HEAT AND MOMENTUM FLUXES AT THE SEA SURFACE

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Introduction

The maintenance of the earth's climate depends on a balance between the absorption of heat from the sun and the loss of heat through radiative cooling to space. For each 100 W of the sun's radiative energy entering the atmosphere nearly 40 W is absorbed by the ocean – about twice that adsorbed in the atmosphere and three times that falling on land surfaces. Much of this oceanic heat is transferred back to the atmosphere by the local sea to air heat flux. The geographical variation of this atmospheric heating drives the weather systems and their associated winds. The wind transfers momentum to the sea causing waves and the wind-driven currents. Major ocean currents transport heat polewards and at higher latitudes the sea to air heat flux significantly ameliorates the climate. Thus the heat and momentum fluxes through the ocean surface form a crucial component of the earth's climate system.

The total heat transfer through the ocean surface, the net heat flux, is a combination of several components. The heat from the sun is the short-wave radiative flux (wavelength 0.3–3 μm). Around noon on a sunny day this flux may reach about $1000\,\mathrm{W\,m^{-2}}$ but, when averaged over 24 h, a typical value is $100–300\,\mathrm{W\,m^{-2}}$ varying with latitude and season. Part of this flux is reflected from the sea surface – about 6% depending on the solar elevation and the sea state. Most of the remaining short-wave flux is absorbed in the upper few meters of the ocean. In calm weather, with winds less than about $3\,\mathrm{m\,s^{-1}}$, a shallow layer may be formed during the day in which the sea is warmed by a few degrees Celsius (a 'diurnal thermocline'). However, under stronger winds or at night the absorbed heat becomes mixed down through several tens of metres. Thus, in contrast to land areas, the typical day to night variation in sea surface sea and air temperatures is small, $<1°\mathrm{C}$. Both the sea and the sky emit and absorb long-wave radiative energy (wavelength 3–50 μm). Because, under most circumstances, the radiative temperature of the sky is colder than that of the sea, the downward long-wave flux is usually smaller than the upward flux. Hence the net long-wave flux acts to cool the surface, typically by $30–80\,\mathrm{W\,m^{-2}}$ depending on cloud cover.

The turbulent fluxes of sensible and latent heat also typically transfer heat from sea to air. The sensible heat flux is the transfer of heat caused by difference in temperature between the sea and the air. Over much of the ocean this flux cools the sea by perhaps $10–20\,\mathrm{W\,m^{-2}}$. However, where cold wintertime continental air flows over warm ocean currents, for example the Gulf Stream region off the eastern seaboard of North America, the sensible heat flux may reach $100\,\mathrm{W\,m^{-2}}$. Conversely warm winds blowing over a colder ocean region may result in a small sensible heat flux into the ocean – a frequent occurrence over the summertime North Pacific Ocean. The evaporation of water vapor from the sea surface causes the latent heat flux. This is the latent heat of vaporization which is carried by the water vapor and only released to warm the atmosphere when the vapor condenses to form clouds. Usually this flux is significantly greater than the sensible heat flux, being on average $100\,\mathrm{W\,m^{-2}}$ or more over large areas of the ocean. Over regions such as the Gulf Stream latent heat fluxes of several hundred $\mathrm{W\,m^{-2}}$ are observed. In foggy conditions with the air warmer than the sea, the latent heat flux can transfer heat from air to sea. In summertime over the infamous fog-shrouded Grand Banks off Newfoundland the mean monthly latent heat transfer is directed into the ocean, but this is an exceptional case.

Measuring the Fluxes

The standard instruments for determining the radiative fluxes measure the voltage generated by a thermopile which is exposed to the incident radiation. Typically the incoming short-wave radiation is measured by a pyranometer which is mounted in gimbals for use on a ship or buoy (**Figure 1**). For better accuracy the direct and scattered components should be determined separately but, apart from at the Baseline Surface Radiation Network stations which are predominantly situated on land, at present this is rarely done. The reflected short-wave

Figure 1 A pyranometer used for measuring short-wave radiation. The thermopile is covered by two transparent domes. (Photograph courtesy of Southampton Oceanography Centre.)

radiation is normally determined from the sun's elevation and lookup tables based on the results of previous experiments. The pyrgeometer used to determine the long-wave radiation is similar to the pyranometer but uses a coated dome to filter out, as far as possible, the effects of the short-wave heating. Because the air close to the sea surface is normally near to the sea temperature, the use of gimbals is less important. However, a clear sky view is required and a number of correction terms have to be calculated for the temperature of the dome and any short-wave leakage. Again, only the downward component is normally measured; the upwards component is calculated from knowledge of the sea temperature and emissivity of the sea surface.

