OBSERVATIONS OF OCEAN SURFACE FLUXES: MEANS AND VARIABILITY

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Abstract: The methods for calculating air-sea heat and momentum fluxes from ship data are described. Systematic errors must be corrected using information on observing techniques. The random errors must be reduced by suitable averaging; their magnitude can be assessed using the "semivariogram" technique. For the SOC climatology (Josey et al. 1997a) we have evaluated the calculated flux fields in terms of the global heat balance, the implied meridional heat transport in the ocean, and by comparison with buoys. The SOC flux estimates were in better agreement with the buoy data than were the fluxes from a sample of numerical models.

1. INTRODUCTION

The air-sea fluxes, the transfers of heat and water, and of momentum, between the sea and the air, are an important component of the earth's climate system. Various studies have provided estimates of the global distribution of these fluxes (e.g. Budyko, 1963; Esbensen and Kushnir, 1981; Hsiung, 1986; Oberhuber, 1988; da Silva et al., 1994). These, and the "SOC climatology" recently calculated by ourselves (Josey et al. 1996, 1997a), are all based on the weather observations provided by merchant ships participating in the Voluntary Observing Ship system of the World Weather Watch. The attraction of these merchant ship observations is that they constitute the only quasi-global data set which extends over a number of decades. Also, the range of observations is such that it has been possible to develop parametrisation formulae with which to calculate all the required fluxes. The major part of this paper will consider the method of calculation (Section 2), the accuracy (Section 3) and the verification (Section 4) for flux fields constructed from ship observations. We will suggest that, despite the many problems, in many cases the ship observations are better than model derived flux estimates. However the geographical distribution and data quality of ship observations is less than optimal, and we will briefly consider the potential of using data from satellites to improve the flux fields (Section 5).

The total heat transfer into the ocean through the air sea interface, F_{tot}, is the combination of several flux components. The source of energy for the ocean-atmosphere system is the heat from the sun, that is the incoming shortwave radiation (0.3 to 3μm wavelength). Part of this flux (around 6% depending on the solar elevation) is reflected from the sea surface. The remainder, the net shortwave flux, F_{SW}, is absorbed by, and warms, the ocean. Both the sea and the sky emit and absorb long wave (3 to 50μm wavelength) radiative energy. Because the effective radiative temperature of the sky is generally colder than that of the sea, the upward longwave flux is greater than the downward flux, and the net longwave flux F_{LW}, normally cools the ocean.

Because the sea and air which is in contact with the sea surface are at different temperatures, viscous and turbulent processes result in a sensible heat flux, F_H. The latent heat flux F_E is the heat transferred due to the evaporation (or condensation) E, of water vapour. We will refer to F_H and F_E as the "turbulent heat fluxes". They may transfer heat in either direction but more usually, compared to the sea, the air is colder and "drier" (see section 2.4.1 below) and this results in cooling of the sea surface. Thus:

$$F_{tot} = F_{SW} - (F_{LW} + F_H + F_E) \tag{1}$$

The mean freshwater flux into the atmosphere is the difference between evaporation, E, and precipitation, P. Most of the freshwater that evaporates from the ocean falls back into the ocean as precipitation. Nevertheless, the 10% which falls over land represents about one third of the terrestrial precipitation (Schmitt, 1994).

The wind stress, τ, is the downward transfer of horizontal momentum through the sea surface. In contrast to land surfaces, the roughness elements which cause the wind stress, the sea surface waves, depend themselves on the magnitude and history of the wind velocity.

There are other, potentially important, fluxes which we shall not consider in this paper. For example trace gas fluxes such as Carbon Dioxide and Dimethyl Sulphide which may directly or indirectly effect the earth's radiation budget. Because the sea is essentially an infinite supply of water vapour, variations in evaporation are largely controlled by atmospheric effects. In contrast, the trace gas fluxes are controlled by the varying sources and sinks of those substances within the ocean. Because these variations occur on relatively small space and time scales, accurate quantification of trace gas fluxes remains a challenge for future research.

2. CALCULATING FLUXES FROM IN SITU DATA

2.1 Introduction - "Classical" and "Sampling" methods of flux calculation

Two methods have generally been used in calculating mean flux values defining the "classical" or the "sampling" means (Esbensen and Reynolds, 1981). The sampling mean is calculated from individual flux estimates each derived from a single weather report. The classical mean is derived by first averaging the observed variables and then calculating the fluxes. The latter method allows the use of more data since all the variables need not be measured simultaneously; it also simplifies the calculation procedures.

Many early flux climatology studies were based on a compilation of marine weather reports assembled at the US National Climate Center (NCC). At great effort, Bunker (1976) and his co-workers used the individual ship reports to calculate sampling means for the fluxes over the Atlantic Ocean. Bunker's area mean values were later used by Isemer and Hasse (1987) to recalculate the Atlantic fluxes using more recent formulae. Other studies (Hastenrath and Lamb, 1978, Esbensen and Kushnir, 1981, Hsiung, 1986) presented classical-mean values based on monthly mean area averaged values prepared by the NCC and other centres. Nowadays, for climate studies, the in situ data have been organised into the Comprehensive Ocean-Atmosphere Data Set (Woodruff et al., 1987). COADS contains observations dating back to the year 1854; although the number of available reports increases substantially after the 1950's. Data are available from COADS as monthly summaries or as the individual reports. Oberhuber (1988) calculated fluxes based on the monthly summaries. However recent studies (e.g. Josey et al., 1995) have suggested that classical mean fluxes may be biased by up to 10% due to correlations between the observed variables. Thus da Silva et al. (1994) present sampling mean fluxes and, as will be described below, in calculating the Josey et al. (1997a) flux climatology we have attempted to correct individual ship observations for biases arising from the observing technique before calculating sampling mean fluxes.

2.2 Sources of in situ observations

2.2.1 Buoys

Buoy data does not allow global estimation of the surface fluxes; however buoys are a good source of verification data for other observing systems. Drifting buoys normally measure a too restricted range of variables to allow flux estimation. Typically, sea temperature, air pressure and possibly wind speed may be available. Moored operational weather buoys normally provide information on wind velocity,

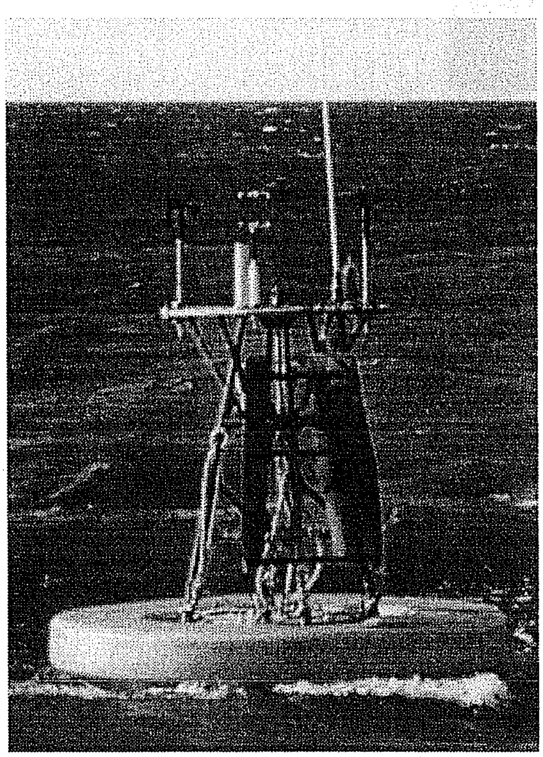

Figure 1. The SOC "Sonic Buoy" carries a Sonic anemometer to provide estimates of the wind stress. Such buoys provide good instrument exposure and are capable of accurate measurements of air-sea interface variables.

air and sea temperature, and humidity; thus allowing estimation of the turbulent fluxes but not the radiative fluxes. With the major exception of the "TAO" buoy array established by the TOGA experiment in the equatorial Pacific (Hayes, et al. 1991), most moored buoys are coastal. Buoys such as the IMET mooring (Hosom et al. 1995) allow all the fluxes to be estimated. Such moorings are expensive but, provided the instruments are well calibrated, they provide an invaluable source of flux verification data (Section 4.3.4). Equipped with fast response sensors, research buoys (Figure 1) can also be used to determine the transfer coefficients (Section 2.4.2)

2.2.2 Ships

Kent and Taylor (1991) described 46 Voluntary Observing Ships (VOS) operating in the North Atlantic, and the meteorological instrumentation with which they were equipped. Typical of this subset would be a container vessel of about 210m length, travelling at about 9 m/s, and loaded with cargo to about 10m to 20m above the main deck (Figure 2). As an instrument platform, this represented a large

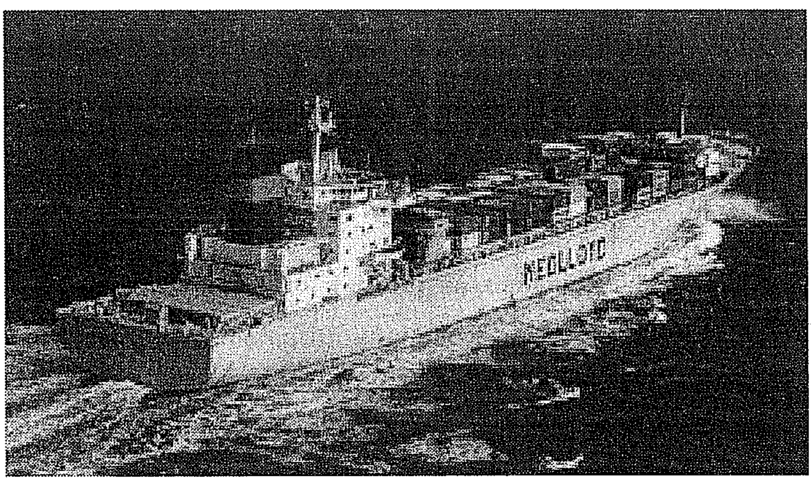

Figure 2. A typical Voluntary Observing Ship. Instrument exposure on many ships is far from ideal.

object capable of significantly altering the environment to which the meteorological sensors are exposed. Also, given that in total there are over 7000 VOS and many ships frequently are sold or go out of service for periods, it is perhaps not surprising that relatively simple meteorological instrumentation was used. Sea temperatures were measured near the surface using a SST bucket, or at depths of 3m to 9m using engine intake or hull contact sensors. Air temperature and humidity observations were taken at about 20m to 30m above the sea using a thermometer screen or hand held psychrometer. If carried, the anemometer was at about 30m to 35m, but two-thirds of the ships reported visually estimated winds.

