Ocean, Energy Flows in

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Glossary

Ekman layer A boundary layer within the top 20 to 30 m of the ocean where the wind stress is balanced by Coriolis and frictional forces.

mixed layer A layer of water with properties that are almost vertically homogenized by wind stirring and convective overturning due to the unstable stratification induced by cooling or salinification.

thermocline A thin layer of water at the depth range of 300 to 800 m in the tropical/subtropical basins, where the vertical temperature gradient is maximum.

thermohaline circulation An overturning flow in the ocean driven by mechanical stirring/mixing, which transports mass, heat, freshwater, and other properties. In addition, the surface heat and freshwater fluxes are necessary for setting up the flow.

wind-driven circulation A circulation in the upper kilometer of the ocean, which is primarily controlled by wind stress. In addition, this circulation also depends on buoyancy forcing and mixing.

Oceanic circulation plays an important role in the climate system because it transports heat and freshwater fluxes meridionally. There is a huge heat flux through the upper surface of the ocean, but the ocean is not a heat engine; instead, the ocean is driven by external mechanical energy, including wind stress and tides. Although mechanical energy flux is 1000 times smaller than the heat flux, it controls the strength of the oceanic general circulation. In particular, energy input through wind stress is the most important source of mechanical energy in driving the oceanic circulation.

1. BASIC CONSERVATION LAWS FOR OCEAN CIRCULATION

Oceanic circulation is a mechanical system, so it obeys many conservation laws as other mechanical systems do, such as the conservation of mass, angular momentum, vorticity, and energy. It is well known that these laws should be obeyed when we study the oceanic circulation either observationally, theoretically, or numerically. However, this fundamental rule has not always been followed closely in previous studies of the oceanic circulation. As a result, our knowledge of the energetics of the ocean circulation remains preliminary.

1.1 Mass Conservation Law

This is one of the most important laws of nature. However, mass conservation has not been applied in many existing numerical models. In fact, most models are based on the Boussinesq approximations, which consist of three terms: mass conservation is replaced by a volume conservation, *in situ* density in the horizontal momentum equations is replaced by a constant reference density, and tracer prognostic equations are based on volume conservation instead of mass conservation. For models based on the Boussinesq approximations, heating (cooling) can induce an artificial loss (gain) of gravitational potential energy. Furthermore, precipitation (evaporation) is often simulated in terms of the equivalent virtual salt flux out of (into) the ocean; thus, an artificial sink of gravitational potential energy is generated in the models.

1.2 Angular Momentum Conservation Law

In a rotating system, the linear momentum is not conserved; instead the angular momentum conservation is more relevant (i.e., the rate of change of the angular momentum of the oceanic general circulation must be balanced by torques). The angular momentum conservation law has been used extensively in the study of the atmospheric circulation. The Antarctic Circumpolar Current is the corresponding part in the oceans where the angular momentum conservation law dictates the balance.

1.3 Potential Vorticity Conservation Law

The potential vorticity equation can be obtained by cross-differentiation of the horizontal momentum equations and subtracting the results; it has been extensively used in oceanographic studies. According to the theory, total potential vorticity is conserved in the interior of a stratified fluid, and the only source of potential vorticity must be on the outer surfaces of the stratified fluid.

1.4 Energy Conservation Law

Motions in the oceans must be associated with friction; thus, to maintain circulation in the ocean, mechanical energy source is required to balance the loss of mechanical energy. Energy source and dissipation are the most important means of controlling the oceanic circulation system. However, over the past years the mechanical energy balance of the oceanic general circulation has been overlooked, and there has been very little discussion about such a fundamental issue. In fact, energy balance equations for the oceanic general circulation in most previously published books or papers are incomplete because one of the most important terms in the mechanical energy balance, energy required for vertical mixing in a stratified environment, is not included. Therefore, there is no complete statement about mechanical energy balance for the oceanic general circulation. In addition, most numerical models do not conserve total energy, especially in the finite difference scheme in time stepping. Thus, energetics diagnosed from these models are unreliable.

2. *ENERGETICS OF THE OCEANIC CIRCULATION*

Energy in the ocean can be classified into four major forms: the gravitational potential energy, kinetic energy, internal energy, and chemical potential, as shown in Fig. 1. Although the chemical potential can be considered as part of the internal energy, it is listed as a separate item in the diagram.

There are four sources (sinks) of gravitational potential energy generated by external forcing. A major source is the tidal force, which is due to the temporal variation of the gravitational forces of the moon and the sun. Although tidal motions have been studied separately from the oceanic general circulation in the past, it is now recognized that a strong

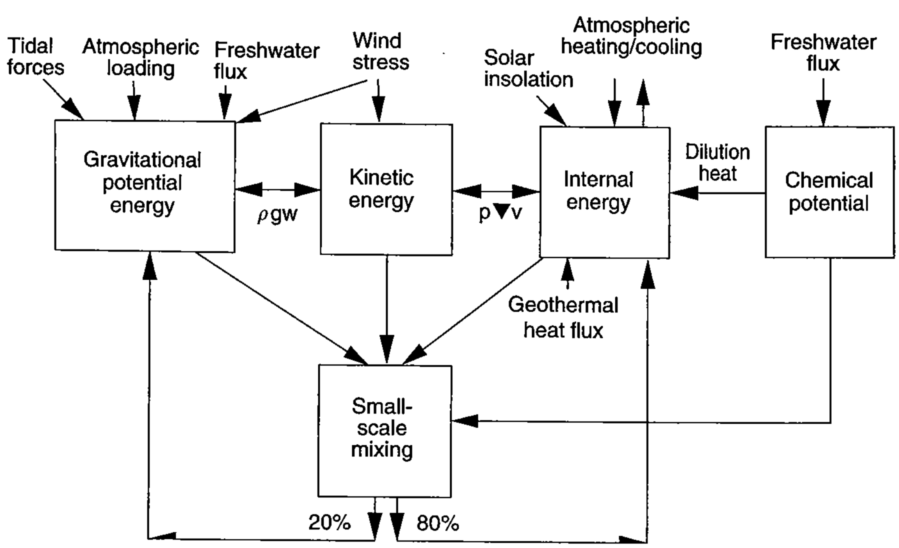

FIGURE 1 Energetics diagram for the ocean circulation.

connection exists between the tidal motions and the oceanic general circulation.

