5
Dynamics of Large-Scale Ocean Circulation

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5.1 Introduction and Summary

The past 30 years have witnessed a rapid evolution of circulation theory. Much of the progress can be attributed to the intuition and physical balance that have emerged from the use of simple models that isolate important processes. Major contributions along these lines were made by Stommel, Welander, and others. An excellent presentation of the ideas together with a number of significant advances appears in Stern's (1975a) book. More recently numerical simulations have provided a different attack on the problem. Processes that are difficult to study with analytical models become accessible through the latter approach. Early, climatological-type studies by Bryan have now been supplemented by numerical models oriented toward the isolation of the effects of individual mechanisms. The papers of Rhines and Holland cited below have been especially instructive.

The development of the theory for the dynamics of large-scale oceanic flows is very recent. One has only to look at the chapter on dynamics in Sverdrup, Johnson, and Fleming (1942) to realize how primitive the theory was in the mid-1940s. Sverdrup's (1947) important demonstration of the generation of planetary vorticity by wind stress was the first step in obtaining explicit information about oceanic flow from a simple external observable. Until that time the dynamic method (i.e., geostrophic–hydrostatic balance) was used to obtain flow information, but this hardly constitutes a theory since one internal property must be used to determine another.

Ekman's (1905) theory for what we now call the Ekman layer was a significant early contribution, but its application to large-scale theory was not understood until Charney and Eliassen (1949) showed the coupling to large-scale flows via the spin-up mechanism. Actually, the generation of large-scale flow by Ekman suction in the laboratory was observed and described by Pettersson (1931), who repeated some of Ekman's (1906) early experiments with a stratified fluid to determine the inhibition of vertical momentum transport by stratification. Pettersson found the large-scale circulation to be an annoying interference, however, in his primary objective, determining vertical transfer of momentum by turbulence, and he discarded the approach as unpromising.

Shortly after Sverdrup's paper Stommel (1948) produced the first significant, closed-basin circulation model showing that westward intensification of oceanic flow is due to the variation of the Coriolis parameter with latitude. Hidaka (1949) proposed a closed set of equations for the circulation including the effects of lateral (eddy) dissipation of momentum. Munk (1950) continued the development by obtaining a solution that resembled Stommel's except for details in the boundary layers near the eastern and western sides of the basin. He applied his solution to an idealized ocean basin with observed wind stresses and related a number of observed oceanic gyres to the driving wind patterns. The first nonlinear correction to these linearized models (Munk, Groves and Carrier, 1950) showed that inertia shifts positive vortices to the south and negative vortices to the north. Nonlinear effects thus introduce the observed north-south asymmetry into a circulation pattern that is predicted by steady linear theory to be symmetric about mid-latitude when the wind driving is symmetric.

Fofonoff (1954) approached the problem from the opposite extreme, treating a completely inertial, nondriven model. His solution exhibits the pure effect of inertia for steady westward flows. The circulation pattern is symmetric in the east-west direction and closes with the center of a cyclonic (anticyclonic) vortex at the south (north) edge of the basin. When linear, frictional effects perturb the nonlinear pattern (Niiler, 1966), the center of the vortex shifts westward. Niiler's model had been proposed independently by Veronis (1966b) after a numerical study of nonlinear effects in a barotropic ocean, and Niiler's solution had been suggested heuristically by Stommel (1965).

The theoretical models leading to these results for wind-driven circulation are discussed below in sections 5.5 and 5.6. More general considerations in section 5.2, based on conservation integrals for the nondissipative equations (Welander, 1971a), prepare the way for the ordered system of quasi-geostrophic equations that are presented in section 5.3. The latter are derived for a fluid with arbitrary stable stratification and for a two-layer approximation to the stratification.[1] A large portion of the remainder of the paper reports results obtained with the simpler two-layer system.[2]

