THE TOGA COARE BULK FLUX ALGORITHM

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1. **Introduction**

The importance of air-sea interaction to the earth's climate is widely appreciated. The pivotal role of the tropical oceans in climate and interannual climate variability led to the establishment of the Tropical Ocean Global Atmosphere (TOGA) program and the subsequent identification of the dominance of the Pacific Ocean in this variability resulted in the Coupled Ocean Atmosphere Experiment--COARE (WCRP, 1990; Webster and Lukas, 1992). The interfacial fluxes are one of three elements of the COARE program. The COARE science plan (WCRP, 1990) identifies several fundamental gaps in our knowledge relevant to fluxes:

> "Atmospheric response in models is extremely sensitive to SST variations, especially where SST is warm. However, ocean models almost universally predict temperatures which are too warm, probably associated with poor assessments of heat, momentum, and moisture fluxes to the ocean and atmosphere."
>
> "The heat balance of the warm pool region of the western Pacific is poorly known with discrepancies as large as 80 Wm^{-2}. The relative involvement of the slowly evolving atmosphere or the higher frequency, more episodic, equatorial events is not understood."

Webster and Lukas (1992) emphasize that "the variation of fluxes between the ocean and the atmosphere is very sensitive to the choice of parameterization, especially in low wind regimes." This has been verified by Miller et al. (1992), who found dramatic improvements in simulated tropical phenomena by strengthening the air-sea coupling in the light wind regime. Furthermore, observational problems in the climatological database--particularly the air-sea temperature difference (Lukas, 1989), the strong boundary-layer diurnal cycle in light winds, and the unknown sensible cooling associated with precipitation represent additional uncertainties in assessing the surface energy balance of the warm pool. In summary, our ability to diagnose, simulate, and predict climate and climate variability is handicapped by a general lack of high-quality data in the region and inadequate parameterizations of air-sea fluxes.

The broad issue of the interfacial turbulent fluxes can be broken down into problems in direct measurement, estimation, and model parameterization. The direct measurement of the fluxes strictly involves computation of the mean covariance of the relevant fluctuating meteorological variables measured near the sea surface. Flux estimation involves the calculation of fluxes from associated data such as the high-frequency part of the meteorological spectrum (the inertial-dissipation method) or the difference in mean surface and mean meteorological values at some reference height (the bulk method). Flux estimations are parameterizations because they rely on similarity theory and empirical functions and coefficients (Geernaert, 1990). Model flux parameterizations are in all cases based on the application of the bulk flux estimation method to model-derived values of the relevant meteorological variables (rather than actual data). Such model-derived variables are generally obtained at space and time scales quite different from the original fundamental measurements (typically 1-h samples at a fixed point) used to obtain the empirical functions and coefficients (although this has not intimidated modelers). Critical study of the disparity between fluxes and flux

estimation methods derived from classical micrometeorological techniques and the spatial/temporal resolution of numerical models that leads to the subgrid-scale variability, vertical resolution, and diurnal variability is hampered by the lack of a consistent and verified 1-h flux algorithm (Hanawa and Toba, 1987; Liu, 1988; Ledvina et al., 1993).

1.1 Bulk flux estimations

Simultaneous flux and bulk meteorological variable measurements combined with laboratory studies of air-sea transfer processes are used to develop the bulk formulae and transfer coefficients. The classic reviews on this subject (Garratt, 1977; Smith, 1988) reveal a substantial midlatitude bias in the field measurements with most of the data obtained in the 4 to 15 ms^{-1} wind speed regime. The majority of the data are also from offshore towers, coastal areas, and other shallow water regimes. Uncertainties in the average neutral coefficients as a function of wind speed remain the critical question. Blanc's (1985) study showed a factor of two variation in suggested values for the humidity transfer coefficient with a consensus uncertainty of about 30%. Note that a 10% uncertainty in this transfer coefficient results in a 10 Wm^{-2} uncertainty in the latent heat flux and, thus, the surface energy budget of the warm pool.

In low wind speed regimes it is necessary to account for hydrostatic stability effects and standard stability-dependent bulk schemes (e.g., Liu et al., 1979; Smith, 1988) have shown good performance in the tropics (Bradley et al., 1991). However, a careful analysis (Godfrey and Beljaars, 1991) has shown these schemes to become singular at winds speeds below about 0.5 ms^{-1}. This occurs when a basic similarity profile assumption (that the roughness length is much smaller than the Monin-Obukhov length) is violated. Godfrey and Beljaars show that this singularity can be eliminated by adding a "gustiness" velocity, w_g, related to the normal convective scaling velocity, which accounts for the fact that the amplitude of the mean wind vector does not properly characterize the mean wind speed in light winds. Because 1-h average point winds of less than 4 ms^{-1} occur about half of the time in the COARE region, particular attention must be paid to this problem. Recent experimental studies in the COARE region have shown unequivocally that the scalar fluxes do not go to zero in the limit of zero mean wind (Bradley et al., 1991; Young et al., 1992; Fujitani, 1992; Bradley et al., 1993)

1.2 Scope of this paper

This paper is concerned with the estimation of air-sea fluxes from bulk variables, with a focus on a specific algorithm developed for the TOGA COARE investigators. Following background material on fluxes and similarity theory, we discuss the representation of the near-surface transfer processes in terms of the surface roughness parameters, theoretical issues associated with extending traditional methods to the light wind, convective regime, the proper thermodynamic constants for the computation of the fluxes, and the estimation of the rain fall contribution to surface cooling (section 2). In section 3 we touch briefly on several measurement issues: the special problem of flux measurements from ships, the accuracy requirements for bulk variables, and a method to correct bulk water temperatures to obtain the true interfacial sea surface temperature. In section 4 we describe the specific implementation of these concepts into the COARE bulk flux algorithm. A comparison with the algorithm with the Moana Wave COARE data is given in section

5 and our conclusions in section 6.

2. Theory

2.1 Background

The turbulent fluxes of sensible heat, latent heat, and stress components are defined by the normal Reynolds averages:

$$\begin{aligned} H_s &= \rho_a c_{pa} \overline{w'T'} = -\rho_a c_{pa} u_* T_*, \\ H_l &= \rho_a L_e \overline{w'q'} = -\rho_a L_e u_* q_*, \\ \tau &= \rho_a \overline{w'u'} = -\rho_a u_*^2 \end{aligned} \qquad \textbf{(1)}$$

where w',T',q' and u' represent the turbulent fluctuations of vertical wind, temperature, water vapor mixing ratio and the streamwise component of horizontal wind; T_*, q_*, and u_* are the related Monin-Obukhov Similarity (hereafter, MOS) scaling parameters (Panofsky and Dutton, 1984; Geernaert, 1990). The overbar denotes an ensemble average, but is usually in practice a time or space average.

