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Latent and sensible heat fluxes under active and weak phases of the summer monsoon of 1986

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Abstract. An analysis of the meteorological data collected by the research vessel ORV Sagarkanya for the mean latent and sensible heat fluxes over the Arabian Sea has indicated appreciable changes between active and weak phases of the southwest monsoon of 1986. We suggest that: (a) the presence of a core of low level winds associated with the Somali jet and its southward shift during the season, along with (b) a ridge in surface pressure over the central Arabian Sea could be responsible for the deficit in monsoon rainfall along the west coast of India in 1986.

Keywords. Latent heat; sensible heat; active phase; weak phase; summer monsoon; Arabian Sea.

1. Introduction

Changes in heat and momentum fluxes over the Indian Ocean during the summer monsoon influence the overlying boundary-layer and, ultimately, the rainfall over the Indian sub-continent. In this context valuable data have been collected through large scale experiments, such as IIOE, ISMEX-73, MONSOON-77 and MONEX-79. Climatological atlases for the surface variables including the monthly heat fluxes (Ramage *et al* 1972; Hastenrath and Lamb 1979a, b) are useful additions to existing knowledge.

Studies on the influence of evaporation and moisture fluxes over the Arabian Sea on the rainfall distribution along the Indian peninsula have been pursued following the work of Pisharoty (1965). These studies have been critically examined by a recent study of Cadet and Greco (1987). Sadhuram and Ramesh Kumar (1988) have emphasized the need for a more broad based systematic study. In this paper, we will discuss the distribution of latent and sensible heat fluxes under active and weak phases of the summer monsoon (table 1). The relevant data were collected during the 1986 summer monsoon by ORV Sagarkanya. The cruise tracks are shown in figure 1.

2. Data and methodology

Continuous observations of the relevant surface meteorological parameters were made at 3-hourly intervals by the research vessel ORV Sagarkanya during the period July 4 to September 7 1986 along the tracks shown in figure 1. The observations were grouped in $4^\circ \times 4^\circ$ grids for the active and weak phases of the monsoon to represent

Table 1. Percentage departures of rainfall along the west coast of India.

Station	Active phase	Weak phase
Bombay	16	−79
Panaji	04	−86
Mangalore	40	−90
Cochin	57	−78
Trivandrum	69	−91
Mean	37	−85

Active phase 9–23 July; 3–14 Aug 1986.
Weak phase: 2–8 July; 24 July–2 Aug; 15 Aug–5 Sept 1986.

