Fofonoff NP (1981) The Gulf Stream system. In: Warren BA and Wunsch C (eds) *Evolution of Physical Oceanography, Scientific Surveys in Honor of Henry Stommel*, pp. 112–139. Cambridge, MA: MIT Press.

Fuglister FC (1960) Atlantic Ocean atlas of temperature and salinity profiles and data from the International Geophysical Year of 1957–1958. *Woods Hole Oceanographic Institution Atlas Series*, 1, 209 pp.

Hogg NG and Johns WE (1995) Western boundary currents. *U.S. National Report to International Union of Geodesy and Geophysics 1991–1994*. Supplement to *Reviews of Geophysics*, pp. 1311–1334.

Johns WE, Shay TJ, Bane JM and Watts DR (1995) Gulf Stream structure, transport and recirculation near 68W. *Journal of Geophysical Research* 100: 817–838.

Krauss W (1996) *The Warmwatersphere of the North Atlantic Ocean*. Berlin: Gebrüder Borntraeger.

Lazier JRN and Wright DG (1993) Annual velocity variations in the Labrador Current. *Journal of Physical Oceanography* 23: 659–678.

Lozier MS, Owens WB and Curry RG (1995) The climatology of the North Atlantic. *Progress in Oceanography* 36: 1–44.

Peterson RG, Stramma L and Korturn G (1996) Early concepts and charts of ocean circulation. *Progress in Oceanography* 37: 1–15.

Reid JL (1994) On the total geostrophic circulation of the North Atlantic Ocean: flow patterns, tracers, and transports. *Progress in Oceanography* 33: 1–92.

Robinson AR (ed.) (1983) *Eddies in Marine Science*. Berlin: Springer-Verlag.

Schmitz WJ Jr. and McCartney MS (1993) On the North Atlantic circulation. *Reviews of Geophysics*, 31, 29–49.

Schmitz WJ Jr. (1996) *On the World Ocean Circulation*, vol. I: *Some Global Features/North Atlantic Circulation*. Woods Hole Oceanographic Institution Technical Report, WHOI-96-03, 141 pp.

Stommel H (1965) *The Gulf Stream, A Physical and Dynamical Description*, 2nd edn. Berkeley, CA: University of California Press.

Worthington LB (1976) On the North Atlantic circulation. *The Johns Hopkins Oceanographic Studies* **6.** Baltimore, MD: The Johns Hopkins University Press.

FLOWS IN STRAITS AND CHANNELS

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Introduction

The term sea strait is used to describe a channel connecting different water bodies. It may refer to narrow passes between an island and the mainland, such as the Strait of Messina, or to channels serving as the primary or sole connection between enclosed seas and the open ocean, such as the Strait of Gibraltar or the Bosphorus. A further distinction exists between straits that are wide relative to the internal Rossby radius of deformation such as the Denmark Strait and the Faroe Bank Channel, for which rotation plays a dominant role, and narrower straits where rotation is of lesser or negligible significance. An intermediate case is the Strait of Gibraltar, where rotation results in a tilted interface but does not dominate the exchange process. In many deep-water channels, such as the Samoan Gap, only the deep water is controlled by the topography. The upper layer in these cases is relatively passive and not dynamically linked to the deeper flow, except by virtue of the imposed density difference between the layers. However, the concept of flow control also applies to these cases and so is appropriately considered here.

Exchange through a strait may also involve movement of water in more than two distinct layers. For example, in Bab el Mandab during summer the surface layer reverses and colder, lower-salinity water intrudes at intermediate depths from the Gulf of Aden into the Red Sea, above the denser, deep-water overflow.

In general, confinement of the flow in straits and channels tends to amplify tidal and atmospherically forced currents. A crucial aspect of circulation in semienclosed seas is the exchange of water with the open ocean through the connecting strait. An important class of flows, referred to as maximal exchange flows, applies when the strait exercises control over the bidirectional transport. The distinction between maximal and submaximal flow conditions, which may change in any single strait on a seasonal or longer-term basis, provides a further characteristic with which to classify straits.

Environmental and strategic considerations have motivated widespread research of straits in recent years and there have been a number of detailed observational and theoretical studies leading to new insights on exchange dynamics, mixing, internal wave generation, implications for climate change, and other aspects. Many estuaries are connected to the ocean through narrow channels and are subject to similar exchange flow constraints. Flow characteristics in these smaller channels have served as excellent surrogates for the study of processes occurring at larger scales.

