1. The unique nature of ACC.

ACC is the only current that flows around the whole world without being interrupted by continents. (Arctic Ocean may be another exception). Our study of wind-driven circulation has been focused on gyre circulation, primarily in the northern hemisphere. Although we have seen great progresses along the line of understanding the structure of the global oceans, many important aspects of this global picture remain unclear. In particular, our understanding of ACC remains incomplete and thus dynamics of ACC is one of the most challenges, and it is a very important and exiting research frontier.

ACC is closely linked with AABW that flows over the bottom of the world oceans. The coldest water at the bottom of the world oceans is formed along the edge of Antarctic Continent. After its formation, it sinks along the continental slope and spreads into the bottom of the world oceans, as shown in Fig. 1.

Fig. 1. Bottom water temperature in the world oceans.

A. Why the AABW can sink to the bottom of the world oceans?

Deep or bottom water is formed mostly in marginal seas and moves to the open ocean in forms of outflow. The major outflows have the essential components in common (Price and Baringer, 1994):

1. Air-sea exchange that produces dense water due to a combination of heat and fresh water flux from the ocean to atmosphere (or saline ejection).
2. Ocean-sea exchange that allows dense water formed in semi-closed marginal seas to flow into the open ocean.
3. Descent and entertainment that modify the properties of the outflow water. In general, the volume of the overflow water doubles. During the process, cabbeling can make the deep water even denser.

For the world oceans there are four major sources of overflow:

- Mediterranean
- Denmark Strait
Properties of these outflows are listed in Table 1. Note that although density of the Mediterranean outflow is the highest among these four sources, the final product from the Mediterranean outflow is the lightest! There are several reasons why the density of outflow reorders: density difference between the outflow and the environment, topographic slope, thermobaric effect*. The outflow can be simulated using simple one-dimensional streamtube model (Price and Baringer, 1994).

*Thermobaric effect: The equation of state is highly nonlinear, so the compressibility of sea water strongly depends on the temperature: cold sea water is more compressible than warm sea water. The thermobaric effect can be seen clearly from the next table.

<table>
<thead>
<tr>
<th>Density at Depth (db)</th>
<th>Mediterranean</th>
<th>Denmark Strait</th>
<th>Feroe bank Channel</th>
<th>Flichner Ice Shelf</th>
</tr>
</thead>
<tbody>
<tr>
<td>T</td>
<td>13.4</td>
<td>0.0</td>
<td>-0.5</td>
<td>-2.1</td>
</tr>
<tr>
<td>S</td>
<td>37.80</td>
<td>34.9</td>
<td>34.92</td>
<td>34.67</td>
</tr>
<tr>
<td>0</td>
<td>28.48</td>
<td>28.03</td>
<td>28.07</td>
<td>27.92</td>
</tr>
<tr>
<td>1000</td>
<td>32.85</td>
<td>32.74</td>
<td>32.79</td>
<td>32.69</td>
</tr>
<tr>
<td>2000</td>
<td>37.12</td>
<td>37.34</td>
<td>37.41</td>
<td>37.37</td>
</tr>
<tr>
<td>3000</td>
<td>41.30</td>
<td>41.84</td>
<td>41.92</td>
<td>41.93</td>
</tr>
<tr>
<td>4000</td>
<td>45.38</td>
<td>46.23</td>
<td>46.33</td>
<td>46.38</td>
</tr>
<tr>
<td>4000*</td>
<td>44.91</td>
<td>45.88</td>
<td>46.07</td>
<td>46.20</td>
</tr>
<tr>
<td>( \Delta \sigma_4 )</td>
<td>0.47</td>
<td>0.35</td>
<td>0.26</td>
<td>0.18</td>
</tr>
</tbody>
</table>

Table 1. Density variation with the imaginary sinking depth for four sites of deepwater formation. The line marked by * indicates the potential density for the final product after mixing, calculated from a stream-tube model by Price and Baringer (1994). \( \Delta \sigma_4 \) is the density decline due to mixing.

