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Estimation of fluxes from wind and temperature profiles in the marine atmospheric surface boundary layer

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सार — 7° 40′ उ., 78° 26′ पू. पर स्थित एक स्थान पर पवन पार्श्विका (प्रोफाइल) पर तथा 12° 35′ उ., 80° 48′ पू. पर स्थित स्थान पर पवन तथा तापमान पार्श्विकाओं पर दिन के समय घंटेवार माप लिए गए हैं। ये माप समुद्र की सतह के ऊपर 3-6 मी. परत में लिए गए। इनके लिए एक 3.2 मी. के ऐसे मस्तूल पर छह स्तरों पर लगे संवेदकों का प्रयोग किया गया जो अनुसंधान जलयान "आर. वी. गवेषणी" के अग्रभाग के आगे की ओर बाहर निकला हुआ था। यह कार्य इस जलयान की मई 1983 में अपनी समुद्री यात्रा के दौरान उस समय किया गया था जब वह क्रमशः 16 या 20 मई को इन स्थानों पर प्रेक्षण लेने के लिए रुका था।

16 मई को पवन प्रोफाइल से कर्ष गुणांक (ड्रैग कोएफिसेएंट) का आकलन किया गया। डायबेटिक प्रभावों के लिए 20 मई में वायु पार्श्विका में संशोधन किया गया। इसके लिए समरूप तापमान पार्श्विका का प्रयोग किया गया है और आंकलित कर्ष गुणांक, संवेग (मोमेंटम) और संवेदी ऊष्मा अभिवाह (फलक्स) को प्रस्तुत किया गया है।

ABSTRACT. Daytime hourly measurements on wind profile at a station 7°40′N, 78°26′ E, and wind and temperature profiles at 12°35′ N, 80°48′E, were made in the 3-6 m layer above sea surface, using sensors fitted at six levels on a 3.2 m mast which was projected forward of the bow of the research vessel "*R.V. Gaveshani*" during her cruise in May 1983 when she was halted at these stations, on 16 and 20 May respectively, for observations.

Drag coefficient was estimated from wind profiles on 16th. Wind profiles on 20th were corrected for diabatic effects using the corresponding temperature profiles and the estimated drag coefficient, momentum and sensible heat flux are presented.

1. Introduction

The turbulent fluxes of momentum, heat and moisture play a key role in the energy transport mechanism of the ocean atmosphere system. An experiment to measure the wind and wind temperature profiles over two sea stations, using a stabilized research vessel '*R.V. Gaveshani*' as measurement platform, is described here. The objective is to estimate and study the stability of marine surface layer and the time variation of drag coefficient, fluxes of momentum and sensible heat from the profiles.

2. Experimental

Micrometeorological measurements were carried out on board the Research vessel, '*R.V. Gaveshani*' of the National Institute of Oceanography, Goa, India during her cruise in May 1983. The ship was halted during the day on 16 and 20 May 1983 at the stations 7°40′N, 78°26′E and 12°35′N, 80° 48′E respectively—Fig. 1, where the environment was an open sea regime. Wind and temperature profiles in the 3-6 m layer above sea surface were measured using a 3.2 m mast fitted with sensors at 6 levels. The mast was fixed to an observational boom extending 2.4 m forward of the bow — Fig. 2. Thornthwaite three cup anemometers and linearised YSI (Yellow Spring Instrurments Company, Ohio) thermistors were used to measure the wind speed and temperature respectively. The sensors were located at 3.1, 3.3, 3.7, 4.5, 5.3 and 6.1 m above sea surface. Thermistor sensors were guarded against sea water spray and direct heating by sunlight by means of a cylindrical wire gauze shield and radiation shield respectively. Anemometers and thermistors were calibrated in the laboratory, prior to the experiment. The sensor mast was oriented to face the wind throughout and the observations were taken at hourly intervals for a sampling period of 10 minutes. Wind speeds recorded by the Thornthwaite counter were averaged over 10 minutes whereas in the case of temperature, absolute value at 6.1 m level and the difference with respect to this at other levels, were noted from the output of a digital voltmeter. The accuracy of wind and temperature sensors is ± 0.05 m/s and $\pm 0.15°$ C respectively. Data on barometric pressure, wet and dry bulb temperature at 5 m level was recorded on board the ship by the National Institute of Oceanography. The ship is stabilized against rolling.