The standard bulk expressions for the scalar fluxes and stress components are:

$$\begin{aligned} H_s &= \rho_a c_{pa} C_h S(T_s - \theta), \\ H_l &= \rho_a L_e C_e S(q_s - q), \\ \tau_i &= \rho_a C_d S(u_{si} - u_i) \end{aligned} \qquad \textbf{(2)}$$

where C_d, C_h and C_e are the transfer coefficients for stress, sensible heat and latent heat; θ is the potential temperature, q the water vapor mixing ratio, and u_i one of the horizontal wind components relative to the fixed earth, each measured at some atmospheric reference height z_r and averaged as in (1). T_s is the sea surface interface temperature, u_{si} the surface current, and q_s the interfacial value of the water vapor mixing ratio which is computed from the saturation mixing ratio for pure water at the sea surface temperature,

$$q_s = 0.98\, q_{sat}(T_s) \qquad \textbf{(3)}$$

Alternatively, we may measure the wind components relative to the sea surface, in which case the u_{si} terms are zero. Following Sverdrup et al. (1942), the factor of 0.98 multiplying the saturation specific humidity of the sea surface temperature takes into account the reduction in vapor pressure caused by a typical salinity of 34 ppth. Note that

$$\theta = T_a + 0.0098 z_r,$$
$$q = RH\ q_s(T_a), \tag{4}$$

where T_a is the air temperature and RH the relative humidity. In (2), S is the average value of the wind speed relative to the sea surface at z_r.

The transfer coefficients in (2) are partitioned into individual profile components

$$C_h = c_T^{1/2} c_d^{1/2},$$
$$C_e = c_q^{1/2} c_d^{1/2}, \tag{5}$$
$$C_d = c_d^{1/2} c_d^{1/2}$$

which are themselves functions of the fluxes in a manner described by MOS surface-layer theory (Panofsky and Dutton, 1984; Geernaert, 1990)

$$c_T^{1/2} = c_{Tn}^{1/2} / [1 - \frac{c_{Tn}^{1/2}}{a\kappa} \psi_h(\xi)],$$
$$c_q^{1/2} = c_{qn}^{1/2} / [1 - \frac{c_{qn}^{1/2}}{a\kappa} \psi_h(\xi)], \tag{6}$$
$$c_d^{1/2} = c_{dn}^{1/2} / [1 - \frac{c_{dn}^{1/2}}{\kappa} \psi_u(\xi)]$$

Here κ is the von Karman constant (0.4), a accounts for the difference in scalar and velocity von Karman constant, ψ is the MOS profile function (assumed the same for temperature and humidity) and $\xi = z_r/L$ where

$$L^{-1} = \frac{\kappa g}{T} (T_* + 0.61 T q_*) / u_*^2 \tag{7}$$

The subscript 'n' denotes the value in neutral conditions (i.e., $\xi=0$) where $\Psi=0$. For a reference height of 10 m, C_{en} and C_{hn} are approximately 1×10^{-3} and have little wind speed dependence (Liu *et al.*, 1979; Smith 1988). The neutral transfer coefficients are related to the roughness lengths, which are defined as the height where the extrapolation of the log-z portion of the variable profile intersects the surface value:

$$c_{Tn}^{1/2} = \frac{a\kappa}{\log(z_r/z_{oT})},$$

$$c_{qn}^{1/2} = \frac{a\kappa}{\log(z_r/z_{oq})}, \qquad \textbf{(8)}$$

$$c_{dn}^{1/2} = \frac{\kappa}{\log(z_r/z_o)}$$

The scaling parameters from (1) can be computed independently from the transfer coefficients given in (6):

$$T_* = -c_T^{1/2}(T_s-\theta),$$

$$q_* = -c_q^{1/2}(q_s-q), \qquad \textbf{(9)}$$

$$u_*^2 = C_d S u$$

where u denotes the mean wind vector magnitudes (relative to the sea surface).

2.2 Surface Characterization

The velocity roughness length, z_o, is often crudely related to the physical roughness of the surface (see page 123 in Panofsky and Dutton, 1984) but the scalar roughness lengths are more complicated. This is discussed in detail in chapters 4 and 5 of Garratt (1992) or chapter 5 of Kraus and Businger (1994), so only brief background will be given here. From laboratory studies, it has proven convenient to characterize the surface and the flow regime by the roughness Reynolds number

$$R_r = \frac{u_* z_o}{\nu} \qquad \textbf{(10)}$$

where ν is the kinematic viscosity of air. When R_r<.13 the flow is said to be "aerodynamically smooth", i.e., the actual roughness elements on the surface are irrelevant and the surface stress is supported by viscous shear. As the wind speed decreases, R_r approaches a constant value of about 0.11 and the relationship between roughness and stress is fixed:

$$z_o = \frac{0.11\,\nu}{u_*} \qquad \textbf{(11)}$$

For R_r>2.0 the flow is "rough" and the stress is dominated by pressure and viscous transfers

associated with the roughness elements.

Over the ocean, smooth flow occurs for 10-m wind speeds less than about 2 ms^{-1} and rough flow occurs for wind speed greater than about 8 ms^{-1} (Fig. 1). The roughness elements over the ocean are primarily surface gravity waves which are themselves generated by the wind/stress. Based on scaling arguments about the slope of the average locally generated seas as a function of the surface stress, Charnock (1955) gave the mean relationship between oceanic roughness and stress for rough flow:

$$z_o = \frac{\alpha u_*^2}{g} \qquad \textbf{(12)}$$

where α is the "Charnock" constant for which values between 0.010 and 0.035 can be found in the literature (e.g, Table 4.1 in Garratt, 1992). The value of the Charnock "constant" can be linked to gross characterizations of the sea state (Geernaert, 1990; Nordberg, 1991) such as the age or slope of the dominant wavelength (from the peak of the gravity wave spectrum). This has been used to explain, for example, the increase of the velocity transfer coefficient in shallow water. The literature on this subject is quite conflicting so at this stage the subject must still be considered exploratory. Broader application also awaits the easy available of wave spectral information. There have also been suggestions (Wu, 1968; Bourassa, 1993) that capillary waves contribute significantly to the stress at intermediate wind speeds. This implies that the surface tension of the surface of the ocean must be considered in wind/stress relationships. So far, clear experimental verification is sketchy.

