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Estimation of drag coefficient over the western desert sector of the Indian summer monsoon trough

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Abstract. In the estimation of momentum fluxes over land surfaces by the bulk aerodynamic method, no unique value of the drag coefficient (C_D) is found in the literature. The drag coefficient is generally estimated from special observations at different parts of the world. In this study an attempt is made to estimate drag coefficient over the western desert sector of India using data sets of Monsoon Trough Boundary Layer Experiment (MONTBLEX) during the summer monsoon season of 1990. For this purpose, the fast and slow response data sets obtained simultaneously from a 30 m high micro-meteorological tower at Jodhpur are used. All the observations used in this study are confined to a wind speed regime of 2·5–9·0 ms^{-1}.

A comparison of momentum fluxes computed by eddy correlation (direct estimation) with profile and bulk aerodynamic ($C_D = 3{\cdot}9 \times 10^{-3}$, Garratt, 1977) methods revealed that though the nature of variation of the fluxes by all these methods is almost similar, both the indirect methods give an under-estimated value of the fluxes. The drag coefficient is estimated as a function of wind speed and surface stability by a multiple regression approach. An average value of the estimated drag coefficient is found to be of the order of $5{\cdot}43 \times 10^{-3}$. The estimated value of C_D is validated with a set of independent observations and found to be quite satisfactory. The recomputed momentum fluxes by bulk aerodynamic method using the estimated drag coefficient are in close agreement with the directly estimated fluxes.

Keywords. Drag coefficient; momentum fluxes; bulk aerodynamic method; eddy correlation method.

1. Introduction

The thermal and dynamic interaction between the atmosphere and the underlying surface (sea or land) occurs through turbulent exchange of momentum, heat and moisture at their interface. The variation of fluxes of momentum, heat and moisture in the surface boundary layer and their distribution in the rest of the planetary boundary layer play a vital role in the energy transport mechanism of the land-atmosphere-ocean system and influence the steady state of the atmosphere.

The surface fluxes are determined by several methods such as eddy correlation, profile, bulk aerodynamic, energy dissipation etc. The bulk aerodynamic method is the most conventional approach and is widely used. In this method, the turbulent transports of momentum, heat and moisture are treated as proportional to the differences of mean meteorological parameters such as wind speed, temperature and moisture between the underlying surface and a reference level. This method is formulated on the basic assumption that the surface wind stress is parallel to the surface wind direction. However, the proportionality constants known as bulk transfer coefficients remain a problem of concern as no unique value is found in the literature.

Most of the studies on the estimation of these transfer coefficients pertain either to open sea, sea-shore, sand-pit or sea-beach interface (Smith and Banke 1975; Sethuraman and Raynor 1975; Francey and Garratt 1978; Vugts and Cannemeijer 1981; Large and Pond 1981 and Tsukamoto *et al* 1990). Only a few works over land regions have been reported in literature.

Garratt (1977) has reported a comprehensive review on the estimation of drag coefficients over ocean and land surfaces. Its dependence on wind speed, stability and nature of the underlying surface has been documented. Beljaars and Holtslag (1991) have briefly described observations and modelling efforts on surface fluxes over land surfaces. However, information on the dependence of drag coefficient over land surfaces on wind speed and stability is still inadequate in many of these reviews.

The most striking feature of all these studies is that there is no unique approach in the estimation of drag coefficient. Garratt (1977) reported average values of neutral drag coefficient over tropical and extra-tropical land surfaces on the basis of the conservation principle of angular momentum. The quoted value of neutral drag coefficient over Asian land mass (20°–50°N) is of the order of $3{\cdot}9 \times 10^{-3}$.

Over the Asian land mass, the summer monsoon exhibits wide variability from region to region. One of the large scale features of the Indian summer (southwest) monsoon is the persistence of surface low pressure extending from the seasonal desert heat low over Pakistan and north west India through the Indo-Gangetic plains and dipping into the Bay of Bengal on the east of the Indian sub-continent. This surface feature, known as the *monsoon trough*, is quasi-permanent during the Indian summer monsoon. The rainfall activity over the Indian sub-continent is known to be largely associated with the positioning of the monsoon trough (Rao 1976). The monsoon trough is understood to influence the planetary boundary layer over India and other regions of tropics.

Atmospheric boundary layer experiments began during the late sixties. In India for the first time, during the Monsoon Experiment (MONEX-79), a 10 m high mast was installed at a coastal station close to the Bay of Bengal to probe the boundary layer. In the summer monsoon season (June – September) of 1990, the Monsoon Trough Boundary Layer Experiment (MONTBLEX) was designed to carry out exclusive surface boundary layer observations over land surfaces along the monsoon trough (Goel and Srivastava 1990). During this experiment, micro-meteorological towers of 30 m height with slow and fast response sensors fixed at six nearly logarithmic levels were installed at four locations along the monsoon trough. These locations represent the dry-convective to moist-convective nature of the atmosphere along the normal axis of the monsoon trough.

In the present study, an attempt is made to estimate the drag coefficient over land surface on the western desert sector of India and to estimate the surface momentum fluxes by direct and indirect methods using the tower observations of MONTBLEX-90.

