Chapter 5

The Oceanic Heat Budget

About half the sunlight reaching Earth is absorbed by the oceans and land, where it is temporarily stored near the surface. Only about a fifth of the available sunlight is directly absorbed by the atmosphere. Of the heat stored by the ocean, part is released to the atmosphere, mostly by evaporation and infrared radiation. The remainder is transported by currents to other areas especially high latitudes in winter. Solar radiation stored in the ocean is therefore available to ameliorate Earth's climate. The transport of heat is not steady, and significant changes in heat transport, particularly in the Atlantic, may have been important for the development of the ice ages. For these reasons, oceanic heat budgets and transports are important for understanding Earth's climate and its short and long term variability.

5.1 The Oceanic Heat Budget

Changes in heat stored in the upper layers of the ocean result from a local imbalance between input and output of heat through the sea surface. The flux of heat to deeper layers is usually much smaller than the flux through the surface. Advection out of the area will be described later, and it too tends to be small, provided the box covers a large enough area. Globally, the flux must balance, otherwise the oceans as a whole would warm or cool.

The sum of the changes in heat fluxes into or out of a volume of water is the *heat budget*. The major terms in the budget at the sea surface are:

- 1. Insolation Q_{SW} , the flux of sunlight into the sea;
- 2. Net Infrared Radiation Q_{LW} , net flux of infrared radiation from the sea;
- 3. Sensible Heat Flux Q_S , the flux of heat through the surface due to conduction;
- 4. Latent Heat Flux Q_L , the flux of heat carried by evaporated water; and
- 5. Advection Q_V , heat carried by currents.

Conservation of heat requires:

$$Q_T = Q_{SW} + Q_{LW} + Q_S + Q_L + Q_V (5.1)$$



Figure 5.1 Specific heat of sea water at atmospheric pressure C_p in joules per gram per degree Celsius as a function of temperature in Celsius and salinity, calculated from the empirical formula given by Millero *et al.* (1973) using algorithms in Fofonoff and Millard (1983). The lower line is the freezing point of salt water.

where Q_T is the resultant heat gain or loss. Units for heat fluxes are watts/m². The product of flux times surface area times time is energy in joules. The change in temperature ΔT of the water is related to change in energy ΔE through:

$$\Delta E = C_p \, m \, \Delta T \tag{5.2}$$

where m is the mass of water being warmed or cooled, and C_p is the specific heat of sea water at constant pressure.

$$C_p \approx 4.0 \times 10^3 \text{ J} \cdot \text{kg}^{-1} \cdot {}^{\circ}\text{C}^{-1}$$

$$(5.3)$$

Thus, 4,000 Joules of energy are required to heat 1.0 kilogram of sea water by 1.0° C (Figure 5.1).

Importance of the Ocean in Earth's Heat Budget To understand the importance of the ocean in Earth's heat budget, let's make a simple comparison of the heat stored in the ocean with heat stored on land during an annual cycle. During the cycle, heat is stored in summer and released in the winter. The point is to show that the oceans store and release much more heat than the land.

To begin, we use (5.3) and the heat capacity of soil and rocks

$$C_{p(rock)} = 800 \text{ J} \cdot \text{kg}^{-1} \cdot {}^{\circ}\text{C}^{-1}$$

$$(5.4)$$

to obtain $C_{p(rock)} \approx 0.2 C_{p(water)}$.

The volume of water which exchanges heat with the atmosphere on a seasonal cycle is 100 m^3 per square meter of surface, i.e. that mass from the surface to a depth of 100 meters. The density of water is 1000 kg/m^3 , and the mass in

contact with the atmosphere is density × volume = $m_{water} = 100,000$ kg. The volume of land which exchanges heat with the atmosphere on a seasonal cycle is 1 m³. Because the density of rock is 3,000 kg/m³, the mass of the soil and rock in contact with the atmosphere is 3,000 kg.

The seasonal heat storage values for the ocean and land are therefore:

$$\begin{split} \Delta E_{oceans} &= C_{p(water)} \, m_{water} \, \Delta T \qquad \Delta T = 10^{\circ} \mathrm{C} \\ &= (4000)(10^5)(10^{\circ}) \, \mathrm{Joules} \\ &= 4.0 \times 10^9 \, \mathrm{Joules} \\ \Delta E_{land} &= C_{p(rock)} \, m_{rock} \, \Delta T \qquad \Delta T = 20^{\circ} \mathrm{C} \\ &= (800)(3000)(20^{\circ}) \, \mathrm{Joules} \\ &= 4.8 \times 10^7 \, \mathrm{Joules} \\ \frac{\Delta E_{oceans}}{\Delta E_{land}} &= 100 \end{split}$$

where ΔT is the typical change in temperature from summer to winter.

The large storage of heat in the ocean compared with the land has important consequences. The seasonal range of air temperatures on land increases with distance from the ocean, and it can exceed 40° C in the center of continents, reaching 60° C in Siberia. Typical range of temperature over the ocean and along coasts is less than 10° C. The variability of water temperatures is still smaller (see figure 6.5).

5.2 Heat-Budget Terms

Let's look at the factors influencing each term in the heat budget.

Factors Influencing Insolation Incoming solar radiation is primarily determined by latitude, season, time of day, and cloudiness. The polar regions are heated less than the tropics, areas in winter are heated less than the same area in summer, areas in early morning are heated less than the same area at noon, and cloudy days have less sun than sunny days.

The following factors are important:

- 1. The height of the sun above the horizon, which depends on latitude, season, and time of day. Don't forget, there is no insolation at night!
- 2. The length of day, which depends on latitude and season.
- 3. The cross-sectional area of the surface absorbing sunlight, which depends on height of the sun above the horizon.
- 4. Attenuation, which depends on:
 - Path length through the atmosphere, which varies as $\csc \varphi$, where φ is angle of the sun above the horizon.
 - Clouds, which absorb and scatter radiation.
 - Gas molecules which absorb radiation in some bands. H₂O, O₃, and CO₂ are all important.