For 10-m wind speeds between 2 and 20 ms^{-1} decades of field programs have not succeeding in clearly demonstrating an open ocean neutral transfer coefficient for heat and moisture that is significantly different than $1.1 \times 10^{-3} \pm 15\%$ (Garratt, 1992). Despite this, the trend in characterizing heat and moisture transfer has been to follow laboratory and over land studies in parameterizing the scalar roughness lengths in terms of the roughness Reynolds. The reasons are that the 15% uncertainty is no longer acceptable plus the roughness structure gives us an approach that can be extended to non-equilibrium wave states and that can be used to deal with air-sea transfer of trace gases. The leading examples are the model of Liu et al. (1979 - hereafter LKB) and Brutsaert (1982). In the laboratory, simple experiments have been done to determine the transfer of heat and moisture by molecular diffusive processes in the thin sublayer directly adjacent to the water surface. Brutsaert assumed that this near surface profile must match the log-profile of turbulent transfer at some matching height. This leads to a relationship between the scalar roughness length (i.e., the log-profile variable) and the velocity roughness length that depends on R_r. The LKB model is quite similar to that of Brutsaert except the sublayer profile is given a specific (exponential) shape and the matching height is determined by equating the slopes of the two profile forms where they intersect. This difference, plus different choices of sublayer constants and different specifications of the velocity behavior leads to substantial differences in the two models (Fig. 2).

2.3 Convective Behavior

Within the framework of MOS, many dynamical variables have clearly defined asymptotic

behavior in the so-called free convection limit (Panofsky and Dutton, 1984; Garratt, 1992) - when u_* approaches zero but the sensible heat flux does not (i..e., L goes to zero). For example, the dimensionless vertical gradients of scalar quantities are expected to exhibit a $\xi^{-1/3}$ dependence. This leads to a scalar profile function of the form

$$\psi_c = 1.5 \ln[\frac{y^2+y+1}{3}] - \sqrt{3} Arctan[\frac{2y+1}{\sqrt{3}}] + \frac{\pi}{\sqrt{3}} \qquad \textbf{(13)}$$

where

$$y = \sqrt[3]{1-\gamma \xi} \qquad \textbf{(14)}$$

and γ is an empirical constant. Note that the convective limit argument says nothing about the form of these functions near neutral stability nor does it apply to the horizontal wind profile function. Numerous overland field programs have determined the forms of the profile functions for near-neutral conditions (Hogstrom, 1988), but the experimental difficulties of measuring the small gradients in the convective limit have prevented clear verification of (13).

It is clear, however, that expressions such as (6) may become singular if ψ becomes too large. Furthermore, Godfrey and Beljaars (1991) have pointed out that (6) cannot be applied in the strict convective limit because it is based on the constraint that the magnitude of $z_{oq}/L << 1$. This has resulted in attempts to sidestep the MOS framework and scale the problem directly (see Liu, 1989, for a discussion). Laboratory studies have indicated that sensible heat flux scales as the 4/3 power of the air-sea temperature difference through scaling arguments that depend on the molecular diffusivities. These concepts have been generalized and applied in limited way to the ocean (Golitsyn and Grachev, 1986). Following an approach based on boundary layer convective similarity, Stull (1994) has developed a scaling model where the sensible heat flux scales as the 3/2 power of temperature difference and the 1/2 power of the depth of the convective boundary layer, z_i. Note that Stull's theory was originally for smooth flow, where the actual physical roughness of the surface is no longer relevant (i.e., z_i is the length scale), but he has suggested that it is also valid for rough flow.

Another approach has been developed by noting that in (2) the parameter S is in fact the average value of the wind speed, not the magnitude of the mean wind vector. Schumann (1988) and Godfrey and Beljaars (1991) have expressed S

$$S^2 = u_x^2 + u_y^2 + w_g^2 \qquad \textbf{(15)}$$

where u_x and u_y are the mean wind components and w_g is proportional to the convective scaling velocity

$$w_g = \beta W_* \qquad \textbf{(16)}$$

and β is an empirical constant, on the order of 1.0, but which depends on the temporal/spatial scale used to compute the averages. We compute W_* as follows,

$$W_*^3 = \frac{g}{T}\left[\frac{H_s}{\rho_a c_{pa}} + 0.61T\frac{H_l}{\rho_a L_e}\right] z_i \qquad \textbf{(17)}$$

A bit of algebra will show that as the mean vector wind approaches zero, S will approach $\beta * w_g$ and (2) will yield a result equivalent to Stull's (1994) scaling theory.

Sykes et al. (1993) have carried this concept further by examining the structure of the local profiles within the gusts. Whereas (2c) properly implies that the average stress vector approaches zero as the average vector wind, the local turbulence intensity that drives the scalar fluxes and determines the local roughness length within the gusts must be scaled by the wind speed

$$u_{*t}^2 = C_d S^2 \qquad \textbf{(18)}$$

The point is that, if the wind blows at 1 ms^{-1} from the east for half and hour and then blows at 1 ms^{-1} from the west for half and hour, it is the average wind speed of 1 ms^{-1} that must be used to compute roughness Reynolds number, the scalar transfer coefficients, and the MOS stability parameters.

2.4 Flux/Moisture Corrections

The flux accuracy guidelines set by the COARE working groups have placed unprecedented demands on measurements and computation. We no longer have the luxury of being able to use "ballpark" figures for geophysical parameters such as the acceleration of gravity, the latent heat of vaporization of water, the vapor pressure of pure water (as opposed to seawater), etc. Businger has discussed at length the issue of the additional contributions to the heat carried by atmospheric moisture (Businger, 1982). He shows that the total enthalphy transported by turbulent correlations is

$$\overline{\rho_a w h} = \rho_a \overline{wT}\,[c_{pa} + q(c_{pv} - c_{pa})] + \rho_a \overline{wq}\,[c_{pv}(T - T_r) + L_e] \qquad \textbf{(19)}$$

where ρ_a is the density of **moist air**, c_{pa} is the specific heat of **dry** air, c_{pv} the specific heat of water vapor, and T_r a reference temperature. This development follows Frank and Emmitt (1981) and corrects an erroneous expression developed by Brook (1978). Businger (1982) points that the proper reference temperature in this application is the sea surface temperature, T_s. The second c_{pv}-term representing the heat required to cool the water vapor from T_s to the air temperature after it is evaporated. Thus, the total sensible heat is

$$H_s = \rho c_{pa}[\overline{w'T'}\,(1 + q(c_{pv} - c_{pa})/c_{pa}) + \overline{w'q'}(T - T_s)c_{pv}/c_{pa}] \qquad \textbf{(20)}$$

For TOGA COARE both of these correction terms are, on average, less than 0.2 W/m^2 and of opposite sign. Because it is unlikely that bulk fluxes will approach this accuracy, even in an average sense, in the near future, we suggest that these correction terms be neglected.