It is clear that water mass density reordering is due to both the T-S properties and mixing. In fact, due to the difference in compressibility Mediterranean outflow has smaller density at large depth. Thus, although the Mediterranean outflow is densest at the time for formation on the surface, the order of density reverses around 1000-1300 meters, even without mixing! Most importantly, mixing is another factor determining the properties of deep water. In fact, mixing for the Mediterranean outflow is much stronger than other three outflows. This is one of the major factors that control the depth where the spreading of the Mediterranean outflow is observed in the Atlantic under the current climate.

An interesting question is whether the Mediterranean outflow could become the source water for the bottom circulation in the world oceans. Assume that deepwater sources from other three sites remain unchanged, an increase of salinity \( \Delta S = 1.83 \) or a decline in temperature \( \Delta T = -4.82^\circ C \) can produce water mass that is densest among all four sites of deepwater formation, so that Mediterranean outflow would flow to the bottom of the world oceans.

<table>
<thead>
<tr>
<th>Case A: ( \delta T = 0, \delta S = 1.83 )</th>
<th>Depth (m)</th>
<th>Denmark Strait</th>
<th>Feroe Bank</th>
<th>Flichner Ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>5000</td>
<td>-0.21</td>
<td>-0.10</td>
<td>-0.0002</td>
<td></td>
</tr>
<tr>
<td>Case B: ( \delta T = 0, \delta S = 1.83 )</td>
<td>0</td>
<td>-1.36</td>
<td>-1.32</td>
<td>-1.46</td>
</tr>
<tr>
<td>Case B: ( \delta T = 0, \delta S = 1.83 )</td>
<td>0</td>
<td>-1.36</td>
<td>-1.32</td>
<td>-1.46</td>
</tr>
</tbody>
</table>

Table 2. Density difference between the Mediterranean outflow and other sites of deepwater formation, assuming changes in the deepwater properties of the Mediterranean outflow only.

| \( \delta T \) | \( \delta S \) | 5000 | -0.21 | -0.10 | -0.002 |

This calculation does not include the effect of mixing. Were the effect of mixing taken into the consideration, the change in either the temperature or salinity required to make the Mediterranean outflow water dense enough to sink to the bottom would be even larger. During the geological past, evaporation rate over the Mediterranean Sea might be much higher than the present day, so such a possibility might be realized under such circumstance.

The discussion above does not include contribution due to cabbeling. In the Southern Ocean the coldest water is formed primarily along the edge of the Antarctic continent, especially in the Weddell Sea and Ross Sea. Cold and relatively fresh water formed in winter mix with relatively warm and more salty water below. Due to the nonlinearity of the equation of state of sea water, density of the mixed water is larger than the density of the original types of water; thus, the newly formed water can sink to even greater depth. This is called cabbeling (Foster and Carmack, 1975). In fact, cold water sinking in the North Atlantic can reach to the depth of roughly 4 kilometers only, and the bottom of the world oceans is occupied by the bottom water formed along Antarctic. The competition of these two types of water is shown in Figs. 2 and 3.

Fig. 2. Temperature distribution along 30°W (in the Atlantic Ocean).
Fig. 3. Salinity distribution along 30°W (in the Atlantic Ocean).

**B. Bipolar modes for the global oceans?**

Although there are two major sources of deep/bottom water in the modern oceans, the situation may change depending on the climate conditions. There are much Paleoclimate data that suggest the shut down of the NADW formation. A most interesting case is the so-called Younger Dryas period, which happened about 12,000 years ago. Although the exact nature of this phenomenon remains a topic of hot debate, it is believed that large amount of ice melting water during the end of last glaciations caused the newly re-established meridional overturning in the North Atlantic to collapse. The shut down of the NADW formation gave rise to a cold period over much of the N. Atlantic basin, as indicated by the grey scale in Fig. 4.

However, there are indicates that the strength of the AABW might be stronger than normal, indicated by the relatively high Δ¹³C in the lower part of Fig. 4. Thus, there might be some kind of compensating the decline of deepwater formation in the North Atlantic. As postulated by W. Broecker (1998), there might be a bipolar mode in the world oceans. The details of such a global model remain fuzzy.
Fig. 4. The contrast between the gray scale (low gray scale, inferred from sediments, indicates low temperature) and carbon content of the surface ocean, which indicates a relatively warm condition during the Younger Dryas period.

