A prognostic scheme of sea surface skin temperature for modeling and data assimilation

Xubin Zeng
Department of Atmospheric Sciences, University of Arizona, Tucson, Arizona, USA

Anton Beljaars
European Centre for Medium-Range Weather Forecasts, Reading, UK

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[1] A prognostic scheme is derived for the computation of sea surface skin temperature in weather forecasting, four-dimensional data assimilation, and ocean-atmosphere coupled modeling. This scheme is then tested using the in situ data over tropical and midlatitude oceans. By implementing this scheme into the ECMWF model, the diurnal variation of sea surface temperature as measured by the geostationary satellite can also be reproduced. **Citation:** Zeng, X., and A. Beljaars (2005), A prognostic scheme of sea surface skin temperature for modeling and data assimilation, *Geophys. Res. Lett.*, *32*, L14605, doi:10.1029/2005GL023030.

1. Introduction

[2] In atmospheric data assimilation, weather forecasting, and atmospheric modeling, the term sea surface temperature (SST) usually refers to the (five-day to monthly) product of blended satellite retrievals and in situ measurements at a depth of a few centimeters to a few meters from buoys and ships [*Reynolds and Smith*, 1994]. In oceanic and ocean-atmosphere coupled modeling, the term SST refers to the mean temperature of the top ocean layer of about 10 meters in depth. Numerous studies [*Fairall et al.*, 1996] have demonstrated that these temperatures are significantly different from the sea surface skin temperature (T_s).

[3] Several approaches have been proposed for determining T_s. *Fairall et al.* [1996] developed separate models for the cooling skin and the warm layer effects. *Clayson and Curry* [1996] and *Gentemann et al.* [2003] developed empirical formulas to estimate the diurnal T_s based on atmospheric conditions (e.g., wind and solar insolation). *Zeng et al.* [1999] derived a theoretical relationship to estimate the diurnal T_s from wind speed and the diurnal variation of bulk temperature measured by buoys. However, these approaches are less suitable for modeling and operational data assimilation. The warm layer model of Fairall et al. is not rigorous because the simple heat and momentum integrals are not handled in a conservative fashion. The shape of the diurnal T_s is fixed in the work by Clayson and Curry and Gentemann et al., while the algorithm of Zeng et al. requires the information of diurnal bulk temperature a priori. In an attempt to develop a T_s scheme for forecasting models, *Beljaars* [1997] reformulated the diagnostic relations of *Webster et al.* [1996] as a prognostic equation for T_s. The purpose of this paper is to develop a new prognostic T_s scheme for weather forecasting, climate modeling, and data assimilation.

2. A Prognostic T_s Scheme

[4] The one-dimensional heat transfer equation in the ocean can be written as

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}(K_w + k_w)\frac{\partial T}{\partial z} + \frac{1}{\rho_w c_w}\frac{\partial R}{\partial z} \tag{1}$$

where the subscript w refers to sea water, T is the sea water temperature and z is the depth defined as positive upward, ρ_w and c_w are the density and volumetric heat capacity of sea water respectively, K_w and k_w are the turbulent diffusion coefficient and molecular thermal conductivity respectively, R is the net solar radiation flux defined as positive downward.

[5] In the oceanic molecular sublayer with depth δ, K_w and $\frac{\partial T}{\partial t}$ are assumed to be negligible, and the top boundary condition at $z = 0$ is

$$\rho_w c_w k_w \frac{\partial T}{\partial z} = Q = LH + SH + LW \tag{2}$$

where LH, SH, and LW are the surface latent and sensible heat fluxes and the net longwave radiation, defined as positive downward, respectively. Integration of (1) then yields

$$\rho_w c_w k_w \frac{\partial T}{\partial z} = Q + R_s - R(z) \tag{3}$$

where R_s is the net solar radiation at ocean surface. Further integration of (3) leads to

$$T_s - T_{-\delta} = \frac{\delta}{\rho_w c_w k_w}(Q + R_s f_s) \tag{4}$$

where f_s is the fraction of solar radiation absorbed in the sublayer [*Fairall et al.*, 1996; *Wick et al.*, 2005]:

$$f_s = \frac{1}{\delta}\int_{-\delta}^{0}\left(1 - \frac{R(z)}{R_s}\right)dz = 0.065 + 11\delta - \frac{6.6\times10^{-5}}{\delta}\left[1 - \exp\left(-\frac{\delta}{8\times10^{-4}}\right)\right] \tag{5}$$