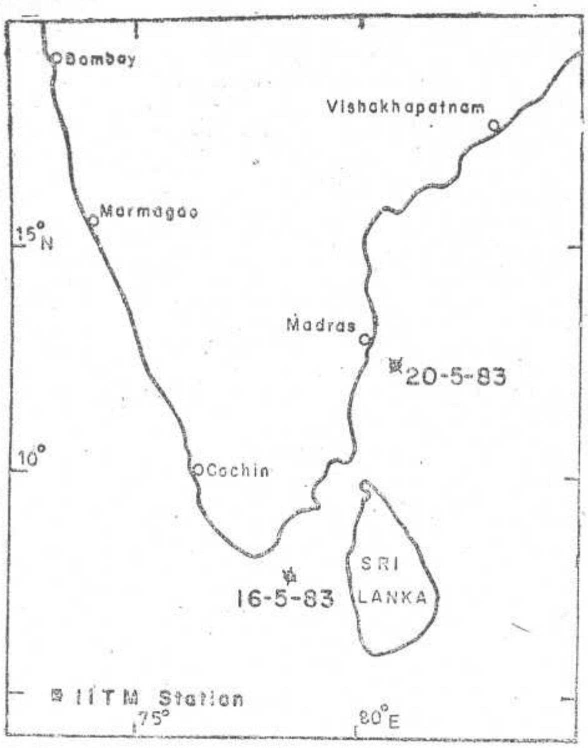

Fig. 1. Position of station and dates of observation from the ship

3. Method of analysis

The method described below considers the sources of errors that might possibly affect the profile measurement accuracy. The stability of the surface layer over ocean estimated from gradient Richardson number has been used to compute the Monin-Obukhov length 'L'. The profiles are corrected for stability effects and the fluxes estimated.

3.1. *Effect due to waves and ship*

The sea waves and the ship's superstructure influence the quality of data obtained. During our observations the wave heights observed visually were not high ($\leqslant 1$ m). The wave influence on the profiles is seemingly restricted to heights below about three wave heights (Krugermeyer *et al.* 1978). Therefore, the wave influence on measured profiles in the present study at 3.3-6.1 m level above sea surface has been assumed to have no effect. Further, for the sensor mast projected by a boom directly in to the wind from the bow of the ship, the effect of the ship's superstructure on airflow has been assumed negligible under a moderate sea state (Mitsuta and Fujitani 1974), Mollo-Christensen (1968, 1979), Jason and Ching (1976).

3.2. *Profile measurement accuracy*

In determining the profile measurement accuracy, the vertical separation of measurement levels and also the number of measurement levels are important. A two level system is the most elementary one. Increasing the number of measurement levels involving the operational complexities is not an efficient method to increase profile measurement accuracy. Blanc (1981) has shown that quadrupling the number of measurement levels (from 2 to 8) results in only a 50% reduction in the profile measurement uncertainty. Hence the five level profile measurement in the present experiment can be considered sufficient.

In our analysis we have considered sensor spacing $\geqslant 0.8$ m for both wind and temperature sensors. Considering other sources of errors, a minimum differential of $\pm 0.10°$ C and ± 0.1 m/sec for temperature and wind respectively has been used. Pairs were formed from amongst the levels. For spacing $\geqslant$ 0.8 m we can have thus a maximum of nine pairs of levels, from the mast described above.

3.3. *Profile flux calculations*

Momentum and sensible heat fluxes were computed from the measured wind and temperature profiles using a scheme proposed by Businger *et al.* (1971), the validity of which over an oceanic area was demonstrated by Badgley *et al.* (1972). The equations based on Businger scheme to compute the fluxes are described below as in Blanc's (1982) work.

The barometric pressure and the wet bulb temperature at 5 m above sea surface (the geometric mean height=5 m) have been used and with the aid of Goff-Gratch formulation (List 1958), the relative and specific humidities, the potential, virtual and virtual potential temperature of the profile array are determined.