2. Methodology

2.1 *Determination of surface fluxes*

The surface fluxes of momentum, heat and moisture are determined widely by three methods, viz., eddy correlation, profile and bulk aerodynamic methods. Each of these methods is briefly described below:

2.1.1 *Eddy correlation method*: It is a direct method for determining the surface (momentum, heat and moisture) fluxes in terms of covariances of vertical velocity component with horizontal velocity components, temperature and specific humidity. For this purpose, fast response observations (of frequency 8–10 Hz) are generally used. The fast response sensors placed on stable platforms need a proper levelling and calibration for a better accuracy and reliability.

The surface momentum flux, in particular, is expressed as:

$$\tau = \rho(\overline{u'w'^2} + \overline{v'w'^2})^{1/2}, \tag{1}$$

where ρ is the density of the surface layer.

2.1.2 *Profile method*: It is an indirect method based mainly on Monin-Obukhov's similarity theory for the determination of surface fluxes. For the construction of wind, temperature and moisture profiles, their measures at two levels in the surface layer are required.

The wind and temperature profiles in the surface layer are described in the following forms (Dyer and Hicks 1970; Businger *et al* 1971):

$$\Delta\bar{u} = \frac{u_*}{k}\left(\ln\frac{z_2}{z_1} - \psi_m(\zeta_2) + \psi_m(\zeta_1)\right), \tag{2}$$

$$\Delta\bar{\theta} = R\frac{\theta_*}{k}\left(\ln\frac{z_2}{z_1} - \psi_h(\zeta_2) + \psi_h(\zeta_1)\right), \tag{3}$$

where $\Delta\bar{u} = u_2 - u_1$ and $\Delta\bar{\theta} = \theta_2 - \theta_1$, the subscripts denoting the respective levels. u_* is the friction velocity and θ_* is the temperature scale. ψ_m and ψ_h are the stability functions associated with wind and temperature profiles respectively. $\zeta(= z/L)$ is the stability function where L is the Monin-Obukhov length given by:

$$L = \frac{\bar{T}}{g}\frac{u_*^2}{k\theta_*}. \tag{4}$$

$R = 0{\cdot}74$, a ratio of eddy diffusivities in the neutral limit. $k = 0{\cdot}4$, the von Karman constant.

The stability functions assume the following form (Paulsen 1970; Barker and Baxter 1975):

In unstable conditions ($\zeta < 0$):

$$\psi_m(\zeta) = \ln\left(\left(\frac{1+x}{2}\right)^2\frac{(1+x^2)}{2}\right) - 2\arctan x + \frac{\pi}{2}, \tag{5}$$

and

$$\psi_h(\zeta) = 2\ln\left(\frac{1+y}{2}\right), \tag{6}$$

where

$$x = (1 - 15\zeta)^{1/4} \text{ and } y = (1 - 9\zeta)^{1/2}, \tag{7}$$

while in stable conditions ($\zeta > 0$):

$$\psi_m(\zeta) = -4{\cdot}7\,\zeta, \tag{8}$$

and

$$\psi_h(\zeta) = \frac{-4{\cdot}7}{R}\zeta. \tag{9}$$

The friction velocity (u_*) and the temperature scale (θ_*) are computed iteratively through the following steps (Berkowicz and Prahm 1982):

Step 1: By assuming a large value of L, u_* and θ_* are computed first in neutral limits.
Step 2: L is recomputed from equation (4) using the neutral estimates of u_* and θ_*
Step 3: u_* and θ_* are recomputed using equations (2) and (3) and L as recomputed in Step 2.
Step 4: Steps 2 and 3 are repeated until the value of L does not change in desired accuracy limits.

For the sake of brevity, the moisture scale q_* is not discussed here.
The surface momentum flux, in particular, is expressed as:

$$\tau = \rho u_*^2. \tag{10}$$

2.1.3 *Bulk aerodynamic method*: This is the most conventional and widely used method for determining the surface fluxes. In this method, the basic assumption is that the surface wind stress is in the direction of surface wind. The surface fluxes are given in terms of the differences in mean meteorological parameters at two lowermost levels in the surface layer.
The surface momentum flux, in particular, is expressed as:

$$\tau = \rho C_D V^2, \tag{11}$$

where C_D is the momentum transfer or drag coefficient and V is the wind speed at a reference height. Over land surfaces, there is no unique value of drag coefficient reported in the literature. It is found to depend on wind, stability and nature of the underlying surface.

2.2 *Determination of surface layer parameters*

The surface layer parameters, viz., friction velocity and temperature scale have been defined in equations (1) and (2) in terms of profiles of wind and temperature at two levels in the surface layer. These parameters can be directly determined in terms of covariances of vertical velocity component with the horizontal velocity components and temperature through the following relations:

$$u_* = (\overline{u'w'^2} + \overline{v'w'^2})^{1/4}, \tag{12}$$

$$\theta_* = -\frac{\overline{w'\theta'}}{u_*}. \tag{13}$$

The drag coefficient (C_D) can be directly determined from the relation:

$$C_D = \left(\frac{u_*}{V}\right)^2, \tag{14}$$

where u_* is as defined in equation (12).

3. Data used and computational procedure

During MONTBLEX-90, surface boundary layer observations were measured from fast and slow response sensors on 30 m high micro-meteorological towers installed at four locations along the normal position of the monsoon trough. These locations are: Jodhpur on the West, Delhi and Varanasi in the centre, and Kharagpur on the East. Jodhpur (26·3°N, 73°E) represents a dry-convective and semi-arid region in the western desert sector of the monsoon trough. The tower observations at Jodhpur are reported to be continuous and extensive (Rudrakumar *et al* 1991).