In addition, changes of the sea level atmospheric pressure (called the atmospheric loading) can generate gravitational potential energy. The total amount of this energy flux remains unclear. Furthermore, freshwater flux across the air-sea interface can generate gravitational potential energy. This turns out to be a small sink of energy, and the global integrated amount of energy flux is about $-0.007TW$ ($1TW = 10^{12}$ *Watts*). The reason for this loss of gravitational potential energy is that evaporation (precipitation) prevails at subtropics (high latitudes) where sea level is high (low) due to warm (cold) temperature. Wind stress is one of the most important sources of mechanical energy for the ocean. In fact, wind stress generates surface waves, thus providing a source of mechanical energy, which is equally partitioned into the gravitational potential energy and kinetic energy. The other source of kinetic energy is the energy input from the wind stress to the surface geostrophic currents and the ageostrophic currents or the Ekman drift.

The external sources of internal energy include three terms: the solar radiation, heat exchange to the atmosphere, and geothermal heat flux. Heat exchange to the atmosphere includes latent heat flux, long wave radiation, and sensible heat flux. Geothermal heat flux through the sea floor is much smaller than the air-sea heat fluxes, but it is one of the important factors controlling the deep circulation in the abyssal ocean.

The source of the chemical potential is associated with the salinity, which is generated by freshwater flux through the air-sea interface in forms of evaporation and precipitation. The general role of chemical potential in the oceanic circulation remains unclear. One of the major roles played by the chemical potential is driving the molecular mixing of salt.

2.1 Energy Conversion

Energy in any given form can be converted into other forms through specific processes, as highlighted in Fig. 1. For example, gravitational potential energy and kinetic energy can be transferred to each other through the vertical velocity conversion term. In addition, kinetic energy can be converted into gravitational potential energy through small scale mixing by internal waves and turbulence. On the other hand, gravitational potential energy can be converted into kinetic energy of meso-scale eddies through baroclinic instability. This may be one of the major pathways of energy flow in the ocean.

2.2 Conversion from Kinetic Energy to Gravitational Potential Energy

In a turbulent ocean, the conversion from kinetic to potential energy is represented by $\overline{\rho w g}$, where the overbar indicates the ensemble mean. This exchange can be separated into several terms:

$$\overline{\rho w g} = \bar{\rho}\bar{w}\bar{g} + \overline{\rho' w'}\bar{g} + (\overline{\rho' w' + \rho' \bar{w} + \bar{\rho} w'}).$$

The first term on the right-hand side represents the kinetic to potential energy conversion associated with the ensemble mean density and velocity fields. The second term on the right-hand side represents the contribution due to turbulent mixing. In a stratified fluid, upward motion is usually correlated with positive density anomaly, so turbulent mixing in a stratified ocean increases the gravitational potential energy of the state, and thus requires a source of kinetic energy. The rest of the terms on the right-hand side represent the kinetic to potential energy conversion due to the time-dependent component of the gravitational force, such as the tidal forces.

In the stratified oceans, the vertical eddy mixing rate κ can be related to the energy dissipation rate through

$$\kappa N^2 = a\varepsilon, N^2 = -\frac{g}{\rho_0}\frac{\partial \rho_\theta}{\partial z},$$

where N^2 is the buoyancy frequency, ε is energy dissipation rate, and $\alpha \approx 0.2$ is an empirical coefficient. Therefore, the decline in turbulent kinetic energy does not entirely become "wasted" heat. In fact, a small percentage of the turbulent kinetic energy "dissipation," represented by the empirical coefficient a, is actually transformed to large-scale potential energy. The energy required to sustain mixing is one of the most important dynamic factors controlling the meridional overturning rate.

2.3 Conversion of Internal Energy to Mechanical Energy

Internal energy can be converted to kinetic energy through the compression term. Since sea water is nearly incompressible, this conversion rate is rather low. In addition, it is very important to note that energy transformation is not unconditional. Instead, the conversion between internal energy and mechanical energy is subject to the second law of thermodynamics—that is, mechanical energy can be converted entirely into internal energy; however, internal energy can be converted into mechanical energy only partially, with the upper bound set by the

Carnot efficiency. As will be discussed later, however, the ocean is not a heat engine; therefore, the equivalent thermal-mechanical energy conversion efficiency is negative.

3. *HEAT FLUX IN THE OCEANS*

Oceans play an important role in the earth's climate system in many ways. In fact, most of the solar radiation penetrates the atmosphere and is first absorbed by the ocean and the land and then sent back to the atmosphere in forms of latent heat, long wave radiation, and sensible heat. To a large degree, these processes are controlled by the oceanic circulation. In the quasi-steady state, all heat flux absorbed by the ocean must be released back into the atmosphere. The spatial distribution of air-sea heat flux is one of the critical components of the earth's climate system.

3.1 Heat Flux through the Upper Surface of the Ocean

The dominating form of energy flux in the oceans is the heat flux. Solar radiation on Earth can penetrate the atmosphere, with only a small fraction being absorbed by the greenhouse gases, such as water vapor and CO_2. Since oceans cover about 70% of the earth, most of the solar radiation is directly absorbed by the oceans (total amount of $52.4PW$).

At low latitudes, solar radiation absorbed by the ocean exceeds the heat flux from the ocean into the atmosphere, including the latent heat flux, long wave radiation, and the sensible heat flux. Thus, the ocean at low latitudes gains net heat through the upper surface, and this extra heat is transported poleward and released at middle and high latitudes. The most important areas of the ocean-to-atmosphere heat release include the Gulf Stream, Kuroshio, and the Northern North Atlantic.

