Computation and Modeling of the Air-Sea Heat and Momentum Fluxes

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1 Introduction

The ocean receives energy through the air-sea interface by exchange of momentum and heat. The turbulent momentum flux is the source of the wind driven circulation of the ocean, smaller scale circulation, mixed later development and wave generation. The transfer of heat across the air-sea interface determines the distribution of temperature in the ocean and without such a transfer there world be no variations in the ocean temperature (except from compression) and that part of the ocean circulation drives by temperature density variations would be absent. The turbulent fluxes of heat, globally averaged, are the largest contributions to the heat loss of the ocean. Ocean-atmosphere fluxes establish the link between ocean-surface temperature changes and atmospheric circulation variability. On the other hand, they provide the mechanism of by which ocean variability is forced by the atmosphere. Thus accurate knowledge of the flux variability is extremely important for understanding climatological and oceanic variations in the coupled ocean-atmosphere system (Pickard and Emery, 1990)

There are many requirements for surface flux values, for example, operational modelling, nowcasting and forecasting; climate change studies; atmospheric and oceanic synoptic scale studies; shelf and coastal sea studies; oceanic, climate and earth observing systems; wave forecasting; marine engineering, etc (Taylor et al., 2000).

Momentum is transferred from the atmosphere to the ocean in the turbulent shear flow at the atmospheric boundary layer near the sea surface. The momentum flux is expressed as a wind stress _ acting on the sea surface. Turbulent transfer of heat and moisture between the atmosphere and the ocean is due to the temperature and humidity gradients near the sea surface. The net heat exchange Q at the surface can be divided into four most effective terms:

$$Q = Q_s + Q_b + Q_H + Q_L \qquad (1)$$

Where Q_s is the net incoming short-wave radiation, Q_b is the net long-wave radiation, Q_H and Q_L are the turbulent fluxes of sensible and latent heat, respectively.

Incoming solar radiation Q_S

The main source of heat flux through the sea surface is incoming solar radiation Q_S, received either directly or by reflection and scattering from the clouds and the atmosphere. The rate at which short-wave solar energy enters the sea, Q_S, depends upon a number of factors including: the length of the day, absorption in the atmosphere, the sun altitude, the cloud cover and the reflection at the sea surface.

Long-wave radiation Q_b

The back radiation term Q_b is the net amount of energy lost by the sea as long-wave radiation. The value of this term is actually the difference between the energy radiation outward from the sea surface in proportion to the fourth power of its absolute temperature and that received by the sea from the atmosphere which also radiates at a rate proportional to the fourth power of its absolute temperature. The outward radiation from the sea is always greater than the inward radiation from the atmosphere and so Q_b always presents a loss of energy from the sea.

Latent heat flux Q_L

The most significant part in terms of heat transfer from the sea to the atmosphere is Q_L The rate of heat loss is the equal to the rate of vaporization times latent heat of evaporation. Q_L can be computed by application of a formula of the type $Q_L = K_e d_e / dz$, where K_e is a diffusion coefficient for water vapor and de/dz is the gradient of water vapor concentration in the air above the sea surface.

Heat conduction Q_H

Sensible heat flux is due to the temperature gradient in the air above the sea. The heat may be lost or gained from the sea surface into the atmosphere. The rate of loss or gain of heat is proportional to the temperature gradient, heat conductivity and the specific heat of air at constant pressure. In this case convection will play a role in assisting the heat loss from the ocean in times of a warmer ocean surface temperature. The sea on the whole is warmer than the air adjacent to it and therefore, Q_H is a heat loss.

2 Radiative fluxes

2-1 Direct measurement of radiative fluxes

Direct measurement of the air-sea fluxes traditionally is best provided by the pyranometer for short wave radiation and the pyrgeometer for long wave radiation. Both are similar in form and face particular problems in use at sea including contamination by salt, and especially motion of the ship or buoy at sea. With the new knowledge and technology, sophisticated radiation instruments for measuring the radiative fluxes are rapidly developing. However, the direct measurement of the air-sea fluxes are too few to contribute directly to the calculation of large scale flux fields and they are used for developing, calibrating, and verifying the parameterization formula used to estimate the fluxes from the basic variables (Taylor et al., 2000).

