

Appendix A

VALUES USED IN THE NEW METHOD

Some of the values necessary for the computation of (3.4-3.20) are known constants and regression coefficients. For other values, appropriate choices have been made in Chapter 3. These choices and known constants are summarized in Table A.1, while the regression coefficients are listed in Table A.2. Table A.3 compiles the temperatures used in conditions (3.22) for a rain mask at different latitudes. Table A.4 lists the standard deviations (or rms errors), s_x , of the variables used to compute emissivity and W variances in (3.24) and (3.26).

Table A.1 Summary of chosen and known initial values in (3.2-3.20)

Variable	Symbol	Value	Units	Reference
Chosen values				
Frequency of radiation	f w	19.35 1.216×10^{11}	GHz rad	Wentz, 1997
Polarization of radiation	h, v	h	-	Section 3.2.2
Water fraction in foam	Q	0.02	-	Section 3.2.1
Known values				
Cosmic background	T_{CB}	2.725	K	Smoot and Scott, 2000
Incident angle	q	53.4	°	Wentz, 1997
Permittivity of free space	$\hat{\epsilon}_0$	8.85×10^{-12}	F m ⁻¹	Stewart, 1985
Dielectric constant at inf frequency	$\hat{\epsilon}_\infty$	4.9	-	Stewart, 1985
Conductivity	$\hat{\sigma}$	5.32	$\Omega^{-1} \text{m}^{-1}$	Hasted, 1988
Empirical constant	\hat{a}	0	-	Klein and Swift, 1977

Table A.2 List of regression coefficients used in (3.6-3.20)

In expression	Symbol	Value	Reference
(3.6)	c_0	240.58	Wentz, 1997
(3.6)	c_1	305.96×10^{-2}	Wentz, 1997
(3.6)	c_2	-764.41×10^{-4}	Wentz, 1997
(3.6)	c_3	885.95×10^{-6}	Wentz, 1997
(3.6)	c_4	-40.8×10^{-7}	Wentz, 1997
(3.6)	c_5	0.6	Wentz, 1997
(3.6)	c_6	-0.16	Wentz, 1997
(3.6)	c_7	-2.13×10^{-2}	Wentz, 1997
(3.6)	c_8	0.8337	Wentz, 1997
(3.6)	c_9	-3.029×10^{-5}	Wentz, 1997
(3.6)	c_{10}	301.16	Wentz, 1997
(3.8)	c_{11}	11.8	Wentz, 1997
(3.9)	c_{12}	2.23×10^{-3}	Wentz, 1997
(3.9)	c_{13}	0	Wentz, 1997
(3.10)	c_{14}	0.2858	Wentz, 1997
(3.10)	c_{15}	0.3751	Wentz, 1997
(3.10)	c_{16}	0.208	Wentz, 1997
(3.10)	c_{17}	-0.026	Wentz, 1997
(3.15)	s_0	87.134	Stewart, 1985
(3.15)	s_1	-0.1949	Stewart, 1985
(3.15)	s_2	-1.276×10^{-2}	Stewart, 1985
(3.15)	s_3	2.491×10^{-4}	Stewart, 1985
(3.15)	a_0	1.0	Stewart, 1985
(3.15)	a_{st}	1.613×10^{-5}	Stewart, 1985
(3.15)	a_1	-3.656×10^{-3}	Stewart, 1985
(3.15)	a_2	3.21×10^{-5}	Stewart, 1985
(3.15)	a_3	-4.232×10^{-7}	Stewart, 1985
(3.16)	t_0	1.768×10^{-11}	Stewart, 1985
(3.16)	t_1	-6.086×10^{-13}	Stewart, 1985
(3.16)	t_2	1.104×10^{-14}	Stewart, 1985
(3.16)	t_3	-8.111×10^{-17}	Stewart, 1985
(3.16)	b_0	1.0	Stewart, 1985
(3.16)	b_{st}	2.282×10^{-5}	Stewart, 1985
(3.16)	b_1	-7.638×10^{-4}	Stewart, 1985
(3.16)	b_2	-7.76×10^{-6}	Stewart, 1985
(3.16)	b_3	1.105×10^{-8}	Stewart, 1985
(3.17)	h_0	0.115	Pandey and Kakar, 1982
(3.17)	h_1	3.8×10^{-5}	Pandey and Kakar, 1982
(3.17)	v_0	0.117	Pandey and Kakar, 1982
(3.17)	v_1	-2.09×10^{-3}	Pandey and Kakar, 1982
(3.17)	v_2	7.32×10^{-2}	Pandey and Kakar, 1982