The tower observations at Jodhpur consist of slow response data at all six levels (1, 2, 4, 8, 15 and 30 m) and fast response data at two levels (4 and 15 m). The slow data were recorded on Campbell data logger with a sampling frequency of 1 Hz and the fast data were recorded on a computer based telemetry system with a sampling frequency of 8 Hz (Rudrakumar *et al* 1991). Data acquisition from slow response sensors was continuous while from fast response sensors, it was 10–15 min. at each observation time. The interval of observations was 3 hours during the normal period of the experiment and hourly during intensive observation periods (IOP). The IOP correspond to specific synoptic situations.

In the present study on the estimation of drag coefficient and surface momentum fluxes, only simultaneously obtained slow (at 1 and 15 m) and fast (at 4 m) response data sets during the entire MONTBLEX period are considered. From 112 of such observations available with us, about 17% of cases with spikes in the data are rejected as a first step. A further quality check on the data sets is carried out through a comparison of surface momentum fluxes estimated by eddy correlation and profile methods. In order to have a homogeneous data base, 10 cases with a difference in momentum fluxes by direct and indirect estimates exceeding by 50% and 14 cases exceeding by 100% are discarded. Finally, 70 sets of homogeneous and good quality data are used to carry out further study.

4. Results and discussion

4.1 *Characteristics of surface layer*

The surface layer parameters, viz., the friction velocity (u_*) and the temperature scale (θ_*) are computed directly from equations (12) and (13) using the fast response data at 4 m from the selected 70 sets of observations. The distributions of these parameters are depicted in figures 1 and 2 respectively. The friction velocity varies in the range of 0·2–0·4 m s^{-1}, with extreme values corresponding to relatively high surface winds (of the order of 7–9 m s^{-1}) of an active spell (2nd–12th July). The temperature scale is in the

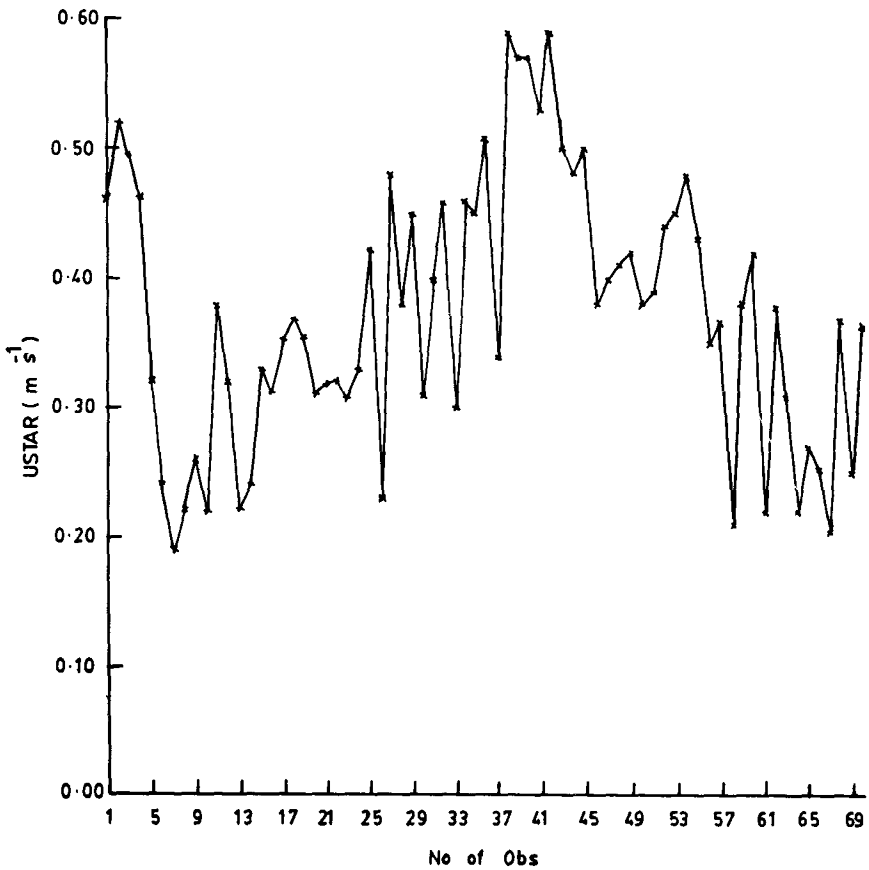

Figure 1. Friction velocity as determined from fast response data at 4 m.

range of $-0{\cdot}2$ and $-0{\cdot}6$, indicating that the surface layer remained unstable over a majority of observation hours.

The standard deviations σ_u, σ_v and σ_w of the three velocity components u, v and w are computed and normalized by the friction velocity (figures not presented). The normalized quantities σ_u/u_* and σ_v/u_* are in the range of 1·8–2·5 while σ_w/u_* in 1·1–1·3 which are nearly in agreement with the general values given by Panofsky and Dutton (1984). They are also comparable with the values obtained by Jiemin *et al* (1990) in their study of surface layer over Gobi desert (39·2°N, 100°E). The fast response observations used in this study are thus of reasonably good quality.

