studies [Geernaert et al., 1986], and to the neglect of the dependence of the drag coefficient on the sea state [Geernaert et al., 1988]. The values of typical mean variations in the sensible heat fluxes between 10 different coefficient schemes was over 30% [Blanc, 1985], with large differences between latent heat fluxes as well.

An alternative bulk method for determining surface turbulent fluxes involves modeling the physical processes that drive the interfacial sublayers of the ocean and atmosphere. In the surface interfacial layers, water vapor and heat fluxes are controlled by molecular diffusion, while momentum is transferred also by pressure forces. One model for describing the physics governing the interfacial layer was described by Bruus (1965, who postulated that transfer at the air-sea interface is by molecular diffusion into internal Kolmogorov-scale eddies, which are renewed intermittently after random times of contact with the evaporating surface. The heat surface flux models described by Liu et al. [1979] (hereinafter referred to as LKB) and Fairall et al. [1996a] have made partial use of surface renewal theory. The LKB model assumes that the log profile in the surface layer can be matched to an exponential profile in the sublayer. The Fairall et al. algorithm differs from the LKB algorithm in that it uses (1) a different specification of the roughness-stress relationship, (2) a gustiness velocity to account for the additional flux induced by boundary-layer-scale variability, (3) profile functions obeying the convective limit, and (4) some adjustments of constants specifying the relationship between the scalar and velocity transfer coefficients.

A new model to determine fluxes at the air-sea interface is described here. The model used in this work takes its inspiration from the work by Bruus (1965, 1975a), incorporating surface renewal theory. Improvements of this model relative to the LKB and Fairall et al. [1996a] models include incorporation of a new timescale parameterization for surface renewal, inclusion of capillary waves in the surface roughness model; derivation of the surface roughness scales of water vapor and heat based solely upon surface renewal theory, and incorporation of a new surface skin temperature model. The model is valid only for nonprecipitating conditions and for wind speeds that are low enough to preclude sea spray. The model is validated using shipborne observations of surface fluxes and surface meteorology that were obtained in the central Pacific Ocean, the western tropical Pacific, the subtropical Pacific, and the midlatitude North Atlantic. An assessment is given of the advantages of the present scheme over the LKB and Fairall et al. schemes. The model results are interpreted in the context of the physical processes involved in determining the surface roughness length and the surface renewal timescale.

2. Description of the Data Sets

Surface fluxes and meteorological measurements were obtained from ships during four separate experiments: (1) the Tropical Instability Wave Experiment (TIWE), November - December, 1991, central Pacific [Sherman et al., 1993]; (2) the Atlantic Stratocumulus Transition Experiment (ASTEX), June 1992, North Atlantic [White et al., 1995]; (3) the Tropical Ocean-Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (TOGA COARE), November 1992 to February 1993 (three legs), tropical western Pacific [Young et al., 1995, Fairall et al., 1996a]; and (4) the San Clemente Ocean Probing Experiment (SCOPE), September 1993, off California [Fairall and Edson, 1994]. Average meteorological values and measured fluxes for each of these experiments are presented in Table 1 (the notation is described in a separate notation section). The meteorological values reported in Table 1 correspond to heights of 10 m for wind speeds in TIWE, ASTEX and SCOPE, and 15 m for COARE, and \(T_a\) and \(q_v\) were measured at 15 m for TIWE and COARE, at 21 m for ASTEX, and at 11 m for SCOPE.

A complete description of the observing system employed during these cruises and their accuracy is given by Fairall et al. [1996a]. Mean and perturbation wind and temperature measurements were made using a sonic anemometer. A dual-wavelength infrared hygrometer was used to measure both mean and perturbation humidity. Sea surface temperature was measured using a thermistor scaled in the top of a floating hose, measuring the temperature at a depth of approximately 5 cm. Data collected from the sonic anemometer, infrared hygrometer, and sea surface thermistor were used to calculate the surface sensible and latent heat fluxes. In this study we use only the flux values determined using the eddy correlation method. The accuracies for these cruises are believed to be similar, although the eddy covariance fluxes measured during SCOPE from the R/P FLIP should be more accurate because of the relative insensitivity of this platform to ship motions.

To reduce the possibility for errors, we have eliminated observations under precipitating conditions, ship maneuvers, or unfavorable wind direction relative to the ship orientation (following Fairall et al. [1996a]). The values used in this study represent 30-min averages. Table 2 summarizes the measurement accuracies for TOGA COARE as estimated by Bradley et al. [1996].

