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Sea brightness temperature and effects of spray and whitecaps

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Abstract. Following the approach of *Tang* [1974], the microwave brightness of the sea surface is recalculated with recent parameterizations in terms of wind velocity for the slope distribution and whitecap coverage of the sea surface and the contribution of sea spray. The difference between this revised calculation with no free parameters and the earlier one becomes more significant as the wind velocity increases; it reaches 10 K or more at a wind velocity of 20 m s^{-1}. Present predictions compare favorably with experimental results, suggesting that this model does include the essential physical mechanisms.

1. Introduction

The sea surface temperature (SST) is an important factor governing energy fluxes between the atmosphere and oceans [*Khalsa*, 1983]. SST is also crucial in parameterizing these fluxes, which are the critical elements in models of air-sea interactions [*Miller et al.*, 1992]. In order to gain insight into these processes on a global scale, it is vitally important to develop remote-sensing techniques along with reliable algorithms to determine SST under various sea surface conditions. As improvements are made in oceanographic and atmospheric models, requirements become more stringent for the determination of SST. The present approach to the measurement of SST is to use the data of microwave brightness temperature, which is simply the apparent temperature of the water body at a given wavelength assuming a blackbody emission. True temperature can be derived from brightness temperature with a known emissivity of the seawater (see, for example, *Reif* [1965, pp. 381–382]). Considerable efforts have been devoted to the understanding of brightness temperature of the sea surface in the microwave regime; these efforts are concentrated on improving the estimates of dielectric constants of the seawater [*Klein and Swift*, 1977] and the understanding of sea surface roughness from passive measurements [*Hollinger*, 1971], as well as the formulation of better inversion algorithms [*Stogryn*, 1967; *Tang*, 1974; *Wilheit*, 1979; *Wentz*, 1983; *Guissard and Sobieski*, 1987; *Rufenach and Shuchman*, 1992].

Experimentally, measurements of the sea surface brightness temperature T_b show a dependence on the wind speed [*Nordberg et al.*, 1971; *Webster et al.*, 1976; *Yueh et al.*, 1995]. Previous theoretical calculations have included effects of whitecaps, sea spray, and the influence of surface slope distributions on the effective emissivity of the sea surface and the attenuation of the near-surface boundary layer [*Tang*, 1974]. However, these calculations used free-parameter fittings to relate the concentration of spray to T_b in order to produce reasonable agreement between experiment and theory. Other works did not include the effects of spray whatsoever [*Stogryn*, 1967; *Wilheit*, 1979; *Guissard and Sobieski*, 1987; *Rufenach and Shuchman*, 1992]. We have revisited this calculation using experimentally established parameterizations resulting from measurements of the spray concentration [*Wu*, 1990a], whitecap coverage [*Wu*, 1979], and sea surface slopes [*Wu*, 1990b] as functions of the wind speed. These parameterizations suggests that physical arguments of *Tang*, [1974] are basically correct. However, our calculations do not use free parameters but distributions of whitecap coverage and sea spray concentrations based on more appropriate experimental results.

2. General Approach

Brightness temperature can be defined via the sum of the radiation from the surface plus the reflected radiation from the sky [*Tang*, 1974]. Since the emissivity is a function of the nadir angle, this rudimentary definition of brightness temperature T_b takes the form

$$T_b(\theta) = E(\theta)T + [1 - E(\theta)]T_s(\theta), \qquad (1)$$

where T is the temperature of the water surface, T_s is the sky temperature, E is the emissivity of the water (hence (1-E) is the reflectivity), and θ is the nadir angle with respect to the mean water surface. Unfortunately, the determination of an emissivity as defined in (1) is not trivial.

In addition to the emitted radiation from the sea surface and the reflected sky radiation, there are also other components which contribute to the total observed brightness temperature which are enumerated by *Tang* [1974]. First of all, the sea surface is not smooth under most conditions. Since the emissivity of seawater is a function of nadir angle, the distribution of surface slopes must be accounted for in the calculation of emissivity. This result could also be extended to the calculation of the albedo via the integration of reflectivity over all wavelengths. Also, there are whitecaps and sea foam which have altogether different emissivities than the smooth seawater. Furthermore, above the sea surface, there is a layer which includes sea spray droplets. As the reflected sky radiation and the emitted surface radiation pass through this layer, they are

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attenuated. In general, there are also multiple reflection events; these contributions, however, will be ignored here since they should be negligibly small.