4.2 *Computation of surface momentum fluxes*

The surface momentum fluxes are first directly estimated by the eddy correlation method and indirectly by the profile method using the 70 sets of observations. As at present a neutral value of drag coefficient over Asian land mass as prescribed by Garratt (1977) is available, the surface momentum fluxes are estimated by bulk aerodynamic method using $C_{DN} = 3{\cdot}9 \times 10^{-3}$. The distributions of the fluxes computed

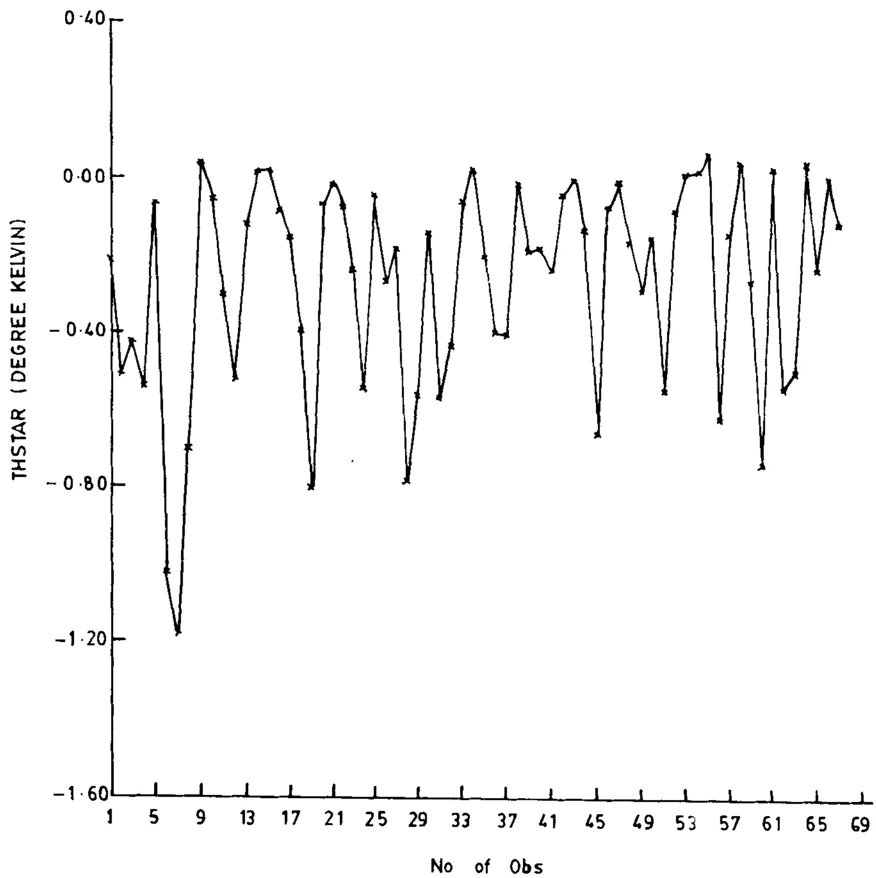

Figure 2. Temperature scale as determined from fast response data at 4 m.

by direct and indirect methods are shown in figure 3. For convenience, the fluxes estimated by the three methods are hereinafter referred to as eddy-fluxes, profile-fluxes and bulk-fluxes and are denoted by MF–E, MF–P and MF–B respectively.

The most striking feature here is that although there is similarity in the nature of variation of these fluxes irrespective of their method of evaluation, the fluxes determined indirectly are under-estimated.

The eddy-fluxes exhibited extreme values during a sequence of observation hours corresponding to an active spell of monsoon rain (2–12 July). The winds in the surface layer during this period are also relatively stronger (7–9 $m\,s^{-1}$) as compared to the winds in any normal period of observation. Such peaks in momentum fluxes are not captured by the profile method and the distribution of profile-fluxes is almost smooth. The standard deviation of profile-fluxes is found to be of lower order (0·05) as compared to that of eddy-fluxes (0·09). As can be seen from figure 3, the bulk-fluxes whose estimation is based on a neutral drag coefficient also exhibited an increasing nature with increasing winds. It is therefore necessary to determine a more realistic drag coefficient from observations and further evaluate it with any statistically significant approach.

4.3 Direct determination of drag coefficient

The under-estimation of bulk-fluxes as compared to eddy-fluxes as seen in § 4.2 may be attributed to the use of a neutral drag coefficient. An alternative approach is to estimate drag coefficient directly, such as by prescribing the observed friction velocity at 4 m level and wind speed at 15 m in equation (14). For convenience, such a directly determined drag coefficient may be referred to as C_D(dir). The estimation gives an average value of the order of $5{\cdot}43 \times 10^{-3}$ within a wind speed range of 2–9 m s^{-1} which is about 35% in excess over Garratt's neutral value of drag coefficient. It is also 3–4 times larger than the normally reported value over ocean surfaces.

In this study, the differences in (potential) temperature between 1 and 15 m levels in the surface layer are in the range of 0·11–3·0°K. In the surface layer, the (potential) temperature difference ($\Delta\theta$) between two lowermost levels is sufficient to determine the stability of the surface layer. In this study $\Delta\theta$ is treated as a stability parameter.