A major disadvantage of the ship observations is that adequate coverage is obtained only in areas that are well covered by shipping lanes. For example, in July there are very few observations which enable a latent heat flux estimate to be made south of 40°S (Figure 3). In contrast the North Atlantic and North Pacific are well covered by observations. The effect on the error distributions will be discussed further in Section 3.3.3.

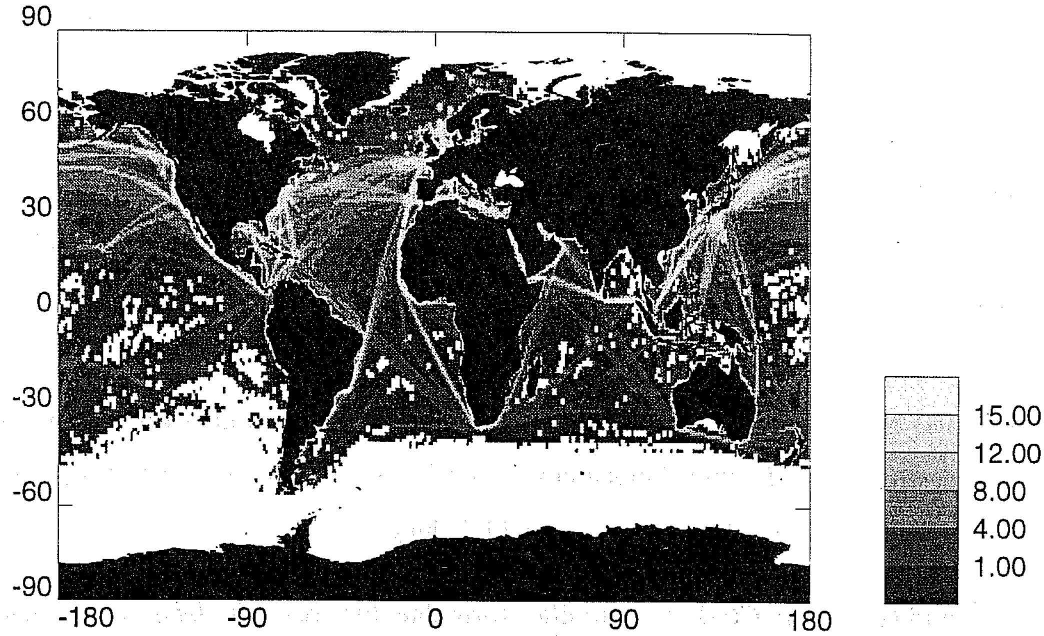

Figure 3. Mean number of latent heat flux estimates in each 1°latitude x 1° longitude cell during the month of July averaged over the 14 years used in the SOC climatology.

2.3 Estimating the radiative fluxes from ship observations

2.3.1 Shortwave

Merchant ships are not equipped to measure the incoming shortwave radiation; it must be estimated from information on the ship's position and the amount of clouds as visually estimated by the ship's officer. However for this flux there is a further potential problem; most ship's report at most 4 times per day and the most reports occur at the standard reporting hour closest to local noon. To avoid introducing bias, the Reed (1977) formulae for the daily mean net shortwave flux F_{sw} uses the calculated monthly mean values for the fractional cloud cover n, and the noon solar elevation in degrees, ϕ. Hence:

$$F_{sw} = F_{sw_0}(1 - c_n n + 0.0019\phi)(1 - \alpha) \tag{2}$$

where $c_n(= 0.62)$ the cloud attenuation factor, and α the albedo (Payne 1972). F_{sw_0} is the shortwave insolation at the surface under clear skies (Seckel and Beaudry 1973), The Reed formula (2) should only be used for $0.3 \le n \le 1$, and $F_{sw} = F_{sw_0}$ should be used for $n < 0.3$ (Gilman and Garrett, 1994) .

Being based only on the mean monthly estimated cloud amount, equation (2) is a crude method of estimation. However it was recommended by the comparative studies of Frouin et al. (1988) and

Dobson and Smith (1988) who found a site dependant long term bias of -1 to +12 Wm^{-2} and a monthly mean rms error of about 8 Wm^{-2}. In contrast, using comparisons with data from several air-sea interaction experiments, Katsaros (1990) found the Reed formula to be biased high by about $20Wm^{-2}$.

2.3.2 Longwave

As for short wave radiation, the incoming longwave radiation is not measured but must be estimated from the visually estimated cloud amount. Other parameters are the air temperature and humidity and, for determining the net longwave flux, the sea temperature. For example (Clark et al., 1974): :

$$F_{LW} = \varepsilon\sigma_{SB}T_S^4(0.39 - 0.05e^{0.5})(1 - \lambda n^2) + 4\varepsilon\sigma_{SB}T_S^3(T_s - T_a) \qquad (3)$$

where ε (= 0.98) is the emittance, σ_{SB}(= 5.7 x 10^{-8} $Wm^{-2}K^{-4}$) the Stefan-Boltzmann constant, e the water vapour pressure, n the fractional cloud cover, and T_a and T_s are the air and sea temperatures in degrees K. The cloud cover coefficient λ varies with latitude.

The global validity of the Clark and similar formulae has recently been questioned. For the Mediterranean, Gilman and Garrett (1994) found that this class of formulae underestimates F_{LW} by about $17Wm^{-2}$ and, based on Mediterranean data, Bignami et al. (1995) proposed a new formulae which, for mid-latitude regions, gives F_{LW} values some 25 Wm^{-2} greater than (3). In contrast the review of Katsaros (1990) found that (3) underestimated the mean F_{LW} by only 5.5 Wm^{-2} and comparisons with ship measurements from the Atlantic and Southern Ocean (Josey et al. 1997b) have shown agreement typically to within 5 Wm^{-2} on timescales of order one month. It may well be that the atmosphere over the Mediterranean is on average different from the open ocean (certainly there are less clouds!) and that a different formula is needed in that region.

2.4 Estimating the turbulent heat fluxes and wind stress

2.4.1 The bulk formulae

The sensible heat (F_H), latent heat (F_E), and wind stress (τ) are calculated from the meteorological observations using the "bulk formulae". These formulae can be derived by using dimensional analysis and boundary layer similarity theory to determine the relationship between the surface flux of a quantity and its vertical gradient, and then integrating between the surface and the observation height (e.g. Geernaert, 1990a). Thus:

$$\tau = \rho C_D(U - U_0)^2 \qquad (4a)$$
$$F_H = \rho C_H(U - U_0)((T_a - \gamma z) - T_s) \qquad (4b)$$
$$F_E = \rho C_E(U - U_0)(q_a - q_s) \qquad (4c)$$

where ρ is the air density, U, T_a, and q_a are the observed wind speed, air temperature, and specific humidity, and U_0, T_s, and q_s are the corresponding values at the sea surface (q_s is the 98% saturation specific humidity at T_s). The air temperature must be corrected for the adiabatic lapse rate, γ, between the surface and the observation height z. Solving the bulk formulae requires iteration because the transfer coefficients, C_D, C_H, and C_E depend on the stability and hence on the fluxes. As an example, the Dalton number, C_E, may be defined by:

$$C_E = k^2 \left(\ln\left(\frac{z_q}{z_{0q}} \right) - \Psi_q \right)^{-1} \left(\ln\left(\frac{z_u}{z_0} \right) - \Psi_m \right)^{-1} \tag{5}$$

where (for humidity and momentum respectively) z_q, z_u are the measurement heights, and z_{0_q}, z_0 are roughness lengths. The stratification functions Ψ_q and Ψ_m (Paulson, 1970) are functions of the stability (defined by the ratio of the observation height to the Monin-Obukhov length. Under neutral conditions $\Psi_q, \Psi_m = 0$, and (5) then defines the 10m neutral value of the transfer coefficient, C_{E10n}. The variation of C_E with wind speed and stability is shown in Figure 4.

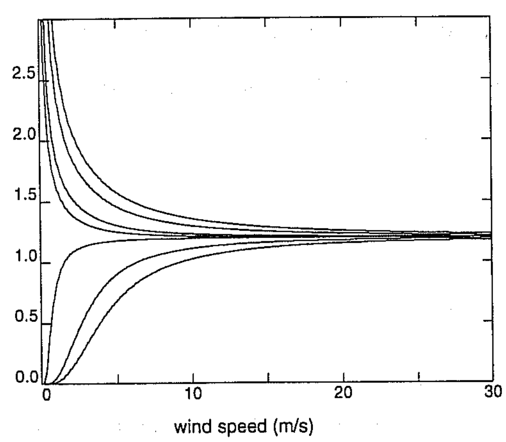

Figure 4. The transfer coefficient for Latent heat (Dalton number) as a function of wind speed and air minus sea temperature difference (bottom to top: 5°C, 1°C, 0.2°C, -0.2°C, -1°C,-5°C,>> -5°C).

2.4.2 *Determining the transfer coefficients*

The transfer coefficients are determined by air-sea interaction experiments in which the flux is measured by the eddy correlation or inertial dissipation method combined with accurate observations of the mean meteorological variables (see, for example, Smith, 1988, 1989, Geernaert 1990b).

The effects of salt spray and other instrumentation problems have limited the number of reliable estimates of C_{E10n} and C_{H10n} over the ocean and there is still at least 10% uncertainty in their values and hence in the derived fluxes. Based on a review of previous studies, Smith (1989) suggested a constant Dalton number ($10^3 C_{E10n} = 1.2 \pm 0.1$) for winds between 4 and 14 m/s. The Humidity EXchange Over the Sea (HEXOS) experiment results (DeCosmo et al., 1996) also suggest a near constant value (any increase being less than 15 to 20%) with ($10^3 C_{E10n} = 1.12 \pm 0.24$) for winds up to 18m/s. For the

Stanton number Smith (1988) suggested ($10^3 C_{H10n} = 1.0$) which is consistent with the ratio of the thermal to species diffusivities (Prandtl/Schmidt numbers), $C_E / C_H = 1.16$, (Friehe and Schmitt 1976), In contrast the HEXOS results (DeCosmo et al., 1995) suggest ($C_E \approx C_H$) to the accuracy of the determination. At wind speeds below about 2 m/s both C_{E10n} and C_{H10n} increase (Bradley et al., 1991).

