UPPER OCEAN HEAT AND FRESHWATER BUDGETS

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Introduction

This article describes the heat and salt balances in the upper ocean. Conservation equations for heat and salt describe how sources and sinks of heat and freshwater at the surface are distributed in the water column as a result of penetrative radiation, internal mixing, and advection.

Heat Budget

The surface layer of the ocean undergoes temperature variations which occur at several different temporal scales: a diurnal cycle reflecting day and night variability, an annual cycle showing seasonal variations in insolation and weather, and inter-annual to decadal cycles, indicating large-scale, long-term climate variability. The strength of a given cycle also depends on the location, for example, seasonal variability in sea surface temperature is smaller in the tropics than at high latitudes.

The heat balance for the near-surface layer is found by vertically integrating the temperature equation from the bottom of the layer, h, to the sea surface to obtain:

$$\begin{aligned}&\text{Rate of change of heat content } (Q_t)\\ &= \text{advection of heat } (Q_a) - \text{net surface heat flux } (Q_{net})\\ &\quad + \text{penetrative solar irradiance at depth } h\ (Q_I(-h))\\ &\quad + \text{vertical turbulent heat flux at depth } h\ (F_T(-h))\end{aligned}$$

The net surface heat flux (Q_{net}) is the sum of the short wave solar irradiance ($Q_I(0)$), the net (in − out) long-wave radiation, sensible and latent heat fluxes, and the heat flux associated with precipitation, e.g., rainfall. By convention, upward heat flux is regarded as positive. The layer depth h can be either fixed or space- and time-varying depth, e.g., a depth of the mixed layer or a depth of any arbitrary isoscalar (isotherm, isohaline, isopycnal). Horizontal mixing is neglected in this discussion.

Diurnal Cycle

The combined effects of net-incoming heat (from short-wave solar) during daytime and net-outgoing heat (due to black-body long-wave radiation – cloud back scatter) during nighttime generates a diurnal cycle of surface heat flux, although the strength of this cycle is modulated by surface winds and varies with seasonal changes in solar radiation and cloud cover. The incoming solar radiation, more than 90% of which is absorbed in the upper 20 m, heats the near-surface water, which decreases the near-surface density, stabilizing the upper water column and inhibiting vertical mixing. During nighttime, the outgoing (primarily long-wave radiative) heat cools the surface and generates turbulent convection, allowing near-surface heat and momentum to mix downward. The uneven cycle of mixing creates a diurnal mixed layer in the upper ocean, which can be observed in profiles of temperature, density, and turbulent kinetic energy (TKE) dissipation rate. The diurnal cycle of convection is typically more visible during weak to moderate winds than during strong wind events when wave breaking and strong shear enhance mixing. If horizontal advection is negligible, then heat balance in the diurnal mixed layer is one-dimensional, with the rate of change of heat coming from the net surface fluxes, Q_{net}, the rate of turbulent entrainment of heat from below, $F_T(-h)$, and the rate of solar heating at depth, $Q_I(-h)$. The entrainment flux at the base of the mixed layer, $F_T(-h)$, may be important during the deepening phase of the diurnal mixed layer, and the penetrative solar irradiance, $Q_I(-h)$, becomes large during the shoaling phase.

Precipitation influences the diurnal cycle, especially in tropical oceans. Tropical rainfall, which is highly variable in space and time, can have a dramatic effect on the ocean surface. The freshwater lenses ($\sim$10–50 km) generated by rainfall can trap heat and momentum near the surface. The extra buoyancy coming from rainfall aids warming near the surface by limiting nighttime convection to a shallow mixed layer. Advection becomes important because, relative to its surroundings, the momentum put in at the surface is trapped by the stratification of the shallow, fresh lens, which consequently moves faster than the surroundings. Thus, three-dimensional circulation must be considered in both heat and salinity budgets.

