# ON STRUCTURE, ENTRAINMENT, AND TRANSPORT IN ESTUARINE EMBAYMENTS

BY

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## ABSTRACT

When fresh water enters the surface of an embayment it moves persistently seaward. Enroute it entrains seawater from below to form a halocline, in which the salinity increases with depth and to seaward. Wind mixing creates a nearly homogeneous zone in the upper part of the halocline. Below the halocline there is a nearly homogeneous lower zone in which sea water intrudes at a rate sufficient to supply the demand for entrainment. This fresh water transport system is superimposed on the tides which ebb and flood at all depths. There is a surface of no net motion in the halocline above which the net transport is seaward and below which it is inwards.

Entrainment is the preferential upward transfer of water from the lower to upper zone. The downward mixing of fresh water is defined by the lower limit of the halocline, which is a surface of unidirectional upward transfer.

The zone structure is accurately defined by plotting salinity as a function of the logarithm of depth.

The seaward volume transport through the upper zone and halocline, the inward transport through the lower zone, and the upward transfer through the interzone boundary may be computed in terms of the continuity of fresh water, where the rate of input is known. Where it is not known, the equations are solved by introducing the difference of dynamic height across the embayment. The features of structure and transport may be defined by application of normal oceanographic data.

This paper is presented in recognition of the inspiration and leadership of Thomas G. Thompson.

## THE ESTUARY MODEL

In an estuarine embayment the input of fresh water exceeds the losses by evaporation and/or freezing.

From his study of Alberni Inlet (Fig. 1), Tully (1949) concluded that fresh water entering the surface of such an embayment moves persistently seaward. Enroute it entrains sea water from below to form a halocline in which the salinity increases with depth and to seaward. Usually the upper part of the halocline, mixed to or nearly to homogeneity by wind, is called the upper zone. Below the halocline there is a nearly homogeneous lower zone in which sea water

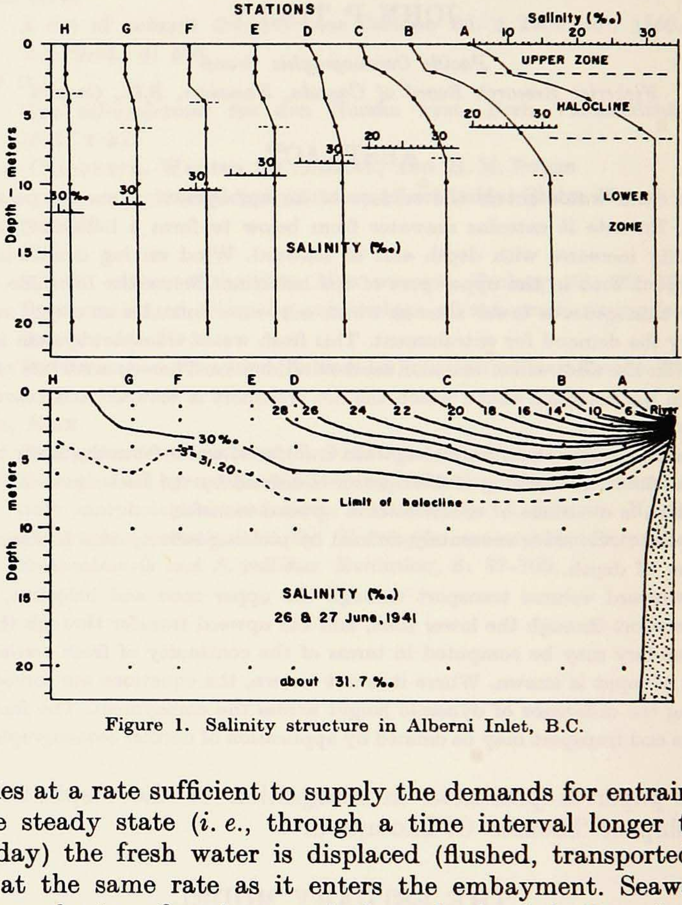


Figure 1. Salinity structure in Alberni Inlet, B.C.

intrudes at a rate sufficient to supply the demands for entrainment. In the steady state (*i. e.*, through a time interval longer than a tidal day) the fresh water is displaced (flushed, transported) seaward at the same rate as it enters the embayment. Seawater is continuously transferred upward through the halocline along the path of transport. It is mixed with the fresh water and conserved in the upper zone and in the halocline, and it moves seaward at the same rate as the fresh water.