A second issue is the proper lower boundary condition for the total mass flux. Webb Pearman and Leuning (1980) have shown that the requirement that the dry mass flux be zero at the surface leads to a small mean vertical velocity in the presence of nonzero sensible and latent heat fluxes:

$$\overline{w} = 1.61\overline{w'q'} + (1 + 1.61q)\overline{w'T'}/T \qquad \textbf{(21)}$$

This leads to a correction that must be added to the latent heat flux

$$H_{lW} = \rho_a L_e \overline{w} q \qquad \textbf{(22)}$$

For the COARE region, H_{lW} has a average value of 4 Wm^{-2}.

2.5 Precipitation cooling

Gosnell et al. (1994) have shown that the sensible heat transferred to the ocean surface by the rain can be represented as

$$H_{sr} = -Rc_{pw}\alpha_w\left(1 + B_o^{-1}\right)\Delta T, \qquad \textbf{(23)}$$

where R is the rain rate (liquid water flux), c_{pw} the specific heat of liquid water, α_w the Clausis-Claperyon wetbulb factor (Fairall et al., 1993), B_o the bulk Bowen ratio, and ΔT the sea-air temperature difference. This equation assumes that the rain is at the wetbulb temperature. Assuming a sea-air temperature difference of 2°C, the mean rainfall for the COARE IOP of 10 mmd^{-1} gives a heat flux of 2.5 Wm^{-2}. Climatologically, this is less than our error in the total heat budget of the warm pool, although in a given 1-h period it can be 30 times larger. We include it as part of our considerations of the bulk flux algorithm because it can be written in terms of simple bulk variables.

3. Measurement Issues

Details on the Moana Wave instruments and various data processing techniques have been provided elsewhere (Fairall et al., 1990; Fairall and Young, 1991; Edson et al., 1991; Young et al., 1992; Hare et al., 1992; Chertock et al., 1993; Fairall and White, 1993; Young et al., 1994). Also, a complete description of the various intercomparisons and instrument calibration and correction procedures for the COARE flux platforms is given by Bradley et al. (1994).

3.1 Flux measurements

The general issue of direct covariance flux measurements is well documented and needs no further discussion here (e.g., Panofsky and Dutton, 1984; Businger 1986; Wyngaard, 1990). Such measurements from ships involve additional complications caused by ship motions, flow distortion, and the contaminating effects of the marine environment. Techniques to correct anemometer velocities for ship motion are derived from earlier aircraft work, with most of the effort being directed toward developing a motion system that does not cost more than the meteorological system. Early efforts in this realm were made by Fujitani (1985), but now several investigators have reported success (e.g., Tsukamoto et al., 1990; Hare et al., 1992). To date, only Hare (1992) has carefully examined the correction equations (this involved the computation of 300 3rd order covariance terms) to determine the sources of motion error in stress measurements and to devise an instrument location strategy to minimize them. Oost et al. (1993) examined the flow distortion effects of both large structures (e.g., a platform or a ship's superstructure) and small objects near the anemometer (such as mounting brackets or other instruments). Unfortunately, there is no theory to describe the first type of distortion effect and wind tunnel simulations, which work nicely for the mean flow distortion, failed to give credible results for the fluxes. Thus, empirical methods such as intercomparisons with other measurement platforms (presumably with no or at least much less distortion) or flux estimation methods are necessary. Naturally, the minimization of motion errors and the minimization of flow distortion are mutually exclusive. Combating the marine environment requires careful selection of robust instruments, extremely attentive monitoring of sensor performance, and frequent cleaning (Fairall et al., 1990; Edson et al., 1991; Larsen et al., 1993). Clearly, ship-ship, ship-buoy, and ship-aircraft intercomparisons are essential to establish the credibility and accuracy of the ship flux measurements. However, the reader is warned that the flux measurements used here (particularly covariance stress measurements) are still not perfect.

3.2 Bulk Measurements

The TOGA/COARE goal of an 10 Wm^{-2} uncertainty in the total surface energy budget of the ocean (including turbulent, radiative, and precipitation heat fluxes) implies certain accuracy requirements for the bulk measurements. Partitioning the error equally between the net radiative components and the turbulent components (neglecting the uncertainty in the precipitation component) and assuming that these errors are independent, suggests an allowable uncertainty in (H_l + H_s) of about 6-7 Wm^{-2}. Assuming that the average uncertainty in the empirical transfer coefficients can be made arbitrarily small, then the uncertainty in the bulk heat fluxes computed via (2) can be approximated (see Blanc, 1986 and 1987, for detailed analyses)

$$\frac{\delta F_x}{F_x} = \delta S/S + \frac{(\delta X_s - \delta X)}{(X_s - X)} \qquad \textbf{(24)}$$

This leads to the following limits on the systematic errors in the mean variables (Fairall and McPhaden, 1993):

$\delta S \approx \delta u = 0.2$ ms^{-1}
$\delta T_s = 0.2$ K
$\delta T = 0.2$ K
$\delta q = 0.2$ gkg^{-1}

Typical claimed accuracies for commercial humidity sensors are 3%. Because the sea-air humidity difference in the COARE region is about 1/3 the mean humidity, this translates into an uncertainty in the derived transfer coefficient of about 10%, not including average errors in the determination of the humidity flux or the SST. Remember that we can reduce the contribution of random errors by averaging, but not systematic errors. To meet the COARE goal of a 10 Wm^{-2} uncertainty in the heat balance of the warm pool, unprecedented accuracies must be obtained in fluxes and mean meteorological variables. This requires examination of sensor calibrations, fast sensor response, flow distortion and ship influence, and processing methods. To accomplish this, we developed a program of post calibrations and analysis of side-by-side field data with other ships, buoys, and aircraft. Details are given in Bradley et al. (1994); estimates of measurement accuracies are given in Table 1.