Figure 5.2 Insolation (spectral irradiance) of sunlight at top of the atmosphere and at the sea surface on a clear day. The dashed line is the best-fitting curve of blackbody radiation the size and distance of the sun. The number of standard atmospheric masses is designated by m. Thus m = 2 is applicable for sunlight when the sun is 30° above the horizon. (From Stewart, 1985).

- Aerosols which scatter and absorb radiation. Both volcanic and marine aerosols are important.
- Dust, which scatters radiation, especially Saharan dust over the Atlantic.
- 5. Reflectivity of the surface, which depends on solar elevation angle and roughness of sea surface.

Solar inclination and cloudiness dominate. Absorption by ozone and water vapor are much weaker.

Figure 5.2 shows insolation above the atmosphere, and at the surface on a clear day with the sun 30° above the horizon. Figure 5.3 gives the insolation at the sea surface for a cloud-free atmosphere, including loss by reflection at the surface and absorption by clear air.

The average annual value for insolation is in the range:

$$30 \text{ W/m}^2 < Q_{SW} < 260 \text{ W/m}^2$$
 (5.5)

Factors Influencing Infrared Flux The sea surface radiates as a blackbody having the same the temperature as the water, which is roughly 290 K. The



Figure 5.3 Monthly average of clear-sky, downward flux of sunlight through the sea surface in W/m^2 during 1989 calculated by the Satellite Data Analysis Center at the NASA Langley Research Center (Darnell et al. 1992) using data from the International Satellite Cloud Climatology Project.

distribution of radiation as a function of wavelength is given by the Planck's equation. Sea water at 290 K radiates most strongly at wavelengths near 10 μ m. These wavelengths are strongly absorbed by clouds, and somewhat by water vapor. A plot of atmospheric transmittance as a function of wavelength for a clear atmosphere but with varying amounts of water vapor (Figure 5.4) shows that the atmosphere has various windows with high transmittance.

The transmittance on a cloud-free day through the window from 8 μ m to 13 μ m is determined mostly by water vapor. Absorption in other bands, such as those at 3.5 μ m to 4.0 μ m depends on CO₂ concentration in the atmosphere. As the concentration of CO₂ increases, these windows close and more radiation is trapped by the atmosphere.

Because the atmosphere is mostly transparent to incoming sunlight, and somewhat opaque to outgoing infrared radiation, the atmosphere traps radiation, keeping Earth's surface 33° warmer than it would be in the absence of an atmosphere but in thermal equilibrium with space. The atmosphere acts like the panes of glass on a greenhouse, and the effect is known as the *greenhouse effect*. See Hartmann (1994: 24–26) for a simple discussion of the radiative balance of a planet. CO₂, water vapor, methane, and ozone are all important greenhouse gasses.

The net infrared flux depends on:

1. The clarity of the atmospheric window, which depends on:

- Clouds thickness. The thicker the cloud deck, the less heat escapes to space.
- Cloud height, which determines the temperature at which the cloud radiates heat back to the ocean. The rate is proportional to T^4 ,



Figure 5.4 Atmospheric transmittance for a vertical path to space from sea level for six model atmospheres with very clear, 23 km, visibility, including the influence of molecular and aerosol scattering. Notice how water vapor modulates the transparency of the 10-14 μ m atmospheric window, hence it modulates Q_{LW} , which is a maximum at these wavelengths. (From Selby and McClatchey, 1975).

where T is the temperature of the radiating body in Kelvins. High clouds are colder than low clouds.

- Atmospheric water-vapor content. The more humid the atmosphere the less heat escapes to space.
- 2. Water Temperature. The hotter the water the more heat is radiated. Again, radiation depends of T^4 .
- 3. Ice and snow cover. Ice emits as a black body, but it cools much faster than open water. Ice-covered seas are insulated from the atmosphere.

Changes in water vapor and clouds are more important for determining back radiation than are changes in surface temperature. Hot tropical regions lose less heat than cold polar regions. The temperature range from poles to equator is $0^{\circ}C < T < 25^{\circ}C$ or 273K < T < 298K and the ratio of maximum to minimum temperature in Kelvins is 298/273 = 1.092. Raised to the fourth power this is 1.42. Thus there is a 42% increase in emitted radiation from pole to equator. Over the same distance water vapor can change the net emitted radiance by 200%.

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The average annual value for net infrared flux is in the narrow range:

$$-60 \text{ W/m}^2 < Q_{LW} < -30 \text{ W/m}^2 \tag{5.6}$$

Factors Influencing Latent-Heat Flux Latent heat flux is influenced primarily by wind speed and relative humidity. High winds and dry air evaporate much more water than weak winds with relative humidity near 100%. In polar regions, evaporation from ice covered oceans is much less than from open water. Open water is often much warmer than the air, hence saturated air near the water surface has much higher water content than air higher in the atmospheric boundary layer. In the arctic, most of the heat lost from the sea is through leads (ice-free areas). Hence the percent open water is very important for the arctic heat budget.

The average annual value for latent-heat flux is in the range:

$$-130 \text{ W/m}^2 < Q_L < -10 \text{ W/m}^2 \tag{5.7}$$

Factors Influencing Sensible-Heat Flux Sensible heat flux is influenced primarily by wind speed and air-sea temperature difference. High winds and large temperature differences cause high fluxes. Think of this as a wind-chill factor for the oceans.

The average annual value for sensible-heat flux is in the range:

$$-42 \text{ W/m}^2 < Q_S < -2 \text{ W/m}^2 \tag{5.8}$$

5.3 Direct Calculation of Fluxes

Before we can describe the geographical distribution of fluxes into and out of the ocean, we need to know how they are measured or calculated.

