Critical Control at Deep Ocean Passages

J. A. Whitehead, Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, USA. jwhitehead@whoi.edu

Flux of oceanic water masses is a fundamental quantity in physical oceanography, important to understanding aspects of the ocean's role in climate, flux of dissolved and particulate chemicals, and movement of biological species, (especially small organisms) in the ocean. Records of past circulation of deep and bottom water from isotope or skeletal analysis are valuable indicators of past climate variations. Flows found at the deepest saddle points between benthic ocean basins (Whitehead, 1989; Whitehead et al., 1974), yield some of the best measurements of volume flux, and produce deep ocean mixing estimates. They are frequently unidirectional as dictated by the source of bottom water from convection far upstream. If the interface between bottom and above-bottom water controls flux, flows should be in a state of critical (maximised) control (Whitehead et al., 1974) so speeds are equal to internal wave speeds (Gill, 1977). This dynamics (called "rotating hydraulics", Whitehead et al., 1974; Pratt and Lundberg, 1991) produces simple and inexpensive predictions of flux for these important benthic water masses. Unfortunately, verification has lagged as it requires extensive and expensive measurements. Moreover, earlier tests used differing approximations and methods (Bethoux, 1979; Borenäs and Lundberg, 1986) so accuracy determination was difficult. Here, flux predictions, admittedly overestimates, (Killworth, 1992; Killworth and McDonald, 1993; Killworth, 1994) are made for the eight most highly measured sill flows in the world oceans using one method (Whitehead, 1989) requiring only data from two CTD hydrocasts and bathymetric data. Comparison with estimates using current meters and geostrophic calculations gives ratios of prediction to measurement from 1.0 to 2.7.

Maps of the Atlantic, Pacific, and Indian Oceans (Fig. 1) with the 4000 m and 5000 m contours separating three shades of grey show that water below 5000 metres lies in distinct isolated basins. Typically, bottom water must rise to depths between 4000 and 5000 metres to flow into the next basin through the deepest connecting passageway, some are labelled in Fig. 1. Flows near the saddle point of those passages are found to be relatively unidirectional and steady. Table 1 contains some estimates of volume flux (from current meter and/or geostrophic calculations) through sill-shaped gaps.

For close to two decades, theory of abyssal ocean circulation pictured a continuous "western boundary current" in conjunction with recirculation gyres in the oceanic interior (Warren, 1981). Many predictions about the western-intensified current are in beautiful accordance with ocean observations. The theory works best for a deep homogeneous layer over a flat bottom in large ocean basins on a rotating, spherical earth. Below this layer, interbasin bottom water flow over saddle points must have dynamics

Table 1. Estimates of volume flux (Q_e), values of density difference ($\Delta\rho/\rho$), height (h), width (W), calculated Rossby Radius (R), and predicted volume flux (Q_p) for deep sills in the world's oceans which have measurements with current meters over a reasonable length of time, and/or geostrophic calculations supplemented with current meters. In three cases the width W is less than R and in the other five it is greater than R. Below these are some other estimates of volume flux over sills based on smaller amounts of geostrophic data.

Sill	Q_e (Sv)	$\Delta\rho/\rho$ $\times 10^4$	h (m)	W (km)	R (km)	Q_p (Sv)
Denmark Straits (Dickson et al., 1990)	2.9	3.0	580	350	14	3.8
Ceara Abyssal Plain (Hall et al., 1997)	2	0.5	430	700	65	4.6
Vema Channel (Speer and Zenk, 1993)	6	1.0	1540	446	25	16.3
Discovery Gap (Saunders, 1987)	0.21	0.1	600	80	4	0.21
Samoa Passage (Johnson et al., 1994)	6	0.3	1050	240	34	7
Vema Gap (Fischer et al., 1996)	2.1	0.5	1000	9	35	3.3
Faeroe Bank Channel (Saunders, 1990)	1.9	5.0	400	20	15	3
Romanche Fracture Zone (Polzin et al., 1996)	1	0.47	380	20	369	2.2
Bornholm Strait (Petrin and Walin, 1976)	0.02					
Iceland-Faeroe Passage (Steele et al., 1962)	~1					
Chain Fracture Zone (Polzin et al., 1996)	0.1					
Hunter Channel (Speer and Zenk, 1993)	0.7					
Charlie-Gibbs Fracture Zone (Saunders, 1994)	2.4					
Strait of Sicily (Grancini and Michelato, 1987)	0.6–0.8					
(Morel, 1971)	0.65					
(Molcard, 1972)	1.23					
(Garzoli and Maillard, 1975)	1.21					
Anagada-Jungfern Passage (Stalcup et al., 1975)	0.056					