3.2 Poleward Heat Fluxes in the Climate System

Heat fluxes in the current climate system consist of three components: the atmospheric sensible heat flux, the oceanic sensible heat flux, and the latent heat flux associated with the hydrological cycle in the atmosphere-ocean coupled system (Fig. 2). The sum of these three fluxes is the total poleward heat flux diagnosed from satellite measurements. The ocean plays the role of heat storage and heat redistributor. The most important part for climate study is the total

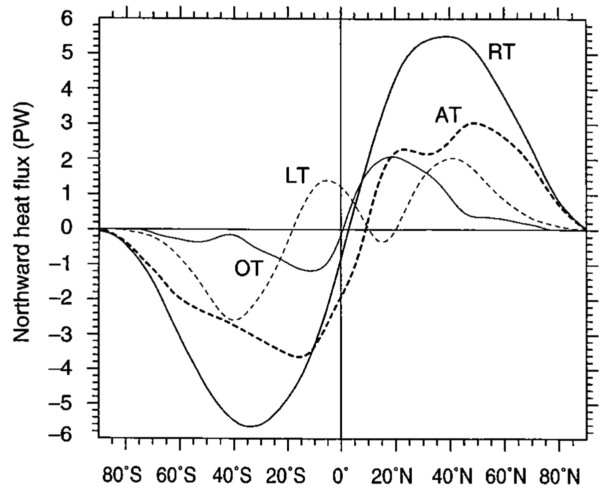

FIGURE 2 Northward heat flux in PW: RT indicates the heat flux required for radiation balance as diagnosed from satellite data; AT indicates the atmospheric sensible heat flux; OT indicates the oceanic sensible heat flux; and LT indicates the latent heat flux associated with the hydrological cycle.

amount of poleward heat flux associated with the oceanic current, which is about $2PW$. Note that the latent heat flux, which has been traditionally associated with the moisture flux in the atmosphere, can be considered as an atmosphere-ocean coupled mode.

Although the atmospheric sensible heat flux is the largest component in both hemispheres, it is not much larger than the other components. Thus, the total oceanic contribution to the poleward heat flux may consist roughly of 50% of the total heat flux.

The poleward sensible heat flux in the ocean can be separated into two major components: the flux associated with the meridional overturning cell and the flux associated with the horizontal gyration. Each component can be further divided into the contribution by the mean Eulerian flow and eddies. For example, poleward heat flux in the North Atlantic reaches its maximum around 25°N, where both the mean meridional overturning and horizontal wind-driven gyre contribute. In comparison, however, the contribution by eddies is relatively small. In the North Pacific, the mean meridional overturning cell induces an equatorward heat flux; however, the sum of heat fluxes is still poleward because it is associated with the domination of the wind-driven gyre.

3.3 Total Heat Flux through the Oceans

The oceanic to atmospheric heat flux consists of three components: the latent heat flux, $31.1PW$; the long wave radiation, $17.8PW$; and the sensible heat flux, $3.5PW$ (Fig. 3). Although air-sea interface heat flux is

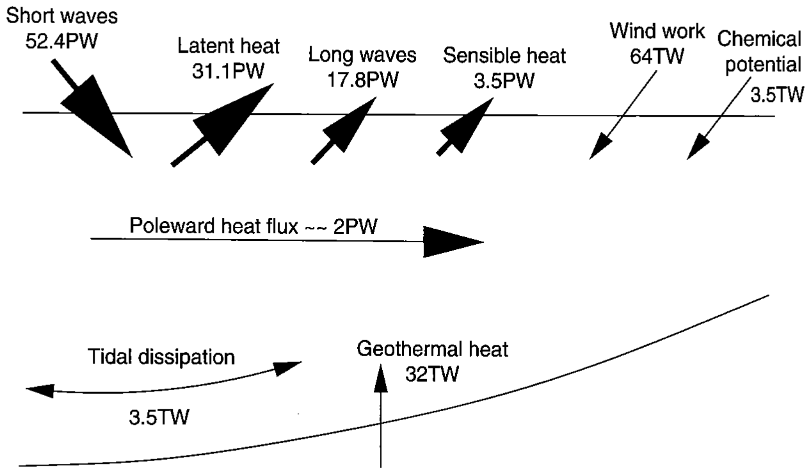

FIGURE 3 A diagram of external sources of energy for the ocean circulation.

of major importance to climate study, it does not directly drive the oceanic general circulation. Instead, air-sea heat flux is controlled by the oceanic circulation or, more accurately, controlled by the atmosphere-ocean-land coupled system. From the energetic point of view, the oceanic circulation is governed by many laws of conservation, in particular the energy conservation law. The thermohaline boundary conditions at the sea surface are preconditions for the thermohaline circulation only, while the ocean circulation is primarily controlled by the sources and sinks of mechanical energy, with the total air-sea heat flux as the by product of the circulation.

In Fig. 3, energy fluxes into the ocean in other forms are also included because this additional energy input eventually turns into internal energy and is released in to the atmosphere in forms of heat flux. The amount of geothermal heat flux is about $32TW$, which is eventually lost in to the atmosphere through the air-sea heat exchange. Furthermore, there are three other fluxes: the tidal dissipation, $3.5TW$; the chemical potential generated through evaporation and precipitation; and the wind work, $64TW$. Thus, the total outgoing heat flux is the sum of the incoming heat flux, $52.4PW$, plus four additional energy sources: 64 (wind) $+$ 32 (geothermal) $+$ 3.5 (tides) $+ 3.5TW$ (chemical potential) $= 103TW$.

4. *THE OCEAN IS NOT A HEAT ENGINE*

4.1 Is the Ocean a Heat Engine?

The atmosphere and oceans work together as a heat engine, if we neglect the small contribution of tidal energy to the circulation in the oceans. The atmosphere can be considered as a heat engine, which is driven by differential heating, with an efficiency of 0.8% (the corresponding Carnot efficiency is about 33%). In this sense although the oceans are subject to differential heating similar to the atmosphere, they are not at all a heat engine. As will be discussed here, the driving force for the oceanic circulation is the mechanical energy in the forms of wind stress and tides. In comparison, differential heating is only a precondition for the thermohaline circulation, and not the driving force of the oceanic general circulation. Thus, the ocean is not a heat engine; instead, it is a machine driven by external mechanical energy that transports thermal energy, freshwater, CO_2, and other tracers.

4.2 Sandstrom's Theorem

Sandstrom theorem states: A closed steady circulation can be maintained in the ocean only if the heating source is situated at a level lower than the cooling source. Sandstrom's theorem can be demonstrated by laboratory experiments, as shown in Fig. 4. In the first experiment, the heating source is put at a level higher than the cooling source. There is no detectable circulation, and a stable stratification can be observed between the heating and cooling levels. Theoretically, there is still a circulation driven by the molecular mixing, but it is so sluggish that there is practically none. In the second experiment, the heating source is put at a level lower than the cooling source. A vigorous circulation can be observed between the levels of heating and cooling, just as we put a pot of water over the kitchen range.