3.3 Sea Surface Temperature

Sea surface temperature presents special problems in bulk flux applications because of the variety of measurement methods. Strictly speaking, the temperature required in (2) is the interfacial temperature of the water that is in direct contact with the atmosphere. Because the sensible, latent, and longwave radiative fluxes are realized in the upper fractions of a millimeter of the surface, they lead to a "cool skin". Thus, the interface is about 0.2 to 0.5 K cooler than the water a millimeter below the surface. Typical bulk thermometers placed in the water to make contact temperature measurements are usually unable to resolve this thin layer and their readings characterize the bulk temperature below the cool skin. A properly calibrated and corrected radiative thermometer can measure the actual interface temperature unambiguously (e.g., Schluessel et al., 1990). About half the solar radiation is absorbed in the upper meter of the ocean, so there are substantial diurnal variations in the water temperature profile in the upper few meters. The details of the profile are determined by the solar absorption characteristics of the water and the convective and wind-driven turbulence in the ocean mixed layer. In light winds the surface temperature can warm by several degrees and this warm layer may be less than a meter deep; for wind speeds in excess of 7 ms^{-1} the effect is negligible for purposes of computed bulk fluxes. However, this means that unless the bulk temperature sensor is placed within a few cm of the sea surface, it may not read the correct surface temperature associated with this "warm layer".

Clearly, radiative sea surface temperature sensors do not require corrections for cool skin and warm layer effects, but bulk sensors do. We have incorporated into the COARE 2.0 algorithm corrections for both effects. The details are described in Fairall et al. (1994). The cool skin and warm layer algorithms require additional data inputs of solar and IR radiative fluxes. The warm layer algorithm requires specification of the sensor depth. The correction is based on integrals of the surface energy and momentum budgets (Price et al. 1986), so a reasonably complete time series is required (i.e., it cannot be applied to a single measurement by itself).

4. Flux Estimation Algorithm

4.1 Description

The COARE 2.0 bulk flux algorithm is presently structured in terms of roughness Reynolds number following the approach of LKB. Based on data from recent measurement programs, including COARE, we have adjusted some of the empirical constants and added additional physics as described in sections 2 and 3. The specifics are as follows.

(i) Combining the formulae of LKB and Charnock for the roughness length to obtain

$$z_o = \alpha \frac{u_*^2}{g} + 0.11 \frac{\nu}{u_*} \qquad \textbf{(25)}$$

as suggested by Smith (1988).

(ii) Altering the dependence on stability of the profiles of temperature, moisture and momentum in highly unstable conditions $\xi = z_r/L << -1$. To do this, we begin with profile functions that obey the $\xi^{-1/3}$ asymptotic convective limit dependence on stability as discussed in section 2.4. This form is blended with the standard Kansas type (Businger et al., 1971) functions (Ψ_{TK} and Ψ_{uK}) in the following manner

$$\psi_x = \frac{1}{1+\xi^2}\psi_{xK} + \frac{\xi^2}{1+\xi^2}\psi_c \qquad \textbf{(26)}$$

(iii) Following Godfrey and Beljaars (1991) we use a gustiness velocity, w_g, in the computation of the mean wind speed as in (15) and (16). For COARE we have used z_i=600 m based on ceilometer and aircraft measurements of the typical boundary-layer cloud base height. A value of β of 1.25 was determined from the Moana Wave measurements of horizontal velocity variance (i.e., $\beta^2=(\sigma_u^2+\sigma_v^2)/W_*^2$) at 50-min time scales.

(iv) Cool skin and warm layer corrections for non-radiative sea surface temperature measurements are incorporated into the algorithm.

(v) The LKB parameterization of R_q and R_T in terms of R_r has been adjusted slightly in the range of $0.12<R_r<1.0$.

(vi) The "Webb" correction to the heat flux is computed as in (22) and the precipitation heat flux is computed as in (23).

4.2 Implementation

The algorithm is presently structured to process data from a given platform with a particular instrument configuration that is operating in the same general time zone. The main program consists of several modules to compute specific variables:

Step 1. Platform inputs.
Input atmospheric measurement heights and water temperature sensor depth.
Input inversion height, surface pressure, approximate latitude of measurements.
Set all predetermined constants (von Karman, etc.)

Main Loop

Step 2. Input a line of data from time series:
u, T_s, q, R, $R_{l\downarrow}$, $R_{s\downarrow}$
Correct T_s and q_s for warm layer from previous main loop step

Step 3. Compute first guess transfer coefficients and u_{*t}, T_*, and q_*
Compute all temperature dependent constants
Set loop count to 0

Step 4. Stability iteration
Compute ξ (eq. 7)
Compute z_o (eq. 25)
Compute R_r (eq. 10)
Compute R_q and R_T (modified LKB formula)
Compute z_{oq} and z_{oT} (invert eq. 10)
Compute neutral transfer coefficients (eq. 8)
Compute ψ functions (eqs. 13 and 26)
Compute u_{*t} (eq. 18), q_*, and T_* (eq. 9)
Compute fluxes (eq. 1)
Compute W_*
If unstable, compute w_g (eq. 16); else w_g=0.1
Compute cool skin correction to T_s and q_s
Increment loop count
If loop count < 20 then go to Step 4

Step 5. Compute Webb and precipitation fluxes. Increment integrals for warm layer.
Compute u_* and stress.

Step 6. Go to Main Loop

4.3 Choice of Constants

The following constants are used in the COARE 2.0 version:

Charnock	α=0.011
Convection	β=1.25
Convective profile	γ=12.87
Von Karman	κ=0.40
Scalar correction	a=1.00
Gravity	g=9.72 ms^{-2}
Inversion height	z_i=600 m

Also, L_e (page 113, Fleagle and Businger, 1980) and ν (page 31, Andreas, 1989) are given well known temperature dependent forms.

