

Thermodynamics of the Oceanic General Circulation – Is the Abyssal Circulation a Heat Engine or a Mechanical Pump?

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1. Introduction

The oceanic general circulation has been investigated mainly from a dynamic perspective. Nevertheless, some important contributions to the field have been made also from a thermodynamic viewpoint. This chapter presents description of the thermodynamics of the oceanic general circulation. Particularly, we examine entropy production of the oceanic general circulation and discuss its relation to a thermodynamic postulate of a steady closed circulation such as the oceanic general circulation: Sandström's theorem. Also in this section, we refer to another important thermodynamic postulate of an open non-equilibrium system such as the oceanic general circulation: the principle of Maximum Entropy Production.

1.1 Outline of oceanic general circulation

Oceanic general circulation is the largest current in the world ocean, making a circuit from the surface to the bottom over a few thousand years. The present oceanic general circulation, briefly speaking, is a series of flows, in which seawater sinks from restricted surface regions in high latitudes of the Atlantic Ocean to the deep bottom ocean. It later comes to broad surface regions of the Pacific Ocean, and returns to the Atlantic Ocean through the surface of the Indian Ocean (see Fig. 1). The atmosphere affects the daily weather, whereas the ocean affects the long-term climate because of its larger heat capacity. Therefore, it is important for our life to elucidate the oceanic general circulation.

The causes generating the oceanic general circulation are momentum flux by wind stress at the sea surface and density flux by heating, cooling, precipitation, and evaporation through the sea surface, except for tides. In general, the oceanic general circulation is explained as consisting of surface (wind-driven) circulation attributable to the momentum flux and abyssal (thermohaline) circulation caused by the density flux. However, the distinction between them is not simple because diapycnal mixing, which is important for abyssal circulation, depends largely on wind, as described in the next sub-section. Moreover, diapycnal mixing depends also on tides.

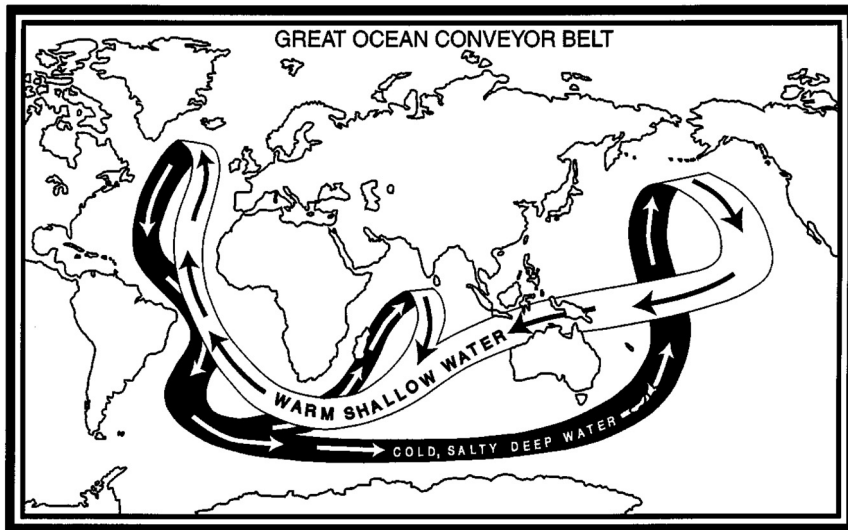


Fig. 1. Illustration of oceanic general circulation (Broecker, 1987).

1.2 Energy sources of abyssal circulation

Sustained abyssal circulation is a manifestation of conversion of potential energy to kinetic energy within the system. Production of potential energy is mainly the result of diapycnal mixing in the ocean interior, geothermal heating through the ocean floor, and the meridional distribution of precipitation, evaporation, and runoff (e.g., Gade & Gustafsson, 2004).

Diapycnal mixing results from turbulent diffusion by wind and tides. The most reasonable mechanism to transfer energy from the surface to the deeper layer is regarded as breaking and wave-wave interaction of internal waves generated by wind and tides (e.g., Muller & Briscoe, 2000). The wind and tidal dissipation quantities have been estimated respectively as about 1 TW (Wunsch, 1998) and 1 TW (Egbert & Ray, 2000). Using these estimates and $R_f = 0.15$ (Osborn, 1980) as the flux Richardson number, $\gamma = R_f / (1 - R_f) = 0.18$ as the ratio of potential energy to available energy, and $S = 3.6 \times 10^{14} \text{ m}^2$ as the total surface area of the ocean, the production of potential energy caused by diapycnal mixing has been estimated as about $1.0 \times 10^{-3} \text{ W m}^{-2}$ ($= 2\text{TW} / (3.6 \times 10^{14} \text{ m}^2) \times 0.18$).

Geothermal heating through the ocean floor causes a temperature increase and a thermal expansion in seawater, and generates potential energy. Production of potential energy caused by geothermal heating has been estimated as about 0.11 (Gade & Gustafsson, 2004) - 0.14 (Huang, 1999) $\times 10^{-3} \text{ W m}^{-2}$.

Precipitation (evaporation) is a flux of mass to (from) the sea surface and consequently a flux of potential energy. On average, the warm (cold) tropics with high (low) sea level are regions of evaporation (precipitation). These therefore tend to reduce the potential energy. The value integrated for the entire ocean shows a net loss of potential energy. Loss of potential energy attributable to precipitation, evaporation, and runoff has been estimated as less than 0.02 (Gade & Gustafsson, 2004) - 0.03 (Huang, 1998) $\times 10^{-3} \text{ W m}^{-2}$. These contributions can be negligible.