Table A.3 Temperatures used in condition (3.22) for a rain mask

Zone	Latitude	T_0 [K]	T_1 [K]
Tropics	0° - 25°	50	175
Mid-latitudes	25° - 55°	50	165
Arctic	55° - 90°	50	130

Table A.4 Variances of the basic variables contributing to the error of emissivities (3.26) and whitecap coverage (3.24).

Standard deviation for variable	Symbol	Value	Units	Reference
Brightness temperature	s_{T_B}	1	K	Wentz, 1997
Sea surface temperature	s_{T_s}	0.3	°C	Data
Water salinity	s_s	0.2	psu	Data
Water vapor	s_V	1.2	mm	Wentz, 1997
Wind speed	s_U	0.9	m s ⁻¹	Wentz, 1997
Incident angle	s_q	0.25	°	Wentz, 1997
Dielectric constant at inf frequency	s_{e_∞}	0.98	-	Stewart, 1985
Conductivity	s_s	4.41	$\Omega^{-1} \text{m}^{-1}$	Chosen
Water fraction	s_Q	1	%	Chosen

Table A.5 Regression coefficients for $Sc(T_s)$ relation (4.7) for Rn and CO₂ (reproduced from Erickson, 1993).

Gas	a	b	c	d
Rn	3147.3	-201.9	5.5	-5.5×10^{-2}
CO ₂	1911.3	-113.7	2.9	-2.9×10^{-2}

Appendix B

DIELECTRIC PROPERTIES OF SEAWATER

Calculations of specular and foam emissivities, e_s and e_f , in (3.2) require the calculation of the real and imaginary components of the dielectric constants of seawater and foam.

The dielectric constant (or permittivity), ϵ , together with the permeability, μ , and the conductivity, σ , are fundamental properties of any material. These three quantities are intrinsic parts of the Maxwell equations, which provide a concise mathematical description of the dynamics of electric and magnetic fields comprising electromagnetic radiation in the presence of matter.

The dielectric constant of a material is a measure of how much the distribution of electric charges within a material can be distorted (or polarized) by an external electric field (Debye, 1929; Hasted, 1973). Because the material absorbs part of the external radiation and uses it for such a distortion, it can also be stated that the dielectric constant is a measure of the absorption of (or losses in) the material.

The dielectric properties are usually considered in terms of electric moments of the molecules composing the material. These electric moments can be both permanent and induced. The molecules of all substances have induced electric moments, which form when a substance is placed in an electric field and the molecules align along the field direction. Only some molecules, however, called polar molecules, have a permanent electric moment. A polar molecule is like a dipole (a system of a

positive and negative charges separated by a distance) and the opposite charges of the molecule create a permanent electric moment. Water molecule is a polar molecule.

Indeed, the remarkable properties of water are not so much a consequence of its elemental composition (one oxygen and two hydrogen atoms), but its structure. The water molecule has an asymmetric architecture with an angle between the two hydrogen atoms of order of 104° . This structure makes the group of hydrogen atoms more electropositive than the oxygen atom, i.e., the water molecules act as dipoles. Being composed of polar molecules, the water possesses its own permanent electric moment. The dielectric constant of water describes these electric properties.