The general starting point for this calculation [*Tang*, 1974] assumes that we can write the measured brightness temperature in the presence of sea spray droplets, T_{bd}, as

$$T_{bd}(\theta) = \{T - [T - T_s(\theta)][1 - E(\theta)]\tau_l^2(\theta)\} \cdot [1 - W] + T_{bf}W \tag{2}$$

T_{bf} is the brightness temperature of whitecaps (foam), W is the coverage of whitecaps, and τ_l is the transmission coefficient of the spray region. Note that for low wind conditions with the absence of spray ($\tau_l = 1$) and whitecaps ($W = 0$), the brightness temperature is reduced to (1). Equation (2) should include all appreciable contributions to T_b. Any further contributions would be the result of multiply reflected components which should be negligibly small. For higher wind speeds it is necessary to consider contributions of whitecaps and spray. Including the effects of whitecaps follows directly from two results, the coverage as a function of the wind speed [*Wu*, 1979] and the emissivity of sea foam at various nadir angles [*Stogryn*, 1972].

Following *Shifrin and Ionina* [1968] and *Tang* [1974], we will also adopt the distribution of sea surface slopes for various wind speeds given by *Cox and Munk* [1954]. This derivation yields the emissivity of the sea surface as a function of θ, the wind velocity at 10-m height U_{10}, the local angle of the sea surface χ, and the slope distribution function $P[\theta_n(U_{10}), \phi_n(U_{10})]$ by Cox and Munk, where θ_n and ϕ_n are the polar and azimuthal angles, respectively, between vertical and the local surface normals. *Shifrin and Ionina* [1968] showed that

$$E = 1 - \sec(\theta)\int P[\theta_n(U_{10}), \phi_n(U_{10})] \cdot \cos(\chi)\sec^4(\theta_n)R(\chi)\sin(\theta_n)\,d\theta_n\,d\phi_n$$
$$= 1 - 0.502\left[\left(1 + \sqrt{\frac{C}{2}}\tan\theta\right)R(\chi^+) + \left(1 - \sqrt{\frac{C}{2}}\tan\theta\right)R(\chi^-)\right], \tag{3}$$

where $C \equiv \frac{1}{2}\overline{s^2}$, with $\overline{s^2}$ being the mean square surface slope; Cox and Munk suggested

$$C = 0.0015 + 0.00254U_{10}, \tag{4}$$

and the angles between the "incident" beam and the true local normal are given by

$$\chi^+ = \cos^{-1}\left[\frac{1}{\sqrt{1+C}}\left(\cos\theta + \sqrt{\frac{C}{2}}\sin\theta\right)\right], \tag{5a}$$

and

$$\chi^- = \cos^{-1}\left[\frac{1}{\sqrt{1+C}}\left(\cos\theta - \sqrt{\frac{C}{2}}\sin\theta\right)\right]. \tag{5b}$$

The reflectivities for horizontal and vertical polarization are, respectively,

$$R_h = \left|\frac{\cos\chi - \left[\dfrac{\varepsilon}{\varepsilon_0(U_{10}, z)} - \sin^2\chi\right]^{1/2}}{\cos\chi + \left[\dfrac{\varepsilon}{\varepsilon_0(U_{10}, z)} - \sin^2\chi\right]^{1/2}}\right|^2 \tag{6a}$$

and

$$R_v = \left|\frac{\dfrac{\varepsilon}{\varepsilon_0(U_{10}, z)}\cos\chi - \left[\dfrac{\varepsilon}{\varepsilon_0(U_{10}, z)} - \sin^2\chi\right]^{1/2}}{\dfrac{\varepsilon}{\varepsilon_0(U_{10}, z)}\cos\chi + \left[\dfrac{\varepsilon}{\varepsilon_0(U_{10}, z)} - \sin^2\chi\right]^{1/2}}\right|^2 \tag{6b}$$

