Supplementary Material to the Manuscript†

"Irreversible Thermodynamics of Seawater Evaporation" by R. Feistel and O. Hellmuth:

Review of Mass Transfer Coefficients Employed in the Dalton Equation

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ABSTRACT

In the present document a compilation of empirical, semiempirical, and theoretical expressions for the water-vapor pressure and specific humidity-based mass transfer coefficients employed in the Dalton equation is presented.

1. Turbulence-induced vertical transport in the atmospheric surface layer

a. Empirical approach

Daily and monthly evaporation rates can be effectively parameterized in terms of routinously measured meteorological observables among which the wind velocity is the key aerodynamic driver for water-vapor mass transfer across the sea-atmosphere interface (next to the relative fugacity or its proxies as the thermodynamic driving force of evaporation) (e.g., Wüst 1920; Jacobs 1942, 1951; Sutton and Simpson 1934; Sverdrup 1936; Penman and Keen 1948; Tomczak 1939; Brogmus 1958, 1959; Budyko 1963; Dammann 1965; Sellers 1965; Richter 1969; Kunz 1972; Richter 1977; Richter 1978; Richter et al. 1979; Dyck and Peschke 1983, pp. 137–141; Richter 1997; Vietinghoff 2000; DWA 2018, pp. 103-122, Table 15). The first trials to directly calculate the evaporation can be probably traced back to Dalton (1802), who proposed the following simple relation:

$$X_{\mathrm{E}} = f(U)\,\Delta e\,, \quad \Delta e = e_{\mathrm{eq}}(S_{\mathrm{A}}, T) - e\,. \tag{1}$$

Here, $X_{\mathrm{E}} = \{E, J_{\mathrm{E}}\}$ denotes the evaporation metrics, which can be either the evaporation velocity (or evaporation rate), E (in units of $\mathrm{m\,s^{-1}}$), or the water-vapor mass flux density, J_{E} (in units of $\mathrm{kg\,m^{-2}\,s^{-1}}$). The quantity Δe denotes the thermodynamic driving force of evaporation, given by the difference between the equilibrium water-vapor pressure, $e_{\mathrm{eq}}(S_{\mathrm{A}}, T)$, as a function of the salinity and temperature and the actual water-vapor pressure, e. The aerodynamic prefactor $f(U)$ is a nonlinear function of the wind velocity, U, and is called "wind function". The unit of $f(U)$ depends on the choice of X_{E}. In the literature, Eq. (1) is frequently given in form of a tailored equation expressed in non-SI units. The aerodynamic prefactor is not a universal (generally valid) function of U, but depends on several factors, such as the local wind field, which is influenced by topography, orographic roughness, shore conditions (morphology, vegation, house building), location of the measurement site etc. This holds true especially for inland waters. Correspondingly, there are different types of wind functions. Already Tomczak (1939) addressed the question, whether it is allowed to extrapolate the local water-vapor mass flux density, J_E, to the whole free water area, A_E, i.e., to determine the total evaporation mass flux from multiplication of the mass flux density with the free-water area, $F_E = J_E \times A_E$. The author denied the answer. Based on the evaporation theory of Sutton and Simpson (1934), Tomczak (1939) analyzed the influence of the fetch on the evaporation rate and derived an analytical expression for the wind function, which depends next to U on the degree of turbulence, and on the geometrical dimensions of the free-water area. A further discussion of this problem can be found in Richter (1969). A consequence of this problem is that the validity of empirical relations is more or less restricted to the special conditions of their derivation. Hence, special care is required when extrapolating empirical relations from one place to another. Compilations and critical reviews of empirical and semiempirical correlations for the estimation of the evaporation of free water areas can be found, e. g., in Vietinghoff (2000, pp. 51–60) and DWA (2018, p. 103-122, Table 15).

b. Rationale of the Monin-Obukhov similarity theory

In contrast to empirical and semiempirical approaches for daily and monthly means, the determination of instantaneous evaporation fluxes, e.g., in model applications, requires the explicit consideration of the atmospheric stability, which controls the turbulence exchange of water vapor in the atmospheric surface layer (ASL). Such approach requires the completion of empirical findings by additional theoretical considerations.

The Monin-Obukhov similarity theory (MOST) serves as a master theory for the treatment of ASL turbulence. This theory was originally published in Russian in 1954, later translated and published in German and English languages (e.g., Monin and Obuchow 1958; Monin and Obukhov 1990). A review of the history, assumptions, rationale, and predictive power of the MOST can be found in, e. g., Foken (2004). Owing to its widespread presence in atmospherically relevant literature, especially on ABL physics (e.g., Pal Arya 1988; Foken 1990; Schmugge and André 1991; Garratt 1992; Kaimal and Finnigan 1994; Stull 1997; Etling 2010; Foken 2016; Emeis 2022), here one can abstain from a comprehensive review of referenced sources. The explanations given below will focus on the rationale of this theory.

The MOST is based on the following assumptions:

1. The application of the MOST is restricted to the ASL, the height of which is denoted as H.
2. The universal laws predicted by the MOST rely on similarity considerations, which are typically applied in aero-hydrodynamics and thermal physics.
3. The flow is assumed to be horizontally homogeneous and to be free of acceleration.
4. Vertical motions are neglected.
5. Turbulence is assumed to be in a quasi-steady state.
6. The turbulent fluxes of momentum and heat are assumed to be independent of height, i.e., the ASL is approximated as a “constant-flux layer”.
7. In the system of equations describing the momentum, mass, and heat budgets of a thermally inhomogeneous medium, terms containing the viscosity and thermal conductivity of the fluid are neglected. Under the condition of fully developed turbulence these terms must only be considered in the description of the details of the microstructure of the wind and temperature field.
8. The differences between the temperature and the potential temperature in the ASL and their vertical gradients are negligibly small. However, in the vicinity of isothermal states, these differences become important and must be considered.