C. Velocity structure of the ACC

ACC is associated with a huge amount of mass transport. The best estimate is about 134 Sv (Rintoul et al., 1998); the current system has a fairly barotropic structure, i.e., the zonal velocity is eastward all the way from sea surface to sea floor (Fig. 5). ACC is associated with strong surface fronts and fast currents. At some places the surface velocity can be as large as 30-50 cm/s. In addition, this current system continues around the whole latitudinal band, Fig. 5. In fact, it is the only current system which circulates around the whole latitudinal circle without interruption in the modern geometry of land-sea distribution.

Fig. 5. Zonal velocity distribution along 170°W, in cm/s, diagnosed from a low-resolution model for the world oceans with 2° × 2° horizontal resolution and 58 layers vertically. The high speed core on the right edge is the Equatorial undercurrent.
Fig. 6. Zonal velocity distribution along 48°S, in cm/s, diagnosed from a low-resolution model for the world oceans with 2° × 2° horizontal resolution and 58 layers vertically.

D. The strength of the meridional overturning cell associated with the AABW

Circulation in the ocean is a complicated 3-D phenomenon; thus, any 2-D mapping of the circulation is always limited. The following figures provide good examples. In these figures, the steady component of the overturning circulation in the world oceans is mapped out in three different vertical coordinates, i.e., the z-coordinates, the $\sigma_s$ coordinates, and the $\theta$ (potential temperature) coordinates, Fig. 6. The reason to choose the $\sigma_s$ instead of $\sigma_0$ is that $\sigma_s$ is a better choice for mapping the deep circulation.

It is readily seen that overturning streamfunction maps plotted in different vertical coordinates show quite different features. These differences reflect the fact that these maps are two dimensional view of the three dimensional structure of the circulation. If you look at the same person from different angle, you will see different feature, of course. In fact, the overturning streamfunction plotted on other coordinates can provide different, or may be better, information of the three-dimensional structure of the circulation in the southern ocean.

A major difference between overturning cells in z-coordinate and sigma or potential temperature coordinates is the existence of the so-called Deacon Cell located between 40-60°S, which is the most outstanding feature in the overturning map in z-coordinate, upper panel in Fig. 6. This difference will be explained shortly.
Fig. 6. Meridional overturning cells in different coordinates, obtained from a model with horizontal resolution of $2^\circ \times 2^\circ$ and 58 levels: a) $Z$; b) $\sigma_4$; and c) Potential temperature, in unit of Sv.

Fig. 7. High resolution maps of meridional overturning cells in different coordinates, obtained from a model with horizontal resolution of $2^\circ \times 2^\circ$ and 58 levels: a) $Z$; b) $\sigma_4$; and c) Potential temperature, in unit of Sv.
From these figures, the deep overturning cell associated with the AABW is about 30-50 Sv, depending on the vertical coordinates used in the calculation. This is consistent with the classical description by W. Schmitz (1995). As shown in Fig. 8, the total amount of bottom water circulation is about 48Sv, including 7 Sv into the Atlantic, 17Sv into the Pacific, and 24Sv into the Indian. Since deep mixing requires external mechanical energy (such as tides and wind stress), it is speculate that mechanical energy supporting mixing in the deep part of the Indian Ocean should be about 3 times of that in the Atlantic in order to maintain the stronger diapycnal mixing in the Indian Ocean.

Fig. 8. Water mass transports in the world oceans, the blue lines for bottom water, green lines for deep, and red for upper ocean (W. Schmitz, 1995).

2. Basic physical processes in the ACC

   A. Can we use the Sverdrup theory?

   The essences of the Sverdrup theory include:

   1) In the gyre interior, the ocean is separated into the upper ocean where wind-driven motions dominate and the abyss which is motionless. The so-called Sverdrup relation applies to the wind-driven circulation in the upper ocean only.

   2) In the western boundary region, the western boundary currents transport water meridionally. Note that the western boundary currents play important roles in balancing mass, vorticity, and energy.