In addition, there can be work done on the ocean by surface heating and cooling. Heating (cooling) causes an expansion (contraction) with a net rise (fall) in the centre of mass and an increase (decrease) in potential energy. The exact estimate of the effect is difficult, but it will be small compared to the effect of the wind forcing. The best recent estimate of work done on the ocean by surface heating and cooling is zero (Wunsch & Ferrari, 2004).

1.3 "Missing mixing" problem

Munk (1966) estimated that the magnitude of diapycnal mixing to drive and maintain abyssal circulation is about $K \approx 10^{-4} \text{ m}^2 \text{ s}^{-1}$. He reached that figure by fitting of vertical profiles of tracers with one-dimensional vertical balance equation of advection and diffusion as

$$K \frac{d^2 T}{dz^2} = w \frac{dT}{dz}, \quad (1)$$

where K is a diapycnal mixing coefficient, T denotes a tracer variable such as temperature, salinity and radioactive tracers, z signifies a vertical coordinate, and w represents the upwelling velocity. The estimated value has been regarded as reasonable because the total upwelling of deep water estimated using the above K is consistent with the total sinking of deep water estimated by observations in the sinking area.

However, some direct observations of turbulence (Gregg, 1989) and dye diffusion (Ledwell et al., 1993) in the deep ocean indicate a diapycnal mixing of only $K \approx 10^{-5} \text{ m}^2 \text{ s}^{-1}$. Moreover, this is consistent with mixing estimated from the energy cascade in an internal wave spectrum (called "background") (McComas & Mullar, 1981). This difference of K is designated as the "missing mixing" problem.

On the other hand, recent observations of turbulence show larger diapycnal mixing of $K \geq 10^{-4} \text{ m}^2 \text{ s}^{-1}$ (Ledwell et al., 2000; Polizin et al., 1997), although such observations are limited to areas near places with large topographic changes such as seamounts (called "hot spots"), where internal waves are strongly generated as sources of diapycnal mixing. Munk & Wunsch (1998) reported that the value averaged over the entire ocean including "background" and "hot spots" can be about $K \approx 10^{-4} \text{ m}^2 \text{ s}^{-1}$, which remains controversial.

1.4 Abyssal circulation as a heat engine or a mechanical pump

Traditionally, the abyssal circulation has been treated as a heat engine (or a buoyancy process) driven by an equatorial hot source and polar cold sources. Broecker & Denton (1990) reported that abrupt changes in the ocean's overturning causes the ocean's heat loss, which might engender large swings in high-latitude climate, such as that occurring during the ice age. They also suggested a descriptive image of abyssal circulation: a conveyor-belt (see Fig. 1). Peixoto & Oort (1992) investigated the atmosphere-ocean system as a heat engine using the concept of available potential energy developed by Lorenz (1955).

Toggweiler (1994) reported that the abyssal formation in the North Atlantic is induced by upwelling because of strong surface wind stress in the Antarctic circumpolar current (a mechanical pump or a mechanical process). This mechanism is inferred from the "missing mixing" problem, as stated in section 1.3. If "background" diapycnal mixing for maintaining abyssal circulation is weaker than Munk's estimate, then another new mechanism to pump

up water from the deep layer to the surface is needed, provided that sinking can occur in the cold saline (i.e. dense) region of the North Atlantic. Drake Passage is located in the region of westerly wind band where water upwells from below to feed the diverging surface flow. Because net poleward flow above the ridges is prohibited (there is no east-west side wall to sustain an east-west pressure gradient in the Antarctic circumpolar current region), the upwelled water must come from below the ridges, i.e., from depths below 1500–2000 m. In addition, very little mixing energy is necessary to upwell water because of weak stratification near Antarctica.

1.5 Sandström theorem

Related to a closed steady circulation such as abyssal circulation, there is an important thermodynamic postulate: Sandström's theorem (Sandström, 1908, 1916)¹.

Sandström considered the system moving as a cycle of the heat engine with the following four stages (see Fig. 2).

1. Expansion by diabatic heating under constant pressure
2. Adiabatic change (expansion or contraction) from the heating source to the cooling source
3. Contraction by diabatic cooling under constant pressure
4. Adiabatic change (contraction or expansion) from the cooling source to the heating source

When the system moves anti-clockwise (expansion in stage 2 and contraction in stage 4), i.e., the heating source ($d\alpha > 0$; α is a specific volume that is equal to the volume divided by the mass) is located at the high-pressure side and the cooling source ($d\alpha < 0$) is located at the low-pressure side (Fig. 2a; $P_{\text{heating}} > P_{\text{cooling}}$), the work done by the system is positive:

$$\oint P d\alpha > 0. \quad (2)$$

In contrast, when the system moves clockwise (contraction in stage 2 and expansion in stage 4), i.e., the cooling source is located at the high-pressure side and the heating source is located at the low-pressure side (Fig. 2b; $P_{\text{heating}} < P_{\text{cooling}}$). Therefore, the work done by the system is negative:

$$\oint P d\alpha < 0. \quad (3)$$

Consequently, Sandström suggested that a closed steady circulation can only be maintained in the ocean if the heating source is located at a higher pressure (i.e. a lower level) than the cooling source.