It is of some interest to re-examine the features of this model in the light of more recent knowledge.

*Fresh Water Transport Mechanism.* Such a transport mechanism can only recognize the difference in density between the zones. It requires a continuous source of light water (fresh water from precipitation and land drainage), a continuous source of dense water (the ocean), and a sink for the mixed light and dense water (the ocean).

The greatest concentration of light (fresh) water is at the head of the system. Hence there is a gradient of increasing mean density seaward. This is compensated by a seaward slope of the sea surface which is the seaward driving force in the upper zone. The driving force in the lower zone is provided by an opposite slope of the boundary between the zones, downwards towards the head of the system, and by the difference of density between the zones.

It is convenient to retain the terms fresh and saline waters because they indicate continuity with their sources. They can be used interchangeably with light and dense in the descriptions. Furthermore, in most small embayments the heat exchange with the atmosphere is uniform over the area, hence the temperature-induced density structure contributes only to the stability, not to the transport mechanism.

The fresh water transport mechanism persists while there is a difference in density between the zones. The entrainment process continually reduces this difference. It may be further reduced by tidal or wind mixing or by loss of fresh water by evaporation or freezing. If it vanishes, the mechanism fails and the waters become homogeneous. Then some of the homogeneous water may be returned to the lower zone of the embayment. Such homogeneity does not normally occur, particularly in embayments where there is no region of complete mixing. Usually there is additional inflow of fresh water along the route of transport.

*Tidal Flushing.* Tidal currents flood and ebb at all depths to transfer the water necessary for the tidal rise and fall. Within the limits of constancy of mean sea level, the flood and ebb excursions are compensating. There is no net transport of sea water. This does not preclude tidal circulation or tidal flushing, since it is not necessary that the water discharged through a cross-section during the ebb excursion be the same water that entered during the previous flood. The inflow and outflow are usually asymmetric. Hence some part of the tidal volume is renewed during each cycle.

*Estuary Mechanism.* Usually there are precipitation and land drainage into a marine embayment. Thus there is zone structure, so that both transport and tidal flushing mechanisms occur. In the upper zone the ebb is accelerated and the flood is retarded by the persistent seaward transport. Conversely, the flood is strongest and the ebb is weakest in the lower zone, because of the persistent inward transport. Consequently there is a surface of *No net motion* above which the net transport is seaward and below which it is inward (Pritchard, 1952).

In such cases it is difficult to distinguish between the purely tidal flushing and the transport mechanism. It is convenient to suppose that the tidal excursions do not contribute to the transport, but rather that it is due solely to the fresh water mechanism, that it occurs as a bias on the tidal excursions, and that it is intermittent with the reversing character of the tides.

*The Halocline and Entrainment.* Turbulence is a consequence of the velocity shear between the upper seaward moving zone and the lower inward moving zone. Turbulence may be regarded as the random exchange of fluid elements. When fluid elements are exchanged vertically in a stratified fluid their displacement is opposed by the stability

$$g\frac{\partial \varrho}{\partial z},$$

where $g$ is gravity, $\varrho$ is density, and $z$ is depth (positive downwards). This acts as a restoring force. The elements tend to return to their original strata. However, having mixed to some extent with surroundings, the densities (salinities) are changed toward intermediate values. Hence they cannot be stable again in their original positions. In effect they must undergo a series of damped oscillations and come to rest in intermediate positions. Thus a halocline is created, in which there is a gradient of salinity from the lower to upper zone.

Entrainment is the preferential upward transfer of lower zone water through the halocline.

It is recognized that fresh water is transferred downward from the upper zone into the halocline. However, if it were transferred below the limit of the halocline, there would be a longitudinal gradient of decreasing salinity toward the head of the embayment in the lower zone. This is not observed. Thus the lower limit of the halocline must be the limit of downward transfer of fresh or mixed water. On the other hand, it is observed that the upper zone and halocline become progressively more saline to seaward. Hence there must be continual upward transfer of sea water through the lower limit of the halocline. It follows that at this boundary the vertical transfer must be unidirectional, upward.