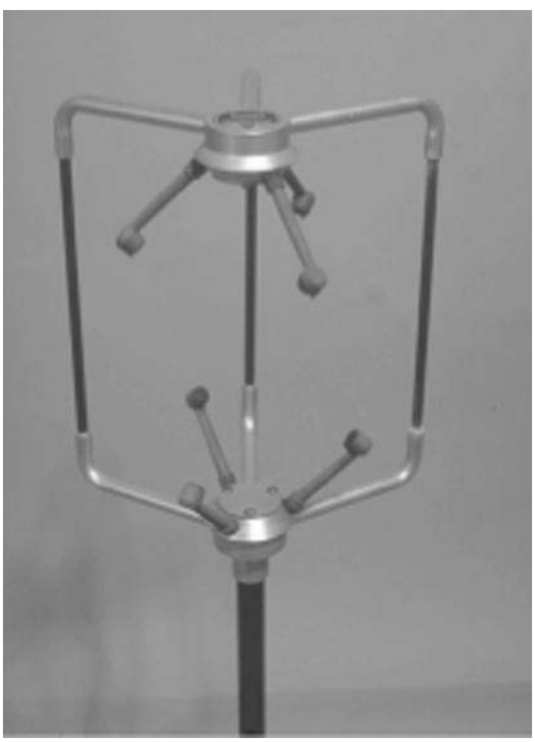

Figure 2 The sensing head of a three-component ultrasonic anemometer. The wind components are determined from the different times taken for sound pulses to travel in either direction between the six ceramic transducers. (Photograph courtesy of Southampton Oceanography Centre.)

The turbulent fluxes may be measured in the near-surface atmosphere using the eddy correlation method. If upward moving air in an eddy is on average warmer and moister than the downward moving air, then there is an upwards flux of sensible heat and water vapor and hence also an upward latent heat flux. Similarly the momentum flux, or wind stress, may be determined from the correlation between the horizontal and vertical wind fluctuations. Since a large range of eddy sizes may contribute to the flux, fast response sensors capable of sampling at 10 Hz or more must be exposed for periods of the order of 30 min for each flux determination. Three-component ultrasonic anemometers (**Figure 2**) are relatively robust and, by also determining the speed of sound, can provide an estimate of the sonic temperature flux, a function of the heat and moisture fluxes. The sensors used for determining the fluctuations in temperature and humidity have previously tended to be fragile and prone to contamination by salt particles which are ever-present in the marine atmosphere. However, improved sonic thermometry, and new techniques for water vapor measurement, such as microwave refractometry or differential infrared absorption instruments, are now becoming available.

Despite these improvements in instrumentation, obtaining accurate eddy correlation measurements over the sea remains very difficult. If the instrumentation is mounted on a buoy or ship the six components of the wave-induced motion of the measurement platform must be measured and removed from the signal. The distortion both of the turbulence and the mean wind by ship, buoy or fixed tower must be minimized and, as far as possible, corrected for. Thus eddy correlation measurements are not routinely obtained over the ocean, rather they are used in special air–sea interaction experiments to calibrate other less direct methods of flux estimation. For example, in the inertial dissipation method, fluctuations of the wind, temperature, or humidity at a few Hertz are measured and related (through turbulence theory) to the fluxes. This method is less sensitive to flow distortion or platform motion, but relies on various assumptions about the formation and dissipation of turbulent quantities, which may not be valid under some conditions. It has been implemented on a semi-routine

Table 1 Typical values (with estimated uncertainties) for the transfer coefficients[a]

Flux	*Transfer coefficients*	*Typical values*
Momentum	Drag coefficient C_{D10n} ($\times 1000$)	$= 0.61\ (\pm 0.05) + 0.063\ (\pm 0.005)\quad U_{10n}$ $(U_{10n} > 3\,\mathrm{m\,s^{-1}}) = 0.61 + 0.57/U_{10n}\ (U_{10n} < 3\,\mathrm{m\,s^{-1}})$
Sensible heat	Stanton no., C_{H10n}	$1.1\ (\pm 0.2) \times 10^{-3}$
Latent heat	Dalton no., C_{E10n}	$1.2\ (\pm 0.1) \times 10^{-3}$

[a]Neither the low wind speed formula for C_{D10n}, nor the wind speed below which it should be applied, are well defined by the available, very scattered, experimental data. It should be taken simply as an indication that, at low wind speeds, the surface roughness increases as the wind speed decreases due to the dominance of viscous effects.

basis on some research ships to increase the range of available flux data.