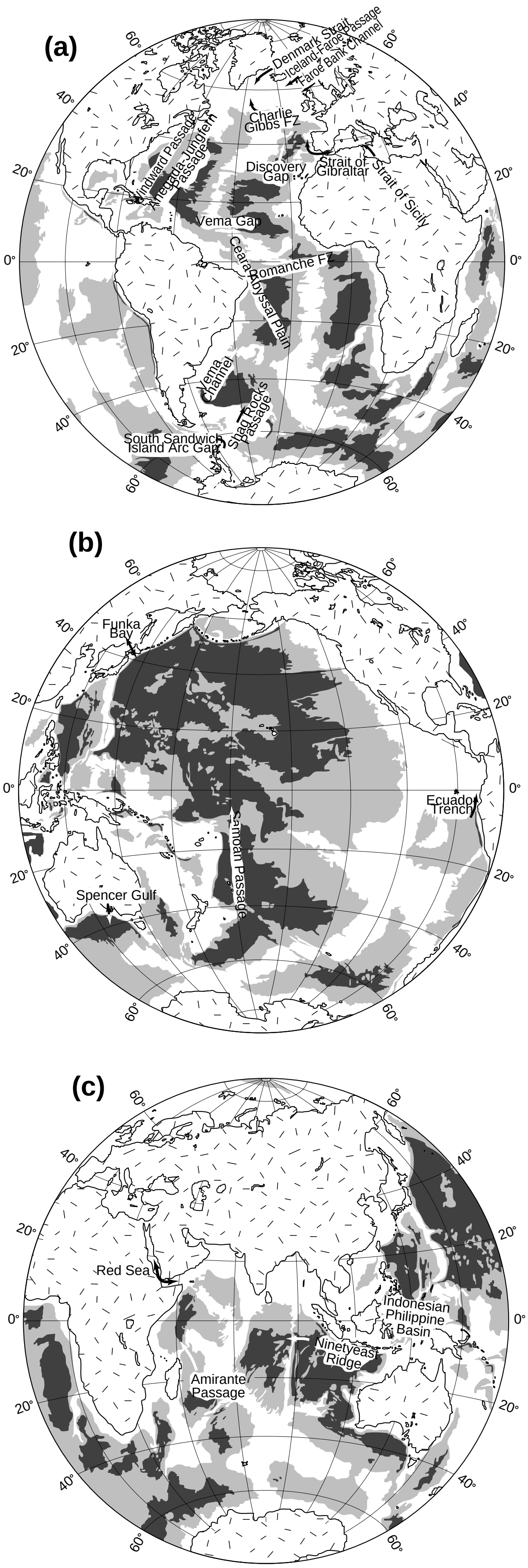


Figure 1. *The oceans with depth below 5000 m in black, and above 4000 m in light grey. A number of known saddle points between deep basins (usually, but not always, in black) are labelled. (a) Atlantic Ocean. (b) Pacific Ocean. (c) Indian Ocean.*

