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Momentum, Heat, and Vapor Fluxes

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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 absorbed in the atmosphere and three times that absorbed by land surfaces. Much of this heat is transferred to the atmosphere by the local sea to air heat flux, a major component of which is caused by the evaporation of water vapor. Although about 90% of the evaporated water falls back into the sea (*see* **Air–Sea Interaction:** Freshwater Flux), the remainder represents about one-third of the precipitation which falls over land. The geographical variation of the 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 poleward and at higher latitudes the sea to air heat transfer significantly ameliorates the climate. Thus the heat, water vapor, and momentum fluxes through the ocean surface form a crucial component of the Earth's climate system.

Having defined the various fluxes and their order of magnitude, this article will review methods of flux measurement and the sources of flux data. The regional and seasonal variation of the fluxes will be summarized. Following a discussion of the accuracy of our present flux estimates, the potential for future improvements will be considered.

Definition of the Fluxes

The momentum flux is the downward transfer of horizontal momentum caused by the drag of the sea surface on the wind. The wave-covered sea surface is continually in motion and, compared with typical land surfaces, appears remarkably smooth to the air flow. For gale force winds the waves may be 10 m or more in height, but the momentum flux is no more than that which would occur over a flat plain. As a result, wind speeds over the ocean tend to be greater than those over land.

The total heat transfer through the ocean surface, the net heat flux, is a combination of several components. The *shortwave radiative flux* (wavelength 0.3–3 μm) is the heat input from the Sun. Around noon on a sunny day this flux may reach about $1000\,\mathrm{W\,m^{-2}}$ but, when averaged over 24 hours, a typical value is $100–300\,\mathrm{W\,m^{-2}}$, varying with latitude and season. Depending on the solar elevation and the sea state, about 6% of the incident radiation is reflected from the sea surface. Most of the remainder is absorbed in the upper few meters of the ocean. In calm weather, with winds less that about $3\,\mathrm{m\,s^{-1}}$, a shallow layer may form during the day in which the sea has been 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 meters. Thus, in contrast to land areas, the typical day to night variation in sea and air temperatures is small, typically less than 1°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 longwave flux* acts to cool the sea surface, typically by $30\,\mathrm{W\,m^{-2}}$ (if cloudy) to $80\,\mathrm{W\,m^{-2}}$ (clear skies).

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 the 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, in regions like the summertime North Pacific Ocean, warm winds blowing over a colder ocean may result in a small sensible heat flux into the ocean. Under most weather conditions the evaporation of water vapor from the sea surface results in a water vapor flux from the sea to the air. The latent heat flux is the heat absorbed on vaporization of the water. This heat is released to warm the atmosphere when the vapor condenses to form clouds. Usually the latent heat 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, condensation may occur on the sea surface, and the vapor flux and latent heat flux are directed from air to sea. In summertime over the fog-shrouded Grand Banks off Newfoundland, the mean monthly latent heat transfer is directed into the ocean, but this is a very exceptional case.

Measuring the Fluxes

For the radiative fluxes, the standard method is to measure the voltage generated by a thermopile exposed to the incident radiation. A pyranometer, mounted in gimbals for use on a ship or buoy, is used to measure the incoming shortwave radiation (**Figure 1**). For better accuracy the direct and diffuse components should be determined separately but at present this is rarely done over the sea. The reflected short-wave radiation is normally calculated using tabulated values of the albedo for different solar elevations. Pyrgeometers, used for longwave radiation, are similar to pyranometers but have a coated dome to filter out the shortwave radiation. For these the use of gimbals is less important, but a clear sky view is required and corrections for the dome temperature and short wave leakage are needed. Again, only the downward component is normally measured; the upward component is calculated from the temperature and the emissivity of the sea surface.

The turbulent fluxes may be measured directly in the near surface atmosphere using the eddy correlation method. For example, if upward-moving air in the turbulent eddies is on average warmer and moister than the downward-moving air, then there is an upward flux of sensible heat and water vapor and hence also of latent heat. Similarly, the momentum

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

flux may be determined from the correlation between horizontal and vertical wind fluctuations. Accurate eddy correlation measurements over the ocean are difficult. 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 order 30 minutes for each flux determination. For instrumentation mounted on a buoy or ship the six components of the wave-induced motion must be measured and removed in the signal processing procedure. The distortion both of the turbulence and of the mean wind by ship, buoy, or fixed tower must be minimized and, if possible, corrected for. While three-component ultrasonic anemometers (**Figure 2**) are relatively robust, the sensors for measuring fluctuations in temperature and humidity have previously been fragile and, in the marine atmosphere, prone to contamination by salt particles and sea spray. Improved sonic thermometry, and water vapor instruments using microwave refractometry or differential infrared absorption, are relatively recent developments. Thus eddy correlation measurements are not routinely obtained over the ocean; rather they are used in air–sea interaction experiments to calibrate other flux estimation methods.