Episodic Wind Events

Episodic wind events play an important role in upper ocean heat and salinity budgets, and they can have both local and remote oceanic responses. Severe storms such as hurricanes in the Atlantic or typhoons in the Pacific Ocean, and westerly wind bursts in the western equatorial Pacific Ocean are examples of such events. These events, which impart large surface stress and buoyancy flux to the ocean, may last a few days to several weeks, and are important local forcing mechanisms. The combined effect of cloud cover and heavy precipitation modifies net surface heat flux and turbulent mixing in the upper ocean. For example, in the western equatorial Pacific, the net incoming solar radiative heat balances the rate of increase of heat content during calm and dry weather (i.e., the near-surface layer heats up). During westerly wind bursts, which may precede the onset of El Niño conditions farther east, vertical heat exchange between the mixed layer and the main thermocline is large, and the net outgoing heat from the surface balances entrainment from below, horizontal and vertical advection, and the rate of change in heat content.

Seasonal Cycle

The surface water is warmer in summer than in winter. For much of the world's ocean, the seasonal variations in upper ocean thermal structure can be described by considering only seasonal variation of the net surface heat flux. However, this balance does not hold in all parts of the ocean. The horizontal advection term plays a crucial role in the heat and salinity budgets during upwelling and downwelling events in coastal regions, in shifts of seasonal currents, and in El Niño events. Apart from these limitations, much of the observed oceanic seasonal temperature cycle can be explained by considering only the net surface heat flux. That is, neglecting the horizontal advection of heat, and assuming a surface layer deep enough such that throughout the year no heat is entrained from below, such that $F_T(-h) \approx 0$, and the solar flux does not extend beyond this depth, so that $Q_I(-h) \approx 0$, then the rate of change of heat in the layer exactly balances the net surface heat flux.

The incoming solar radiation has an annual cycle. In the Northern Hemisphere, incoming heat from the sun has a maximum in mid-June and a minimum in mid-December. The outgoing heat from the ocean peaks during August–September, and drops to a minimum during mid-March. As a result, there is a net gain of heat to the surface layer during mid-February to mid-August, and a net loss during the other six months. The seasonal change in net surface heat flux sets up a seasonal thermocline, which is reflected in the annual variation in the depth of the mixed layer, the weakly stratified layer above the seasonal thermocline.

A typical growth and decay cycle of seasonal thermocline for a mid-latitude, Northern Hemisphere site is shown in **Figure 1**. The mixed-layer temperature is colder and the thermocline is deeper during winter, whereas the temperature is warmer and the thermocline is shallower during summer. The decreasing incoming solar radiation and increasing wind forcing associated with mid-latitude weather patterns during fall intensify turbulent mixing, which in turn generates a deep winter mixed layer. The seasonal thermocline disappears by end