<table>
<thead>
<tr>
<th>Variable</th>
<th>TIWE</th>
<th>ASTEX</th>
<th>COARE</th>
<th>SCOPE</th>
</tr>
</thead>
<tbody>
<tr>
<td>(N )</td>
<td>39</td>
<td>151</td>
<td>822</td>
<td>193</td>
</tr>
<tr>
<td>(u_0)</td>
<td>6.9</td>
<td>5.2</td>
<td>3.9</td>
<td>4.0</td>
</tr>
<tr>
<td>(T_a^0)</td>
<td>27.2</td>
<td>20.7</td>
<td>29.2</td>
<td>19.1</td>
</tr>
<tr>
<td>(T_a^0)</td>
<td>26.7</td>
<td>19.7</td>
<td>27.9</td>
<td>16.5</td>
</tr>
<tr>
<td>(q_v^0)</td>
<td>22.1</td>
<td>14.9</td>
<td>25.0</td>
<td>13.4</td>
</tr>
<tr>
<td>(q_v^0)</td>
<td>17.0</td>
<td>10.8</td>
<td>18.2</td>
<td>10.2</td>
</tr>
<tr>
<td>(H)</td>
<td>1.5</td>
<td>5.5</td>
<td>6.4</td>
<td>14.6</td>
</tr>
<tr>
<td>(LE)</td>
<td>107.8</td>
<td>72.0</td>
<td>96.0</td>
<td>50.7</td>
</tr>
<tr>
<td>(\tau)</td>
<td>0.033</td>
<td>0.038</td>
<td>0.033</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Cruise acronyms are identified in text. \(N\) is number of samples.
Table 2. TOGA COARE R/V Moana Wave Measurement Accuracies as Estimated by Bradley et al. [1996]

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>50-min rms</th>
<th>Bias</th>
</tr>
</thead>
<tbody>
<tr>
<td>u</td>
<td>m s⁻¹</td>
<td>0.3</td>
<td>±0.2</td>
</tr>
<tr>
<td>T, day</td>
<td>K</td>
<td>0.3</td>
<td>±0.2</td>
</tr>
<tr>
<td>T, night</td>
<td>K</td>
<td>0.2</td>
<td>±0.1</td>
</tr>
<tr>
<td>q</td>
<td>g kg⁻¹</td>
<td>0.3</td>
<td>±0.2</td>
</tr>
<tr>
<td>Tₘ</td>
<td>K</td>
<td>0.1</td>
<td>±0.2</td>
</tr>
<tr>
<td>H</td>
<td>W m⁻²</td>
<td>±20%</td>
<td>±2</td>
</tr>
<tr>
<td>E</td>
<td>W m⁻²</td>
<td>±20%</td>
<td>±1.4</td>
</tr>
<tr>
<td>τ</td>
<td>N m⁻²</td>
<td>0.015±30%</td>
<td>0.002</td>
</tr>
</tbody>
</table>

3. Surface Renewal Theory

Determination of the turbulent fluxes involves modeling the physical processes that drive the interfacial sublayers of the ocean and atmosphere, which are controlled by molecular diffusion only. The interfacial sublayers have been studied mainly in the aqueous region, where the existence of a strong temperature gradient across the sublayer was first examined by Ewing and McAlister [1960]. Subsequent descriptions of the nature of the aqueous sublayer are provided by McAlister and McLeish [1969], Wu [1971, 1984, 1985], Khunzhua and Andreyev [1974], Liu and Businger [1975], McLeish and Pudtland [1975], Katsaros et al. [1977], Katsaros [1977, 1980], and Wick [1995].

To formulate a theoretical model for the water vapor transfer coefficient at a rough surface, Brutsaert [1965] postulated that at the air-sea interface the transfer is by molecular diffusion into internal Kolmogorov-scale eddies which are renewed intermittently after random times of contact with the evaporating surface. The turbulence in the interfacial sublayer can thus be parameterized using a timescale, referred to as the surface renewal timescale \( t_s \), which can be used to determine the water vapor and heat flux within the interfacial sublayer (a complete description of the notation used and values of constants is given in the section noted).

Brutsaert [1975a] suggested that \( t_s \) is equivalent to the timescale of the Kolmogorov eddies, i.e.,

\[
    t_s \propto \left( \frac{v_{z_0}}{u_*} \right)^{1/2}
\]

where \( z_0 \) is the surface roughness length and \( u_* \) is the friction velocity. This form of \( t_s \) was used also by LKB. A different renewal timescale parameterization was developed by Soloviev and Schlussel [1994], in which different renewal timescales were adopted for low, moderate, and high wind speed regimes.

Wick [1995] has established a relationship for the surface renewal timescale that incorporates both shear-driven and free convection timescales. His relationship for the surface renewal time is

\[
    t_s = t_{\text{shear}} \left( t_{\text{conv}} - t_{\text{shear}} \right) \exp \left( \frac{R_{f_c}}{R_{f_e}} \right)
\]

where \( t_{\text{shear}} \) is the shear-driven timescale, \( t_{\text{conv}} \) is the convective timescale, \( R_{f_c} \) is a critical value of the surface Richardson number, and \( R_{f_e} \) is the surface Richardson number [Kudryavtsev and Soloviev, 1985]. This surface Richardson number determines the transition between free convection, in which the cyclic nature of replacement of the water in the sublayer is driven by convection, and forced convection, in which the surface renewal is driven by viscous surface stress variations associated with rollers on breaking wavelets [Soloviev and Schlussel, 1994]. Based upon examination of available sea “skin” surface temperature measurements and flux data, Wick [1995] has determined that the transition between the shear-driven and free convection regimes is smooth and that also, the convective timescale should apply only as the wind speed nears zero and \( R_{f_e} \) becomes infinite; the shear-driven timescale should dominate however at other wind speed regimes. Based on these considerations, the exponential relationship of (2) satisfies these requirements and also matches best with available data. The shear-driven timescale is given by