The most commonly used method of flux estimation is variously referred to as the bulk (aerodynamic) formulae. These formulae relate the difference between the value of temperature, humidity or wind ('x' in [1]) at some measurement height, z, and the value assumed to exist at the sea surface – respectively the sea surface temperature, 98% saturation humidity (to allow for salinity effects), and zero wind (or any nonwind-induced water current). Thus the flux F_x of some quantity x is:

$$F_x = \rho U_z C_{xz}(x_z - x_o) \qquad [1]$$

where ρ is the air density, and U_z the wind speed at the measurement height. While appearing intuitively correct (for example, blowing over a hot drink will cool it faster) these formulae can also be derived from turbulence theory. The value for the transfer coefficient, C_{xz}, characterizes both the surface roughness applicable to x and the relationship between F_x and the vertical profile of x. This varies with the atmospheric stability, which itself depends on the momentum, sensible heat, and water vapor fluxes, as well as the measurement height. Thus, although it may appear simple, Eqn [1] must be solved by iteration, initialized using the equivalent neutral value of C_{xz} at some standard height (normally 10 m), C_{x10n}. Typical neutral values (determined using eddy correlation or inertial dissipation data) are shown in **Table 1**. Many research problems remain. For example: C_{D10n} is expected to depend on the state of development of the wave field, but can this be successfully characterized by the ratio of the predominant wave speed to the wind speed (the wave age), or by the wave height and steepness, or is a spectral representation of the wave field required? What are the effects of waves propagating from other regions (i.e., swell waves)? What is the behavior of C_{D10n} in low wind speed conditions? Furthermore C_{E10n} and C_{H10n} are relatively poorly defined by the available experimental data, and recent bulk algorithms have used theoretical models of the ocean surface (known as surface renewal theory) to predict these quantities from the momentum roughness length.

Sources of Flux Data

Until recent years the only source of data for flux calculation routinely available from widespread regions of the world's oceans was the weather reports from merchant ships. Organized as part of the World Weather Watch system of the World Meteorological Organisation, these 'Voluntary Observing Ships (VOS)' are asked to return coded weather messages at 00 00, 06 00, 12 00, and 18 00 h GMT daily, also recording the observation (with further details) in the ship's weather logbook. The very basic set of instruments provided will normally include a barometer and a means of measuring air temperature and humidity – typically wet and dry bulb thermometers mounted in a hand swung sling psychrometer or a fixed, louvered 'Stevenson' screen. Sea temperature is obtained using a thermometer and an insulated bucket, or by reading the temperature gauge for the engine cooling water intake. Depending on which country recruited the VOS an anemometer and wind vane might be provided, or the ship's officers might be asked to estimate the wind velocity from observations of the sea state using a tabulated 'Beaufort scale'. Because of the problems of adequately siting an anemometer and maintaining its calibration, these visual estimates are not necessarily inferior to anemometer-based values.

Thus the VOS weather reports include all the variables needed for calculating the turbulent fluxes using the bulk formulae. However, in many cases the accuracy of the data is limited both by the instrumentation and its siting. In particular, a large ship can induce significant changes in the local temperature and wind flow, since the VOS are not equipped with radiometers. The short-wave and long-wave fluxes must be estimated from the observer's estimate of the cloud amount plus (as appropriate) the solar elevation, or the sea and air temperature and humidity. The unavoidable observational errors and the crude form of the radiative flux formulae imply that large numbers of reports are needed, and correction schemes must be applied, before satisfactory flux estimates can be obtained. While there are presently nearly 7000 VOS, the ships tend to be concentrated in the main shipping lanes. Thus whilst coverage in most of the North Atlantic and North Pacific is adequate to provide monthly mean flux values, elsewhere data is mainly restricted to relatively narrow, major trade routes. For most of the southern hemisphere the VOS data is only capable of providing useful values if averaged over several years, and reports from the Southern Ocean are very few indeed. These shortcomings of VOS-derived fluxes must be borne in mind when studying the flux distribution maps presented below.

Satellite-borne sensors offer the potential to overcome these sampling problems. They are of two types, passive sensors which measure the radiation emitted from the sea surface and the intervening atmosphere at visible, infrared, or microwave frequencies, and active sensors which transmit microwave radiation and measure the returned signal. Unfortunately these remotely sensed data do not allow all of the flux components to be adequately estimated. Sea surface temperature has been routinely determined using visible and infrared radiometers since about 1980. Potential errors due, for example, to changes in atmospheric aerosols following volcanic eruptions, mean that these data must be continually checked against ship and buoy data. Algorithms have been developed to estimate the net surface short-wave radiation from top of the atmosphere values; those for estimating the net surface long wave are less successful. The surface wind velocity can be determined to good accuracy by active scatterometer sensors by measuring the microwave radiation backscattered from the sea surface. Unfortunately scatterometers are relatively costly to operate, since they demand significant power from the spacecraft and, to date, few have been flown. The determination of near-surface air temperature and humidity from satellite is hindered by the relatively coarse vertical resolution of the retrieved data. A problem is that the radiation emitted by the near-surface air is dominated by that originating from the sea surface. Statistically based algorithms for determining the near-surface humidity have been successfully demonstrated. More recently neural network techniques have been applied to retrieving both air temperature and humidity; however, at present there is no routinely available product. Thus the satellite flux products for which useful accuracy has been demonstrated are presently limited to momentum, short-wave radiation, and latent heat flux.