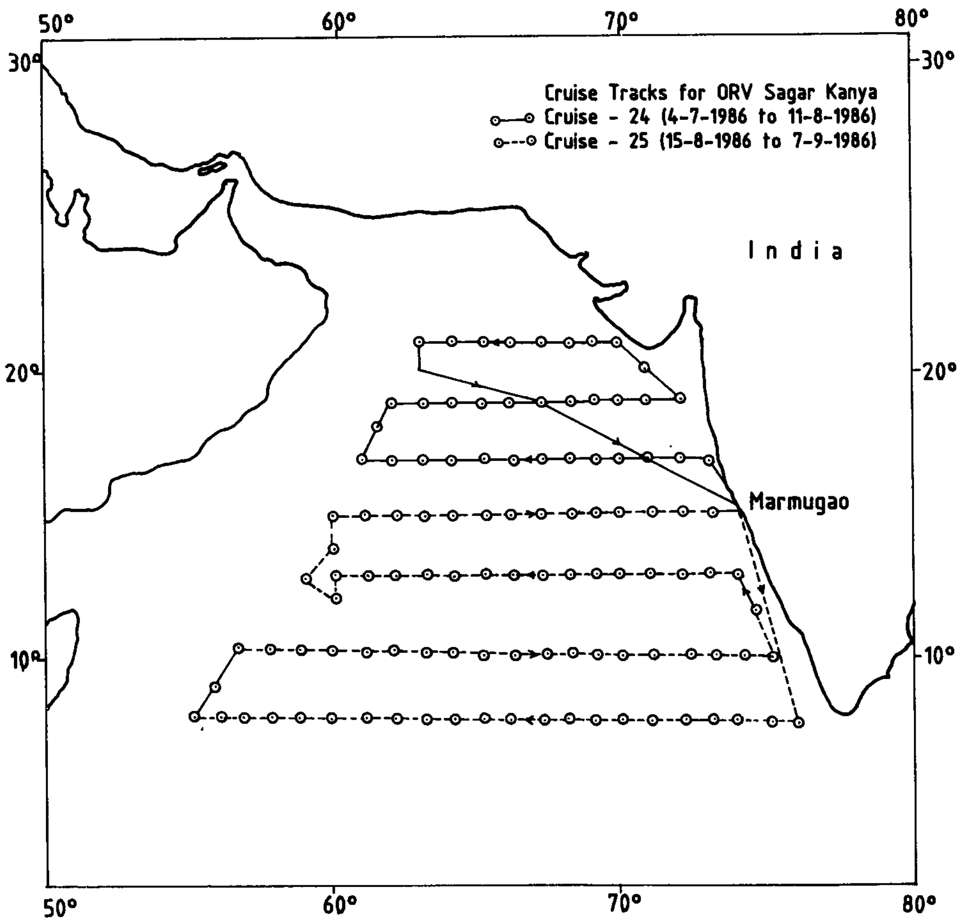

Figure 1. Cruise tracks in the Arabian Sea during 1986 summer monsoon.

the mean conditions of the two contrasting phases of the monsoon. During the survey period the weather was free from any large scale transient disturbance.

The fluxes (Wm^{-2}) of latent heat and sensible heat were computed by the bulk-aerodynamic method. We put

$$Q_e = \rho C_e L U(q_s - q_a) \tag{1}$$

$$Q_s = \rho C_p C_h U(\theta_s - \theta_a) \tag{2}$$

where Q_e is the latent heat flux, Q_s the sensible heat flux, ρ the density of air ($1{\cdot}16\,kgm^{-3}$), L the latent heat of vaporization ($2440 \times 10^3\,J\,kg^{-1}$), C_p the specific heat of air at constant pressure ($1020\,J\,kg^{-1}\,{}^{\circ}K^{-1}$), U the wind speed (ms^{-1}), q_s and q_a are specific humidities ($kg\,kg^{-1}$) at sea surface temperature and at the observational level (22 m from the sea surface) respectively, θ_s and θ_a, the potential temperature of the sea surface and air respectively and C_e and C_h are the transfer coefficients for latent and sensible heat fluxes respectively.

Values of these coefficients were taken from Masagutov (1981) which were tabulated by Blanc (1985). Masagutov solved a series of five transcedental equations by successive approximations, on a computer, to determine the explicit form of dependence of exchange coefficients upon wind speed and stability. He made use of the eddy correlation measurements of meteorological parameters for the computation of coefficients. The coefficients for a wind speed range of 2 to 21 ms^{-1} and sea-air virtual potential temperature difference of -4 to $+4$°C are presented in the form of a nomogram. The coefficients are the same for heat and moisture and decrease with stability.

As the data were collected on a ship one would expect some influence of the spatial and temporal variations in the meteorological variables while estimating the heat fluxes (Weare and Strub 1981). In addition, the height of the measurement on a ship also influences the estimates if the data were not collected at standard anemometric height of 10 m (Blanc 1986). In the present study, the data were collected at a height of 22 m. The wind speed, the sea-air temperature difference and the specific humidity gradients were accordingly reduced to 10 m height following the procedure given by Stevenson (1982) and the details are presented in Appendix 1. From table 2 one can see that the positive bias due to height correction was about 4% for wind speed, and 6% for latent heat flux (Q_e). Height corrections were applied in the final presentation of the mean fields.