Topographic details can have a profound effect on the exchange. Many straits have a shallower sill in addition to being narrower than the adjacent water bodies. The location of the shallowest section relative to the narrowest portion of the strait can determine whether or not variability in the stratification of adjoining water bodies can propagate into the strait and thus affect the rate at which exchange takes place.

History

It has long been recognized that straits are often characterized by persistent flows in one direction or another, and these have motivated scientific investigation and analysis. Two examples for which scientific comment dates back many centuries are the Straits of Gibraltar and the Bosphorus or Strait of Istanbul. It is a striking fact that both of these straits exhibit a persistent flow inward to the Mediterranean. Attempts to explain this apparent violation of continuity included the proposed existence of underground channels through which the surplus drained into the interior of the earth. It was recognized that evaporation played a role, although evaporation alone removed insufficient water to account for the exchange. Towards the end of the seventeenth century, Luigi Marsigli carried out a beautiful laboratory experiment with salt and fresh water that exchanged as density currents past a barrier. This demonstrated the mechanism of bidirectional flow, consistent with Marsigli's own observations in the Bosphorus. In 1755 Waitz explained a similar exchange flow in the Strait of Gibraltar, subsequently confirmed by drogue measurements, thus accounting for the large surface flow into the Mediterranean.

The fact that water moves in opposite directions at different depths through the Strait raises the question of possible controls on the rate of exchange. This is a problem of quite general significance in fluid mechanics, having application, for example, to calculation of the temperature of thoroughly mixed air in a heated room communicating through an open door with outside air of lower temperature. Stommel was one of the first to develop a modern explanation of internal exchange controls and in the 1950s described the situation in which thoroughly mixed water in an estuary was exchanged with the open ocean through a strait. The estuary was said to be 'overmixed' in the sense that the bidirectional exchange had achieved a maximum rate for the given density difference and channel geometry owing to the presence of internal flow control. For a given influx of fresh water, the salinity of the estuary was then determined. This flow state is now recognized as an important special case of bidirectional exchange and may occur in any strait that is not so wide as to be dominated by Coriolis effects, provided a suitable set of bounding conditions exist at either end.

Two-layer Exchange Flows in Straits

In the simplest case, where rotation is negligible and the strait is short enough for mixing and friction to be unimportant, the two-layer exchange flow can be analyzed in terms of the frictionless theory of layered flow. We refer to the flow as being controlled when the combined Froude number G^2 is unity:

$$G^2 = F_1^2 + F_2^2 = 1 \qquad [1]$$

where

$$F_i^2 = \frac{u_i^2}{g' y_i} \qquad [2]$$

u_i is flow speed and y_i is the thickness of the upper ($i = 1$) and lower ($i = 2$) layer respectively, and

$$g' = g\frac{\rho_2 - \rho_1}{\rho_2} \qquad [3]$$

is the reduced gravity due to the differences in density ρ_i of each layer. F_i^2 is referred to as the layer Froude number. When the flow is controlled, adjustments in the depth of the interface can travel in one direction, but not in the other. Thus an internal control acts as an information gate that blocks adjustments of interface depth, in the form of long internal waves, from propagating against the flow. The control condition [1] corresponds to the point at which such a wave is just arrested and it always separates subcritical from supercritical flow. Subcritical flow occurs when the combined Froude number is less than unity such that waves can travel in both directions. Supercritical flow occurs when the Froude number is greater than unity such that waves can only travel downstream. Steady-state solutions to the frictionless exchange flow can be found from the Bernoulli and continuity equations and have been discussed in detail in the literature.

Maximal exchange flow constitutes the steady-state limit when water of differing densities is free to move in opposite directions. In general, maximal flow requires two separate locations where eqn [1] is satisfied. Where there is both a sill and a contraction, the relative locations of these topographic features are important in determining whether or not the reservoir conditions, specifically the interface depths in each adjoining water body, can influence stratification within the controlling region, thus eliminating at least one control and preventing maximal exchange from taking place. If the narrowest section is located toward the end of the strait bounded by denser water and the shallowest section is toward the end bounded by less-dense water and if controls [1] occur at both of these locations, the requirements for maximal exchange are met. Additional controls may also occur, for example, in the form of a sequence of deeper sills downstream of the first, but these have no direct influence on the exchange rate.