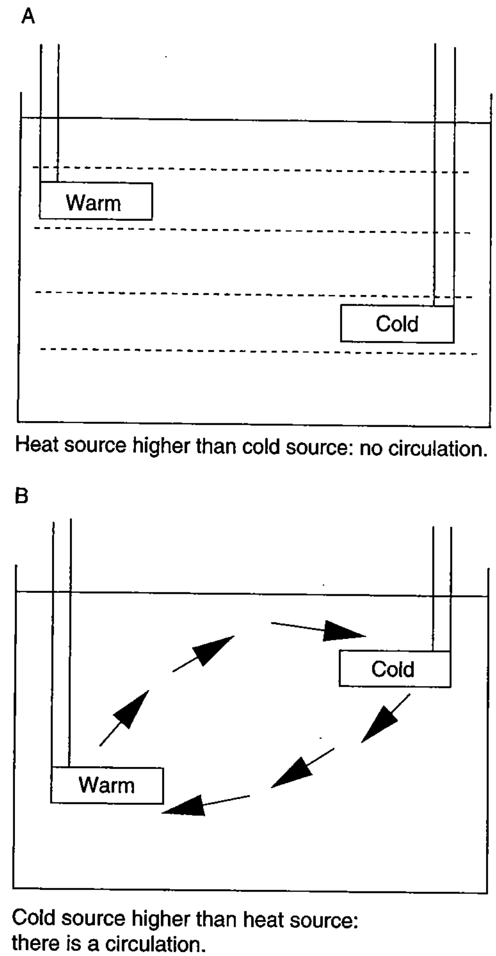

FIGURE 4 Laboratory experiments demonstrating the Sandstrom theorem.

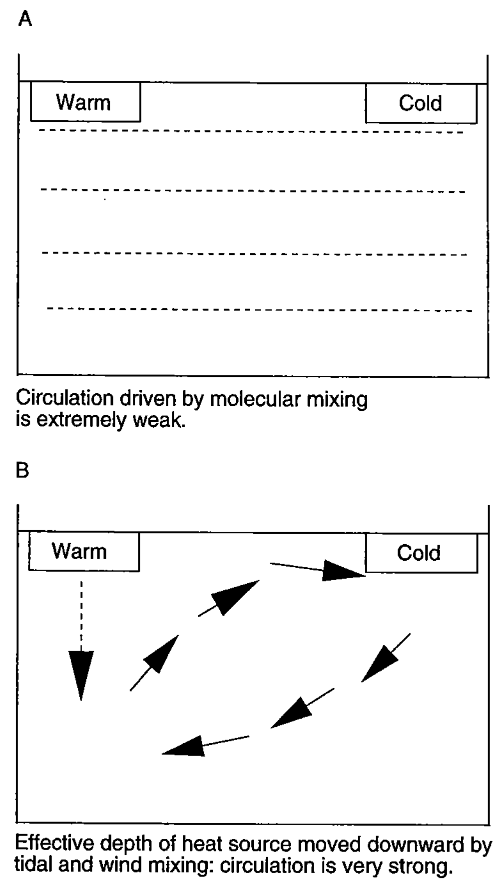

FIGURE 5 Application of the Sandstrom theorem to the world's oceans.

Sandstrom's theorem applies to both the atmosphere and oceans. The atmosphere is heated from below and cooled from above; thus, it is a heat engine. On the other hand, the ocean is heated and cooled from the sea surface. Since heating and cooling take place at the same level, there is no circulation driven by the thermal forcing (Fig. 5A). Under such a condition, thermal forcing can create a very thin thermocline, on the order of meters on the surface of the oceans. Below such a thin thermocline, temperature is virtually constant, equal to the coldest temperature set up by the deepwater formation at high latitudes. Since there is no density difference at depth, there is only a very sluggish circulation. However, the effective depth of heating is moved downward by wind and tidal mixing. The large horizontal density gradient thus created drives a strong circulation in the world oceans (Fig. 5B).

The essential point is that to maintain a quasi-steady circulation in the ocean, external sources of mechanical energy are required to balance the loss of mechanical energy caused by friction associated with motion. Although there is a tremendous amount of thermal energy going through the ocean, it cannot be converted into mechanical energy needed to sustain the oceanic circulation, as dictated by the fundamental law of thermodynamics. Therefore, instead of utilizing the huge amount of thermal energy going through the ocean, the maintenance of the oceanic circulation depends on the availability of external mechanical energy, with a flux rate of three to four orders of magnitude smaller than the thermal energy fluxes (Fig. 3).

5. MECHANICAL ENERGY FLOWS IN THE OCEAN

Mechanical energy sources and sinks for the oceans are illustrated in Fig. 6. Wind stress applies to the