The dielectric constant of seawater, and any material, in general, is a complex number, $\mathbf{e} = \mathbf{e}' - i\mathbf{e}''$, and the Debye equation gives its explicit form:

$$\mathbf{e} = \mathbf{e}_\infty + \frac{\mathbf{e}_s - \mathbf{e}_\infty}{1 + (i\omega\tau)^{1-b}} - i \frac{\mathbf{S}}{\omega\epsilon_0} \quad (\text{B.1})$$

where $\omega = 2\pi f_r$ is the angular frequency of the radiation interacting with the material, f_r is the frequency in Hz, ϵ_∞ is the dielectric constant at infinite frequency, ϵ_0 is the static dielectric constant, τ is the relaxation time, σ is the ionic conductivity of water, \hat{a} is an empirical constant, and ϵ_0 is the permittivity of free space. With the real and imaginary components given by (3.14), this equation becomes:

$$\mathbf{e} = \mathbf{e}_\infty + \frac{\mathbf{e}_s - \mathbf{e}_\infty}{1 + \omega^2\tau^2} - i \left[\frac{\omega\tau(\mathbf{e}_s - \mathbf{e}_\infty)}{1 + \omega^2\tau^2} + \frac{\mathbf{S}}{\omega\epsilon_0} \right] \quad (\text{B.2}).$$

While the real component, \mathbf{e}' in (B.2) represents the actual magnitude of the dielectric constant, the imaginary part, \mathbf{e}'' (in the brackets), is a measure of the attenuation of incident radiation by the material; $\mathbf{e}'' = 0$ for nonabsorbent material. In general, the effect of losses within the material, expressed with \mathbf{e}'' , is to decrease its dielectric

constant. The representation (B.2) of the Debye equation assists in understanding the nature of the dielectric constant at different electromagnetic frequencies.

The static dielectric constant, $\hat{\alpha}_s$, is the dielectric constant in a static electric field: at $\hat{\omega} = 0$, $\mathbf{e}' = \hat{\alpha}_s$. The static dielectric constant consists of the permanent moments, if present, of the molecules of a given material, and the moments induced in each molecule by the surrounding molecules.

When an external electric field acts upon the medium, its frequency introduces changes in the relative contribution of permanent and induced moments to the dielectric constant. For low-frequency variations of the electric field, the dielectric constant is still formed by the permanent and induced moments—these are the two terms in the real component in (B.2). With increasing frequency of the applied field, the contribution of the induced electric moment to the dielectric constant of the medium increases. Also, the role of the static dielectric moment decreases, because it becomes more difficult for the dipoles to follow the field variations.

Eventually, some lag develops between the changes of the electric field and the orientation of the dipoles. This causes some of the field energy to be absorbed by the medium, i.e., losses appear, which decreases the dielectric constant. Mathematically this is described with the appearance of the first term in the imaginary part of (B.2). This term grows as the frequency, $\hat{\omega}$, increases. At some specific sufficiently high frequency, the contribution of the static term to the dielectric constant is completely dominated by the induced moment.

At infinitely high frequencies, the dielectric constant forms only from the induced moments and this is the dielectric constant at infinite frequency, $\hat{\alpha}_\infty$.

The frequency at which the contribution of the induced electric moment starts to prevail over the permanent one is determined by the condition $\hat{\omega}\hat{\delta} = 1$. Each

material has his own relaxation time, $\hat{\delta}$, of the order of 10^{-12} s, which is the time necessary for the dipoles of the material to either return from aligned back to random distribution after removing the electric field, or to re-orient as the electric field alternates. The losses start to appear when the relaxation time, $\hat{\delta}$, is not short enough to assure dipole re-alignment as the frequency of the field dictates. The relaxation time does not have one explicit value. Instead, different dipoles within a substance create a spread of values for $\hat{\delta}$. The parameter \hat{a} in (B.1) is a measure of this spreading of values of the relaxation time, $\hat{\delta}$.

The second term in the imaginary component in (B.2) is of great importance for conducting materials, such as seawater. Seawater, being an electrolyte solution with many ions like Na^+ and Cl^- , has higher conductivity than pure water. The role of conductivity is to add to the absorption losses and further decrease the dielectric constant. The physical reason for conductivity to lower the dielectric constant of seawater is that in electrolyte solutions the dipoles are already slightly oriented and when an external field is applied the ability of the dipole to orient along its direction is diminished; in other words, more energy is necessary to prevail over this initial orientation and make the dipoles follow the electric field variations.

A decrease in dielectric constant increases the reflectivity, r , of the medium, and with that, its emissivity, e , decreases. Free ions in seawater make it a good conductor with relatively high conductivity. This high conductivity of seawater increases the complex component of the dielectric constant giving the relatively low emissivity of seawater.