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.)

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 regarding the formation and dissipation of turbulent quantities that may not always be valid. It has been implemented on some research ships to increase the range of available flux data.

The bulk (aerodynamic) formulas are the most commonly used method of flux estimation. The flux is determined from the difference between the temperature, humidity or wind 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 non-wind-induced water current). Thus the flux F_x of some quantity x is given by eqn [1], where ρ is the air density and U_z is the wind speed at the measurement height.

$$F_x = \rho U_z C_{xz}(x_z - x_0) \qquad [1]$$

While appearing intuitively correct (for example, blowing over a hot drink will cool it faster) these formulas 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. The transfer coefficient 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 are shown in **Table 1**.

Many research problems remain. For example: the increase of C_{D10n} at higher wind speeds must depend on the varying sea state, but can the latter be successfully characterized by the ratio of the predominant wave speed to the wind speed (the wave age), or by factors like the wave height and steepness, or is a complete spectral representation of the wave field required? What are the effects of waves propagating from other regions at varying angles to the wind (i.e., swell waves)? What is the exact behaviour of C_{D10n} in low wind speed conditions? Since C_{E10n} and C_{H10n} are not well defined by the available experimental data, recent implementations of the bulk algorithms have used theoretical models of the ocean surface (known as surface renewal theory) to predict these quantities from the momentum roughness length.

Table 1 Typical values (with estimated uncertainties) for the transfer coefficients. 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

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

Sources of Flux Data

Until recent years the only routinely available sources of data were the weather reports from merchant ships. Organized as part of the World Weather Watch system of the World Meteorological Organization, these voluntary observing ships (VOS) are asked to transmit coded weather messages at 00.00, 06.00, 12.00 and 18.00 GMT daily, and to record a more detailed observation in a weather logbook. The very basic instruments used normally include a barometer and a means of measuring air temperature and humidity: wet and dry bulb thermometers mounted in a hand-swung sling psychrometer or in a fixed, louvered Stevenson screen. Sea temperature is obtained using a thermometer and an insulated bucket, or by reading the temperature of 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 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 considered inferior to anemometer-based values.

The bulk formulas are used to calculate the turbulent fluxes from the VOS observations. However, in many cases the accuracy is poor. In particular, a large ship can produce significant changes in the local temperature and wind flow. The radiative fluxes must be estimated from the observer's estimate of the cloud amount plus, for short wave, the solar elevation, or for long wave, the sea and air temperature and humidity. The unavoidable observational errors and the crude form of the radiative flux formulas imply that large numbers of reports are needed, and correction schemes must be applied, before satisfactory flux estimates can be obtained. Though there are presently nearly 7000 VOS, they tend to be concentrated in the main shipping lanes. While coverage over 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 are only capable of providing useful values if averaged over several years, and reports from the Southern Ocean are very few indeed. These problems must be borne in mind when studying the flux distribution maps shown in marine climatological atlases, of which examples are presented below.

Satellite-borne sensors can overcome these sampling problems. 'Passive' sensors measure the radiation emitted from the sea surface and the intervening atmosphere at visible, infrared, or microwave frequencies; 'active' sensors transmit microwave radiation and measure the returned signal. The problem is to develop methods of determining the fluxes from the various satellite data that can be obtained. For example, sea surface temperature has been routinely determined using visible and infrared radiometers since about 1980. However, the data must be frequently checked against ship and buoy values to avoid errors due to changes in atmospheric aerosol content that may follow volcanic eruptions. Satellite-derived fields of net surface shortwave radiation are available; values for the net surface longwave radiation are less accurate. The surface wind velocity can be determined to good accuracy by active scatterometer sensors by measuring the microwave radiation back-scattered from the sea surface. The determination of near-surface air temperature and humidity from satellites is hindered by the relatively coarse vertical resolution of the retrieved data. Thus 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 developed. More recently, neural network techniques have been applied to retrieving both air temperature and humidity; however, there is presently no routinely available product. Thus the satellite flux products for which useful accuracy has been demonstrated on a global basis 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 part of the calculations. Assimilating much of the available data from the World