Regarding the atmosphere, the heating source is located at the ground surface and the cooling source is located at the upper levels because the atmosphere is almost transparent to shortwave radiation of the sun, which heats the ground surface directly. Then heat is transferred from the heated surface by vertical convection. Therefore, the atmosphere can be regarded as a heat engine.

¹ An English translation of Sandström (1906) is available as an appendix in Kuhlbrodt (2008), but the Sandström papers are written in German, and are not easy to obtain. Other explanations of Sandström's theorem can be found in some textbooks of oceanic and atmospheric sciences: Defunt (1961), Houghton (2002), and Huang (2010).

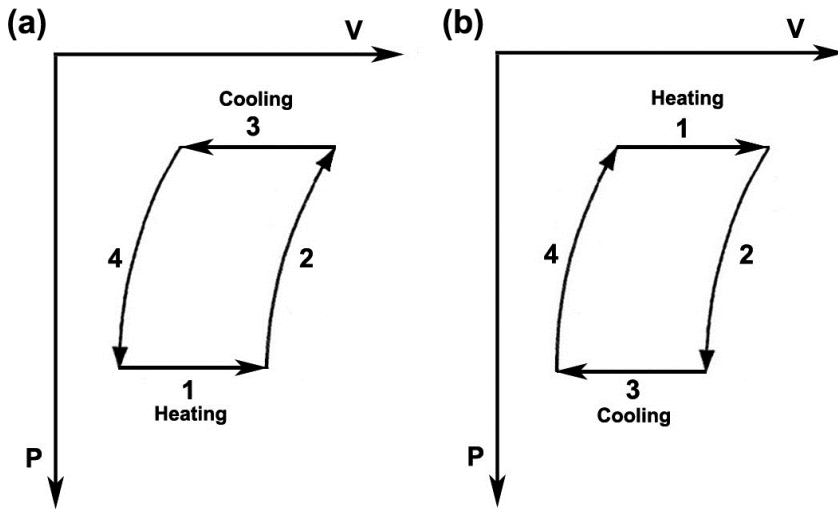


Fig. 2. Heat engines of two types discussed by Sandström (1916): (a) anti-clockwise and (b) clockwise.

1.6 Principle of maximum entropy production and oceanic general circulation

In this sub-section, we briefly explain another important thermodynamic postulate of stability of a nonlinear non-equilibrium system such as the oceanic general circulation, the principle of the maximum Entropy Production and consider the stability of oceanic general circulation from a global perspective because local processes of generation and dissipation of kinetic energy in a turbulent medium remain unknown.

The ocean system can be regarded as an open non-equilibrium system connected with surrounding systems mainly via heat and salt fluxes. The surrounding systems consist of the atmosphere, the Sun and space. Because of the curvature of the Earth's surface and the inclination of its rotation axis relative to the Sun, net gains of heat and salt are found in the equatorial region; net losses of heat and salt are apparent in polar regions. The heat and salt fluxes bring about an inhomogeneous distribution of temperature and salinity in the ocean system. This inhomogeneity produces the circulation, which in turn reduces the inhomogeneity. In this respect, the formation of the circulation can be regarded as a process leading to final equilibrium of the whole system: the ocean system and its surroundings. In this process, the rate of approach to equilibrium, i.e., the rate of entropy production by the oceanic circulation, is an important factor.

Related to the rate of entropy production in an open non-equilibrium system, Sawada (1981) reported that such a system tends to follow a path of evolution with a maximum rate of entropy production among manifold dynamically possible paths. This postulate has been called the principle of Maximum Entropy Production (MEP), which has been confirmed as valid for mean states of various nonlinear fluid systems, e.g., the global climate system of the Earth (Ozawa & Ohmura, 1997; Paltridge, 1975, 1978), those of other planets (Lorenz et al., 2001), the oceanic general circulation including both surface and abyssal circulations (Shimokawa, 2002; Shimokawa & Ozawa, 2001, 2002, 2007), and thermal convection and shear turbulence (Ozawa et al., 2001). Therefore, it would seem that MEP can stand for a

universal principle for time evolution of non-equilibrium systems (see reviews of Kleidon and Lorenz, 2005; Lorenz, 2003; Martyushev & Seleznev, 2006; Ozawa et al., 2003; Whitfield, 2005). However, although some attempts have been made to seek a theoretical framework of MEP (e.g., Dewar, 2003, 2005), we remain uncertain about its physical meaning.

1.7 Main contents of this chapter

As described above, the problem of whether the abyssal circulation is a heat engine or mechanical pump and how it is related to the Sandström theorem are important for better understanding of the oceanic general circulation. In the following sections, we discuss the problem referring to the results of numerical simulations of the oceanic general circulation. In section 2, a numerical model and method are described. In section 3, a calculation method of entropy production rate in the model is explained. In section 4, details of entropy production in the model are described. In section 5, by referring to the results, the problem of whether the abyssal circulation is a heat engine or mechanical pump and how it is related to the Sandström theorem is discussed.

