The Dynamics of the Oceanic Subtropical Gyres

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Recent theoretical advances in physical oceanography have produced plausible models for the structure of the circulation in the oceans' subtropical gyres. These theoretical ideas are reviewed and assessed in the light of present observational evidence. The dominant role of potential vorticity is emphasized in describing the strong interplay between three-dimensional distributions of density and velocity. In particular, the complementary processes of subduction and recirculation jointly act to shape the circulation pattern of the subtropical gyre.

THE DOMINANT FEATURE OF THE LARGE-SCALE OCEANIC circulation is the great, swirling subtropical gyres in each ocean. The gyres extend poleward from about 10° north and south of the equator to about 45° in each hemisphere, and the water circulating in these massive pools reaches to nearly 2 km beneath the oceanic surface.

A largely schematic depiction of the gyre as represented by the field of surface velocity is shown in Fig. 1, which emphasizes an important observation. The vigor of the circulation in each of the gyres is intensified greatly in the western parts—where, for example, in the Atlantic the Gulf Stream is formed—and is relatively weak in its eastern parts outside the domain of the western boundary currents; the ratio of velocities in the western to eastern parts is at least 10:1. The broad region of relatively weak flow occupies most of the gyre and is called the Sverdrup regime. Many of the advances of the past decade in developing an understanding of the oceanic circulation have dealt with the physics of the Sverdrup regime. In this article, I outline some of the key dynamical ideas recently developed to explain the structure of this immense part of the ocean's general circulation and which form the impulse behind a set of imaginative field experiments designed to examine these ideas and to help clarify still obscure aspects of the physics.

The oceans are very shallow (~5 km) compared with their horizontal extent so that they are slender shells whose proportions closely resemble the thin pages of this magazine. Nevertheless, there is a rich structure for the velocity field and the fields of temperature and salinity (and thus density) with depth. Beneath a surface layer that is mixed to homogeneity by the action of the winds and local cooling to the atmosphere, the oceanic temperature decreases rapidly with depth in the main thermocline until uniform temperatures of near freezing are reached at great depth. The upper mixed layer is about 100 m thick, the thermocline is about 1 km thick, and the bulk of the remaining 3 or 4 km is filled with cold water of clearly polar origins (Fig. 2).

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The oceans, in their present configurations, are millions of years old. The radical departure of the temperature from a linear conductive gradient, which there was clearly sufficient time to establish, implies that a dynamic process is at work in forming the nonuniform temperature structure of the thermocline. Indeed, the temperature-depth curves (Fig. 2) are observed to vary geographically. Both the ocean's surface temperature and the depth of the thermocline vary with latitude and longitude. Such nonuniform temperatures imply that the density is horizontally nonuniform. The ocean's thinness allows us to calculate the pressure hydrostatically; the vertical accelerations are so limited by the thinness of the fluid that the pressure at any point is almost precisely equal to the weight of water above the point, aside from certain high-frequency fluctuations, which are not discussed further. Thus lateral nonuniformities in density lead to similar variations in pressure, and lateral variations in pressure will, in turn, lead to lateral motions. The motion induced by the lateral pressure gradients will act to rearrange the density field as well. Furthermore, over 60 years ago, Montgomery (*1*) and Iselin (*2*) observed that the vertical distribution of properties at depth were systematically related to their lateral distributions on the sea

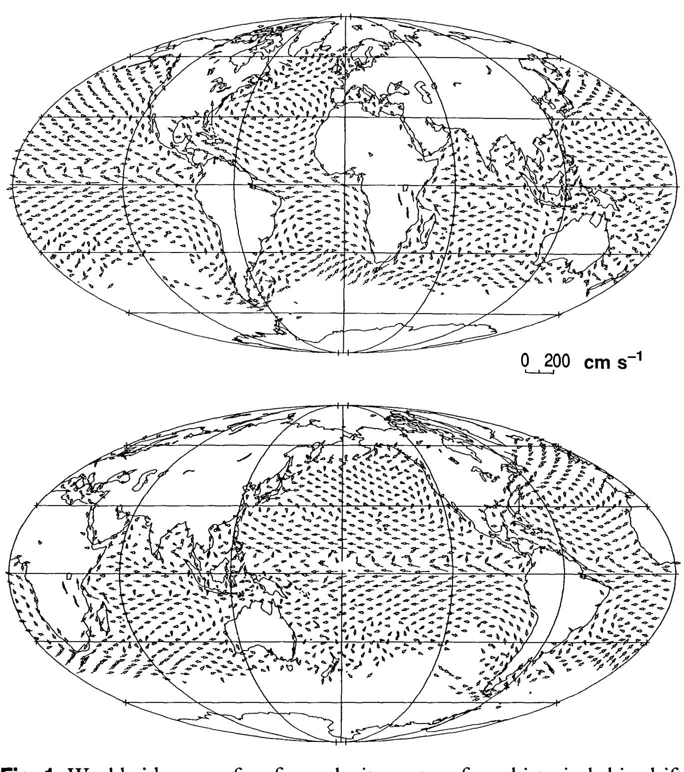

Fig. 1. Worldwide map of surface velocity vectors from historical ship-drift measurements. In each hemisphere, each ocean is dominated by the large, swirling subtropical gyres clockwise (counterclockwise) in the Northern (Southern) Hemisphere. [Courtesy of P. L. Richardson]