Section 5.7 concludes the discussion of simple models of steady, wind-driven circulation with a suggested simple explanation of why the Gulf Stream and other western boundary currents leave the coast and flow out to sea (Parsons, 1969; Veronis, 1973a). Separation of the Gulf Stream from the coast occurs within an anticyclonic gyre at a latitude where the Ekman drift due to an eastward wind stress in the interior must be returned geostrophically in the western boundary layer. If the mean thermocline depth is sufficiently small, i.e., if the amount of upper-layer water is sufficiently limited, the thermocline surfaces on the onshore side of the Gulf Stream and separation occurs. The surfacing of the thermocline is enhanced by the poleward transport by the Gulf Stream of upper-layer water that eventually reaches polar latitudes and sinks.

A review of models of thermohaline circulation is given in section 5.8. The open models introduced by Welander (1959) and Robinson and Stommel (1959) and the subsequent developments by them as well as other authors are described. The section concludes with a description of a closed, two-layer model in which the heating and cooling processes are parameterized by an assumed upwelling of lower-layer water across the thermocline (Veronis, 1978). The closure of the model leads to an evaluation of the magnitude of upwelling of $1.5 \times 10^{-7}\ \mathrm{m\,s^{-1}}$, in agreement with values obtained from chemical tracers and the estimated age of deep water.

The normal modes for a two-layer system are derived in section 5.9 and the free-wave solutions are obtained for an ocean of constant depth. The derivation is a generalization of the treatment by Veronis and Stommel (1956) but the method is basically the same. The results include barotropic and baroclinic modes of inertiogravity and quasi-geostrophic Rossby waves. Brief mention is made of observations of these waves and the roles they play in developed flows.

Topography introduces a new class of long-period wave motions. Quasi-geostrophic analysis leads to the three types of waves described by Rhines (1970, 1977) as topographic-barotropic Rossby waves, fast baroclinic (bottom-trapped) waves, and slow baroclinic (surface-trapped) waves. The properties of slow baroclinic waves are independent of topography, yet the creation of these waves may be facilitated by steep topography that inhibits deep motions. For purposes of comparison the analysis is carried out with stratification approximated by two layers and by a vertically uniform density gradient.

Baroclinic instability in a two-layer system is described in section 5.11. The model (Phillips, 1951; Bretherton, 1966a) has convenient symmetries (equal layer depths and equal and opposite mean flows in the two layers) that simplify the analysis and show the nature of the instability more clearly. The stabilizing effect of β is evident after the simpler model has been analyzed. After a discussion of the energetics and of the relative phase of the upper- and lower-layer motions required for instability, the study of linear processes ends with a brief review of the stability study made by Gill, Green, and Simmons (1974) for a variety of mean oceanic conditions.

The last section extends the discussion to include the effects of turbulence and strong nonlinear interactions. Batchelor's (1953a) argument that two-dimensional turbulence leads to a red cascade in wavenumber space is followed by a description of several of Rhines's (1977) numerical experiments exhibiting the red cascade for barotropic quasi-geostrophic flow and the inhibition of the red cascade by lateral boundaries and topography. An initially turbulent flow in a two-layer fluid will evolve toward a barotropic state followed by the red cascade when nonlinear interactions or baroclinic instability generate motions on the scale of the internal radius of deformation. The latter scale is the window leading to barotropic behavior. Rough topography can inhibit the tendency toward barotropy by scattering the energy of the flow away from the deformation scale.

The generation of deep motions in wind-driven flows by upper-layer eddies that evolve from barotropic and baroclinic instabilities leads to a mean flow that is very different from the one predicted by the linear theories of the earlier sections. The closed-basin circulation obtained in a two-layer quasi-geostrophic numerical experiment by Holland (1978) and analyzed by Holland and Rhines (1980) shows how many of the processes described earlier come together to generate the mean flow. Simple balances for some of the results are suggested. A significant result of this experiment (and others mentioned) is the enhancement of the mean transport by the circulation resulting from the eddy interactions. A similar enhancement is made possible when topography and baroclinic effects are present (Holland, 1973). A brief discussion of several other numerical studies concludes the review.