2. Numerical model and method

The numerical model used for this study is the Geophysical Fluid Dynamics Laboratory's Modular Ocean Model (Pacanowski, 1996). The model equations consist of Navier-Stokes equations subject to the Boussinesq, hydrostatic, and rigid-lid approximations along with a nonlinear equation of state that couples two active variables, temperature and salinity, to the fluid velocity. A convective adjustment scheme is used to represent the vertical mixing process. Horizontal and vertical diffusivity coefficients are, respectively, $10^3 \text{ m}^2 \text{ s}^{-1}$ and $10^{-4} \text{ m}^2 \text{ s}^{-1}$. The time-step of the integration is 5400 s.

The model domain is a rectangular basin of 72° longitude by 140° latitude with a cyclic path, representing an idealized Atlantic Ocean (Fig. 3(a)). The southern hemisphere includes an Antarctic Circumpolar Current passage from 48°S to 68°S . The horizontal grid spacing is 4 degrees. The ocean depth is 4500 m with 12 vertical levels (Shimokawa & Ozawa, 2001). All boundary conditions for wind stress, temperature and salinity are arranged as symmetric about the equator (Figs. 3(b), 3(c), and 3(d)). The wind stress is assumed to be zonal (eastward or westward direction, Fig. 3(b)). A restoring boundary condition is applied: The surface temperature and salinity are relaxed to their prescribed values (Figs. 3(c) and 3(d)), with a relaxation time scale of 20 days over a mixed layer depth of 25 m. The corresponding fluxes of heat and salt are used to calculate F_h and F_s at the surface. The initial temperature distribution is described as a function of depth and latitude. The initial salinity is assumed to be constant (34.9‰). The initial velocity field is set to zero. Numerical simulation is conducted for a spin-up period of 5000 years.

Figure 4 shows a zonally integrated meridional stream function at years 100, 1000, 2000, 3000, 4000, and 5000, after starting the calculations. At year 100, the circulation pattern is almost symmetric about the equator. The sinking cell in the southern hemisphere does not develop further because of the existence of the Antarctic Circumpolar Current. In contrast, the sinking cell in the northern hemisphere develops into deeper layers, and the circulation pattern becomes asymmetric about the equator. The oceanic circulation becomes statistically steady after year 4000. Temperature variations are shown to be less than 0.1 K after year 4000. In the steady state, the northern deep-water sinking cell is accompanied by an Antarctic bottom-water sinking cell and by a northern intrusion cell from the south. The flow pattern is apparently a basic one in the idealised Atlantic Ocean.

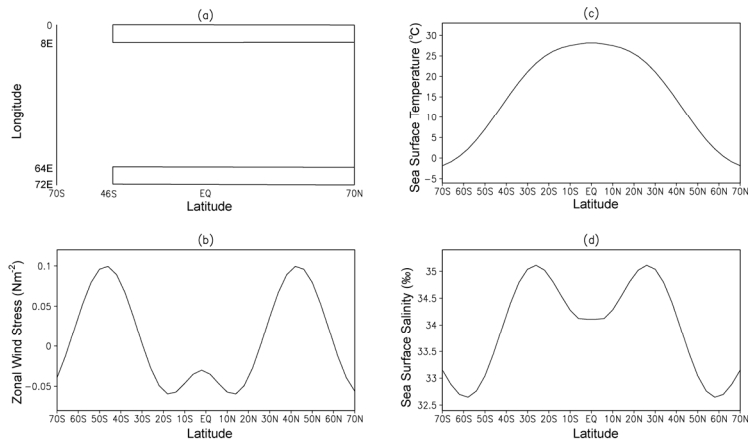


Fig. 3. (a) Model domain, and forcing fields of the model as functions of latitude, (b) forced zonal wind stress (N m^{-2}) defined as positive eastward, (c) prescribed sea surface temperature ($^{\circ}\text{C}$), and (d) prescribed sea surface salinity (‰).

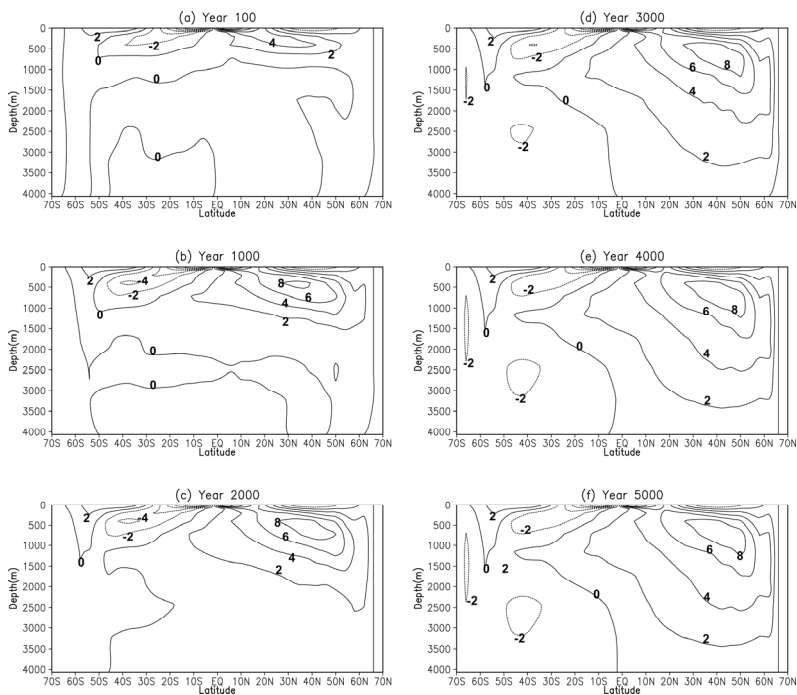


Fig. 4. The zonally integrated meridional stream function at years (a) 100, (b) 1000, (d) 2000, (e) 3000, (d) 4000, and (e) 5000 after starting the numerical calculations. The contour line interval is 2 SV ($10^6 \text{ m}^3 \text{ s}^{-1}$). The circulation pattern reached a statistically steady-state after year 4000.