An internal control separates supercritical from subcritical flow. The depth of the interface between the two layers is asymmetrical about a control position (i.e., the interface depth increases or decreases continuously as one moves through the control) and thus easily recognized in oceanographic data. Although the control acts on both layers, in the more general case of a strait containing both a sill and contraction, a sill acts primarily on the lower layer, whereas the contraction acts primarily on the upper layer. In the case of exchange flows through a strait, maximal exchange requires that the supercritical flow be directed away from the control and towards the nearest reservoir; that is, the two control locations for which eqn [1] is satisfied are separated by flow that is subcritical ($G^2 < 1$). The presence of supercritical conditions on either side of a subcritical interior portion, the 'control section' lying between the two controls, ensures that adjustments in the level of the interface in the adjoining water bodies cannot propagate into the strait and thereby influence the exchange, although adjustments in interface depths at one control can communicate through the subcritical part of the flow with the other control. The exchange within a strait is therefore determined entirely by the local geometry and the densities of the exchanging water masses and is therefore maximal. The bounding supercritical states isolate the control from the adjacent seas. Of course, the supercritical flow must always match the subcritical conditions in each reservoir far away from the strait and generally does so through an internal jump. In the special case of a simple contraction with no sill and in the absence of barotropic forcing, such as that due to the tide, fresh water discharge or atmospheric pressure differences, the two controls can be thought of as having coalesced, so that the subcritical portion vanishes and the location at which eqn [1] is satisfied separates two supercritical flows.

The above description covers maximal exchange. If the interface (or usually the thermocline) is sufficiently shallow in the less dense reservoir (i.e., the Atlantic Ocean in the case of the Strait of Gibraltar) or sufficiently deep in the denser reservoir (the Alboran Sea in the case of Gibraltar) to prevent formation of a supercritical flow, the nearest control location is said to be 'flooded' and flow in this portion of the strait is subcritical. The exchange is then subject only to a single control, for example, at the sill if the contraction is flooded, and is therefore submaximal. In this case the exchange rate is a function not only of the densities of the exchanging water masses but also of the stratification depth in the reservoir adjacent to the flooded contraction. Both maximal and submaximal conditions have been observed in the Strait of Gibraltar. However, the frequency with which transitions take place between one state and the other, which has important implications for the interpretation of longer-term variability in the Mediterranean, remains to be established.

Barotropic Forcing of Exchange Flows

Natural flows in straits are rarely steady: barotropic forcing by tides, meteorological effects and changes in the stratification at either end may modify the rate of exchange. Quasi-steady solutions of the exchange equations may still be valid, provided the influence of the forcing is properly accommodated. Steady solutions for maximal exchange in the presence of a sill and a contraction remain valid if interfacial depth adjustments at one control communicate through the subcritical portion of the strait with the other control, subject to a delay that is short relative to the time scale of forcing. In the Strait of Gibraltar this condition is met approximately for tidal forcing.

Barotropic forcing modifies the exchange within the maximal state, but may also be strong enough to arrest one of the layers, thus inhibiting the bidirectional flow altogether and temporarily overiding the internal control. This situation is common in many coastal environments with large tidal currents. In the quasi-steady case the effect of barotropic forcing can be investigated by appropriate modification of the exchange equations to accommodate relative differences in the transport of each layer. The resulting exchange falls naturally into three regimes (**Figure 1**). In the 'moderate' regime both layers continue to exchange, but with an adjustment of interface depth to maintain control at the sill and the contraction. If the forcing is from the less-dense towards the denser reservoir, the interface drops; for forcing in the other direction, it rises. If the forcing is great enough to cause the interface to intersect the seafloor or the surface, one of the layers is arrested.

'Strong' forcing occurs when the barotropic pressure gradient from one reservoir to the other is great enough to push the interface downstream of the sill. In this case control over the sill crest is lost, leaving just a single layer over the crest, which is therefore uncontrolled (see **Figure 1A** and **C**). The steeply inclined stratification intersects the surface in a front, a phenomenon commonly observed in coastal waters. The front is characterized by sharp changes in water properties and may entrain bubbles to considerable depth. In general, tidal forcing tends to increase the overall exchange through the strait beyond the maximal exchange rate in the absence of tides. This is particularly true in straits such as Gibraltar, where the dependence of flow rate on tidal modulation is strongly nonlinear.

