

Methods for the determination of the fluxes of momentum and heat in the surface layer over the sea

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ABSTRACT. The different direct and indirect methods for the determination of the fluxes of momentum and heat in the surface layer over the sea are reviewed. Flux-gradient relationships and bulk transfer coefficients in diabatic conditions have been discussed. Comparative discussions have been made about R_i and Z/L .

1. Introduction

During the last decades, determinations of the fluxes of momentum and heat in the surface layer over the sea, by different methods, have been intensified through international co-operations, with the occurrences of IIOE, BOMEX, GATE, ATEX, AMTEX, JASIN, JONSWAP, MONEX, etc. To a reasonable agreement, most of these attempts have led to the refinement of the instrumental techniques (Schwerdtfeger 1976) as well as to that of the theoretical aspects of the methods (Lumley and Panofsky 1964, Roll 1965, and Kraus 1972), reviewed by Hasse (1970), Pond (1972) and Businger (1975).

When we consider the available methods, the selection of an appropriate method depends on the purpose for which the fluxes are to be estimated, e.g., for determinations of heat transfer at sea surface (McAlister and McLeish 1969), sea surface temperature deviations (Hasse 1971), estimation of heat flux and precipitation (Ostapoff *et al.* 1973), wave growth and swell decay (Hasselmann *et al.* 1973), sea surface temperature anomaly in relation to thermocline variability (Frankignoul and Hasselmann 1977), effect of sea surface temperature on monsoon (Shukla 1975), etc. Further, the oceanic response to large scale wind stresses is related to upwelling. While considering the wide range of applications, the reliability of the estimated fluxes depends on the methods chosen for the estimation of fluxes, which in turn depends on sea conditions, wind speed, atmospheric stratifications, instrumental

and operational problems, recording units, operational costs, etc.

The aim of this study is to review the direct and indirect methods for the determination of fluxes of momentum and heat and the bulk transfer coefficients in different conditions of the sea, wind speed and diabatic conditions of the atmosphere. Thus this review evaluates the methods suitable for the Indian Ocean region in order to investigate the air-sea interaction problems like estimation of heat flux and precipitation during the Indian monsoons.

Most of the methods discussed here are confined to the measurements of fluxes in the surface layer above the sea where the maximum transfer of kinetic energy from mean wind into turbulent motions takes place. For all practical purposes, to allow assuming constant fluxes, a height of 10-12 m has been accepted as the surface layer height or the height of the so-called constant flux layer. The definition, structure, characteristics and vertical extent of the surface layer have already been investigated in detail (Kraus 1972, Kaimal *et al.* 1972, Busch 1973, Panofsky 1974, Hasse *et al.* 1978, Wipperman 1975, Augstein 1976, Aliguseynov *et al.* 1976).

2. Theoretical aspects of available methods for the investigations on the fluxes of momentum and heat

2.1. Eddy correlation method (Cross-correlation or direct method)

Businger (1975) has reviewed the successful applications of this eddy correlation method from fixed platforms, buoys and aircraft. Pacquin

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List of symbols

c_D	drag coefficient	α	coefficient in the linearized stability dependence of the bulk transfer coefficients, with index E, H, M for latent, sensible heat and momentum
C_E, C_H	bulk transfer coefficient for latent and sensible heat	γ	Bowen ratio
c_p	specific heat with constant pressure	ϵ	dissipation of kinetic energy into heat
E	vertical flux of water vapour	θ	potential temperature
f	Coriolis parameter or frequency	κ	von Karman's constant, =0.40 approximately
H	vertical flux of sensible heat	ρ	density of the air
i	index substituting for M, H, E	τ	magnitude of the horizontal stress vector
K	eddy diffusivity, with index E, H, M for latent and sensible heat and momentum	τ_{xz}, τ_{yz}	shearing stress components flux of resp. x and y momentum in z direction
k	radian wavenumber	ϕ	stability function with index E, H, M for latent and sensible heat and momentum
K'	Kolmogoroff-constant	ϕ	spectral density, e.g., with index u for downwind velocity component
l	heat of vapourisation	$\phi_E(k)$	three-dimensional energy spectrum of turbulence
L	Monin-Obukhov-length	$\phi_1(k)$	one-dimensional energy spectrum (for the downwind component)
q	specific humidity	ψ	integrated stability function with index E, H, M for latent, sensible heat and momentum
R	radiation balance		
R_i	Richardson number		
T	temperature		
u, v, w	velocity components; with index g for geostrophic		
u_*	friction velocity		
z_0	roughness length, with index u, θ, E for wind, temperature, water vapour		
z_1, z_2	upper level of integration with ageostrophic method		