Compared to the Dalton and Stanton numbers, there have been many more estimates of the open ocean drag coefficient, C_{D10n}. Recently Yelland et al. (1997) reanalysed a large data set from the Southern Ocean (Yelland & Taylor, 1996) taking into account the effects of wind flow distortion around the research ship. Their results (Figure 5) were not significantly different from the formula proposed by Smith (1980):

$$10^3 C_{D10n} = 0.61 + 0.063 U_{10n} \tag{6}$$

Figure 5 also illustrates the significant scatter which is typical of most determinations of C_{D10n}. Yelland et al. (1997) suggest that this scatter can be explained by random measurement errors. A correlation with the sea state could not be found and the C_{D10n} anomalies did not appear to be coherent over periods of an hour or more as might be expected if they were caused by the changing wave field. The important inference is that, for climatological purposes, the mean wind stress can be estimated from the mean wind without any requirement to know the state of development of the surface wave field. However other papers in this seminar will present an opposing argument (Janssen,1977).

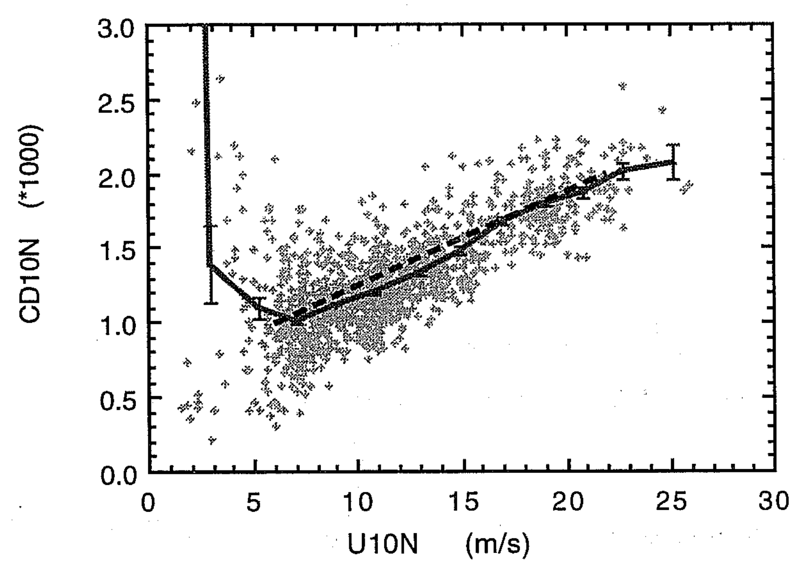

Figure 5. Individual and mean values for the drag coefficient as a function of 10m neutral wind speed as measured from the RRS Discovery in the Southern Ocean (Yelland et al. 1997). Also shown is the Smith (1980) formula (dashed line).

As the wind decreases below about 2m/s, Smith (1988) suggested that C_{D10n} would increase due to viscous effects. However, the significantly greater C_{D10n} values for winds below about 5m/s predicted by Wu (1994) appear to be a better fit to the data of Bradley et al. (1991) and Yelland and Taylor

(1996). While the exact form of C_{D10n} at low wind speeds does not significantly change the value of τ (because U_{10n}^2 is small) it may be important for estimating the stability and hence the correct values of C_E and C_H for estimating the heat fluxes in regions such as the western Pacific warm pool.

2.4.3 The TOGA COARE parametrisation

As a result of the TOGA COARE experiment, which took place in the western tropical Pacific (Webster & Lukas, 1992), a new algorithm for the turbulent fluxes has been developed (Fairall et al., 1996). This algorithm differs from or extends equation (4) in a number of ways. The values of C_E and C_H follow the theoretical model of Liu et al. (1979) with the C_{D10n} value defined as the sum of a Charnock formula and a smooth flow limit (Smith 1988), The Monin-Obukhov dimensionless profile functions were given a form that asymptotically approached the proper convective limit as wind speed goes to zero. The wind speed in the bulk equation was augmented by a gustiness velocity, proportional to the convective scaling velocity, W*. The Webb et al. (1980) correction to the latent heat flux, the sensible heat flux due to rainfall (Gosnell et al., 1994), and the rain momentum flux are all calculated within the algorithm.

It was recognised that, although the sea temperature used in equation (4) is normally measured perhaps a few metres below the surface, it is the surface skin temperature that physically determines the values of the turbulent and longwave heat fluxes. Typically those fluxes result in a surface skin which is a few tenths colder than the bulk SST measurement. This cool skin effect was estimated using a modified Saunders (1967) model. A different effect occurs for light wind conditions with high solar radiation, when a diurnal warm layer may form which is too shallow to be detected by bulk SST measurements. This was modelled using a simplified scaling version of the Price et al. (1986) mixed layer model. Warming as great as 4 K may be produced with a warm layer depth of about 0.25 m (Fairall et al., 1995). The depth and intensity of this warm layer depend on the integrated effects of the surface fluxes. Thus, this part of the TOGA algorithm requires a complete time series of data throughout the diurnal cycle and can not be applied to the intermittent data obtained from VOS. In Section (4.3.4) below we will compare fluxes calculated using the TOGA algorithm to fluxes obtained from the basic form of equation (4).

2.5 Estimating precipitation from ship observations

The precipitation is not measured on ships but is determined from the "present weather" descriptions reported in the ships' weather observations (Tucker, 1961). Weather codes which indicate that precipitation is occurring are modelled by formulae which describe the precipitation rate based on the contribution from three continuous rain categories: light, moderate, or heavy. The relationship between these categories, and the rain rate which they represent, was found empirically. Dorman and Bourke (1978) adapted Tucker's scheme for tropical regions by taking temperature into account. The accuracy of all these "present-weather" based precipitation estimates is not likely to be high.

3. THE ACCURACY OF GLOBAL FLUX FIELD ESTIMATES

3.1 Introduction

Although many flux climatologies have been presented there are few estimates of the accuracy of the calculated fields. Perhaps the most thorough study to date is that of Gleckler and Weare (1995) who took as an example the Oberhuber climatology and calculated uncertainties for 2° latitude x 2° longitude regions. They suggested that random uncertainties (which can be reduced by averaging over larger areas) ranged from 5 to 15 Wm^{-2} in the northern oceans to 25 to 50 Wm^{-2} in the tropics and subtropics. Systematic uncertainties (for example due to uncertainties in the transfer coefficients) were between 15 and 30 Wm^{-2} in the northern oceans and between 30 to 45 Wm^{-2} elsewhere. In the western boundary currents they were in excess of 45 Wm^{-2}.

While some of the assumptions used by Gleckler and Weare may be questionable, that they found large possible errors is not surprising. Examination of Equation (4) shows that, if the heat fluxes are to be obtained to, say, 10 Wm^{-2} then, the required order of accuracy for the observations is about ±0.2°C for the SST, dry and wet bulb temperatures (or about 0.3 g/kg for specific humidity) and the winds should be estimated to ±10% or better, say about 0.5 m/s (see, for example, Taylor, 1984, 1985). These are stringent requirements which will rarely be met by an individual VOS observation. In calculating a climatology, the aim must be to remove systematic errors in the data set and ensure that enough observations are averaged to reduce the errors to the required level. In this section we will describe how we are attempting to assess the accuracy of the SOC flux climatology.

3.2 Systematic errors

3.2.1 The VSOP-NA programme

The VSOP-NA (Voluntary Observing Ship Special Observing Programme - North Atlantic) project (Kent et al., 1993a) was designed to quantify errors in the VOS data. A subset of 46 VOS was chosen, the instrumentation used on each of the participating ships documented (Kent & Taylor, 1991), and extra information was obtained with each report, for example the relative wind at the time of observation. The output from an atmospheric forecast model was used to compare one ship observation against another. The results were then analysed according to instrument type and exposure, ship size and nationality, and other factors.

Bias errors identified by the VSOP-NA were: that SST values from engine intake thermometers were biased high (by 0.35°C); that daytime air temperatures were too warm, the bias being a function of F_{sw} and the relative wind (Kent et al. 1993b); that the dew point temperature was not biased by the air temperature error (Kent and Taylor, 1996) but, compared to aspirated psychrometer readings, was biased high when obtained from fixed thermometer screens. The results suggested that correction of measured winds to allow for the height of the anemometer was advantageous. For visually estimated winds later work by Kent and Taylor (1997) suggested that the form of the Beaufort scale devised by Lindau (1995) gave best agreement with the anemometer winds.

3.2.2 Application of the VSOP-NA corrections - the WMO47 metadata

The COADS data do not contain all the information needed to apply the VSOP-NA corrections. For example, the method of SST measurement is only included in COADS for log book reports, and is often absent. The method of humidity measurement is not included. The additional source of such information is the International List of Selected, Supplementary and Auxiliary Ships ("WMO 47", e.g. WMO, 1990) which is published annually by the World Meteorological Organisation (WMO). The details of the instrumentation carried on most of the 7000 plus Voluntary Observing Ships include the method of SST, air temperature and humidity measurement, whether the ship carried an anemometer and if so, the height of the anemometer. To use these data it is necessary to merge the COADS and WMO47 datasets by using the ships call sign, Figure 6 shows a time series of the number of successfully matched reports for the SOC climatology. The composition of COADS can be seen to change with time, the ship reports dominating in 1980 and the buoy reports dominating by 1993 (the number of ship reports in the later period may increase as more log book data becomes available for the later years). The success rate in matching the COADS and WMO47 data increased with time as the information contained in WMO47 became more comprehensive.

	North Pacific (30° to 50°N, 180° to 150°W)			North Atlantic (30° to 50°N, 40° to 20°W)		
Year	Mean Height (m)	Standard deviation (m)	Fraction (%)	Mean Height (m)	Standard deviation (m)	Fraction (%)
1980	28.7	5.9	69	18.4	7.3	35
1986	33.7	6.4	81	21.5	8.9	44
1990	35.2	8.4	82	24.2	10.9	38

Table 1. Mean and standard deviation of the distribution of anemometer heights during January of the years indicated for the North Pacific and the North Atlantic. Also shown is the fraction of wind observations which were measured by anemometer.

The merged data set showed that the mean height at which anemometers were mounted varied both regionally and with time (Table 1), with a trend to increasing heights, especially in the North Pacific.

Anemometer height corrections changed the monthly mean wind speeds by about 1 m/s in the northern oceans in winter.