Here ε is the dielectric constant of seawater, and $\varepsilon_0(U_{10}, z)$ is the dielectric constant for the air at height z above the mean sea surface. The value $\varepsilon_0(U_{10}, z)$ can be associated with ε via

$$\varepsilon_0 = 1 + (\varepsilon - 1)V_s \tag{7}$$

where V_s is the concentration of water due to spray droplets. In *Tang*'s [1974] calculation this concentration was assumed to behave as

$$V_s = a_1 + a_2U_{10} + a_3U_{10}^2 \tag{8}$$

where a_i are constants that account for observations with no appreciable spray existing below 5–7 m s^{-1}. In the above expression, droplet production is assumed to be roughly proportional to U_{10}, and whitecap coverage is roughly proportional to U_{10}^2. Additionally, V_s was forced to be zero for $U_{10} \leq 5$ m s^{-1}. Otherwise, the parameters a_2 and a_3 were allowed to be free in order to best fit the experimental brightness measurements. *Tang* [1974] also adopted *Stogryn*'s [1972] form for whitecap coverage

$$W = 7.75 \times 10^{-6}U^{3.231}. \tag{9}$$

We must also know the brightness temperature as a function of angle for sea foam T_{bf}. This was derived by *Stogryn* [1972] to be

$$T_{bf}(\theta) \equiv E_p(f, \theta)T_w \tag{10a}$$

$$E_p(f, \theta) \equiv E(f, 0)F_p(\theta), \qquad p = h \text{ or } v, \tag{10b}$$

$$E(f, 0) = 208 + 1.29f, \tag{10c}$$

$$F_h(\theta) = 1 - 1.748 \times 10^{-3}\theta - 7.336 \times 10^{-5}\theta^2 + 1.044 \times 10^{-7}\theta^3, \tag{10d}$$

$$F_v(\theta) = 1 - 9.946 \times 10^{-4}\theta + 3.218 \times 10^{-5}\theta^2 - 1.187 \times 10^{-6}\theta^3 + 7 \times 10^{-20}\theta^{10}. \tag{10e}$$

In these equations, θ is the nadir angle in degrees, T_w is the bulk water temperature, f is the receiving frequency of the measuring device, and F_p are the fitted polynomials with h and v denoting horizontal and vertical polarizations, respectively.

We will now consider each of the remaining components of (2), which have not yet been defined. As shown in the simplified relationship (1), the contribution to the observed brightness temperature of the sea surface is just the sum of the emitted radiation from the surface ET and the reflected radiation from the sky, $RT_s \equiv (1 - E)T_s$. In the work by *Tang* [1974], T_s was suggested to follow

$$T_s = 268(1 - e^{(-0.065\sec\theta)}) + 2.7. \tag{11}$$

With the additional formulation of τ_l from Tang given as

$$\tau_l = \exp\left(\frac{4l \sec\theta}{\lambda(J+1)} \frac{\varepsilon_{0i}(U_{10}, 0^+)}{\sqrt{\varepsilon_{0r}(U_{10}, 0^+)}}\right) \quad (12)$$

where λ is the wavelength of the detected radiation, l is an attenuation length, J is an integer greater than 1, and ε_{0i} and ε_{0r} are the imaginary and real parts of ε, respectively, we can now calculate (2). Note that $J = 3$ and $l = \lambda$ are convenient choices. It was shown by Tang that the final results are only weakly dependent on these parameters.

3. New Parameterizations

Now we will address some of the updated assumptions that have been made for this calculation. First of all, we include explicit results for both droplet concentration and whitecap coverage which we will show to be the two dominant components in the model. Note that these observations have been updated in the past 20 years and more accurate estimates have been produced for the parameterization of both of these components on wind speed. For the droplet concentration we will use the relationship due to *Wu* [1990a],

$$V_s = 8.46 \times 10^{-8} U_{10}^{2.65} \quad (13)$$

In addition, we include spume drop production, the tearing of wave crests by the wind. By using the production rate as parameterized by *Wu* [1993],

$$V_{\text{spume}} = 8.7 \times 10^{-5} \exp(0.875\, U_{10}), \quad (14)$$

we have estimated this contribution and shown the total spray concentration in the near-surface atmosphere; see Figure 1. Note that the contribution due to spume drops is strongly dependent on wind speed and does not become appreciable until U_{10} exceeds 20 m s^{-1} in our current estimate of the production rate. However, this rate has not been clearly established for field experimental results and is therefore a potential source of error in our estimation.