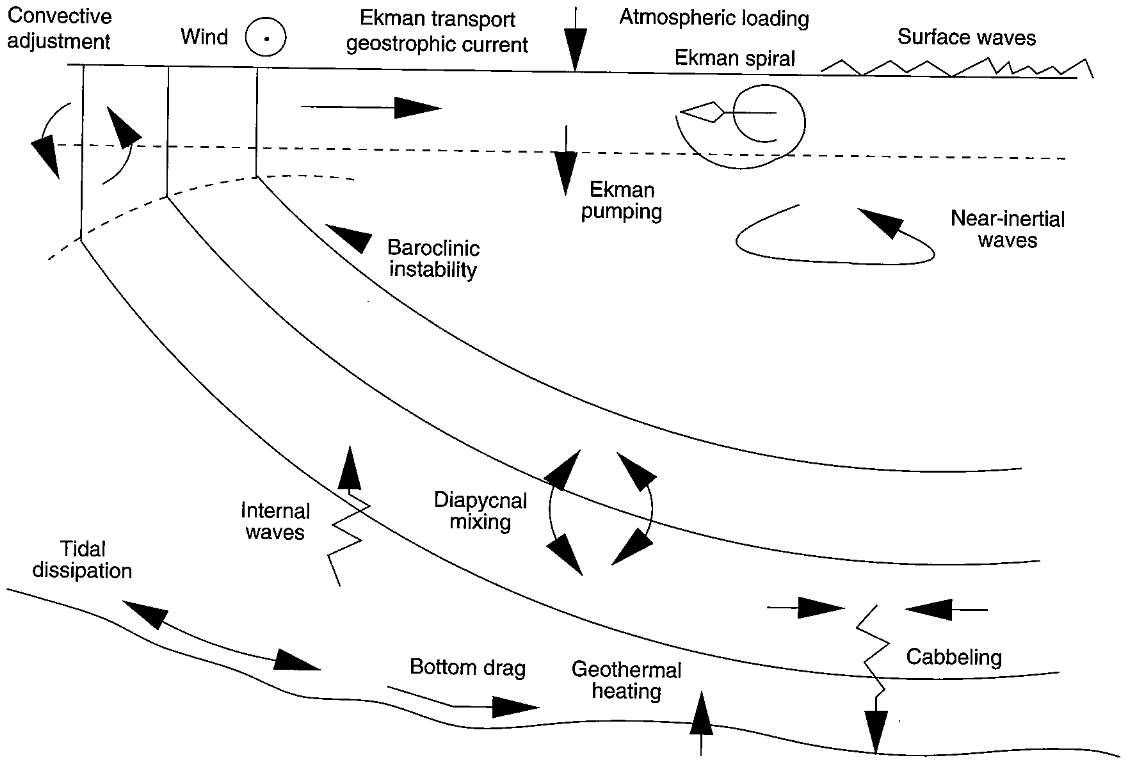

FIGURE 6 Mechanical energy diagram for the ocean circulation.

surface of the ocean and drives both surface currents and waves. The mechanical energy input from wind stress into the ocean W_{wind} is defined by

$$W_{wind} = \langle \sigma_{ij} \rangle \cdot \vec{u} = \vec{\tau} \cdot \vec{U}_0 + \overline{\vec{\tau}'\vec{u}'_0} + \overline{p'w'_0},$$

where $\vec{\tau}$ and $\vec{U}_0$ are the spatially averaged tangential stress and surface velocity respectively, $\vec{\tau}'$ and $\vec{u}'_0$ are the perturbations, and p' and w'_0 are perturbations of the surface pressure and velocity component normal to the surface. The perturbations are defined in terms of the scale of the surface wavelength.

The first term on the right-hand side is the wind stress work of the quasi-steady current on the surface, and quasi-steadiness is defined in comparison to the timescale of the typical surface waves. $\vec{U}_0$ has two components: the geostrophic current and the Ekman drift:

$$\vec{U}_0 = \vec{U}_{0,G} + \vec{U}_{0,Ekman.}$$

5.1 Geostrophic Current

Energy input through the surface geostrophic current can be calculated as the scaler product of wind stress and surface geostrophic velocity. The recent calculation by Carl Wunsch puts the global total as $1.3TW$. This energy is directly fed into the large-scale current, so it can be efficiently turned into gravitational potential energy through the vertical velocity conversion term.

5.2 Surface Drift

There is a frictional boundary layer (Ekman layer) on the top of the ocean, in which the wind stress is balanced by frictional force. The surface drift is called the Ekman drift. The exact amount of wind stress energy input through the Ekman spiral is unclear, and the preliminary estimate made by Wei Wang and Rui Xin Huang is about $2.4TW$ for the frequency higher than 1/(2 days). In addition, there is a large amount of energy input through the near-inertial waves, which is due to the resonance at a frequency of $\omega = -f$; that is, for inertial frequency of the clockwise (anticlockwise) wind stress in the Northern (Southern) Hemisphere. Recent calculations by Watanaba and Hibiya gave an estimate of $0.7TW$.

5.3 Geopotential and Kinetic Energy of the Wind-Driven Circulation

The convergence of the Ekman flux gives rise to a pumping velocity at the base of the Ekman layer W_E, which is responsible for pushing the warm water into the subsurface ocean and thus forming the main thermocline in the subtropical ocean. The main thermocline is a relatively thin layer of water at the depth of 500 to 800 m in the upper ocean where the vertical temperature gradient is maximum. Since salinity contrition to the density is relatively small in most places of the oceans, the thermocline is also close to the pycnocline where the vertical gradient of density is maximum. Ekman pumping sets up the wind-driven circulation and the associated bow-shaped main thermocline in the world's oceans. In this process, gravitational potential energy is increased, which is a very efficient way of converting kinetic energy into gravitational potential energy. Simple scaling analysis shows that the amount of available gravitational potential energy associated with the main thermocline is about 1000 times larger than the amount of kinetic energy associated with the mean flow of the wind-driven circulation.

5.4 Loss of Gravitational Potential Energy through Convective Adjustment

Another important process taking place in the mixed layer is the convective adjustment due to cooling and salinification. According to the Sandstrom theorem, if both heating and cooling apply to the same level (the sea surface), there is no gravitational potential energy being generated by the thermal forcing alone. However, heating and cooling in the mixed layer is not a linear process. During the cooling process, water at the sea surface becomes heavier than water at the depth, thus a gravitational unstable stratification appears. Due to this unstable stratification, a rapid convective adjustment process takes place and density becomes nearly homogenized in the upper part of the water column. During this process, gravitational potential energy of the mean state is converted into kinetic energy for turbulence and internal waves (Fig. 7A).

As a result, the effective center of cooling is not at the sea surface; instead it is located halfway of the well-mixed layer's depth. Because of this asymmetry associated with cooling/salinification, surface buoyancy forcing gives rise to a sink of gravitational potential energy. The total amount of energy loss through this process remains unclear. A preliminary estimate based on the monthly mean climatology by Rui Xin Huang and Wei Wang for this sink term is about 0.24*TW*. However, calculation in which the diurnal cycle is resolved may give rise to a much larger value.