   In a word, the wind-driven circulation theory, including its Sverdrup interior and the associated frictional western boundary currents, is a theory of linear vorticity balance. (For the case of inertial western boundary layer, the theory works for the equator half of the western boundary only. As discussed in previous lecture, a pure inertial western boundary layer cannot provide a complete balance of energy and vorticity.) The establishment of Sverdrup flow requires the existence of the meridional pressure gradient. First, since the lower layer is motionless the momentum input due to wind stress must be balanced by pressure difference in the upper layer.
Second, the meridional motions in the subsurface layers are geostrophic flows which also require a zonal pressure gradient for the momentum balance. Such zonal pressure difference can be built up only in the presence of meridional boundary in the basin. Lack of meridional barriers around ACC makes the dynamic balance of ACC quite different from gyre circulation discussed in the classical theory of wind-driven gyre in subtropical basins.

The essential point is to find the right mechanism to balance the eastward momentum input due to the wind stress. Started from Munk and Palmen (1951), it was realized that both lateral friction and bottom friction from an ocean without bottom topography are insufficient for balancing the wind stress input. Thus, the only possible candidate for balancing the momentum is the form drag.

Form drag is a very useful dynamic quantity which can be used to interpret the momentum balance for flow over topography. It is well known that pressure at the downstream side of topography is higher than the upstream side; thus, the along-stream depth-pressure integration gives rise to the so-called form drag. Note that form drag is due primarily to pressure difference associated with lee waves and other large-scale motions, not the bottom friction. The disadvantage of the form drag is that it is a diagnostic quantity that can be calculated only when the flow field is known. In the following sections we discuss some conceptual models which can be used to highlight the wind-driven circulation component of ACC.

**B. Stommel’s model for the ACC**

A simple conceptual model for the ACC was formulated by Stommel (1957). In this model the Southern Ocean is approximated by a flat bottom, with idealized coastline, shown in Fig. 9. Applying the Sverdrup relation to section AB gives rise to a simple relation between the volume flux of the ACC, $T$, and the line integral of wind-stress curl

$$\beta T = \int_A^B \nabla \times \tilde{\tau}_w \, dx$$

(1)

Fig. 9. Sketch to illustrate Stommel's model for the ACC.

In the Stommel model the circulation is in the classical Sverdrup balance. There is an eastern boundary current north of B, but there is no western boundary current south of B. The volume flux predicted by the above formulas must go through the gap BA and complete the loop of motion.

**C. A simple homogeneous model to introduce the concept of form drag**

The concept of form drag has been used to diagnose the momentum balance in OGCM simulation of ACC (Gille, 1995). Using the results from the Semtner-Chervin model simulation,
Gille showed that the dominating terms in momentum balance of ACC is indeed between the wind stress input and the form drag. In fact, form drag in ACC is primarily generated near three ridges. The basic concept of form drag can be illustrated clearly through the following simple model of a one-layer homogeneous fluid for the wind-driven circulation in a \(\beta\)-plan channel. The linear potential vorticity equation can be used for the wind-driven circulation in ACC (Wang and Huang, 1995; Krupitsky and Cane, 1995). First, we discuss a simple case on an f-plan with constant wind stress \(\tau\). The potential vorticity equation is

\[
J\left(\psi, \frac{f_0}{H} h\right) = \kappa \nabla^2 \psi
\]

with the boundary conditions:

\[
\psi \big|_{y=0} = 0; \quad \psi \big|_{y=D} = \psi_0
\]

where the total streamfunction \(\psi_0\) is determined through the averaging of the zonal momentum equation:

\[
\rho_0 f_0 h \frac{\partial \psi}{\partial x} - \kappa \rho_0 \frac{H}{D} \psi_0 + \nabla^2 \psi = 0
\]

where \(\bar{Z}^{xy}\) is the area mean

\[
\bar{Z}^{xy} = \frac{1}{LD} \int_0^L \int_0^D Z dx dy
\]

The first term on the left-hand side of (4) is the form drag. In the inviscid limit, \(\kappa \rightarrow 0\), the bottom friction term (the second term) vanishes, so wind stress input must be balanced by the form drag.