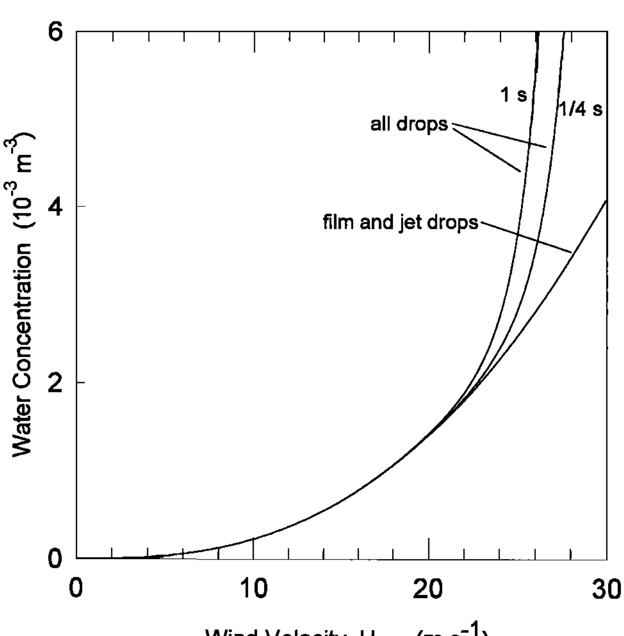

Figure 1. Water concentrations due to spray produced by bursting bubbles (film and jet drops) as well as spume drops with 1/4- and 1-s lifetimes versus the wind velocity.

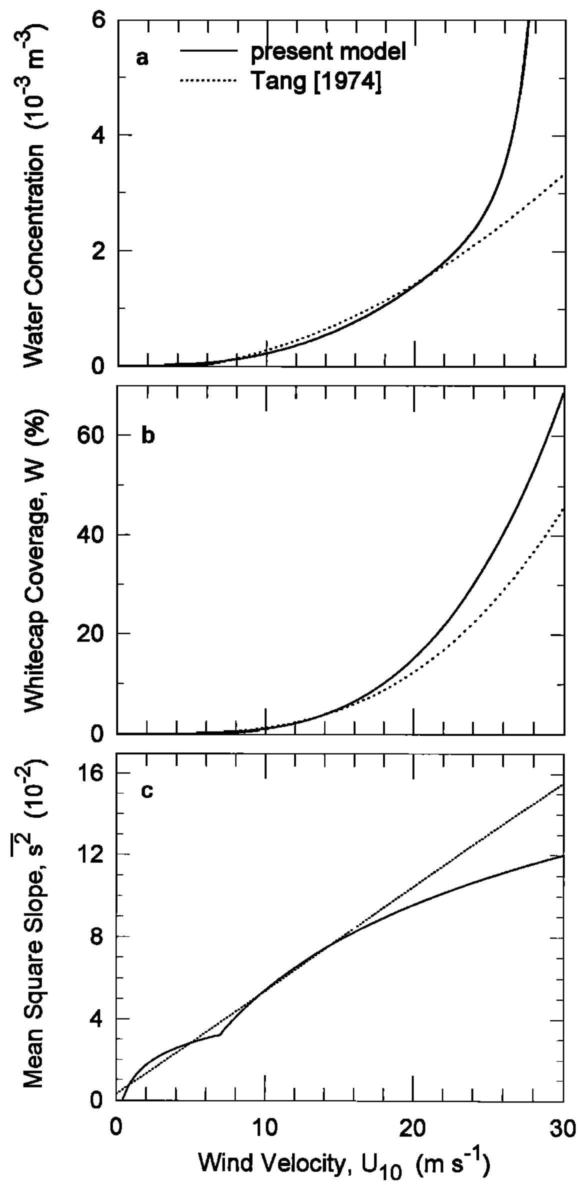

Figure 2. Comparisons of (a) water concentration, (b) whitecap coverage, and (c) mean square slope between the previous [*Tang*, 1974] and the current models.