Above the surface of no net motion the fresh water and entrained sea water are transported persistantly seaward and are eventually lost from the embayment. Fresh water mixed downward to or below this surface must be retarded or transported toward the head of the system. Since it cannot escape from the embayment in these positions and cannot be dissipated in the lower zone, it must remain between the level of no net motion and the limit of the halocline and must be preferentially available for entrainment. It follows that the level of no net motion must be in the halocline.

Thus the necessary entrainment and transition of properties from the lower to the upper zone are achieved. The halocline is the zone of entrainment. The presence of a halocline is sufficient proof that entrainment rather than any other process of upward transfer, such as upwelling (Sverdrup, *et al.*, 1942), is occurring.

*Energy of Turbulence.* Entrainment depends primarily on the occurrence of turbulent mixing. This can persist only if there is a continuing source of energy. Normally, energy is obtained from the mean flow by the action of turbulent stresses. The intensity of turbulence varies with the energy available in the mean flow, which in turn depends on potential energy or head in the upper and lower zones. La Croix and Tully (1954), in their study of Seymour Narrows, B.C., showed that about 50 $^0/_0$ of the available potential energy of the tides was dissipated in turbulence. Evidently, at least part of the energy required for turbulent mixing originates with the tides. This is effective throughout the depth of tidal action, which is usually the whole depth in regions over a continental shelf.

The wind must also be recognized as a source of energy which can lead to turbulent mixing. Eckman (Sverdrup, *et al.*, 1942) showed that the depth ($D$) of wind influence is

$$D = \frac{6.7\,W}{\sqrt{\sin \Phi}}$$

where $W$ is the wind speed (m/sec) and $\Phi$ is the latitude. Within this depth the wind energy is added to the tidal energy, creating further turbulence. All evidence shows that the depth ($D$) of wind mixing coincides closely with the depth of the nearly homogeneous upper zone. A gradient develops in this layer when the winds are small (Fig. 1, St. A). In the limit, the halocline becomes continuous from the lower zone to the surface.

It is at once evident that the lower limit of the halocline must be determined by the mixing forces. If the depth of the estuary is greater than this limit, the halocline and lower zone waters must be distinct in the structure. This is the case in the deep British Columbia inlets observed by Tully and others (Fig. 1). If, however, the depth of the estuary is less than this limit, the halocline structure must reach the bottom. That is, there must be more or less uniform salinity gradient to the bottom. This was observed by Pritchard (1952) and others studying the plains estuaries on the Atlantic coast of United States.

## STRUCTURE

Utilization of these concepts depends on definition of the zones, which in turn requires an accurate description of the structure. This may be accomplished by the method outlined by Tully (1952, 1957). The observed salinity data are plotted as functions of the logarithm of depth (Fig. 2). In this presentation the zones are defined, within small limits of error, by straight lines of constant slope ($k$) such that

$$S = k \ln \frac{z}{z_1} + S_1,$$

where $z$ is depth (positive downwards), $z_1$ is unit depth, and $S_1$ is a constant.

The first diagram in Fig. 3 shows such a plot of the salinity data from Alberni Inlet, corresponding to Fig. 1. The subsequent diagrams show salinity data from Juan de Fuca Strait.

Examination of more than 1000 examples from regions dominated by fresh water showed that this logarithmic plot could be applied with an apparent error of less than 0.1 $^0/_{00}$ in the halocline and of less than 0.05 $^0/_{00}$ in the lower zone in 90 $^0/_0$ of the cases (Tully, 1957).

The limits of the zones are defined by the intersections of the lines. According to this interpretation of structure there should be discontinuities at the upper and lower limits of the halocline ($D$ and $L$). These have not been observed and there is some doubt that they exist. This uncertainty cannot be resolved until adequate instrumentation is available to examine the salinity structure in detail. In the meantime these limits are regarded as "virtual" values.