Consider a simple topography in the form of a ridge with height \(h\). For this f-plan channel, the potential vorticity is

\[
q = \frac{f_0}{H} h + f_0
\]

where the first term is the topographic vortex stretching term. Potential vorticity in different zones has different values. As fluid particles move from one zone to other zones, potential vorticity has to be changed. In the inviscid limit, there is no vorticity source in the interior, and the only places where vorticity can be altered are the thin frictional boundary layers, where friction can be very strong. The topographic form drag in this case is in a close form

\[
\tau_{\text{form}} = -\rho_0 \int \int f_0 \psi \frac{\partial h}{\partial x} dx dy = \rho_0 H \psi_0 \Delta q D
\]

where \(\Delta q = -f_0 h_0 / H\) is the potential vorticity barrier imposed by the ridge. Using (4), we obtain the zonal transport in the inviscid limit

\[
T'_r = \frac{\tau_0 HL}{\rho_0 |f_0| h_0}
\]

This approach can be generalized to the case of a \(\beta\)-plan channel with non-constant wind stress. The major weakness of this approach is the assumption of homogeneous fluid. Although this can be extended to multiple layer models, it is not clear how to modify this approach to continuously stratified fluid.

**D. Interfacial form stress in a continuously stratified model**
For a continuously stratified model we have to find out a mechanism that transfers horizontal momentum from wind stress applied on the sea surface to sea floor. This can be analyzed by examining the interface form stress (IFS) between isopycnal layers as follows. Assume there are two layer interfaces, \(-d_1(x)\) and \(-d_2(x)\), in a periodic channel. Note that

\[
0 = \oint dx \partial \left( \frac{\partial d_1}{\partial x} \right) - \oint p \partial \left( \frac{\partial d_1}{\partial x} \right) + \oint p \partial \left( \frac{\partial d_2}{\partial x} \right)
\]

Thus, the pressure gradient force integrated over the volume along the whole channel is

\[
-\oint \partial \left( \frac{\partial p}{\partial x} \right) dz = -\oint \left\{ p \left[ x, z = -d_1(x) \right] \frac{\partial d_1}{\partial x} - p \left[ x, z = -d_2(x) \right] \frac{\partial d_2}{\partial x} \right\}
\]

\[
= \oint \left\{ d_1 \frac{\partial}{\partial x} \right\} \left[ p \left[ x, z = -d_1(x) \right] - d_2 \frac{\partial}{\partial x} \right\] dx
\]

Therefore, the interfacial form stress \(\overline{dp_x}\) (the over bar indicates average in space and time) is produced by the out-of-phase arrangement of layer thickness and pressure gradient. In order to transfer momentum from the wind stress applied to the upper surface to the sea floor, a similar out-of-phase arrangement of isopycnal layer thickness and pressure gradient must exist over the water column. A sketch shown Fig. 10 illustrates the suitable arrangement of pressure and layer depth: \(p_x\) is positive over places where layer is deep, but it is negative over places where layer is shallow. This arrangement gives rise to a positive contribution to \(\overline{dp_x}\).

It is readily seen that the mean layer thickness does not contribute, so the IFS can be rewritten as \(d^* p_x^*\), and the contribution to an infinitesimal layer is the vertical divergence of

\[
IFS = -d^* p_x^*
\]

Note that * indicates contribution due to time-independent standing eddies and time-varying eddies.

---

**Fig. 10. Sketch of relationship between the layer thickness and pressure which can produce a non-zero interfacial form drag (Rintoul et al., 2001).**

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**E. Connection between interface form stress and poleward heat flux**

The close connection between interfacial form stress and poleward heat flux can be inferred as following. The meridional geostrophic velocity is \(f v_g^* = p_x^*\); layer thickness perturbation is linked
to potential temperature anomaly, 

\[ \frac{d^r}{\partial z} = -\frac{\theta^*}{\partial z} \]  

(Note eddies are not exactly geostrophic; however, they are nearly geostrophic, so the following scaling analysis can provide a good relation.) Therefore, IFS is linked to the meridional eddy heat flux,

\[ -d^r p_s^* = v^* \theta^* \cdot f / \partial z \]  

(11)

As a result, the southward eddy heat flux implies an eastward IFS, and the decline of eddy heat flux in the vertical direction suggests a gradually decline of IFS with increase of depth (Fig. 11).

Due to this relation between poleward heat flux and eddy activity, one may expect that stronger eddy activity may lead to larger poleward heat flux. This argument will be used to the recent warming of subsurface water in ACC shortly.