Appendix C

ERROR ANALYSIS PROCEDURES

The procedure of deriving analytical expressions for the sensitivity coefficients necessary for the error analysis (§3.2.3) involves determination of the functional dependencies of the investigated variables and decision which covariant terms to include. This appendix gives details for these considerations.

C.1 Functional dependencies

According to (3.2), whitecap coverage is a function of seawater emissivity, e , specular emissivity, e_s , foam emissivity, e_f , and rough sea emissivity, Δe_r :

$$W = f(e, e_s, e_f, \Delta e_r) \quad (\text{C.1})$$

According to (3.4), the seawater emissivity, e , is a function of brightness temperature, T_B , water temperature, T_s , upwelling and downwelling atmospheric radiations, T_{BU} and T_{BD} , and atmospheric transmittance, t :

$$e = f(T_B, T_s, T_{BU}, T_{BD}, t) \quad (\text{C.2})$$

Here T_B and T_s are measured, while T_{BU} , T_{BD} , and t are calculated with (3.5 – 3.10).

Using equations (3.5 – 3.10), the functional dependencies of T_{BU} , T_{BD} , and t can be determined. From (3.5):

$$T_{BU} = f(T_U, t) \quad (\text{C.3a})$$

and

$$T_{BD} = f(T_D, t) \quad (\text{C.3b})$$

Next, (3.6) shows that the downwelling and upwelling effective air temperatures, T_D and T_U , are functions of water vapor, V , and water temperature, T_s :

$$T_D = f(V, T_s, T_V) \cong f(V, T_s) \quad (\text{C.4a})$$

and

$$T_U = f(T_D, V) \cong f(V, T_s) \quad (\text{C.4b})$$

using the fact that the water temperature typical for water vapor V is $T_V = f(V)$.

From (3.7a), the atmospheric transmittance depends on the incident angle, \mathbf{q} , and the attenuation coefficient, k :

$$t = f(\mathbf{q}, k), \quad (\text{C.5})$$

where \mathbf{q} is known, while (3.7b) shows that the attenuation coefficient is a function of the attenuation coefficients of oxygen, a_O , water vapor, a_V , and cloud liquid water, a_L :

$$k = f(a_O, a_V, a_L). \quad (\text{C.6})$$

From (3.8), $a_O = f(T_D)$, which with (C.4a) transforms to:

$$a_O = f(V, T_s) \quad (\text{C.7a})$$

From (3.9), the attenuation coefficient of water vapor, a_V , is a function of water vapor content, V , only:

$$a_V = f(V) \quad (\text{C.7b}).$$

From (3.10), the attenuation coefficient of cloud liquid water, a_L , depends on the mean temperature between the surface and the freezing level, T_L , and cloud liquid water, L :

$$a_L = f(T_L, L).$$

Water temperature, T_s , determines T_L (§3.2.3), i.e., $T_L = f(T_s)$, yielding:

$$a_L = f(T_s, L) \quad (\text{C.7c})$$

Equations (C.7) modify (C.6), i.e., $k = f(V, T_s, L)$, which, in turn, modifies (C.5) to:

$$t = f(\mathbf{q}, V, T_s, L) \quad (\text{C.8})$$

Using (C.4a), (C.4b), and (C.8), the seawater emissivity, e , in (C.2) is found to depend on five basic variables:

$$e = f(T_B, T_s, V, L, \mathbf{q}).$$

The application of a clear-sky mask to all data (§3.2.2, *Data preparation*) sets $L = 0$, which reduces the variables to four:

$$e = f(T_B, T_s, V, \mathbf{q}). \quad (\text{C.9a})$$

Analogous tracking of the functional dependencies for the other emissivities in (C.1) gives:

$$e_s = f(\mathbf{e}_\mathbf{x}, T_s, S, \mathbf{s}, \mathbf{q}) \quad (\text{C.9b})$$

$$e_f = f(\mathbf{e}_\mathbf{x}, T_s, S, \mathbf{s}, \mathbf{q}, Q) \quad (\text{C.9c})$$

$$\Delta e_r = f(U_{10}, T_s, \mathbf{q}) \quad (\text{C.9d})$$

Equations (C.9) show that all emissivities and W are dependent on nine basic variables, namely T_B , V , U_{10} , T_s , S , $\mathbf{e}_\mathbf{x}$, \mathbf{s} , \mathbf{q} , and Q . The variances of these basic variables would contribute to the variance of each of the emissivities and W .