Gust-Probe Measurements of Turbulent Fluxes There is only one accurate method for calculating fluxes of sensible and latent heat and momentum at the sea surface: from direct measurement of turbulent quantities in the atmospheric boundary layer made by gust probes on low-flying aircraft or offshore platforms. Very few such measurements have been made. They are expensive, and they cannot be used to calculate heat fluxes averaged over many days or large areas. The gust-probe measurements are used only to calibrate other methods of calculating fluxes.

- 1. Measurements must be made in the surface layer of the atmospheric boundary layer (See §4.3), usually within 30 m of the sea surface, because fluxes are independent of height in this layer.
- 2. Measurements must be made by fast-response instruments (gust probes) able to make several observations per second on a tower, or every meter from a plane.
- 3. Measurements include the horizontal and vertical components of the wind, the humidity, and the air temperature.

Symbol	Variable	Value and Units
$\overline{C_p}$	Specific heat capacity of air	$1030 \text{ J} \cdot \text{kg}^{-1} \cdot \text{K}^{-1}$
C_D	Drag coefficient (see 4.3)	$(0.60 + 0.070 U_{10}) \times 10^{-3}$
C_L	Latent heat transfer coefficient	1.2×10^{-3}
C_S	Sensible heat transfer coefficient	1.0×10^{-3}
L_E	Latent heat of evaporation	$2.5 \times 10^6 \text{ J/kg}$
q	Specific humidity of air	kg (water vapor)/kg (air)
q_a	Specific humidity of air 10 m above the sea	kg (water vapor)/kg (air)
q_s	Specific humidity of air at the sea surface	kg (water vapor)/kg (air)
Q_S	Sensible heat flux	W/m^2
Q_L	Latent heat flux	W/m^2
T_a	Temperature of the air 10 m above the sea	K or $^{\circ}C$
T_s	Sea-surface temperature	K or $^{\circ}C$
t'	Temperature fluctuation	$^{\circ}C$
u'	Horizontal component of fluctuation of wind	m/s
u_*	Friction velocity	m/s
U_{10}	Wind speed at 10 m above the sea	m/s
w'	Vertical component of wind fluctuation	m/s
ho	Density of air	1.5 kg/m^3
T	Vector wind stress	Pa

 Table 5.1 Notation Describing Fluxes

Fluxes are calculated from the correlation of vertical wind and horizontal wind, humidity, or temperature: Each type of flux is calculated from different measured variables, u', w', t', and q':

$$T = \langle \rho \, u' w' \rangle = \rho \, \langle u' w' \rangle \equiv \rho \, u_*^2 \tag{5.9a}$$

$$Q_S = C_p \left\langle \rho \, w' t' \right\rangle = \rho \, C_p \left\langle w' t' \right\rangle \tag{5.9b}$$

$$Q_L = L_E \left\langle w'q' \right\rangle \tag{5.9c}$$

where the brackets denotes time or space averages, and the notation is given in Table 5.1. Note that *specific humidity* mentioned in the table is the mass of water vapor per unit mass of air.

Radiometer Measurements of Radiative Fluxes Radiometers on ships, offshore platforms, and even small islands are used to make direct measurements of radiative fluxes. Wideband radiometers sensitive to radiation from 0.3 μ m to 50 μ m can measure incoming solar and infrared radiation with an accuacy of around 3% provided they are well calibrated and maintained. Other, specialized radiometers can measure the incoming solar radiation, the downward infrared radiation, and the upward infrared radiation. Usually, however the upward infrared radiation is calculated from the measured surface temperature of the sea. This is more accurate than measuring the radiation.

Radiometer errors are due to salt spray and rime on the aperture, failure to keep the instrument horizontal, and variations in heat loss due to wind on the instrument (Hinzpeter, 1980).

5.4 Indirect Calculation of Fluxes: Bulk Formulas

The use of gust-probes is very expensive, and radiometers must be carefully maintained. Neither can be used to obtain long-term, global values of fluxes. To calculate these fluxes from practical measurements, we use observed correlations between fluxes and variables that can be measured globally.

For fluxes of sensible and latent heat and momentum, the correlations are called *bulk formulas*. They are:

$$T = \rho C_D U_{10}^2 \tag{5.10a}$$

$$Q_S = \rho C_p C_S U_{10} (T_s - T_a)$$
(5.10b)

$$Q_L = \rho L_E C_L U_{10} (q_s - q_a)$$
(5.10c)

Air temperature T_a is measured using thermometers on ships. It cannot be estimated from space using satellite instruments. T_s is measured using thermometers on ships or from space using infrared radiometers such as the AVHRR.

The specific humidity of air is the mass in kilograms of water vapor in a kilogram of air. The specific humidity of air at 10 m above the sea surface q_a is calculated from measurements of relative humidity made from ships. Gill (1982: pp: 39–41, 43–44, & 605–607) describes equations relating water vapor pressure, vapor density, and specific heat capacity of wet air. The specific humidity at the sea surface q_s is calculated from T_s assuming the air at the surface is saturated with water vapor. U_{10} is measured or calculated using the instruments or techniques described in Chapter 4.

The coefficients C_D , C_S and C_L are calculated by correlating direct measurements of fluxes made by gust probes with the variables in the bulk formulas. Smith (1998) investigated the accuracy of published values for the coefficients, and the values for C_S and C_L in table 5.1 are his suggested values. Smith (1998) also gives fluxes as a function of observed variables in tabular form. Liu, Katsaros, and Businger (1979) discuss alternate bulk formulas. Note that wind stress is a vector with magnitude and direction. It is parallel to the surface in the direction of the wind.