In its orginal form, the MOST is formulated for a dry ASL (Monin and Obuchow 1958), but it is commonly agreed, that the rules for the heat flux can also be applied to specific humidity or any other passive tracer, hence the heat-flux relations are applicable to scalar fluxes. Under these conditions, the MOST describes the turbulence in a thermally inhomogeneous medium by only four independent observables (e. g., Foken and Richter 1991, p. 9 therein), namely

1. the screening height z,
2. the buoyancy parameter $\beta_\mathrm{B} = g/T$ with g denoting the constant of gravity, and T the temperature at the screening height,
3. the momentum flux $J_U = -|\tau_\mathrm{Reynolds}| = -\sqrt{\tau_{xz}^2 + \tau_{yz}^2} = -\varrho_\mathrm{AV} u_\star^2$ (in units of $\mathrm{N\,m^{-2}}$), with τ_{xz} and τ_{yz} denoting the xz- and yz-component of the Reynolds stress tensor, ϱ_AV the mass density of humid air, and $u_\star$ the friction velocity, and finally
4. the sensible heat flux $J_T = \varrho_\mathrm{AV} c_\mathrm{p} \overline{(w'T')_\mathrm{s}}$ (in units of $\mathrm{W\,m^{-2}}$), with c_p denoting the specific isobaric heat capacity of humid air, $\overline{(w'T')_\mathrm{s}} = -u_\star T_\star$ the kinematic heat flux (in units of $\mathrm{K\,m\,s^{-1}}$), $T_\star$ the kinematic scaling temperature, $w = \overline{w} + w'$ the vertical velocity, and $T = \overline{T} + T'$ the potential temperature employing Reynolds decomposition in a mean and deviatoric part.

These four quantities define a dimensionless stability parameter ζ,

$$\zeta = \frac{z}{L}\,, \quad L = -\frac{u_\star^3}{\kappa \beta_\mathrm{B} \overline{(w'T')_\mathrm{s}}}\,, \tag{2}$$

with a characteristic turbulence length scale, L, serving as a key scaling length of the MOST, which found entrance in the literature as Monin-Obukhov length (MOL). The quantity κ is the Von Kármán constant. The upper limit of the ASL height is estimated to amount

$$H \approx \alpha \times 250\,\mathrm{m}\,, \quad \frac{u_\star^2(0) - u_\star^2(H)}{u_\star^2(0)} \le \alpha\,,$$

where α is a prescibed parameter controlling the vertical gradient of the momentum flux (vanishing vertical gradient at $\alpha = 0$). For $\alpha = 0.2$ the ASL height amounts $H = 50\,\mathrm{m}$. The most important prediction of the MOST is the existence of similarity laws for the vertical gradients of the horizontal wind velocity, $\vec{U} = u\,\vec{\mathrm{i}} + v\,\vec{\mathrm{j}}$, the temperature, T, and the specific humidity, q:

$$\begin{aligned} \frac{\kappa z}{\chi_\star}\frac{\partial\overline{\chi}}{\partial z} &= \Phi_\chi(\zeta)\,, \quad \overline{\chi} = \{|\overline{\vec{U}}|, \overline{T}, \overline{q}\}\,, \quad \chi_\star = \{u_\star, T_\star, q_\star\} \\ \rightsquigarrow \quad \frac{\kappa z}{u_\star}\frac{\partial|\overline{\vec{U}}|}{\partial z} &= \Phi_U(\zeta)\,, \quad \frac{\kappa z}{T_\star}\frac{\partial\overline{T}}{\partial z} = \Phi_T(\zeta)\,, \quad \frac{\kappa z}{q_\star}\frac{\partial\overline{q}}{\partial z} = \Phi_q(\zeta)\,. \end{aligned} \tag{3}$$

The quantities $\Phi_U(\zeta)$, $\Phi_T(\zeta)$, and $\Phi_q(\zeta)$ are universal similarity functions for momentum, heat, and humidity, which must be empirically determined. In the limiting case of neutral stratification one has $J_T = 0$, resulting in $L \longrightarrow \infty$ and $\zeta = 0$. The restoration of the logarithmic wind profile under neutral conditions requires $\Phi_U(0) = 1$. The quantities $u_\star$, $T_\star$, and $q_\star$ are independent of altitude and serve as characteristic scaling properties. A direct consequence of the MOST is the mutual interdependence of momentum, heat, and evaporation fluxes.

Now it is assumed, that the relations given in Eq. (3) are valid in turbulent layers of thicknesses $\Delta z|_U = z = z - z_0$, $\Delta z|_T = z = z - z_{0,T}$, and $\Delta z|_q = z = z - z_{0,q}$ for momentum, heat, and moisture transport, respectively. The quantitites z_0, $z_{0,T}$, and $z_{0,q}$ are the aerodynamic roughness lenghts for momentum, temperature, and specific humidity and define the lower height of the applicability of the MOST scaling laws. Adding a nutritive zero $1 - 1 = 0$ to the right-hand side of the first scaling relation in Eq. (3), the integration over the height from the aerodynamic roughness length z_0 to the screening height z delivers:

$$\begin{aligned}
\Delta|\overline{\vec{U}}(z)| &= |\overline{\vec{U}}(z)| - |\overline{\vec{U}}(z_0)| = \frac{u_\star}{\kappa}\int\limits_{z_0}^{z} \frac{\Phi_U(\zeta(z'))}{z'}\,\mathrm{d}z' = \frac{u_\star}{\kappa}\Bigg[\ln\frac{z}{z_0} + \underbrace{\int\limits_{z_0}^{z}\frac{\Phi_\mathrm{M}(\zeta(z'))-1}{z'}\,\mathrm{d}z'}_{=I}\Bigg] \\
I &= \int\limits_{\zeta_0}^{\zeta}\frac{\Phi_U(\zeta(z'))-1}{\zeta'}\,\mathrm{d}\zeta' = -\Bigg[\underbrace{\int\limits_{0}^{\zeta}\frac{1-\Phi_U(\zeta(z'))}{\zeta'}\,\mathrm{d}\zeta'}_{=\Psi_U(\zeta)} - \int\limits_{0}^{\zeta_0}\frac{1-\Phi_U(\zeta(z'))}{\zeta'}\,\mathrm{d}\zeta'\Bigg] \\
\rightsquigarrow \quad \Delta|\overline{\vec{U}}(z)| &= \frac{u_\star}{\kappa}\left[\ln\frac{z}{z_0} - \Psi_U\left(\frac{z}{L}\right) + \Psi_U\left(\frac{z_0}{L}\right)\right].
\end{aligned} \tag{4}$$

Analogously, carrying out the integration of the second relation in Eq. (3) from $z_{0,T}$ to z with $\overline{T}(z_{0,T}) = \overline{T}_\mathrm{s}$, and of the third relation from $z_{0,q}$ to z with $\overline{q}(z_{0,q}) = \overline{q}_\mathrm{s}$ one arrives at the following integrals:

$$\begin{aligned}
\Delta\overline{T} = \overline{T}(z) - \overline{T}_\mathrm{s} &= \frac{T_\star}{\kappa}\int\limits_{z_{0,T}}^{z}\frac{\Phi_T(\zeta(z'))}{z'}\,\mathrm{d}z' = \frac{T_\star}{\kappa}\left[\ln\frac{z}{z_{0,T}} - \Psi_T\left(\frac{z}{L}\right) + \Psi_T\left(\frac{z_{0,T}}{L}\right)\right], \\
\Psi_T(\zeta) &= \int\limits_{0}^{\zeta}\frac{1-\Phi_T(\zeta(z'))}{\zeta'}\,\mathrm{d}\zeta',
\end{aligned} \tag{5}$$

$$\begin{aligned}
\Delta\overline{q} = \overline{q}(z) - \overline{q}_\mathrm{s} &= \frac{q_\star}{\kappa}\int\limits_{z_{0,q}}^{z}\frac{\Phi_q(\zeta(z'))}{z'}\,\mathrm{d}z' = \frac{q_\star}{\kappa}\left[\ln\frac{z}{z_{0,q}} - \Psi_q\left(\frac{z}{L}\right) + \Psi_q\left(\frac{z_{0,q}}{L}\right)\right], \\
\Psi_q(\zeta) &= \int\limits_{0}^{\zeta}\frac{1-\Phi_q(\zeta(z'))}{\zeta'}\,\mathrm{d}\zeta'.
\end{aligned} \tag{6}$$

Considering $\Delta|\overline{\vec{U}}(z)|$, $\Delta\overline{T}$, and $\Delta\overline{q}$ as given, and inserting $u_\star$ from Eq. (4) together with $T_\star$ from Eq. (5) into Eq. (2), one obtains a transcendental equation for the determination of L, the knowledge of which allows the determination of $u_\star$, $T_\star$, and $q_\star$:

$$u_\star = \frac{\kappa\Delta|\overline{\vec{U}}(z)|}{\ln\frac{z}{z_0} - \Psi_U\left(\frac{z}{L}\right) + \Psi_U\left(\frac{z_0}{L}\right)}, \tag{7}$$

$$T_\star = \frac{\kappa \Delta \overline{T}}{\ln \dfrac{z}{z_{0,T}} - \Psi_T\left(\dfrac{z}{L}\right) + \Psi_T\left(\dfrac{z_{0,T}}{L}\right)}\,, \tag{8}$$

$$q_\star = \frac{\kappa \Delta \overline{q}}{\ln \dfrac{z}{z_{0,q}} - \Psi_q\left(\dfrac{z}{L}\right) + \Psi_q\left(\dfrac{z_{0,q}}{L}\right)}\,. \tag{9}$$

By virtue of $J_U = -\varrho_{\mathrm{AV}} u_\star^2$, $J_T = -\varrho_{\mathrm{AV}} c_p u_\star T_\star$, and analogously, $J_q = -\varrho_{\mathrm{AV}} L_{\mathrm{V}} u_\star q_\star$, with L_{V} denoting the specific heat of evaporation, one arrives at the following flux representations for the turbulent transfer of momentum, sensible heat, and latent heat:

$$\begin{aligned} J_U &= -\varrho_{\mathrm{AV}}\, C_U \left[\Delta |\overline{\vec{U}}(z)|\right]^2 , \\ C_U &= \frac{\kappa^2}{\left[\ln \dfrac{z}{z_0} - \Psi_U\left(\dfrac{z}{L}\right) + \Psi_U\left(\dfrac{z_0}{L}\right)\right]^2} , \end{aligned} \tag{10}$$