Weather Watch system, including satellite data, radiosonde profiles, and surface observations, NWP models are potentially the best source of flux data. However, there are a number of problems. The vertical resolution of these models is relatively poor and many of the near-surface processes have to be represented in terms of larger-scale parameters. Improvements to NWP models are judged on the resulting quality of the weather forecasts, not on the accuracy of the surface fluxes, which may become worse. Indeed, the continual introduction of model changes results in time discontinuities in the output variables. Thus the determination of interannual variations is difficult and, for that reason, centers such as the European Centre for Medium Range Weather Forecasting (ECMWF) and the US National Centers for Environmental Prediction (NCEP) have recently reanalyzed the past weather over 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 are more accurate. Particular weaknesses for the models are in the shortwave radiation and latent heat fluxes. New reanalyses are underway and efforts are being made to improve the flux estimates; eventually these reanalyses will provide the best source of flux data for many purposes.

The 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. The major features of the wind stress variability derived from ship observations from the period 1980 to 1993 are summarized here, using plots for January and July to illustrate the seasonal variation (**Figure 3**).

In Northern Hemisphere winter (**Figure 3A**) large wind stresses due to strong mid-latitude westerly winds occur in the North Atlantic and the North Pacific west of Japan. To the south, the extratropical high-pressure zones result in low wind stress values, and south of these is the belt of north-east trade winds. The very light winds of the Intertropical Convergence Zone (ITCZ) lie close to the Equator in both oceans. In the summertime Southern Hemisphere the south-east trade wind belt is less pronounced. The extratropical high-pressure regions are extensive but, despite it being summer, high winds and significant wind stress exist in the mid-latitude Southern Ocean. The north-east monsoon dominates the wind patterns in the Indian Ocean and the South China Sea (where it is particularly strong). In the latter regions the ITCZ is a diffuse area 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 mid-latitude 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, which mainly lies north of the Equator. The south-east trades are particularly strong in the Indian Ocean and feed air across the Equator into a very strong south-westerly monsoon flow in the Arabian Sea. These ship data indicate very strong winds in the Southern Ocean south-west of Australia. Such winds 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 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; this is a good example of the care needed in interpreting the flux maps available in many atlases.

The Regional and Seasonal Variation of the Heat Fluxes

The global distribution of the mean annual net heat flux is shown in **Figure 4A**. 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. It is only because the tropical oceans are heated throughout the year, and atmospheric moisture from trade wind zones converges in the ITCZ, that the tropics are so important with regard to driving the atmospheric circulation. The major regions of ocean cooling occur in winter over the Gulf Stream and the Kuroshio currents. However, in summer the long period of daylight in these mid-latitude regions results in mean short-wave radiation values similar to or larger than those observed in equatorial regions. Thus the mean monthly short-wave flux is greater than the cooling induced by the combined latent heat and net long-wave fluxes. At a more typical mid-latitude site the ocean cools in winter and warms in summer, in each case by around $100\,\mathrm{W\,m^{-2}}$. The annual mean flux is