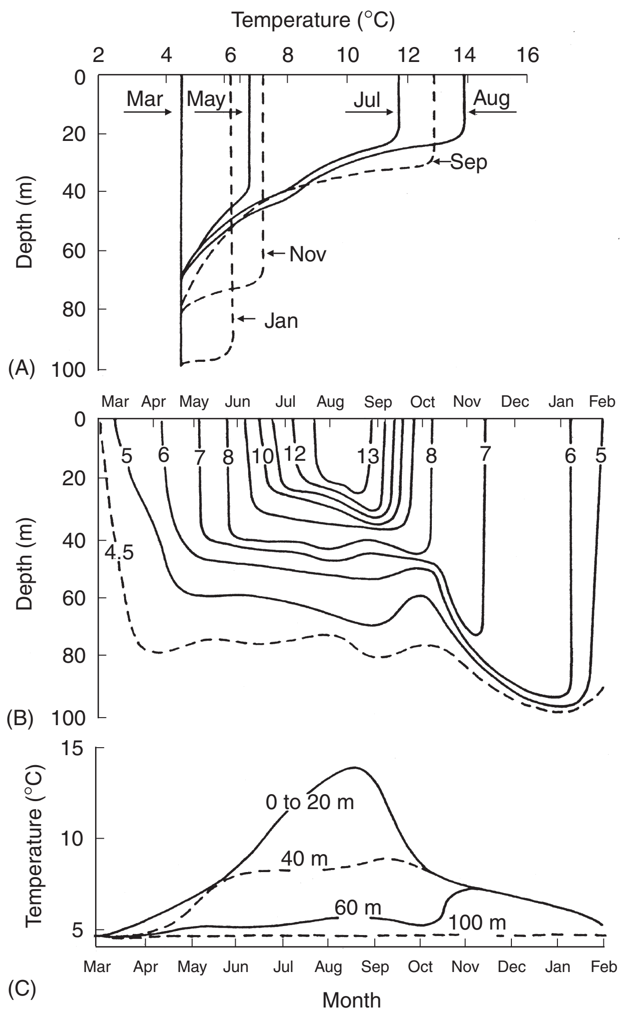

Figure 1 Seasonal variability of upper ocean thermal structure at a mid-latitude site in the Northern Hemisphere (50°N, 145°W). The same information is shown in three different ways: (A) vertical profiles of temperature; (B) temperature contours in depth and time; and (C) temperature at four depths. There is no seasonal change below 100 m depth.

of winter, but reappears in spring as the annual radiation cycle progresses. Turbulent mixing becomes weak during summer because of weak summer winds and the entrainment flux from below the mixed layer is minimal due to strong stability created by net heating and weak summer winds.

Long-term Balance and Advection

Long-term changes in temperature at a given location in the upper ocean are small compared to diurnal and seasonal variability. Averaged over several years, the time rate of change can be neglected in the upper ocean heat budget, and horizontal advection of heat balances the net surface heat flux.

The net surface heat flux, which results in warming at lower latitudes and cooling at higher latitudes, is balanced by a net poleward transport of heat by ocean currents. In this redistribution, both warm poleward currents (such as the Gulf Stream, Agulhas, and Kuroshio currents) and cold equatorward currents (such as the Labrador, Canary, Benguela, California, and Peru currents) play important roles. For example, there is a net oceanic heat loss of order $100\,\mathrm{W\,m^{-2}}$ in the vicinity of major western boundary currents (such as the Gulf Stream, and Kuroshio), and there is an oceanic net heat gain of order $50\,\mathrm{W\,m^{-2}}$ in the vicinity of major eastern boundary currents (such as the Canary and Peru currents). In the subtropics, waters on the west sides are warmer than those on the east, whereas in high latitudes the opposite is true. The eastern side of the Norwegian Sea, for example, is anomalously warm as a result of poleward transport of heat by the Gulf Stream.

In the trade wind regions of the Atlantic and Pacific Oceans, a long-term average surface heat flux of order $100\,\mathrm{W\,m^{-2}}$ is balanced by lateral advection and upwelling. This contrasts strongly with the region of the western equatorial Pacific Ocean that is home to the westerly wind bursts, which presage the onset of El Niño conditions in the central and eastern Pacific. Annual mean winds in this region are weak and surface waters are the warmest ($>29°\mathrm{C}$) in the world's oceans, yet the surface receives almost negligible heat flux ($<20\,\mathrm{W\,m^{-2}}$) in a long-term average, indicating that horizontal heat transport is also small on average.

It is crucial to point out that changes in the ocean circulation, such as a small shift in the strength and the position of the western Pacific Pool warm pool or meanders in the Gulf Stream and any changes in major currents will result changes in ocean–atmosphere heat balance. Such remote changes in the heat budget will change the weather patterns, since the ocean–atmospheric system is strongly coupled.

Salt and Fresh Water Budgets

Distribution of fresh water in the upper ocean can be studied by examining salinity budget. The major sources and sinks of surface salinity are evaporation (at a rate E), rainfall (P), freezing of sea water (F), melting of sea ice (M), and river run-off (R). Although salt is removed from the ocean by aerosols and spray (at least when they do not fall back to the ocean but to land), evaporation does not remove salt, only water. In the process of freezing, most of the salts are excluded.

A governing equation for salinity conservation can be written as

$$\begin{aligned}&\text{Rate of change of salinity } (S_t)\\&\quad= \text{advection of salinity } (S_a)\\&\quad- \text{net surface salinity flux } (S_{net})\\&\quad+ \text{vertical turbulent salinity flux at depth } h(F_S(-h))\end{aligned}$$

The salinity loss or gain due to evaporation and precipitation is defined as an upward flux of magnitude $S_0(P-E)$, where S_0 is the salinity of sea water, which for much of the world's ocean is close to 35 PSU (practical salinity units, which for most purposes is nearly identical to parts per thousand). When the ocean is ice covered, the equivalent upward salinity flux is $(S_0 - S_i)(M - F)$, where S_i is the equivalent salinity of sea ice (if melted) and is about 4 PSU. Loss of salinity due to river runoff can be approximated as S_0R. For most regions of world oceans, $S_{net} = S_0(P-E) + S_0R$, and for polar oceans, $S_{net} = (S_0 - S_i)(M-F) + S_0R$.