Averaging is denoted by an overbar, deviation from the average by a dash; vectors are indicated by bold types.

(1972) has dealt with some theoretical aspects of this method.

By measuring the fluctuations of vertical velocity component (w) and of horizontal wind components (u, v), temperature (θ) and specific humidity (q) the vertical fluxes can be obtained as covariances:

$$\text{Momentum flux } \tau_{xz} = -\rho \overline{u'w'}; \quad \tau_{yz} = -\rho \overline{v'w'}$$

$$\text{Sensible heat flux } H = \rho c_p \overline{T'w'}$$

$$\text{Moisture flux } E = \rho \overline{q'w'}$$

where ρ is air density, c_p specific heat at constant pressure; bar and prime denote mean and fluctuating deviations from the mean respectively. With l heat of vapourisation, the moisture flux yields the latent heat flux lE . In case of the momentum flux, (τ_{xz}, τ_{yz}) are also called the Reynolds stresses.

The physical interpretation is quite clear from the covariances itself. If, for example, excess temperature ($\theta' > 0$) is predominantly connected with upward velocities ($w' > 0$) and low temperature ($\theta' < 0$) with downward velocity ($w' < 0$), then heat is transported upwards in the average.

The covariances are almost independent of height in the constant flux layer. Under diabatic conditions the covariances are functions of stability, as can best be seen from the respective cospectra (Busch 1973).

To use this method, it is necessary to measure the fluctuating quantities in a broad frequency range, for example, 0.01 to 10 Hz, depending on height above the ground and stability. This implies a large amount of data reduction. The measurement of especially the fluctuating vertical component is difficult at sea. The method is sometimes called the direct one, since it is free

of assumptions and the above equations may be used as definition for the fluxes.

2.2. Aerodynamic methods

The aerodynamic methods are based on the Austausch or exchange concept. This is the formulation of our experiences, that the turbulent mixing, if a gradient exists, produces a transport which is down the gradient and *inter alia* proportional to the magnitude of the gradient. The coefficient of the proportionality is called the Austausch or exchange coefficient (ρ, K) or the eddy diffusivity (K). The eddy diffusivity lumps together all the complicated mechanics of turbulent transport.

For practical applications, *a priori* knowledge of the appropriate eddy diffusivity as a function of height is necessary. To avoid this difficulty, Prandtl (1932) on base of similarity arguments developed the mixing length hypothesis, applicable under condition of neutral stability. This has been generalized for non-neutral stability by Monin and Obukhov (1954) on dimensional arguments through the stability functions.

2.2.1. Aerodynamic profile method

Making use of the measured profiles and defining equations for the eddy diffusivity coefficients, the fluxes can be estimated (Roll 1965 and Hasse 1970):

$$H = -\rho c_p K_H \frac{\partial \bar{\theta}}{\partial Z}, \quad E = -\rho K_E \frac{\partial \bar{q}}{\partial Z},$$

$$\tau = \rho K_M \frac{\partial \bar{u}}{\partial Z}$$

Integration of the above equations yields the profile, *i.e.*, the variation of wind speed of temperature or humidity as function of height, where the roughness-length Z_0 enters as an integration constant (Prandtl 1932). Brocks and Krugermeyer (1972) have made a thorough review for determination of Z_0 through profile measurements under non-neutral conditions. For non-neutral stability Monin and Obukhov (1954) introduced the flux-gradient relationships:

$$K_H u_* \kappa Z \phi_H = K_E = u_* \kappa Z / \phi_E, \quad K_M = u_* \kappa Z / \phi_M$$

(κ = von Karman's constant)