The impact of the various data corrections on the calculated net ocean to atmosphere heat flux varied from area to area by about ±15 Wm^{-2}. Fluxes were decreased in the northern oceans in winter (mainly due to wind and sea temperature correction) and increased in the Tropical oceans (mainly due to the temperature and humidity corrections (Josey et al. 1997a)).

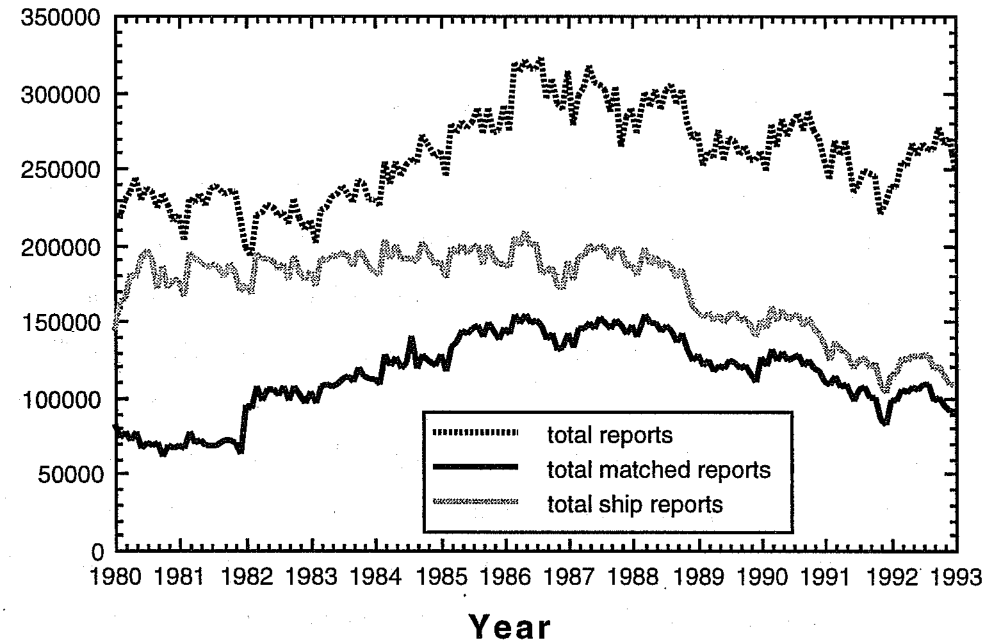

Figure 6. A time series of the number of reports used in the SOC climatology for which matching WMO47 data was available. The top dotted line shows the total number of reports in each month. The centre (grey) line shows the number of reports from ships (rather than buoys, fixed platforms, etc.). The lower line shows the number of these ship reports matched.

3.3 Random errors

3.3.1 Position errors

In preparing to calculate the SOC climatology it was found that about 2 to 3% of the VOS weather reports in COADS could be identified as having incorrect position information . Typically the position is incorrect by 10° or is in the wrong quadrant. Often these data exist in COADS as a duplicates, one report having the correct position (Launder and Morrissey, 1987). Position errors are potentially serious because the error may place the ship away from the shipping lanes in a data sparse region where the erroneous report may be given undue weight. For example in January, 1984, ship reports from near Iceland appeared as erroneous duplicates near Antarctica.

3.3.2 *Random errors for a single observation*

The random errors for ship observations of a given variable can be determined (following Lindau, 1995) by regressing the square of the difference between pairs of ship observations as a function of their spatial separation. The gradient of this regression relates to the spatial variability of the variable, the intercept gives the scatter for a single observation (Figure 7). This method is common in geological analysis where a 'semivariogram' is used to analyse correlated spatial data such as that from boreholes. Kent et al. (1997a) have analysed samples of the COADS data set for different ocean regions and defined RMS error values. For example for wind speed (Figure 8a) the standard deviation of a single observation varies from 1.7 m/s (ITCZ in July) to 2.5 m/s or greater (Gulf stream, North Pacific, and Arctic in January).

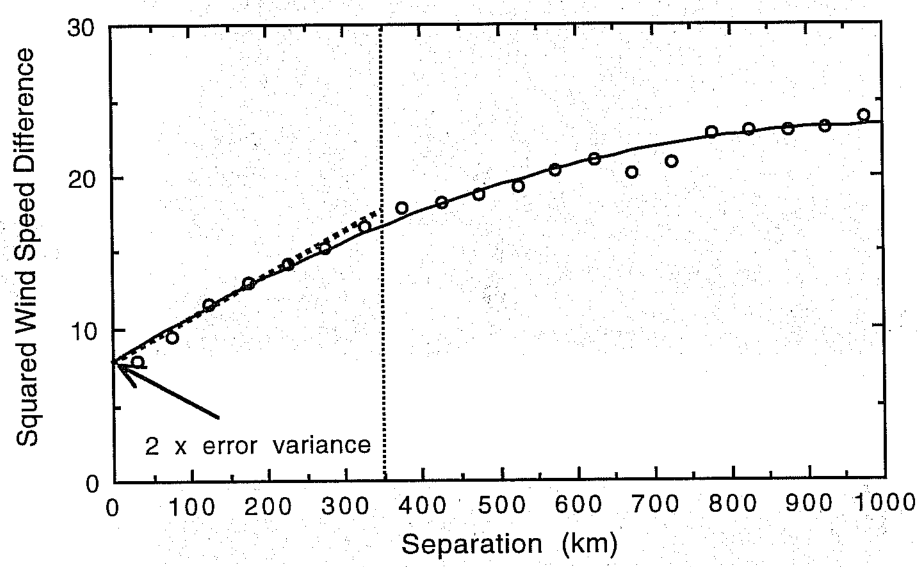

Figure 7. Example of a semivariogram for wind speed observations. The intercept is a measure of the error variance for a single observation.

3.3.3 *Accuracy of mean estimates*

Once the RMS for a single observation is known, the accuracy of the basic 1° latitude x 1° longitude mean values can be calculated from the known density of observations. For example, for wind speed the random errors are reduced in the regions of the North Atlantic to less than 0.8 m/s (Figure 8b). However for areas like the Southern Ocean with few observations, the RMS error remains high.

The final stage is to determine the effect of the objective analysis method used to interpolate and smooth the 1° mean values. For the SOC climatology a successive correction scheme of the type developed by Cressman (1959) and Barnes (1964) was applied using influence radii of 1541, 1211, 771 and 331 km on four successive passes and a zonal mean background field (Josey, 1996). A similar method has been used by Levitus (1982), da Silva et al. (1994), and others. The effect of this scheme can be estimated by using it to analyse artificial data consisting of fields of white noise (with or without varying fractions of simulated missing data). Use of a semivariogram analysis shows that, whereas the random field has constant variability at all scales, for the smoothed field the variance increases with separation for distances up to 500 km or more; information from each observation has been spread over that distance. This implies that the error estimates (shown for example in Figure 8b) should be reduced, perhaps by a factor of around 4.

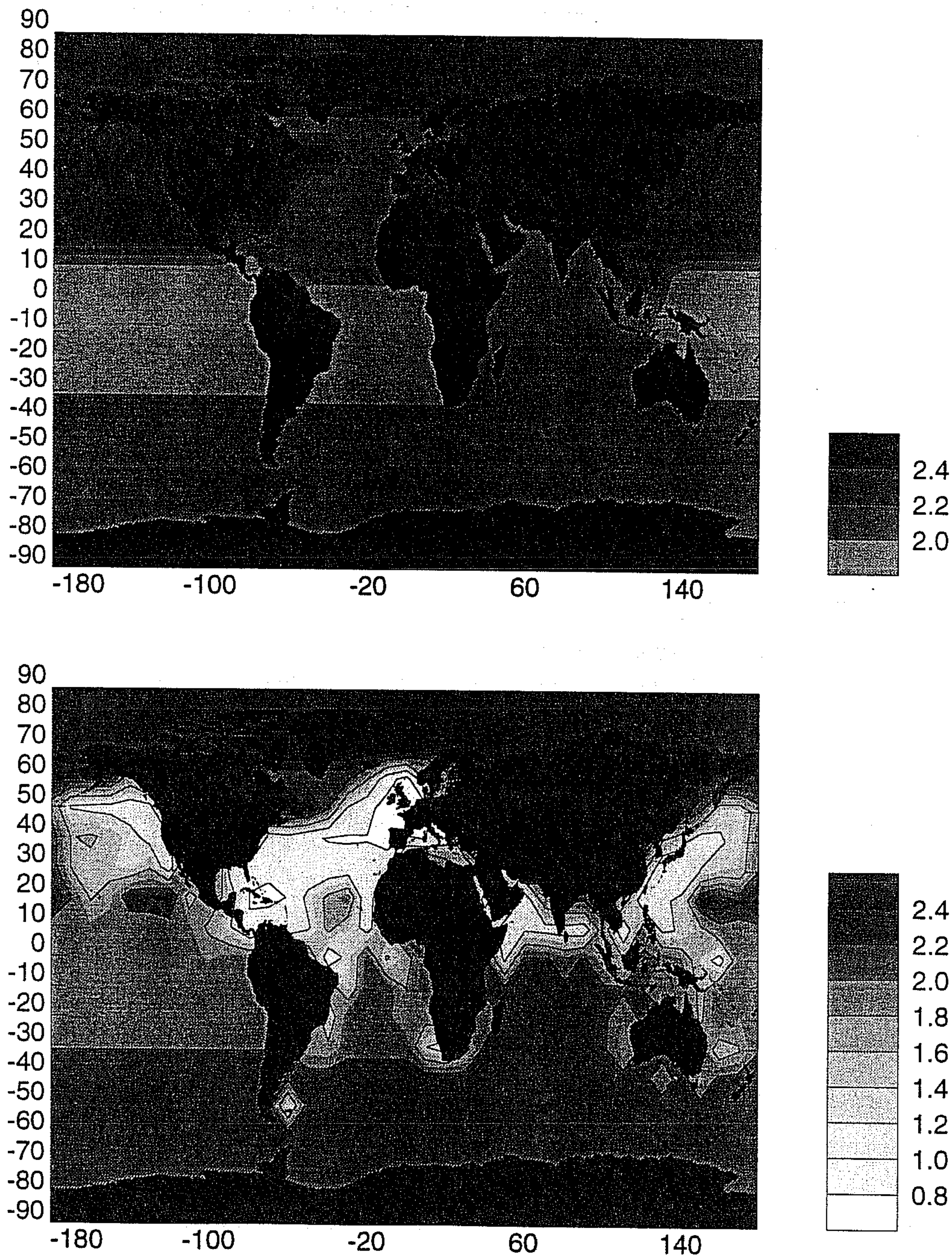

Figure 8 RMS error for wind speed in January (from Kent et al. 1997). (a) The value for a sampling density of 1 observation in each 1° grid cell has been calculated for a number of discrete ocean areas. (b) The expected error for each 1° grid cell given the number of observations during the month.