Long-term Budget

The ocean becomes locally saltier when surface water evaporates, whereas it becomes fresher, when rain falls. The net effect of such imbalance can be seen from charts of surface salinity and charts of rainfall (P) minus evaporation (E). The melting of sea ice minus freezing of sea water in polar oceans plays a role analogous to $P-E$ in lower latitudes.

A band of low-salinity stretches across the Pacific coincident with high rainfall in the inter-tropical convergence zone, where rising moist air causes condensation and precipitation. Also notable are the low-salinity, high-rainfall western equatorial Pacific warm pool and the region south west of southern Mexico. At the latitudes of the subtropical oceans, descending limbs of the Walker circulation bring cold, dry air to the surface, and evaporation exceeds precipitation. The surface salinity of the subtropics is consequently higher than average. The Atlantic

Ocean is more saline than the Pacific mainly due to the difference in $P - E$, with greater evaporation from the Atlantic. More localized minima are found in regions of major river run-offs such as the Amazon, Congo, and Ganges.

Heat and Freshwater Balances in Polar Oceans

Because salt is rejected upon freezing of sea water, the heat and salt (or equivalently, freshwater) balances of the seasonally and permanently ice-covered regions of the ocean are intimately connected. Because the Arctic Ocean is salt stratified, rather than temperature stratified as is much of the world's oceans, the addition, removal, and redistribution of salt or fresh water can have a significant impact on dynamics. With a significant input from Eurasian rivers, the Arctic basin receives an excess of precipitation and runoff over evaporation, and exports the excess fresh water southward to the Greenland Sea in the form of fresh water and ice. Melting of the exported sea ice and sensible loss to the atmosphere represent the endpoints of the large-scale poleward transport of heat mentioned earlier. Similarly, shedding of ice from the Antarctic ice shelves represents an equatorward transport of fresh water and a sink for poleward transport of heat.

The Arctic Ocean is essentially a mediterranean sea composed of a mixture of salty water flowing in from the Atlantic Ocean, slightly fresher water flowing in from the Pacific Ocean, and fresh river water that flows into the central basin after flowing across the broad shelves that surround much of the basin (*see* **Arctic Basin Circulation**). The Atlantic inflow is the end of the Gulf Stream system, which warms the eastern side of the Greenland and Norwegian Seas in its transit northward. Most of the ice found in the Arctic region is sea ice of thickness 1–8 m; glacial ice is introduced around the periphery, notably from Greenland into the Greenland, Icelandic and Labrador Seas. In contrast to the Arctic, the Antarctic is a region of a large central land mass surrounded by open ocean. This fundamental difference results in katabatic winds, that can rapidly thicken the sea ice, and a large marginal ice zone where the ice is exposed to warm water and wave action. In addition, massive sections of the Antarctic ice shelves (e.g., the Ross Ice Shelf) up to about 10 000 km^2 in area, break off occasionally, effecting equatorward transport of freshwater sequestered in glacial ice.

Seasonal changes in the ice cover in the Arctic modulate the surface heat flux, with most of the heat exchange to the atmosphere occurring at the periphery, in polynyas, and in leads. In the interior, the cold halocline layer isolates the surface sea ice from the heat of the warm Atlantic water, and consequently limits the upward heat flux to the ice. The cold halocline layer originates in wintertime convection north of the Barents Sea, evolving from a deep mixed layer that is capped by modified river water input from the Russian shelves, and which subsequently experiences the seasonal cycle of sea-ice melting and freezing. Additionally, the formation of ice over the continental shelves produces brine that drains off the shelves into the interior to maintain the halocline.