$$\begin{aligned} J_T &= -\varrho_{\mathrm{AV}}\, c_{\mathrm{p}}\, C_T\, \Delta |\overline{\vec{U}}(z)|\, \Delta \overline{T} , \\ C_T &= \frac{\kappa^2}{\left[\ln \dfrac{z}{z_0} - \Psi_U\left(\dfrac{z}{L}\right) + \Psi_U\left(\dfrac{z_0}{L}\right)\right]\left[\ln \dfrac{z}{z_{0,T}} - \Psi_T\left(\dfrac{z}{L}\right) + \Psi_T\left(\dfrac{z_{0,T}}{L}\right)\right]} , \end{aligned} \tag{11}$$

$$\begin{aligned} J_q &= -\varrho_{\mathrm{AV}}\, L_{\mathrm{V}}\, C_q\, \Delta |\overline{\vec{U}}(z)|\, \Delta q , \\ C_q &= \frac{\kappa^2}{\left[\ln \dfrac{z}{z_0} - \Psi_U\left(\dfrac{z}{L}\right) + \Psi_U\left(\dfrac{z_0}{L}\right)\right]\left[\ln \dfrac{z}{z_{0,q}} - \Psi_q\left(\dfrac{z}{L}\right) + \Psi_q\left(\dfrac{z_{0,q}}{L}\right)\right]} . \end{aligned} \tag{12}$$

Here, C_U, C_T, and C_q denote the drag coefficient, and the Stanton and Dalton numbers, respectively. The application of Eqs. (10), (11), and (12) requires the specification of the similarity functions, Φ_χ, and the boundary conditions at z_0, $z_{0,T}$, and $z_{0,q}$. The availability of the Dalton number, C_q, allows a unique determination of the vapor-pressure and specific humidity-based transfer coefficients, D_e and $D_q = C_q\, \Delta |\overline{\vec{U}}(z)|$, appearing in the Dalton equation.

c. Applicational aspects and compilation of transfer coefficients

The MOST serves as the theoretical fundament of countless meteorological and metrological applications in parameterizing and measuring near-surface turbulent fluxes. For example, the MOST is part of the flux parameterization in several community models for numerical weather prediction (NWP) and global circulation models (GCM), such as,

- the NWP model COSMO of the German Weather Service (e. g., Doms et al. 2013),
- the Integrated Forecast System (IFS) of the European Center for Medium-Range Weather Forecasts (ECMWF) (e. g., ECMWF-IFS 2021, `https://www.ecmwf.int/en/publications/ifs-documentation`),
- the NCAR-GCM (e. g., Large and Pond 1981, 1982; Large et al. 1997; Large and Yeager 2004, 2009; Brodeau et al. 2017), and
- the TOGA-COARE-GCM (e. g., Webster and Lukas 1992; Fairall et al. 1996b,a, 1997; Andreas 2003; Brunke et al. 2003; Fairall et al. 2003a,b; Zeng et al. 2003; Andreas et al. 2008; Fairall et al. 2011; Edson et al. 2013; Yusup et al. 2018).

While retaining the basic physical assumptions of the theory, the specific implementation is subject to manifold modifications and enhancement to continuously ensure best agreement with available state-of-the-art empirical and theoretical findings and to remove still existing biases in the flux parameterization (e.g., Yu 2019). The variety of specifications concerns, e.g.,

- the specification of the empirical similiarity functions, Φ_χ, underlying the determination of the stability functions, Ψ_χ, and the drag coefficient and the Stanton and Dalton numbers (e. g., Dyer and Hicks 1970; Paulson 1970; Businger et al. 1971; Kaimal et al. 1972, 1990; Dyer 1974; Skeib 1980; Foken and Skeib 1983; Skeib and Richter 1984; Holtslag 1987; Högström

1988; Foken 1990, 1991; Garratt 1992; Kaimal and Finnigan 1994; Fairall et al. 1996b; ECMWF-IFS 2021),

- the parameterization of the surface-roughness lengths, z_0, $z_{0,T}$, and $z_{0,q}$ (e. g., Smith 1988; Miller et al. 1992; Beljaars 1995; Fairall et al. 1996b,a, 1997, 2003a; Large and Yeager 2009; Fairall et al. 2011; Doms et al. 2013; Edson et al. 2013; Liu et al. 2013), and

- the refinement of the description of the exchange processes in the molecular boundary layer and the viscous intermediate (buffer) layer, i. e., at $0 \le z \le \{z_0, z_{0,T}, z_{0,q}\}$ (e. g., Owen and Thomson 1963; Kondo 1975; Foken et al. 1978; Liu et al. 1979; Foken 1979b,a, 1984, 1986; Foken and Richter 1991; Foken and Skeib 1991; Richter and Skeib 1991), e.g., within the framework of the surface-renewal theory (e. g., Brutsaert 1975; Soloviev and Schlüssel 1994, 1998; Clayson et al. 1996; Fairall et al. 1996a; Zappa et al. 1998; Mengistu and Savage 2010; Horvath and Chatterjee 2018; Hu et al. 2018).

Table 2 contains a compilation of selected empirical and semi-empirical expressions of the vapor-pressure and specific humidity-based transfer coefficients, D_e and D_q, appearing in the Dalton equation.

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LIST OF TABLES

Table 1. List of Symbols.