Dyer (1974) has reviewed the flux-profile relationships and the following relationships were found to be preferable:

$$\phi_H = \phi_E = \phi_M = 1 + 5 \left(\frac{Z}{L} \right) \text{ for stable conditions}$$

$$\left. \begin{aligned} \phi_M &= \left(1 - 16 \frac{Z}{L} \right)^{-\frac{1}{4}} \\ \phi_H = \phi_E &= \left(1 - 16 \frac{Z}{L} \right)^{-\frac{1}{2}} \end{aligned} \right\} \text{ for unstable conditions}$$

The following relationships are used for the estimation of fluxes:

$$\frac{Z}{L} = - \frac{g}{\theta} \frac{H}{c_p \rho u_*^3} \kappa Z$$

$$R_i = \frac{\gamma}{\theta} \frac{\frac{\partial \bar{\theta}}{\partial Z}}{\left(\frac{\partial \bar{u}}{\partial Z} \right)^2}$$

$$= \frac{Z}{L} \frac{\phi_H(Z/L)}{\phi_M(Z/L)}$$

$$\frac{\partial \bar{\theta}}{\partial Z} = - \frac{H}{c_p \rho u_*^3 \kappa Z} \phi_H \left(\frac{Z}{L} \right)$$

$$\frac{\partial \bar{q}}{\partial Z} = - \frac{E}{\rho u_* \kappa Z} \phi_E \left(\frac{Z}{L} \right)$$

$$\frac{\partial \bar{u}}{\partial Z} = \frac{u_*}{\kappa Z} \phi_M \left(\frac{Z}{L} \right) \text{ where } u_* = \sqrt{\frac{\tau}{\rho}}$$

2.2.2. Bulk aerodynamic method

For the extensive studies on atmospheric and oceanic conditions, the estimation of fluxes at a number of stations over the sea, for example by the direct or the profile method will not be feasible. The only possibility is to parameterise the fluxes through bulk aerodynamic formulae, making use of the routine meteorological and oceanographic observations. In this method the fluxes can be estimated from single level observations above the sea surface and the observations near the sea surface. The bulk aerodynamic method may be conceived as an integration of the profile from the surface to the observation height (Panofsky 1963). Essentially, the complicated physics of the turbulent transport is lumped together in the bulk transfer coefficients C_D, C_H, C_E . Fricke and Schmitt (1976) have reviewed the parameterisations for sensible heat and moisture fluxes by bulk aerodynamic formulae, considering the diabatic conditions also. Duncel *et al.* (1974) have successfully applied this method to measurements from the high sea, under neutral and diabatic conditions, for determination of bulk transfer coefficients. This parameterisation uses the average values of wind speed (\bar{u}); the average air *minus* sea temperature difference ($\Delta \bar{\theta}$) and the average humidity difference between air and the saturated air at the water surface ($\Delta \bar{q}$). For measuring heights above the sea surface greater than mean wave height are preferred (Krugermeyer 1976).

$$\tau = C_D \rho \bar{u} |\bar{u}|, \quad H = - C_H \bar{c}_p \rho \bar{u} \Delta \bar{\theta}, \quad E = - C_E \rho \bar{u} \Delta \bar{q}$$

$$C_D = \kappa^2 \left[\ln \frac{Z}{Z_{0u}} + \psi_M \left(\frac{Z}{L} \right) \right]^{-2}$$

$$C_H = \kappa^2 \left[\ln \frac{Z}{Z_{0u}} + \psi_M \left(\frac{Z}{L} \right) \right]^{-1} \times$$

$$C_E = \kappa^2 \left[\ln \frac{Z}{Z_{0e}} + \psi_E \left(\frac{Z}{L} \right) \right]^{-1} \times$$

$$\text{where } \psi_i \left(\frac{Z}{L} \right) = \int_{Z_{0i}}^Z \frac{\phi_i(Z/L) - 1}{Z} dZ, \quad i = M, H, E$$

With an error of less than 3 per cent the stability functions can be linearly approximated in the stability range $-0.75 \leq R_{iv} \leq 0$ by the following relationships (Krugermeier 1976):

$$C_D = C_{D,N}^* (1 + \alpha_M R_{iv}) \quad \text{with} \quad \alpha_M = -0.331$$

$$C_H = C_{H,N}^* (1 + \alpha_H R_{iv}) \quad \alpha_H = -0.455$$

$$C_E = C_{E,N}^* (1 + \alpha_E R_{iv}) \quad \alpha_E = -0.394$$

and index N indicating neutral stability.