The problem now becomes: How to calculate the fluxes across the sea surface required for studies of ocean dynamics? The fluxes include: 1) stress; 2) solar heating; 3) evaporation; 4) net infrared radiation; 5) rain; 5) sensible heat; and 6) others such as CO₂ and particles (which produce marine aerosols). Furthermore, the fluxes must be accurate. We need an accuracy of approximately $\pm 15 \text{ W/m}^2$. This is equivalent to the flux of heat which would warm or cool a column of water 100 m deep by roughly 1°C in one year. Table 5.2 lists typical accuracies of fluxes measured globally from space. Now, let's look at each variable.

Wind Speed and Stress Stress is calculated from wind observations made from ships at sea and from scatterometers in space as described in the last chapter. Beaufort observations give mean wind velocity and wind stress, and scatterometers measurements yield global maps of day to day variability of the winds used to produce maps of monthly-mean wind stress. The two largest

Variable	Accuracy	Comments
Wind Speed	$\pm 1.5 \text{ m/s} \\ \pm 1.5 \text{ m/s}$	Instrument Error Sampling Error (Monthly Average)
Wind Stess	$\begin{array}{c} \pm 10 \ \% \\ \pm 14 \ \mathrm{Pa} \end{array}$	Drag Coefficient Error Assuming 10 m/s Wind Speed
Insolation	$^{\pm 5}$ % $^{\pm 15}$ W/m ² $^{\pm 10}$ %	Monthly Average Monthly Average Daily Average
Rain Rate	$\pm 50~\%$	
Rainfall	$\pm 10~\%$	$5^\circ \times 5^\circ$ area for TRMM
Net Long Wave Radiation	$\pm 4-8 \%$ $\pm 15-27 \text{ W/m}^2$	Daily Average
Latent Heat Flux	$\begin{array}{c} \pm 35 \ \mathrm{W/m^2} \\ \pm 15 \ \mathrm{W/m^2} \end{array}$	Daily Average Monthly Average

 Table 5.2 Accuracy of Wind and Fluxes Observed Globally From Space

sources of error are: 1) lack of sufficient measurements of wind in time and space (sampling error); and 2) uncertainty in C_D (Chelton and Freilich, 1985).

Insolation Insolation is calculated from cloud observations made from ships and from visible-light radiometers on meteorological satellites. Satellite measurements are far more accurate than the ship data because it's very hard to estimate cloudiness from below the clouds. Satellite measurements processed by the International Satellite Cloud Climatology Project ISCCP are the basis for maps of insolation and its variability from month to month (Darnell et al. 1988; Rossow and Schiffer 1991).

The basic idea behind the calculation of insolation is very simple. Sunlight at the top of the atmosphere is accurately known from the solar constant, latitude, longitude, and time. Sunlight is either reflected back to space by clouds, or it eventually reaches the sea surface. Only a small and nearly constant fraction is absorbed in the atmosphere. Thus insolation is calculated from:

Insolution = S(1 - A) - C

where $S = 1365 \text{ W/m}^2$ is the solar constant, A is albedo, the ratio of incident to reflected sunlight, and C is a constant which includes absorption by ozone and other atmospheric gases and by cloud droplets. Insolation is calculated from cloud data (which also includes reflection from aerosols) collected from instruments such as the AVHRR on meteorological satellites. Ozone and gas absorption are calculated from from known distributions of the gases in the atmosphere. Q_{SW} is calculated from satellite data with an accuracy of 5–7%.

Recent work by Cess et al. (1995) and Ramanathan et al. (1995) suggest that the simple idea may be wrong, and that atmospheric absorption is a function of cloudiness. Schmetz (1989) gives a good review of the technique, and Taylor (1990) describes some of the relationships between satellite observations and terms in the radiation budget.



Figure 5.5 Rainfall in m/year calculated from data compiled by the Global Precipitation Climatology Project at NASA's Goddard Space Flight Center using data from rain gauges, infrared radiometers on geosynchronous meteorological satellites, and the SSM–I. Contour interval is 0.5 m/yr, light shaded areas exceed 2 m/yr, heavy shaded areas exceed 3 m/yr.

Errors come from lack of knowledge of the angular distribution of sunlight reflected from clouds and the surface and the daily variability of insolation, which is needed when data from polar-orbiting satellites are used for calculating insolation (See Salby et al. 1991).

Water Flux (Rainfall) Rain rate is another variable that is very difficult to measure from ships. Rain collected from gauges at different locations on ships and from gauges on nearby docks all differ by more than a factor of two. Rain at sea falls mostly horizontally because of wind, and the ship's superstructure distorts the paths of raindrops. Rain in many areas falls mostly as drizzle, and it is difficulty to detect and measure.

The most accurate measurements of rain rate in the tropics $(\pm 35^{\circ})$ are calculated from microwave radiometer and radar observations of rain at several frequencies using instruments on the Tropical Rain Measuring Mission TRMM launched in 1997. Rain for other times and latitudes can be calculated accurately by combining microwave data with infrared observations of the height of cloud tops and with raingauge data (Figure 5.5). Rain is also calculated from the reanalyses of the output from numerical weather forecast models (Schubert, Rood, and Pfaendtner, 1993), from ship observations (Petty, 1995), and from combining ship and satellite observations with output from numerical weather-prediction models (Xie and Arkin, 1997).

The largest source of error is due to conversion of rain rate to cumulative rainfall, a sampling error. Rain is very rare, it is log-normally distributed, and most rain comes from a few storms. Satellites tend to miss storms, and data must be averaged over areas up to 5° on a side to obtain useful values of rainfall. Errors also arise when the raining area doesn't fill the radiometer's beam.