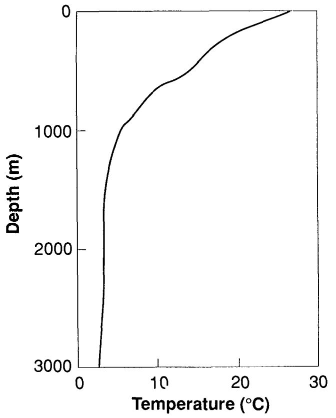

Fig. 2. A schematic rendering of a typical temperature-depth profile in the middle of the North Atlantic gyre. Note the relatively strong temperature gradient in the upper 1 km.

surface. Some advective process is therefore required to carry the surface signal to great depth in the thermocline, but, for good reason, the dynamics of the process resisted dynamical explanation until recently. The fields of density and velocity are nonlinearly coupled, and the problem of understanding the remarkable distribution of density in the ocean is identical to understanding the entire structure, lateral and vertical, of the oceanic general circulation. This is a problem in classical physics of great complexity.

Key Dynamical Ideas

The extraordinary slimness of the ocean has already been noted. For motions limited to the upper 1 or 2 km in a gyre thousands of kilometers in horizontal extent, the ratio of depth to width is on the order of 10^{-3}. Not only does this imply that the momentum balance is hydrostatic in the vertical direction,

$$\frac{\partial p}{\partial z} = -\rho g \tag{1}$$

where p is pressure, ρ is density, z is the local vertical coordinate, and g is the acceleration due to gravity, but also that the primary circulation velocities will be horizontal, that is, nearly tangent to the earth's surface. Where the ocean currents are weak, the dominant horizontal acceleration is the Coriolis acceleration suffered by each fluid element on the rotating earth, $2\mathbf{\Omega} \times \mathbf{u}$, where $\mathbf{\Omega}$ is the earth's rotation vector and $\mathbf{u}$ is the velocity of oceanic fluid. Because the velocity is nearly horizontal, the horizontal component of the Coriolis acceleration involves only the local normal component of the earth's rotation. If $\mathbf{F}$ is the sum of the applied horizontal forces per unit volume, then the horizontal momentum balance implies that

$$\rho f \hat{k} \times \mathbf{u} = \mathbf{F} \tag{2}$$

where $f = 2\Omega\sin\theta$, $\hat{k}$ is a unit vertical vector parallel to the local z axis, and θ is geographic latitude. In the broad regions of the ocean outside swift boundary currents, the force $\mathbf{F}$ is composed of the joint action of horizontal pressure gradients and, in the layers of fluid near the surface, the horizontal forces due to the turbulent stresses induced in the fluid by the frictional wind stress at the air-sea interface. Thus

$$F = -\nabla p + \frac{\partial \boldsymbol{\tau}}{\partial z} \tag{3}$$

where $\boldsymbol{\tau}$ is the turbulent shear stress in the fluid and where $\boldsymbol{\tau}$ (surface) $= \boldsymbol{\tau}_w$, the applied atmospheric wind stress, which is considered a known, externally given force.

A quantity of particular dynamical importance is the vorticity, which is the curl of the velocity. The vorticity consists of two parts. The first is the vorticity each fluid element possesses by sharing in the overall rotation of the planet. The component perpendicular to the earth's surface is

$$f = 2\Omega\sin\theta = [\nabla \times (\mathbf{\Omega} \times \mathbf{r})] \cdot \hat{k} \tag{4}$$

and, in the broad reaches of the Sverdrup regime, is many times greater than the vorticity due to the currents themselves, which make up the second part.

In the Northern Hemisphere, the eastward winds in mid-latitudes combine with the westward trade winds to reduce continuously the vorticity of fluid in the general circulation. The fluid responds by moving equatorward where f is smaller. This idea was quantified by Sverdrup (*3*) with the relation

$$\beta \int v\,dz = \hat{k} \cdot \frac{\nabla \times \boldsymbol{\tau}_w}{\rho} \tag{5}$$

where v is the meridional (northward) velocity, and the integral in Eq. 5 extends over the entire depth of the moving water. The Sverdrup relation, as Eq. 5 is known, requires that there are very weak currents at great depth. The key parameter in Eq. 5 is

$$\beta = \frac{2\Omega\cos\theta}{R} \tag{6}$$

where R is the earth's radius and β is the planetary vorticity gradient, that is, the northward increase with distance of f. In the subtropical gyres $\hat{k} \cdot \nabla \times \boldsymbol{\tau}_w$ is negative, and the total transport must move equatorward. This general southward drift must be compensated by northward flow, and Stommel (*4*) showed that the planetary vorticity gradient simultaneously selected the western side of the ocean as the location for the swift ocean return currents (for example, the Gulf Stream) in which the constraints of Eqs. 2 and 5 are broken. The western boundary current absorbs the clockwise Sverdrup circulation impinging on the western boundary and returns it to the northern part of the gyre. The overall pattern of the circulation is complex, and the dynamics of the merging of the gentle Sverdrup flow with the intense but geographically limited western boundary current flow is still poorly understood. However, from the point of view of understanding the Sverdrup regime in the eastern reaches of the ocean, the restriction of swift currents with their complex dynamics to the western part of the oceans permits great dynamical simplification. This geometry of the gyre has striking consequences for the circulation.