that differ from abyssal circulation theory for many reasons. First, the region is localised rather than global so f-plane approximations are more relevant than beta plane ones, especially since bathymetry has large local changes. Second, bottom shoaling also means that the flow must accelerate locally. Third, the vertical variation of density with depth must come into play since the fluid remains stably stratified in such regions. By retaining these three aspects, local rotation rate, acceleration, and stratification, one can find three dynamic relations (Whitehead et al., 1974; Gill, 1997): (1) a geostrophic relation, (2) the conservation of potential vorticity and (3) Bernoulli's law. By combining these, assorted solutions have been found. All involve considerable simplifications, for example continuous variation of density with depth is invariably abandoned and incorporating a simplifying assumption for upstream potential vorticity is virtually required to get uncomplicated solutions (Pratt and Lundberg, 1991). Even so, volume flux is usually predicted by complicated relations (Whitehead et al., 1974; Gill, 1977; Pratt and Lundberg, 1991), so we use the simplest one (zero potential vorticity (infinite depth upstream) and one layer in a rectangular channel, Whitehead et al., 1974) here:

$$Q_p = \frac{g\Delta\rho h^2}{4\rho Q\sin\theta'} W > R\left(= \left(\frac{g\Delta\rho h}{2\rho(Q\sin\theta)^2}\right)^{1/2}\right)$$

otherwise

$$Q_p = \left(\frac{2}{3}\right)^{3/2} W\sqrt{g\Delta\rho}\left[h - \frac{\rho Q^2 \sin\theta^2 L^2}{2g\Delta\rho}\right]^{3/2}$$

where g is gravity, Qsinθ is the local vertical angular earth rotation vector, h is height of the interface of the fluid above the floor of the channel, W is width of the channel, $\rho + \Delta\rho$ is density of a layer of fluid that is infinitely deep upstream, and a motionless fluid of density ρ lies above the flowing fluid layer. Such flux differs by less than 22% from that with upstream depth constant (Whitehead, 1989).

Do these formulae reliably predict oceanic measurements like those listed in Table 1? A consistent method (Whitehead, 1989) was used in 1988, to generate values of Δρ,h, and L (values of *g* and θ are obvious) using features of the bifurcation diagram in Fig. 2. Predictions were 1.6 to 4.2 times greater than observations for the only four sills with sufficient current meter measurements and geostrophic calculations to estimate volume flux. Here, we review three predictions at old locations which have new measurements. The fourth old location (Whitehead, 1995) is deleted as the current source region was not obvious. Five new current meter-based estimates augmented with geostrophic