While considering the effect of the objective analysis scheme it is of interest to examine its behaviour in data sparse (or unsampled) areas such as the Southern Ocean. One method is to attempt to map a variable of known value using data values only at the positions of ship observations. For example, Figure 9 shows an attempt to map longitude over the globe. The contoured, analysed field should consist of the series of vertical stripes, as it does over most of the ocean. However for this July example (southern hemisphere winter) the field becomes distorted south of 30°S and is highly distorted beyond 45°S. In those regions information is being spread over large distances round the globe. If the latitude value is similarly analysed it is found that information from further north has been spread into the data sparse Southern Ocean. Such errors are not particular to the SOC climatology; similar errors will be found in all other studies which attempt to fill in data sparse regions using objective analysis techniques.

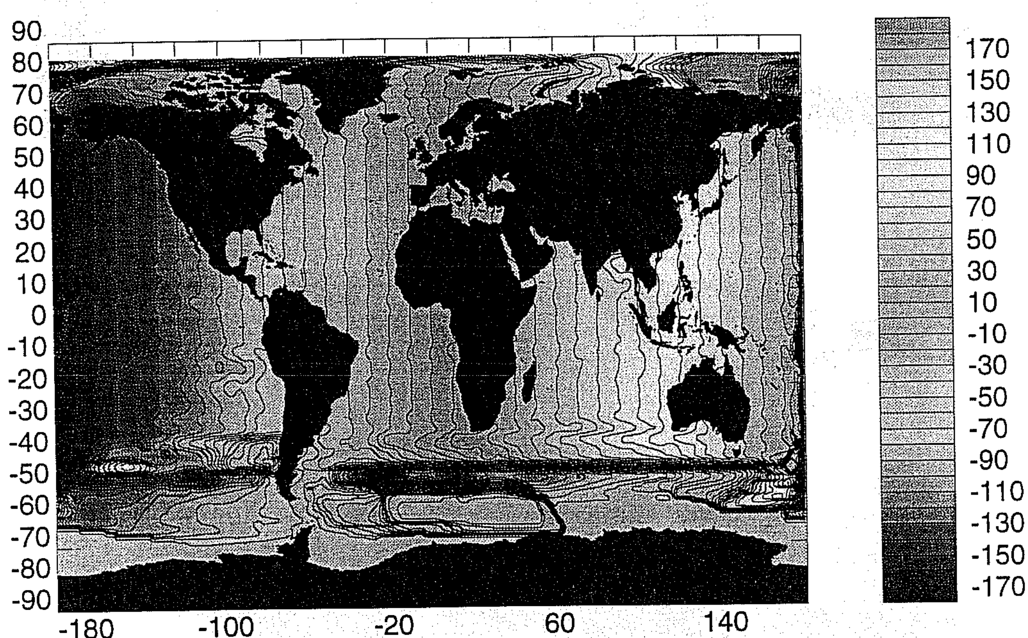

Figure 9. Successive correction analysis of the longitude value based on observation positions for latent heat estimates in July, 1993, (see text).

4. A SURFACE FLUX CLIMATOLOGY

4.1 Introduction

In the previous sections we have gone to some length to describe the problems, and potential sources of error, when calculating a global air-sea flux climatology. We believe that it is important that these limitations are understood by those wishing to make use of the fields. However in this section we will now attempt to show that the calculated heat flux fields not only represent qualitatively the major climatic regions, but also provide quantitatively useful estimates of the fluxes. The fields have been derived from the COADS 1a (Woodruff et al, 1993) which has been merged with additional metadata from the WMO47 list of ships as described in 3.2.2. A full description of the method employed and of the climatological fields is given in Josey et al (1997a), here we shall show as examples, fields for net heat flux and wind stress.

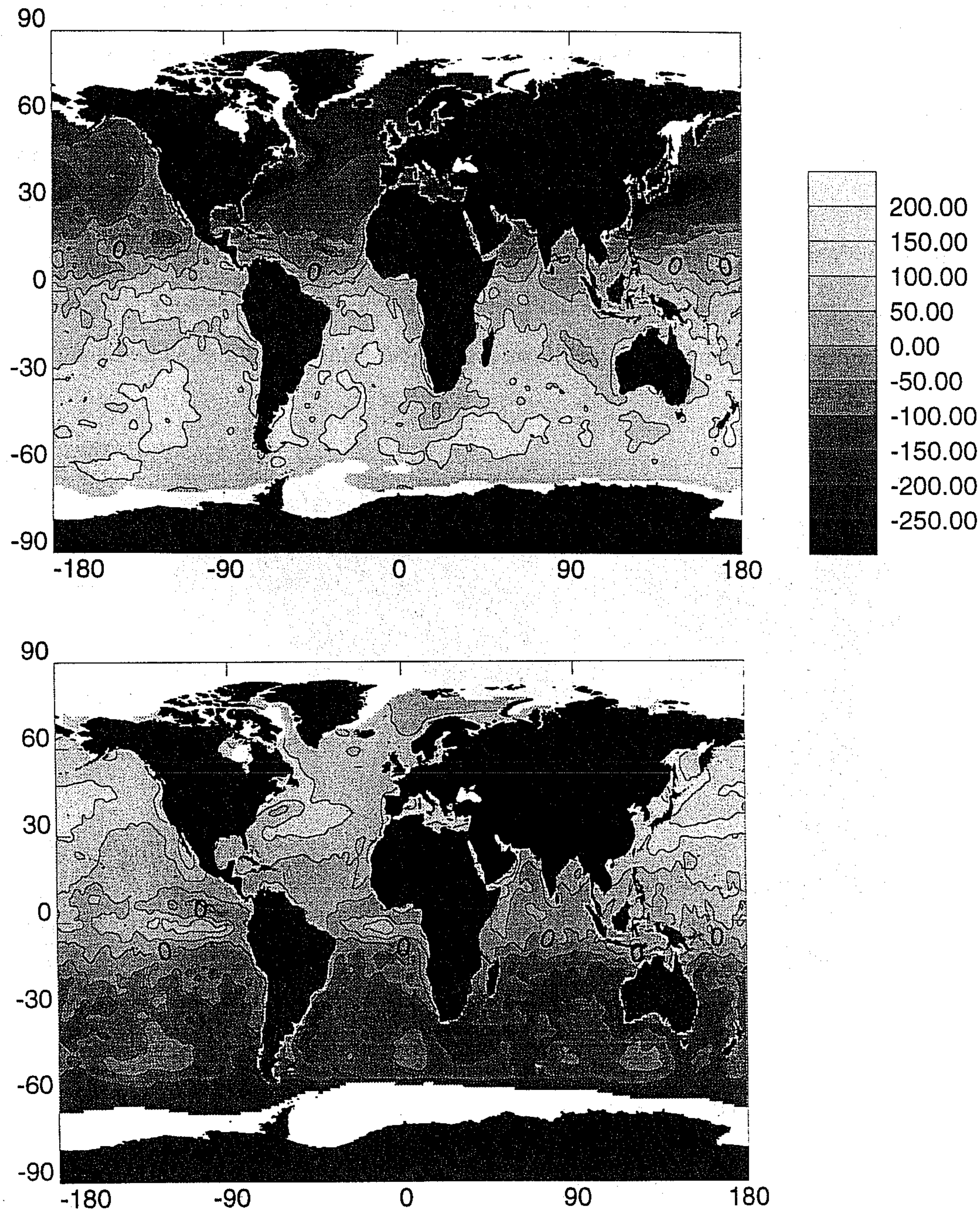

Figure 10. Mean values for the net heat flux into the ocean for January and July based on the years 1980 to 1993 (Josey et al. 1997a).

4.2 The geographical and seasonal variation of the mean fluxes

4.2.1 The net heat flux

Plots of the net heat flux into the ocean in January and July are shown in Figure 10. These plots are similar to those produced by other climatologies. They are shown here to emphasise that in almost all ocean regions the annual mean flux is a balance between summer heating and winter cooling. Thus at the peak of summer, net surface heating extends over the whole summer hemisphere and into the inter-tropical region of the winter hemisphere. Maximum surface heating occurs at about latitude 30°. In contrast cooling occurs over most of the winter hemisphere with the greatest cooling rates over the northern hemisphere western boundary currents.

The relative contribution of the different flux components is illustrated in Figure 11 by zonal means for January and July. The net longwave flux is relatively constant over the globe. The latent heat flux becomes small in the higher latitudes of the summer hemisphere. Compared to the other fluxes, the sensible heat flux is small except in the northern hemisphere in winter. That similar sensible heat fluxes are not observed in the southern hemisphere in July is presumably because of the lack of very cold, dry continental air (also the ice edge extends northwards to 60°S).