5.5 Baroclinic Instability

The steep isopycnal surface in the oceans, such as that along the meridional edges of the wind-driven gyre, is unstable due to baroclinic instability. Because of this, a large amount of gravitational potential energy of the mean state can be converted into the eddy kinetic/potential energy (Fig. 7B). It is estimated that the amount of eddy kinetic energy is about 100 times larger than the kinetic energy of the time-mean flow. Despite great effort in carrying out field observations and numerical simulation, there is no reliable estimate on the total amount of eddy kinetic energy in the world's oceans. However, satellite observations have provided global distribution of both the mean and eddy kinetic energy on the sea surface. The ratio of these two forms of kinetic energy is on the order of 100, consistent with

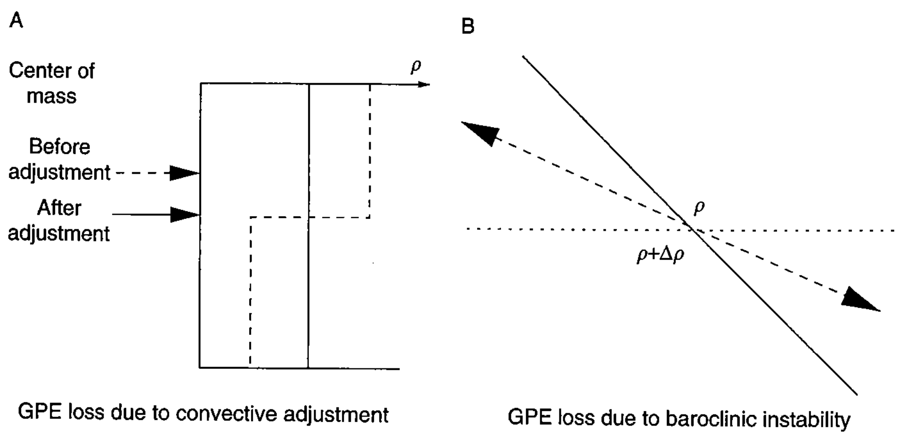

FIGURE 7 Gravitational potential energy loss due to convective adjustment and baroclinic instability.

theoretical prediction based on scaling. The rate of energy conversion over the world's oceans through the baroclinic instability remains unknown, and the recent estimate by Rui Xin Huang and Wei Wang is about 1.1*TW*. Since most eddy energy is dissipated through small-scale processes, the conversation from the mean state to eddies is considered as a sink of the gravitational potential energy of the mean state.

5.6 Surface Waves

Another major source of energy caused by wind stress is fed through the surface waves. Wind stress drives surface waves in the oceans, and this energy input can be treated as the form drag for the atmospheric boundary layer. The rate of energy input through surface waves remains unclear; however, the preliminary estimate by Wei Wang and Rui Xin Huang is about 60*TW*. Although this seems to be a large amount of energy, it is believed that most of it is transformed into long waves and propagates away from the source area. It is likely that only a small fraction of this energy is dissipated locally through wave breaking and white capping, but a major fraction of this incoming energy may be dissipated to remote places in forms of swell. These long waves have a rather low dissipation rate, their energy is likely to dissipate cascading into other forms of energy or through wave breaking along the beaches in the world's oceans; however, the details of this dissipation mechanism remain unclear.

5.7 Diapycnal and along Isopycnal Mixing

Tracers, including temperature and salinity, are mixed in the oceans through internal wave breaking and turbulence. Mixing can be classified into diapycnal mixing and along isopycnal mixing. Since along isopycnal mixing involves the least amount of gravitational potential energy, it is the dominating form of mixing on large scales.

On the other hand, in stratified fluid vertical mixing increases gravitational potential energy because light fluid is pushed downward and heavy fluid is pushed upward. Although molecular diffusion can play the role of mixing, the corresponding rate is too small and thus the result is irrelevant to the oceans. In the upper ocean, below the mixed layer, diapycnal mixing rate is on the order of $10^{-5} m^2 s^{-1}$, which is much larger than the mixing rate due to molecular mixing. In other places of the oceans, the mixing rate can be much higher than this background rate. The strong diapycnal (or vertical) mixing in the oceans is driven by strong internal wave and turbulence.

The energy sources supporting diapycnal mixing in the oceans include wind stress input through the sea surface and tidal dissipation, which will be discussed in the next section. Turbulent mixing associated with fast current, especially that associated with flow over sill and the down-slope flow afterward, can provide a strong source of energy supporting mixing. There is much observational evidence indicating that mixing can be on the order of 10^{-3}–$10^{-1} m^2 s^{-1}$. The mixing rate can be as large as $0.1\, m^2 s^{-1}$ within the Romanche Fracture Zone, where water drops more than 500 m within 100 km of a downward flow.

Internal lee waves generated by flow over topography, such as the Antarctic Circumpolar Current, are probably one of the most important contributors.

The total amount of energy supporting diapycnal mixing in the oceans remains unclear because mixing is highly nonuniform in space and time. According to the theory of thermohaline circulation, the meridional overturning rate is directly controlled by the strength of the external mechanical energy available for supporting mixing in the oceans, including mixing in the mixed layer and mixing in the subsurface layers. Therefore, understanding the physics related to the spatial and temporal distribution of mixing is one of the most important research frontiers in physical oceanography.

5.8 Tidal Dissipation

The primary source of mechanical energy supporting mixing likely comes from both the wind stress applied to the ocean's surface and tidal dissipation in the deep ocean. The total amount of tidal dissipation in the world's oceans is 3.5*TW* (Fig. 8), which is calculated from accurate tracking of the moon's orbit. The spatial distribution of tidal energy dissipation remains inaccurate. Based on satellite altimeter data assimilation, Walter Munk and Carl Wunsch estimated that 2.6*TW* is dissipated within the shallow seas of the world's oceans, and the remainder 0.9*TW* is believed to be distributed in the deep ocean (Fig. 6). Tidal dissipation in the deep ocean takes place primarily over rough topography, where barotropic tidal energy is converted into energy for internal tides and internal waves that sustain the bottom-intensified diapycnal mixing on the order of $10^{-3} m^2 s^{-1}$. The conversion of kinetic energy to gravitational potential energy through