Correlation coefficients of C_D(dir) obtained with wind speed and stability parameter are found in the order of − 0·39 and 0·12 respectively. Thus a rapid decreasing trend in

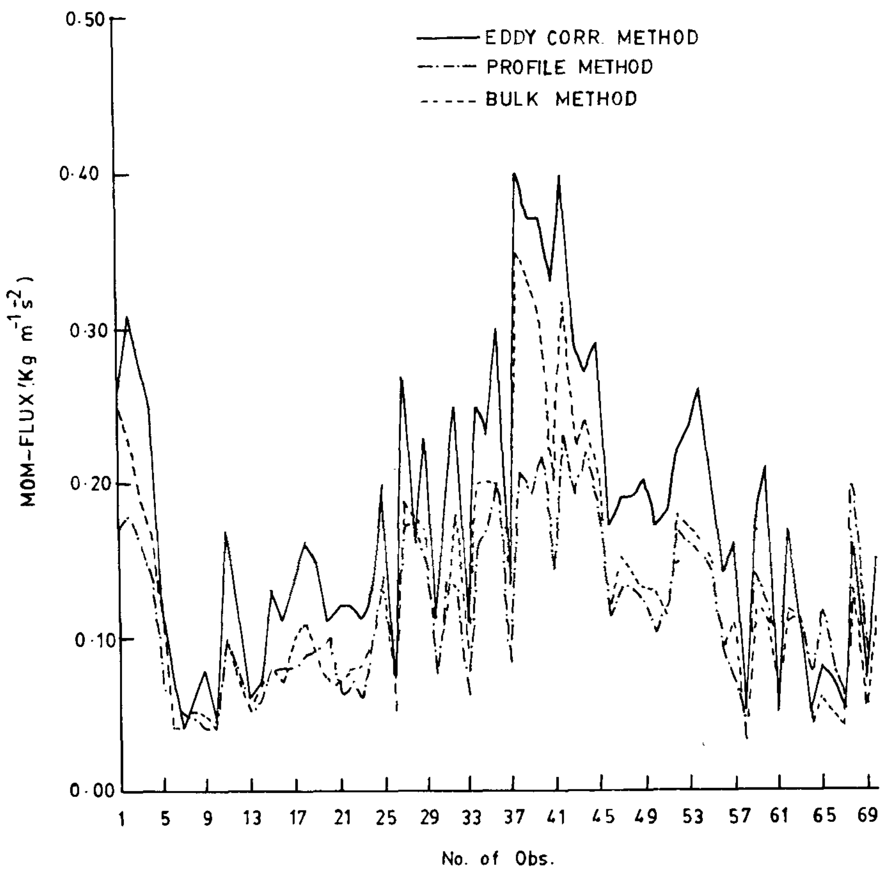

Figure 3. Momentum fluxes computed by eddy correlation, profile and bulk aerodynamic (with C_{DN}, Garratt) methods.

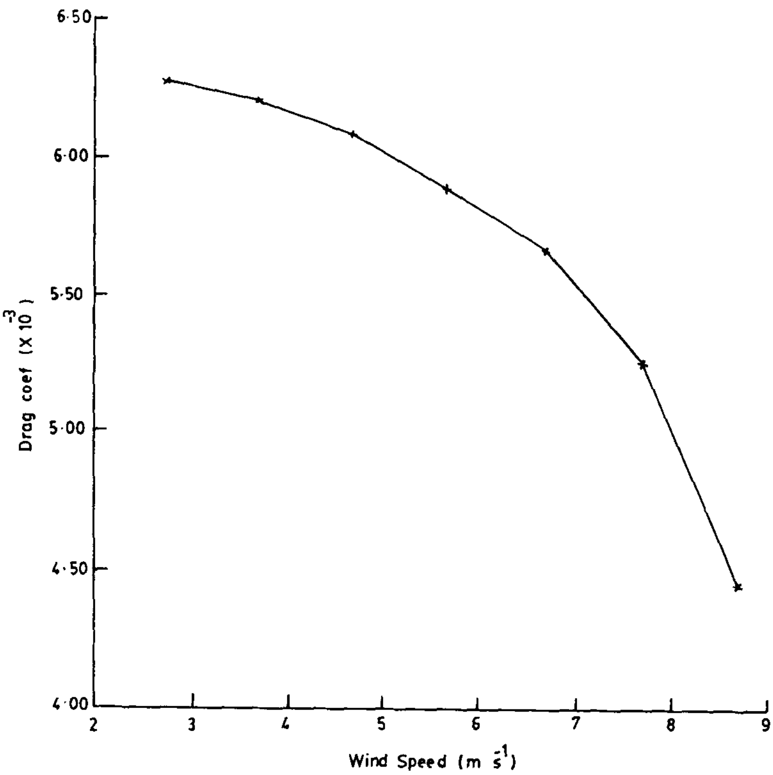

Figure 4. Variation of drag coefficient (direct) with wind speed.

drag coefficient with increasing wind speed, and a slowly increasing trend with increasing instability is evident within the observed ranges of wind speed and (potential) temperature differences. The variation of C_D(dir) with wind speed is shown in figure 4.

The foregoing analysis thus suggests a reexamination of drag coefficient in a more statistically significant approach keeping in view its dependence on wind speed and surface stability.

4.4 *Estimation of drag coefficient from wind speed and stability in the surface layer*

Over ocean surfaces, the estimation of drag coefficient is normally made considering it either as a linearly varying function or as power function (the power is normally confined to an order of 2) of wind speed at a reference height above the sea surface. Kondo (1975) instead considers a stability parameter in the more generalized form of Richardson number.