Table 2. Percentage error in the estimation of latent heat flux due to height correction.

Phase	Wind speed (m/s)	$T_s - T_a$ (°C)	$q_s - q_a$ (kg/kg)	Q_e (W/m^2)
Active	4·3	1·2	1·0	5·6
Weak	4·2	0·9	1·7	5·8

$T_s - T_a$: Sea-air temperature difference; $q_s - q_a$: specific humidity gradient. Meteorological parameters measured at 22 m height are corrected for 10 m height following the procedures given by Stevenson (1982).

3. Results

3.1 *Surface meteorological conditions*

3.1a *Surface pressure and wind speed*: The meridional gradients of surface pressure and wind speed are greater during an active phase compared to those in a weak phase (figures 2a–d). Over the eastern and central parts of the Arabian Sea, the wind speed was greater during an active phase and less in the weak phase while the wind intensification is moderate in the western part (figures 2c and d). A ridge existed across the southern Arabian Sea around 65°E, and a trough was noticed off the west coast of India during a weak phase. This influenced the wind field over the central and eastern parts. The axis of the core of the maximum winds (~ 15 m/s) was aligned in a zonal direction rather than the usual southwest-northeast direction—a feature often observed under both active and normal monsoon conditions. This departure reflects a southward shift in the axis of the Somali jet from its normal position. A similar feature was reported during the month of June in a drought year (Krishnamurti *et al* 1979). The split in the Somali jet occurs around 10°N; 50°E with the southwest-northeast oriented northern dome striking the west coast of India normally around 18°N; 73°E, while the southern part extends across the Arabian Sea up to Sri Lanka (Findlater 1971). However, in the present study, the Somali jet seemed to extend right across the southern Arabian Sea with less normal component to the west coast of India.

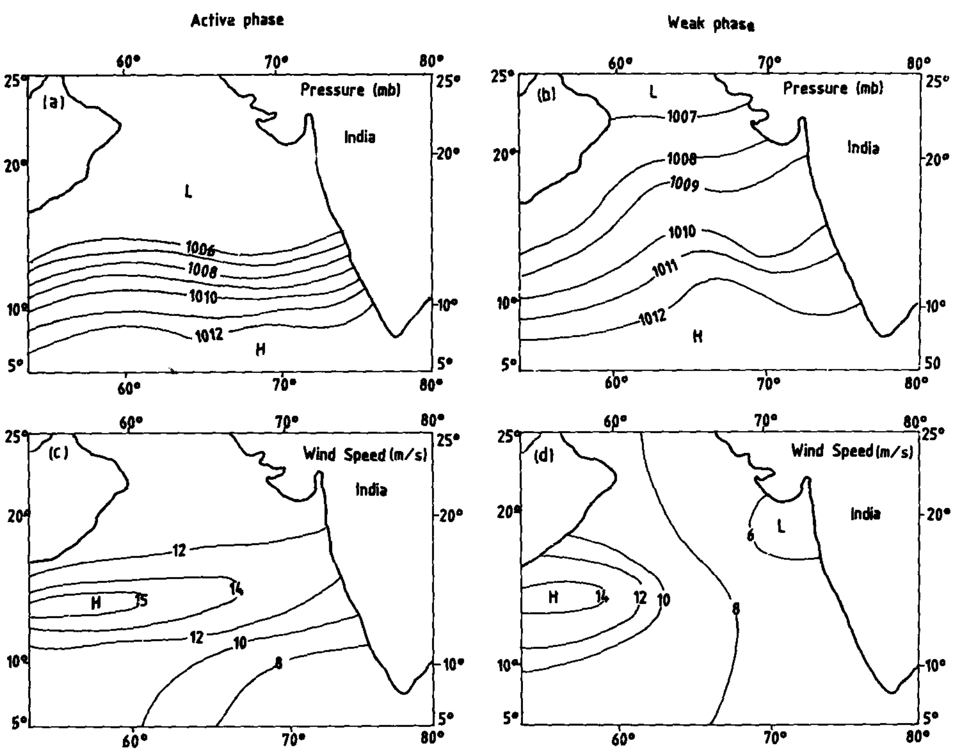

Figure 2. Mean fields of surface pressure (**a** and **b**) and wind speed (**c** and **d**) under active and weak phases of the monsoon.