2-2 Parameterization of radiative fluxes

2-2-1 Parameterization of short wave radiative flux

Short wave radiation flux on the sea surface in general may be parameterize as:

$$Q_S = T_f Q_{top} \quad (2)$$

where Q_{top} is the short wave radiation at the top of the atmosphere, equal to $I_0\left(\frac{r_0}{r}\right)^2\cos(z)$ where I_0 is the solar constant, r is the Earth-Sun distance, r_0 is the mean Earth-Sun distance and Z is the Zenit angle of Earth. T_f represents the fraction of the solar radiation at the top of the atmosphere that reaches the open surface. T_f is parameterized in terms of cloud cover and thermodynamic parameters of the atmosphere. It is preferable that T_f be divided into two terms. One represents the modification of short wave radiation under clear sky condition (astronomy, temperature, humidity and aerosols) and the other the cloud modification of the clear sky radiation. In this case, the general formula for short wave radiation becomes $Q_S = Q_0 T_f'$ where Q_0 is solar radiation at the sea surface under clear sky and is assumed to be a function of the astronomy and of the transmission for the clear sky atmosphere. T'_f is the empirical function of the fractional cloud cover, air temperature, and solar altitude. Reed (1977) formulae for the daily mean net shortwave flux using the calculated mean values for the fractional cloud cover n, and the noon solar elevation in degree, ϕ is presented as: $Q_S = Q_0(1 - C_n n + .0019\phi)(1 - \alpha)$ where C_n=0.62 the cloud attenuation factor, and α is the albedo. Q_0 is the short wave insolation at the surface under clear skies.

2-2-2 Parameterization of long wave radiative flux

The amount of long wave flux is dependent to the surface water temperature, atmosphere temperature, humidity and cloud cover. For example (Clark et al, 1974)

$$Q_{LW} = \varepsilon\ \sigma\ T_S^4(039 - 0.05\ e^{0.5})(1 - \lambda\ n^2) + 4\ \varepsilon\ \sigma\ T_S^3(T_S - T_a) \quad (3)$$

Where ε =0.98, σ is the Stefan-Boltzman constant, e is the water vapor pressure, n the fractional cloud cover, and T_a and T_s are the air and sea temperatures in K. The cloud cover coefficient λ varies with latitude.

2-2-3 Radiative Fluxes by Remote Sensing

The determination of the short wave and long wave fluxes from satellite data involves measuring the radiative fluxes at the top of the atmosphere and accounting for the effects of the atmosphere using a radiative transfer model. The problem is allowing for the effects of clouds. It has been proved easier to model the scattering and absorption of short wave radiations than long wave radiation.The problem for long wave measurement is that the surface budget is dependent on the height of the lowest cloud, a quantity not easily determined by satellite.

The different algorithms differ in the cloud information used and the sophistication of the RTM. While various methods have been proposed that use the narrow band (visible) satellite radiances directly together with an RTM (Pinker and Laszlo, 1992)

3 Turbulent fluxes

Exchange between the atmosphere and ocean are most easily measured in the atmospheric surface layer where the fluctuating vertical velocity transports fluid properties up and down (Large and Pond, 1982). Turbulent fluxes of momentum and heat can be defined by normal Reynolds averages.

$$
\begin{aligned}
\tau &= \rho_a \langle u' w' \rangle \\
Q_H &= \rho_a C_{pa} \langle t' w' \rangle \qquad (4) \\
Q_L &= \rho_a L_e \langle q' w' \rangle
\end{aligned}
$$

σ is Wind stress Q_H Sensible heat flux Q_L Latent heat flux, u', w', t', q' are Turbulent fluctuations horizontal wind vertical wind, temperature, and specific humidity ρ_a Air temperature C_{pa} Specific heat of air L_e Latent heat of vaporization

Using Monin-Obukhov similarity theory

Wind stress $\tau = \rho_a \langle u' w' \rangle = \rho_a u_*^2$

Sensible Heat (5) $Q_H = \rho_a C_{pa} \langle t' w' \rangle = \rho_a C_{pa} t_* u_*$

Latent Heat $Q_L = \rho_a L_e \langle q' w' \rangle = \rho_a L_e q_* u_*$

where

Shear velocity (6) $u_* = (\frac{\tau}{\rho_a})^{1/2} = \left| \langle u' w' \rangle \right|^{1/2}$

3-1 Direct measurement of turbulent fluxes

The most direct flux measurement is the Reynolds flux method or eddy correlation method. Integration of the u', w' cospectrum $\Phi_{u'w'}(f)$ over all contributing frequencies f gives the covariance and hence momentum flux. The limitation in the extend of the measurement time and large amount of data required, and the sensitivity of the method to instrument orientation, which is a great problem on ships and buoys, are disadvantage of this method. Although the Reynolds flux method is not easily applicable to remote open sea operation, it has become the standard to which the other methods are compared (Large, and Pond, 1982).