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The thickness of the skin layer (δ) is taken from *Fairall et al.* [1996]:

$$\delta = 6\left[1 + \left(\frac{-16 g \alpha_w \nu_w^3}{u_{*w}^4 k_w^2 \rho_w c_w}(Q + R_s f_s)\right)^{3/4}\right]^{-1/3} \tag{6}$$

where g is gravity, α_w is the thermal expansion coefficient, ν_w is the kinematic viscosity, and the friction velocity in the water $u_{*w} = u_{*a}\sqrt{\rho/\rho_w}$ with u_{*a} being the friction velocity in the atmosphere and ρ being the air density.

[6] Below the skin layer, k_w is not as important as K_w. Integration of (1) along with the use of (3) at $z = -\delta$ results in

$$\frac{\partial}{\partial t}\int_{-d}^{-\delta} T dz = \frac{Q + R_s - R(-d)}{\rho_w c_w} - K_w \frac{\partial T}{\partial z}\Big|_{z=-d} \tag{7}$$

where d is the measurement depth at which the diurnal cycle can be omitted, and $R(-d)/R_s = \sum_{i=1}^{3} a_i \exp(-d\, b_i)$ with $(a_1, a_2, a_3) = (0.28, 0.27, 0.45)$, and $(b_1, b_2, b_3) = (71.5, 2.8, 0.07)$ m^{-1} [*Soloviev*, 1982]. Following *Large et al.* [1994],

$$K_w(z) = k u_{*w}(-z)/\phi_t\left(\frac{-z}{L}\right) \tag{8}$$

where $k = 0.4$ is the Von Karman constant, z is negative in the ocean, and the stability function

$$\phi_t\left(\frac{-z}{L}\right) = \begin{cases} 1 + 5\frac{-z}{L} & \text{for } \frac{-z}{L} \geq 0 \\ \left(1 - 16\frac{-z}{L}\right)^{-1/2} & \text{for } \frac{-z}{L} < 0. \end{cases} \tag{9}$$

The Monin-Obukhov length is

$$L = \rho_w c_w u_{*w}^3/(kF_d), \text{and } F_d = g\alpha_w[Q + R_s - R(-d)] \tag{10}$$

Furthermore, we assume $T = T_{-\delta} - [(z + \delta)/(-d + \delta)]^{\nu}(T_{-\delta} - T_{-d})$ with $d \gg \delta$, and the exponent ν is an empirical parameter. Under these conditions, (7) can be simplified as

$$\frac{\partial}{\partial t}(T_{-\delta} - T_{-d}) = \frac{Q + R_s - R(-d)}{d\rho_w c_w \nu/(\nu + 1)} - \frac{(\nu + 1)k u_{*w}}{d\phi_t(d/L)}(T_{-\delta} - T_{-d}) \tag{11}$$

[7] In the blended SST analysis product [*Reynolds and Smith*, 1994], the highest weight is given to the nighttime buoy and ship measurements and the diurnal cycle is omitted. For global ocean-atmosphere coupled models, the diurnal cycle of the temperature in the top oceanic layer (usually about 10 m in depth) is omitted (if the coupling is done once a day, as in most models) or very small (with hourly coupling). These temperatures can be directly taken as T_{-d}. The diurnal variation of ocean temperature is usually small at $d = 2$–4 m, so we take $d = 3$ m and $R(-d) = 0.36\ R_s$ using the *Soloviev* [1982] formulation. The parameter ν was taken as 1.0 by *Fairall et al.* [1996]. In general, it should be less than unity due to a stronger near-surface solar heating. We take $\nu = 0.3$ so that for the peak