Numerical weather prediction (NWP) models (as used in weather forecasting centers) estimate values of the air–sea fluxes as a necessary part of their calculations. Since these models assimilate most of the available data from the World Weather Watch system, including satellite data, radiosonde profiles, and surface observations, it might be expected that NWP models represent the best source of flux data. However, there are other problems. The vertical resolution of these models is relatively poor and many of the near-surface processes which affect the fluxes have to be represented in terms of larger-scale parameters. Improvements to these models are normally judged on the resulting quality of the weather forecasts, not on the accuracy of the surface fluxes; sometimes these may become worse. Indeed, the continual introduction of model changes results in time discontinuities in the output variables. This makes the determination of interannual variations difficult. Because of this, NWP centres such as the European Centre for Medium Range Weather Forecasting (ECMWF) and the US National Centers for Environmental Prediction (NCEP) have reanalyzed the past weather and have gone back several decades. The surface fluxes from these reanalyses are receiving much study. Those presently available appear less accurate than fluxes derived from VOS data in regions where there are many VOS reports; in sparsely sampled regions the model fluxes may be more accurate. There are particular weaknesses in the short-wave radiation and latent heat fluxes. New reanalyses are planned and efforts are being made to improve the flux estimates; eventually these reanalyses will provide the best source of flux data for many purposes.

Regional and Seasonal Variation of the Momentum Flux

The main features of the wind regimes over the global oceans have long been recognized and descriptions are available in many books on marine meteorology (see Further Reading). The major features of the wind stress variability derived from ship observations from the period 1980–93 will be summarized here, using plots for January and July to illustrate the seasonal variation. The distribution of the heat fluxes will be discussed in the next section.

In northern hemisphere winter (**Figure** 3A) large wind stresses due to the strong midlatitude westerly winds are obvious in the North Atlantic and the