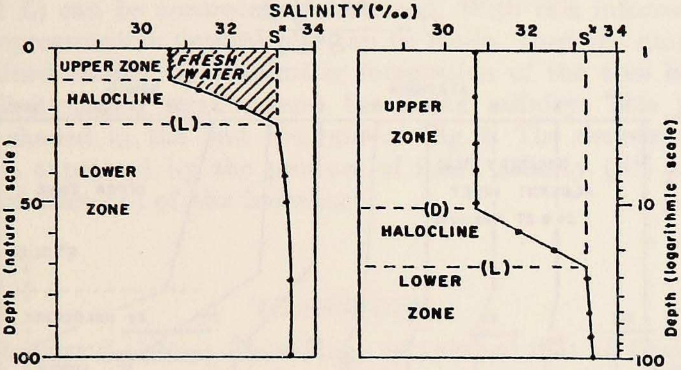


Figure 2. Comparison of natural and logarithmic plots.

The mixing mechanism which creates this structure has not been demonstrated. However, it is intuitively apparent that the mechanism is constant in each zone of constant slope ($k$). In the halocline this must be the entrainment process.

*Index Salinity.* The salinity at the lower limit ($L$) of the halocline defines the water which is about to be entrained. It is called the index salinity ($S^*$). This has been found to be remarkably constant in any region where the halocline was continuous and where the lower zone water did not change.

Herlinveaux (1959, in preparation), studying Juan de Fuca Strait, and Waldichuk (1957), studying the Strait of Georgia, found that it followed an annual cycle associated with the seasonal variation of sea water intruding the lower zone. Tully and Dodimead (1957) and Dodimead (1958), working in the eastern sub-Arctic Pacific Ocean, found it to be constant ($33.8 \pm 0.1\ ^0/_{00}$) both winter and summer during four years (1955–58), despite major changes in the upper zone. From examination of these data it became evident that index salinity is determined solely by the properties and processes in the lower zone. It is independent of the character of the upper zone. This confirms the conclusion that the lower limit ($L$) of the halocline is a surface of unidirectional upward transfer.

*Determination of the Fresh Water Fraction* ($C$). From the foregoing discussions, the concentration ($C$) of fresh water at any position in the upper zone and halocline is

$$C = \frac{S^*L - \int_0^L s dz}{S^*L}.$$

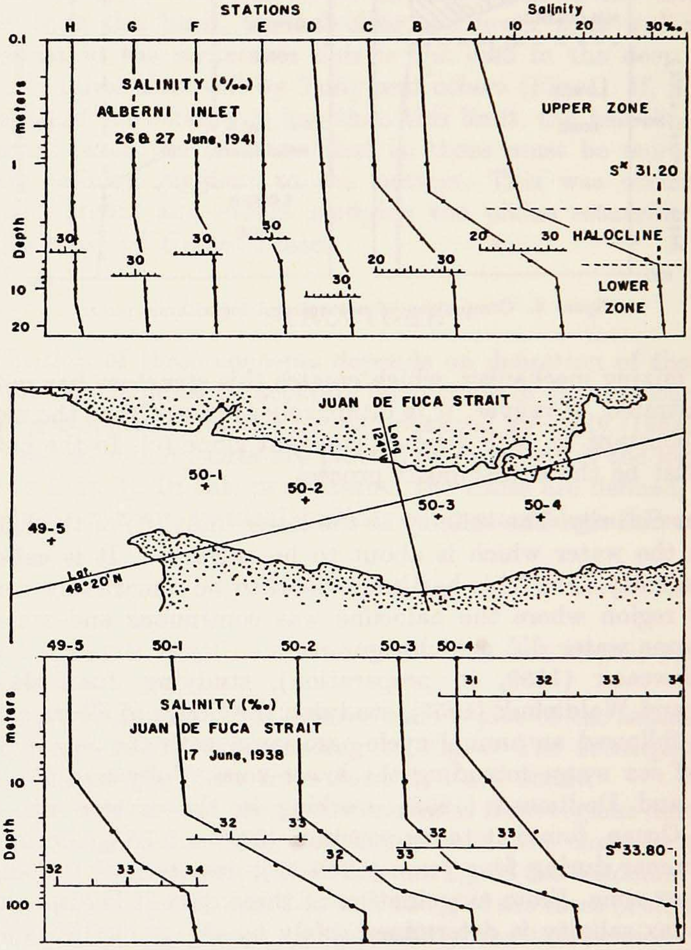


Figure 3. Logarithmic plots of salinity structure in Alberni Inlet and Juan de Fuca Strait, B.C.