3. Entropy production rate calculation

According to Shimokawa & Ozawa (2001) and Shimokawa (2002), the entropy increase rate for the ocean system is calculable as

$$\frac{dS}{dt} = \int \frac{1}{T} \left[\frac{\partial(\rho c T)}{\partial t} + \text{div}(\rho c T v) + p \text{div}(v) \right] dV + \int \frac{F_h}{T} dA - ak \int \left[\frac{\partial C}{\partial t} + \text{div}(C v) \right] \ln C dV - ak \int F_s \ln C dA \quad (4)$$

where ρ stands for the density, c denotes the specific heat at constant volume, T signifies the temperature, $a = 2$ is van't Hoff's factor representing the dissociation effect of salt into separate ions (Na^+ and Cl^-), k is the Boltzmann's constant, C is the number concentration of salt per unit volume of seawater, F_h and F_s are the heat and salt fluxes per unit surface area respectively, defined as positive outward, and dV and dA are the small volume and surface elements, respectively.

If we can assume that the seawater is incompressible ($\text{div } v = 0$) and that the volumetric heat capacity is constant ($\rho c = \text{const.}$), then the divergence terms in (4) disappear. In this case, we obtain

$$\frac{dS}{dt} = \int \frac{\rho c}{T} \frac{\partial T}{\partial t} dV + \int \frac{F_h}{T} dA - ak \int \frac{\partial C}{\partial t} \ln C dV - ak \int F_s \ln C dA \quad (5)$$

The first two terms in the right-hand side represent the entropy production rate attributable to heat transport in the ocean. The next two terms represent that attributable to the salt transport. The first and third terms vanish when the system is in a steady state because the temperature and the salinity are virtually constant ($\partial T / \partial t = \partial C / \partial t = 0$). In the steady state, entropy produced by the irreversible transports of heat and salt is discharged completely into the surrounding system through the boundary fluxes of heat and salt, as expressed by the second and fourth terms in equation (5).

The general expression (4) can be rewritten in a different form. A mathematical transformation (Shimokawa and Ozawa, 2001) can show that

$$\frac{dS}{dt} = \int F_h \cdot \text{grad} \left(\frac{1}{T} \right) dV + \int \frac{\Phi}{T} dV - ak \int \frac{F_s \cdot \text{grad}(C)}{C} dV, \quad (6)$$

where F_h and F_s respectively represent the flux densities of heat and salt (vector in three-dimensional space) and Φ is the dissipation function, representing the rate of dissipation of kinetic energy into heat by viscosity per unit volume of the fluid. The first term on the right-hand side is the entropy production rate by thermal dissipation (heat conduction). The second term is that by viscous dissipation; the third term is that by molecular diffusion of salt ions. Empirically, heat is known to flow from hot to cold via thermal conduction, and the dissipation function is always non-negative ($\Phi \geq 0$) because the kinetic energy is always dissipated into heat by viscosity. Molecular diffusion is also known to take place from high to low concentration (salinity). Therefore, the sum should also be positive. This is a consequence of the Second Law of Thermodynamics.

4. Results – details of entropy production in the model

We describe here the details of entropy production in the model from the final state of the spin-up experiment (Fig. 4(f)). Because entropy production due to the salt transport is negligible (Shimokawa and Ozawa, 2001), local entropy production can be estimated from the first term in equation (6) as

$$A = \frac{\rho C}{T^2} (A_x + A_y + A_z), A_x = D_h \left(\frac{dT}{dx} \right)^2, A_y = D_h \left(\frac{dT}{dy} \right)^2, A_z = D_v \left(\frac{dT}{dz} \right)^2, \quad (7)$$

where D_h denotes horizontal diffusivity of $10^3 \text{ m}^2 \text{ s}^{-1}$, D_v stands for vertical diffusivity of $10^{-4} \text{ m}^2 \text{ s}^{-1}$ (see section 2), and other notation is the same as that used earlier in the text. It is assumed here that $F_h = -k \text{ grad}(T) = -\rho c D_E \text{ grad}(T)$, where $k = \rho c D_E$ signifies thermal conductivity and where D_E represents the eddy diffusivity (D_h or D_v). Figure 5 shows zonal, depth and zonal-depth averages of each term in equation (7). The quantities not multiplied by dV represent the values at the site, and the quantities multiplied by dV represent the values including the effect of layer thickness.