The dimensional barotropic transport per unit width is defined as

$$Q = u_{1s}y_{1s} + u_{2s}y_{2s} \qquad [4]$$

where y_{is} is the dimensional thickness of each layer $(i = 1, 2)$ and u_{1s}, u_{2s} are the corresponding flow speeds; the subscript 's' refers to values above the sill crest (see **Figure 1**). Considering the case illustrated in **Figure 1D** ($Q > 0$), the barotropic transport for which two-layer exchange is lost and the interface intersects the surface is then shown to be

$$Q = (2/3)^{3/2} h_s \sqrt{g' h_s} \qquad [5]$$

where $h_s = y_{1s} + y_{2s}$. The corresponding limit for flow in the other direction (in **Figure 1A**) is

$$Q = -h_s \sqrt{g' h_s} \qquad [6]$$

for which the interface intersects the sill crest and the dense water layer is arrested. At still stronger forcing in this direction, the point of intersection moves down the lee face of the sill (**Figure 1G**).

The unsteady character of tidal forcing in straits where the flow is controlled can have a further effect leading to strong surface signatures in remotely sensed images. The release of potential energy stored in the deformation of the interface downstream of the control as the tide slackens can generate large-amplitude nonlinear internal waves. These propagate away from the sill in the form of an undular bore. In the Strait of Gibraltar, for

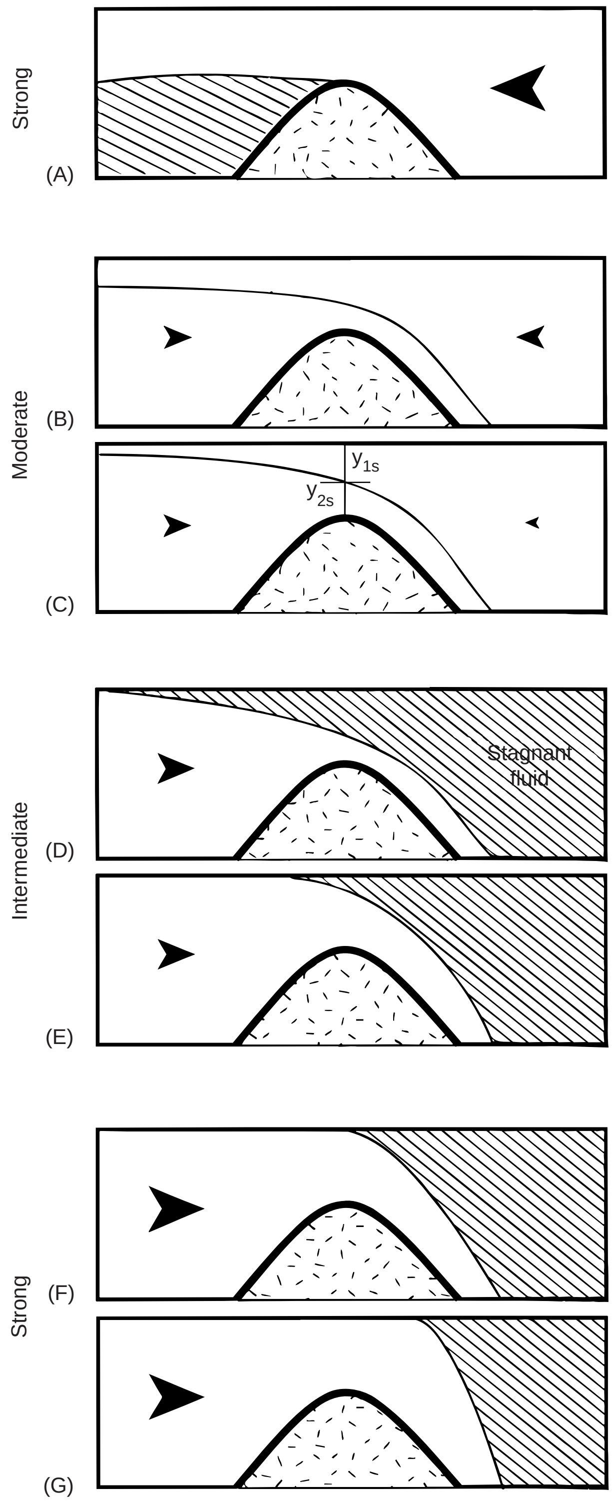

Figure 1 Quasi-steady response of a two-layer exchange flow through a strait subject to barotropic forcing. With strong enough forcing from the less dense reservoir on the right, the denser layer is arrested and the interface intersects the crest of the sill (A). A similar effect occurs with strong forcing from the dense reservoir on the left, in which case the interface intersects the surface (D) and may be pushed downstream of the sill (G). (Adapted from Farmer & Armi 1986.)