3.1b *Sea surface temperature (SST) and sea-air potential temperature difference.* SST variations ($\sim 4°C$) across the Arabian Sea with the incidence of high SST values in the east and low values in the west indicate the influence of upwelling in the western region. The spreading of isotherms in the western region together with slight surface warming in the eastern region during the weak phase of the monsoon is an outcome of weakening in the wind regime (figures 3a and b). In general the difference in potential temperature is less than zero over most of the region and especially in the western part where high negative values ($>1{\cdot}0$) occur. This suggests the advection of cold upwelled waters from Somali coast. In a weak phase, the difference is positive and lies between 1 and 1·4°C in the southeastern region (figures 3c and d).

3.2 *Vapour pressure gradient and latent heat flux (Q_e)*

From the distribution of vapour pressure gradient (figures 4a and b), one could see that they were controlled by the SST, especially in the weak phase. Low values were confined to the western region during both phases while high values prevailed over the southeastern section. Figures 4c and d depict the distribution of latent heat flux (Q_e) under active and weak phases of the monsoon. Under an active phase, a zone of maximum Q_e ($>260\,\mathrm{Wm^{-2}}$) existed around 14°N; 65°E and the flux decreases both northward and southward of this location. The magnitude of the flux is low in the southeastern part where the influence of wind speed, rather than the vapour

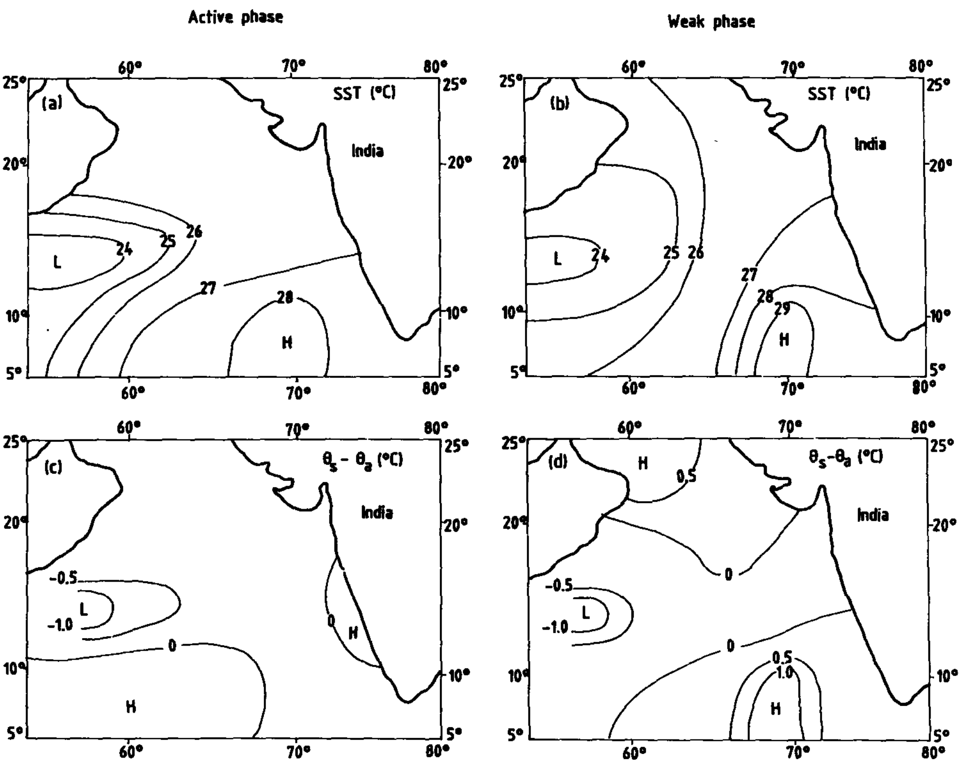

Figure 3. Same as figure 2 but for SST (**a** and **b**) and sea-air potential temperature difference (**c** and **d**).