Net Long-Wave Radiation Net Long-wave radiation is not easily calculated because it depends on the height and thickness of clouds, and the vertical distribution of water vapor in the atmosphere. It can be calculated by numerical weather-prediction models or from observations of the vertical structure of the atmosphere from atmospheric sounders. The net flux is:

$$F = \langle e \rangle [F_d - S T^4] \tag{5.11}$$

where $\langle e \rangle$ is the average emissivity of the surface, F_d is downward flux calculated from satellite, microwave-radiometer data or numerical models, T is sea-surface temperature, and S is the Stefan-Boltzmann constant. The first term is the downward radiation from the atmosphere absorbed by the ocean. Frouin, Gautier, and Morcrette (1988) describe how F_d can be calculated. The second term is the radiation emitted from the ocean. Both terms are large, and the net flux is the difference between two large quantities (see figure 5.6).

Schluessel et al. (1995) estimated the accuracy of monthly averaged values is $\pm 5-15 \text{ W/m}^2$. Improvements will come from more data, which reduces sampling error, and from a better understanding of daily cloud variability. Note, however, that the flux tends to be relatively constant over space and time; so much-improved accuracy may not be necessary.

Latent Heat Flux Latent heat flux is calculated from ship observations of relative humidity, water temperature, and wind speed using the bulk formula. Latent heat fluxes are also calculated by numerical weather models or from the COADS data set described below. The fluxes are difficult to calculate from space because satellite instruments are not very sensitive to water vapor close to the sea. Liu (1988) however showed that monthly averages of surface humidity are correlated with monthly averages of water vapor in the column of air extending from the surface to space. This is easily measured from space; and Liu used monthly averages of microwave-radiometer observations of wind speed, water vapor in the air column, and water temperature to calculate latent heat fluxes with an accuracy of $\pm 35 \text{ W/m}^2$. Later, Schulz et al (1997) used AVHRR measurements of sea-surface temperature together with SSM/I measurements of water vapor and wind, to calculate latent heat flux with an accuracy of $\pm 30 \text{ W/m}^2$ or $\pm 15 \text{ W/m}^2$ for monthly averages.

Sensible Heat Flux Sensible heat flux is calculated from observations of airsea temperature difference and wind speed made from ships, or from the output of numerical weather models. Sensible fluxes are small almost everywhere except offshore of the east coasts of continents in winter when cold, Arctic air masses extract heat from warm, western, boundary currents. In these areas, numerical models give perhaps the best values of the fluxes. Historical ship report give the long-term mean values of the fluxes.

5.5 Global Data Sets for Fluxes

Ship and satellite data have been processed to produce global maps of fluxes. Observations from ship measurements made over the past 150 years yield maps

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of the long-term mean values of the fluxes, especially in the northern hemisphere. Ship data, however, are sparse in time and space, and they are increasingly supplemented by data from space. Space data give the variability of some of the fluxes. Output from numerical weather models is also used. The most useful maps are perhaps those that result from combining level 3 and 4 satellite data sets and observations from ships, using numerical weather models. Let's look first at the sources of data, then at a few of the more widely used data sets.

Comprehensive Ocean-Atmosphere Data Set Data collected from observers on ships are the richest source of marine information. Slutz et al. (1985) describing their efforts to collect, edit, summarize, and publish all marine observations write:

Since 1854, ships of many countries have been taking regular observations of local weather, sea surface temperature, and many other characteristics near the boundary between the ocean and the atmosphere. The observations by one such ship-of-opportunity at one time and place, usually incidental to its voyage, make up a marine report. In later years fixed research vessels, buoys, and other devices have contributed data. Marine reports have been collected, often in machine-readable form, by various agencies and countries. That vast collection of data, spanning the global oceans from the mid-nineteenth century to date, is the historical ocean-atmosphere record.

These marine reports have now been edited and published as the *Comprehensive Ocean-Atmosphere Data Set* COADS (Woodruf et al. 1987) available through the National Oceanic and Atmospheric Administration.

The first COADS release includes 70 million reports of marine surface conditions collected by observers on merchant ships from 1854–1979. A second release of data is based on reports from 1980–1986. The data set include fully qualitycontrolled (trimmed) reports and summaries. Each of the 70 million unique reports contains 28 elements of weather, position, etc., as well as flags indicating which observations were statistically trimmed. Here, statistically trimmed means outliers were removed from the data set. The summaries included in the data set give 14 statistics, such as the median and mean, for each of eight observed variables: air and sea surface temperatures, wind velocity, sea-level pressure, humidity, and cloudiness, plus 11 derived variables.

The data set consists of an easily-used data base at three principal resolutions: 1) individual reports, 2) year-month summaries of the individual reports in 2° latitude by 2° longitude boxes, and 3) decade-month summaries.

Duplicate reports judged inferior by a first quality control process designed by the National Climatic Data Center NCDC were eliminated or flagged, and "untrimmed" monthly and decadal summaries were computed for acceptable data within each 2° latitude by 2° longitude grid. Tighter, median-smoothed limits were used as criteria for statistical rejection of apparent outliers from the data used for separate sets of *trimmed* monthly and decadal summaries. Individual observations were retained in report form but flagged during this second quality control process if they fell outside 2.8 or 3.5 estimated standarddeviations about the smoothed median applicable to their 2° latitude by 2° longitude box, month, and 56–, 40–, or 30–year period (*i.e.*, 1854–1990, 1910–1949, or 1950–1979).

The data are most useful in the northern hemisphere, especially the North Atlantic. Data are sparse in the southern hemisphere and they are not reliable south of 30° S. Gleckler and Weare (1997) analyzed the accuracy of the COADS data for calculating global maps and zonal averages of the fluxes from 55°N to 40°S. They found that systematic errors dominated the zonal means. Zonal averages of insolation were uncertain by about 10%, ranging from $\pm 10 \text{ W/m}^2$ in high latitudes to $\pm 25 \text{ W/m}^2$ in the tropics. Long wave fluxes were uncertain by about $\pm 7 \text{ W/m}^2$. Latent heat flux uncertainties ranged from $\pm 10 \text{ W/m}^2$ in some areas of the northern oceans to $\pm 30 \text{ W/m}^2$ in the western tropical oceans to $\pm 50 \text{ W/m}^2$ in western boundary currents. Sensible heat flux uncertainties tend to be around $\pm 5 - 10 \text{ W/m}^2$.