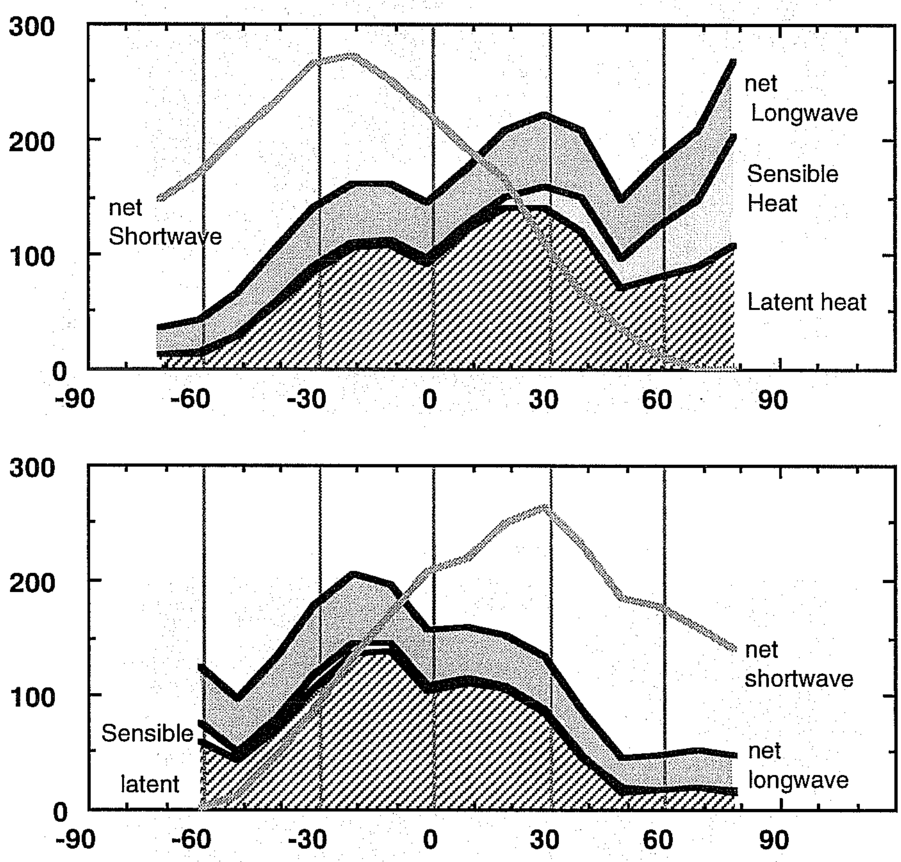

Figure 11 Zonal mean values for the heat flux components calculated from the SOC climatology (Josey et al. 1997a). The shaded area represent the contributions to surface heat loss from (bottom to top) latent heat, sensible heat, and net longwave. The total should be compared to the net shortwave line.

4.2.2 Wind Stress

The most widely employed climatology of wind stress forcing over the ocean has been that of Hellerman and Rosenstein (1983) which is based on observations from 1870-1976. A simple polynomial approximation was used for the drag coefficient with the air-sea temperature difference taken as a measure of atmospheric stability. However, as discussed by Harrison (1988), this resulted in an unrealistically high drag coefficient (compared for example to Smith (1980) - see section 2.4.2). As a result the Hellerman and Rosenstein wind stress estimates are probably too large over much of the globe. This has potentially significant consequences for derived quantities such as the Ekman transport and hence for estimates of the ocean heat transport. Figure 12 shows lower values of wind stress fields from the SOC climatology. Although not as marked as the seasonal shift in the net heat flux there is a clear weakening of the wind stress field between winter and summer in each hemisphere.

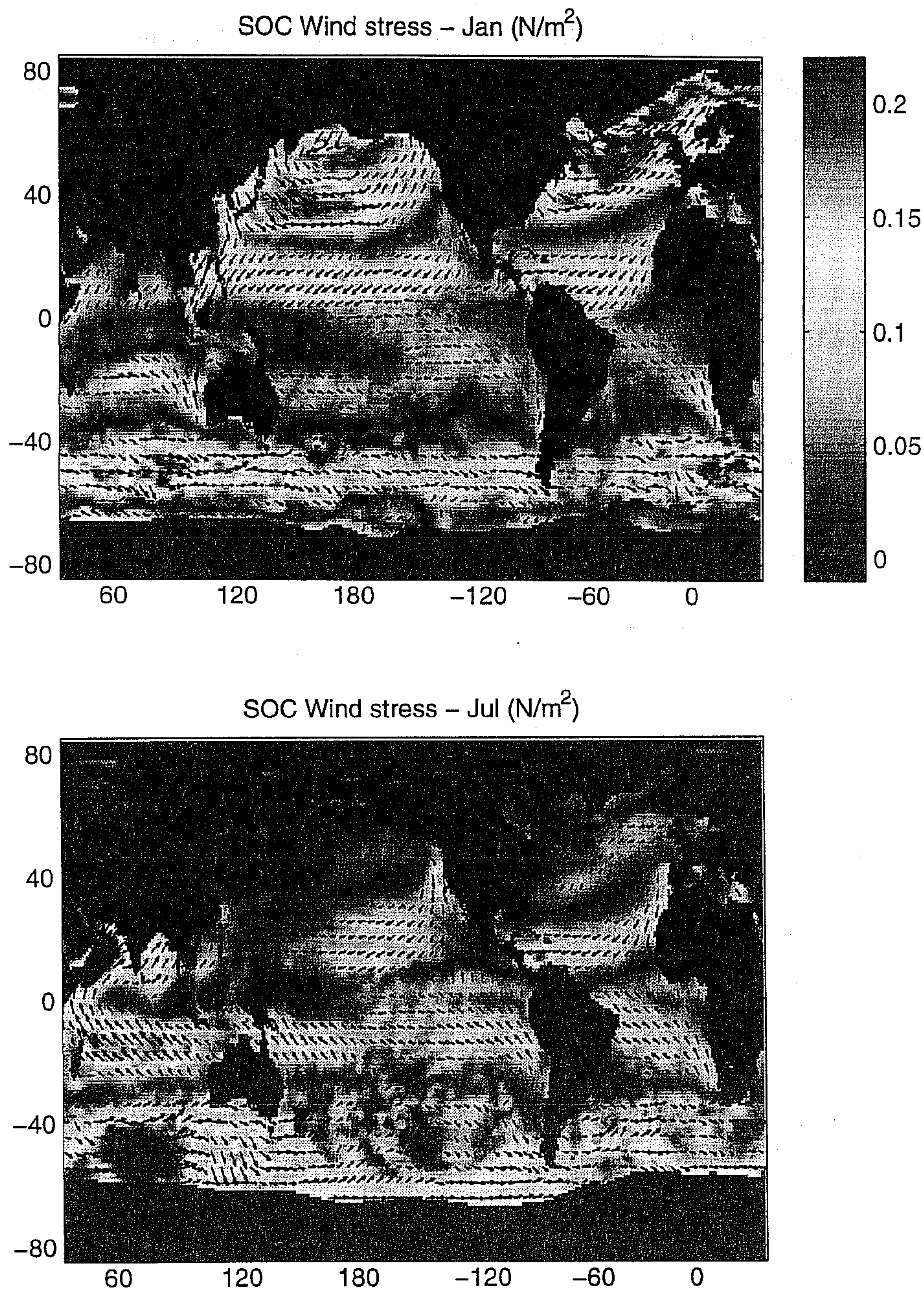

Figure 12. Mean values for the wind stress for January and July based on the years 1980 to 1993 (Josey et al. 1997a).

4.3 Verifying the flux estimates

4.3.1 The global heat balance

How accurate are the net heat flux distributions shown in Figure 10? The aim of applying the VSOP-NA derived corrections (Section 3.2) was to attempt to remove any systematic bias from the observations. However, if the net flux is integrated over the entire ocean area the result is a net heat flux into the ocean of 28 Wm^{-2}. Such a flux would cause the ocean to heat by 2.8^{o} C over the period sampled by the climatology (1980 -1993) if mixed down to a mean depth of 1000 m. Clearly, the true global mean must be much closer to zero. Similar results have been obtained in previous studies and corrections devised, usually by increasing the transfer coefficients significantly above the values found from air-sea interaction experiments. For example, Bunker and Worthington (1976) used transfer coefficients increased by 10% from the then accepted experimental values (which were already large compared to modern values). In recalculating Bunker's fluxes for the North Atlantic, Isemer and Hasse (1987) initially chose lower values for C_E compared to Bunker and Worthington (1976), however consideration of the implied meridional heat flux (Isemer et al., 1989) led to a revision upwards. Oberhuber (1988) increased his transfer coefficients for similar reasons.

Kent and Taylor (1995) compared these various flux calculation methods by applying each to the VSOP-NA (Section 3.2.1) data set. Compared to values calculated with the Smith (1988) transfer coefficients and uncorrected ship observations, they showed that the increase in the fluxes ranged from 21% (Oberhuber, 1988), through 31% (Hsiung, 1985; Isemer and Hasse, 1987) to 35% (Bunker, 1976),. These increases were significantly greater than the 6% to 15% increase that might be justified by the measurement errors detected during the VSOP-NA. Only Esbensen and Kushnir (1981) did not explicitly increase their flux estimates, however their use of the skin transfer coefficients of Liu et al. (1979) with the bulk sea temperatures was, in effect, a 16% increase.

A more recent climatology, that of da Silva et al. (1994), showed a global net heat flux of 30 Wm^{-2}, very similar to the SOC value. Adjustments were suggested based on various constraints on the heat and freshwater budgets. The required flux adjustments were almost independent of the constraints chosen. Insolation was decreased by about 10%, longwave cooling increased by about 2%, latent heat flux increased by about 14%. While these changes are not outside the possible errors in the flux calculations, the validity of applying adjustments on a global basis is questionable (see Section 6 and Josey et al., 1997a).

4.3.2 Meridional heat transport in the Atlantic Ocean

Oceanographic estimates of the ocean heat transport by the Atlantic Ocean exist for a number of different latitudes. The ocean heat transport implied by an earlier version of the SOC climatology

(Josey et al, 1996) is shown in Figure 13. It was calculated by integrating the net heat flux over successive zonal strips both northward and southward from 24° N where the Bryden (1993) value of 1.25 PW (based on three separate surveys) was used as a boundary value. Compared to the hydrographic estimates, the climatology implies too much heat (0.8 - 1.0 PW) into the polar oceans and too rapid a transition to southward heat transport in the South Atlantic compared to the Saunders and King (1995) value. Integrating from a different boundary value would merely recast the problem. For example, integrating southward from the Aagaard and Greisman (1975) value of 0.1 PW at 80° N would result in too small a northward heat transport at 24°N.