\[
    t_{\text{shear}} = C_{\text{shear}} \left( \frac{v_{z_0}}{u_*} \right)^{1/2}
\]

and the convective timescale is given by

\[
    t_{\text{conv}} = C_{\text{conv}} \left( \frac{v_{z_0}}{u_*} \right)^{1/2}
\]

where \( C_{\text{shear}} \) and \( C_{\text{conv}} \) are empirically determined constants. The variable \( U_{ni} \) represents the sum of the latent and sensible heat fluxes plus the radiative fluxes (for details, see Wick [1995]). A radiative transfer model is used to distribute the shortwave radiation within the upper ocean, so the only portion of solar radiation absorbed in the upper millimeter of the ocean is included in \( U_{ni} \). Examination of (1) indicates that the original \( t_s \) determined by Brutsaert [1975a] was effectively a shear-driven timescale.

In the context of surface renewal theory, the rate of molecular diffusion of water vapor normal to the interface is given by [e.g., Brutsaert, 1965]

\[
    E = \frac{\rho_e c_p}{t_s} \int_0^\infty \exp \left( -\frac{t}{t_s} \right) \left( \frac{\partial q}{\partial z} \right)_{z=0} \, dt
\]

where \( E \) is the rate of evaporation as a mass flux per unit area of the interface into eddies of all ages. A similar expression may be written for the sensible heat flux density at the interface:

\[
    H = \frac{\rho_e c_p \kappa}{t_s} \int_0^\infty \exp \left( -\frac{t}{t_s} \right) \left( \frac{\partial T}{\partial z} \right)_{z=0} \, dt
\]

where \( H \) is the corresponding sensible heat flux density.

Integration of (5) and (6) requires evaluation of the gradients of humidity and temperature at the interface. Using the diffusion equations and assuming that an eddy in contact with a rough surface is temporarily stagnant [e.g., Brutsaert, 1975a], we can determine the gradients at the surface to be

\[
    \left( \frac{\partial q}{\partial z} \right)_{z=0} = (q_0 - q_s) \left( \varepsilon \pi t \right)^{1/2}
\]

\[
    \left( \frac{\partial T}{\partial z} \right)_{z=0} = (T_0 - T_s) \left( \kappa \pi t \right)^{1/2}
\]
where the subscript $s$ denotes values at the interface and the subscript $h$ denotes values in the interfacial layer. Utilization of these expressions allows the integrals in (5) and (6) to be evaluated:

\[ E = \rho_a \left( \frac{q_s - q_h}{\mathcal{K}} \right) \left( \frac{c}{t_*} \right)^{1/2} \]

\[ H = \rho_a c_p \left( \frac{T_s - T_h}{\mathcal{K}} \right) \left( \frac{K}{t_*} \right)^{1/2} \]

(8)

Following Brutsaert [1975a], dimensionless numbers can be defined for the evaporative and sensible heat transfer. The interfacial Dalton number $Da_0$ and the interfacial Stanton number $St_0$ can then be defined as:

\[ Da_o = \frac{K^{1/2}}{\mathcal{K}^{1/2}} \left[ \frac{c_{\text{shear}}}{\mathcal{K}} \left( \frac{v z_o}{u_*} \right)^{1/2} \right] + \left[ \frac{c_{\text{conv}}}{\mathcal{K}} \left( \frac{v z_o}{u_*} \right)^{1/2} \right] \left[ \frac{v z_o}{u_*} \right] \exp \left( \frac{R_{e_0}}{R_{f_0}} \right) \right]^{1/2} \]

(9)

\[ St_0 = \frac{K^{1/2}}{\mathcal{K}^{1/2}} \left[ \frac{c_{\text{shear}}}{\mathcal{K}} \left( \frac{v z_o}{u_*} \right)^{1/2} \right] + \left[ \frac{c_{\text{conv}}}{\mathcal{K}} \left( \frac{v z_o}{u_*} \right)^{1/2} \right] \left[ \frac{v z_o}{u_*} \right] \exp \left( \frac{R_{e_0}}{R_{f_0}} \right) \right]^{1/2} \]

(10)

These equations can be compared to other empirical and theoretical parameterizations of the interfacial Dalton and Stanton numbers, such as those given by Owen and Thomson [1963], Yaglom and Kader [1974], and Brutsaert [1975b] for rough surfaces. As was noted above, the Brutsaert [1975b] model has the same form as the shear part of the above Dalton number. On the basis of this result, for this model we will take $C_{\text{shear}}$ to be 7.32. The empirical constants determined by Wick [1995] were for the oceanic sublayer, and thus we would expect these numbers to vary for the atmospheric sublayer. However, we take the proportion between the shear and the convective constants to remain the same for the atmosphere, giving us a $C_{\text{conv}}$ value of 0.8.