2.3.1. Inertial method

The inertial technique is based on the knowledge that in turbulent flow the eddies are generated at lower frequencies and dissipated at higher frequencies. For large Reynolds number flow, according to the two Kolmogoroff hypotheses there exists an inertial subrange, where the energy of turbulence is handed over to higher frequency bands without sources or sinks, and where the spectral density depends only on frequency and dissipation rate.

The indirect method, by application of Kolmogoroff's hypothesis (Wucknitz 1976 and Pacquin 1972) also known as spectral density technique (Clancy 1970, Hicks and Dyer 1972 and Garatt and Hyson 1975) can be employed for the estimation of momentum flux. In order to derive the dissipation ϵ from the spectral energy, the assumptions of this method are the conditions for isotropic turbulence and existence of inertial subrange. Then in the inertial subrange, the spectral density is determined by

$$\varphi_u = K' \epsilon^{2/3} k^{-5/3}$$

where K' is the one-dimensional Kolmogoroff's constant and k the radian wave number, obtained from frequency ($k = 2\pi f/\bar{u}$). Also, to arrive at the momentum flux, use is made of the equation for the kinetic energy of turbulence with different degrees of truncation (Wucknitz 1976). Keeping only the main terms, for neutral conditions,

$$\frac{\tau}{\rho} \cdot \frac{\partial \bar{u}}{\partial Z} = \epsilon$$

and for non-neutral stratification,

$$\epsilon = \frac{\tau}{\rho} \cdot \frac{\partial \bar{u}}{\partial Z} + \frac{g}{T} \frac{H}{c_p \rho}$$

2.3.2. Dissipation method

The dissipation may also be measured directly since (e.g., Hinze 1959, p. 179)

$$\epsilon = 2 \nu \int_0^\infty k^3 \varphi_E(k) dk$$

where ν is kinematic viscosity, k modulus of the vector radian wave number, and $\varphi_E(k)$ the three dimensional energy-spectrum. Considering, that due to the factor k^3 in the integral, the main contribution is with high wave numbers (typical length scale in the marine atmosphere few millimetres), the assumption of isotropy may be involved and the integral rewritten in terms of the one-dimensional spectrum $\varphi_1(k)$ (for an incompressible fluid):

$$\epsilon = 15 \nu \int_0^\infty k^2 \varphi_1(k) dk$$

(Monin and Yaglom II page 56). ($\varphi_1(k)$ is usually determined by Taylors Hypothesis from a time series of the downwind fluctuations.

For determination of the momentum flux from the dissipation the energy balance equation is used as in Section 2.3.1.

2.4. Budget method

For such quantities where a balance equation exists (e.g., conservation of energy for sensible heat or conservation of mass for water vapour) the flux at the sea surface may be determined for a larger volume of atmospheric air by measurement of the fluxes through the sides and the top of the volume and of the change of state within the volume. This would involve to determine one quantity (the flux at the sea surface) as a residuum of a sum of other quantities of at least equal magnitude. In praxi, this is not so much a method to determine the flux at the sea surface, but rather a method to balance the uncertainties of rather difficult measurements of different kinds in order to obtain a consistent and hopefully meaningful data set. This method has been a tool in the analysis of the IIOE, APEX, BOMEX and GATE.

2.5. Bowen ratio

This is not really a method of measurement, but a helpful device. The Bowen ratio is defined as the ratio between sensible and latent heat flux, $\gamma = H/L_E$. This at sea reasonably can be estimated to be:

$$\gamma \sim \frac{c_p \Delta \theta}{l \Delta q}$$

from the bulk aerodynamic formula.