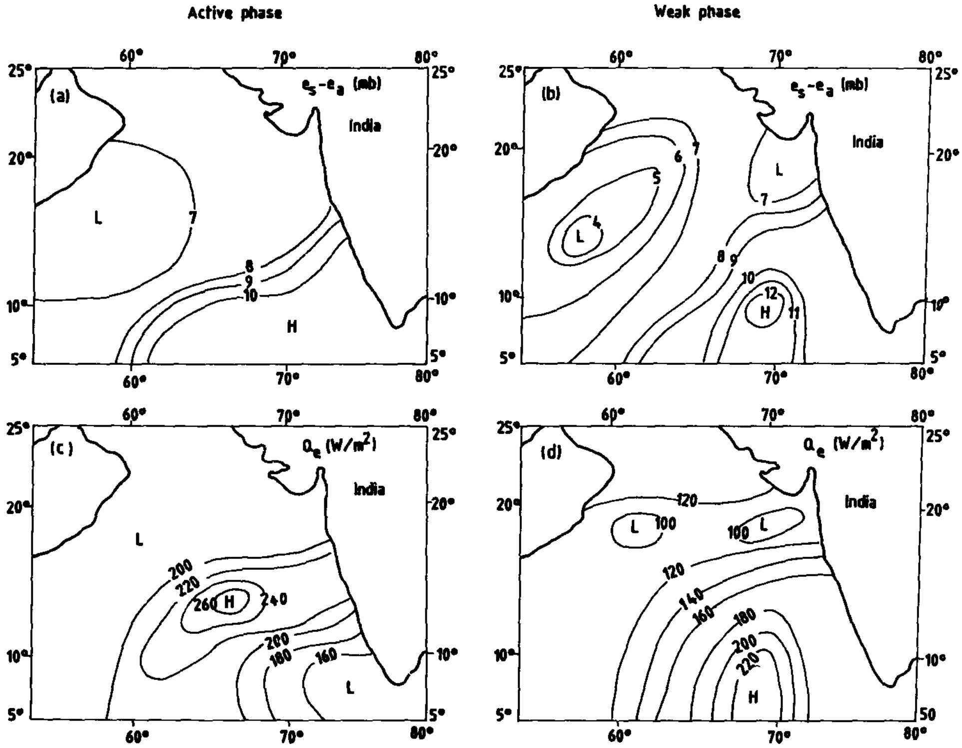

Figure 4. Same as figure 2 but for vapour pressure gradient (**a** and **b**) and latent heat flux (**c** and **d**).

pressure is important. On the other hand, during weak phase the pattern is changed due to the dominance of vapour pressure gradient, especially in the southeastern section. The zone of maximum Q_e is shifted to southeastern part with a heat flux of more than 220 Wm^{-2}. The latent heat flux is nearly doubled over the central Arabian Sea from active phase to weak phase.

3.3 *Sensible* (Q_s) *and total* ($Q_e + Q_s$) *heat flux*

The sensible heat flux is negative in the western region and positive over the southern Arabian Sea. This is an outcome of the distribution of wind speed and sea-air temperature difference. The observed minimum value (15 Wm^{-2}) is fairly close to the climatological value. Positive values of 5 to 10 Wm^{-2} are observed in the southeastern region during a weak phase (figures 5a and b).

Under an active phase the maximum value of total heat flux ($Q_e + Q_s$) is around 260 Wm^{-2} over the central Arabian Sea, while the zone of maximum is shifted to the southeastern region under a weak phase (figures 5c and d).