5. COARE Analysis

5.1 Background

The COARE bulk algorithm began evolving with the COARE Pilot cruise in 1990 (Young et al., 1992) and continued with a number of pre-COARE cruises by Australian and US research groups. This early work provided valuable insights on some mean sensor calibration problems and uncovered several false steps in the ship motion removal problem. A post-COARE deployment of the R/P FLIP off California (Fairall and Edson, 1994) proved invaluable in sorting out some of the small inconsistencies in the Environmental Technology Laboratory system (see Bradley et al., 1994, for further discussion). The results we present here will emphasize the measurements made on the R/V Moana Wave during COARE. This is because the Wave fluxes were computed in real time and were available for algorithm evaluation as soon as COARE ended. Thus, despite this long development, the results we present here must be considered to be preliminary. It is also important to realize that the Moana Wave operated primarily in a "drift" mode while deploying ocean microstructure instruments and in light winds the relative ventilation of the ship's atmospheric sensors was quite weak. Light and variable winds may lead to contamination by the ship's structure and heat island effects. Also, light relative winds greatly increase the sampling uncertainty, a particular problem for covariance fluxes (Wyngaard, 1990). These effects, plus extreme sensitivity to imperfections in the ship motion corrections, lead to the large uncertainty in 50-min covariance stress measurements noted in Table 2. Since several of the flux measuring ships in COARE took data while underway, more definitive information on this issue will soon be available.

The Moana Wave operated in the COARE intensive flux array for three cruise legs between November 1992 and February 1993. Except for two stops at the equator (at the start of leg 2 and the end of leg 3), the ship was stationed within 10 km of the Woods Hole Oceanographic Institution IMET buoy nominally located at 2 S latitude and 155 W longitude. A summary of the measurements for each leg and the average of all three is presented in Table 2. Note that the values for the turbulent fluxes used to compute these averages are taken from the covariance measurements unless they were deemed compromised by experimental conditions which interfered with the instruments or otherwise invalidated the results. Conditions that were considered included: precipitation, unfavorable relative

wind direction, sun or salt contamination of the fast humidity sensor, excessive flow tilt, and ship maneuvers. If the covariance measurement for a particular sample period was deemed invalid, the bulk value was substituted for computing the cruise average. Also, invalid periods were not used in the bulk-turbulence comparisons discussed later in this section.

The summary presented in Table 2 indicates that the November to early December period was characterized by light winds, little precipitation and clear skies; the later December through January period by stronger winds, heavy precipitation and cloudy skies; the February period was typical of the overall average. Also, the grand average of the total heat input to the ocean was 13.8 Wm^{-2}, which includes 6.4 Wm^{-2} cooling by precipitation and the Webb effect. The total heating/cooling rates for the individual legs are approximately consistent with the changes in the sea surface temperature observed.

5.2 Bulk Algorithm Evaluation

The COARE data offer a number of different possibilities for evaluating the flux algorithm. First we attempt to examine specific aspects and constants used in the algorithm. For this stage of the analysis we averaged the Moana Wave data in mean wind speed bins with 1 ms^{-1} bin width. Using the velocity-bin-average variables (of fluxes, air-sea temperature difference, and various bulk-derived variables such as R_r, R_q, u_{*t}^2, and $c_{dn}^{1/2}$, etc.), we computed various parameters of interest. For example, mean values of inertial-dissipation measurements of u_{*t}^2 were used to compute the roughness Reynolds number and compared with the bulk values yielded by (25) as used in the COARE 2.0 (see Fig. 1).

We also used covariance measurements of latent heat flux to directly estimate the neutral transfer coefficient and roughness Reynolds number, R_q, for moisture. Because the bulk-derived values of R_r, u_{*t} and $c_{dn}^{1/2}$ agree, on average, with the turbulence measurements, we used them for the velocity aspects of this analysis. Note that this involves the following steps:

a. Using the average bulk-derived values of u_{*t}^2 (to account for gustiness), q_* is computed from the mean covariance values of H_l via (1).
b. Using q_* and the sea-air humidity difference, $c_q^{1/2}$ is computed via (9).
c. Using u_{*t} and the mean covariance sensible heat flux, ξ is computed via (7).
d. Using this value for ξ, (6) is inverted to compute $c_{qn}^{1/2}$.
e. Using $c_{qn}^{1/2}$, z_{oq} is computed via (8). Using z_{oq}, the 10-m reference height value for $c_{qn}^{1/2}$ is computed.
f. R_q is computed via (10).
g. The 10-m neutral transfer coefficient, C_{en}, is computed as $c_{dn}^{1/2} * c_{qn}^{1/2}$.

Because the velocity parameter is essentially "taken out" in step a and then "put back in" in step g, it has virtually no effect of the final value of C_{en}. Thus, this result is not dependent on whether we use bulk-derived or directly measured turbulence parameters in the analysis. However, it does affect the values of R_q. We chose this approach to alleviate the highly noisy velocity turbulence measurements at low wind speeds and we are comfortable with (25) as a representation of the mean roughness of the ocean during COARE. The bin-averaged results for C_{en} are shown in Fig. 2 along with curves from the original LKB model and the Garratt/Brutsaert model. The solid curve in this

figure is the present form of the COARE 2.0 algorithm. Fig. 3 shows R_q as a function of R_r for this same data. This comparison looks quite different because the effects of the different z_o parameterizations used in the three models shown in Fig. 2 are removed. Here the change we have made in the LKB parameterization of R_q is apparent as the extended constant range for $0.16<R_r<1.0$. Note also that the small statistical scatter (±0.03) in the neutral transfer coefficient leads to rather large scatter in R_q.

The bin-averaged data have been used to estimate the Charnock constant. Here we simply take inertial-dissipation values of u_{*t} to compute z_o in the manner described above. Using (25) we can compute a value for α for each bin. These data are shown as function of R_r in Fig. 4 and are consistent with the value of 0.011 we are using. In a similar fashion, we have taken bin-averaged values of the sum of the horizontal wind component variances and computed W_*^2 from the bin-average of the sensible heat flux. The square root of the ratio of these two variables yields an estimate of β for the 50-min time scale (see Fig. 5). In the low wind speed (small R_r) regime, the value is consistently very close to 1.25. It increases slightly with wind speed outside of the convective regime, but there the gustiness effect is negligible so a slight underestimate of β will not matter.

Before going on to the flux intercomparisons it is of interest to ponder the relative importance of the processes we have attempted to represent with this approach. To do this, let us write the latent heat flux in the following form

$$H_l = \rho_a L_e [C_{en} u (q_s - q)] * stability * gustiness \qquad \textbf{(27)}$$

where

$$stability = \frac{1}{[1 - \frac{c_{dn}^{1/2}}{\kappa} \psi_u(\xi)][1 - \frac{c_{qn}^{1/2}}{a\kappa} \psi_h(\xi)]} \qquad \textbf{(28)}$$

and

$$gustiness = [1 + (\beta W_* / u)^2]^{1/2} \qquad \textbf{(29)}$$

Here in (27) the first two terms are thermodynamic constants, the term in brackets represents the small scale processes attacked in the laboratory investigations, and the last two terms represent surface-layer and boundary-layer scale processes.