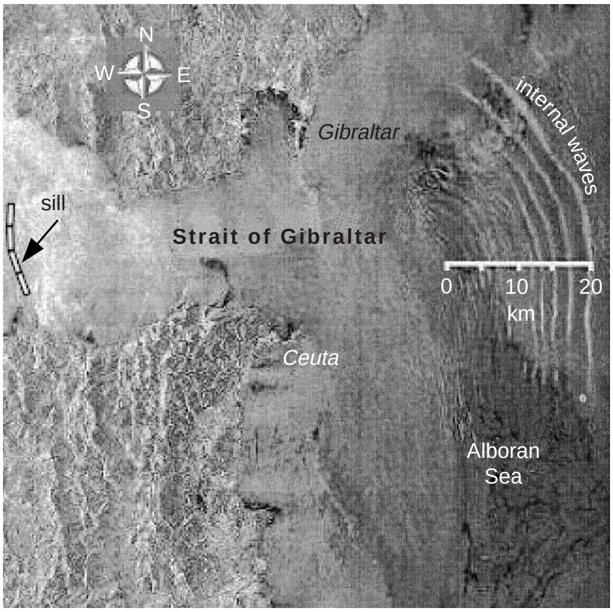

Figure 2 An ERS-1 synthetic aperture radar (SAR) image, showing internal waves at 22.39 UT, January 1, 1993, formed over the sill of the Strait of Gibraltar, radiating eastward into the Alboran Sea. (Photograph: European Space Agency.) The strait is approximately 27 km wide at the sill and 14 km wide in the narrowest section to the east of the sill.

example, they are generated toward the end of the ebb flow as the internal hydraulic control over the sill is lost. They are observed to travel east along the strait and into the Alboran Sea where they spread radially before dissipating (**Figure 2**).

Mixing in Straits and over Sills

Exchange flows in straits experience enhanced shear between the exchanging layers. Under certain circumstances this leads to instability and mixing. In longer straits, such as the Bosphorus, mixing can produce significant changes in the layer densities. Combined with frictional effects, this results in an internal response that differs from the frictionless results discussed above. In contrast to the frictionless prediction, there can be a marked slope within the subcritical portion of the flow and the exit control of the upper layer is displaced downstream with respect to the active layer.

Mixing can also result from tidal effects. Small-scale processes leading to mixing have been examined in detail over some sills where they are also seen to play a role in the establishment of controlled flow. A particularly well-studied example is Knight Inlet, British Columbia, where instabilities form on the interface over the sill crest (**Figure 3**).