It is apparent from the zonal average of A (Fig. 5(a)) that entropy production is large in shallow-intermediate layers at low latitudes. This is apparent also in the zonal-depth average of $A \times dV$ (Fig. 5(c)). However, it is apparent from the depth average of $A \times dV$ (Fig. 5(b)) that entropy production is large at the western boundaries at mid-latitudes and at low latitudes. Consequently, entropy production is greatest at the western boundaries at mid-latitudes as the depth average, but it is highest at low latitudes as the depth-zonal average. It is apparent as the figures show of A_x , A_y and A_z (Figs. 5(d), (g) and (j)) that A_x is large in shallow layers at mid-latitudes, A_y is large in shallow-intermediate layers at high latitudes, and that A_z is large in shallow-intermediate layers at low latitudes. It is also apparent that as the figures show of $A_x \times dV$, $A_y \times dV$ and $A_z \times dV$ (Figs. 5(e), 5(f), 5(h), 5(i), 5(k) and 5(l)) that $A_x \times dV$ is large at the western boundaries at mid-latitudes, $A_y \times dV$ is large at high latitudes, and $A_z \times dV$ is large at low latitudes. Additionally, it is apparent that the values of A_z ($A_z \times dV$) is the largest, and those of A_x ($A_x \times dV$) are smaller than those of A_y ($A_y \times dV$) and A_z ($A_z \times dV$).

Consequently, there are three regions with large entropy production: shallow-intermediate layers at low latitudes, shallow layers at the western boundaries at mid-latitudes, and shallow-intermediate layers at high latitudes. It can be assumed that the contribution of shallow-intermediate layers at low latitudes results from the equatorial current system. That of western boundaries at mid-latitudes results from the western boundary currents such as Kuroshio, and that of intermediate layers at high latitudes results from the meridional circulation of the global ocean. It is apparent that high dissipation regions at low latitudes expand into the intermediate layer in the zonal averages of $A \times dV$ and $A_z \times dV$. These features appear to indicate that equatorial undercurrents and intermediate currents in the equatorial current system are very deep and strong currents which can not be seen at other latitudes (Colling, 2001). It is also apparent that high dissipation regions at high latitudes in the northern hemisphere intrude into the intermediate layer in the zonal averages of $A \times dV$ and $A_y \times dV$, and the peak of northern hemisphere is larger than that of southern hemisphere in the zonal-depth averages of A and A_y . These features appear to represent the characteristics of the circulation with northern sinking (Fig. 4(f)).

Strictly speaking, we should consider dissipation in a mixed layer and dissipation by convective adjustment for entropy production in the model. Dissipation in a mixed layer can be estimated from the first term in (6) as

$$B = \frac{\rho C}{T^2} \frac{(T_r - T_s)}{\Delta t_r}, \tag{8}$$

where T_r signifies restoring temperature (Fig. 3(c)), T_s is the sea surface temperature in the model, and Δt_r stands for the relaxation time of 20 days (see section 2). It is assumed here that $F_h = -k \text{ grad}(T) = -\rho c D_M \text{ grad}(T)$, where $k = \rho c D_M$ is thermal conductivity, $D_M = \Delta z_r^2 / \Delta t_r$ represents diffusivity in the mixed layer, and Δz_r is the mixed layer thickness of 25 m (see section 2). The estimated value of B is lower than that of A by three or four orders: it is negligible. Dissipation by convective adjustment can be estimated from the first term in (5) such that

$$C = \frac{\rho C}{T_b} \frac{(T_b - T_a)}{\Delta t}, \tag{9}$$

where T_b is the temperature before convective adjustment, T_a is the temperature after convective adjustment, and Δt is the time step of 5400 s (see section 2). In fact, T_b is identical to T_a at the site where convective adjustment has not occurred. The value of C is negligible because the effect of convective adjustment is small in the steady state.

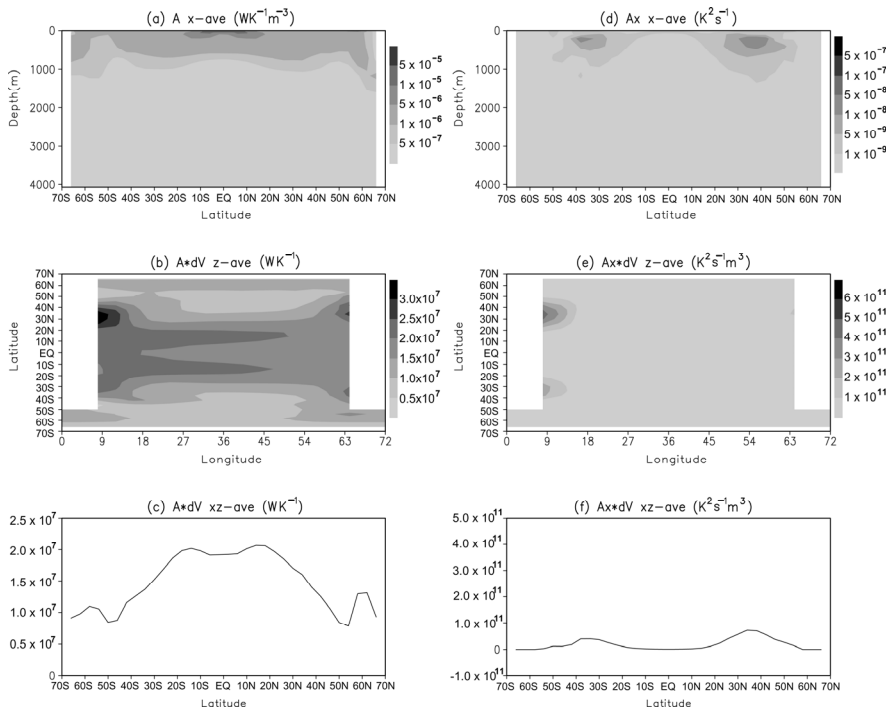


Fig. 5. Entropy production in the model.