In Fig. 6a we depict the balance of these processes as represented in the COARE 2.0 algorithm for the conditions found by the Moana Wave during COARE. The solid line shows the neutral transfer coefficient, the dotted line the gustiness, the circles the stability effect, and the dashed line the product of these three (as per Eq. 27). The lowest wind speed bin shown here is for speeds between 0 and 0.5 ms^{-1}. Only 11 usable 50-min samples from the Moana Wave fell in this bin, with an average wind vector magnitude of 0.33 ms^{-1} and an average latent heat flux of 32±5

Wm^{-2}. The COARE 2.0 algorithm yielded a value of 28 Wm^{-2}. Notice that gustiness and stability effects were about equal, leading to a combined enhancement of the flux of about a factor of four. We can return to the primitive days of yore by setting $\beta \approx 0$ and forcing all of the convective effects into the MOS term (Fig. 6b). Notice that this has only a modest effect until the wind speed becomes less than 2-3 ms^{-1}. Now for u=0.33 ms^{-1}, the algorithm yields a flux of 21 Wm^{-2}. However, as Godfrey and Beljaars (1992) have pointed out, we know that at sufficiently low wind speed, this version of the model must become pathological. By going to the other extreme (setting β=2.5), we can demonstrate that the gustiness and stability effects tend to be compensating. Note in Fig. (6c) that the increase in the gustiness term is significantly cancelled by a reduction in the stability term. Thus a doubling of β only leads to an increase in the latent heat flux of 46% (H_l=41 Wm^{-2}). This is expected because the increased gustiness leads to an increase in u_{*t} and therefore a reduction in the magnitude of ξ.

5.3 Flux Comparisons

The comparison of covariance and bulk latent heat flux is shown in Fig. 7a for the 876 samples that passed the various rejection criteria (relative wind direction within 90° of the bow, no ship maneuvers, ship not underway full speed, and no sun, radio, or salt contamination on the fast hygrometer). The mean covariance flux for this restricted set is 94.8 Wm^{-2}; the mean bulk flux is 93.3 Wm^{-2}. The RMS scatter is slightly less than 20%. Much of this scatter is sampling uncertainty in the covariance measurement. In Fig. 7b we compare the bulk flux with the average of the covariance and inertial-dissipation fluxes. This reduces the RMS scatter to about 12%, a truly remarkable level of consistency.

In Fig. 8a we show a comparison of bulk and covariance sensible heat values. Here we have cheated by adding data from the R/P FLIP cruise. We do this because the sea-air temperature differences in COARE were so small that it was difficult get a reasonable test of the parameterization. Also, because of the sensitivity of temperature measurements to contamination by the ship's heat island, we restricted the data to relative winds within 30° of the bow. Note the great clot of points in the region between 0 and 5 Wm^{-2}. The cloud of points scattered outside this clot are primarily caused by radiation induced errors in the air temperature. The Moana Wave system had an aspirated radiation shield but intercomparisons with the R/V Franklin showed that it was subject to errors as large as 1.5 °C in light winds. A crude correction was developed, but it still leaves some residual uncertainty. In a similar graph for just night time data (not shown) much of this outer layer of points disappears. Also, it has been demonstrated that ship motion significantly affects covariance values for sensible heat flux, so some of this scatter may be due to inadequate corrections. For the 618 values from COARE meeting the quality control criteria, the average covariance sensible heat flux was 8.2 Wm^{-2} and the average bulk value was 5.7 Wm^{-2}. If we restrict the computation to night time data only, the difference narrows to 1.5 Wm^{-2}. Whilst the scatter shown in Fig. 8a is disturbing, the overall effect on the uncertainty of the total surface energy budget is well within our limits. This is illustrated in Fig. 8b where the sensible heat flux is plotted on the same scale as the latent heat flux (Fig. 7).

A comparison of inertial-dissipation and covariance stress measurements with bulk estimations is shown in Fig. 9. The log-log format emphasizes the dynamic range of stress and the astounding consistency of the 1025 inertial-dissipation and bulk values (Fig. 9a). There are more

points here than in Figs. 7 and 8 because hygrometer contamination is irrelevant to this comparison. The covariance results (Fig. 9b) are, to put it mildly, unimpressive (covariance data are restricted to within 30 degrees of the bow and tilt in the fixed frame less than 10° for a total of 676 points). The same data is depicted in Fig. 10 in linear format. The clot of covariance points very near the origin is characteristic of fixed-site measurements of stress over the ocean, where the sampling uncertainty is considerable. However, measurements from fixed platforms and FLIP do not show the large scatter seen here, particularly at higher wind speeds (e.g., Edson et al., 1991). Whereas occasional spotty GPS reception and subtle ship maneuvers may cause some of the obvious outliers, the majority of this scatter is believed to be due primarily to inadequate ship motion corrections, particularly for the horizontal wind components, although some contributions from flow distortion cannot be ruled out. The average stress magnitude for the 1025 points is 0.031, 0.034, and 0.033 Ntm^{-2} for covariance, inertial-dissipation, and bulk methods, respectively. This comparison suggests that the average bias in the covariance stress measurements is no more than 0.002 Ntm^{-2}. Incidently, for this data the average horizontal-transverse stress was 0.005 Ntm^{-2}, indicating the stress vector is, on average, rotated about 10° relative to the wind vector.