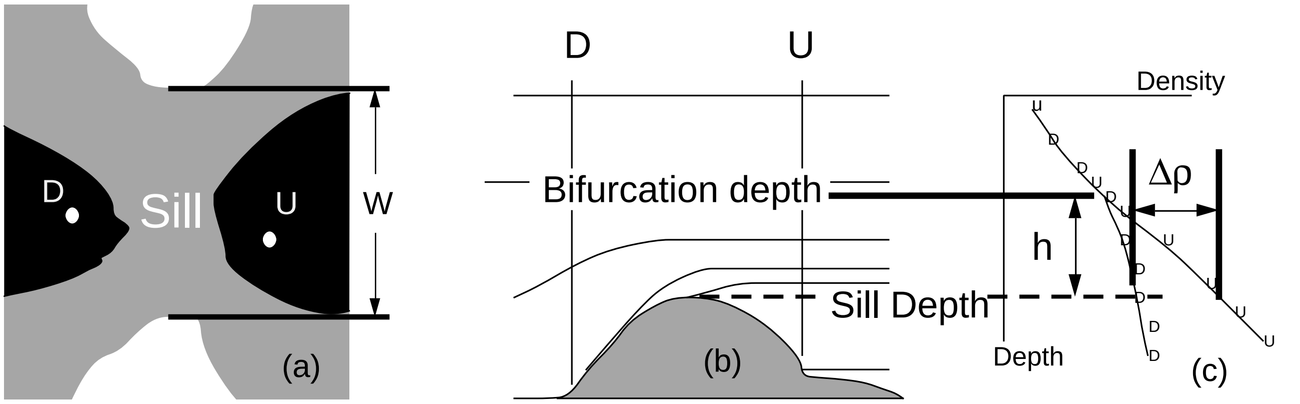


Figure 2. A sketch of the major features of critical control at a sill and outline of the method to produce values to use in the equations. (a) Plan view of bathymetry near a saddle point (sill) (b) Elevation view of a saddle point with isopleths of constant density sketched in the asymmetry of the isopleths denote flow from right (upstream U) to left (downstream Δ). (c) Two stations of density versus depth, one in the upstream basin and one downstream. Stations must be in basin interiors as sketched in (a) so that interior density is sampled. At some height h above the sill depth, upstream and downstream curves of density versus depth bifurcate. Width W is bathymetric width at bifurcation depth (a) and Δ_ρ is upstream and downstream density difference at sill depth (b). Using g=9.8 m/s and local latitude, values were used in the above equations to predict volume flux Q_p in Table 1.

calculations come from major ocean programmes. Most estimates required a large effort, with numerous moorings set for long intervals using two or more oceanographic cruises.

Here we compare the eight sets of oceanic measurements against predictions repeating the simple method outlined above. Bifurcation diagrams and bathymetric information of seven have been presented (Whitehead, 1995) elsewhere. The eighth is in press (Whitehead, 1997). Predicted volume fluxes are displayed along with estimates in Table 1 and Fig. 3. Ratios of prediction to estimates have decreased to the range from 1.0 to 2.7, the decrease from 4.2 being from better ocean data. Essentially the simple predictions are within the range of volume flux estimates over a broad range of parameters.

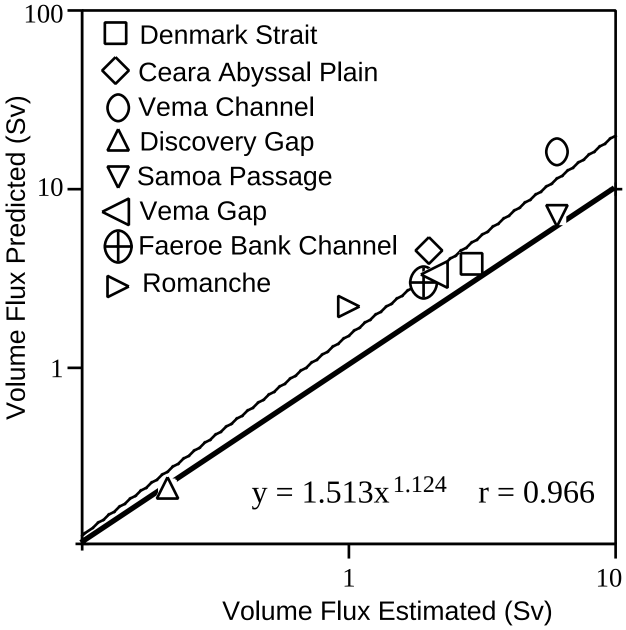


Figure 3. Comparison of Q_p with Q_e for the eight cases with best measurements listed in Table 1. The light line is a least squares power law fit to the data. The equation is the best fit equation, and r is the least squares regression value. The heavy line locates perfect fit between prediction and measurement. The predictions and estimates form a linear trend over a wide range of parameters. Predictions consistently form an overestimate as expected.

Of course, disagreement is still sizeable and there are many sources of difference. The admittedly simplified theory is both an overestimate (Whitehead, 1989; Whitehead et al., 1974; Gill, 1977; Pratt and Lundberg, 1991; Killworth, 1992; Killworth and McDonald, 1993; Killworth, 1994) and ignores continuous stratification, friction, complex bathymetry and turbulence. None have been incorporated in theories of rotating hydraulic control of unidirectional flow over a sill. In addition, measurements, never completely resolve a current, either in space or time. Errors of both theoretical shortcomings and measurement resolution each probably exceed 30%.