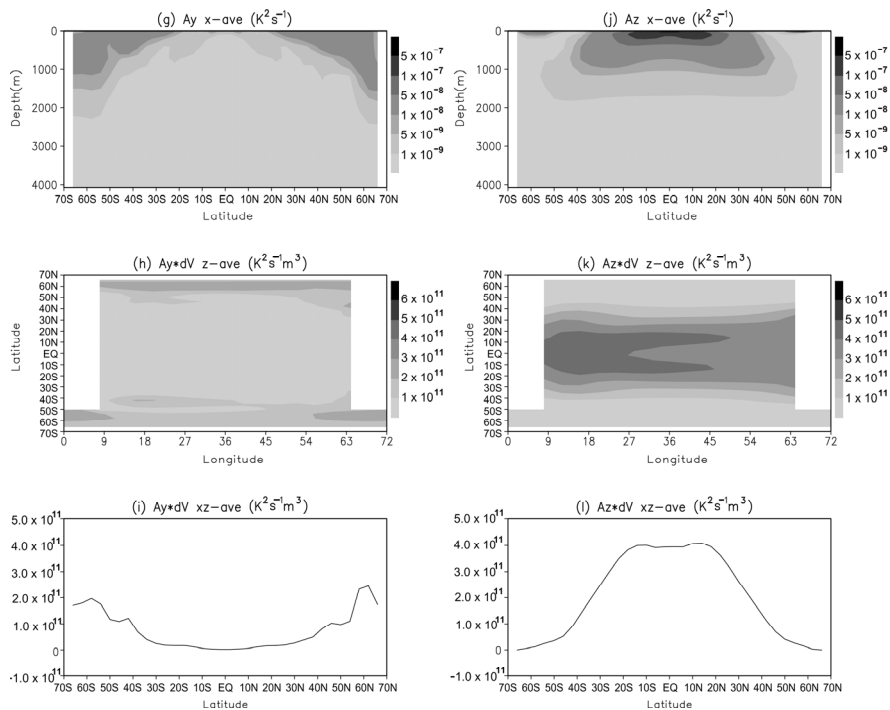


Fig. 5. (continued)

(a) zonal average of A , (b) depth average of $A \times dV$, (c) zonal-depth average of $A \times dV$, (d) zonal average of A_x , (e) depth average of $A_x \times dV$, (f) zonal-depth average of $A_x \times dV$, (g) zonal average of A_y , (h) depth average of $A_y \times dV$, (i) zonal-depth average of $A_y \times dV$, (j) zonal average of A_z , (k) depth average of $A_z \times dV$, (l) zonal-depth average of $A_z \times dV$. The unit for A is $W K^{-1} m^{-3}$. The unit for $A \times dV$ is $W K^{-1}$. The unit for A_x , A_y , and A_z is $K^2 s^{-1}$. The unit for $A_x \times dV$, $A_y \times dV$, and $A_z \times dV$ is $K^2 s^{-1} m^3$. The contour interval is indicated at the right side of each figure.

5. Discussion – Sandström theorem and abyssal circulation

As stated in section 1.5, Sandström suggested that a closed steady circulation can only be maintained in the ocean if the heating source is located at a higher pressure (i.e. a lower level) than that of the cooling source. Therefore, he suggested that the oceanic circulation is not a heat engine.

Huang (1999) showed using an idealized tube model and scaling analysis that when the heating source is at a level that is higher than the cooling source such as the real ocean, the circulation is mixing controlled, and in the contrary case, the circulation is friction-controlled. He also suggested that, within realistic parameter regimes, the circulation requires external sources of mechanical energy to support mixing to maintain basic stratification. Consequently, oceanic circulation is only a heat conveyer, not a heat engine. Yamagata (1996) reported that the oceanic circulation can be driven steadily as a heat engine only with great difficulty, considering the fact that the efficiency as a heat engine of the

oceanic circulation calculated heating and cooling sources at the sea surface is very low, in addition to a view of Sandström's theorem. He therefore concluded that the oceanic circulation might not be driven steadily as a heat engine, but that it shows closed circulation by transferral to mechanically driven (e.g. wind-driven) flow on the way: the oceanic circulation might be sustained with a mixture of the buoyancy process and mechanical process.

However, these arguments are based on the assumption that the heating source is located only at the sea surface. If a diabatic heating because of turbulent diffusion takes place in the ocean interior (and the cooling source is placed at the sea surface), then Sandström's theorem is not violated. The important quantity in this respect is diapycnal diffusion, as stated in section 1, which corresponds to A_z in our model. As stated in section 4, A_z in our model showed high entropy production attributable to turbulent diapycnal diffusion down to 1000 m in the whole equatorial region (<30 deg). By contrast, the diapycnal diffusion at high latitude is very small and is confined to the surface in Fig. 5(j). Although there also exists dissipation caused by convective adjustment in the polar region, it can be negligible as the regional average: the region of adiabatic heating at low latitudes extends into the deeper layer (i.e. a higher pressure), but the region of adiabatic cooling at high latitudes is confined to the surface (i.e. a lower pressure). These results support the inference described above. In addition, the real ocean is also affected by dynamic interaction among tides, topography, and the resultant diabatic heating, which has not been considered in our model.