![Fig. 11. Cross-stream eddy heat flux observed in AUSSAF experiments (Phillips and Rintoul, 2000).](image)

**F. Bottom form stress**

Similar to the interfacial form stress, bottom stress can be derived from the same formula (10), except that the bottom is now treated as an interface; thus, bottom form drag is 

\[ \frac{-hp_{b,s}}{h_s} = \frac{h_s}{p_b} \]

which acts to transfer momentum from the fluid to the sea floor. Similar to the IFS, bottom form drag requires an out-of-phase arrangement of the bottom topography and bottom pressure. As shown in Fig. 10, there must high pressure at the rising site of the topography, but low pressure at the falling site of the bottom topography.

Although one may expect that surface wind stress is gradually compensated by IFS downward, diagnosis of numerical models for the Southern Ocean reveals that the basic momentum balance of the Southern Ocean takes a quite unexpected form, Fig. 12.
Fig. 12. Momentum balance of in the Southern Ocean. A) Balance of wind stress (solid line) and bottom form drag; b) balance of baroclinic form stress (line 1), barotropic form stress (line 2) and wind stress (line 3) (Stevens & Ivchenko, 1997).

As seen from this figure, wind stress is indeed balanced primarily by bottom form stress Fig. 12a; however, the baroclinic form stress has a sign opposite to that required for direct compensating wind stress. In fact, the barotropic form stress (related to the sea surface elevation) and baroclinic form stress are one order of magnitude larger than that of wind stress, and their sum gives a small residual which balances the wind stress. The balance of zonal momentum of ACC is illustrated in Fig. 13. The dynamics leading to such a balance is closely related with resonance of Rossby waves associated with bottom topography, the readers are referred by the review by Olbers at al. (2004).

Fig. 13. Sketch of the balance of zonal momentum of ACC: There is high barotropic pressure and low baroclinic pressure upstream of the ocean ridge, and low barotropic pressure and high baroclinic pressure downstream of the ridge. There is a nearly compensation of barotropic and baroclinic pressure, with a small residual which balances the wind stress input. Surface wind stress drives a northwards flux within the surface Ekman layer, and there is southward flux in the valleys upstream and downstream of the ridge (Olbers et al., 2004).

G. The “Drake Passage Effect” and the Deacon Cell

The latitude band of Drake Passage is a unique topographic feature in the world oceans. Except the Arctic Ocean, this latitude band is the only place in the modern world where the ocean occupies the whole latitudinal circle non-interruptedly. The shallowest depth of this channel is
about 2500m. Due to the lack of meridional boundary at this latitudinal band, there is no net zonal pressure gradient force; thus, any meridional currents above this sill depth must be ageostrophic in nature.

At 50°S, wind stress in the southern westerly reaches the maximum, and there is about 30Sv volume flux in the Ekman layer, which moves northward under the wind stress. A strong upwelling which pulls water from subsurface is associated with this Ekman flux. In order to conserve mass, there must be the same amount of water flowing southward at depth. A first choice of such southward flow of subsurface is the meridional geostrophic flow below the sill depth (2500m). Such a deep flow must have relatively low temperature; thus, it is unlikely from place in middle latitude Southern Hemisphere. A more logical choice would be the NADW. Thus, the Drake Passage Effect implies the connection between the Ekman transport in ACC and the deep overturning cell associated with NADW.

However, such a connection is not the only possible connection. First, the northward volume flux in the upper ocean at 40°S is much smaller than this value of 30Sv (Schmitz, 1995, 1996), indicating that substantial part of the returning flow may take place at shallow depths above topography. In fact, if the streamfunction is plotted in the z-coordinate, there is an outstanding overturning cell, the so-called Deacon Cell, Figs. 6a and 14.

In principle, the meridional circulation in the world ocean can be presented in different coordinates. A traditional way is to show the meridional streamfunction in the geopotential height coordinates (the z-coordinates). When the circulation associated with the Southern Ocean, there appears a large meridional cell in the upper three km, with upwelling south of 45°S and downwelling north of 45°S and a total volume flux on the order to 25-30 Sv. This is called the Deacon Cell, left panels in Fig. 14. However, there is very little density change associated with this overturning cell. Thus, if the meridional overturning streamfunction is presented in the density coordinate (or the potential coordinate), there is no such overturning cell, right panels Fig. 14.