The relation is difficult to solve in a natural plot because the structure is often irregular and requires subjective interpretation. Also, the lower limit of the halocline and the index salinity are indefinite. It is more precise and convenient to first plot the salinity data as a function of the logarithm of depth, as has been demonstrated. Here the index salinity ($S^*$) and the limits of the halocline ($D$ and $L$) can be accurately determined. With this information a truly representative natural plot can be made. Then the numerator is obtained readily by planimeter integration of the area between the salinity-depth relation and the index salinity. This area is shown shaded in the first diagram of Fig. 2. The denominator is the area expressed by the product of index salinity ($S^*$) and the depth of limit ($L$) of the halocline.

## TRANSPORT

*Fresh water Equations.* It has been established that all fresh water together with the entrained sea water is transported seaward through the upper zone and halocline. Then to maintain continuity the fresh water fraction ($C$) of the total volume transport through any cross-section of the upper zone and halocline ($T'$) must equal the rate of inflow of fresh water ($Q$) from the head to that section. Taking the positive direction seaward

$$CT' = Q, \tag{1}$$

whence the total transport through the upper zone and halocline is

$$T' = \frac{Q}{C} \tag{2}$$

in units of volume per unit time. Because of the diurnal cycle of the tides it is implied that the unit of time is the tidal day (24 hours, 52.2 minutes) regardless of the units used in calculation.

It is also evident that the seaward transport of the sea water fraction $(1 - C)$ of the upper zone and halocline must equal the upward transport $(-W)$ through the interzone boundary in the area between the section and the head of the embayment. This in turn must equal the inward transport through the cross-section in the lower zone ($T''$), else sea water would accumulate or decrease in the embayment

$$-W = T'' = -(1-C)T' = \frac{1-C}{C}Q. \tag{3}$$

It is stipulated in these equations that the zones are defined in the structure (Fig. 2) and that transport is primarily dependent on the fresh water inflow ($Q$). This is the total of precipitation, river discharge, and seepage, inwards of the cross-section being considered. These quantities vary considerably and are difficult or impossible to measure accurately. Hence it is necessary to introduce another concept to solve the equations.

*Transverse Forces.* During the flooding tide, the maximum velocity is in the lower zone. In the northern hemisphere these waters must tend to the right more than those in the upper zone. Hence the greatest depth of lower zone (dense) water must be on the right of the inward flow. Conversely, during the ebbing tide the greatest velocity is in the upper zone. These waters must tend to the right more than those in the lower zone. Thus the greatest depth of upper zone (light) water must be on the right of the seaward flow.

In a situation where the flood and ebb flows are opposite, there must be a cross-channel slope of the isopycnal surfaces, representing the gravity forces required to balance the Coriolis forces which are associated with the difference of velocity between the upper and lower zones. This is the case in Juan de Fuca Strait, illustrated in Fig. 4. Then

$$\frac{d_2 - d_1}{\Delta y f} = \bar{V}' - \bar{V}'', \tag{4}$$

where $d_1$ and $d_2$ are the anomalies of dynamic height across the channel, $\Delta y$ is the width, $f$ is the Coriolis parameter ($2\omega \sin \Phi$), and $\bar{V}'$ and $\bar{V}''$ are the mean velocities in upper and lower zones.

The velocity difference can be related to the transports in

$$\bar{V}' - \bar{V}'' = \frac{T'}{A'}\,\frac{T''}{A''}, \tag{5}$$

where $A'$ and $A''$ are the cross-sectional areas of the upper and lower zones which can be defined from the structure.

*Transport.* Combining the concepts it is possible to evaluate the transport.