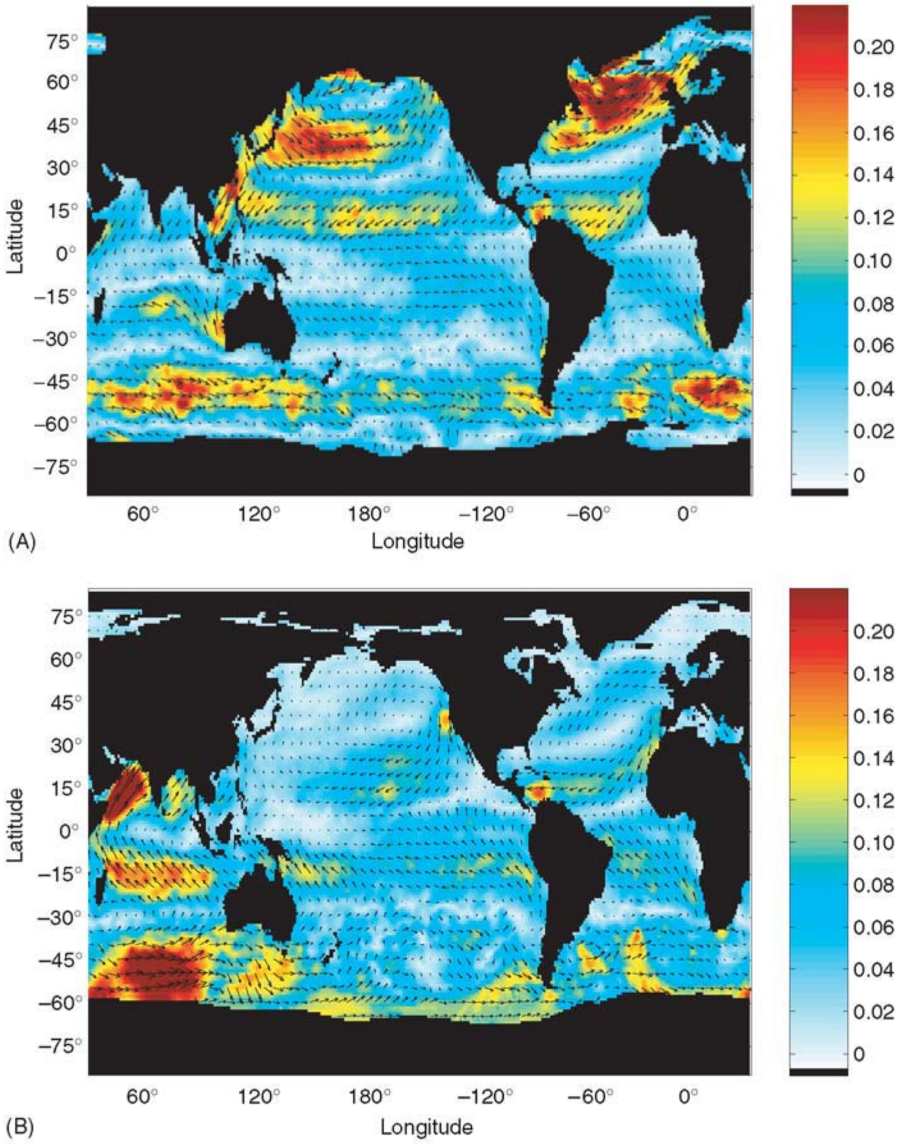

Figure 3 Monthly vector mean wind stress (N m^{-2}) for (A) January and (B) July calculated from voluntary observing ship weather reports for the period 1980 to 1993. (Adapted from Josey SA, Kent EC, and Taylor PK (1998). *The Southampton Oceanography Centre* (*SOC*) *Ocean–Atmosphere Heat, Momentum and Freshwater Flux Atlas*, SOC Report No. 6.)

small, of order 10 W m^{-2}, but cannot be neglected because of the very large ocean areas involved.

Considering now 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. In the eastern equatorial Pacific the change in the net heat flux under El Niño conditions is around 40 W m^{-2}. For extra-tropical and mid-latitude regions the interannual variability of the summertime net heat flux is typically about 20–30 W m^{-2}, being dominated by the variations in latent heat flux. In winter the typical variability increases to about 30–40 W m^{-2}, although in particular areas (such as over the Gulf Stream) variations of up to 100 W m^{-2} may 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 mid-latitude westerly zone.

Discussion: Accuracy of Flux Estimates and Future Trends

We have seen that, although the individual flux components are of the order of hundreds of W m^{-2},