The errors in measuring or estimating any of the basic variables come from different sources. The accuracy of measuring brightness temperature, T_B , with SSM/I depends on the radiometer noise, the performance and stability of the calibration system, which periodically reads data from hot and cold reference targets, the implementation of the APC accounting for spurious radiation falling within the side lobes of the antenna, and the precision of the geolocation algorithm, which assigns latitude/longitude coordinates to the measurements within the satellite footprint (Holinger, 1990). The uncertainty in the incident angle, \mathbf{q} , comes from variations in the spacecraft altitude, misalignment between the SSM/I sensor and the satellite, and mechanical inaccuracies in the rotation and positioning of the scanning system. Wind speed, U_{10} , and atmospheric water vapor, V , are retrieved using the so called T_B model function, which is a simple close-form expressions of the RTE relating uniquely T_B , U_{10} , V , and other variables (Wentz, 1997). The retrieval of U_{10} and V is accomplished by varying their values until the T_B model function matches the SSM/I observations. Thus,

the accuracy of U_{10} and V depends on the accuracy of the T_B model. Sources of error in retrieving U_{10} and V are the atmospheric model used to represent the atmospheric conditions through which the radiation propagates, the variability in wind direction, the radiometer noise, the sampling mismatch both spatial and temporal between satellite and buoy data used for deriving coefficients for the model, and other errors whose cause is not well determined (Wentz, 1997). Errors degrading the accuracy of AVHRR retrievals of T_s come from sensor calibration, detector noise, atmospheric correction, cloud contamination, aerosol contamination, and the “skin” effect at the sea surface (Harris and Saunders, 1996). Data for salinity have undergone extensive quality control checks (Levitus et al., 1998). The main accuracy in S , therefore, comes from inaccuracies in the parameters of the applied objective analysis algorithm (recall section **WOA98 S data**). The objective analysis algorithm is essentially an interpolation routine applied to salinity measurements from different surface stations. The nonuniform distribution of these measurements—clustered in some places and completely missing in other regions—is the main cause for the errors in the objective analysis algorithm (Smith et al., 1986). The errors in seawater conductivity, σ , and the dielectric constant of seawater at infinite frequency, ϵ_∞ , is due to their choice as constants, while they depends on T_s . Finally, the error introduced via the choice of Q is due to the lack of more knowledge about the relation between emissivity and air content within whitecaps.

Some of these errors could be improved with the technological development. For instance, the inclusion of novel design features in the along-track scanning radiometer (ATSR) leads to more accurate SST retrievals compared to those from AVHRR (Harris and Saunders, 1996). ATSR has lower calibration error, lower detector noise, and improved atmospheric correction. This results in improvement of

SST accuracy by more than a factor of 2. A new NASA program aims the launch of a sensor measuring S on a global scale with an accuracy as low as 0.1 psu.

C.2 Covariant terms

In deriving error expressions for the emissivities in (3.2), variances and covariances of the basic variables (T_B , V , U_{10} , T_s , S , e_{v} , \mathbf{s} , \mathbf{q} , and Q) appear. There is a set of covariant terms for the variance of each emissivity. For example, using (C.9a), the possible covariant terms entering the variance \mathbf{s}_e^2 for seawater emissivity, e , couple the following pairs of variables: T_B — T_s , T_B — V , T_B — \mathbf{q} , T_s — V , T_s — \mathbf{q} , and V — \mathbf{q} . Knowing how each of the basic variables is measured or calculated helps to identify which among these pairs are independent and which are correlated. The reasoning for the \mathbf{s}_e^2 example is as follows.