Gradient Richardson number (Ri_{gmh}) has been computed, at the geometric mean height (gmh) using the criterion in section 3.2 for accuracy and spacing of sensors;

$$Ri_{gmh} = Ri_{nm} = g\,(\partial\theta v/\partial z)/(Tv_{nm}+273.16)\,(\partial u/\partial z)^2 \quad (1)$$

$$(\partial\theta v/\partial z) = (\theta v_m - \theta v_n)/gmh \times \ln(Z_m/Z_n) \quad (2)$$

$$(\partial u/\partial z) = (u_m - u_n)/gmh \times \ln(Z_m/Z_n) \quad (3)$$

$$gmh = (Z_m\,Z_n)^{\frac{1}{2}} \quad (4)$$

where g is the acceleration due to gravity (9.8 m/s^2), Z is the height above sea surface (metres), subscripts n and m denote the two different levels in the profile array, u is the wind speed, Tv and θv are respectively the virtual temperature and virtual potential temperature.

The Richardson number stability parameter can be used to compute the Monin-Obukhov stability parameter (Z/L or ζ) by the following transformations :

In an unstable atmosphere $Ri \leqslant 0$ and the Monin-Obukhov stability (ζ_{nm}) is calculated from

$$\zeta_{nm}\ (-0.01 < Ri < 0) \simeq 1.3\ Ri_{nm} \quad (5)$$

$$\zeta_{nm}\ (-1.5 \leqslant Ri \leqslant -0.01) \simeq -10^{C_1+C_2+C_3} \quad (6)$$

where $C_1 = 2.844\times10^{-2}$

$$C_2 = 0.96125\ [\log\ (-Ri_{nm})]$$

$$C_3 = 1.3655\times10^{-3}\ [\log\ (-Ri_{nm})]^2$$

$$\zeta_{nm}(-\infty < Ri < -1.5) \simeq 1.05\ Ri_{nm} \quad (7)$$

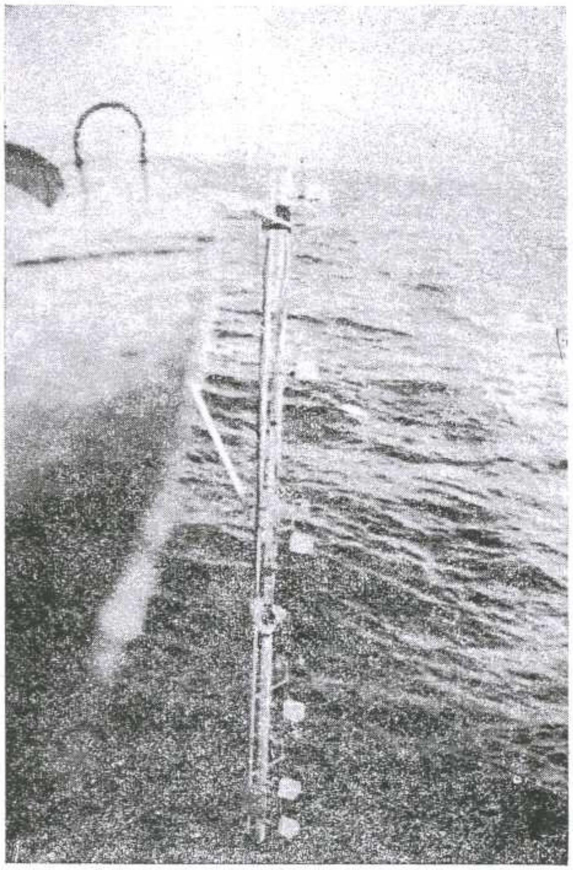

Fig. 2. Cup anemometers and thermistor temperature sensors mounted on a boom projecting outwards from the bow of the ship