The Vertical Partition of the Motion

Quite near the upper surface, the turbulent stresses, $\boldsymbol{\tau}$, in Eq. 3 are important, whereas within the main body of the thermocline they are negligible. Vertical integration of Eqs. 2 and 3 over this uppermost layer yields for the direct stress-driven horizontal transport

$$\mathbf{U}_E = \int \mathbf{u}\,dz = -\frac{\hat{k} \times \boldsymbol{\tau}_w}{\rho f} \tag{7}$$

where the integral is over the upper layer in Fig. 3 in which $\boldsymbol{\tau}$ diminishes from its surface value $\boldsymbol{\tau}_w$ to zero. In the subtropical gyres the Ekman transport $\mathbf{U}_E$ is laterally convergent. Because at oceanic speeds the water is incompressible, this convergence produces a downwelling velocity

$$w_e = \hat{k} \cdot \nabla \times \left(\frac{\boldsymbol{\tau}_w}{\rho f}\right) < 0 \tag{8}$$

that is responsible for setting the deeper layers of the ocean in

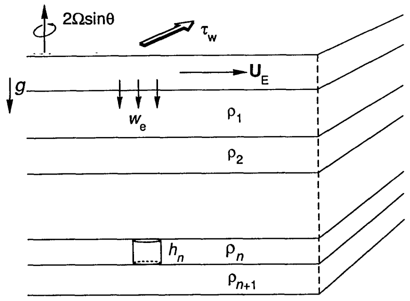

Fig. 3. A schematic of a density-stratified ocean driven by the surface wind stress τ_w. The transport directly driven by τ_w in the mixed layer is U_E. Convergence of U_E leads to a downwelling, w_e, into the upper layer of the ocean. Cylinder in the nth layer is in motion and conserves $q_n = 2\Omega\sin\theta/h_n$.

motion. This process is called Ekman pumping. The perplexing question that faced oceanographers was to understand how deep motion could take place. Reference to Fig. 3 makes the problem clear. The ocean may be considered as a set of layers, each of uniform density. By an increase in the number of layers, any desired degree of vertical resolution of the dynamical fields can be achieved. To a first approximation, beneath the stress-driven layer, the fluid will tend to preserve its density and hence remain in the layer in which it was initially. It is easy to see how the first layer beneath the surface layer can be set into motion. Fluid pumped downward according to Eq. 8 will force motion in the upper layer (labeled 1 in Fig. 3). Indeed, by Kelvin's circulation theorem applied to a circuit in layer 1 enclosing an area A, the product of the vorticity and A will be conserved. But because the vorticity is very nearly just f, the product fA is fixed. Because seawater is nearly incompressible, the pumping down from the surface layer will increase A. To preserve the product fA, f must decrease, that is, the fluid must move southward, as indicated by Eq. 5. At first glance, it appears that in the absence of strong frictional coupling between layer 1 and the lower layers, the lower layers would remain at rest. This catastrophe would compress the entire Sverdrup transport (Eq. 5) into the uppermost layer, and in the limit of increasing vertical resolution, the velocity would attain a delta function character with depth and hardly be realistic.

The resolution of this puzzle starts with the realization that beneath layer 1, the volume of any small cylinder bounded by density surfaces is nearly conserved. Let h_n be the height of the cylinder spanning a layer's thickness, then conservation of fA implies that

$$q_n = \frac{f}{h_n} \tag{9}$$

must be conserved on streamlines of flow. The parameter q_n is called the potential vorticity, and its expression in Eq. 9 is a special case of an elegant theorem (*5*) of wide application in both meteorology and oceanography. Streamlines of the subsurface flow must therefore coincide with lines of constant q_n in each layer. Were h_n constant or nearly so, the lines of q_n would be latitude circles and the flow in the deeper layer would be purely zonal. The streamlines would then intersect the eastern boundary. But because the eastern wall cannot support intense boundary currents that break the constraints leading to Eq. 5, no flow can slide along the streamlines out of the eastern oceanic boundary. The q_n contours hitting the eastern wall are "blocked" and will not support a q_n conserving flow. The fluid flow in all such regions will thus be trivially small unless directly forced by other mechanisms that destroy conservation of potential vorticity, a circumstance that will be discussed below.

The only way the layers below layer 1 can move in response to the action of the wind is if their thicknesses vary geographically to such an extent that the q_n contours depart significantly from latitude circles and avoid intersecting the eastern boundary of the oceanic basin. Of course, such variations of thickness will arise only if the layers are in motion so that, in general, determining the isolines of q_n is a fundamentally nonlinear problem, that is, the q_n isolines, which determine the pathways of flow, are part of the solution of the problem as well as determining its character. Before being concerned about the detailed nature of the q_n isolines, one could ask what alternatives are there for isolines of q_n that have starting points aside from the eastern oceanic wall. Two possibilities have been suggested, and the working out of the consequences of the two possibilities has led to two complementary pictures of the dynamics of the circulation.

Ventilated Layers in the Thermocline

The surface of the ocean is not uniform in temperature and salinity. Roughly, the temperature decreases poleward so that surface isotherms (see Fig. 4) run nearly east-west along the ocean's surface. The continuation of those isolines beneath the surface forms the topography of the density surfaces in the thermocline. In other words, Fig. 3 should more accurately be redrawn as Fig. 5, which shows upper, warmer, less dense layers in the thermocline reaching upward to the surface. In the regions in which a layer is in contact with the surface layer (the mixed layer), it ingests fluid pumped down by the wind in accordance with Eq. 8; this pumping sets the layer into motion. It is plausible, at least, that layer 3, for example, can be in motion even for $\theta < \theta_1$, where it is shielded from the direct pumping from the mixed layer. In this case, the deep motion is due to the surface ventilation of each layer and the subsequent subduction of the flow beneath warmer waters as the flow is driven southward in general agreement with Eq. 5.