A crucial element for determining the surface heat and evaporative fluxes correctly is the determination of $\varepsilon$. A number of possibilities exist for determining this roughness length. We employ the parameterization developed by M.A. Bourassa et al. [A stress parameterization including shear due to short waves, submitted to J. Atmos. Sci., 1996] for open ocean conditions:

\[ \varepsilon = \begin{cases} \frac{b \sigma}{\mu \rho_{\omega}} + 0.48 \frac{u^2}{w_{\omega}}, & \text{for } u_w > c_p \min \\ 0.11 \frac{\nu}{u_*}, & \text{for } u_w < c_p \min \end{cases} \]

(12)

where $b=0.019$ is an empirically derived dimensionless constant, $\sigma$ is the surface tension (which depends on sea surface temperature), $w_{\omega}$ is the wave age, $\nu$ is the molecular viscosity, and $c_p \min$ is the minimum phase speed for water waves which determines whether surface waves exist. The minimum phase speed is defined as a function of the surface tension and the density of water and is approximately equal to 23.2 cm s$^{-1}$.

Determination of these parameters is based on a sea state parameterization that has a non-arbitrary wave age; wave age is determined from the wind speed and the sea state model which describes perturbations from a condition of local wind-wave equilibrium using standard meteorological observations [Bourassa et al., A sea state parameterization for low wind speeds and a non-arbitrary wave age, submitted to J. Phys. Oceanogr., 1996]. This parameterization is a significant improvement over previous models of the sea state because of the strong influence of wave age on the shape and size of the modeled waves. This model includes effects of capillary as well as gravity waves in the determination of the surface roughness. The necessity of including these effects in the description of the surface roughness length has been outlined by Blanc [1985], Geernaert et al. [1986, 1988], and Wu [1994].

4. Formulation of the Surface Flux Model

Equation (8) is difficult to use in determining the surfaces fluxes of heat and moisture, because measurements of temperature, humidity, and winds are rarely made in the interfacial layer. A more practical approach is to devise a surface flux formulation that can utilize conventional meteorological observations that are made in the atmospheric surface layer. LKB combined Monin-Obukhov similarity theory with a surface renewal model to evaluate the surface turbulent fluxes, with further improvements to the LKB model described by Fairall et al. [1996a]. The present model adopts this general approach but returns to the Brutsaert [1975a] formulation of the surface evaporative and sensible heat flux and does not utilize the temperature profile in the ocean interfacial layer described by Liu and Businger [1975].

Using Monin-Obukhov similarity theory, the turbulent fluxes of momentum ($\tau$), sensible heat ($H$), and moisture ($E$) are defined as:

\[ \tau = \rho_a u^2 \]

\[ H = -\rho_a c_p u_* T_* \]

\[ F = -\rho_a u_* q_* \]

(13)

where $T_*$, $q_*$, and $u_*$ are the Monin-Obukhov similarity scaling parameters for temperature, water vapor mixing ratio, and
horizontal wind. The fluxes here are defined such that a flux in to the ocean is positive; flux out of the ocean is negative. The surface latent heat flux is obtained from the product of the evaporative flux and the latent heat of vaporization (LE). In the surface layer, the following profiles for velocity, temperature and humidity have been determined empirically [e.g., Businger, 1979]

\[
\frac{T - T_e}{T_o} = \frac{Pr_l}{k} \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_T \right]
\]

\[
\frac{q - q_o}{q_o} = \frac{Sc_l}{k} \left[ \ln \left( \frac{z}{z_{eq}} \right) - \Psi_q \right]
\]

\[
\frac{u - u*}{u*} = \frac{1}{k} \left[ \ln \left( \frac{z}{z_0} \right) \right. \Psi_u
\]

where \( k \) is the von Karman constant, and \( Pr_l \) and \( Sc_l \) are the turbulent Prandtl and Schmidt numbers, respectively. The terms \( z_0, 5q \) and \( 5T \) represent the surface roughness lengths for momentum, moisture and heat, respectively. The terms \( \Psi_u, \Psi_q, \) and \( \Psi_T \) represent the respective dimensionless stability functions. The form of the dimensionless stability functions used in this study follow Beijaars and Holtslag [1991] for stable conditions

\[
\Psi_u = a \left( \frac{z}{z_0} \right)^b + b \left( \frac{z}{z_0} \right)^{3/2} \exp \left( -d \left( \frac{z}{L} \right) \right)
\]

\[
\Psi_{T,q} = \left( 1 + \frac{z}{z_0} \right)^{3/2} + b \left( \frac{z}{z_0} \right)^{3/2} \exp \left( -d \left( \frac{z}{L} \right) \right)
\]

and Benoit [1977] for unstable conditions:

\[
\Psi_u = \ln \left( \frac{z}{z_0} \right) + \left[ \frac{(\zeta^{1/3} + 1)(\zeta + 1)}{(\zeta^{1/3} + 1)(\zeta + 1)} \right] + 2 \tan^{-1}(\zeta) \tan^{-1}(\zeta_0)
\]

\[
\Psi_{T,q} = \ln \left( \frac{z}{z_0} \right) + 2 \ln \left( \frac{\zeta_0 + 1}{\zeta + 1} \right)
\]