Josey et al (1999) compared averaged fluxes calculated from COADS with fluxes calculated from observations made by carefully calibrated instruments on some ships and buoys. They found that mean flux into the oceans, when averaged over all the seas surface had errors of $\pm 30 \text{ W/m}^2$. Errors vary seasonally and by region, and global corrections proposed by DaSilva, Young, and Levitus (1995) shown in figure 5.7 may not be useful.

Satellite Data Raw data are available from the many satellite projects. Mostly, these data must be further processed to be useful. Various levels of processed data from satellite projects are recognized (Table 5.3):

Table 5.3 Levels of Processed Satellite Data

Level	Level of Processing	
Level 1	Unprocessed data from the satellite in engineering units (volts)	
Level 2	Data processed into geophysical units (wind speed) at the time and place	
	the satellite instrument made the observation	
Level 3	Level 2 data interpolated to fixed coordinates in time and space	
Level 4	Level 3 data averaged in time and space or further processed	
-		

Data are available from the instruments on operational meteorological satellites including:

- 1. NOAA series of polar-orbiting, meteorological satellites;
- 2. Defense Meteorological Satellite Program DMSP polar-orbiting satellites, which carry the Special Sensor Microwave/ Imager (SSM/I);
- 3. Geostationary meteorological satellites operated by NOAA (GOES), Japan (GMS) and the European Space Agency (METEOSATS).

Data are also available from instruments on experimental satellites such as:

- 1. Nimbus-7, Earth Radiation Budget Instruments;
- 2. Earth Radiation Budget Satellite, Earth Radiation Budget Experiment;
- 3. The European Space Agency's ERS-1 & 2;
- 4. The Japanese ADvanced Earth Observing System (ADEOS).

5.5. GLOBAL DATA SETS FOR FLUXES

Data from satellites are collected, processed, and archived by government organizations. Archived data are further processed to produce useful flux data sets.

International Satellite Cloud Climatology Project Ship-based observations of insolation and clouds are being supplemented and replaced more and more by data from space. The International Satellite Cloud Climatology Project is an ambitious project to collect observations of clouds made by dozens of meteorological satellites from 1985 to 1995, to calibrate the the satellite data, to calculate cloud cover using carefully verified techniques, and to calculate surface insolation (Rossow and Schiffer, 1991). The clouds were observed with visible-light instruments on polar-orbiting and geostationary satellites.

Global Precipitation Climatology Project uses three sources of data to calculate rain rate (Huffman, et al. 1995, 1997):

- 1. Infrared observations of the height of cumulus clouds from GOES satellites. The basic idea is that the more rain produced by cumulus clouds, the higher the cloud top, and the colder the top appears in the infrared. Thus rain rate at the base of the clouds is related to infrared temperature.
- 2. Measurements by rain gauges on islands and land.
- 3. Radio-frequency emissions from from water droplets in the atmosphere observed by the SSM–I.

Accuracy is about 1 mm/day. Data from the project are available on a 2.5° latitude by 2.5° longitude grid from July 1987 to December 1995 from the Global Land Ocean Precipitation Analysis at the NASA Goddard Space Flight Center.

Xie and Arkin (1997) produced a 17-year data set based on seven types of satellite and rain-gauge data combined with the output from the NCEP/NCAR reanalysed data from numerical weather models. The data set has the same spatial and temporal resolution as the Huffman data set.

Reanalyzed Data From Numerical Weather Models Surface heat fluxes have been calculated from weather data using numerical weather models by various renalysis projects described on page 48 of the last chapter. The fluxes are consistent with atmospheric dynamics, they are global, and they are available for many years on a uniform grid. For example, the NCAR/NCEP reanalyzed data are available on a CD-ROM including daily averages of wind stress, sensible and latent heat fluxes, net long and short wave fluxes, near-surface temperature, and precipitation. Data on tape include values every six hours.

The reanalysed data sets are just becoming available, and their accuracy is still uncertain. Recent studies (WCRP, 1998) suggest:

- 1. The fluxes are biased because they were calculated using numerical models optimized to produce accurate weather forecasts. The time-mean values of the fluxes may not be as accurate as the time-mean values calculated directly from ship observations.
- 2. The fluxes are probably more accurate in the northern hemisphere where ship ship observations are most common.

- 3. The fluxes have zonal means that differ significantly from the same zonal means calculated from COADS data. The differences can exceed 40 W/m^2 .
- 4. The atmospheric models do not require that the net heat flux averaged over time and Earth's surface be zero. The ECMWF data set averaged over fifteen years gives a net flux of 8 W/m^2 into the ocean.

Thus reanalysed fluxes are most useful for forcing ocean, general-circulation models needing actual heat fluxes and wind stress. COADS data are most useful for calculating time-mean fluxes except perhaps in the southern hemisphere.

Data From Numerical Weather Models Some projects require flux data a few hours after after observations are collected. The surface analysis from numerical weather models is a good source for this type of data.

5.6 Geographic Distribution of Terms in the Heat Budget

Various groups have used ship and satellite data together with numerical models of the atmospheric circulation to calculate globally averaged values of the terms for Earth's heat budget. The values give an overall view of the importance of the various terms (Figure 5.6). Notice that insolation balances infrared radiation at the top of the atmosphere. At the surface, latent heat flux and net infrared radiation tend to balance insolation, and sensible heat flux is small.