Symbol	Unit	Value	Meaning
a	1		Regression parameter
a_q	1		Regression parameter
A_E	m^2		Evaporation surface area
b	1		Regression parameter
b_q	1		Regression parameter
b_corr	$\mathrm{m\,s^{-1}}$		Regression parameter
c_q	1		Regression parameter
c_p	$\mathrm{J\,kg^{-1}\,K^{-1}}$		Specific heat capacity of humid air
C_q	1		Bulk-transfer coefficient for moisture (Dalton number)
C_q^N	1		Bulk-transfer coefficient for moisture for neutral stratification
C_T	1		Bulk-transfer coefficient for temperature (Stanton number)
C_U	1		Bulk-transfer coefficient for momentum (drag coefficient)
C_U^N	1		Bulk-transfer coefficient for momentum for neutral stratification
D_e	$\mathrm{m^2\,s\,kg^{-1}}$		Vapor-pressure based mass-transfer coefficient
D_q	$\mathrm{m\,s^{-1}}$		Specific-humidity based mass-transfer coefficient
e	Pa		Actual water-vapor pressure
e_eq	Pa		Equilibrium water-vapor pressure
E	$\mathrm{m\,s^{-1}}$		Evaporation velocity (evaporation rate)
f_U	$\mathrm{m^2\,s\,kg^{-1}}$		Wind function
F_E	$\mathrm{kg\,s^{-1}}$		Water-vapor mass flux (evaporation mass flux)
F_q	1		Stability correction function for moisture
g	$\mathrm{m\,s^{-2}}$		Gravitational constant
H	m		Height of the ASL

Continuation of Table 1.

Symbol	Unit	Value	Meaning
J_q	$\mathrm{W\,m^{-2}}$		Turbulent flux of latent heat
J_T	$\mathrm{W\,m^{-2}}$		Turbulent flux of sensible heat flux
J_U	$\mathrm{N\,m^{-2}}$		Turbulent flux of momentum
J_E	$\mathrm{kg\,m^{-2}\,s^{-1}}$		Water-vapor mass flux density (evaporation mass flux density)
K	$\mathrm{Pa^{-1}}$		Pressure coefficient
K_m	$\mathrm{m^2\,s^{-1}}$		Molecular exchange coefficient
$K_{\mathrm{m},U}$	$\mathrm{m^2\,s^{-1}}$		Molecular exchange coefficient for momentum
L	m		Monin-Obukhov length
L_V	$\mathrm{J\,kg^{-1}}$		Specific heat of evaporation
n	1		Parameter in the Skeib-similarity function
n_turb	1		Metric for the degree of turbulence
p_q	1		Regression parameter
$\mathrm{Pr_m}$	1		Molecular Prandtl number
q	$\mathrm{kg\,kg^{-1}}$		Specific humidity (mass fraction of water vapor in humid air)
q_{10}	$\mathrm{kg\,kg^{-1}}$		Specific humidity at $z = 10\,\mathrm{m}$
q_eq	$\mathrm{kg\,kg^{-1}}$		Equilibrium specific humidity
$q_\star$	$\mathrm{kg\,kg^{-1}}$		Kinematic scaling specific humidity
$\mathrm{Ri_0}$	1		Bulk Richardson number between the surface and the first model layer
S	$\mathrm{K\,s\,m^{-2}}$		Stability parameter
S_R	$\mathrm{K\,s\,m^{-2}}$	1	Reference stability parameter
$\widehat{S}$	1		Dimensionless stability parameter
S_A	$\mathrm{kg\,kg^{-1}}$		Salinity (mass fraction of sea salt in seawater)

Continuation of Table 1.

Symbol	Unit	Value	Meaning
T	K		Temperature
T_{A}	K		Air temperature
T_{SST}	K		Sea surface temperature
$T_{\star}$	K		Kinematic scaling temperature
u	$\mathrm{m\,s^{-1}}$		x-component of wind vector
$u_{\star}$	$\mathrm{m\,s^{-1}}$		Friction velocity
U	$\mathrm{m\,s^{-1}}$		Wind speed
U_{R}	$1\,\mathrm{m\,s^{-1}}$	1	Reference wind speed
U_1	$\mathrm{m\,s^{-1}}$		Wind speed at the first model level
U_{10}	$\mathrm{m\,s^{-1}}$		Wind speed at $z = 10\,\mathrm{m}$
$\widehat{U}$	1		Normalized wind speed
v	$\mathrm{m\,s^{-1}}$		y-component of wind vector
w	$\mathrm{m\,s^{-1}}$		z-component of wind vector
x	1		Auxiliary function
z	m		Height
z_0	m		Surface-roughness length for momentum (aerodynamic roughness)
$z_{0,\mathrm{q}}$	m		Surface-roughness length for specific humidity
$z_{0,\mathrm{T}}$	m		Surface-roughness length for temperature
z_1	m		Height of the first model level above the surface

Continuation of Table 1.

Symbol	Unit	Value	Meaning
α	1		Relative change of the momentum flux in the ASL
α_{m}	$\mathrm{m^2\,s^{-1}}$		Thermal diffusivity of air
β	1		Parameter
β_{B}	$\mathrm{m\,s^{-2}\,K^{-1}}$		Buoyancy parameter
γ	1		Parameter
δ_{m}	m		Thickness of the molecular diffusion layer
$\delta_{\mathrm{m},q}$	m		Thickness of the molecular boundary layer for water vapor
$\widehat{\delta}_q$	1		Dimensionless thickness of the molecular boundary layer for water vapor
ΔT	K		Temperature difference
ε	1		Tubulence parameter
ζ	1		Stability parameter at z
ζ_0	1		Stability parameter at $z = z_0$
ζ_{c}	1		Critical stability parameter defined in the Skeib-similarity function
η			Physical parameter
κ	1	0.4	von-Kármán constant
ν	$\mathrm{m^2\,s^{-1}}$		Kinematic viscosity of air
ξ	1		Wind function
ϱ_{AV}	$\mathrm{kg\,m^{-3}}$		Mass density of humid air
ϱ_{W}	$\mathrm{kg\,m^{-3}}$		Mass density of water
ϑ	°C		Celsius temperature

Continuation of Table 1.