The Bowen ratio is used to determine evaporation at sea from measurements of radiation balance. If R is radiation balance, neglecting storage and advection in the sea (!), $R = lE + H$. The partition of R between lE and H is done with aid of the estimate of the Bowen ratio. Since H in the tropics is of order 0.1 lE , a rough estimate of γ is good enough to obtain reasonable estimates of evaporation, provided the radiation balance is known.

2.6. Ageostrophic method

In the planetary boundary layer, the equations of motion contain three main terms: pressure gradient, coriolis acceleration and friction. Due to friction, a deviation from the geostrophic balance produces momentum which is finally given to the sea surface. If the wind profile is measured through the planetary boundary layer (say up to 500 m or so at sea), then integration of momentum production from a suitable height down to the surface yields the stress at the surface (with f coriolis-parameter):

$$\int_{z_1}^0 \frac{\partial \tau_{xz}}{\partial z} dz = \int_{z_1}^0 \rho f (v_g - v) dz$$

$$\int_{z_2}^0 \frac{\partial \tau_{yz}}{\partial z} dz = \int_{z_2}^0 \rho f (u - u_g) dz$$

(from the stationary, homogeneous equations of motions in the planetary boundary layer). Here, u_g, v_g are the horizontal components of the geostrophic wind. In case of the instationary, advective boundary layer u_g, v_g could be replaced by a generalized geostrophic wind (Lettau 1957, Hasse 1976), thus keeping the same equations as above. It is necessary to allow the geostrophic wind to vary linearly with height.

The upper boundary of integration z_1, z_2 can be either the height of the planetary boundary layer, where the wind is assumed to be geostrophic, or intermediate heights, if the profiles of the wind components have extreme at such heights so that the stress vanishes.

Various schemes have been used to obtain the stress with aid of these equations from the wind profile, the most successful being the approaches by Lettau (1957) and Lettau and Høber (1964), also Schriever (1966).

The data for this method are usually obtained from a series of pilot balloon ascents. Since double theodolite measurements are necessary, theodolites are usually positioned on small islands while balloons are started at sea. In old years, single theodolite observations have been taken from board of research vessels and have yielded reasonable estimates of drag coefficients.

About 20 ascents are needed to obtain an average profile. For ease of interpretation, stationary conditions are preferred.

3. Discussion

3.1. Eddy correlation method

The eddy correlation method is the only method to obtain a picture of fluctuations—variances and covariances—thereby to inform about the turbulent structure. With the aid of cospectra for these fluctuations, it can be verified whether all scales contributing to the fluxes are included or not. The outstanding attraction of the method is that it is independent of assumptions. Measurements by the covariance method can be considered as a standard by which the results obtained with other methods can be assessed (Kraus 1972, p. 153). Considering the economical aspects, the experimental set-up is extremely expensive because the measurements of variances of quantities require fast-response instruments. From the practical point of view for the collection, reduction and processing of data experienced scientists are required.

3.2.1. Aerodynamic profile method

This method mainly depends on the flux-gradient relationship and hence is not an independent method. Compared to eddy correlation method, this method is more feasible for operations, but still elaborate. It has to be considered, that to measure a profile, several (4 to 7) instruments have to be used with at least a very good relative accuracy. For practical reasons only a height range 1.5 m through 10 m can be used, where the variation of any variable is only of order 1/10 of its total difference between water surface and 10 m height. Measurements should be made from a buoy drifting, in order to keep the instruments undisturbed and oriented into the wind (see, e.g., Dunckel *et al.* 1974). Since the slope and curvature of the profiles depends on stability, a reliable determination of stability is necessary.

It will now be interesting to discuss about R_i and Z/L , the two stability terms most often used in the estimation of fluxes. The stability parameter R_i gives a measure of the onset of turbulence, whereas Z/L is a measure of relative importance of convective to mechanical turbulence. A negative sign of Z/L implies that the

stratification produces convective turbulence and a positive sign implies that the stratification acts to diminish turbulence. If Z/L is large and negative, both mechanical and convective turbulence are produced, but convective turbulence dominates, resulting into a state of free convection. On the other hand, if Z/L is large and positive the stratification acts to destroy the turbulence.