5.4 Wind Speed Dependence of the Latent Heat Flux

The mean transfer coefficient comparisons of Fig. 2 and the point-by-point flux comparisons of Fig. 6 give most of the picture, but the importance of latent heat flux makes a look at the mean wind speed dependence worthwhile. Using the velocity binning method described earlier, we have computed the mean latent heat flux with the restrictions corresponding to Fig. 7a. The mean wind speed dependence for bulk, covariance, and inertial-dissipation data is shown in Fig. 11a. Except for an overestimate of the flux in the convective limit by the inertial-dissipation method, the agreement is quite close (as one would expect since we adjusted the R_q parameterization to fit this data). The same data is shown in Fig. 11b for the Garratt/Brutaert bulk model, which tends to fall slightly below the data at low wind speeds and above the data at high wind speeds (the crossover point is at about 7 ms^{-1}). The average Garratt/Brutsaert bulk flux is about 7 Wm^{-2} less than the average covariance value. Finally, to get a better look at the higher wind speed region, we relax the salt contamination restrictions on the fast hygrometer data. This adds about 200 points to the comparison, mostly at wind speeds above 6 ms^{-1} (see Fig. 11c). Because the salt contamination has a first-order effect on the measured humidity variance spectrum, the inertial-dissipation values for the highest wind speeds fall significantly above the other two. There is slight reshuffling of the relative positions of the covariance and bulk points at higher wind speeds, but basically the comparison is unchanged.

6. Conclusions

7. Bibliography

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Table 1. TOGA COARE R/V Moana Wave measurement accuracies as estimated in Bradley et al. (1994)			
Variable	Units	50-min RMS	Rls 2.0 Bias
u	ms^{-1}	0.3	±0.2
T_a, day	K	0.3	±0.2
T_a, night	K	0.2	±0.1
q	gkg^{-1}	0.3	±0.2
T_s	K	0.1	±0.2
q_s	gkg^{-1}	0.1	±0.2
H_s, cov	Wm^{-2}	3 ±20%	±2
H_l, cov	Wm^{-2}	5 ±20%	±4
τ, cov	Ntm^{-2}	0.015 ±30%	0.002
$\|\tau\|$, ID	Ntm^{-2}	15%	0.002
R_s	Wm^{-2}	1%	±4
R_l	Wm^{-2}	5	±2.5
Precip	mmh^{-1}	15%	?

BIAS = Average sensor - Average correct value

Table 2. Data summary from the Moana Wave during COARE.				
Variable	Leg 1	Leg 2	Leg 3	Total
Dates	11/11-12/3	12/16-1/11	1/28-2/16	11/11-2/16
# 50-min	589	648	385	1622
u, ms^{-1}	3.40	5.49	4.77	4.56
T_s, ∘C	29.45	28.98	29.11	29.18
T, ∘C	27.96	27.38	27.56	27.63
q, gkg^{-1}	18.06	18.08	18.06	18.07
q_s, gkg^{-1}	25.23	24.53	24.72	24.83
H_s, Wm^{-2}	6.6	10.9	9.6	9.1
H_l, Wm^{-2}	92.7	119.0	114.7	108.5
H_{lW}, Wm^{-2}	3.2	4.3	4.1	3.9
H_{rain},Wm^{-2}	0.9	3.8	2.6	2.5
$R_{s\downarrow}$, Wm^{-2}	235.4	176.3	200.9	203.6
$R_{l\downarrow}$, Wm^{-2}	408.3	417.5	412.6	413.0
R_{net}, Wm^{-2}	160.9	115.9	134.0	136.5
R_{net}-H_l-H_s	60.9	-11.1	10.3	20.2
H_{tot}, Wm^{-2}	56.8	-19.3	3.6	13.8
Rain, mmd^{-1}	4.0	18.5	11.8	11.6
τ, Ntm^{-2}	0.030	0.055	0.040	0.040
u^2, $(ms^{-1})^2$	14.8	40.7	27.1	28.1

Figure Captions

Fig. 1. Roughness Reynolds number, R_r, as a function of the 10-m wind speed for the Moana Wave COARE data. The data have been averaged in wind speed bins with 1 ms^{-1} bin widths. The solid line is the COARE 2.0 alogrithm result and the x's are computed from the mean inertial-dissipation values for stress.

Fig. 2. Neutral stability values for the 10-m moisture transfer coefficient, C_{en}, as a function of 10-m wind speed bin-averaged as in Fig. 1. The solid line is the COARE 2.0 model, the dotted line the Garratt/Brutsaert model, and the dashed line the original LKB model (see text). The crosses are derived from the Moana Wave covariance latent heat flux measurements.

Fig. 3. Moisture rouhness Reynolds number, R_q, versus R_r derived from the fluxes averaged in wind speed bins as in Fig. 1. The solid line is the COARE 2.0 model, the dotted line the Garratt/Brutsaert model, and the dashed line the original LKB model (see text). The crosses are derived from the Moana Wave covariance latent heat flux measurements.

Fig. 4. Charnock constant, α, versus R_r derived from the fluxes averaged in wind speed bins as in Fig. 1. The dashed line is the value of 0.011 used in the COARE 2.0 model and the crosses are derived from the Moana Wave inertial-dissipation stress measurements.

Fig. 5. Gustiness constant, β, versus R_r derived from the horizontal velocity variances and the sensible heat flux (used to compute convective scaling velocity) averaged in wind speed bins as in Fig. 1. The dashed line is the value of 1.25 used in the COARE 2.0 model and the crosses are derived from the Moana Wave measurements.

Fig. 6. Various components of the COARE 2.0 moisture transfer coefficient as a function of wind speed for the COARE data. The solid line is C_{en} multiplied by 1000; the dotted line the gustiness parameter (eq. 29); the circles are the stability parameter (eq. 28); and the dashed line is stability*gustiness*C_{en}*1000.

a. For β=1.25.
b. For β=0.1.
c. For β=2.5.

Fig. 7. Moana Wave turbulence latent heat flux data versus COARE 2.0 bulk values:

a. The vertical axis is covariance data.
b. The veritcal axis is the average of covariance and inertial-dissipation data.

Fig. 8. Moana Wave and FLIP covariance sensible heat flux data versus COARE 2.0 bulk

values:

a. Expanded scale.
b. The same scale as Fig. 7.

Fig. 9. Log-Log plot of Moana Wave turbulence stress data versus COARE 2.0 bulk values:

a. The vertical axis is inertial-dissipation turbulence data.
b. The vertical axis is covariance turbulence data.

Fig. 10. As in Fig. 9 but for linear-linear axes.

Fig. 11. Moana Wave latent heat flux versus wind speed, bin averaged as in Fig. 1. The solid line is the bulk value, the cross the inertial-dissipation value and the circle the covariance value.

a. Data for winds within 90 degress of the bow, no ship maneuvers, and no contamination of the fast hygrometer. Bulk model is COARE 2.0.

b. As in a, but for the Garratt/Brutsaert bulk model.

c. As in a, but relaxing the fast hygrometer contamination restriction.