Symbol	Unit	Value	Meaning
τ	$\mathrm{N\,m^{-2}}$		Reynolds stress
θ_{v}	K		Absolute virtual potential temperature
$\theta_{\mathrm{v,s}}$	K		Absolute virtual potential temperature at the surface
$\theta_{\mathrm{v,1}}$	K		Absolute virtual potential temperature at the first model level
Φ_q	1		Similarity function for specific humidity
Φ_T	1		Similarity function for temperature
Φ_U	1		Similarity function for momentum
Φ_χ	1		Similarity function for quantity χ
χ	$[\chi]$		Placeholder for any physical quantity
$\chi_\star$	$[\chi]$		Kinematic scaling value of quantity χ
Ψ_q	1		Stability function for specific humidity
Ψ_T	1		Stability function for temperature
Ψ_U	1		Stability function for momentum

Table 2. Compilation of water-vapor mass transfer coefficients from the literature (selection).

Vapor-pressure based formulation of Daltons evaporation formula (e.g., Wüst 1920; Sutton and Simpson 1934; Sverdrup 1937; Tomczak 1939; Jacobs 1942, 1951; DWA 2018)	
$E = \frac{J_E}{\varrho_W} = \frac{J_q}{\varrho_W L_V} = D_e[e_{eq}(S_A, T) - e]$	
$D_e = a\left(1 + 3.66 \cdot 10^{-3}\vartheta/°C\right)\left(1 + bU/(km\,h^{-1})\right)$, $a = (4.6 - 5.64) \cdot 10^{-11}\,m^2\,s\,kg^{-1}$, $b = 0.075 - 0.11$	(Wüst 1920, pp. 39-41)
$D_e = a\,(U/U_R)$, $a = (0.94 - 1.24) \cdot 10^{-11}\,m^2\,s\,kg^{-1}$	(Jacobs 1942, 1951)
$D_e = \omega\left[a + b(U/U_R)^{\varepsilon}\right]$, $\varepsilon = \frac{2 - n_{turb}}{2 + n_{turb}}$	(Sutton and Simpson 1934; Tomczak 1939)
$a = 0.16$, $b = 0.18$, $\varepsilon = 0.84$, $\omega = 8.7 \cdot 10^{-11}\,m^2\,s\,kg^{-1}$	(Richter 1969, p. 16, Stechlinsee, Brandenburg);
$D_e = f_U(U)$	(Compilation of different wind functions: Vietinghoff 2000, pp. 51-60; DWA 2018, p. 117, Table 15)
Specific-humidity based formulation of Daltons evaporation formula (e.g., Budyko 1963, pp. 74-76, Eqs. (68), (69), (73), Sellers 1965, p. 145, Eq. (10.9), Kondo 1975, Eq. (2), Liu et al. 1979, Eq. (1b), Foken 1984, Eq. (10), Smith 1988, Eq. (3), Pal Arya 1988, p. 188, Eq. (12.13), Kondo et al. 1990, Eqs. (1), (2), Ruprecht and Simmer 1991, Section 2, Schmugge and André 1991, p. 72, Eqs. (4.6)-(4.8), Garratt 1992, p. 56, Eq. (3.51), Miller et al. 1992, Eq. (1), Kraus and Businger 1994, pp. 180–181, Kaimal and Finnigan 1994, p. 13, Eq. (1.17); p. 70, Eq. (3.13), Zhang and McPhaden 1995, Eq. (1b), Stull 1997, p. 263, Eq. (7.4.1e), Fairall et al. 2003a, Liu et al. 2007, Yu 2007, Eq. (1), Stewart 2008, p. 58, Eq. (5.10c), Table 5.1, Liu et al. 2013, Eq. (5), Rosenberg 2016, Eq. (2), Kumar et al. 2017, Eq. (1), Cronin et al. 2019, Eq. (1.6b), Yu 2019, Eq. (3), Hogan 2020, Reeves Eyre et al. 2021, Eq. (3), Table 1, Song et al. 2022, Eq. (1))	
$J_q = \varrho_{AV} L_V D_q[q_{eq}(S_A, T) - q]$, $D_q = C_E\,U$	
$D_q = \frac{K_m U}{U\,\delta_m + \frac{K_m}{\kappa^2}\left(\frac{z_1 + z_0}{z_0}\right)^2}$, $\delta_m = (1 - 1.5) \cdot 10^{-3}\,m$	(Sverdrup 1936, Budyko 1963, Eq. (69))
$D_q = a + b\,(U/U_R)$, $a = (7 - 45) \cdot 10^{-4}\,m\,s^{-1}$, $b = 2 \cdot 10^{-3}\,m\,s^{-1}$ ($a = 7 \cdot 10^{-4}\,m\,s^{-1}$ for large lakes)	(Sellers 1965, p. 159, Eq. (11.2))
$D_q = 5 \cdot 10^{-3}\,m\,s^{-1}$	(Kaimal and Finnigan 1994, pp. 73-76, Eq. (3.21), Table 3.1)
$D_q = \frac{\kappa u_\star}{\kappa\,\mathrm{Pr}_m\,\widehat{\delta}_{m,q} + 4\kappa + \ln\left(\frac{\zeta_c\,L u_\star}{20 K_{m,U}}\right) + \frac{1}{n}\left[1 - \left(\frac{z}{\zeta_c\,L}\right)^{-n}\right]}$	(Foken 1984, Eq. (10))
$\widehat{\delta}_{m,q} = \frac{\delta_{m,q}\,u_\star}{K_{m,U}} = 6$, $\{\zeta_c, n\} = \begin{cases} \{-0.075, 0.5\}, & -2.5 \le \zeta \le -0.075 \\ \{0.16, -2\}, & 0.16 \le \zeta \le 1 \end{cases}$	(Skeib and Richter 1984)

Continuation of Table 2.