Moreover, the inference is supported by some experimental studies that the circulation is possible if external heating and cooling are placed at the same level (Park & Whitehead, 1999), or even if external heating is placed at a higher level than external cooling (Coman et al. 2006). Coman et al. (2006) reported that heat diffusion (whether by molecular conduction or turbulent mixing) allows heat to enter and leave the fluid at the boundary and causes the heating to be distributed throughout at least the depth of the boundary layer. Warmed water ascends towards the surface after having warmed and expanded at higher pressures than the surface pressure. Positive work is available from the heating and cooling cycle, even when the heating source is above the cooling source. Therefore, they concluded that Sandström theorem cannot be used to discount the formation of a deep convective overturning in the oceans by the meridional gradient of surface temperature or buoyancy forcing suggested by Jeffreys (1925). In addition, the driving force of the circulation in these experiments is only internal diabatic heating by molecular conduction or turbulent diffusion: the real ocean includes stronger diabatic heating due to external forcing of wind and tide, as explained in sections 1.2 and 1.3. In the equatorial region, the flow structure consisting of equatorial undercurrents and intermediate currents is organized such that forced mixing by wind stress at the surface accelerates turbulent heat transfer into the deeper layer. However, in the polar regions, forced mixing by wind stress at the surface does not reach the deeper layer, and adiabatic cooling is confined to the surface. For that reason, seawater expands at the high-pressure intermediate layer in the equatorial region because of heating and contracts at the low-pressure surface in the polar regions because of cooling. Consequently, mechanical work outside (i.e. kinetic energy) is generated and the circulation is maintained. The above inference will be strengthened in consideration of the real ocean.

Using numerical simulations, Hughes & Griffiths (2006) showed that by including effects of turbulent entrainment into sinking regions, the model convective flow requires much less energy than Munk's prediction. Results obtained using their model indicate that the ocean

overturning is feasibly a convective one. Therefore, they suggested that there might be no need to search for “missing mixing.” As stated in section 1.4, the idea of the ocean as “mechanical pump” was the idea derived to solve the “missing mixing” problem: the “mechanical pump” was introduced as another new mechanism of diapycnal mixing to maintain abyssal circulation. If their conclusion is correct in the real ocean, then the assumption of a “mechanical pump” (i.e. “missing mixing”) is not necessary. Small “background” diapycnal mixing might be sufficient to maintain abyssal circulation.

It is possible that the idea of the ocean as a “heat engine” is not fully contradicted by the idea of the ocean as a “mechanical pump”: it can be considered that a circulation driven as a “heat engine” is strengthened by a pump-up flow driven as a “mechanical pump”. In a sense, the idea of a mixture of buoyancy processes and mechanical processes by Yamagata (1996) might be right on target.

As stated in section 1.3, although recent observations of turbulence show large diapycnal mixing, such observations are limited to a few locations. It is not clear how much is the value of diapycnal mixing averaged in the entire ocean. Although global mapping of diapycnal diffusivity based on expendable current profiler surveys has been tried (Hibiya et al., 2006), the observed places remain limited. To verify the thermodynamic structure of the oceanic general circulation suggested in this chapter, the entire structure of adiabatic heating and cooling should be resolved. Particularly, observations of the following are recommended: 1) the structure of turbulent heat transfer into the intermediate layer because of forced mixing by wind stress at the surface and the resultant adiabatic heating in the equatorial region, 2) the process of adiabatic cooling confined to the surface and the subsequent concentrated sinking in the polar regions. In addition, direct observations of sinking and upwelling, not inferred from other observations, are important because the inferred value might include the effects of assumptions and errors. The observation of sinking is difficult because of severe climates in polar winter, with the worst conditions occurring when the sinking occurs. Moreover, observation of the upwelling itself is extremely difficult because of the low velocity. Future challenges must include technical improvements of observational instruments.

6. Conclusion

This chapter presented discussion of the problem of whether the abyssal circulation is a heat engine or a mechanical pump. We also discussed how it is related to the Sandström theorem, referring to results of numerical simulations of the oceanic general circulation. The results obtained using our model show high-entropy production due to turbulent diapycnal diffusion down to 1000 m in the entire equatorial region (<30 deg). By contrast, diapycnal diffusion at high latitude is very small and is confined to the surface: the region of adiabatic heating at low latitudes extends into the deeper layer (i.e. a higher pressure), but the region of adiabatic cooling at high latitudes is confined to the surface (i.e. lower pressure). In this case, Sandström’s theorem is not violated. In the equatorial region, the flow structure consisting of equatorial undercurrents and intermediate currents is organized such that forced mixing by wind stress at the surface accelerates turbulent heat transfer into the deeper layer. However, in polar regions, forced mixing by wind stress at the surface does not reach the deeper layer, and adiabatic cooling is confined to the surface. Consequently, seawater expands at a high-pressure intermediate layer in the equatorial region because of

heating and contracts at a low-pressure surface in polar regions because of cooling. Therefore, mechanical work outside (i.e. kinetic energy) is generated and the circulation is maintained. The results suggest that abyssal circulation can be regarded as a heat engine, which does not contradict Sandström's theorem.

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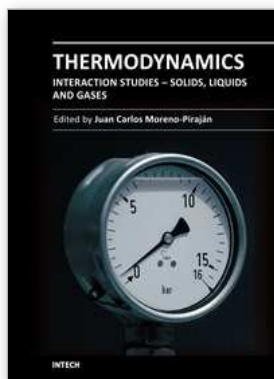
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