Luyten *et al.* (*6*) described in detail how the process of ventilation and subduction allows q_n to be determined, for each layer, in regions where the layer is shielded from Ekman pumping by warmer layers above it. For each layer below the surface layer in which $\boldsymbol{\tau}$ is important, Eqs. 2 and 3 yield

$$f\mathbf{u}_n = \hat{k} \times \frac{\nabla p_n}{\rho} \tag{10}$$

so that streamlines for the horizontal motion are isobars as on a weather map. Hence the condition that q_n is conserved for all but the uppermost layer in contact with w_e yields

$$q_n = \frac{f}{h_n} = Q_n(p_n),\ m - 1 < n < N \tag{11}$$

where N is the deepest moving layer and m is locally the shallowest layer subject to pumping from the mixed layer. The hydrostatic equation (Eq. 1) relates p_n to the h_n so that Eq. 10 becomes $N - m$ equations for the $N - m + 1$ unknown layer depths. The remaining equation that closes the system is Eq. 5, which in the layered form is

$$\beta \sum_{n=m}^{N} v_n h_n = f w_e \tag{12}$$

That ventilation and subduction determine the potential vorticity along with its subsequent conservation is at the heart of this view of the circulation. One dramatic prediction of the theory is that it is impossible for flows that conserve potential vorticity to remain attached to the eastern boundary. In order to avoid flow through the eastern wall, Eq. 10 implies that the pressure and hence each layer thickness should remain constant along the eastern boundary. But at the same time $q_n = f/h_n$ must be conserved. Were h_n to remain constant as the fluid progressed southward under the influence of downward Ekman pumping, the decrease of f would decrease q_n. To preserve q_n the only alternative is for the flow to pull away from the eastern boundary as fluid moves southward. This opens up a