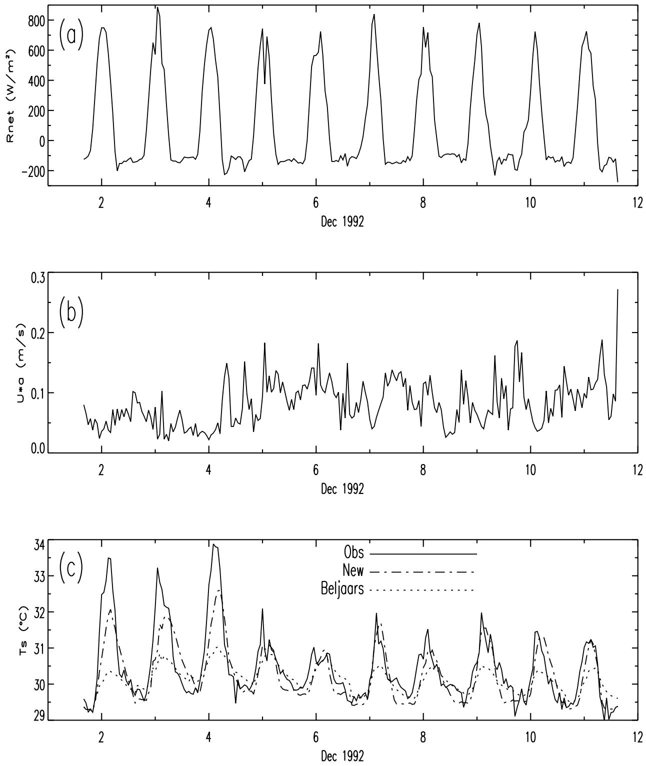

Figure 1. (a) Observed net surface flux ($Q + R_s$) over the western Pacific warm pool for ten days in December 1992; (b) observed air friction velocity; and (c) observed and simulated skin temperatures.

insolation of about 1000 W m^{-2} and assuming the balance of the last two terms in (11), $(T_s - T_{-d})$ is about 3 K under weak wind conditions. Note that, if a significantly different d is used, ν should also be adjusted under the above constraint.

[8] The last term in (11) represents the relaxation of $(T_{-\delta} - T_{-d})$ towards zero with the e-folding time $\tau_e = 0.5 d\phi_t(d/L)/(k u_{*w})$. Under strong wind conditions, τ_e is very small so that $(T_{-\delta} - T_{-d})$ is effectively zero. Under weak wind conditions, the solar heating term is correctly dominant during the day in (11). Furthermore, observations indicate that the residual warm layer can still exist long after sunset [*Fairall et al.*, 1996; *Gentemann et al.*, 2003]. However, this behavior cannot be simulated if (10) is used directly to compute L, because the stable stratification as represented by a positive $(T_{-\delta} - T_{-d})$ is not in equilibrium with the negative buoyancy flux F_d in (10) near or after sunset. Mathematically, a negative F_d in (10) decreases ϕ_t in (9). This, in turn, decreases τ_e and hence leads to the rapid destruction of the residual warm layer after sunset. To derive a more appropriate expression for F_d, we omit the first term in (11) and assume $\phi_t\,(d/L) \approx 5d/L$. Then (10) and (11) yield

$$F_d = \left(\frac{\nu g \alpha_w}{5d}\right)^{1/2} \rho_w c_w u_{*w}^2 \sqrt{T_{-\delta} - T_{-d}} \tag{12}$$

and it replaces the F_d formulation in (10) in the computation of L for $(T_{-\delta} - T_{-d}) > 0$. Equations (4) and (11) represent