The reason that the Deacon Cell is so weak, or nearly does not exist in the density coordinates is due to the following dynamics: Flows on different isopycnal surfaces can nearly cancels each other and thus leaves very little net contribution to the meridional transport, Fig. 15. The return flow is primarily in forms of eddies, which motions are nearly adiabatic. Therefore, eddy activity plays a critically important role in the dynamics of ACC.

Recent numerical experiments in GFDL (Hallberg and Gnanadesikan, 2006) indicate the so-called eddy saturation phenomenon. This is to say that in model with low resolution, meridional overturning cell is enhanced in response to increase in wind stress. However, for model with high resolution (1/6°) further increase in wind stress may bring the system to a state of eddy saturation, i.e., increase of wind stress can only enhance the eddy activity, without much change in the overturning circulation. Thus, results from model with low resolution must be interpreted with caution.
Fig. 14. Sketch of the Deacon Cell identified from a numerical model. Meridional circulation maps produced on (left panels) geopotential height coordinates and (right panels) potential coordinates.

Fig. 15. Sketch of the circulation on two isopycnal surfaces, demonstrating why the Deacon Cell does not appear in the isopycnal coordinates.

3. **What really controls the structure of ACC?**

   Although the Sverdrup theory gives a concise constraint over the volume flux in the wind-driven gyre interior, it does not apply to the recirculation. The zonal volume transport of ACC is a similar case, where no simple theory can give a clear connection between the volume transport and the potential dynamic forcings. To summarize, the main indexes of the volume transport of ACC include:
a) The wind stress  
b) The wind stress curl  
c) The meso-scale eddies  
d) The thermohaline forcing  
e) The topography  

Wind stress seems to be the primarily controlled of the zonal transport in ACC. Thus, one may expect that as wind stress associated with the southern westerly is enhanced, the zonal volume flux of ACC should be increased. However, this may not the case.  

In fact, the southern westerly has been enhanced substantially over the past several decades. One possible dynamic reason for such enhancement is due to the ozone depletion which is much more severe in the southern pole, although the exact reason of this enhancement remains unclear. Increase of wind stress in the Southern Ocean implies that more mechanical energy is input into the ocean, and hence a possibility of stronger circulation, especially the ACC. However, a low-resolution model (1° × 1° ) HIM model, forced by the NCEP wind stress over the past 50 years gave no hint of enhancement of the mean velocity, Fig. 16.  

This is not inconsistent with the in-situ observations. In fact, analysis of data collected through ISOS (International Southern Ocean Study, 1976-1982) and WOCE (1991-2001) gave no indication of significant trends in the zonal transport of ACC (Olbers et al., 2004). The measurements taken during 1990s are shown in Fig. 17.  

Similarly, numerical results from ECCO (a data-simulation model, Wunsch, personal communication, 2005) and an eddy-permitted model running in GFDL also gave similar results (Halberg and Gnanadesikan, personal communication, 2005), i.e., stronger westerly wind does not drive a stronger zonal volume transport in ACC; instead, under stronger wind force, the eddy activity is enhanced.

Fig. 16. Time evolution of the normalized wind-energy input through surface currents, surface waves, and Ekman layer and the variability of the mean horizontal velocity at ACC (black dotted line), (Huang et al., 2006).
The enhancement of eddy activity in ACC implies a stronger poleward heat flux in the Southern Ocean, and this is not inconsistent with the apparent warming of subsurface water temperature in the Southern Ocean. In fact, over the past 50 years, water temperature at the depth of 700-1100 m in the Southern Ocean has warmed up 0.17°C, and this speed is doubled of the global rate of temperature increase over the same period (Gille, 2002). Apparently, stronger wind stress puts more mechanical energy into the upper ocean, there the baroclinic instability transfer most of this energy into the meso-scale eddies, without much enhancement of the zonal volume flux. As discussed above, stronger eddy activity means more poleward heat flux, and a much stronger warming trend at high southern latitude ocean.
4. The pathway of meso-scale energy

A. Partition of energy between barotropic and baroclinic modes

Energy partition among eddy motions in the barotropic and baroclinic modes is roughly in the ratio of: 1: 1: 1/2: 1/4: 1/8…(Wunsch, 1997). Since the existence of shallow thermocline in the ocean, the kinetic energy of the first baroclinic mode is surface trapped; thus, signals diagnosed from satellite altimetry data are primarily associated with the first baroclinic mode.