where

\[
\zeta = \left( 1 - 16 \frac{z}{L} \right)^{1/4}; \quad \xi = \left( 1 - 16 \frac{z_{eq}}{L} \right)^{1/4}
\]

\[
\lambda = \left( 1 - 16 \frac{z}{L} \right)^{1/2}; \quad \lambda_0 = \left( 1 - 16 \frac{z_{eq}}{L} \right)^{1/2}
\]

and \( L \) is the Obukhov length. The Beijaars and Holtslag [1991] stability functions are preferred over the usual stable function, since this has been shown to produce realistic results within a weather forecast model [Louis, 1979]. The Beijaars and Holtslag [1991] formulation is derived by using the budget constraints on buoyancy destruction and shear production and has been shown to be more consistent with critical Richardson number constraints than other formulations [Beijaars and Holtslag, 1991].

To determine the surface turbulent fluxes using measurements of atmospheric temperature, water vapor mixing ratio, and wind speed in the atmospheric surface layer, while at the same time incorporating surface renewal theory, continuity in \( q \) and \( T \) is assumed at the top of the interfacial sublayer, and the surface layer is matched to the interfacial layer. The matching for the water vapor equation is shown here; the heat transfer equation is similar. Using (14), the surface layer profile can be written as

\[
\frac{q - q_o}{q_o} = \frac{q - q_{eq}}{q_{eq}} + \frac{q_{eq} - q_o}{q_o} = \frac{Sc_l}{k} \left[ \ln \left( \frac{z}{z_{eq}} \right) - \Psi_q \right]
\]

where \( h \) is the height of the interfacial layer. The profile in the atmospheric surface layer is given by:

\[
\frac{q - q_{eq}}{q_{eq}} = \frac{Sc_l}{k} \left[ \ln \left( \frac{z}{h} \right) - \Psi_q \right]
\]

where

\[
\ln \left( \frac{z}{h} \right) = \ln \left( \frac{z}{z_0} \right) - \ln \left( \frac{h}{z_0} \right)
\]

and \( h \) is the height of the interfacial sublayer. Incorporating (9), (17) and (18) into (16), we have

\[
\frac{Sc_l}{k} \left[ \ln \left( \frac{z}{z_{eq}} \right) - \Psi_q \right] = \frac{Sc_l}{k} \left[ \ln \left( \frac{z}{z_0} \right) - \ln \left( \frac{h}{z_0} \right) - \Psi_q \right] - D a_o^{-1}
\]

We can now solve for the roughness length for water vapor, \( z_{eq} \)

\[
z_{eq} = z_0 \exp \left[ k \left( \frac{u_h}{u_s} - \frac{1}{Sc_l D a_o} \right) \right]
\]

The quantity \( (u_h/u_s) \) is here taken to be equal to 5 [Brutsaert, 1975a]. Similarly, the roughness length for heat becomes:

\[
z_{eq} = z_0 \exp \left[ k \left( \frac{u_h}{u_s} - \frac{1}{Pr_l S t_{eq}} \right) \right]
\]

The quantity \( (u_h/u_s) \) is here taken to be equal to 5 [Brutsaert, 1975a], thus assuming that the boundary between the two layers lies at the level where \( U(\hat{u}*+5) = 5 \).

Using equations (10), (11), (12), (14), (20), and (21), we have a closed system of eight equations and eight unknowns (\( D a_o, S t_{eq}, u_s, T_s, q_{eq}, \varphi, z_{eq} \) and \( z_{eq} \)), which are solved iteratively. Values of \( T_s, u_s, \) and \( q_{eq} \) are then used in (13) to solve for the surface fluxes of momentum, heat and moisture, using measurements obtained in the atmospheric surface layer and at the ocean surface. Specific measurements required as input into the model are \( T_s, q_{eq}, u_s \), and either \( T_s \) or the ocean mixed layer temperature.

The surface fluxes should be evaluated using the surface skin temperature, since it is the ocean skin that interacts directly with the atmosphere. The skin temperature differs from the temperature of the ocean just below the surface (-1 cm) by -0.1°C to +0.1°C, due to the heat loss/gain experienced by this thin sublayer. For temperatures measured by ships (5 m depth) or buoys (0.5 m depth), the difference from the skin temperature can be even larger. Webster et al. [1996] calculated that for average conditions during TOGA COARE, a 1°C error in SST would result on average in an error in sensible heat flux of 2.4 W m⁻² (23% of the average value) and an error in latent heat flux of 18.7 W m⁻² (16% of the average value). From these numbers it can be inferred that errors exceeding 10 W m⁻² in the surface heat flux can be made by using a bulk value of sea surface temperature rather than the skin temperature.
Table 3. Comparison of Model Results With Covariance Latent Heat Flux