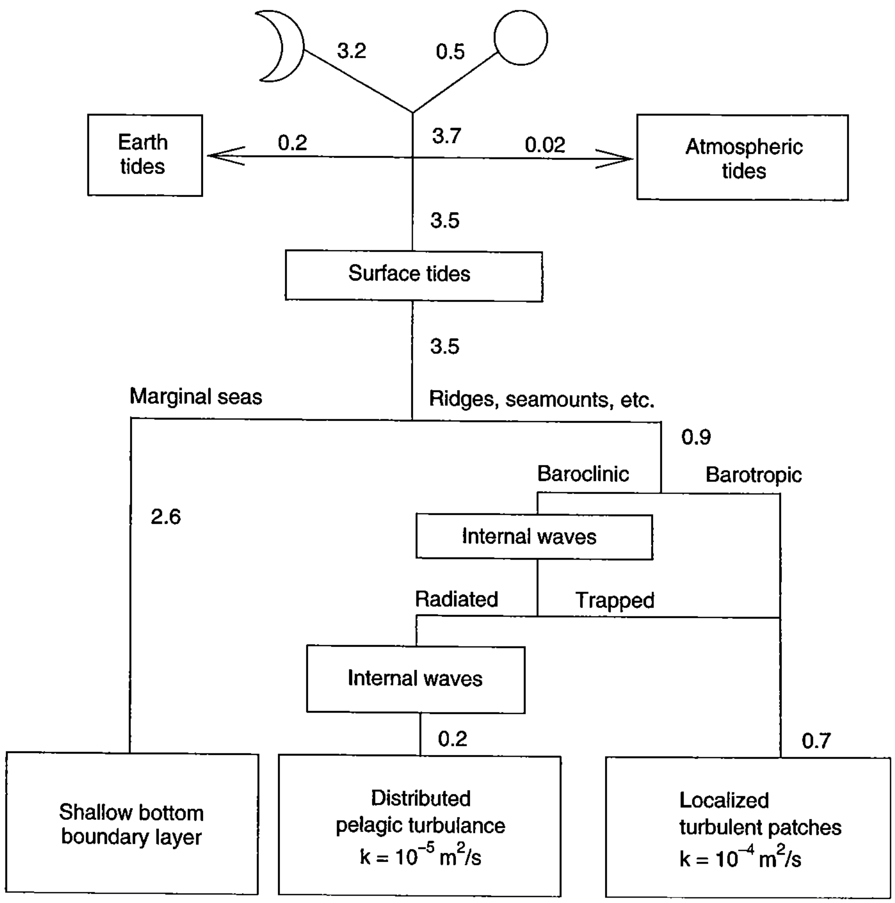

FIGURE 8 Tidal dissipation diagram, unit in *TW*. Modified from Munk and Wunsch (1998).

internal waves and turbulence is of rather low efficiency, typically on the range of 20%; thus, the amount of gravitational potential energy generated by tidal dissipation in the deep ocean is about 0.18*TW*.

Tidal dissipation has varied greatly over the geological past, thus the energy supporting diapycnal mixing may vary as well. The reader is referred to the review by Kagan and Sundermann.

5.9 Geothermal Heat

In addition, geothermal heat or hot plumes provide a total heat flux of 32*TW*. Although this is much smaller than the heat flux across the air-sea interface, it may be a significant component of the driving force for the abyssal circulation. Since the geothermal heating is applied at great depth, and the corresponding cooling takes place at the sea surface, geothermal heat can be more efficiently converted into gravitational potential energy. The rate of this conversion is about 0.05*TW*, which is a small term compared to other major terms; nevertheless it is not negligible, especially for the abyssal circulation and temperature distribution in the abyss.

5.10 Bottom Drag

Oceanic currents moving over rough bottom topography must overcome bottom or form drag, as shown in Fig. 6. The total amount of bottom drag for the world's oceans circulation remains unclear; however, the preliminary estimate for the open oceans is about 0.4*TW*.

5.11 Atmospheric Loading

Finally, sea level atmospheric pressure varies with time, and the sea surface moves in the vertical direction in response. As a result, changes in sea level atmospheric pressure can input mechanical energy into the oceans. The total amount of energy due to this source for the world's ocean circulation remains unclear; however, the preliminary estimate for the open oceans is about 0.04*TW*.

5.12 Miscellaneous

Many other mechanisms contribute to the mechanical energy balance in the oceans.

Cabbeling. Due to the nonlinearity of the equation of state, water density is increased during both diapycnal and isopycnal mixing. The newly formed water with higher density sinks is called cabbeling, as shown in Fig. 6. As a result, gravitational potential energy is converted into energy for internal waves and turbulence. The total amount of gravitational potential energy loss associated with cabbeling in the oceans remains unclear.

Double diffusion: Seawater contains salt, thus it is a two-component chemical mixture. On the level of molecular mixing, the heat diffusion is 100 times faster than the salt diffusion. The difference in heat and salt diffusion for laminar fluid and turbulent fluid environment is one of the most important aspects of the thermohaline circulation in the oceans.

Double diffusion in the oceans primarily manifests itself by two types: salt fingers and diffusive convection. Salt fingers appear when warm and salty water lie over cold and freshwater. The mechanism that drives the instability is illustrated in Fig. 9A, assuming a vertical pipe connects the cold and freshwater in the lower part of the water column with the warm and salty water in the upper part of the water column. The wall of the pipe is very thin, so it allows a rather efficient heat flux into the pipe thus warming up the cold water. At the same time, the low salt diffusivity preserves the freshness of the ascending water parcel. Thus, the buoyancy difference drives the upward motion of water inside the pipe, and a self-propelled fountain can be built in the ocean.

The subtropical gyre interior is a salt-finger's favorite place, where strong solar insolation and excessive evaporation leads to warm and salty water above the main thermocline. In this salt-finger favorite regime, warm and salty fingers move downward and cold and fresh plumes move upward. Since heat is 100 times more easily diffused between the salt fingers and the environment, salt fingers lose their buoyancy and continue their movement downward. In this way gravitational potential energy released is used to drive the mixing. According to the new insite observations in the upper thermocline of the subtropical North Atlantic, the mixing rate for the temperature (salinity) is on the order of $4 \times 10^{-5} m^2 s^{-1} (8 \times 10^{-5} m^2 s^{-1})$, and the equivalent density mixing rate is negative. The total amount of gravitational energy loss due to salt fingering in the world's oceans remain unknown.

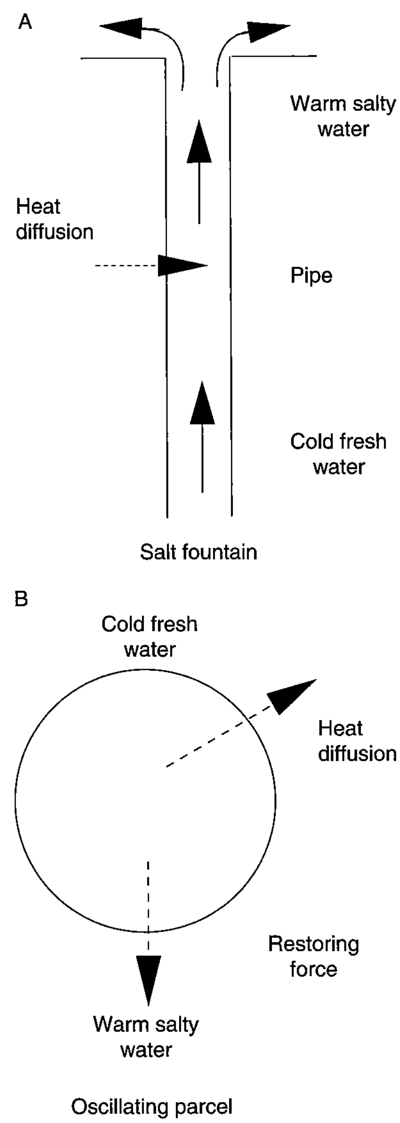

FIGURE 9 Two possible cases for double diffusion in the oceans.