In the present study, the estimation of drag coefficient is proposed in the following manner:

- Representing drag coefficient as a function of wind speed alone such that

$$10^3 C_D(V) = a_0 + a_1 V + a_2 V^2. \tag{15}$$

■ Representing drag coefficient as a second degree polynomial in both wind speed and stability parameter such that

$$10^3 C_D(V, \Delta\theta) = b_0 + b_1 V + b_2 V^2 + b_3 \Delta\theta + b_4 \Delta\theta^2 + b_5 V\Delta\theta, \qquad (16)$$

where V is the wind speed at 15 m and $\Delta\theta$ is the difference in potential temperature at 1 and 15 m levels in the surface layer. $a_i (i = 0$ to $2)$ and $b_j (j = 0$ to $5)$ are the regression coefficients to be determined.

The regression coefficients of either form can be obtained by solving equation (15) or (16) by the method of least squares using the wind and temperature data and the directly determined drag coefficients. For the estimation purpose, the first 50 sets of observations out of a total number of 70 are used. The 50 sets of observations are treated as dependent. The remaining 20 sets of observations are treated as independent for validating the estimation.

The regression coefficients a_i and b_j are presented in table 1. The regression coefficients a_0 to a_2 and b_0 to b_2 are nearly of the same order showing that the estimation of drag coefficient is largely influenced by wind speed. The standard error of the estimation of drag coefficient as a function of wind speed alone i.e. $C_D(V)$ is found to be 0·80 while that of $C_D(V, \Delta\theta)$ is 0·72. Equation (16) thus represents a more general variation of drag coefficient applicable under all wind and stability conditions.

The estimated drag coefficient is hereinafter referred to as drag coefficient (regression) and denoted by C_D(reg). The variation of C_D(reg) is shown in figure 5. On the average, C_D(reg) is of the order of $5{\cdot}38 \times 10^{-3}$.

Correlation analysis revealed a correlation coefficient of C_D(reg) with wind speed (V) of the order of − 0·76 and with stability parameter ($\Delta\theta$) about 0·24, indicating a general decreasing trend of drag coefficient with increasing winds and a slowly increasing trend with increasing instability, as were also obtained with C_D(dir) (cf. § 4.3). The trends of C_D(reg) are also in agreement with the analysis of Hsu (1974), Liu *et al* (1979) and Large and Pond (1981) over ocean surfaces.

Interestingly, Kondo (1975) observes two opposite natures of the variation of drag coefficient on sea surface, (i) decreasing under light wind regime ($< 2\,\mathrm{m\,s^{-1}}$) and (ii) increasing under high wind regime ($> 2\,\mathrm{m\,s^{-1}}$) in association with the sea surface roughness. In the present study, as mentioned above, light winds ($< 2{\cdot}5\,\mathrm{m\,s^{-1}}$) are not found with the data sets and hence the nature of the variation of the drag coefficient under such light winds is not examined.

As a further step, the momentum fluxes are computed by the bulk aerodynamic method using the drag coefficient (regression) corresponding to the 50 dependent sets of observations. A comparative distribution of the bulk-fluxes with the eddy-fluxes is shown in figure 6. It can be noticed that the bulk-fluxes are quantitatively in good agreement with the eddy-fluxes as compared to the bulk-fluxes depicted in figure 3.

Table 1. Regression coefficients in the estimation of drag coefficient.

C_D(reg)	a_0/b_0	a_1/b_1	a_2/b_2	b_3	b_4	b_5
$C_D(V)$	5·55345	0·15217	−0·03148	—	—	—
$C_D(V, \Delta\theta)$	5·27371	0·09504	−0·01663	0·68692	0·06837	−0·11917

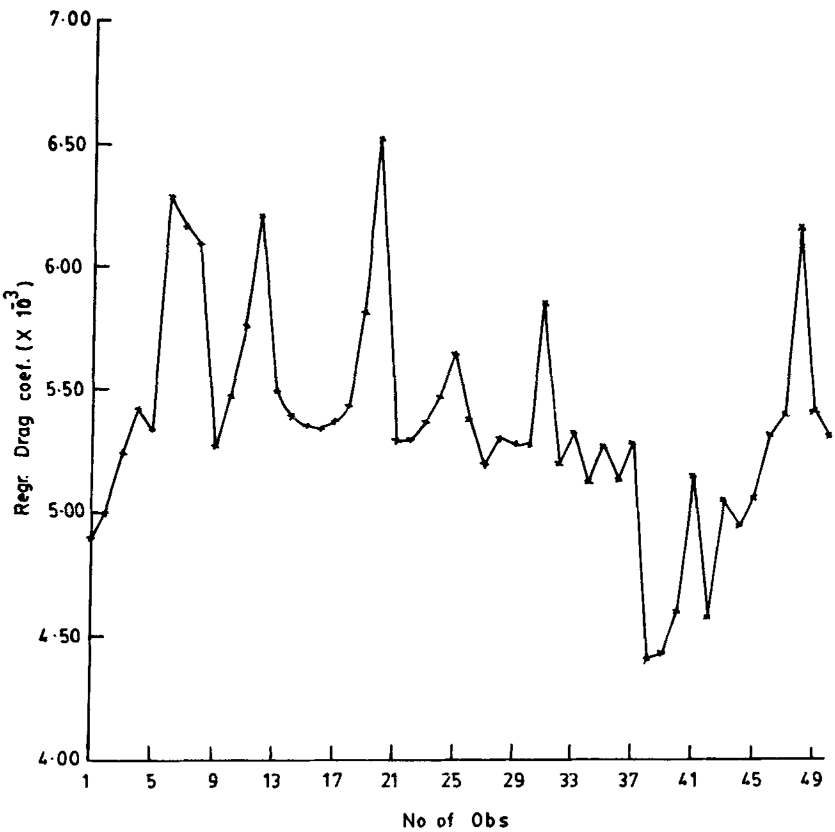

Figure 5. Variation of drag coefficient (regression).