<table>
<thead>
<tr>
<th></th>
<th>TIWE</th>
<th>ASTEX</th>
<th>COARE</th>
<th>SCOPE</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Bias</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>-0.3</td>
<td>5.2</td>
<td>-3.7</td>
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<tr>
<td>Fairall</td>
<td>6.3</td>
<td>7.1</td>
<td>-0.7</td>
<td>4.8</td>
</tr>
<tr>
<td>LKB</td>
<td>59.0</td>
<td>41.8</td>
<td>45.9</td>
<td>28.3</td>
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<td><strong>Root-Mean-Square-Error</strong></td>
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<td>20.7</td>
<td>16.3</td>
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<td>9.5</td>
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<tr>
<td>Fairall</td>
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<td>LKB</td>
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<td></td>
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<tr>
<td>Model</td>
<td>0.61</td>
<td>0.86</td>
<td>0.90</td>
<td>0.96</td>
</tr>
<tr>
<td>Fairall</td>
<td>0.61</td>
<td>0.85</td>
<td>0.90</td>
<td>0.96</td>
</tr>
<tr>
<td>LKB</td>
<td>0.62</td>
<td>0.83</td>
<td>0.88</td>
<td>0.84</td>
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Fairall, Fairall et al. [1996a]; LKB, Liu et al. [1979].

Although the skin temperature is measured from satellite, skin temperatures are rarely available from ship or buoy measurements. Instead, measurements are taken of “surface” temperature at depths ranging from a few centimeters to a meter or more. In fact, most drag coefficients are “tuned” to the use of a bulk sea surface temperature. However, the skin temperature is the correct temperature to use in evaluating the surface fluxes. Thus corrections need to be made to the bulk temperature to obtain a true surface skin temperature measurement. For temperatures obtained at depths greater than a few centimeters, substantial corrections due to daytime solar heating need to be included [e.g., Fairall et al., 1996b]. However, for temperatures measured at depths of a few centimeters, only a skin-bulk temperature difference model needs to be applied. An approach to determining this temperature difference based on surface renewal theory has been taken by LKB, Soloviev and Schluesel [1994], and Wick [1995]. LKB showed on the basis of surface renewal theory, that the temperature difference AT between the ocean skin and bulk temperatures should be related to the net surface heat flux QN, surface friction velocity, and the surface roughness Reynolds number Re. This theory has been extended by Soloviev and Schluesel [1994], who determined the following for the temperature difference between the ocean skin and bulk temperatures:

$$-\Delta T = \frac{Q_N}{\rho_w c_p} \left( \frac{u^*}{K} \right)^{1/2}$$  \hspace{1cm} (22)

Further details on this parameterization, including validation of the parameterization, are described by Soloviev and Schluesel [1994] and Wick [1995]. Soloviev and Schluesel [1996] have examined the evolution of the cool skin during daytime. During the daytime, absorption of solar radiation within the thermal molecular sublayer of the ocean can modify the temperature difference across the cool skin. Under low wind speed conditions, the solar heating damps the convective instability, strongly increasing the renewal time. During strong insolation, the skin temperature may be warmer than the true bulk temperature. Under low wind conditions, convective instability caused by salinity flux due to evaporation limits the surface temperature increase. Once $T_s$ has been determined, a value of $q_s$ is determined following Fairall et al. [1996a]

$$q_s = 0.98 q_{sat}(T_s)$$  \hspace{1cm} (23)

where $q_{sat}$ is the saturation vapor pressure. This expression accounts for the reduction in vapor pressure associated with a salinity of 34%. $T_s$ is not given, (22) is solved iteratively with (10), (11), (12), (14), (20), and (21).

Table 4. Comparison of Model Results With Covariance Sensible Heat Flux

<table>
<thead>
<tr>
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<th>ASTEX</th>
<th>COARE</th>
<th>SCOPE</th>
</tr>
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</tr>
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<tr>
<td>Model</td>
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<td>4.4</td>
<td>4.2</td>
<td>4.3</td>
</tr>
<tr>
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<td>2.4</td>
<td>4.1</td>
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<td>4.4</td>
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<tr>
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<td>0.88</td>
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<td>0.85</td>
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<td>0.67</td>
<td>0.83</td>
<td>0.76</td>
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</table>

Table 5. Comparison of Model Results With Covariance Momentum Flux

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<th>COARE</th>
<th>SCOPE</th>
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<td></td>
<td></td>
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</tr>
<tr>
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<td>-0.006</td>
<td>0.004</td>
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<td>0.006</td>
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<tr>
<td>Model</td>
<td>0.031</td>
<td>0.031</td>
<td>0.028</td>
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<tr>
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<td>0.038</td>
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<tr>
<td>Fairall</td>
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<td>0.72</td>
<td>0.96</td>
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<tr>
<td>LKB</td>
<td>0.75</td>
<td>0.83</td>
<td>0.71</td>
<td>0.96</td>
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5. Comparisons of Flux Parameterizations With Data

In this section the model results are compared with the eddy correlation fluxes obtained during TOGA COARE (referred to in the figures and tables as COARE), ASTEX, TIWE, and SCOPE. Hourly-averaged values of the observations are used. The results from the present model are also compared to results obtained using the LKB and Fairall et al. [1996a] models. It is noted here that the surface temperatures in these databases were measured at a depth of 1 cm; thus (22) is used to evaluate $T_s$.