Changes in the circulation of the upper layers of the Arctic Ocean and associated variations in heat and freshwater transports are related to the North Atlantic Oscillation (NAO), a robust mode of large-scale atmospheric variability. Phases of the NAO persist for several decades. The most recent phase change of the NAO, referred to as an NAO high-index phase, is associated with decreased salinity (primarily through increased precipitation associated with changing winter storm tracks in the European Arctic and sub-Arctic), increased temperature, and increased transport of the Atlantic inflows to the Arctic. During this period, the cold halocline layer disappeared from the European side of the Arctic Ocean due to a shift of the injection point of fresh, riverine water from the Russian continental shelves into the interior. Southward ice flux from the Arctic Ocean to the Greenland Sea shows a strong correlation to the NAO, with positive phases of the NAO corresponding to surges in ice flux. That the ice flux/NAO relationship is not simple, however, is illustrated by the Great Salinity Anomaly (GSA) of the 1960s, in which some 2000 km^3 of excess ice led to a cessation of deep-water formation in the Greenland Sea, in the midst of the negative phase of the NAO. Thus, variations in the upper ocean heat and salt balance of the Arctic Ocean have the potential to impact the deep thermohaline circulation of the temperate oceans.

See also

Air–Sea Transfer: Dimethyl Sulphide, COS, CS_2, NH_4, Non-methane Hydrocarbons, Organo-halogens. Arctic Basin Circulation. Ekman Transport and Pumping. El Niño Southern Oscillation (ENSO). El Niño Southern Oscillation (ENSO) Models. Evaporation and Humidity. Freshwater Transport and Climate. Heat and Momentum Fluxes at the Sea Surface. Heat Transport and Climate. Ocean Circulation. Open Ocean Convection. Pacific Ocean Equatorial Currents. Penetrating Shortwave Radiation. River Inputs. Sea Ice: Overview; Variations in Extent and Thickness. Upper Ocean Mean Horizontal

Structure. Upper Ocean Mixing Processes. Upper Ocean Responses to Strong Forcing Events. Upper Ocean Time and Space Variability. Upper Ocean Vertical Structure. Wind and Buoyancy-forced Upper Ocean.

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UPPER OCEAN MEAN HORIZONTAL STRUCTURE

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Introduction

The upper ocean is the most variable, most accessible, and dynamically most active part of the marine environment. Its structure is of interest to many science disciplines. Historically, most studies of the upper ocean focused on its impact on shipping, fisheries, and recreation, involving physical and biological oceanographers and marine chemists. Increased recognition of the ocean's role in climate variability and climate change has led to a growing interest in the upper ocean from meteorologists and climatologists.

In the context of this article the upper ocean is defined as the ocean region from the surface to a depth of 1 km and excludes the shelf regions. Although the upper ocean is small in volume when compared to the world ocean as a whole, it is of fundamental importance for life processes in the sea. It determines the framework for marine life through processes that operate on space scales from millimeters to hundreds of kilometers and on timescales from seconds to seasons. On larger space and timescales, its circulation and water mass renewal processes span typically a few thousand kilometers and several decades, which means that the upper ocean plays an important role in decadal variability of the climate system. (In comparison, circulation and water mass renewal timescales in the deeper ocean are of the order of centuries, and the water masses below the upper ocean are elements of climate change rather than climate variability.)

The upper ocean can be subdivided into two regions. The upper region is controlled by air–sea interaction processes on timescales of less than a few months. It contains the oceanic mixed layer, the seasonal thermocline and, where it exists, the barrier layer. The lower region, known as the permanent thermocline, represents the transition from the upper ocean to the deeper oceanic layers. It extends to about 1 km depth in the subtropics, is somewhat shallower near the equator and absent poleward of the Subtropical Front. These elements of the upper ocean will be defined and described in more detail, following an introductory overview of some elementary property fields.

Horizontal Property Fields

The annual mean sea surface temperature (SST) is determined by the heat exchange between ocean and atmosphere. If local solar heat input would be the only determinant, contours of constant SST would extend zonally around the globe, with highest values at the equator and lowest values at the poles. The actual SST field (**Figure 1**) comes close to this simple distribution. Notable departures occur for two reasons.

1. Strong meridional currents transport warm water poleward in the western boundary currents along the east coasts of continents. Examples are the Gulf Stream in the North Atlantic Ocean and the Kuroshio in the North Pacific Ocean. In contrast, cold water is transported equatorward along the west coast of continents.