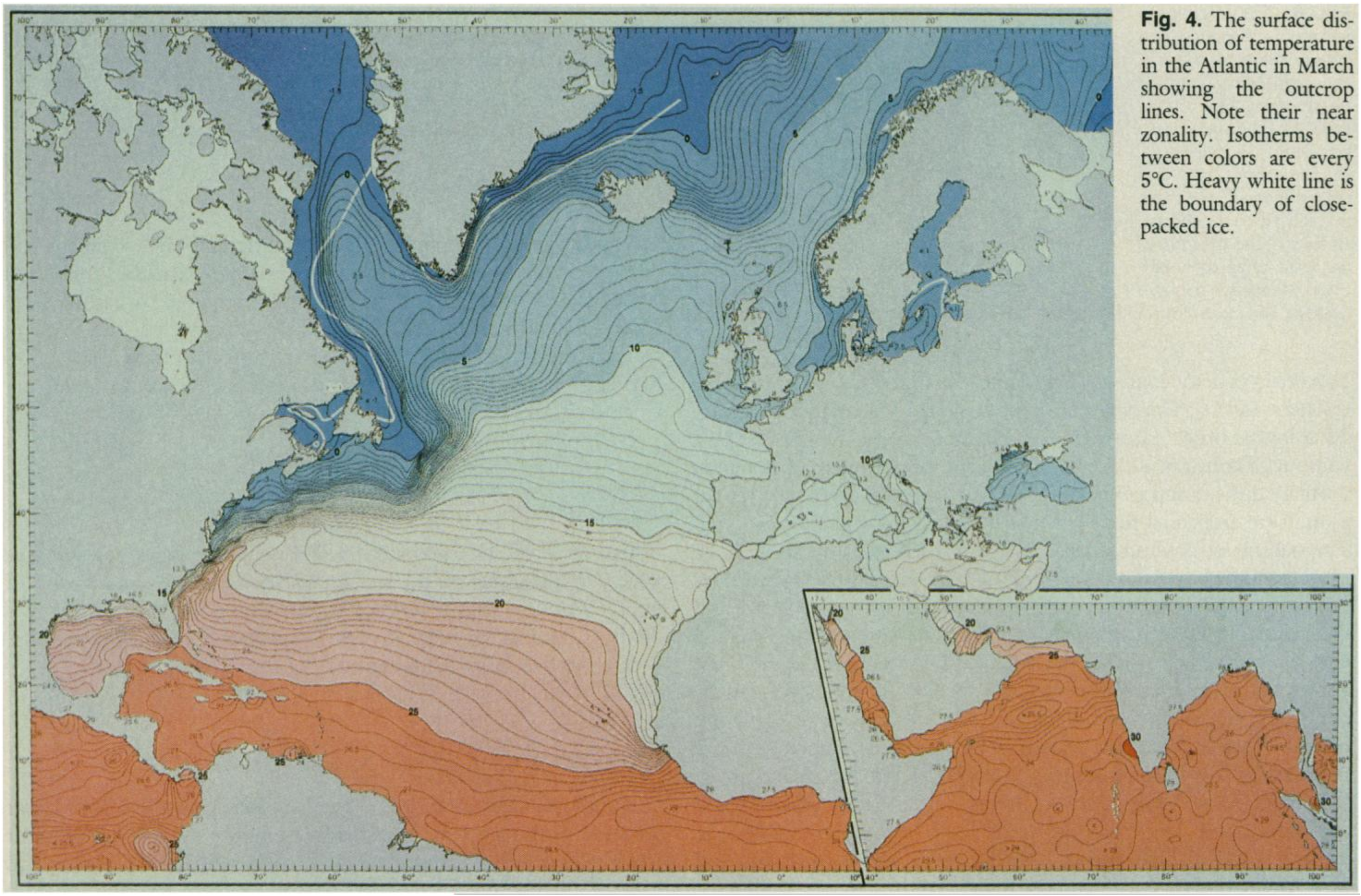

Fig. 4. The surface distribution of temperature in the Atlantic in March showing the outcrop lines. Note their near zonality. Isotherms between colors are every 5°C. Heavy white line is the boundary of close-packed ice.

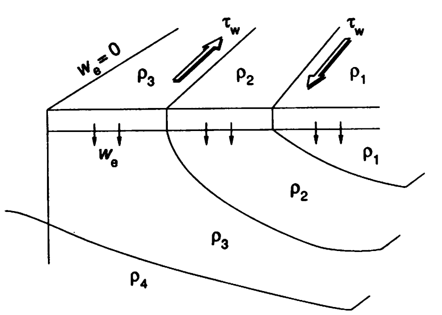

Fig. 5. A schematic showing the subsurface extensions of the outcrop geometry (Fig. 4). Determining the structure of the density interfaces below the surface is equivalent to the determination of the thermocline's thermal structure.

"shadow zone" adjacent to the eastern boundary that cannot be reached by subducted fluid (Fig. 6). This region remains unventilated. The presence of the shadow zone is a robust feature of the ventilation theory because it does not rely on exact conservation of potential vorticity. The absence of flow through the eastern boundary fixes h_n to be constant, and only special and accidental sources of potential vorticity would be consistent with the decrease of q_n imposed by the decrease of f. A general confirmation of these ideas is in the analysis by Jenkins (*7*) of the tracer, tritium, in the eastern regions of the ocean. The increase of age of the tritium with distance south of the outcrop line agrees well with the detailed predictions of ventilation models, in which the younger ages signify regions refreshed recently by subduction. The observations also raise further questions because the volume of observed subducted water is considerably greater than can be accounted for by Eq. 8.

There are other zones in the thermocline that cannot be ventilated by the subduction process. As Fig. 5 shows, deeper layers, but still within the main thermocline, actually outcrop at latitudes much farther north, that is, in the subpolar gyre, where the Ekman pumping velocity is upward and therefore unable to be a source of ventilation. Even in those layers that are ventilated, some of the streamlines in the characteristic clockwise swirl of streamlines predicted by the theory will have origins in the western boundary region. On those streamlines, q_n is not determined by subduction at outcrop lines. The fluid in such pools and in the deeper layers, if in motion, is entirely recirculating, without any refreshment at outcrop locations.

Recirculation Theories

In a series of important papers, Rhines and Young (*8–10*) considered quite a different escape mechanism by which the q_n contours can avoid striking the oceanic eastern boundary, which would consequently prohibit motion. They imagined the case where deeper layers do not outcrop but where the potential vorticity isolines, which are typically along latitude circles, are so distorted by the motion that they actually close upon themselves entirely within the unventilated density layer. If the motion is nearly inviscid, any flow circulation along such closed streamlines is a nonlinear, steady free mode of the system

$$\mathbf{u}_n \cdot \nabla q_n = 0 \qquad (13)$$

because $\mathbf{u}_n$ is then trivially perpendicular to ∇q_n. The smallest amount of frictional coupling to the layers above, directly driven by the wind, is then sufficient to produce substantial motion by a steady resonance of the free, nonlinear mode. Motion will therefore occur within a pool whose outer boundary is the outermost closed

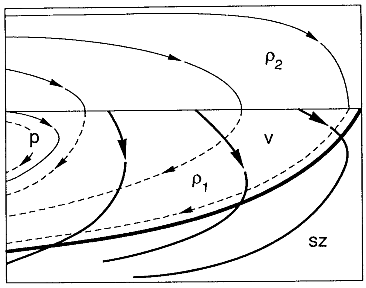

Fig. 6. The circulation pattern, in plan view of a simple two–moving layer ocean with a single outcropping line at latitude θ. The pool (p), ventilated region (v), shadow zone (sz), and streamlines of the flow in each layer are shown. In layer 2, the streamlines are solid until they subduct, whereafter they are dashed. The streamlines in layer 1 are heavy solid lines.