For $Ri < -2$, ζ is of questionable validity

In the case of a neutral atmosphere $Ri_{nm} = 0$ and $\zeta_{nm} = 0$

In a stable atmosphere ζ_{nm} is given by

$$\zeta_{nm}(0.2 \geqslant Ri \geqslant 0) = -d_2-(d_2{}^2-4d_1d_3)^{\frac{1}{2}}/2d_1 \qquad (8)$$

where, $d_1 = (22.09\, Ri_{nm}) - 4.7$

$d_2 = (9.4\ Ri_{nm}) - 0.74$

and $d_3 = Ri_{nm}$

ζ is undefined for $Ri > 0.2$

The Monin-Obukhov length (L) in metres is calculated from the relation

$$L_{nm} = Z_{nm}/\zeta_{nm} \qquad (9)$$

for the various levels n with respect to m satisfying the accuracy and separation criterion mentioned in section 3.2. Mean of various L_{nm} is obtained from the pairs of levels and is denoted as L. Using L the Monin-Obukhov stability function ζ_n for any level Z_n has been computed by

$$\zeta_n = Z_n/L \qquad (10)$$

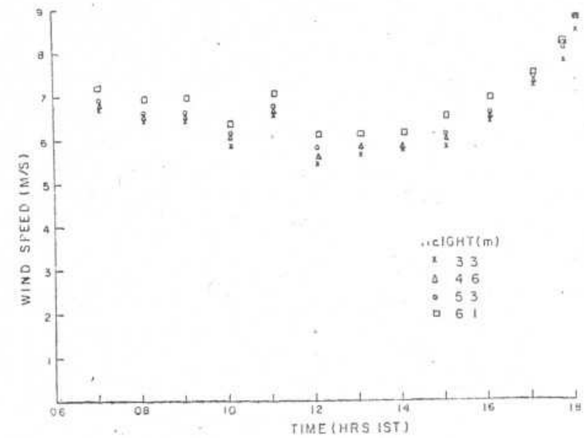

Fig. 3. Time variation of wind speed at different levels above sea surface at 7° 40'N, 78° 26'E on 16 May 1983

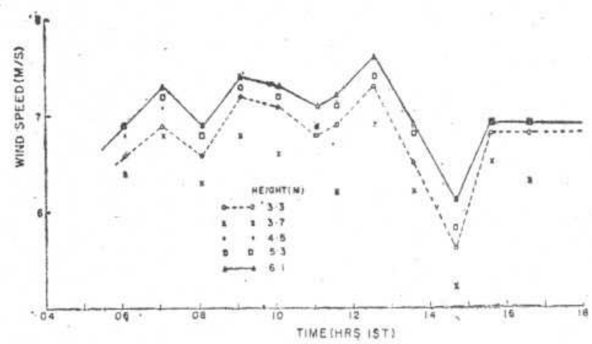

Fig. 4. Time variation of wind speed at different levels above sea surface at 12° 35'N, 80° 48'E on 20 May 1983

Log-linear profile relationship in the surface layer is valid strictly under neutral stability conditions. With increased stability or instability the non-linearity of the wind profile increases. To correct for this stability dependent non-linearity ψ corrections are applied to the measured profiles. The stability and altitude dependent profile corrections for wind speed (ψ_u), and potential temperature (ψ_θ) have been calculated at each level n using the corresponding relations :

$$\psi_{un}(\zeta<0) = \left\{ \ln\left[(1+x_n)/2\right]^2 \left[(1+x^2{}_n)/2\right]\right\} - 2\tan^{-1}x_n + \frac{\pi}{2} \qquad (11)$$

$$\psi_{un}(\zeta>0) = -4.7\ \zeta_n \qquad (12)$$

$$\psi_{\theta n}(\zeta<0) = \ln\left[(1+y_n)/2\right] \qquad (13)$$

$$\psi_{\theta n}(\zeta>0) = -4.7\zeta_n/0.74 \qquad (14)$$

where,

$$x_n = (1-15\zeta_n)^{\frac{1}{4}} \qquad (15)$$

$$y_n = (1-9\zeta_n)^{\frac{1}{2}} \qquad (16)$$

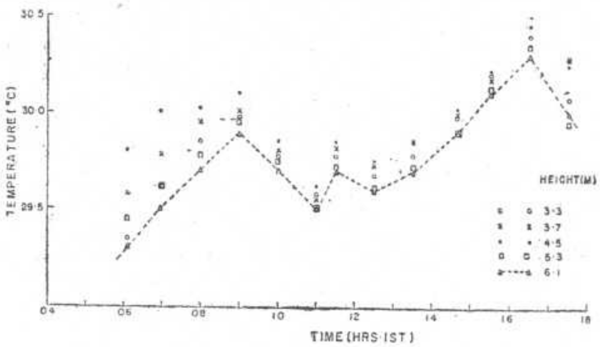

Fig. 5. Time variation of temperature at different levels above sea surface at 12° 35′N, 80° 48′E on 20 May 1983