If we regard $\partial\theta/\partial Z$ and $\partial u/\partial Z$ as dependent variables then R_i would seem to be much less suitable than Z/L which contains the two independent and controlling variables u and H (Swinbank, 1968). R_i has a practical advantage over Z/L that it can be measured from profile information only, it has the disadvantage that its variation with height in the surface layer is not known *a priori*. A relation between Z/L and R_i and other stability parameter can be constructed using the stability functions of the flux-gradient-relationship given under 2.2.1. As a practical device to obtain a reliable stability parameter, Hasse *et al.* (1978) have used a bulk Richardson number, made up from the temperature difference air-sea and the mean wind speed squared. Again, a relationship to other stability parameters is obtained from the integrated flux gradient relationship given under 2.2.2.

3.2.2. Bulk aerodynamic method

Bulk aerodynamic method provides a unique approach to empirical parameterisations, thereby making the large scale studies of oceanic and atmospheric circulations comparatively easier.

The bulk aerodynamic method is applicable even under rapidly changing adjustments of the surface layer to new conditions, for time scales of longer than 10 minutes (Hasse *et al.* 1978). Since the bulk transfer coefficients are available within 20 per cent accuracy, except perhaps for high wind speeds, these parameterisations are promising for the modeling and other large scale studies.

3.3.1. Inertial method

The indirect estimation of momentum flux through this method is simpler than the other methods of measurement, provided the assumptions are fulfilled. It is especially attractive for use at sea because the estimations are independent of the ship or buoy motions, if high enough frequencies are used. As the frequency interval (*e.g.*, 2-20/sec) of spectrum is separated from dominant sea wave frequencies, the wave induced wind fluctuations as well as sensor motions do not have an effect (Wucknitz 1976).

The reliability of the method is debated for the existence of an inertial subrange. The conditions for the local isotropy in the wave number

regions mostly used for this purpose, are doubtful (Busch and Panofsky 1968).

Isotropy cannot be established from the measurement of only one component, so that an essential advantage of this method is lost (Hasse 1970). It requires further investigations, over a wide range of conditions, through comparisons with other techniques.

3.3.2. Dissipation method

As in the case of the inertial method, this method is independent of ship or buoy motions. A drawback is the necessary high frequency resolution (a few kHz). The conditions of isotropy are probably better fulfilled in this frequency range than with the inertial method. Since dissipation may be as variable as the energy input into the spectrum at the lower, energy containing frequencies, sampling over a larger time interval is necessary.

For both the inertial and the dissipation method, it is a drawback, that these methods primarily yield the dissipation. To obtain the stress, the kinetic energy balance of turbulence is involved. This requires knowledge of the simultaneous wind profile or at least knowledge of the appropriate flux gradient relationship in the presence of waves. Also, other terms of the energy equation are neglected, for example the transport of turbulent energy. For both methods it, therefore, still has to be shown that they are more reliable or more independent than the much simpler bulk aerodynamic method.

3.4. Budget

This is certainly a valuable tool for analysis of large scale experiments. In order to achieve reasonable accuracy of balances in the given volume, considerable experimental effort is necessary of ships and airplanes.

3.5. Bowen ratio

The Bowen ratio is simple to estimate. Provided that the radiation balance is known (*i.e.*, at least short wave incoming solar and sky radiation measured, which is possible on board of a ship) reasonable estimates of evaporation are obtained. The main problem is the neglect of the energy going into the sea. Although storage may be neglected in a long term annual average, advection with the currents may not be negligible, depending on the oceanographic conditions.

3.6. Ageostrophic method

This method is very valuable, since the experimental effort is small. Theodolites are easy to

use, almost any type of theodolite may be selected. Since the ageostrophic components are integrated over height, inaccuracies in the theodolite readings are not too detrimental if systematic biases are avoided. The main difficulty with optical tracking of the balloon is that the balloon is lost in low clouds. A ceiling of 500 m is aimed for.

4. Conclusion

For large scale problems, the bulk aerodynamic method together with measurements of radiation is most suitable to obtain the heat flux and evaporation at the sea surface. The momentum flux can be determined also from the bulk aerodynamic formula, additionally, the ageostrophic method is suggested for use at suitable places.

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