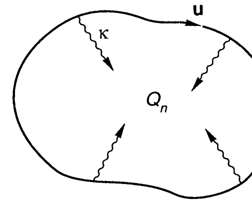

Fig. 7. A closed streamline on which q_n is the constant Q_n will encircle a domain of homogenized potential vorticity due to the slow but inexorable inward diffusion of q_n (diffusion coefficient κ) from the girdling contour.

potential vorticity contour (Fig. 7) on which, by Eq. 13, q_n is some constant, say Q_0. The presence of weak dissipation will not affect Eq. 13 at lowest order. However, once the pool in Fig. 7 is girdled by a contour of constant q_n, the weak diffusion of q_n inward, if potential vorticity diffuses according to Fick's law (*11*), will eventually render q_n uniform and equal to its boundary value everywhere in this pool. The outermost contour is determined in each layer by an iterative process. If the circulation is limited only to layers directly forced by ventilation, the q_n contours can be calculated in the layers beneath the ventilated region. If closed contours are found, the potential vorticity everywhere within the outermost contour is known by the homogenization hypothesis, and the motion in that layer can be calculated. Because it can be shown that the deeper recirculating pools are nested such that each deeper layer is contained within a smaller lateral domain than the pool above, the argument can be continued sequentially until at great depths layers with closed contours are no longer found. Such layers remain motionless. In realistic oceanic situations, it is unlikely that the q_n contours will close on themselves entirely in the middle of the ocean. Usually, the q_n contours must instead thread through western boundary current regions where the validity of Eq. 13 is unclear. In a bold move, Rhines and Young (*9*) advocated applying the homogenization argument to these flows as well.

When the two theories, the one involving subduction and the other homogenization, are combined (*12*), an elegantly formed shape is predicted for the thermocline circulation. Starting with the deepest layers that do not outcrop in the subtropical gyre, the circulation occurs in a small bowl nestled in the northwest corner of the subtropical gyre. As we proceed to density surfaces higher in the water column, the bowl expands both in longitude and latitude but is well removed from the eastern boundary. In this part of the bowl, the potential vorticity is homogeneous. Higher in the water column, density layers that are ventilated in the subtropical gyre are met and the circulation is laterally more extensive. The potential vorticity is inhomogeneous in these layers except in western pools of homogenized potential vorticity that shrink westward in extent as ever higher layers in the gyre are considered. The whole of the circulation, aside from its shallowest part, is detached from the eastern boundary of the ocean, which is the domain of the shadow zone. The overall depth δ_a of the anticyclonic gyre is given by the Welander advective scale (*13*)

$$\delta_a = \left[\frac{2\Omega R^2}{g\Delta\rho/\rho} W\right]^{1/2} \tag{14}$$

where W is the characteristic Ekman downwelling velocity and $\Delta\rho$ is the characteristic variation of the density in the thermocline (for example, its range of surface values). For example, for $\Omega = 7.3 \times 10^{-5}\ \mathrm{s}^{-1}$, $W = 2 \times 10^{-4}\ \mathrm{cm\ s}^{-1}$, $\Delta\rho/\rho = 5 \times 10^{-4}$, and $R = 6 \times 10^3$ km, all of which are typical of the oceanic subtropical gyres, the scale δ_a turns out to be about 1.5 km. This particular scaling law, which results from the momentum and vorticity balances described above, sets the overall scale. Actually, numerical values in each gyre will be modified by the spatial structure of the wind field and surface density distributions as well as the size and configuration of each basin. The advective scale, obtained with Eq. 8, is equivalent to a simple energy relation for the vertical scale, that is, with Eq. 8, we have the equivalent scaling relation

$$\frac{\tau_w}{\rho} \cdot R = g \frac{\Delta\rho}{\rho} \delta_a^2 \tag{15}$$

Equation 15 can be interpreted as balancing the work per unit area added to the gyre by the wind, which results in an increase of potential energy that is realized by deepening the thermocline to a depth δ_a, against the reduced gravitational force $g\Delta\rho/\rho$. In this form, δ_a is independent of the magnitude of the earth's rotation.

Observational Tests of the Theory

The theory described above may be considered the lowest order theory. It is innocent of a substantial number of recent elaborations and modifications discussed below. However, it is the theory in this form that has so far been subject to observational scrutiny. Because the potential vorticity in the Sverdrup regime can be evaluated from standard hydrographic data, it is fairly straightforward to prepare maps of potential vorticity for particular density surfaces, which are identified as the mid-point densities of the layers in the theory. What is a good deal more subtle is the specification of the appropriate outcrop latitude for a particular density or temperature surface. The theory outlined above is a steady-state theory that purports to deal with a conceptual time average of the circulation. Over the course of the year in actuality each outcrop line moves northward in summer as the ocean surface layer is warmed and then southward again in

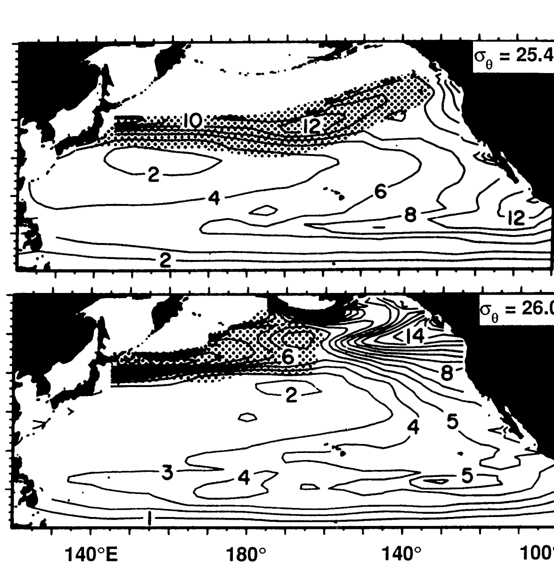

Fig. 8. Observations of potential vorticity on the density surfaces σ_θ = 25.4 and 26.0 in the North Pacific [from (*16*)].

winter. Stommel (*14*) made the ingenious observation that the position of the outcrop in late winter (that is, March) is the appropriate position to use for the steady theory rather than the annual average. His compelling argument was that fluid pumped down into the thermocline earlier than that in the year is overtaken and recaptured by the cooled, wind-stirred upper layer as it deepens during winter. The velocity with which the upper layer deepens is great enough so that only water that escapes the onrush of this "Ekman express" during the winter is free to subduct.