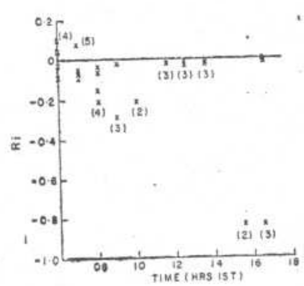

Fig. 6. Time variation of gradient Richardson number over sea surface on 20 May 1983 at 12° 35′N, 80° 48′E (Number in parenthesis indicates the number of levels used to obtain the number of pairs)

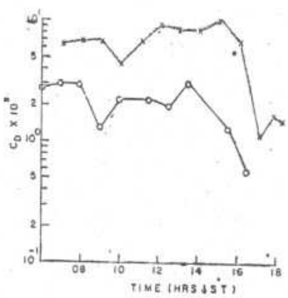

Fig. 7. Time variation of neutral drag coefficient above sea surface on 16 (x-x) and 20 (o—o) May 1983 at 7° 40′N, 78° 26′E and 12° 35′N 80° 48′E respectively

The mean stability corrected slope ($<S_p>$) (where p denotes the wind or potential temperature) of the profile has been calculated as

$$S_p(n, m) = [(\ln Z_m - \psi_{pm}) - (\ln Z_n - \psi_{pn})]/(p_m - p_n) \quad (17)$$

$$<S_p> = \left[\sum_{n=1}^{m-1} S_p(n, m)\right] \Big/ (m-1) \quad (18)$$

The scaling parameters for wind speed (u_*) and potential temperature (θ_*) have been computed by

$$u_* = k/<S_u> \quad (19)$$

$$\theta_* = k/0.74 <S\theta> \quad (20)$$

where k the Von-Karman constant is equal to 0.40. Hence from the scaling parameters the fluxes have been estimated using the following equations. The momentum flux (M) in Newton/m^2 is

$$M = -\varrho u_*^2 \quad (21)$$

where ϱ is the density of moist air in kg/m^3 and is

$$\varrho(P, Tv) = \frac{(3.4838 \times 10^{-3})P}{(Tv_{nm} + 273.16)} \quad (22)$$

where P is the barometric pressure.

The sensible heat fiux (H_s) in watt/m^2 is

$$H_s = -\varrho c_p \theta_* u_* \quad (23)$$

where c_p is the specific heat of moist air in J/kg °K.

The sign negative indicates downward flux and positive indicates upward flux.

The non-dimensional drag coefficient at the geometric mean height is computed from the frictional velocity by

$$C_D = (u_*/u_m)^2 \quad (24)$$

4. Results

4.1. *Wind and temperature profile*

In the 3-6 m layer above sea surface the time variation of wind speed is shown in Figs. 3 and 4 corresponding to the two stations. On 16th, wind speed varies smoothly during 0700-1800 hrs IST. Between 1100 & 1200 hrs on 16th there is a sudden but only a small change in wind speed ($\approx$ 1 ms^{-1}). This can be due to a light gust around 1100 hrs. It increases in the afternoon hours (1200-1800 hrs IST). Relatively the wind speed increases with height in this layer. The time variation of wind speed is not quite smooth on 20th. It varies from 5-7.5 ms^{-1}. The wind on 20th fluctuates successively in time within ± 1 ms^{-1} between 0600-1300 hrs. There is a lull around 1500 hrs. The density and period of observation being small, we can only say that it is the state of the observed wind field at that time. Hence no specific reason could at this stage be attributed to the changes.