Different techniques and sensors measure brightness temperature, T_B , and water temperature, T_s . Their variances, therefore, are independent of each other. The covariant term coupling T_B and T_s would not contribute to the variance of seawater emissivity, \mathbf{s}_e^2 , and could be omitted in the expression for \mathbf{s}_e^2 . On the basis of the same logic, terms for T_s — \mathbf{q} and T_s — V pairs also could be dropped out. It is noteworthy here that water temperature, T_s , and water vapor, V , are physically dependent (the amount of V does depend on T_s) but since T_s and V values are derived from different sensors, their variances would not interact. The water vapor, V , is derived from brightness temperature, T_B , thus the variances of both are correlated. Therefore, the covariant term T_B — V has to be included in the error expression. Covariant terms for the T_B — \mathbf{q} and V — \mathbf{q} pairs also have to be accounted for because brightness temperature, T_B , and the water vapor, V , are obtained from measurements of SSM/I at incident angle, \mathbf{q} , having some variance, \mathbf{s}_q^2 . As a result of this reasoning, the variance of seawater emissivity, \mathbf{s}_e^2 , must include three of the possible six

covariant terms. These are the terms containing variances \mathbf{s}_{BV}^2 , \mathbf{s}_{Bq}^2 , and \mathbf{s}_{Vq}^2 in the expression for \mathbf{s}_e^2 .

A similar train of reasoning for the covariant terms in variances $\mathbf{s}_{e_s}^2$, $\mathbf{s}_{e_f}^2$, and $\mathbf{s}_{e_r}^2$ leads to the expressions shown in (3.26). An additional point for the consideration of these variances is that the variances of dielectric constant at infinite frequency, ϵ_∞ , seawater conductivity, \mathbf{s} , and water fraction of foam, Q , are either known or chosen, and are, therefore, independent.

The expression (3.24) for the variance of whitecap coverage, \mathbf{s}_w^2 , includes two out of six possible covariant terms representing the coupling of $e-\Delta e_r$ and e_s-e_f . It is necessary to include the covariance of seawater emissivity, e , and rough sea emissivity, Δe_r , because the computations of both use variables measured by SSM/I. Specular and foam emissivity, e_s and e_f , are correlated because the computation of foam dielectric constant, needed for e_f , uses the seawater dielectric constant included in the computation of e_s as well.

Appendix D

DEPOSITION VELOCITY

Following the suggestion of Smith et al. (1993), equation (5.6) converts the number flux of sea-salt aerosols to a sea-salt number concentration using the deposition velocity, V_d . This appendix describes the procedure of calculating V_d .

Generally, the deposition velocity of a particle is a sum of dry and wet deposition velocities (Slinn, 1978). Only the dry component of V_d is used here for two reasons: i) sea-salt aerosols are relatively large and for their deposition the gravitational settling is more important than the scavenging by hydrometeors such as rain and snow; ii) the considerations in this study are for a non-raining atmosphere.

The calculations use the Slinn and Slinn (1980) model, which considers V_d as a result of particle deposition through a two-layer system with a constant flux layer and a diffusion layer near the water surface. The constant flux layer is assumed to have a RH of 80% and the deposition layer, which is in equilibrium with seawater, has a RH of 98.2% (Erickson and Duce, 1988). Thus, the sea-salt particle radii r_{80} and r_0 are used for the corresponding layers. The transfer velocity through the constant flux layer, k_c , and the transfer velocity through the diffusion layer, k_d , contribute to V_d , (Slinn and Slinn, 1980):

$$\frac{1}{V_d} = \frac{1}{k_c} + \frac{1}{k_d} - \frac{V_g(r_{80})}{k_c k_d} \quad (\text{D.1})$$

where

$$k_c = k'_c + V_g(r_{80}) \quad (\text{D.2})$$

accounts for gravitational settling with term $V_g(r_{80})$ and turbulent transfer at high wind speeds with term k'_c , while

$$k_d = k'_d + V_g(r_0) \quad (\text{D.3})$$

considers gravitational settling, $V_g(r_0)$, and transfer via diffusion and impaction, k'_d .