In summary, this simple and cheap method can apparently produce a crude prediction of flux over a range of parameters using only archived data in many regions. Additional data would clarify many questions concerning critical control, friction, and the many simplifications needed to analyse rotating hydraulics. With improvement this may produce an accurate, cheap estimate of flow through numerous oceanic sills.

References

Bethoux, J. P., 1979: Budgets of the Mediterranean Sea. Their dependence on the local climate and the characteristics of the Atlantic waters. Oceanol. Acta., 2, 157–163.

Borenäs, K., and P. Lundberg, 1986: On the Deep-Water Flow Through the Faeroe Bank Channel. J. Fluid Mech., 167, 309–326.

Dickson, R. R., E. M. Gmitrowicz, and A. J. Watson, 1990: Deep Water Renewal in the Northern Atlantic. Nature, 344, 848–850.

Fischer, J., M. Rhein, F. Schott, and L. Stramma, 1996: Deep Water Masses and Transports in the Vema Fracture Zone. Deep-Sea Res., 43, 1067–1074 (1996).

Garzoli, S., and Maillard, 1976: Hydrologie et circulation hivarnales dans les canaux de Sicile et de Sardaigne. C. Rapp. Int. Lab. Oceanogr. Paris. Museum National d'Histoire Naturelle, Paris, 21 pp.

Gill, A. E., 1977: The hydraulics of rotating channel flow. J. Fluid Mech., 80, 641–671.

Grancini, G. F., and Michelato, 1987: Current Structure and variability in the Strait of Sicily and adjacent area. A. Annales Geophysicae, 5B, (1) 75–88.

Hall, M. M., M. E. McCartney, and J. A. Whitehead, 1997: Antarctic Bottom Water Flux in the Equatorial Western Atlantic. J. Phys. Oceanogr. In press.

Killworth, P. D., 1992: Flow properties in rotating, stratified hydraulics. J. Phys. Oceanogr., 22, 997–1017.

Killworth, P., 1994: On reduced-gravity flow through sills. Geophys. Astrophys. Fluid Dyn., 75, 91–106.

Killworth, P. D., and N. R. McDonald, 1993: Maximal reduced-gravity flux in rotating hydraulics. Geophys. Astrophys. Fluid Dyn., 70, 31–40.

Morel, A., 1971: Caracteres hydrologiques des eaux echanges entre le bassin oriental et le bassin occidental de la Mediterranée. Cah. Oceanogr., 22, 4, 329–42.

Molcard, R., 1972: Preliminary results of current measurements in the Strait of Sicily in May 1970. Proc. Saclant Conf., 7, 82–95.

Petrin, O., and G. Walin, 1976: Some observations of the deep flow in the Bornholm Strait during the period June 73–December 74. Tellus, 28, 74–87.

Polzin, K. L., K. G. Speer, J. M. Toole, and R. W. Schmitt, 1996: Intense Mixing of Antarctic Bottom Water in the equatorial Atlantic Ocean. Nature, 380, 54–57.

Pratt, L. J., and P.A. Lundberg, 1991: Hydraulics of rotating strait and sill flow. Ann. Rev. Fluid Mech. 23, 81–106.

Rudnick, D. L., 1997: Direct Velocity Measurements in the Samoan Passage. J. Geophys. Res., 102, 3293–3302.

Saunders, P. M., 1987: Flow through Discovery Gap. J. Phys. Oceanogr., 17, 631–643.

Saunders, P. M., 1990: Cold Outflow from the Faeroe Bank Channel. J. Phys. Oceanogr., 20, 29–43.

Saunders, P. M., 1994: The flux of overflow water through the Charlie-Gibbs Fracture Zone. J. Geophys. Res., 99, 12343–12355.