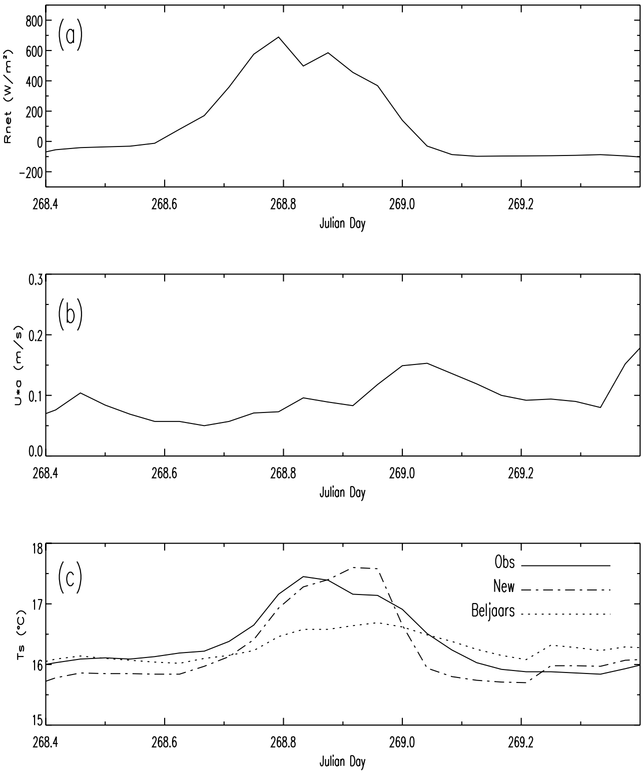

Figure 2. Same as Figure 1 except using the in situ data off the coast of Monterey, California for one day in September 2000.

our new scheme for T_s. For numerical stability, (11) can be solved using an implicit scheme.

3. Validation of the New Scheme

[9] First the radiometric T_s measurements from the R/V *Franklin* over the western Pacific warm pool region are used to evaluate our scheme and that of *Beljaars* [1997]. To mimic the intended applications of these schemes, the early morning averaged bulk temperature, measured by the ship's thermosalinograph taking water at a depth of 2.4 m, is used as T_{-d} in (11).

[10] Figure 1 shows that the peak net surface flux $R_{net} = Q + R_s$ does not vary much during the 10-day period, but the diurnal amplitude (i.e., daytime maximum minus nighttime minimum T_s) for the first three days is nearly twice as large as that for other days. This is primarily caused by the abrupt increase of wind after the first three days. Both the new scheme and the *Beljaars* [1997] scheme can simulate the diurnal cycle of T_s due to solar heating. However, the new scheme produces a more realistic daytime peak T_s throughout the period in Figure 1. In particular, the diurnal amplitude using the Beljaars scheme is insensitive to wind, which is inconsistent with observations. The mean absolute deviation between the computed and observed T_s values is 0.39 K and the correlation is 0.85 using the new scheme, while they are 0.50 K and 0.72, respectively, using the Beljaars scheme. For the averaged diurnal cycle over this 10-day period, the observed amplitude is 2.3 K, while the new and Beljaars schemes give 2.0 K and 0.88 K, respectively. If d is changed from 3 m by ±0.5 m in the new scheme, the amplitude would be changed by less than 0.15 K.

[11] Over midlatitude oceans, the skin temperature was measured with the calibrated infrared in situ measurement system (CIRIMS) radiometer [*Jessup et al.*, 2002] aboard the Research Platform *Flip* off the coast of Monterey, California in September – October 2000 [*Wick et al.*, 2005]. Figure 2 evaluates the two schemes using this dataset. The net solar flux and wind for this day over this midlatitude site are similar to the last few days over the tropical site in Figure 1. Hence the observed diurnal T_s amplitudes are also similar (i.e., about 1.5 K). The amplitude simulated using the new scheme is similar to the observed value, while the amplitude from the Beljaars scheme is just about half of the observed value.