Note that only 20% of insolation reaching Earth is absorbed directly by the atmosphere while 49% is absorbed by the ocean and land. What then warms the atmosphere and drives the atmospheric circulation shown in figure 4.3? The answer is rain and infrared radiation from the ocean absorbed by the moist tropical atmosphere. Here's what happens. Sunlight warms the tropical oceans



Figure 5.6 The mean annual radiation and heat balance of the earth. From Houghton et al. (1996: 58), which used data from Kiehl and Trenberth (1996).

which must evaporate water to keep from warming up. It also radiates heat to the atmosphere, but the net radiation term is smaller than the evaporative term. Trade winds carry the heat in the form of water vapor to the tropical convergence zone where it falls as rain. Rain releases the latent heat evaporated from the sea, and it heats the air in cumulus rain clouds by as much as 125 W/m^2 averaged over a six-year period (See figure 14.1).

At first it may seem strange that rain heats the air, after all, we are familiar with summertime thunderstorms cooling the air at ground level. The cool air from thunderstorms is due to downdrafts. Higher in the cumulus cloud, heat released by rain warms the mid-levels of the atmosphere causing air to rise rapidly in the storm. Thunderstorms are large heat engines converting the energy of latent heat into kinetic energy of winds.

The zonal average of the oceanic heat-budget terms (Figure 5.7) shows that insolation is greatest in the tropics, that evaporation balances insolation, and that sensible heat flux is small. *Zonal average* is an average along lones of constant latitude. Note that the terms in figure 5.7 don't sum to zero. The areal-weighted integral of the curve for total heat flux is not zero. Because the



Figure 5.7 **Upper:** Zonal averages of heat transfer to the ocean by insolation Q_{SW} , and loss by infrared radiation Q_{LW} , sensible heat flux Q_S , and latent heat flux Q_L , calculated by DaSilva, Young, and Levitus (1995) using the COADS data set. **Lower:** Net heat flux through the sea surface calculated from the data above (solid line) and net heat flux constrained to give heat and fresh-water transports by the ocean that match independent calculations of these trasports. The area under the lower curves ought to be zero, but it is 16 W/m² for the unconstrained case and -3 W/m² for the constrained case.



Total Sky Net Insolation (W/m²)

Figure 5.8 Annual-mean insolation Q_{SW} (top) and infrared radiation Q_{LW} (bottom) through the sea surface during 1989 calculated by the Satellite Data Analysis Center at the NASA Langley Research Center (Darnell et al., 1992) using data from the International Satellite Cloud Climatology Project. Units are W/m², contour interval is 10 W/m².

net heat flux into the oceans averaged over several years must be less than a few watts per square meter, the non-zero value must be due to errors in the various terms in the heat budget.

Errors in the heat budget terms can be reduced by using additional information. For example, we know roughly how much heat and fresh water are transported by the oceans and atmosphere, and the known values for the transports can be used to constrain the calculations of net heat fluxes (Figure 5.7). The constrained fluxes show that the ocean gains heat in the tropics and loses heat at high latitudes.

Maps of the regional distribution of fluxes give clues to the processes pro-



Figure 5.9 Annual-mean latent heat flux from the sea surface Q_L in W/m² during 1989 calculated from data compiled by the Data Assimilation Office of NASA's Goddard Space Flight Center using reanalysed output from the ECMWF numerical weather prediction model. Contour interval is 10 W/m².

ducing the fluxes. Net downward short-wave radiation (insolation) at the sea surface (Figure 5.8 top) shows that the large flux into the tropical region is modulated by the distribution of clouds, and that heating is everywhere positive. The net infrared heat flux (Figure 5.8 bottom) is largest in regions with the least clouds, such as the central gyres and the eastern central Pacific. The net infrared flux is everywhere negative. Latent heat fluxes (Figure 5.9) are dominated by evaporation in the trade wind regions and the offshore flow of cold air masses behind cold fronts in winter offshore of Japan and North America. Sensible heat fluxes (Figure 5.10 top) are dominated by cold air blowing off continents. The net heating gain (Figure 5.10 bottom) is largest in equatorial regions and net heat loss is largest downwind on Asia and North America.

5.7 Meridional Heat Transport

Overall, Earth gains heat at the top of the tropical atmosphere, and it loses heat at the top of the polar atmosphere. The atmospheric and oceanic circulation together must transport heat from low to high latitudes to balance the gains and losses. This north-south transport is called the *meridional transport*.

How much heat is carried by the ocean and how much by the atmosphere? The sum of the meridional heat transport by the ocean and atmosphere together is calculated accurately from the divergence of the zonal average of the heat budget measured at the top of the atmosphere by satellites. To make the calculation, we assume steady state transports over many years so that any longterm net heat gain or loss through the top of the atmosphere must be balanced by a meridional transport and not by heat storage in the ocean or atmosphere. So let's start at the top of the atmosphere.



Figure 5.10 Annual-mean upward sensible heat flux Q_S **Top** and constrained, net, downward heat flux **Bottom** through the sea surface in W/m² calculated by DaSilva, Young, and Levitus (1995) using the COADS data set from 1945 to 1989. Contour interval is 2 W/m² (top) and 20 W/m² (bottom).

Heat Budget at the top of the Atmosphere Heat flux through the top of the atmosphere is measured with useful accuracy using radiometers on satellites.

- 1. Insolation is calculated from the solar constant and observations of reflected sunlight made by meteorological satellites and by special satellites such as the Earth Radiation Budget Experiment Satellite.
- 2. Back radiation is measured by infrared radiometers on the satellites.
- 3. The difference between insolation and net infrared radiation is the net heat flux across the top of the atmosphere.

Errors arise from calibration of instruments, and from inaccurate information about the angular distribution of reflected and emitted radiation. Satellite instruments tend to measure radiation propagating vertically upward, not radiation at large angles from vertical, and radiation at these angles is usually calculated not measured.

The sum of the meridional heat transported by the atmosphere and the oceans is calculated from the top of the atmosphere budget. First average the satellite observations in the zonal direction, to obtain a zonal average of the heat flux at the top of the atmosphere. Then calculate the meridional derivative of the zonal mean flux to calculate the north-south flux divergence. The divergence must be balanced by the heat transport by the atmosphere and the ocean across each latitude band.