The other possible double diffusive process is the diffusive convection which takes place if cold and freshwater overlay warm saltwater. The system is relatively stable and allows an oscillatory instability (Fig. 9b).

5.13 Attempt at Balancing the Mechanical Energy in the Ocean

The balance of mechanical energy, including both the kinetic energy and gravitational potential energy, is shown in Fig. 10. It is clear that at this time we do not know even the lowest-order balance of mechanical energy. There is much kinetic energy input into

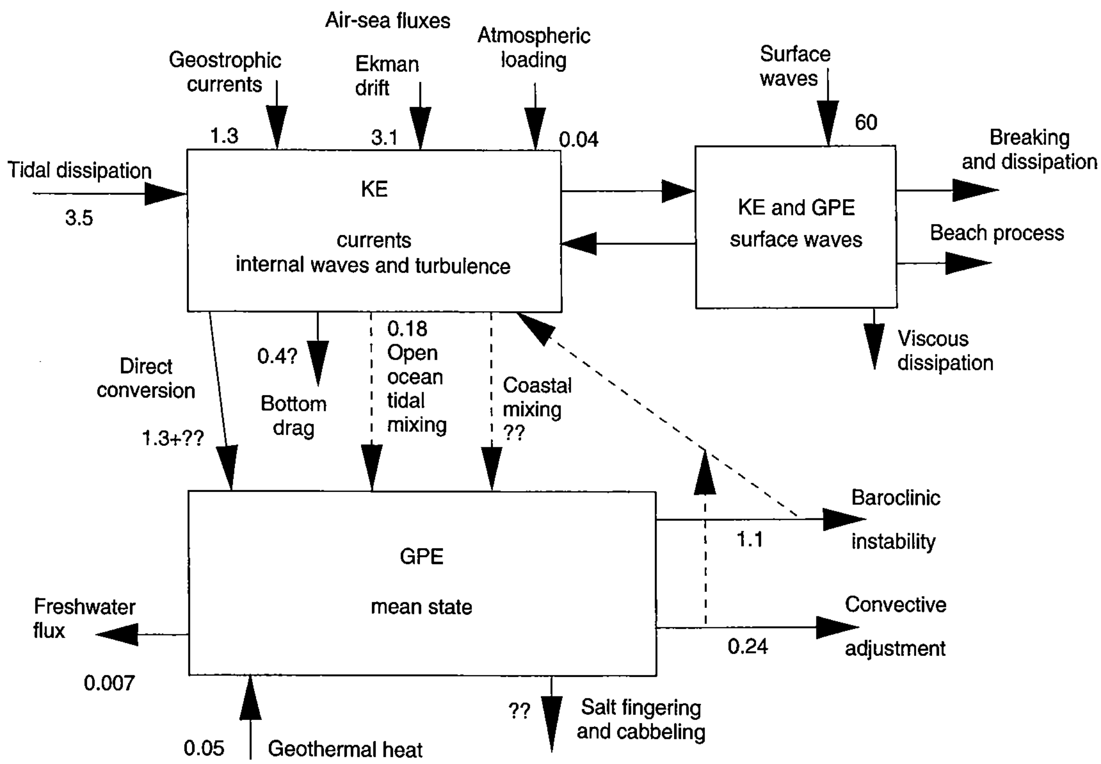

FIGURE 10 Mechanical energy balance for the world's oceans, unit in *TW*.

the ocean, but it is not clear how such energy is distributed and eventually dissipated in the oceans. On the other hand, there are two major sinks of gravitational potential energy, but it is not clear how energy is transported and supplied to such energy sinks. It is fair to say that most of the energy fluxes listed in Fig. 10 are accurate to the factor of 2 only. More accurate energy pathways and estimates require further study.

6. *MAJOR CHALLENGE ASSOCIATED WITH ENERGETICS, OCEANIC MODELING, AND CLIMATE CHANGE*

6.1 Scaling Laws Governing the Meridional Circulation

Scaling analysis shows that the meridional mass flux and heat flux are linearly proportional to the mechanical energy available for mixing. Thus, the amount of energy available for supporting mixing is what really controls the strength of circulation, so more energy available for mixing means stronger circulation.

6.2 Energetics and Numerical Simulation of Ocean Circulation

Since we do not well know the distribution of mechanical energy available for mixing, many oceanic general circulation models of the current generation are based on rather simple parameterization of subgrid mixing. A common practice in oceanic general circulation modeling is to choose a diapycnal mixing rate and let the models run; the diapycnal mixing rate remains the same all the time, even under different climatic conditions. In fact, no modelers have ever checked how much energy is required to support mixing and circulation and, most important, whether this amount of energy is really available.

An interesting and important question is how to parameterize mixing in climate-related models. Should we use the same mixing rate or same mixing energy when the climate is drifting? Clearly, both a fixed mixing rate and a fixed mixing energy represent the two extremes, and the reality is likely to be between these two assumptions.

It is speculated that when we have the new generation of the Oceanic General Circulation Model (OGCM), in which the mixing parameterization is more rational, climate variability induced by something like the CO_2 doubling may be different from whatever we have learned based on existing models.

7. CONCLUSION

In summary, energetics of oceanic circulation is of capital importance in understanding ocean circulation and climate. Energetics of the oceanic circulation is one the most important research frontiers. There are many unanswered fundamental questions, and at this time, we do not have the lowest-order balance. Most flux terms cited here should be treated as preliminary estimates, which will certainly be improved in the near future. Since circulation in the oceans is primarily controlled by energy available for mixing, further study of energetics will bring about major breakthroughs in our understanding of the ocean circulation and climate.

SEE ALSO THE FOLLOWING ARTICLES

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Further Reading

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