Various workers examining the thermocline circulation (*15–19*), have generally taken Stommel's suggestion as their starting point in preparing analyses of potential vorticity maps and relating the results to the processes of ventilation, recirculation, and homogenization. Perhaps the most detailed comparison between observation and the first-order theory was made by Talley (*16*) for the circulation of the Northern Hemisphere Pacific subtropical gyre. In spite of some important deficiencies addressed in more recent models, Talley showed that the order one theories were remarkably successful in predicting the domains of ventilation and recirculation. Important structural predictions such as the compression of the domain of recirculation with depth were also clearly observed in the data. The actual flow paths inferred from hydrographic measurements also seemed to agree, at least in broad outline, with the predictions of the theory. There are, however, some ambiguities present that tend to confuse interpretation. The observation that potential vorticity is largely homogenized on some apparently unventilated density surfaces may actually reflect a subduction process (*19*) driven by shallow convection in which the subducted water fortuitously has nearly uniform potential vorticity at the site of subduction. This is an important issue that is discussed further below. Above all, it is often difficult from superficial examination of the data to distinguish between (i) water that is ventilated in the sense that its potential vorticity is reset at subduction on each circuit and (ii) purely recirculating water that is shielded from sources of potential vorticity.

Puzzles and New Progress

The overall successes of the order one theory in delineating the various dynamical zones and the overall sweep of the great gyres' circulation do not conceal some obvious deficiencies in the predictions of the theory. Maps of potential vorticity (*16, 19*) commonly show closed contours of potential vorticity on density surfaces and extensive tongues of potential vorticity (Fig. 8). The implication is that the flow lines are crossing isolines of potential vorticity, because these patterns are inconsistent with the known direction of flow in those regions. Thus potential vorticity is not truly conserved. The lack of exact conservation of potential vorticity is not surprising; it is related to the nonadiabatic character of the flow. Buoyancy is not ignored in the purely wind-driven theories; clearly the variations in density produced by the motion acting on the imposed surface density field is essential to the theory. However, the idealization of the motion as preserving density, which makes the potential vorticity a conservative quantity, is an oversimplification. It is less easy to rectify the situation than recognize it. Although several illuminating studies have been made of the effect that deep heating or cooling would have on the thermocline motion and structure (*20–22*), in all cases, the distribution and intensity of the buoyancy forcing was largely arbitrary. Whereas Eq. 8 is a fairly robust result relating the known external forcing, τ_w, to the downwelling that drives the subtropical gyres, there is no similar trustworthy relation between observed or deduced surface oceanic heat flux and deep buoyancy forcing. A proper and useful formulation of such processes remains an unfulfilled challenge.

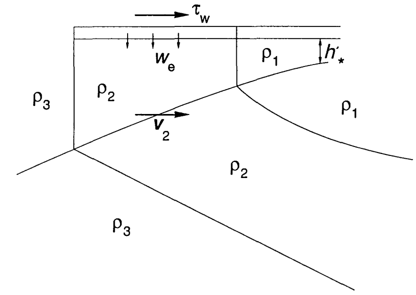

Fig. 9. The mixed layer in which convection and stirring eliminate vertical density gradients has a depth h_* which increases northward. The velocity in the mixed layer dramatically enhances the amount of fluid being subducted into the thermocline.

Some progress has been made in dealing with the effects of surface heating in the uppermost layers of the ocean. Woods (*23*) has emphasized the importance of the role of surface heat exchange in producing a mixed layer structure that is more complex than the simplistic, fixed thickness model shown in Fig. 5. Rather than capping a geostrophic region where potential vorticity is conserved with an extremely thin, constant thickness layer from which fluid is pumped directly into the thermocline at the rate given by Eq. 8, an alternative picture, shown in Fig. 9, has been proposed. Observations show that because of the unequal distribution of atmospheric cooling over the gyre the depth of the mixed layer thickens considerably from south to north. In March, during the turnaround period of the Ekman express, the mixed layer deepens from about 100 m at 20°N in the North Atlantic to over 400 m at the most northern extent of the subtropical gyre. Beneath a relatively shallow layer, the velocities in the mixed are geostrophic. The Ekman pumping acting on this layer produces a large meridional velocity by the same mechanism that leads to Eq. 12 so that the fluid subducted into the region where potential vorticity is (more nearly) conserved exceeds the fluid pumped out of the Ekman layer. This situation is consistent with tracer studies of the thermocline circulation that suggest that ventilation rates are in excess of the rate at which water is directly pumped downward from the upper Ekman layer. In this newer picture, the thermocline in large areas of the ocean is ventilated laterally by water that has its potential vorticity set by deep convection (which sets the local mixed-layer depth) before being laterally inducted into the thermocline circulation. Such lateral ventilation has important consequences for our understanding of the partitioning of water in the thermocline.