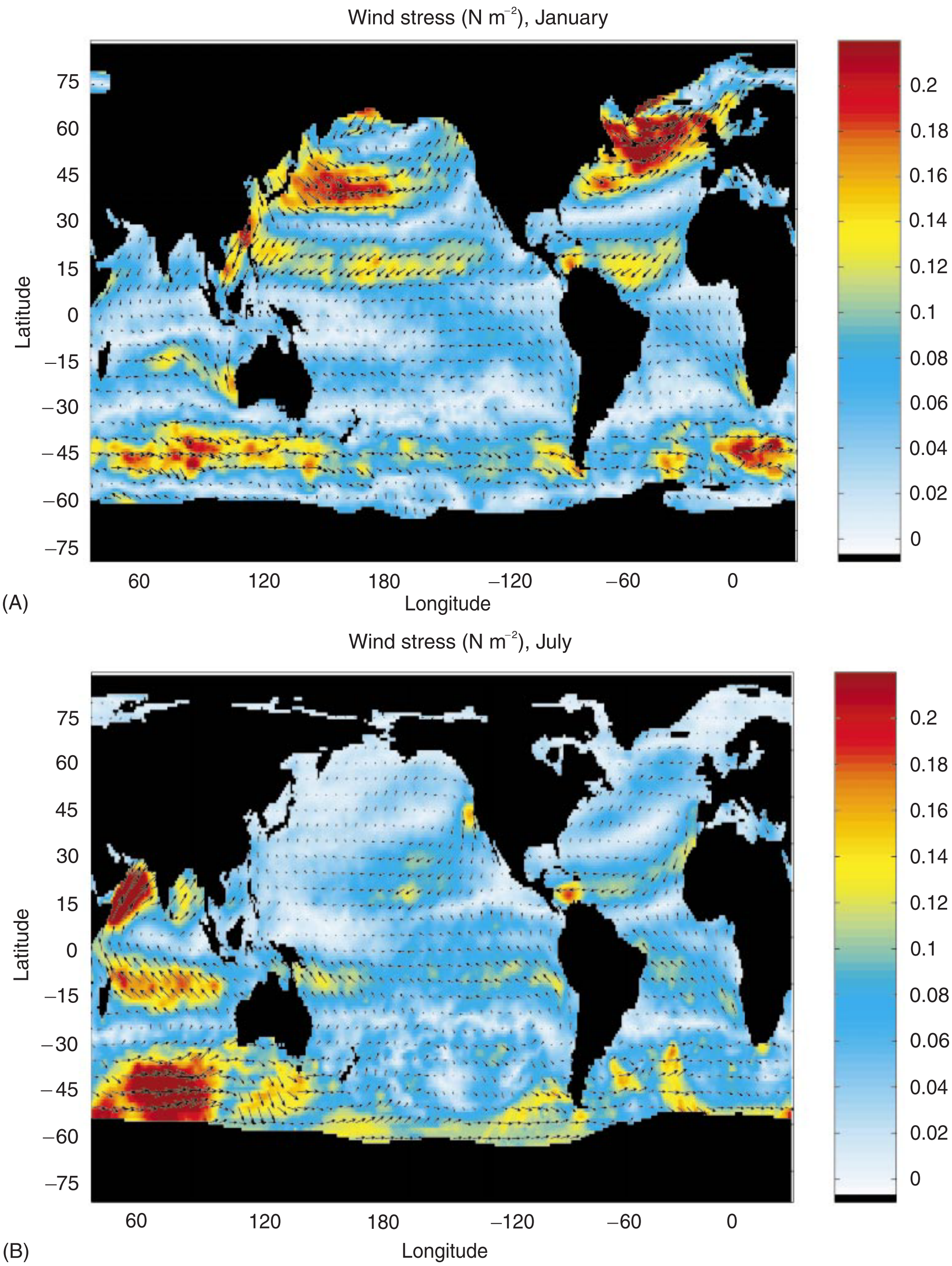

Figure 3 Monthly vector mean wind stress ($N\,m^{-2}$) for (A) January and (B) July calculated from Voluntary Observing Ship weather reports for the period 1980–93. (Adapted with permission from Josey SA, Kent EC and Taylor PK (1998) *The Southampton Oceanography Centre* (*SOC*) *Ocean–Atmosphere Heat, Momentum and Freshwater Flux Atlas*. SOC Report no. 6.)

North Pacific west of Japan. To the south of these regions the extratropical high pressure zones result in low wind stress values, south of these is the north-east trade wind belt. The Inter-Tropical Convergence Zone (ITCZ) with very light winds is close to the equator in both oceans. In the summertime southern hemisphere the south-east trade wind belt is less well marked. The extratropical high pressure regions are extensive but, despite it being summer, high winds and significant wind stress exist in the midlatitude southern ocean. The north-east monsoon dominates the wind patterns in the Indian Ocean and the South China Sea (where it is particularly strong). The ITCZ is a diffuse region south of the equator with relatively strong south-east trade winds in the eastern Indian Ocean.

In northern hemisphere summer (**Figure 3B**) the wind stresses in the midlatitude westerlies are very much decreased. Both the north-east and the south-east trade wind zones are evident respectively to the north and south of the ITCZ. This is predominantly north of the equator. The south-east trades are particularly strong in the Indian Ocean and feed into a very strong south-westerly monsoon flow in the Arabian Sea. The ship data indicate very strong winds in the Southern Ocean south west of Australia. These are also evident in satellite scatterometer data, which suggest that the winds in the Pacific sector of the Southern Ocean, while still strong, are somewhat less than those in the Indian Ocean sector. In contrast the ship data appear to show very light winds. The reason is that in wintertime there are practically no VOS observations in the far south Pacific. The analysis technique used to fill in the data gaps has, for want of other information, spread the light winds of the extratropical high pressure region farther south than is realistic; a good example of the care needed in interpreting the flux maps available in many atlases.

Regional and Seasonal Variation of the Heat Fluxes

The global distribution of the mean annual net heat flux is shown in **Figure 4A**. The accuracy and method of determination of such flux distributions will be discussed further below, here they will be used to give a qualitative description. Averaged over the year the ocean is heated in equatorial regions and loses heat in higher latitudes, particularly in the North Atlantic. However, this mean distribution is somewhat misleading, as the plots for January (**Figure 4B**) and July (**Figure 4C**) illustrate. The ocean loses heat over most of the extratropical winter hemisphere and gains heat in the extratropical summer hemisphere and in the tropics throughout the year. The relative magnitude of the individual flux components is illustrated in **Figure 5** for three representative sites in the North Atlantic Ocean. At the Gulf Stream site (**Figure 5A**) the large cooling in winter dominates the incoming solar radiation in the annual mean. However, even at this site the mean monthly short-wave flux in summer is greater than the cooling. Indeed the effect of the longer daylight periods increases the mean short-wave radiation to values similar to or larger than those observed in equatorial regions (**Figure 5C**). The mid-latitude site (**Figure 5B**) is typical of large areas of the ocean. The ocean cools in winter and warms in summer, in each case by around $100\,\mathrm{W\,m^{-2}}$. The annual mean flux is small – around $10\,\mathrm{W\,m^{-2}}$ – but cannot be neglected because of the very large ocean areas involved. At this site, and generally over the ocean, this annual balance is between the sum of the latent heat flux and net long-wave flux which cool the ocean, and the net short-wave heating. Only in very cold air flows, as over the Gulf Stream in winter, is the sensible heat flux significant.