Substituting for $T''$ from eq. (3),

$$\frac{d_2-d_1}{\Delta yf}=\frac{T'}{A'}+\frac{1-C}{A''}T',$$

whence the seaward transport in the upper zone is

$$T'=\frac{d_2-d_1}{\Delta yf}\cdot\frac{1}{\dfrac{1}{A'}+\dfrac{1-C}{A''}} \tag{6}$$

and the inward transport in the lower zone is

$$T''=-\frac{d_2-d_1}{\Delta yf}\cdot\frac{1}{\dfrac{1}{A'(1-C)}+\dfrac{1}{A''}}. \tag{7}$$

In eqs. (6) and (7) all terms on the right side of the equations can be accurately determined in normal oceanographic data.

The application of these equations is demonstrated in a study of Juan de Fuca Strait (Tully and Herlinveaux, 1960).

*Comments.* Cameron (1951) observed the cross channel slope of the isopycnal surfaces in Portland Inlet, B.C. He showed that the difference of velocity between the upper and lower zones agreed well with that deduced from direct current observations. Pritchard (1956) made further studies in the James River. He observed that the lateral gradient of pressure is not exactly balanced by the Coriolis force, as calculated from the observed mean flow. He ascribed the residual force to the gradient of a lateral Reynold's shear stress. Stewart (1957) showed that the discrepancy was probably due to curvature of the flow path.

These studies show that the technique is applicable if care is taken to choose a cross-section in a straight reach of the embayment, or where curvature of the flow path is not a consequence of boundary limitations.

In many cases the velocity-depth gradient remains nearly constant throughout the cycles of tidal flow, as in Juan de Fuca Strait (Fig. 4). In such cases the transverse slope of the isopycnals must remain almost constant. Hence any measurement of the difference of dynamic heights may be utilized, regardless of the stage of the tide during observations. However, in those cases where the velocity gradient, and hence the slope of the isopycnals, varies with the tide, it is implied that the mean values through a tidal day should be used in the equations.

The dynamic heights may be evaluated from any columns in the section which extend to or below the deepest level of the interzone boundary ($L$). It is unlikely that there would be significant differences of dynamic heights below this level.

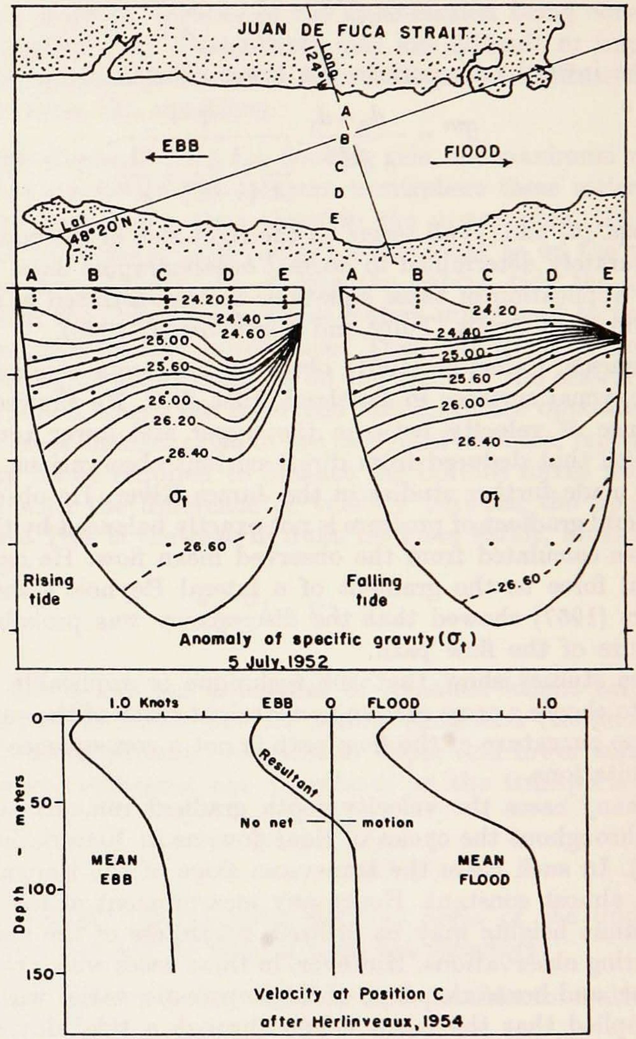


Figure 4. Density and velocity structure in Juan de Fuca Strait, B.C.

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