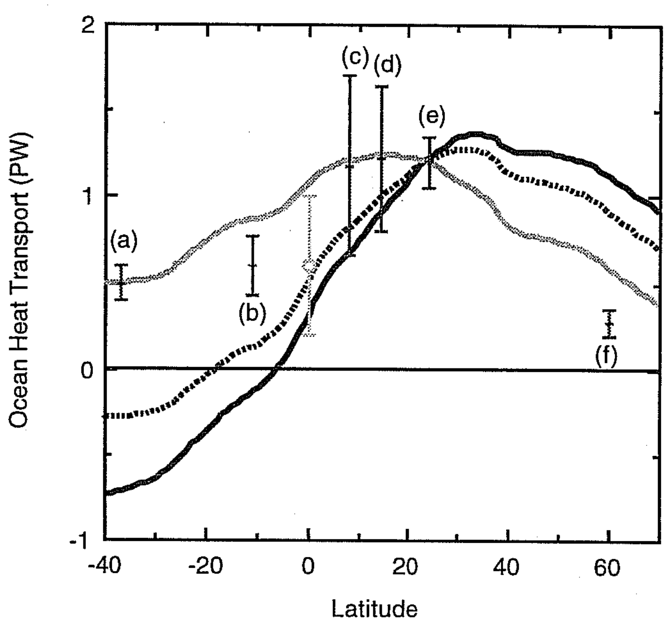

Figure 13. Meridional heat transport in the Atlantic Ocean. The dark line was calculated from the unadjusted fluxes. The dotted line shows the effect of increasing the latent heat flux by 10%. The grey line also has a "aerosol correction" added (see text). The hydrographic estimates shown are: (a) Saunders and King (1995), (b) Speer (1996), (c,d) Klein et al. (1995), (e) Hall and Bryden (1982), (f) Bacon (1995). The value on the equator is from Wunsch, 1984.

Also shown, for illustration, in Figure 12 is the effect of two possible adjustments. Firstly of increasing the latent heat loss by 10%, which results in a curve which is closer to the hydrographic estimates but with slopes which are still too steep in the southern hemisphere and too flat in the northern. Secondly of also reducing the shortwave flux by 15% between the equator and 30°N in a crude attempt to model dust loading from sub-Saharan regions; this ad hoc correction brings reasonable agreement.

Unfortunately what can be learnt from these comparisons with the ocean heat transport is limited. The hydrographic estimates have significant uncertainty. Most studies used section data obtained at a given time of year which may not be representative of the annual average. Using a combined analysis of the surface heat loss fields and internal heat storage calculated from sounding data, Lamb and Bunker (1982) suggested that there is a significant seasonal variation in the heat transport in the North Atlantic. Attempts have begun to synthesise heat transport estimates from the growing global set of hydrographic sections within the framework of an inverse model (Macdonald & Wunsch, 1996). Analyses such as this provide heat flux divergence estimates which have the potential to define particular regions in which the surface flux values are biased. Unfortunately for these estimates too, the likely errors are large.

4.3.4 Comparison with buoy data

The establishment of "Flux calibration sites" for verifying the fluxes from climatologies and from models was recommended by WCRP (1989). Perhaps the nearest realisation of such sites has been the series of mooring deployments undertaken by the Woods Hole group. Here we will make use of their moorings from the Subduction experiment in the North Atlantic (Moyer et al., 1996), in the west Pacific during the TOGA COARE experiment (Weller and Anderson, 1996), and in the Arabian Sea (Weller et al., 1997). The monthly mean data for the particular months of the buoy deployment has been extracted from the SOC flux dataset for the grid cells closest to the buoy position (for a more detailed discussion of the comparison see Josey et al., 1997a).

	F_{SW}		F_{LW}		F_E		F_H		F_{tot}	
	Buoy	SOC	Buoy	SOC	Buoy	SOC	Buoy	SOC	Buoy	SOC
N. Atlantic	192	176	-66	-57	-96	-92	-9.3	-6.3	21	21
SOC - Buoy	-16		9		4		3		0	
Arabian Sea	243	240	-59	-51 (-65)	-123	-111	-1.7	-1.2	60	77 (62)
SOC - Buoy	-3		8 (-6)		12		0.5		17 (2)	
Trop. Pac.	196	219	-58	-50	-108	-115	-8.8	-9.7	21	44
SOC-Buoy	23		8		-7		-0.9		23	

Table 3. Comparison of statistics from Buoys and SOC Fields. The positions and deployment periods were North Atlantic: 33°N 22°W, July 1991 - May 1993; Arabian Sea: 15.5°N 61.5°E, October 1994 - October 1995; Tropical Pacific: 1.75°S 156°E, October 1992 - March 1993. Figures in parentheses indicate the modified results if the Bignami formula is used in the Arabian Sea.

Overall the mean comparison (shown in the last row of Table 3) shows good correspondence between the buoy and SOC flux values. The SOC analysis showed agreement with the buoy at the Subduction site in the net heat flux to within 1 Wm^{-2} ; at the other two sites the SOC heat gain was of order 20 Wm^{-2} greater than the buoy value. The climatic conditions in the Arabian Sea were such that the Bignami et al. (1995) formula for the net longwave cooling might be justified in preference to the Clark et al (1974) formula; in that case the SOC and buoy net heat fluxes would be in agreement to 2 Wm^{-2}. There appears to be a systematic underestimate of the longwave cooling (by about 8 Wm^{-2}) at all three sites with the Clark et al. formula. However, for the other components of the heat flux, the comparisons varied significantly when the individual sites are examined. The climatology had 16 Wm^{-2} less shortwave heating at the North Atlantic site, but 23 Wm^{-2} more at the COARE site. At the Arabian Sea site the climatology underestimated the latent heat cooling by 12 Wm^{-2}; at the COARE site it over-estimated by 7 Wm^{-2}. The latter result would appear to justify our use of the standard bulk formulae with the VOS observations for areas such as the Pacific warm pool, rather than attempting to implement the more sophisticated TOGA COARE algorithm (Section 2.4).

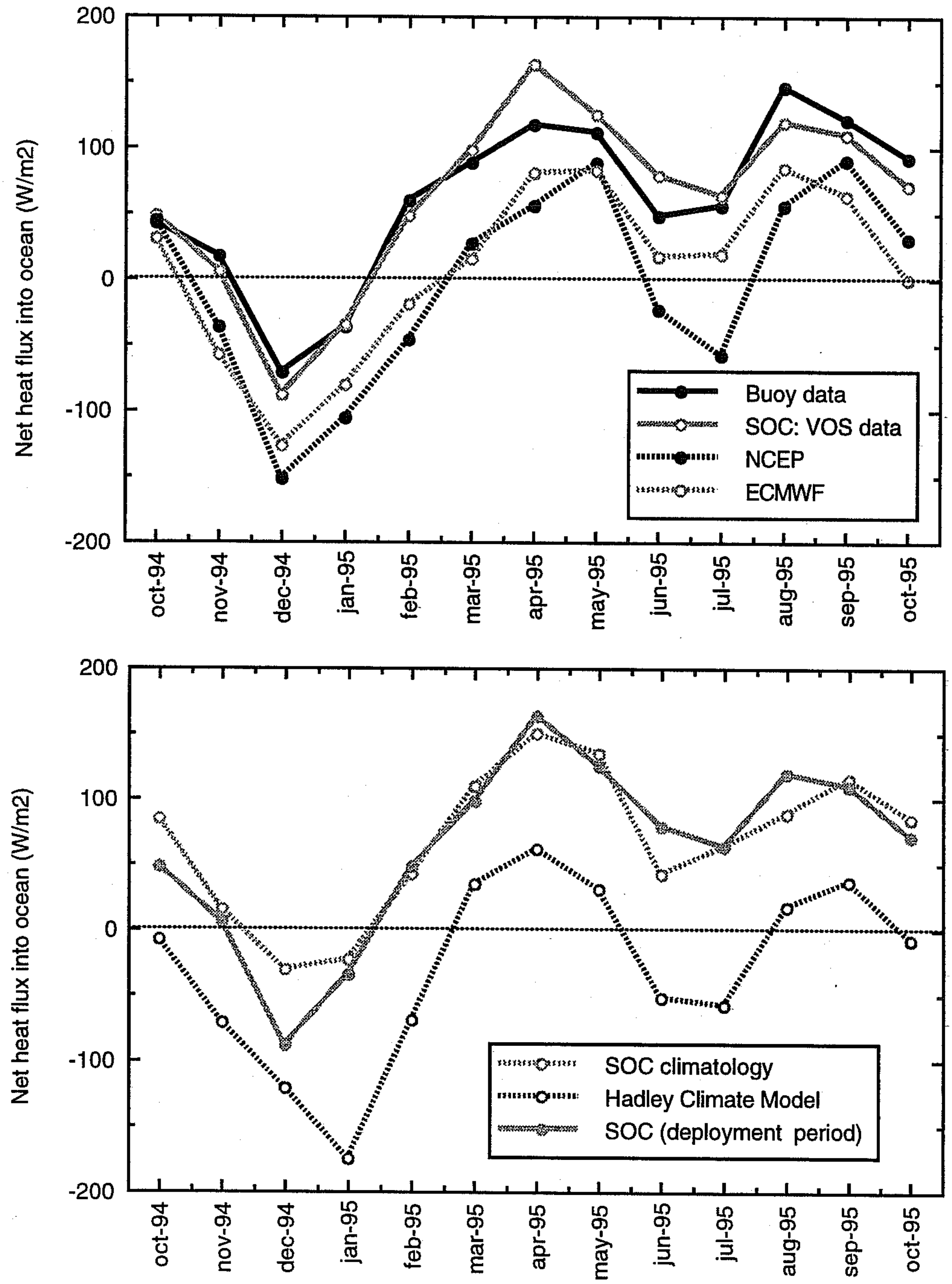

Figure 14. Net heat flux at the Arabian Sea buoy site (adapted from Weller et al. 1997). Top: the measured buoy values, the values from the SOC flux fields, and fluxes from the ECMWF and NCEP models. In each case the data are for the actual buoy deployment period. Bottom: The SOC flux values for the deployment period and for the 1980 to 1993 mean values. Also the fluxes from an AMIP run of the Hadley Centre climate model (Hall, C. D., R. A. Stratton and M. L. Gallani, 1995: *Climate simulations with the Unified Model: AMIP runs.* Climate Research Technical Note 61, Hadley Centre, Bracknell, UK).

The Arabian Sea data set has been chosen for further analysis since this region shows a large seasonal variation due to the effects of the two monsoons. Periods of substantial net heating and cooling are seen in the time series of the buoy data from Weller et al (1997) (Figure 14a). Also shown are the SOC climatology values for the specific months of the buoy deployment. The climatology is seen to track the variations in net heating well, the standard deviation, σ_{obs}, of the monthly differences is about 22 Wm^{-2} (implying that the 12 month mean difference from the SOC climatology is known to around ±6 Wm^{-2}).