As regards the interannual variation of the surface fluxes, the major large-scale feature over the global ocean is the El Niño-Southern Oscillation system in the equatorial Pacific Ocean. The changes in the net heat flux under El Niño conditions are around $40\,\mathrm{W\,m^{-2}}$ in the eastern equatorial Pacific. For extra-tropical and midlatitude regions the interannual variability of the summertime net heat flux is typically about $20\text{–}30\,\mathrm{W\,m^{-2}}$, being dominated by the variations in latent heat flux. In winter the typical variability increases to about $30\text{–}40\,\mathrm{W\,m^{-2}}$, although in particular areas (such as over the Gulf Stream) variations of up to $100\,\mathrm{W\,m^{-2}}$ can occur. The major spatial pattern of interannual variability in the North Atlantic is known as the North Atlantic Oscillation (NAO). This represents a measure of the degree to which mobile depressions, or alternatively near stationary high pressure systems, occur in the midlatitude westerly zone.

Accuracy of Flux Estimates

It has been shown that, although the individual flux components are of the order of hundreds of $\mathrm{W\,m^{-2}}$, the net heat flux and its interannual variability over much of the world ocean is around tens of $\mathrm{W\,m^{-2}}$. Furthermore it can be shown that a flux of $10\,\mathrm{W\,m^{-2}}$ over 1 year would, if stored in the top 500 m of the ocean, heat that entire layer by about 0.15°C. Temperature changes on a decadal time scale are at most a few tenths of a degree, so the global mean budget must balance to better than a few $\mathrm{W\,m^{-2}}$. For these various reasons there is a need to measure the flux components, which vary on many time and space scales, to an accuracy of a few $\mathrm{W\,m^{-2}}$. Given the available data sources and methods of determining the fluxes described in the previous sections, it is not surprising that this level of accuracy cannot be achieved at present.

To take an example, in calculating the flux maps shown in **Figure 4** from VOS data many corrections were applied to the VOS observations to attempt to remove biases caused by the methods of observation. For example, air temperature measurements were corrected for the heat island caused by the ship heating up in sunny, low wind conditions. The wind speeds were adjusted depending on the anemometer heights on different ships. Corrections were applied to sea temperatures calculated from engine room intake data. Despite these and other corrections, the