[12] Using the Geostationary Operational Environmental Satellite (GOES) SST data, *Wu et al.* [1999, Figure 10] showed that, for a three day period in May 1998, the SST difference between 2000 UTC and 1200 UTC is as large as (and even larger than) 3 K along a zonal band from the Gulf of Mexico to North Atlantic where the surface wind is weak. Note that, because the GOES SST data are derived from regression against subsurface (bulk) temperatures, their diurnal cycle cannot be unambiguously associated with the skin layer. To compare with the GOES data, we have implemented our scheme into the ECMWF operational model. Specifically, T_{-d} in (11) is replaced by the ECMWF SST analysis and T_s is computed from (4) and (11) at each time step. The ECMWF model hindcasts for 3 days, starting from 1200 UTC using existing ECMWF analysis as the

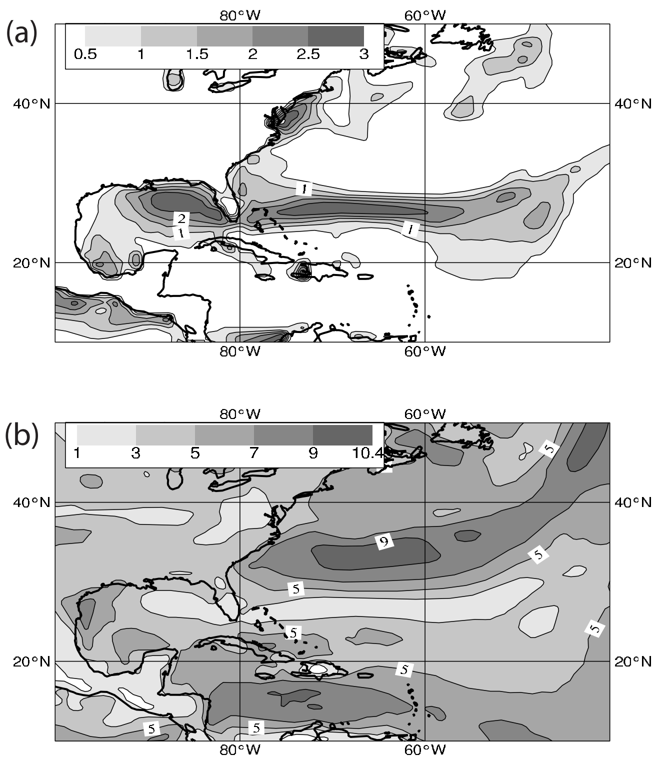

Figure 3. (a) The averaged T_s difference (K) between 2000 UTC and 1200 UTC, 20–22 May 1998 based on the ECMWF model along with the new T_s scheme; and (b) the averaged surface wind (m/s).

initial conditions. Figure 3 shows the three-day averaged T_s difference between 2000 UTC and 1200 UTC over a domain similar to that in Figure 10 of Wu et al. It is remarkable that the ECMWF model reproduces the zonal band of observed weak wind (Figure 3b versus Figure 10c of Wu et al.). Accordingly, the new scheme also realistically reproduces the large T_S variation along this band. Quantitatively, the temperature difference in Figure 3a is smaller than that in Figure 10b of Wu et al. primarily for two reasons. First, later studies [*Wick et al.*, 2002] have found that the systematic bias in the GOES SST retrieval also has a diurnal cycle, leading to an overestimate of the diurnal variation. Second, only the clear-sky GOES composite can be provided, while Figure 3a gives the three-day averaged variation (with or without clouds).

[13] We have also run the ECMWF model with three ensemble members for one year (August 2000–July 2001). As an example, the new scheme changes the ensemble annual mean surface latent heat flux by more than 10 W m^{-2} over several regions (figure not shown). It is a future task to do a detailed analysis of the impact of the T_s scheme on weather forecasting and atmospheric modeling (such as the ensemble simulations above). This scheme may also affect the four-dimensional atmospheric data assimilation, particularly over regions where SST has a significant diurnal variation. It can also be directly implemented into ocean-atmosphere coupled models. A recent study (G. Danabasoglu et al., Diurnal ocean-atmosphere coupling, submitted to *Journal of Climate*, 2005) has demonstrated that large-scale ocean-atmosphere coupling is a prime mechanism for amplifying the impact of solar diurnal variations on the daily mean SST.

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A. Beljaars, European Centre for Medium-Range Weather Forecasts, Reading RG2 9AX, UK. (anton.beljaars@ecmwf.int)

X. Zeng, Department of Atmospheric Sciences, University of Arizona, Tucson, AZ 85721, USA. (zeng@atmo.arizona.edu)