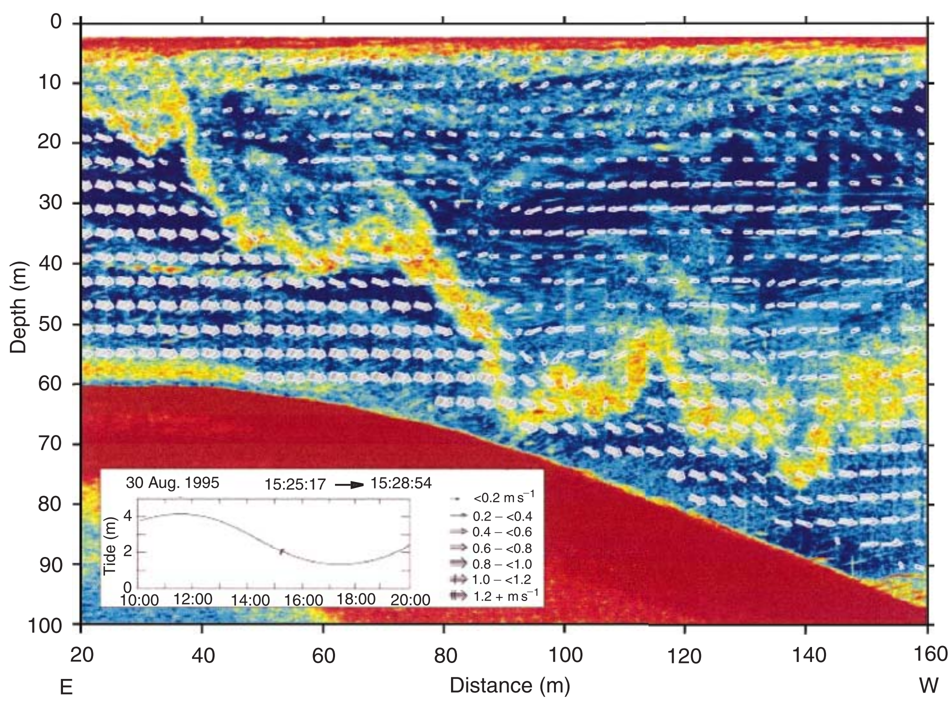

Figure 3 Instabilities on the sheared interface of controlled flow over the sill in Knight Inlet. The instabilities are asymmetrical, leading to injection of water from the supercritical lower layer into the slowly moving upper layer. The flow state corresponds to the bottom illustration in **Figure 1**. The tidal current is from left to right, with arrows indicating current vectors; a weak recirculation exists within the upper layer. The inset indicates the phase of the tide at the time of measurement. Adapted from Farmer DM and Armi L (1999) *Proceedings of the Royal Society*, *Series A* 445: 3221–3258.

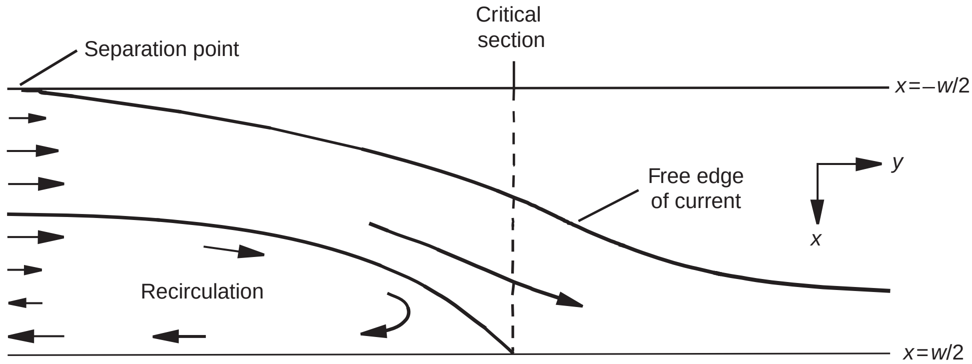

Figure 4 Plan view of exchange flow in a rotating strait, showing separated single-layer flow at the critical section of a strait, with a corresponding stagnation point on the opposite wall. Recirculation occurs upstream. Adapted with permission from Pratt and Lundberg (1991).

Acceleration of the flow in the supercritical layer downstream of the crest creates an asymmetric instability that ejects fluid upward from the deeper layer. This in turn forms an intermediate layer of weakly mixed fluid that fills in as the downslope flow becomes established. When the tidal current slackens, the mixed layer can intrude upstream just beneath the fresher surface layer. While Knight Inlet is perhaps unique in the extent to which it has been studied, similar processes can be expected wherever stratified flow occurs over abrupt topography such as commonly found in straits.

Effects Due to Rotation

The hydraulics of flow with uniform potential vorticity are very similar to classical hydraulics, provided there is no separation from the sidewalls. Scale analysis based on the geostrophic relation suggests that separation occurs if the channel width is greater than the distance L,

$$L \sim \frac{2(g'\bar{Y})^{1/2}f^{-1}}{\bar{U}/(g'\bar{Y})^{1/2}} \qquad [7]$$

where f is the Coriolis parameter and $\bar{Y}$ and $\bar{U}$ are representative depth and velocity scales within the strait. In the absence of separation, maximal exchange in a rotating channel flow can still occur, with the control being exercised through a Kelvin wave. The situation becomes more complicated, however, when the width is sufficient for the flow to separate. Control is then exercised by a frontal wave with strong cross-stream velocities.

As with all hydraulic approaches, irrespective of rotation, the calculation involves integration over layers, for example, from the seafloor to the interface and from the interface to the surface in a two-layer flow. For rotational flows, once the potential vorticity is assumed, integration can be carried out and the cross-stream structure of the flow is fully determined. It has been shown that when the flow is critical with respect to the frontal wave, a stagnation point occurs on the right sidewall (in the northern hemisphere), independent of the potential-vorticity distribution. The two-layer portion of the flow crosses over the strait in a distance of order the internal Rossby radius; under certain conditions recirculation may occur as shown in **Figure 4**. Time dependent adjustment of strait flows where rotation is important gives rise to internal bores or shock waves that have a two-dimensional structure. Even where rotation is of minor importance, transverse variability and flow separation can also occur owing to abrupt changes in channel width or direction. This is observed, for example, in the Bosphorus, where the channel geometry leads to marked transverse variability.