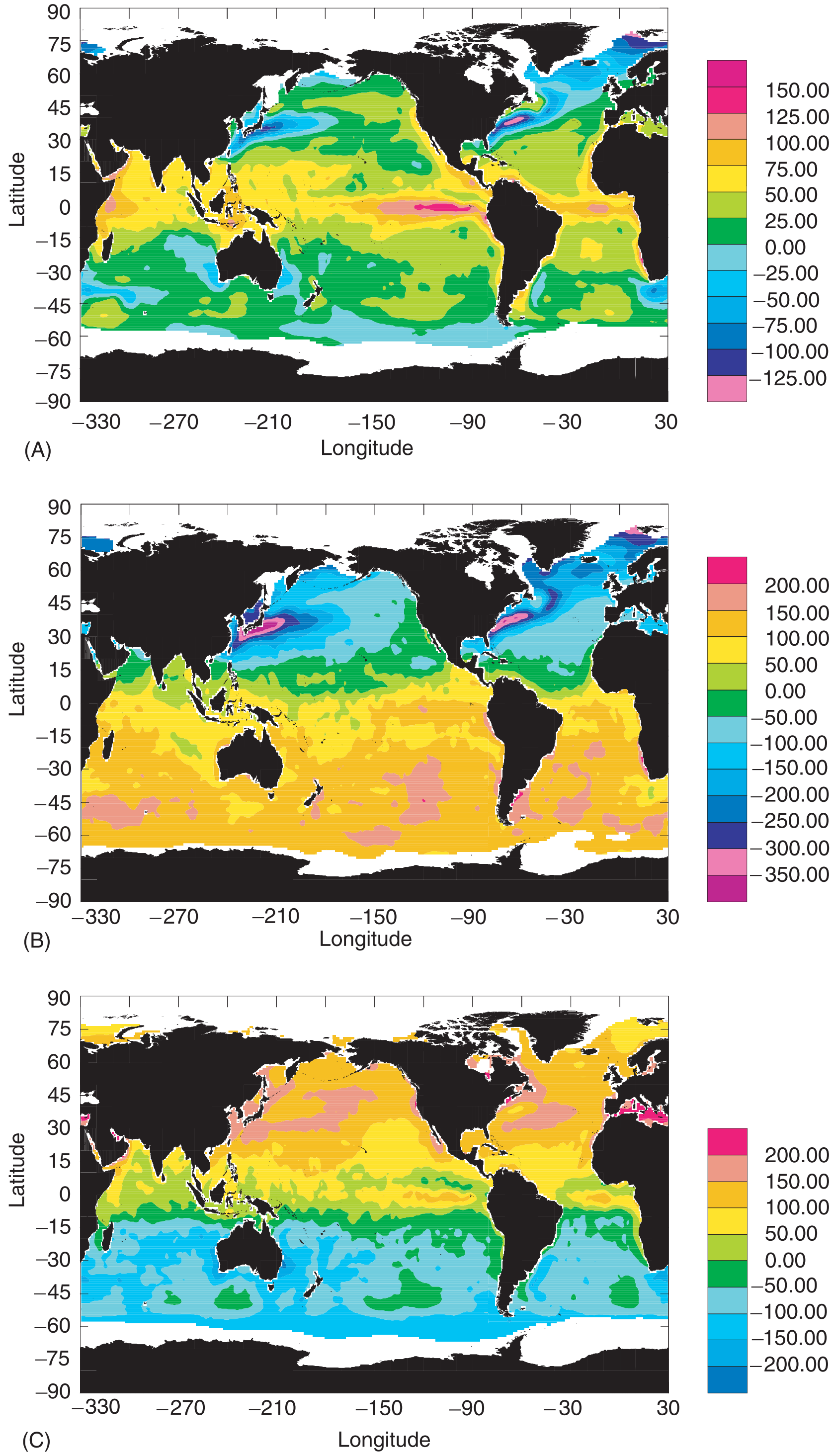

Figure 4 Variation of the net heat flux over the ocean, positive values indicate heat entering the ocean: (A) annual mean, (B) January monthly mean, (C) July monthly mean. (Adapted with permission from Josey SA, Kent EC and Taylor PK (1998) *The Southampton Oceanography Centre* (*SOC*) *Ocean–Atmosphere Heat, Momentum and Freshwater Flux Atlas*. SOC Report no. 6.)

global annual mean flux showed about 30 $W\,m^{-2}$ excess heating of the ocean. Previous climatologies calculated from ship data had shown similar biases and the fluxes had been adjusted to remove the bias, or to make the fluxes compatible with estimates of the meridional heat transport in the ocean. However, comparison of the unadjusted flux data with accurate data from air–sea interaction buoys

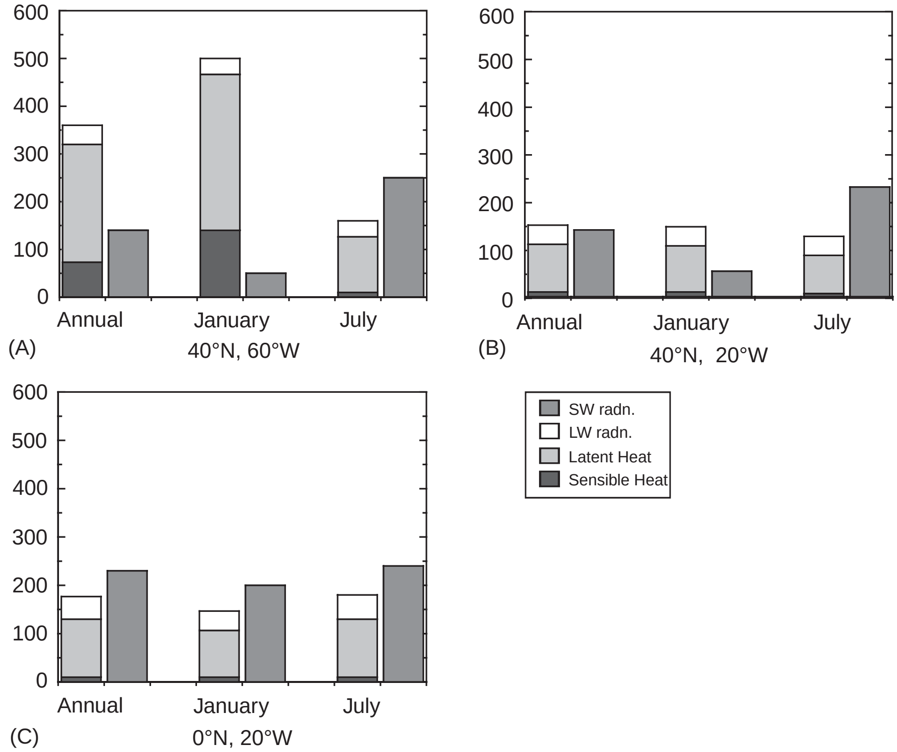

Figure 5 Mean heat fluxes at three typical sites in the North Atlantic for the annual mean, and the January and July monthly means. In each case the left-hand column shows the fluxes which act to cool the ocean while the right-hand column shows the solar heating. (A) Gulf Stream site (40°N, 60°W), (B) midlatitude site (40°N, 20°W), (C) equatorial site (0°N, 20°W).

showed good agreement between the two. This suggests that adjusting the fluxes globally is not correct and that regional flux adjustments are required; however, the exact form of these corrections is presently not shown.