Most of the emphasis in this paper is on linear processes and on the remaining features of the dynamics that can be used as building blocks to synthesize the involved, interactive flows observed in the ocean. Only a selected few of the many numerical studies that have emerged in the past few years are discussed, and even for those only some of the generalizable results are mentioned. Some important topics, such as the use of diagnostic models (Sarkisyan, 1977) and the generation of mean circulation by fluctuating winds (Pedlosky, 1964a; Veronis, 1970; Rhines, 1977), are omitted only because time limits forced me to draw the line somewhere. Most of the references are to the literature in the English language because that is the literature with which I am most familiar.

5.2 The Equations for Large-Scale Dynamics

The complete equations for conservation of momentum, heat, and salt are never used for studies of large-scale oceanic dynamics because they are much too complicated, not only for analytical studies but even for numerical analyses. Justification for use of an appropriate set of simplified equations requires a much more extensive argument than is feasible here so we shall confine ourselves to a short discussion with references to publications that discuss the different issues. It is appropriate, however, to mention a general result for a fluid with a simple equation of state.

If dissipative processes are ignored, the conservation of momentum for a fluid in a rotating system can be written as

$$\frac{\partial \mathbf{v}}{\partial t} + \mathbf{v}\cdot\nabla\mathbf{v} + 2\boldsymbol{\Omega}\times\mathbf{v} = -\frac{1}{\rho}\nabla p - \nabla\Phi, \tag{5.1}$$

or equivalently as

$$\frac{\partial \mathbf{v}}{\partial t} + (2\boldsymbol{\Omega} + \nabla\times\mathbf{v})\times\mathbf{v} = -\frac{1}{\rho}\nabla p - \nabla\left(\frac{1}{2}\mathbf{v}\cdot\mathbf{v}\right) - \nabla\Phi \tag{5.2}$$

where $\mathbf{v}$ is the three-dimensional velocity vector, $\boldsymbol{\Omega}$ is the angular rotation vector of the system, ρ the density, p the pressure, and $\nabla\Phi$ the total gravity term (Newtonian plus rotational acceleration).

Conservation of mass is described by

$$\frac{d\rho}{dt} + \rho\nabla\cdot\mathbf{v} = 0, \qquad \frac{d}{dt} = \frac{\partial}{\partial t} + \mathbf{v}\cdot\nabla. \tag{5.3}$$

Furthermore, if a state variable $s(p,\rho)$ is conserved along a trajectory, it satisfies the equation

$$\frac{ds}{dt} = 0. \tag{5.4}$$

These equations can be combined to yield the conservation of potential vorticity (Ertel, 1942):

$$\frac{d}{dt}\left[\frac{(2\boldsymbol{\Omega} + \nabla\times\mathbf{v})\cdot\nabla s}{\rho}\right] = 0. \tag{5.5}$$

This general result for a dissipation-free fluid does not apply precisely to sea water where the density is a function not only of temperature and pressure but also of the dissolved salts. The effect of salinity on density is very important in the distribution of water properties. However, for most dynamic studies the effect of the extra state variable is not significant and (5.5) is valid.

Circulation of waters in the world ocean involves trajectories from the surface to the deep sea and from one ocean basin to another. The relative densities of two parcels of water formed at the surface in different locations can be inverted when the parcels sink to great depths. Thus, surface water in the Greenland Sea is denser than surface water in the Weddell Sea; yet when these water masses sink and flow to the same geographic location, the latter (Antarctic Bottom Water) is denser and lies below the former (North Atlantic Deep Water). This inversion is due in large part to the different amounts of thermal expansion of waters of different temperatures and salinities.[3]

Neither compressibility nor individual effects of temperature and salinity on the density are included in the treatment that follows. Use of potential density (not only in the equations but in boundary conditions as well) together with the Boussinesq approximation (Spiegel and Veronis, 1960) makes it possible to treat the dynamic effects of buoyancy forces in a dynami-