How well do the σ_{obs} values from the buoy deployment agree with our prediction, σ_{pred}, based on the estimates of the random errors (Section 3.3)? A rigorous treatment (Josey et al. 1997a) requires objective analysis of the generated error fields; here we will attempt a rough estimate. Using latent heat as an example, semivariogram analysis indicated that in the Arabian Sea area the standard deviation for a single ship observation would be large, 77 Wm^{-2} in January and 91 Wm^{-2} in July. The buoy was deployed in a poorly sampled area, however the successive correction method produced an effective sampling density of about 1 observation per 1° grid square. The analysed SOC flux fields from a 5° by 5° area surrounding the buoy were used for comparison. This will reduce the σ_{pred} values; based on the sampling density we will assume a factor of 5. Then, using the mean of January and July values, both σ_{pred} and σ_{obs} agree at 17 Wm^{-2}. Similar calculations for the sensible heat flux gave values for both σ_{pred} and σ_{obs} of about 4 Wm^{-2}. While this exact agreement must be fortuitous, the results illustrate how the accuracy observed in the buoy comparisons is possible, despite large random errors in individual observations.

For the longwave radiation σ_{obs} was significantly more than predicted (σ_{pred} = 3 Wm^{-2} compared to σ_{obs} = 13 Wm^{-2}, or 10 Wm^{-2} if the Bignami, (1995) formula were used). A possible explanation is that, while the ships may be providing consistent estimates of total cloud amount, that parameter is not sufficient to determine the downwelling radiation. For F_{SW} we are unable to predict a value because of the use of monthly means in the formulae, however for this variable, σ_{obs} = 25 Wm^{-2}.

4.4 Comparison with Numerical Models

The buoy data can also be used to evaluate flux products from numerical models. Shown in Figure 14a are the net heat fluxes obtained from the ECMWF and NCEP weather prediction models. The model mean values showed less heat input to the ocean by 20 to 30 Wm^{-2}. This was due to lower shortwave heating and greater latent heat loss (Weller et al. 1997). That the year in which the observations were obtained, 1995, was typical of conditions in the region is illustrated in Figure 14b, which compares the SOC values for 1994-95 to the climatological mean values. Again there is good agreement. However also shown on this Figure are flux values from a climate model (an AMIP run of the UK Hadley Centre model). As for the other models, these show too much heat loss from the ocean, by about 30 Wm^{-2}. This was mainly caused by excessive latent heat loss in the model. For this area at least the SOC climatology is superior to a sample of 3 numerical models!

5. NEXT STEPS: THE POTENTIAL OF SATELLITE DATA

5.1 Introduction

In future the best flux estimates will be produced by combining in situ data with remotely sensed information from satellite sensors, or will be obtained from numerical models which have assimilated the available observational data. Other papers during this meeting will discuss model derived fluxes. In this section we will briefly discuss the potential value of satellite derived data. Satellite sensors have the potential for providing consistent observations over the global ocean. However, although the sampling density is far superior to the ship observations over many ocean regions, sampling studies (Salby, 1982a,b) suggest that wide swath sensors flown on a pair of polar orbiting satellites are needed to provide routinely the coverage sought (Taylor, 1984).

5.2 Radiative flux determination from satellite

5.2.1 Shortwave

Global fields of the surface shortwave radiative flux obtained from satellite radiance data have been available for several years. The primary analysis effort has been the WCRP Surface Radiation Budget (SRB) Project (Whitlock et al., 1995) from which fields are available on a 280 km grid for the period March 1985 - December 1988. Radiance parameters from the International Satellite Cloud Climatology Project (ISCCP) were input to the "Pinker" and "Staylor" radiative transfer algorithms. By taking these two quite different approaches, it was hoped to check the accuracy of the estimates in regions lacking ground truth. Several sources of uncertainty remain, particularly those related to aerosol attenuation and the effects of absorption by water vapour (WCRP/GEWEX, 1996). Nevertheless, (Whitlock et al., 1995) claim that the estimated shortwave irradiances have typical bias values in the range ± 20 W/m^2 and rms errors of around 25 W/m^2 in the monthly mean (although there are regions with much larger uncertainties). At present a revised version of the WCRP/SRB product is being prepared using the Pinker algorithm which will cover an extended period (July 1983 - June 1995) and take as input a revised ISCCP dataset (WCRP/GEWEX, 1996).

5.2.2 Longwave

Less progress has been made in the retrieval of the surface longwave flux using satellite data. Various algorithms have been developed which make use of a combination of temperature and humidity profiles and cloud amount (Katsaros, 1990) but these remain at the experimental stage (Gupta et al., 1992) and further intercomparison work needs to be carried out before reliable estimates of the longwave can be obtained (WCRP/GEWEX, 1996).

5.3 Turbulent fluxes

5.3.1 Introduction

Satellite sensors have the potential to determine most of the variables needed to determine the turbulent heat fluxes and wind stress. One major omission is the near surface air temperature which is needed both to determine the sensible heat flux and to estimate the atmospheric stability.

5.3.2 Sea surface temperature

Remotely sensed sea surface temperature data are from infra-red radiometers have now been available for some 20 years. McClain et al., (1985) found an accuracy (rms scatter) for Advanced Very High Resolution Radiometer (AVHRR) data of about ±0.6°C. Despite improvements to the retrieval algorithms (see for example, Wick et al. 1992), the global accuracy of the AVHRR data remains little changed (Barton 1995). The error budget is dominated by the instrument calibration accuracy and atmospheric transmission effects. The most reliable global data set remains a combination of satellite and in situ data (e.g. Reynolds and Smith, 1994).

5.3.3 Near surface humidity

Liu and Niiler (1984) suggested that monthly mean values for near surface atmospheric humidity can be estimated by using the total atmospheric water vapour content data measured by passive microwave radiometers. In thus calculating evaporation over the global oceans (Esbensen et al., 1993) found biases of order 2g/kg due to the variations in the mean atmospheric humidity profile. Jourdan and Gautier (1995) achieved improved accuracy by blending ship and satellite estimates in order to calculate a global latent heat flux. An alternative technique is to estimate directly the water content in the lower 500m of the atmosphere from radiometer data (Schulz et al. 1993). An accuracy of 1.2g/kg was achieved in comparisons with individual radiosonde ascents and averaging would reduce this error significantly.

5.3.4 Wind stress

Wind velocity can be estimated using a satellite scatterometer. For example, the ERS scatterometer samples a 500km swath with 50km spatial resolution. The retrieval algorithm has been fitted empirically to buoy wind data (Quilfen, 1994). For wind speed the bias was less than 10cm/s and the rms difference was 1.2 m/s over a range of 2 to 20 m/s. For wind direction the rms difference was 15°. Using semivariogram analysis, Kent et al. (1997b) demonstrated that, whereas the quality controlled ship data had a rms error of about 2 m/s, the rms for the scatterometer winds was 0.5 m/s. Taking this difference into account, the unbiased regression was:

$$U_{10n}(\text{ship}) = 1.025 \times U_{10n}(\text{satellite}) + 0.255 \qquad (??)$$

Although the scatterometer offers great promise for global wind velocity determination, at present there are no plans to mount a scatterometer on an operational satellite system. Research satellite systems do not guarantee continuity as was recently demonstrated by the failure of the Advanced Earth Observing System (ADEOS) satellite.

Wind speed data for estimating heat fluxes may also be obtained from passive microwave sensors or radar altimeters.

5.4 Precipitation

Global climatological fields of monthly mean precipitation over both ocean and land, derived from a combination of satellite and terrestrial observations made over the period July 1987 to June 1995, have recently been produced within the Global Precipitation Climatology Project (GPCP, Huffman et al, 1997). The satellite values are a combination of passive microwave and infrared techniques. The infrared techniques rely on a correlation between cloud top brightness temperatures and rain rate and are more appropriate for use with deep convective systems. Used within 40 degrees of the equator, they exploit the more frequent sampling afforded by geostationary satellites. The microwave technique detects the radiative effects of hydrometeors and has the stronger correlation with surface rainfall. It relies on sensors on polar orbiting satellites (to obtain adequate spatial resolution) and hence has a relatively low sampling rate, but can be employed to higher latitudes in the GPCP analysis.

The major problem facing satellite estimates of precipitation is the lack of alternative fields for validation. Attempts at validation thus far have been restricted to comparisons with climatologies obtained from ship present weather reports which also suffer from large uncertainties (Legates, 1995).

6. DISCUSSION AND SUMMARY

We have described the methods used to calculate air-sea heat and momentum fluxes from ship data. It has been emphasised that individual flux estimates may have large errors and that to obtain accurate flux estimates, systematic errors must be corrected and a large number of observations averaged. We have described correction formulae based on the results from the VSOP-NA and other experiments which aim to remove measurement the biases. We have also demonstrated how to estimate the random errors in the observed quantities. Taking the SOC climatology (Josey et al. 1997a) as an example, we have evaluated the calculated flux fields in different ways.

Considering first the global heat balance, despite the attempts to remove biases from the individual observations, the SOC climatology exhibits a feature common to other climatologies; it indicates a net heating of the ocean. The imbalance, about 30 Wm^{-2} is similar to that found by da Silva et al. 1994. Consideration of the meridional heat transport in the North Atlantic again suggests that the calculated net heat input to the ocean must be decreased. Preliminary comparisons with three buoy deployments suggest that the Clark et al. (1974) formula underestimates the longwave cooling by several Wm^{-2} (and there is other evidence that, in some regions of the ocean, the Bignami et al. (1995) formula is preferable). However the buoy comparisons do not give evidence for a consistent bias in any of the other flux components. Based on this and other evidence, Josey et al. (1997a) will suggest that the global heat balance should not be corrected purely by applying global correction factors, rather that regional corrections to the heat budget may be necessary. The buoy comparisons also suggest that the flux values from the climatology are still, at least in some regions, closer to reality than the fluxes derived from numerical models.

When examining flux fields both the spatial and temporal variability is of interest. Over large areas of the globe, the ship observations are not sufficient in number to determine year to year variability in any detail. However the observations do define the climatological mean variation in the seasonal cycle which, arguably is the most important temporal variation to be understood (Wallace, 1996).

A major weakness in using the ship data is the lack of adequate sampling in much of the southern hemisphere, and especially in the Southern Ocean. The use of data from satellite sensors will be of particular value in these regions. Sea surface temperature, wind, shortwave radiation, and precipitation products are available. Techniques for longwave radiation, and near surface humidity exist. A method to obtain sensible heat flux estimates remains to be developed.

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