If the entire subtropical gyre is considered as a whole, the amount of water pumped downward into the gyre is of the order of $|w_e|L_yL_x$, where $|w_e|$ is a characteristic value for the Ekman pumping and L_x and L_y are the longitudinal and latitudinal extents of the gyre. In the order one model, this is the amount of fluid that has its potential vorticity reset as it is subducted. On the other hand, the volume of water moving southward across the gyre is, by Eq. 11, of the order $f/\beta\ |w_e|L_x$. The ratio of the two water masses is the recirculation index introduced by Rhines (*24*)

$$Rh = \frac{f}{\beta L_y} = \frac{R}{L_y}\tan\theta \qquad (16)$$

which is about 2 or 3. Such a value would imply that the bulk of the fluid in the gyre endlessly recirculates without readjustment of its potential vorticity, while only a small fraction of the fluid is ventilated. Yet as the observations (*16, 19*) indicate, a rather larger part of the recirculating fluid has its potential vorticity set by subduction in regions where the mixed layer is deep. The reason why Eq. 16 gives a large value for *Rh* is the gyroscopic efficiency of the circulation. A relatively weak vertical velocity can give rise, by Eq. 12, to large horizontal motions in the same manner as pumping a child's spinning top produces rapid rotation. However, in the situation described in Fig. 9, the same mechanism will act and lead to subduction into the open slots of the thermocline's intersection with the mixed layer at rates that far exceed the Ekman pumping, w_e. Huang (*25*) and Williams (*26*) have independently reconsidered the order one theory, taking into account the northward deepening of the mixed layer. Both found that a strikingly enhanced proportion of the circulation is in the ventilated part of the flow. Huang, for example, estimates that about 70% of the circulation has its potential vorticity reset by subduction, while the remainder is equally partitioned between the upper mixed layer and the deeper unventilated thermocline. Given the observations of extensive zone of nearly homogenized potential vorticity, normally taken as a sign of lack of ventilation, consistency suggests, as have both Keffer (*19*) and Talley (*16*), that diabatic effects control and render uniform the potential vorticity over much of the subduction zone. Hence, the uniformity in potential vorticity may be a misleading characterization of homogenization versus subduction and instead reflect a uniformity in the source region.

The dynamics I have described apply only to the interiors of the gyres away from western boundary currents. Yet it is clear that the fluid circulating in the gyre completes its circuit through a western boundary current where the dynamics are much different than in the interior. The role that the western boundary currents have in conditioning the interior is unclear. Although numerical models of the general circulation (*27, 28*) perforce include western boundary currents and, indeed, at least qualitatively, reproduce the elements of the theory and observations, a numerical model cannot yet be accepted as a surrogate for the natural ocean. There are theoretical suggestions (*29*) that the western boundary currents may react back onto the interior, at least in its western reaches, rather than simply passively serving as a northward conduit to close the gyre's mass flux.

Overall, a few firm steps clearly have been taken in understanding the dynamical structure of one of our planet's most fascinating natural phenomena. These are exciting days in oceanography. The hint of progress carried in the present theories serves mainly as a spur to confront the dynamical uncertainties that remain. Indeed, an extensive set of field experiments involving several oceanographic institutions is scheduled for the early 1990s (the Subduction Experiment). These experiments are designed to clarify the dynamical processes associated with subduction. One novel feature of the experiment will be an attempt to seed the ocean surface with "Bobbers." These are floats whose buoyancy is adjusted continuously so that the float cycles vertically between two preset temperature surfaces. Because the float moves horizontally with the fluid, monitoring the trajectories of a Bobber cluster should give oceanographers their first detailed picture of the kinematics of subduction. In addition, the determination of the vertical distance between temperature surfaces should give an important measure of the degree to which potential vorticity is set and conserved during and following subduction. A variety of complementary meteorological, hydrographic, and chemical tracer measurements will be taken and these data will vastly enlarge our understanding of the mechanisms involved in the ventilation process. If the history of oceanography is a reliable guide, these new data will provide provocative surprises. Oceanographers can be confident that rich dynamical challenges lie ahead.

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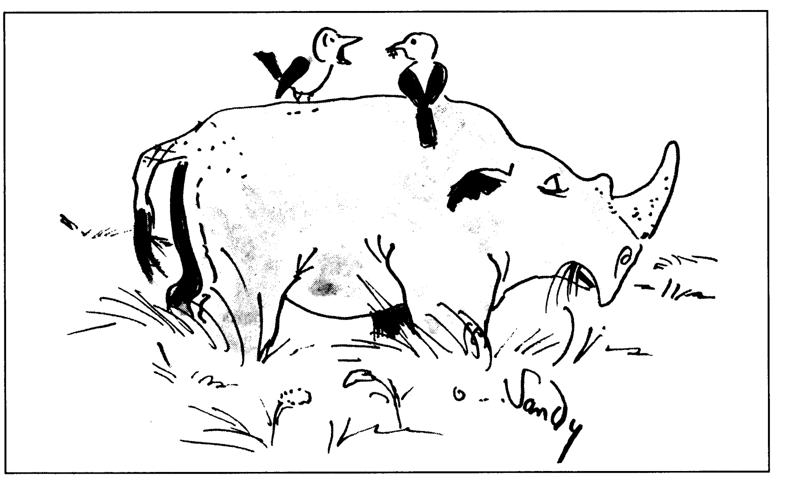

"From time to time I worry that our relationship is not* truly *symbiotic."