Specific-humidity based formulation of Daltons evaporation formula with correction for subsurface sea-temperature effects according to Kruspe (1977, Eq. (3), Fig. 2)
$J_q = \varrho_{AV} L_V \left[C_q U_{10} [q_{eq}(S_A, T) - q_{10}] + b_{corr} \right], \quad C_q \times 10^3 = 1.36 \pm 0.25, \quad b_{corr} = -0.7 \cdot 10^{-5}\,\mathrm{m\,s^{-1}}$
Formulations of the Dalton number for neutral, stable, and unstable stratification from different sources

$$C_q^N \times 10^3 = \begin{cases} 1.15 & \text{(Large and Pond 1982, Table 2, for unstable conditions)} \\ 1-5 & \text{(Stull 1984, p. 263, Eq. (7.4.1e))} \\ 1.1\ (\pm 15\,\%) & \text{(Garratt 1992, p. 101-104, Fig. 4.9)} \\ 1.2 & \text{(Smith 1988, best estimate from compilation of data, Kraus and Businger 1994, p. 181 for } U = (5-20)\,\mathrm{m\,s^{-1}}\text{, Stewart 2008, Table 5.1, Hogan 2020, Eq. (29))} \\ (1.1-1.2)@U < 10\,\mathrm{m\,s^{-1}},\ 1.45@U \approx 20\,\mathrm{m\,s^{-1}} & \text{(Fairall et al. 2003a, Fig. 3)} \\ 1.2-1.5 & \text{(Liu et al. 2007, Table 3)} \\ 1.3 & \text{(Kumar et al. 2017, Eq. (1))} \end{cases}$$

$$C_q = \begin{cases} \text{LUT as function of } U \text{ and } \Delta T = T_A - T_{SST} & \text{(Bunker 1976, Table 4)} \\ C_q^N \times \left\{ \begin{array}{ll} (1-\zeta)\left[1+10^{-2}\beta^{3/4}\right] & \text{at } \zeta < 0 \text{ (unstable)} \\ \left[\dfrac{1}{1+3.5\zeta}\right]\left[1+10^{-2}\beta^{3/4}\right] & \text{at } \zeta > 0 \text{ (stable)} \\ \zeta = \dfrac{z}{L},\ z_0 = z\exp\left(-\dfrac{\kappa U}{u_\star}\right),\ \beta = \dfrac{z_0 u_\star}{\nu} & \end{array} \right\} & \text{(Panin et al. 2006, for deep-water air-sea interaction)} \\ \dfrac{\kappa^2}{\left[\ln\dfrac{z}{z_0} - \Psi_U(\zeta)\right]\left[\ln\dfrac{z}{z_0} - \Psi_q(\zeta)\right]}, \quad \zeta = \dfrac{z}{L} & \text{(Pal Arya 1988, p. 167, Eq. (11.16), Garratt 1992, p. 55, Eqs. (3.47)-(3.51), Stull 1997, p. 267, Eq. (7.4.1m))} \\ \dfrac{\kappa^2}{\left[\ln\dfrac{z}{z_0} - \Psi_U(\zeta) + \Psi_U(\zeta_0)\right]\left[\ln\dfrac{z}{z_0} - \Psi_q(\zeta) + \Psi_q(\zeta_0)\right]}, \quad \zeta_0 = \dfrac{z_0}{L} & \text{(Liu et al. 2013, Eqs. (7), (8))} \end{cases}$$

Stability function (e.g., Paulson 1970, Pal Arya 1988, p. 167, Eq. (11.14), Zhang and McPhaden 1995):

$$\Psi_U(\zeta) = \begin{cases} -5\zeta & \text{for } \zeta \geq 0 \\ \ln\left[\left(\dfrac{1+x^2}{2}\right)\left(\dfrac{1+x}{2}\right)^2\right] - 2\arctan x + \dfrac{\pi}{2}, \quad x = (1-15\zeta)^{1/4} & \text{for } \zeta < 0 \end{cases}$$

$$\Psi_T(\zeta) = \Psi_q(\zeta) = \begin{cases} -5\zeta & \text{for } \zeta \geq 0 \\ 2\ln\left(\dfrac{1+x^2}{2}\right), \quad x = (1-15\zeta)^{1/4} & \text{for } \zeta < 0 \end{cases}$$

Continuation of Table 2.

Parameterization of the Dalton number according to Kondo (1975) on the base of the Owen-Thomson theory of evaporation (Owen and Thomson 1963) (see also Garratt (1992, pp. 90-93) for interfacial sublayer relations)

(A) Parameterization for neutral (adiabatic) stratification (Kondo 1975, Fig. 5, Appendix 1, Table A1):

$$C_q^{\mathrm{N}}(10\,\mathrm{m})\times 10^3 = a_q + b_q\left(\frac{U_{10}}{U_{\mathrm{R}}}\right)^{p_q} + c_q\left[\frac{U_{10}}{U_{\mathrm{R}}} - 8\right]^2$$

$U_{10}/(\mathrm{m\,s^{-1}})$	a_q	b_q	c_q	p_q
0.3 – 2.2	0	1.23	0	−0.16
2.2 – 5	0.969	0.0521	0	1
5 – 8	1.18	0.01	0	1
8 – 25	1.196	0.008	−0.0004	1
25 – 50	1.68	−0.016	0	1