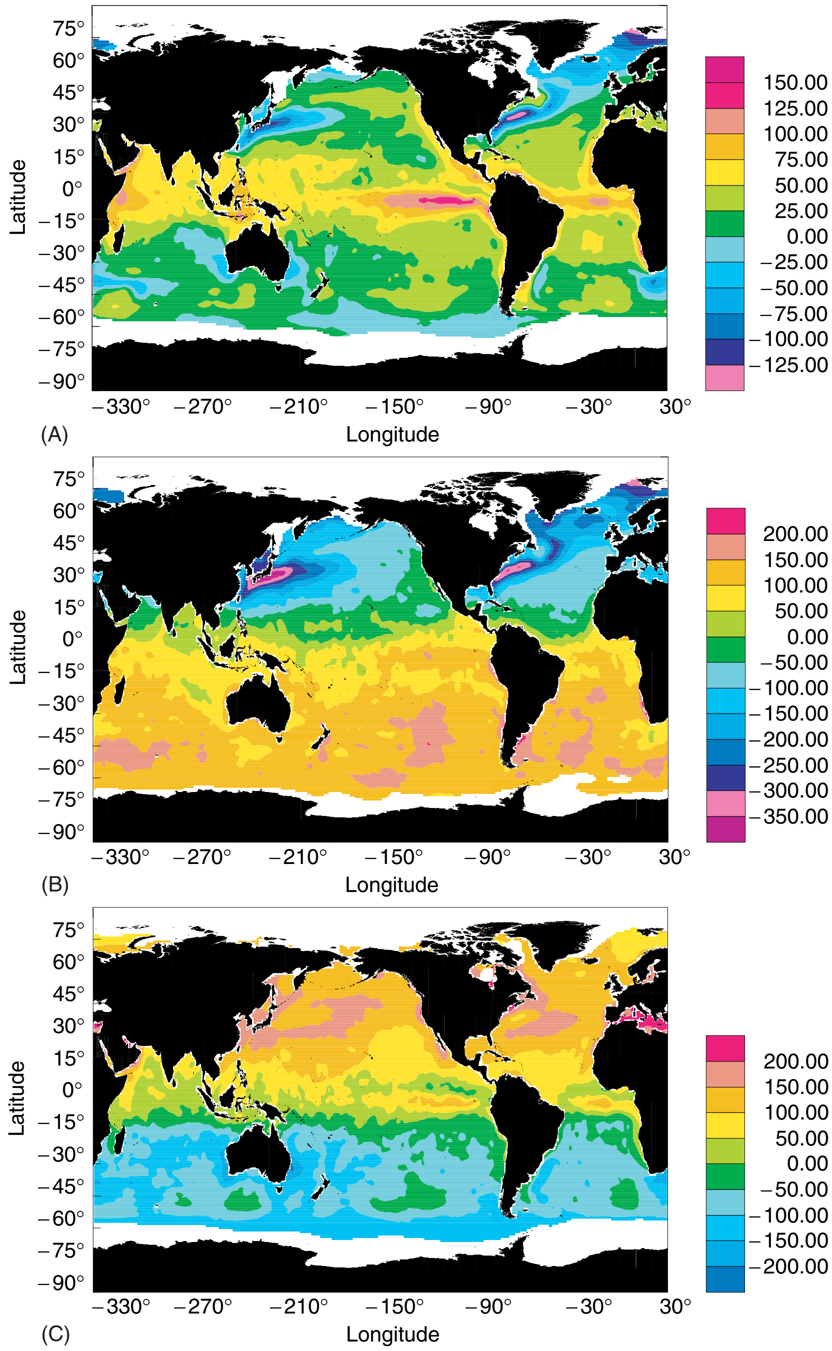

Figure 4 Variation of the net heat flux over the ocean; positive values indicate heat entering the ocean: (A) annual mean; (B) January monthly mean; (C) July monthly mean. (Adapted from Josey SA, Kent EC, and Taylor PK (1998). *The Southampton Oceanography Centre* (*SOC*) *Ocean–Atmosphere Heat, Momentum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

the net heat flux and its interannual variability over much of the world ocean is of the order of tens of $\mathrm{W\,m^{-2}}$. Furthermore, it can be shown that a flux of $10\,\mathrm{W\,m^{-2}}$ over one 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 above, it is not surprising that this accuracy at present cannot be achieved.

To take an example, in calculating the flux maps shown in **Figure 4** many corrections were applied to the VOS observations in an attempt to remove biases caused by the observing methods. 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 global annual mean flux showed about $30\,\mathrm{W\,m^{-2}}$ 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 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 known.

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 at present the flux values are realistic. Errors in the short-wave and latent heat fluxes may compensate one another; indeed, in a typical simulation the sea surface temperature stabilized to a value that 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 have occurred by the time this article has been published. 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 air–sea flux estimates will be from numerical models of the coupled ocean–atmosphere system.

See also

Aerosols: Observations and Measurements; Role in Radiative Transfer. **Air–Sea Interaction:** Freshwater Flux; Sea Surface Temperature; Surface Waves. **Boundary Layers:** Observational Techniques *In Situ*; Observational Techniques–remote; Surface Layer. **Buoyancy and Buoyancy Waves:** Optical Observations. **Climate Variability:** North Atlantic and Arctic Oscillation. **Coupled Ocean–Atmosphere Models**. **El Niño and the Southern Oscillation:** Observation. **Reflectance and Albedo, Surface**. **Weather Prediction:** Regional Prediction Models.

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Sea Surface Temperature

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Introduction

As the controlling variable of heat, momentum, salt, and gas fluxes between the ocean and the atmosphere,