The statistical measures of parameters (C_D(reg), MF–E, MF–P, MF–B) associated with the 50 dependent sets of observations are shown in table 2. The standard deviation of the profile-fluxes is relatively on the lower side (0·05) as compared to that of the eddy-fluxes (0·09).

The lower diagonal matrix of correlation coefficients associated with the dependent sets of observations is presented in table 3. The drag coefficient shows its decreasing tendency with increasing wind and increasing tendency with increasing instability. The bulk fluxes are nearly perfectly correlated with the eddy-fluxes (correlation coefficient is of the order of 0·97).

4.6 *Performance with independent data sets*

Using wind and temperature data corresponding to 20 independent sets of observations in equation (16) in which the regression coefficients (b_j) are as given in table 1, the drag coefficient at every independent hour of observation is determined. For convenience, the drag coefficient so determined is represented by C_D(ind). The mean value of C_D(ind) over 20 sets of observations is $5{\cdot}57 \times 10^{-3}$ which is of comparable magnitude with the mean value of C_D(reg) over 50 (dependent) sets of observations ($5{\cdot}38 \times 10^{-3}$).

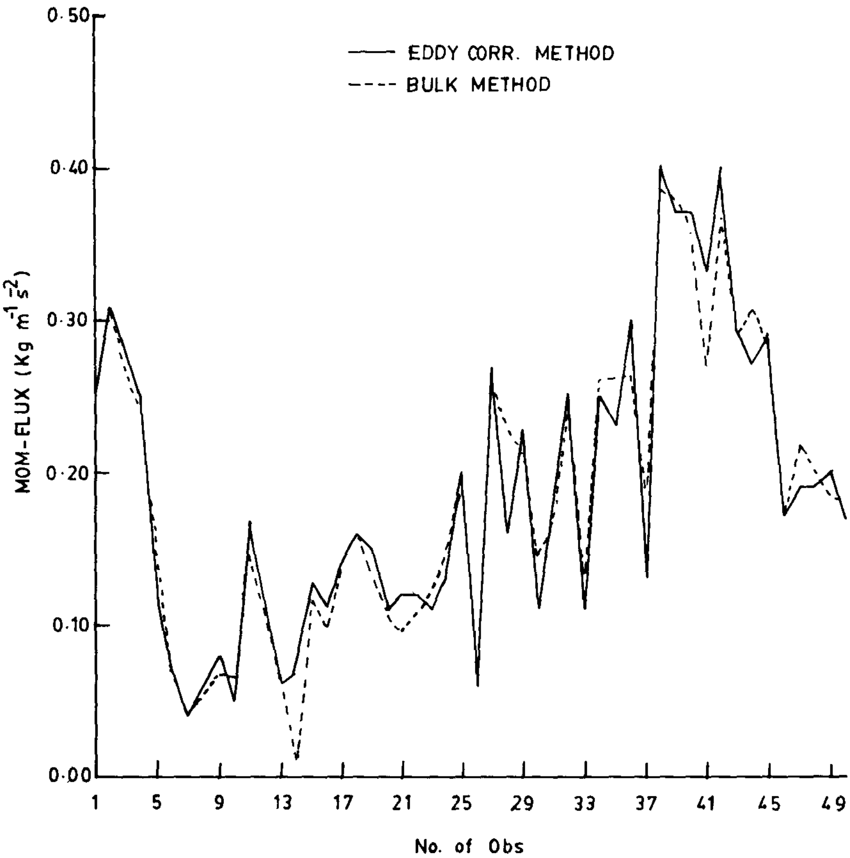

Figure 6. Momentum fluxes computed by eddy correlation and bulk aerodynamic (with C_D(reg)) methods (50 dependent cases).

Table 2. Statistical measures of parameters associated with 50 dependent sets of observations.

Parameter	Mean	Std. Dev.	RMS error
$10^3 C_D$(reg)	5·380	0·430	0·723
MF–E	0·186	0·096	—
MF–P	0·120	0·055	0·083
MF–B	0·186	0·093	0·024

Table 3. Lower diagonal matrix of correlation coefficients associated with 50 dependent sets of observations.

	C_D(reg)	V	$\Delta\theta$	MF–E	MF–P	MF–B
C_D(reg)	1·0					
V	−0·76	1·0				
$\Delta\theta$	0·24	0·34	1·0			
MF–E	−0·70	0·96	1·0			
MF–P	−0·59	0·93	0·48	0·93	1·0	
MF–B	−0·72	0·99	0·41	0·97	0·95	1·0