Figure 2a shows a comparison of (13) plus (14) with the parameterization (8) by *Tang* [1974]. Note that *Wu*'s [1990a] model is based on measurements of sea spray, whereas the quadratic fit is from *Tang* [1974] and simply represents the optimized agreement. The reasonable comparison between the two at lower winds suggests that the physical arguments of *Tang* [1974] regarding the influence of sea spray are probably valid, even though the calculation using the spray concentration as a free parameter is not. The high wind divergence is due to the inclusion of the spume production rate.

In order to include the effects of whitecap coverage, we use the formulation by *Wu* [1979] for this parameter W,

$$W = 2 \times 10^{-6} U_{10}^{3.75}, \quad (15)$$

to weight the emissivity values of *Stogryn* [1972]. We compare *Tang*'s [1974] parameterization for W (9) with (15) in Figure 2b.

In the previous formulation by *Tang* [1974], he used the

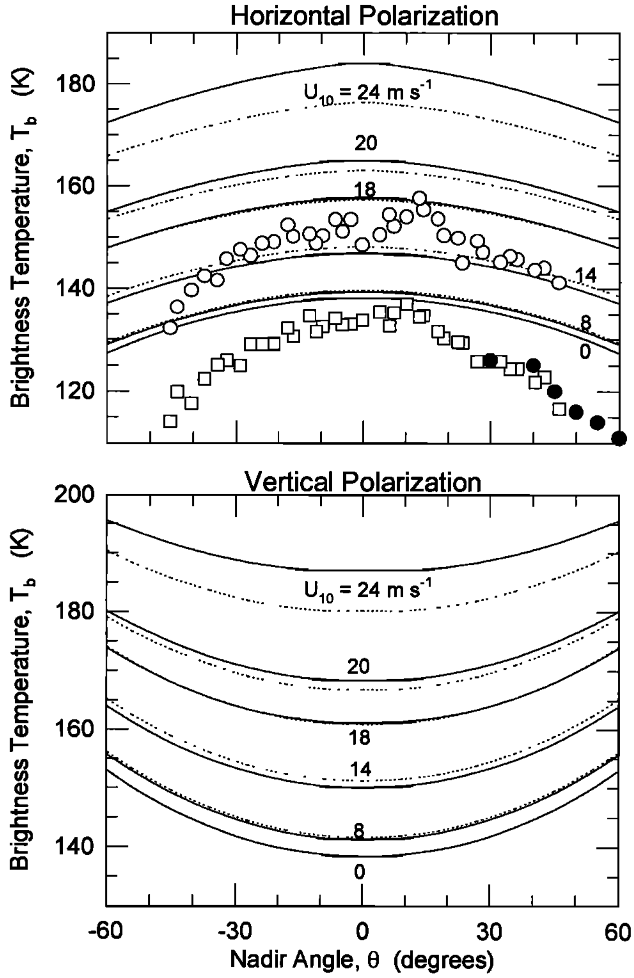

Figure 3. Brightness temperature of 19.35-GHz horizontally and vertically polarized microwaves versus nadir angle. The present model is shown as the solid line, and *Tang*'s [1974] model is shown as the dashed line with experimental data by *Nordberg et al.* [1971] at calm conditions (open squares) and 14 m s^{-1} (open circles) and *Yueh et al.* [1995] (solid circles).

results directly due to *Cox and Munk* [1954] for the mean square slope as reflected in (4). *Wu* [1990b] has an improved parameterization for the sea surface slope distribution which we will incorporate into the final product.

$$\overline{s^2} = (0.90 + 1.20 \ln U_{10}) \times 10^{-2} \qquad U_{10} < 7 \text{ m s}^{-1}, \qquad (16a)$$

$$\overline{s^2} = (-8.40 + 6.00 \ln U_{10}) \times 10^{-2} \qquad U_{10} > 7 \text{ m s}^{-1} \qquad (16b)$$

Figure 2c shows comparisons between the current and previous parameterizations of the mean square slope as a function of the wind velocity.