The variation of temperature in time at different levels is shown in Fig. 5 at the station 12° 35′N, 80°48′ E. Variation of temperature is within a degree centigrade during 0600-1630 IST.

Fig. 6 depicts the time variation of gradient Richardson number (*Ri*) determined from different pairs of levels. Positive (*Ri*) values are shown by two pairs at 0600 IST and a pair at 0700 IST. Except these, *Ri* is negative throughout showing that the layer is unstable. A pronounced instability is seen at 1530 and 1630 IST with $Ri = -0.85$. An overall picture that could be derived from this is that the layer is mixed during the day between 0900-1630 IST.

4.2. *Drag coefficient*

The drag coefficient (C_D) at 6 m estimated using Eqn. (24) is shown varying with time in Fig. 7, for the case of assumed neutral wind profile for the observation on 16th, and the wind profile corrected for diabatic effects on 20th. Obviously the drag coefficient is higher on 16th by a factor of 3-5 compared to that on 20th. On 20th the lowest and the highest estimated value of C_D is 0.00057 and 0.003 respectively. Mean drag coefficient is 0.0022 which is greater by a factor of about two compared to the typical value of 0.0013 for wind speeds below 12 ms^{-1} reported in the literature (Ernest Augstein 1978).

4.3. *Momentum flux*

The momentum transferred downwards through the surface layer exerts a drag force per unit area on the sea surface which is called the surface shear stress expressed in Newton per square metre. Fig. 8 depicts the time variation of momentum flux at the station 12° 35′N, 80° 48′E on the Bay of Bengal; the negative sign in the ordinate indicates downward flux. Estimated mean of the momentum flux is -0.125 Nm^{-2}.

4.4. *Flux of sensible heat*

Time variation of sensible heat flux is shown in Fig. 9. The flux is positive during the day on 20 May 1983, implying transmission of heat away from surface. During 0700-0800 hrs the flux is relatively high. During 1130-1630 hrs it stabilises around a mean value of 30 Wm^{-2}. Estimated mean value of sensible heat flux during the entire period of observations is 49 Wm^{-2}.

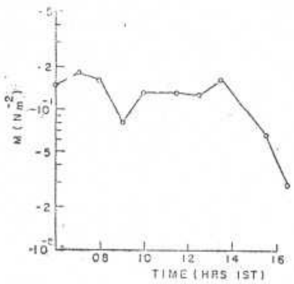

Fig. 8. Time variation of momentum flux above sea surface on 20 May 1983 at 12° 35′ N, 80° 48′E

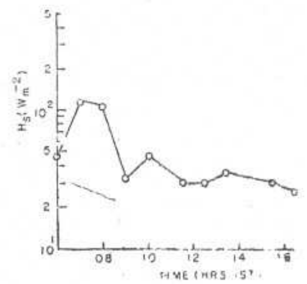

Fig. 9. Time variation of sensible heat flux over sea surface on 20 May 1983 at 12° 35′N, 80° 48′E.

5. Discussion and conclusion

The accuracy of sensors, the separation between levels in the profile array and the limit defined for the gradient Richardson number ($-2.0 \leqslant Ri \leqslant 0.2$) contain the number of pairs used to determine the Monin-Obukhov length (L). In the present analysis the number of pairs and levels used are shown in Fig. 6. The results obviously depend on this prime factor.

In Fig. 5 it can be seen that the temperature distribution at different levels in the 3-6 m layer is so random during 0600-0800 IST as to result in sharp temperature gradients which yield both positive and negative Richardson number in Fig. 6. This reveals that a transition from stability to instability occurs in the layer. Relatively high value of sensible heat flux is observed during 0700-0800 IST. The sky was observed to be partly cloudy during this period with *Ci* (3/8) and *Sc* (1/8) clouds which gives rise to an increase in flux.

Estimated mean value of the drag coefficient, flux of momentum and sensible heat from the small sample of observations during pre-monsoon period is found to be within a factor of two when compared to the values measured elsewhere by other workers (Badgley *et al.* 1972, Deacon 1956). Elaborate experiments in future are needed to provide a good statistics of data which will aid in understanding the meso-scale systems that govern the Indian monsoon.

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