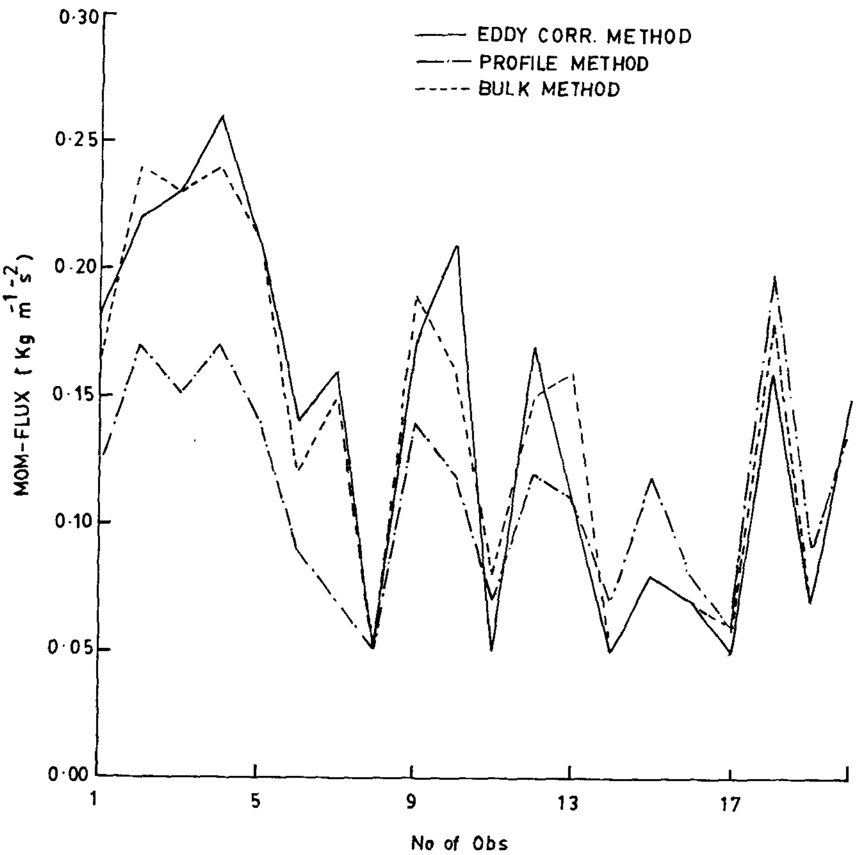

Figure 7. Momentum fluxes computed by eddy correlation, profile and bulk aerodynamic (with C_D(ind)) methods (20 independent cases).

Thus the regression coefficients (b_j) and equation (16) may be used for any arbitrary data sets.

The bulk-fluxes (using C_D(ind)), the profile-fluxes and the eddy-fluxes are presented in figure 7. The bulk-fluxes are again found to be in good agreement with the eddy-fluxes and the profile-fluxes are again found to be under-estimated in most of the cases.

Table 4. Statistical measures of parameters associated with 20 independent sets of observations.

Parameter	Mean	Std. Dev.	RMS error
$10^3 C_D$(ind)	5·572	0·290	0·842
MF–E	0·140	0·066	—
MF–P	0·107	0·040	0·054
MF–B	0·140	0·062	0·021

Table 5. Lower diagonal matrix of correlation coefficients associated with 20 independent sets of observations.

	C_D(ind)	V	$\Delta\theta$	MF–E	MF–P	MF–B
C_D(ind)	1·0					
V	−0·25	1·0				
$\Delta\theta$	0·67	0·49	1·0			
MF–E	−0·19	0·94	0·52	1·0		
MF–P	−0·06	0·84	0·59	0·79	1·0	
MF–B	−0·15	0·99	0·60	0·95	0·85	1·0

The statistical measures of parameters (C_D(reg), MF–E, MF–P, MF–B) associated with the independent sets of observations are shown in table 4. The standard deviation of the profile-fluxes is on the lower side (0·04) as compared to that of the eddy- and bulk-fluxes (0·07 and 0·06). A low value of standard error of the bulk-fluxes (0·02) indicates a relatively better performance of the estimated drag coefficient. The lower diagonal matrix of correlation coefficients is shown in table 5. Although stability dominance in the estimation of the drag coefficient is found with the independent sets of observations, the respective decreasing and increasing trends with increasing wind and increasing instability are still persistent.

5. Conclusions

An estimation study on the drag coefficient over the western desert sector of the monsoon trough using simultaneous fast and slow response observations of 30 m high micro-meteorological tower during MONTBLEX-90 brought out the following features:

- The drag coefficient exhibits a decreasing tendency with increasing winds in a wind regime of 2·5–9 $\mathrm{m\,s^{-1}}$.
- The drag coefficient increases in general with increasing instability of the surface layer. However, the tendency with light winds ($< 2{\cdot}5\,\mathrm{m\,s^{-1}}$), as they were not available with the present data sets, is not known.
- An average value of the estimated diabatic drag coefficient is found to be of the order of $5{\cdot}43 \times 10^{-3}$ which is about 3–4 times larger than the reported average value over ocean surfaces but does not exceed by more than 40% the neutral value ($3{\cdot}9 \times 10^{-3}$) prescribed by Garratt (1977) for the Asian region.
- In general, the momentum fluxes computed by the profile method are quantitatively under-estimated as compared to the fluxes directly determined by the eddy correlation method. The fluxes computed by the bulk aerodynamic method using the estimated drag coefficient are in good agreement with the eddy-fluxes.

The conclusions regarding the estimation of drag coefficient are confined to a land region on the western desert sector of the monsoon trough. Similar studies at other major locations along the monsoon trough using MONTBLEX-90 data are desirable to establish the spatial and temporal variation of drag coefficient during the summer monsoon.

The estimation of drag coefficient over tropical land surfaces along the monsoon trough is useful in the parameterization of surface layer in large scale numerical models.

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