4. Discussion

In general, the results show that the mean fields of SST, wind speed, vapour pressure gradient and sea-air potential temperature difference are nearly uniform over the

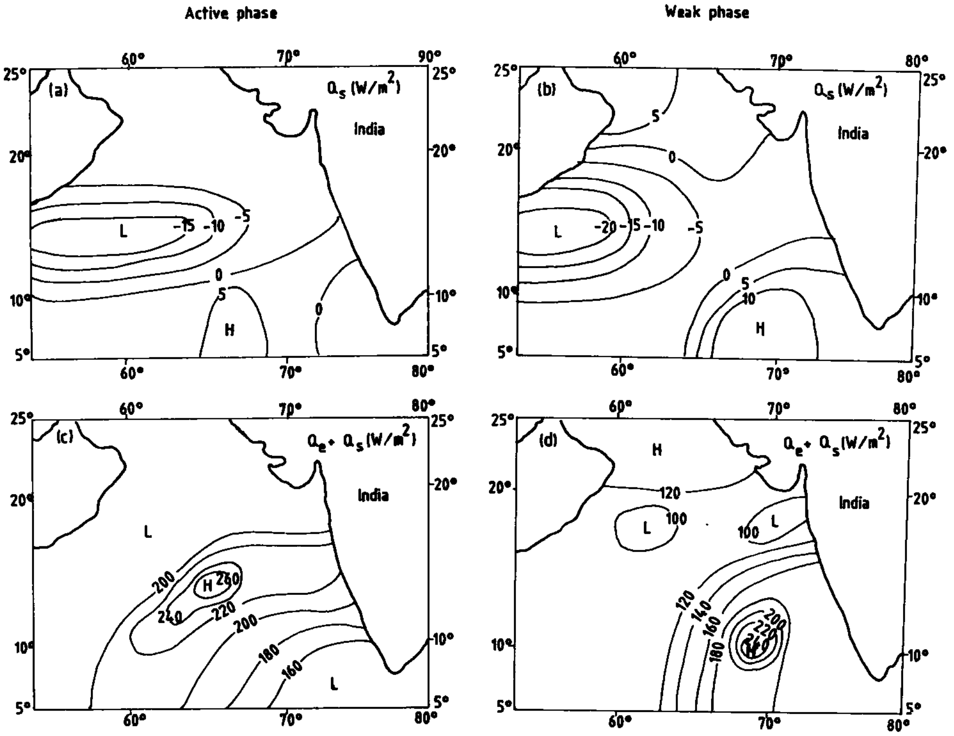

Figure 5. Same as figure 2 but for sensible heat flux (**a** and **b**) total heat flux (**c** and **d**).

western Arabian Sea during the monsoon irrespective of its intensity. The warmer SST and higher vapour pressure gradient observed in the southeastern Arabian Sea under a weak phase leads to a larger heat loss by about 33% from that during active phase. Earlier studies of Howland and Sikdar (1983) and Cadet and Greco (1987) have indicated lower rates of evaporation from an active to a weak monsoon in the central and northeastern regions of the Arabian Sea. The present analysis shows an increase in Q_e in the southeastern Arabian Sea.

Presuming higher evaporation under an active phase Ghosh *et al* (1978) and Murakami *et al* (1984) stressed the contribution of evaporation from the Arabian Sea towards the moisture transport across the west coast of India. The present data set is not sufficient to enable us to infer the total evaporation over the entire Arabian Sea. Recently, Ramesh Kumar and Sadhuram (1989) reported that evaporation from the Arabian Sea remains unaltered under a weak ($3{\cdot}66 \times 10^{10}$ tons/day)) or an active ($3{\cdot}59 \times 10^{10}$ tons/day) monsoon. Consequently the contribution from higher evaporation is less likely to add to the moisture transport across the west coast of India. Another study, on a seasonal scale, by Sadhuram and Ramesh Kumar (1988) indicates that the contribution from evaporation towards the moisture transport across the west coast of India was less significant. This leads one to infer the importance of cross equatorial flux as pointed out in earlier studies (Hastenrath and Lamb 1980; Cadet and Reverdin 1981; Howland and Sikdar 1983; Cadet and Greco 1987). However, changes in the synoptic conditions over the Arabian Sea between active and weak phases need not be ignored. For example, a ridge in the pressure field over the central