Oceanic Heat Transport Oceanic heat transport are calculated three ways:

- 1. Surface Flux Method uses measurements of wind, insolation, air, and sea temperature, cloudiness, and bulk formulas to estimate the heat flux through the sea surface. Then the fluxes are integrated to obtain the zonal average of the heat flux (Figure 5.7). Finally, the meridional derivative of the net flux gives the flux divergence, which must be balanced by heat transport in the ocean.
- 2. Direct Method uses measured values of current velocity and temperature from top to bottom along a zonal section spanning an ocean basin. The values are used to calculate the flux from the product of northward velocity and heat content derived from the temperature measurement.
- 3. Residual Method uses atmospheric observations or the output of numerical models of the atmospheric circulation to calculate the heat transport in the atmosphere. This is the direct method applied to the atmosphere. The atmospheric transport is subtracted from the total meridional transport calculated from the top-of-the-atmosphere heat flux to obtain the oceanic contribution as a residual (Figure 5.11).

Oceanic heat transports calculated from the various methods were summarized by Charnock (1989), Talley (1984) and Trenberth and Solomon (1994). They found that it is difficult to calculate oceanic heat transport acurately by any method, and that errors were probably larger than estimated by the earlier studies. The most recent calculations, such as those shown in Figure 5.11, are now in better agreement and the error bars shown in the figure are realistic.

5.8 Meridional Fresh Water Transport

The Earth's water budget is dominated by evaporation and precipitation over the ocean. Baumgartner and Reichel (1975) report that 86% of global evaporation and 78% of global precipitation occur over the ocean. A map of the net evaporation (Figure 5.12) shows that evaporation exceeds a meter per year in the trade wind regimes in the eastern parts of the oceans.

The transport of fresh water by the ocean can be calculated in the same ways heat transports are calculated, with similar uncertainties (Figure 5.13).



Figure 5.11 Northward heat transport for 1988 in each ocean and the total transport summed over all oceans calculated by the residual method using atmospheric heat transport from ECMWF and top of the atmnosphere heat fluxes from the Earth Radiation Budget Experiment satellite. From Houghton et al. (1996: 212), which used data from Trenberth and Solomon (1994).

Knowledge of water fluxes and transports is important for understanding the global hydrological cycle, ocean dynamics, and global climate. For example, the variability of fresh water fluxes may have played an important role in the ice



Figure 5.12 Precipitation minus evaporation in meters per year calculated from global rainfall by the Global Precipitation Climatology Project and latent heat flux calculated by the Data Assimilation Office, both at NASA's Goddard Space Flight Center. Precipitation exceeds evaporation in the shaded regions, contour interval is 0.5 m.



Figure 5.13 Meridional transport of fresh water by the Atlantic from three surface-flux calculations: BR—Baumgartner and Reichel (1975); SBD—Schmitt et al. (1989); and IH—Isemer and Hasse's (1987) evaporation estimates combined with Dorman and Bourke's (1981) precipitation values. Also shown are direct measurements at 24°N by Hall and Bryden (1982) and 11°N by Friedrichs and Hall (1993). All are summed relative to an estimated Arctic southward export due to the Bering Strait throughflow and the water budget of the Arctic itself. From Schmitt (1994).

ages as discussed in $\S13.3$.

5.9 Variations in Solar Constant

We have assumed so far that the solar constant, the output of light and heat from the sun, is steady. Recent evidence based on variability of sunspots and faculae (bright spots) shows that the output varied by $\pm 0.2\%$ over centuries (Lean, Beer, and Bradley, 1995), and that this variability is correlated with changes in global mean temperature of Earth's surface of ± 0.4 °C. (Figure 5.14). In addition, White and Cayun (1998) found a small 12 yr, 22 yr, and longerperiod variations of sea-surface temperature measured by bathythermographs and ship-board thermometers over the past century. The observed response of Earth to solar variability is about that calculated from numerical models of the coupled ocean-atmosphere climate system. Many other changes in climate and weather have been attributed to solar variability. The correlations are somewhat controversial, and much more information can be found in Hoyt and Schatten's (1997) book on the subject.

5.10 Important Concepts

- 1. Sunlight is absorbed primarily in the tropical ocean. The amount of sunlight is modulated by season, latitude, time of day, and cloud cover.
- 2. Most of the heat absorbed by the oceans in the tropics is released as water vapor which heats the atmosphere when water is condenses as rain. Most of the rain falls in the tropical convergence zones, lesser amounts fall in mid-latitudes near the polar front.



Figure 5.14 Changes in solar constant (total solar irradiance) and global mean temperature of Earth's surface over the past 400 years. Except for a period of enhanced volcanic activity in the early 19th century, surface temperature is well correlated with solar variability (From Lean, personal communication).

- 3. Heat released by rain and absorbed infrared radiation from the ocean are the primary drivers for the atmospheric circulation.
- 4. The net heat flux from the oceans is largest in mid-latitudes and offshore of Japan and New England.
- 5. Heat fluxes can be measured directly using fast response instruments on low-flying aircraft, but this is not useful for estimating heat budgets of the ocean.
- 6. Heat fluxes through large regions of the sea surface can be estimated from bulk formula. Seasonal, regional, and global maps of fluxes are available based on ship and satellite observations.
- 7. The most widely used data sets for studying heat fluxes are the Comprehensive Ocean-Atmosphere Data Set and the reanalysis of meteorological data by numerical weather prediction models.
- 8. The oceans transport about one-half of the heat needed to warm higher latitudes, the atmosphere transports the other half.
- 9. Solar output is not constant, and the observed small variations in output of heat and light from the sun seem to produce the changes in global temperature observed over the past 400 years.