The gravitational settling, be it at r_{80} or r_0 in μm , is calculated from Stokes' law in cm s^{-1} (Batchelor, 1967, p.234):

$$V_g(r_{80}) = \frac{2\mathbf{r}_p g r_{80}^2}{9\mathbf{m}} C \times 10^{-8} \quad (\text{D.4})$$

where \mathbf{r}_p is the density of sea-salt particle ($\mathbf{r}_p = 1027 \text{ kg m}^{-3}$ at RH of 98.2% and $\mathbf{r}_p = 1197 \text{ kg m}^{-3}$ at RH of 80%), $g = 9.8 \text{ m s}^{-2}$ is the acceleration due to gravity, and $\mathbf{m} = 1.79 \times 10^{-4} \text{ g cm}^{-1} \text{ s}^{-1}$ is the absolute viscosity of the air (Erickson and Duce, 1988). The factor 10^{-8} converts the particle size from cm to μm . The factor C is the “slip” or Cunningham factor applicable for small particles with radii $r_0 < 2.5 \mu\text{m}$. Stokes' law applies to particles large compared to the mean free path of the air molecules, \mathbf{I} . As particle size decreases and approaches \mathbf{I} , this factor accounts for the increase of their settling velocity due to two effects: 1) particles participate in Brownian motion, which helps them to migrate from regions of higher to lower particle concentrations; 2) the drag force on small particles decreases as they are able to move or “slip” through the intermolecular voids. The slip factor can be calculated for r_{80} or r_0 as (Zhang et al., 2001):

$$C = 1 + \frac{\mathbf{I}}{r_{80}} \left[1.257 + 0.4 \exp\left(-\frac{1.1 r_{80}}{\mathbf{I}}\right) \right] \quad (\text{D.5}).$$

The mean free path of air molecules is calculated as a function of air temperature, T , and pressure, P , (McPherson, 1993):

$$I = \frac{m}{0.499P \left(\frac{8}{pRT} \right)^{\frac{1}{2}}}$$

where $R = 287.04 \text{ K kg}^{-1} \text{ K}^{-1}$ is the gas constant. At standard atmosphere conditions with $T = 20 \text{ }^\circ\text{C}$ and $P = 10^5 \text{ Pa}$, $I = 6.52 \times 10^{-8} \text{ m}$.

The transport by turbulence across the constant layer in cm s^{-1} is (Slinn and Slinn, 1980):

$$k'_c = \frac{100}{1-k} C_D U_{10} \quad (\text{D.6})$$

where $k = 0.4$ is the von Karman constant and $C_D = 1.3 \times 10^{-3}$ is the drag coefficient. In the calculations of (D.6), fields of monthly U_{10} in m s^{-1} are used. The factor 100 converts the units of k'_c from m s^{-1} to cm s^{-1} .

The transfer velocity across the diffusion layer in cm s^{-1} is calculated as (Slinn and Slinn, 1980):

$$k'_d = \frac{100}{k} C_D U_{10} (Sc^{-0.5} + 10^{-3/St}) \quad (\text{D.7})$$

where Sc is the Schmidt number representing diffusion resulting from the Brownian motion, and St is the Stokes number accounting for the impaction processes.

The Schmidt number is the ratio between air kinematic viscosity, $\nu = 0.146 \text{ cm}^2 \text{ s}^{-1}$, and particle diffusivity, D in $\text{m}^2 \text{ s}^{-1}$, calculated for r_{80} or r_0 in m as (McPherson, 1993):

$$D = \frac{k_B TC}{6\pi\eta r_{80}} \quad (\text{D.8})$$

where $k_B = 1.38 \times 10^{-23} \text{ J K}^{-1}$ is the Boltzmann constant.

The Stokes number is calculated as (Erickson and Duce, 1988):

$$St = \frac{V_g(r_0)u_*^2}{g\mathbf{n}} \quad (\text{D.9})$$

where $u_* \cong 0.036U_{10}$ is in cm s^{-1} and U_{10} is in m s^{-1} .

Employing this procedure, the deposition velocity, V_d , is calculated for each bin size used for the calculation of sea-salt aerosol flux. Figure D.1 plots V_d for $U_{10} = 15 \text{ m s}^{-1}$ over a wide range of sizes in order to compare with others' calculations. In the figure, V_d ranges from 10^{-2} to 10^2 cm s^{-1} and these values are comparable to the values reported by Slinn and Slinn (1980).