Arabian Sea blocks the moisture transport towards the west coast of India during a weak phase. Under an active phase the low level strong winds associated with the Somali jet, in the absence of a pressure ridge, increases the transport of moisture towards the west coast. The influence is seen in the distribution of observed rainfall at Mangalore, Cochin and Trivandrum where large differences prevail from active to weak phase (table 1).

5. Conclusions

It appears that the Somali jet was aligned in a zonal direction rather than a southwest-northeast orientation under both active and weak monsoons. The presence of a high pressure ridge during a weak phase seems to block the transport of moisture towards the west coast of India. These factors may be responsible for the drastic deficit in rainfall along the west coast of India during the 1986 summer monsoon.

A systematic data set with proper space and time converges is needed to obtain the mean fields more accurately. This could help to eliminate the bias due to the temporal and spatial variations of meteorological variables.

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Appendix I

A brief summary of Stevenson's method of applying correction for height is as follows. The transfer coefficients at 10 m height for heat, water vapour and momentum are approximated by $C_{H10} = 1{\cdot}0 \times 10^{-3}$, $C_{E10} = 1{\cdot}3 \times 10^{-3}$ and $C_{D10} = 1{\cdot}2 \times 10^{-3}$ respectively. Then the Richardson number is calculated using the formula

$$R_i = (gz/T_0 U_z^2)[(T_z - T_s) + 0{\cdot}61\, T_0 (C_{E10}/C_{H10})\,(q_z - q_s)] \tag{A.1}$$

where g is acceleration due to gravity, z the height of measurement above the sea surface and $T_0 = 298{\cdot}16°\mathrm{K}$. The stability of the atmosphere is given by

$$z/L = k\,(C_{H10}/C_{D10}^{3/2})\,R_i \tag{A.2}$$

where k is the von Karman constant $(=0{\cdot}40)$.

Finally, the ratio of wind speed at 10 m to the wind speed at the measurement height, the humidity difference at 10 m to the humidity difference at the measurement height and the temperature difference at 10 m to the temperature difference at the measurement height are computed using the equations

$$\frac{U_{10}}{U_z} = \frac{1 - (C_{D10}^{1/2}/k)\psi_U(10/L)}{1 + (C_{D10}^{1/2}/k)(\ln(z/10) - \psi_U(z/L))} \tag{A.3}$$

$$\frac{q_s - q_{10}}{q_s - q_z} = \frac{1 - C_{E10}/(kC_{D10}^{1/2})\psi_E(10/L)}{1 + C_{E10}/(kC_{D10}^{1/2})(\ln(z/10) - \psi_E(z/L))} \tag{A.4}$$

$$\frac{T_s - T_{10}}{T_s - T_z} = \frac{1 - C_{H10}/(kC_{D10}^{1/2})\psi_H(10/L)}{1 + C_{H10}/(kC_{D10}^{1/2})(\ln(z/10) - \psi_H(z/L))} \tag{A.5}$$

where,

$$\psi_U(x) = 2\ln[(1+x)/2] + \ln[(1+x^2)/2] - 2\tan^{-1}x + \pi/2,$$

$$\psi_E(x) = \psi_H(x) = 2\ln((1+x^2)/2)$$

where $x = (1 - 16z/L)^{1/4}$ for an unstable atmosphere (i.e. $z/L < 0$) and $\psi_U = \psi_E = \psi_H = -5z/L$ for stable atmosphere (i.e. $z/L > 0$).

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