Comparison statistics of all three models with each of the four data sets are presented in Tables 3, 4, and 5 (bias, root mean-square error, and correlation). The bias is defined as the
modeled value minus the observed value. The relatively poor performance of the LKB model for the sensible and latent heat fluxes arises partly because of the values chosen for the ratios of the turbulent diffusivities of momentum, heat, and moisture; changing these values would result in an improved performance. The performance of the new model and that of the Fairall et al. model are comparable, both showing very good agreement with the covariance fluxes. The biases of the sensible and latent heat fluxes for the new model are within the stated accuracy of the measurements (Table 2), except for ASTEX. The biases of the modeled latent heat flux for the Fairall et al. algorithm are slightly lower than for the new model for the COARE data, although the biases for the Fairall et al. model are larger than the new model for the other data sets. Since the Fairall et al. model was tuned to the COARE data, we expect that particular bias to be small. The modeled values of the momentum flux for both the new model and Fairall et al. model exceed slightly the measurement accuracy. Perhaps the most stringent test of the model accuracies is given by comparison with the SCOPE data, where the measurement accuracies are the highest. The correlations for both the new model and Fairall et al. model are the highest for this date set, with values of 0.96 for both the latent heat and momentum fluxes. The bias and root-mean-square error of the latent heat flux determined from the new model are slightly better than for the Fairall et al. model.

Figure 1 shows comparisons of the new model with covariance fluxes, for all four data sets combined. The mean biases of the combined data set are latent heat flux: -1.6 W m$^{-2}$, sensible heat flux: 0.3 W m$^{-2}$, momentum flux: -0.04 N m$^{-2}$. Median biases are: -0.2 W m$^{-2}$, 0.6 W m$^{-2}$, and 0.005 N m$^{-2}$. Much of the scatter in Figure 1 is associated with sampling uncertainty in the covariance measurement. It is seen from Figure 1 that the model shows good agreement with the observations over the entire flux regime. The poorest agreement between the model and observations is for some of the lowest flux values, where the modeled values are lower than observed.

A special area of concern is the reliability of the model at low wind speeds, where fluxes have been shown to have unusual characteristics (e.g., Large and Pond, 1982; Bradley et al., 1991; Greenhut and Khalsa, 1995). Additionally, a concern exists at the higher wind speeds because of the possible influence of sea spray. A comparison of the model bias with respect to wind speed is shown in Figure 2. As can be seen from these figures, the model performs well at these lower wind speeds, and the biases in latent heat flux, sensible heat flux, and momentum flux are all nearly totally uncorrelated with wind speed. At the highest wind speeds ($u > 9$ m s$^{-1}$), a bias is seen in the sensible heat flux. Because of the small sample size at the higher wind speeds, it is difficult to interpret the results. The Fairall et al. [1996a] algorithm also performed well for low wind speeds for the TOGA COARE data set (as cited by Fairall et al. [1996a]), although the good agreement was accomplished by specifically tuning the gustiness parameterization at low wind speeds against the TOGA COARE observations.

**Figure 1.** Modeled surface fluxes versus observed fluxes determined from covariance data, for the complete data set: (a) latent heat flux, (b) sensible heat flux, (c) momentum flux.
6. Sensitivity Studies

In order to gain some understanding as to the relative importance of the improvements that have been made to the new model, a number of sensitivity studies were performed with the goal of understanding the role the new physics plays in influencing the surface fluxes. Specifically, we examine the sensitivity to surface roughness length and surface renewal timescale.

The effects of changing the surface roughness length and surface renewal timescale are evident primarily at lower wind speeds. Therefore comparisons are made for wind speeds less than 3 m s$^{-1}$. The comparisons are illustrated only for the latent heat flux. Figure 3 shows the model bias for the new model latent heat flux as a function of wind speed, for $u < 3$ m s$^{-1}$. Note that Figure 3 is essentially the same as Figure 2a, except for the range of wind speeds examined. For the low wind speed values shown in Figure 3, the model biases are latent heat flux, $-3.1$ W m$^{-2}$; sensible heat flux, $0.7$ W m$^{-2}$; and momentum flux, $0.002$ N m$^{-2}$. It is noted here that the ship measurements of wind are believed to be accurate to with $0.3$ m s$^{-1}$ [Bradley et al., 1996].

In the LKB and Fairall et al. models, a Charnock plus smooth flow $z_0$ parameterization [Smith, 1988] is used. The primary difference between the Charnock formulation and BVW formulation (BVWa, b) used in the new model is the incorporation of capillary waves in the Bourassa et al. formulation of $z_0$. Capillary waves are of importance primarily in the wind speed range 1.2 to 4 m s$^{-1}$. Neglecting capillary waves reduces the values of the roughness length in this wind speed range, and thus the modeled fluxes. Figure 4 shows the model biases for $u < 3$ m s$^{-1}$ when the value of $z_0$ from Charnock [1955], which is a function of $u_*, g$, and $z_0$, is substituted for the value of $z_0$ described by BVW. Comparing the results from Figure 4 with Figure 3 shows a strong bias for the modeled latent heat fluxes determined using Charnock's model for $z_0$, with the model underestimating the observations. The biases of the modeled fluxes using the Charnock $z_0$ for wind speeds less than 3 m s$^{-1}$ are latent heat flux, $-19.9$ W m$^{-2}$; sensible heat flux, $-1.3$ W m$^{-2}$; and momentum flux, $-0.002$ N m$^{-2}$. The consistent model underestimate of the fluxes at low wind speeds using the Charnock $z_0$ arises from the model producing values of $z_0$ that are too small, which results in values of $z_0Q$ and $z_0T$ which are too small, resulting in an underestimation of the heat fluxes.
7. Conclusions