Speer, K. G., and W. Zenk, 1993: The flow of Antarctic Bottom Water into the Brazil Basin. J. Phys. Oceanogr., 23, 2667–2682.

Stalcup, M. C., W. G. Metcalf, and R. G. Johnson, 1975: Deep Caribbean inflow through the Anegada-Jungfern Passage. Marine Res. (Suppl.), 33, 15–35.

Steele, J. H., J. R. Barrett, and L. V. Worthington, 1962: Deep currents south of Iceland. Deep-Sea Res., 9, 465–474.

Warren, B. A., 1981: Deep Circulation of the World Ocean. In Evolution in Physical Oceanography, pp. 6-41, Warren, B. A. and Wunsch, C., Eds., MIT Press.

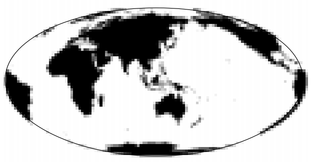
Whitehead, J. A., 1989: Internal hydraulic control in rotating fluids – applications to oceans. Geophys. Astrophys. Fluid Dyn., 48, 169–192.

Whitehead, J. A., 1995: Critical control by Topography - Deep Passages, Straits and Shelf Fronts. In Topographic Effects in the Ocean, pp. 141–156, Müller, P., and Henderson, D. eds., SOEST Special Publication.

Whitehead, J. A., 1997: Critical Control by Ocean Passages. Rev. Geophys. In press.

Whitehead, J. A., A. Leetmaa, and R. A. Knox, 1974: Rotating hydraulics of strait and sill flows. Geophys. Fluid Dyn., 6, 101–125.

Topography and Barotropic Transport Control by Bottom Friction



William K. Dewar, Department of Oceanography, Florida State University, Tallahassee, USA. bill@ocean.ocean.fsu.edu

Introduction

Saunders and King (1995), in an analysis of WOCE cruise A11 South Atlantic observations, have inferred the existence of a major anticyclonic flow. This flow appears to be centred on the so-called Zapiola Drift, a bottom topographic feature consisting of a local high. The transport associated with the anticyclone was estimated to be roughly 100 Sv, and bottom flows on the order of 10 cm/s were found. Evidence from other sources (Flood and Shor, 1988; Weatherly, 1993) argue the Zapiola anticyclone is a mean feature of the circulation in the area. Here we present some simple theoretical ideas which potentially apply to the Zapiola anticyclone. A fuller accounting can be found in Dewar (1997).

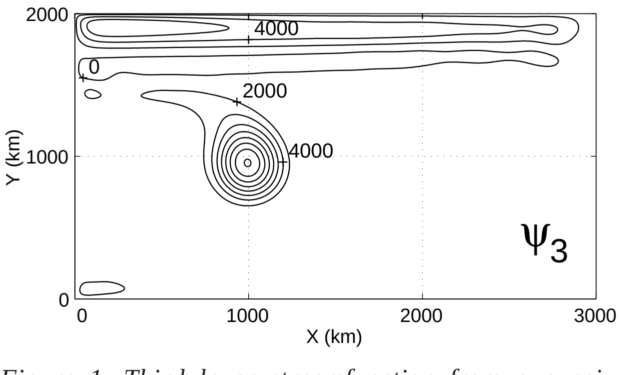


Figure 1. Third layer streamfunction from a quasi-geostrophic model. A strong anticyclonic circulation occurs over a Gaussian bump. CI = 2000 m^2/s.

Theoretical development

The equations of motion for a quasigeostrophic two layer ocean are:

$$J(\psi_i, q_i) = \frac{f_0 w_e}{H_1}\delta_{i,1} + R\nabla^2 q_i - D\nabla^2 \Psi_2 \delta_{i,2}$$

$$q_i = \beta y + \frac{f_0^2(\Psi_2 - \Psi_1)(-1^{i+1})}{g'H_i} + \frac{f_0 h_b \delta_{i,2}}{g'H_2} \qquad (1)$$