3-2 Parameterization of turbulent fluxes

For many reasons the expressions of turbulent fluxes based on turbulent fluctuation products are not very useful, and bulk expressions relating the fluxes to more easy measurable atmospheric variables have been developed (Smith et al, 1996). In this method the turbulent fluxes are determined from formulae using the basic variables such as wind speed, air temperature, etc. The same "bulk formula" are applicable whether the basic variables have been oriented by in situ measurements or by remote sensing or have been calculated by a numerical atmosphere model. The parameterized bulk formula for momentum and heat fluxes would be presented as follows:

$$
\begin{aligned}
\tau &= C_D \rho_a u_z^2 \\
Q_H &= C_H \rho_a C_{pa} u_z (t_z - t_0) \\
Q_L &= C_E \rho_a L_e u_z (q_z - q_0)
\end{aligned}
\tag{7}
$$

Where C_D (drag coefficient), C_H (Stanton number), and C_E (Dalton Number) are transfer coefficients. The transfer coefficients have been traditionally obtained by measuring the surface fluxes, using one of several techniques, together with measurement of the mean physical variables required.

For wind stress, the drag coefficient values used in many past studies were those obtained by Large and Pond (1981, 1982) or Smith (1980, 1988). For 10m wind speeds u_{10} less than 10 m/s they found a constant value ($10^3 C_{D10n}$ = 1.12 ± 0.2), and for u_{10} between 10 and 25 m/s suggest a linear increase given by ($10^3 C_{D10n}$ =0.49 + 0.065 u_{10}) Other schemes suggest a linear relation over the entire range of validity; Garratt (1977) gives ($10^3 C_{D10n}$ = 0.75 + 0.067 u_{10} for wind from 4-21 m/s and Smith (1980) gives ($10^3 C_{D10n}$ =0.61 + 0.063 u_{10}) from 6-22 m/s (Figure 1).

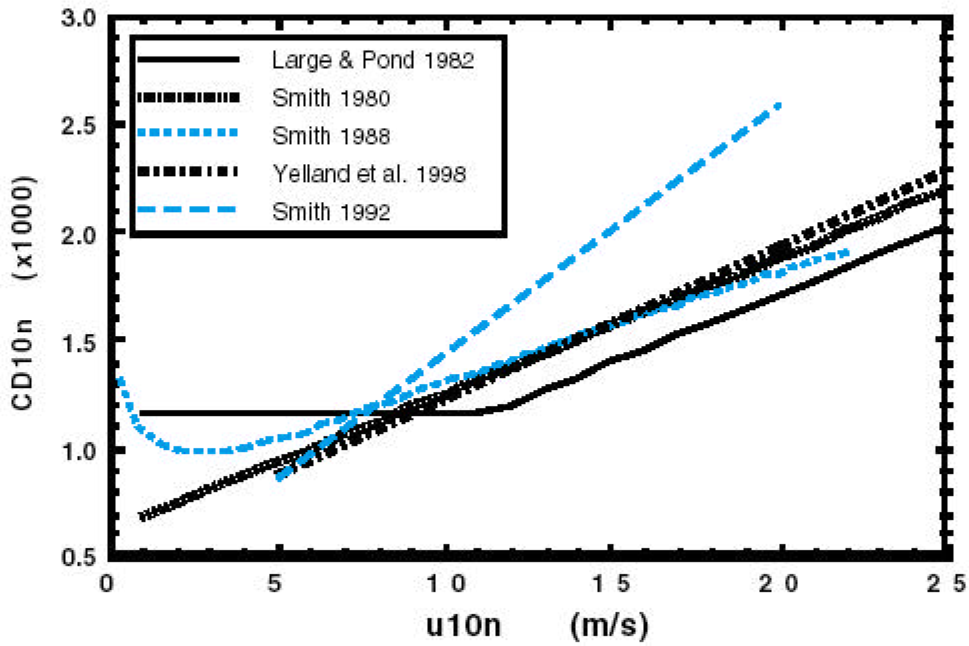

Figure 1: Examples for the Drag Coefficient, C_{D10n} plotted as a function of wind speed, *U10n*

3-4 Transfer Coefficient for Sensible and Latent Heat

Unlike the drag coefficient, traditional estimates of CE_{10n} and CH_{10n} over the ocean tend to support a fairly constant value over a wide range of wind speed. Friehe and Schmitt (1976) recommend a constant Dalton number (10^3 C_{E10n} =1.32 ± 0.07) on the basis of several turbulence-based datasets where the highest wind speed was 4 m/s. After a critical assessment of previous studies, Smith (1989) also suggested a constant "consensus" value (10^3 C_{E10n} =1.2 ± 0.1) for winds between 4 and 14 ms-1. DeCosmo *et al.*(1996) also suggest a near constant value with ($10^3 CE10n$, = 1.12 ± 0.24) for winds up to 18 m/s. For the Stanton number, Friehe and Schmitt (1976) obtained slightly different values for unstable and stable conditions (10^3 C_{H10n} = 0.97 and 0.86 respectively. Smith (1988) suggested $10^3 C_{E10n}$ =1.0.

4 References

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