In the future, computer models are expected to provide a major advance in flux estimation. Recently coupled numerical models of the ocean and of the atmosphere have been run for many simulated years during which the modeled climate has not drifted. This suggests that the air–sea fluxes calculated by the models are in balance with the simulated oceanic and atmospheric heat transports. However, it does not imply that the presently estimated flux values are realistic. Errors in the shortwave and latent heat fluxes may compensate one another; indeed in a typical simulation the sea surface temperature stabilized to a value which was, over large regions of the ocean, a few degrees different from that which is observed. Nevertheless the estimation of flux values using climate or NWP models is a rapidly developing field and improvements will doubtless occur in the next few years. There will be a continued need for *in situ* and satellite data for assimilation into the models and for model development and verification. However, it seems very likely that in future the most accurate routine source of the air–sea flux data will be from numerical models of the coupled ocean–atmosphere system.

See also

El Niño Southern Oscillation (ENSO). El Niño Southern Oscillation (ENSO) Models. Evaporation and Humidity. Freshwater Transport and Climate. Heat Transport and Climate. IR Radiometers. North Atlantic Oscillation (NAO). Satellite Passive Microwave Measurements of Sea Ice. Satellite Remote Sensing Microwave Scatterometers. Satellite Remote Sensing of Sea Surface Temperatures. Sensors for Mean Meteorology. Sensors for Micrometeorology and Wind Stress. Turbulence Sensors. Upper Ocean Heat and Freshwater Budgets. Wind Driven Circulation. Wave Energy. Wave Generation by Wind. Wind and Buoyancy-forced Upper Ocean.

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HEAT TRANSPORT AND CLIMATE

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Introduction

It has long been recognized that the ocean plays a significant role in determining the Earth's climate through its interaction with the atmosphere. Since it is the radiation from the sun heating the Earth which drives motion in both these fluid regimes, an understanding of climate and how it evolves is necessarily based upon an understanding of how energy is transported within, and exchanged between, the ocean and the atmosphere.

The Earth receives energy from the sun in the form of short-wave (ultraviolet, visible, and infrared) radiation. Due to the shape and tilt of the Earth, and simple geometry, the intensity of this incoming solar radiation is greatest near the equator and least (but not zero) near the poles (**Figure 1**). The same is true of the intensity of the outgoing long-wave (infrared) radiation returning to space. However, the equator to pole difference is far less for the outgoing radiation (about 50 Wm^{-2}) than for the incoming (>200 Wm^{-2}). Therefore, equatorward of about 35°, the Earth experiences a net warming, while poleward of this latitude, it experiences a net cooling. Because the distribution of incident solar radiation over the Earth is both spatially and temporally uneven, the two systems work together in advecting and exchanging energy to produce an equilibrium, which is why the equatorial regions do not become steadily warmer and the polar regions do not become steadily colder.

To maintain the balance, the ocean–atmosphere system carries a maximum of $\sim$5–6 PW of energy poleward (**Figure 2**). Whether it is the atmosphere or the ocean which is the dominant contributor to this poleward transfer and at what latitudes, is a source of continued debate and further research. However, because the heat capacity and density of the atmosphere are so much less than those of water, the top 2.5 m of the ocean holds as much heat as the entire depth of the atmosphere. It would only take a temperature change of 0.01°C to change the ocean heat content by 1 PW. It is then not surprising that available estimates clearly suggest that the oceanic heat budget is a major component of the climate system, supporting at least half the poleward transfer of energy at some latitudes.

Historically, estimates of oceanic heat transport have presented a broad range of values, but more recently, there appears to be some convergence both in direct estimates (especially in the North Atlantic), as well as in indirect estimates. Satellite data and numerical models now allow us to look at some aspects of ocean heat transport variability, and together the various methods give a broad view of how the oceans contribute to the poleward transfer of energy.

Basic Theory

Estimates of meridional oceanic heat transport derived from observations (i.e., direct estimates) are usually computed from the data using the following formula:

$$\int \rho(T,S,a)c_p\Theta(T,S,a)v(a)\,\mathrm{d}a \qquad [1]$$

where, a is a function of horizontal distance and pressure such that the integral is taken over a vertical slice of the ocean, ρ is density, T is temperature, S is salinity, c_p is specific heat at constant pressure (in practice taken as a single value calculated at a standard T and S), Θ is the potential temperature, and v is the absolute velocity.

Note, however, that what is generally referred to as oceanic heat transport is actually the transport of internal energy. The internal energy is part of the total energy, E, whose transport per unit mass