See also

Estuarine Circulation. Overflows and Cascades.

Further Reading

Armi L (1986) The hydraulics of two flowing layers with different densities. *Journal of Fluid Mechanics* 163: 27–58.

Armi L and Farmer DM (1988) The flow of Atlantic water through the Strait of Gibraltar. *Progress in Oceanography* 21(1): 1–105.

Assaf G and Hecht A (1974) Sea straits: a dynamical model. *Deep-Sea Research* 21: 947–958.
Briand F (ed.) (1996) Dynamics of Mediterranean straits and channels. *Bulletin de l'Institut océanographicque, numéro spécial 17*. Monaco: Musee océanographique.
Deacon M (1985) An early theory of ocean circulation: J. S. Von Waitz and his exploration of the currents in the Strait of Gibraltar. *Progress in Oceanography* 14: 89–101.
Farmer DM and Armi L (1986) Maximal two-layer exchange over a sill and through the combination of a sill and contraction with barotropic flow. *Journal of Fluid Mechanics* 164: 53–76.
Farmer DM and Armi L (1988) The flow of Mediterranean water through the Strait of Gibraltar. *Progress in Oceanography* 21(1): 1–105.
Murray SP and Johns W (1997) Direct observations of seasonal exchange through the Bab el Mandab Strait. *Geophysics Research Letters* 24: 2557–2560.
Pratt LJ (ed.) (1990) *The Physical Oceanography of Sea Straits*, NATO ASI Series, Series C: Mathematical and Physical Sciences, vol. 318.
Pratt LJ and Lundberg PA (1991) Hydraulics of rotating strait and sill flow. *Annual Review of Fluid Mechanics* 23: 81–106.
Stommel H and Farmer HG (1953) Control of salinity in an estuary by a transition. *Journal of Marine Research* 12: 13–20.

FLUID DYNAMICS, INTRODUCTION AND LABORATORY EXPERIMENTS

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Introduction

Laboratory experiments have provided considerable insight and quantitative information about many of the physical processes which affect the fluid ocean. Although often made with the purpose of investigating some fundamental process in fluid dynamics, motivation for making laboratory experiments frequently comes directly from a need to improve understanding of processes in the oceans, or in some other geophysical fluid such as the fluid interior or atmosphere of the Earth and other planets, and such studies consequently belong to the broad field of geophysical fluid dynamics. Laboratory experiments are particularly valuable in testing theory and in providing quantitative, if empirical, estimates of, for example, constants of proportionality which cannot presently be determined by theory or numerical computations. They are therefore an essential component of geophysical fluid dynamics in relating theory to reliable application.

Laboratory experiments as a means of illuminating oceanographic processes have a long history which can be traced back at least as far as the experiment by Marsigli, reported in 1681, which demonstrated the way in which density differences drive exchange flows in the Bosphorus between the Mediterranean and the less dense Black Sea. The purpose of making laboratory experiments is rarely, however, to reproduce some aspect of ocean circulation. More often it is to study a particular process in isolation from others which occur in the natural environment. In addition to density differences or stratification, laboratory studies have been made of processes which result, for example, as a consequence of the Earth's rotation (including the β effect) and from the effect of free or fixed boundaries (e.g., promotion of turbulence or waves).

Several general objectives in making laboratory experiments may be identified and some are briefly described in the following sections. The particular experiments mentioned as examples are perhaps not always the best which might be chosen, but they are ones (among many) which demonstrate some particular value of making laboratory studies.

Testing Predictions

A beautiful example is the study made by Mowbray and Rarity of internal gravity wave propagation in a tank filled with salt-stratified water. Waves were generated by the slow oscillation of a horizontal cylinder and made visible using an optical Schlieren system. The experiments demonstrated beautifully the theoretical prediction that internal waves propagate away from the cylinder in a vertical plane at an angle to the horizontal given by $\sin^{-1}(\sigma/N)$, where σ is the frequency of the cylinder and N the buoyancy frequency in the stratified fluid (*see* **Internal Waves**).