(B) Parameterization for diabatic stratification (Kondo 1975, Appendix 3):

$$C_q \approx C_q^{\mathrm{N}} \times \begin{cases} 0.01 + 0.03\,\widehat{S} + 0.9\exp(4.8\,\widehat{S}) & \text{for} \quad T_{\mathrm{SST}} - T < 0 \text{ and } -3.3 < \widehat{S} < 0 \text{ (stable stratification)} \\ 0 & \text{for} \quad T_{\mathrm{SST}} - T < 0 \text{ and } \widehat{S} \leq -3.3 \text{ (very stable stratification)} \\ 1.0 + 0.63\,\widehat{S}^{1/2} & \text{for} \quad T_{\mathrm{SST}} - T > 0 \text{ (unstable stratification)} \end{cases}$$

$$\widehat{S} = \widehat{S}_0 \frac{|\widehat{S}_0|}{|\widehat{S}_0| + 0.01}\,, \quad \widehat{S}_0 = \frac{S_0}{S_{\mathrm{R}}}\,, \quad S_0 = \frac{T_{\mathrm{SST}} - T}{\left(U\left[1.0 + \log_{10}\left(\frac{10\,\mathrm{m}}{z}\right)\right]\right)^2}$$

Continuation of Table 2.

Parameterization of the Dalton number according to Miller et al. (1992, Eqs. (2), (3), Fig. 3)

$$C_q = C_q^{\mathrm{N}} F_q\left(\mathrm{Ri}_0, \frac{z_1}{z_0}, \frac{z_1}{z_{0,T}}, \frac{z_1}{z_{0,q}}\right), \quad C_q^{\mathrm{N}} = \frac{\kappa^2}{\left[\ln\frac{z_1}{z_0}\right]\left[\ln\frac{z_1}{z_{0,q}}\right]}$$

Rough-wind regime (Charnock relation):

$$z_0 = \max\left(0.018\,\frac{u_\star^2}{g}, 1.5\cdot 10^{-5}\,\mathrm{m}\right), \quad z_0 = z_{0,T} = z_{0,q}$$

Weak-wind regime (smooth surface):

$$z_0 = 0.11\,\frac{\nu}{u_\star} + 0.018\,\frac{u_\star^2}{g}, \quad z_{0,T} = 0.40\,\frac{\nu}{u_\star} + 1.4\cdot 10^{-5}\,\mathrm{m}, \quad z_{0,q} = 0.62\,\frac{\nu}{u_\star} + 1.3\cdot 10^{-4}\,\mathrm{m}$$

Free-convection limit of ocean evaporation:

$$J_q = \varrho_{\mathrm{AV}} L_{\mathrm{V}} D_q' [q_{\mathrm{eq}}(S_{\mathrm{A}}, T) - q_1], \quad D_q' = 0.17\left(\frac{g\,\alpha_{\mathrm{m}}^2}{\nu\,\theta_{\mathrm{v}}}\right)^{1/3} \left(\theta_{\mathrm{v,s}} - \theta_{\mathrm{v,1}}\right)^{1/3}$$

Dalton number with consideration of the free-convection limit:

$$C_q = C_q^{\mathrm{N}}\left(1 + C_{\mathrm{R}}^{\gamma}\right)^{1/\gamma}, \quad C_q^{\mathrm{N}} = \left(\frac{\kappa}{\ln(z_1/z_0)}\right)^2, \quad C_{\mathrm{R}} = \frac{\eta}{C_q^{\mathrm{N}} U_1}\left(\theta_{\mathrm{v,s}} - \theta_{\nu,1}\right)^{1/3}$$

$$\gamma = 1.25, \quad \eta = 0.0016\,\mathrm{m\,s^{-1}\,K^{-1/3}}$$

z_0 = from the Charnock formula

Continuation of Table 2.

Parameterization of Dalton number according to Large and Yeager (2004, Eq. (6))
$C_q \times 10^3 = \begin{cases} 18.0\sqrt{C_U} & , \quad \zeta > 0 \\ 32.7\sqrt{C_U} & , \quad \zeta \leq 0 \end{cases} \quad , \quad C_U \times 10^3 = \dfrac{2.7\,\mathrm{m\,s^{-1}}}{U_{10}} + 0.142 + \dfrac{U_{10}}{13.09\,\mathrm{m\,s^{-1}}}$
Parameterization of Dalton number according to Large and Yeager (2009)
$C_q^{\mathrm{N}} = \dfrac{\kappa\sqrt{C_U^{\mathrm{N}}}}{\ln\dfrac{10\,\mathrm{m}}{z_{0,q}}} , \quad C_U^{\mathrm{N}} = \begin{cases} \dfrac{a_1}{\widehat{U}_{10}} + a_2 + a_3\,\widehat{U}_{10} + a_8\left[\widehat{U}_{10}\right]^6 & , \quad \widehat{U}_{10} \geq 33 \\ 0.00234 & , \quad \widehat{U}_{10} < 33 \end{cases}$ $\widehat{U}_{10} = \dfrac{U_{10}}{U_{\mathrm{R}}} , \quad a_1 = 2.7\cdot 10^{-3} , \; a_2 = 1.42\cdot 10^{-4} , \; a_3 = 7.64\cdot 10^{-5} , \; a_8 = -3.14807\cdot 10^{-13}$ $\dfrac{\kappa}{\ln\dfrac{10\,\mathrm{m}}{z_{0,q}}} = 0.0346$
Parameterization of the wind function ξ in dependence on the wind velocity, the difference between water and air temperatures, and on the humidity deficit according to Babkin (2023, Eq. (1))
$E = \xi(U)\,U\left[q_{\mathrm{eq}}(S,T) - q\right]$