A new bulk model that incorporates surface renewal theory to determine fluxes at the air-sea interface has been described. The recently developed surface renewal timescale described by "Wick [1995] has been used to determine new expressions for the interfacial Dalton and Stanton numbers. The surface roughness model described by RVW has been incorporated into the model. To make use of conventional meteorological observations in determining the surface fluxes, the surface renewal model is combined with Monin-Obukhov similarity theory by matching the solutions at the top of the atmospheric interfacial layer. Improvements of this model relative to the LKB and "Fairall et al. [1996a] models include incorporation of a new timescale parameterization for surface renewal, inclusion of capillary waves in the surface roughness model, derivation of the surface roughness scales of water vapor and heat based solely upon surface renewal theory, and incorporation of a new surface skin temperature model.

The new model was compared with ship observations of fluxes determined from the eddy correlation method that were obtained in the central Pacific Ocean, the western tropical Pacific, the subtropical Pacific, and the midlatitude North Atlantic. It is only recently that a significant body of accurate surface fluxes over the ocean has become available. The new model compared very well with observations, with the biases for the most part within the limits of accuracy of the observations. The model showed substantially better agreement with the observations than did the LKB model. The "Fairall et al. [1996a] model also compared very well with the observations, particularly for the TOGA COARE data set obtained in the tropical western Pacific, although the new model overall has slightly better comparison statistics than the Fairall et al. model.

Sensitivity studies were conducted to examine the influence of the new surface renewal timescale and surface roughness length parameterizations. These new parameterizations influence the fluxes primarily at low wind speeds, through the addition of capillary waves to the surface roughness model and convective timescale to the surface renewal timescale. When capillary waves are not included in the surface roughness model, the modeled fluxes at low wind speeds are too low. When the convective renewal timescale is not included, the modeled fluxes at low wind speeds are also too low. By including the convective renewal timescale and the improved surface roughness model, the new model performs well under conditions of low wind speeds, without tuning.

The appeal of the new model over the LKB and Fairall et al. models is that the new model is more firmly grounded in surface renewal theory. The new values of the surface renewal timescale and surface roughness length improve the model fluxes, particularly at low wind speeds, and tuning is not required. The parameterization for surface skin SST is also consistent with surface renewal theory, using the same surface renewal timescale used in the flux model. Unlike the Fairall et al. model, a "gustiness" factor is not needed to increase the fluxes at low wind speeds.

This model does not include the effects of whitecapping and sea spray, which are large contributors to the latent heat flux at higher wind speeds (above 15 m s⁻¹) [Andreas et al., 1995], and the applicability of the surface renewal approach to the atmospheric surface layer has been questioned by Wu [1992] at wind speeds above 7 m s⁻¹. The validation results suggest that
the model may be valid up to wind speeds of 12 m s$^{-1}$ although the sample size at wind speeds greater than 8 m s$^{-1}$ is too small to be conclusive. It should also be noted that the effects of precipitation are ignored in the model. The role of precipitation in forcing the ocean surface is generally restricted to its impact on surface fresh water flux. However, Gosnell et al. [1995] and Fairall et al. [1996a] have highlighted the importance of the sensible heat flux associated with rain. Additionally, rain imparts a momentum flux to the ocean associated with both the horizontal motion of the rain and the gravitational fall speed of the rain [e.g., Caldwell and Elliot, 1971; Manton, 1973]. The presence of rain significantly complicates the physics occurring at the ocean surface and surface sublayer, since rain effectively destroys the ocean skin and this sublayer. Rain also modifies the surface roughness length by damping the longer-wavelength surface waves and increasing the capillary waves [e.g., Caldwell and Elliot, 1971]. Extension of the surface flux model to include the effects of sea spray and precipitation will be the subject of future work.

Notation

\begin{align*}
\rho_a & \quad \text{density of air} \\
\rho_w & \quad \text{density of water} \\
\sigma & \quad \text{surface tension} \\
v & \quad \text{kinematic viscosity} \\
\end{align*}

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References


C. A. Clayson, Department of Earth and Atmospheric Science, 1397 Civil Building, Purdue University, West Lafayette, IN 47907-1397. (email: clayson@thunder.atmos.purdue.edu)

J.A. Curry, Program in Atmospheric and Oceanic Sciences, University of Colorado, Boulder, CO 80302. (curryja@cloud.colorado.edu)

C.W. Fairall, NOAA Environmental Technology Laboratory, R/EW/PJ, 325 Broadway, Boulder, CO 80303. (email: cwf@etl.noaa.gov.)

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