As discussed in Lecture 5, there is a huge amount of mechanical energy from the mean state GPE associated with the geostrophic current converted through baroclinic instability to the GPE and kinetic energy of the meso-scale baroclinic eddies, especially in the Southern Oceans. These eddies have the horizontal scale on the order of first deformation radius, which is on the order of 10-20km for the high-latitude ocean. The corresponding wave length $2\pi L_{\delta_j}$ is about 100-150km. At low latitudes the deformation radius is larger. Baroclinic eddies are three-dimensional in nature; thus, their energy and enstrophy tend to cascade down scale.

B. Energy kinetic energy spectrum diagnosed from observations

On the other hand, due to stratification and rotation large-scale motions in the oceans and atmosphere are quasi two dimensional. Thus, barotropic eddies evolve with time, closely following the rules of 2-dimensional turbulence, i.e., energy cascades upscale (moves toward large scale), but enstrophy cascades down scale. However, due to the existence of planetary vorticity gradient, $\beta$, this cascade continues till the eddies reach the Rhines (1977) scale $l_{\text{Rhines}} = \sqrt{2U / \beta}$ where the upscale cascade cannot continue. For the ACC we assume $U=60\text{cm/s}$, $\beta = 10^{-13} / \text{cm/s}$, $l_{\text{Rhines}} \approx 300\text{km}$. This would corresponds to a rather large wave length, on the order of 1500-2000 km. This would mean there might be many large eddies there in the Southern Oceans with such a typical horizontal dimension. With the sparse data collected in the Southern Ocean, it remains rather difficult to compare the data with the theory.
With the advance of satellite data, we are now able to see such a comparison. Note that the satellite data come primarily in forms of sea surface elevation anomaly. As discussed above, at middle and lower latitudes, SSH signals primarily reflect the contributions from the first baroclinic mode. On the other hand, at high latitudes the stratification is weakened, thus the SSH may also include substantial contribution from the barotropic mode. The separation of signals into different components remains a great challenge.

Note that a positive slope of the spectral kinetic energy indicates a source of eddy kinetic energy at that frequency domain. For example, at 55°S, there is kinetic energy source for wavelength 150-250km, which is near, but slightly larger than, the length scale corresponding to the first deformation radius, as shown in E) of Fig. 19. For this latitude band, the arrest wavelength is about 250 km, which is much smaller than that predicted from the $\beta$-arrest scale of Rhines (1977). At lower latitudes, the range of scale of the eddy kinetic energy source and the arrest wavelength increase. Apparently, results from satellite data analysis at lower latitudes are close to the theory of Rhines (1977), but at high latitudes, there is a major discrepancy between the theory and observations.

![Time-mean spectral kinetic energy](image)

Fig. 19. Time-mean spectral kinetic energy diagnosed from satellite altimetry data from latitude band of A) 15°S, B) 25°S, C) 35°S, D) 45°S, E) 55°S. (Scott and Wang, 2005)

A major question remains open: How do the meso-scale eddies loose their energy? There are at least two possible channels:

a) Through the interfacial form stress and eventually reaching the sea floor where it is dissipated through the bottom drag.

b) The geostrophic balance in the meso-scale eddies may break down through some energetic episodes, and thus feed their energy to turbulence and internal waves; however, this remains very unclear at this time. This is currently a new research frontier, and the reader who is
interested in this subject may search new papers related to balance equations, such as Molemaker et al. (2005).

5. What controls the properties of AABW, such as the volume flux and the T, S properties?
a) The thermohaline forcing condition at the sea surface
b) Wind stress and its curl
c) Baroclinic instability
d) Cabbeling

References
Foster and Carmack, 1975.
Huang, R. X. et al., 2006.
Krupitsky and Cane, 1995.
Schmitz, W., 1995.
Stommel, H. M., 1957.