4. Results

Figure 3 shows the calculated horizontal and vertical polarized brightness temperature as a function of nadir angle at various wind speeds with the resultant changes in spray, surface slope, and whitecap coverage being included. The water temperature is assumed to be 300 K. The superimposed experimental data are due to *Nordberg et al.* [1971] and *Yueh et al.* [1995]. The measurements shown were made at horizontal polarization and were taken along with ground truth measurements. There is a fair agreement between the predictions and the data. *Yueh et al.* [1995] found that the measured brightness temperature also correlated with wind direction, so their data represent an azimuthal angle average. This azimuthal effect is not considered in the current model. Earlier measurements by *Webster et al.* [1976] at a nadir angle of 38° showed an increase in the brightness temperature of roughly 30 K at a wind speed of 20 m s^{-1}, again showing good agreement with the current model. It is clear from the comparison of the current calculation with that done previously [*Tang*, 1974] that the essential physical grounds have not changed. In fact, the central difference between the two models is the inclusion of more updated parameterizations of various components near the air-sea interface that change the effective emissivity of the ocean surface. The advantage of the current model is that no free parameters are necessary to provide a reasonable agreement.

In order to establish the relative importance of the various air-sea interface components which affect the brightness temperature, we have calculated the changes in T_b due to each one separately. These results are shown in Figure 4 where we have plotted the brightness temperature change at nadir as a function of the wind velocity. The effects of spray and whitecap coverage are clearly the dominant terms, with both having comparable contributions. Note that the spume component of spray is displayed separately and does not appear to be a major factor for wind velocities less than 20 m s^{-1}. It is reasonable that for much higher winds, the spume component is important as suggested by the results shown in Figure 1. Also, from Figure 4, note that the effect of sea surface slope on the total change in T_b is very small. The change in T_b due to sea surface slope is also polarization dependent unlike the spray and whitecap contributions; increasing sea surface slope causes an increase in the vertically polarized T_b, while the horizontally polarized T_b is reduced.

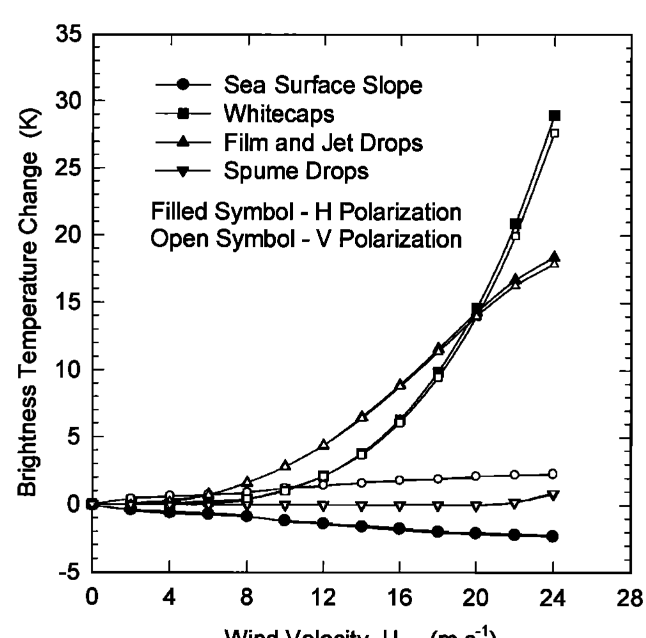

Figure 4. Contributions to brightness temperature change as a function of wind velocity for sea surface slope, whitecaps, film and jet drops, and spume drops. Both horizontal and vertical polarization cases are shown.

5. Conclusions

We have updated the approach of *Tang* [1974] in order to provide a more realistic algorithm for calculating SST from radiometry data. By using parameterizations of sea surface slope, whitecap coverage, and spray production, we are able to arrive at a reasonable prediction of the brightness temperature without fitting parameters. These parameterizations are all in terms of the wind velocity, a quantity which can be obtained from scatterometer data. Therefore a combination of radiometer brightness temperature data and scatterometer wind velocity data should provide sufficient input to derive SST by inverting the present type of wind-dependent correction approach. Using such a technique, it is possible that more accurately determined SST can be obtained with remote sensing.

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