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A Study of the Oceanographic Structure in British Columbia Inlets and some of the Determining Factors.

Author

R.W. Trites

ATLANTIC OCEANOGRAPHIC GROUP

A STUDY OF THE OCEANOGRAPHIC STRUCTURE IN BRITISH COLUMBIA INLETS AND SOME OF THE DETERMINING FACTORS

by

R. W. Trites

N O T E

This is the reproduction of a Thesis submitted in Partial Fulfilment of the requirements for the degree of Doctor of Philosophy in Physics at the University of British Columbia. The work was carried out at the Institute of Oceanography while the author was on leave of absence from the Fisheries Research Board and while seconded to the Pacific Oceanographic Group.

ABSTRACT

Fresh water entering the inlets moves seaward, mixing with and entraining salt water from below. On the assumption that horizontal advection of salt is balanced by vertical diffusion, the magnitude and variation in time and space of the diffusion coefficient are determined by numerical evaluation of the differential equation describing this process. A method of systematically smoothing the data is established by solving the differential equation analytically. These solutions yield further information on the diffusion coefficient which is found to be nearly constant in the upper reaches of an inlet but increases rapidly towards the mouth in the brackish layer.

A technique is developed for determining the total fresh water inflow to the inlets using precipitation observations and available river flow measurements. The results indicate that a significant proportion of fresh water enters from the sides at all times. The importance of this in theoretical studies is emphasized. The results are also analysed for a possible relationship between inlet dimensions and intensity of mixing.

A new method, based on the heat budget, is developed to determine the mean seaward movement of the brackish layer. This method is also applied to determine the depth from which salt water is entrained into the surface layer.

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I. INTRODUCTION

A knowledge of the nature and process of mixing within a fluid is one of the fundamental requirements for an understanding of the circulation and ditribution of properties in the sea. Until recently the studies of the characteristics and dynamics of flow in the sea have been largely directed to those of the open ocean. However in the past few years there has been a growing interest in inshore waters and particularly in estuarial systems. An estuary may be defined as a semi-enclosed coastal body of water having a free connection with the open sea and containing a measurable quantity of sea salt (1). The inlets in the British Columbia coast are estuaries within this definition, and are of particular interest and concern to the Institute of Oceanography at the University of British Columbia.

Classification and Description of Inlets

Pritchard (1) has presented a general discussion on estuarine hydrography and has classed estuaries in terms of fresh water inflow and evaporation. When precipitation and fresh water inflow exceeds evaporation an estuary is termed positive. This is the classification into which the British Columbia inlets fall. In addition Pritchard classes them according to their geomorphological structure, i.e., they may be estuaries that have been formed by drowning of former river valleys, either from subsidence of the land or from a rise in sea level. These are usually relatively shallow estuaries. A second class of estuary is the deep basin type exemplified by the glacial cut fiords in the Norwegian and British Columbia coastlines. These estuaries are elongated indentations of the coastline, containing a relatively deep basin and steep sides, with or without a shallow sill at the mouth.

The inlets in the British Columbia coast vary in length from 5 to 90 miles and in width from one-half to 3 miles (see figs. 1, 2, 3 and 4). All of the inlets receive fresh water drainage which exceeds evaporation and consequently in the inlets there is a net movement of the water to seaward. Recent investigations of a number of the mainland inlets have been reported briefly by Pickard (2), and a detailed quantitative study of the behavior of fresh water entering the sea through Alberni Inlet has been reported by Tully (3). As a part of this study he constructed a hydraulic model of the harbour and head of the inlet and from this he was able to evaluate the effect of river discharge, wind, and tide on the rate of dissipation of fresh water seaward.

Only about five of the mainland inlets had been investigated from an oceanographic standpoint, even superficially, prior to 1951. During that summer however, a general survey of all the major mainland inlets was undertaken. The principal object of this survey was to gain information on the general characteristics of the water, the distribution of properties, distribution and abundance of plankton, and the circulation.

During each summer since 1951 there have been surveys in the inlets in the southern half of the British Columbia coast. Attention has been directed towards gaining a more detailed knowledge of the inlet characteristics than could be carried out during the extensive survey of 1951. In particular direct current measurements were taken in Bute and Knight Inlets during the summers of both 1952 and 1953, and during each survey the distribution of temperature, salinity and oxygen has been observed.

A brief summary of the water characteristics, currents, dissolved oxygen distribution, and internal waves has been reported by Pickard (2). Le Brasseur (4) has reported the general distribution and abundance of plankton. In addition a fairly detailed analysis of the seasonal variations of physical oceanographic variables in Bute Inlet, based on eleven surveys between August 1950 and August 1952, has been made by Tabata (5a, b).

The information on depths in the inlets is incomplete. Only eight of the inlets have been fully surveyed by the Hydrographic Service. To supplement the available charts, soundings were made during the oceanographic surveys. These were usually made along the centerline of the inlets. The average mid-channel depth is approximately 180 fathoms but 300 fathoms is common and the greatest depth recorded is 400 fathoms. The sides of the inlets above the water surface are mostly steep and rugged, and angles of 45° or more are not uncommon. Below the surface the slope usually continues at about the same angle to a fairly flat level bottom.

In a simple model of an inlet, which consists of an indentation perpendicular to a straight coastline, the 'mouth' and 'head' can easily be defined. The mouth is taken to be at the point where the inlet suddenly widens, and the head as the fartherest point in the inlet away from the mouth. However an outline of the British Columbia coast (figs. 2, 3, and 4) indicates a wide range in shapes and lateral boundaries of the inlets, with many of them being very irregular and sinuous in plan form. A number of them form a system of channels that are joined in a complicated network. This complication makes it difficult or impossible to define the location of the mouths of some of the inlets.

In many respects the British Columbia coastline resembles that of Norway. But the presence of a very shallow sill at or near the mouth is a common feature of most Norwegian fiords (6), whereas most British Columbia inlets do not have this feature. The shallow sill in the Norwegian fiords separates the deep water in the fiords from that of the open ocean. The sill may be only 2 or 3 fathoms in depth while the depth in the fiord may be 40 or 50 times as great. By contrast only 7 or 8 of the 25-30 British Columbia inlets posses sills. The deep water in the other 20 or more have free communication with the coastal waters, although in all cases the depth between the inlets and the edge of the continental shelf decreases to less than 100 fathoms.

Only three of the main inlets have sills. In Knight and Observatory Inlets the sill depth is 25 to 30 fathoms; in the Seymour and Belize Inlet system the sill is at a depth of 15 fathoms. The only very shallow sills are in the smaller inlets. Call Creek has a sill depth of about 10 fathoms. Princess Louisa Inlet, a small arm off the upper reach of Jervis Inlet (fig. 2), has an entrance sill of 3 fathoms, while Sechelt Inlet, another arm off Jervis Inlet, is only 7 fathoms at its mouth. These, then, constitute the extent of the sills in British Columbia Inlets.

General Distribution of Properties

1) Salinity

Because of the free connection with the sea, the bulk of the water in the inlets is saline. However, fresh water entering the inlets overrides the heavier salt water, but as it proceeds to seaward it mixes with and entrains salt water from below. A schematic diagram of the structure and circulation in the inlets, for summer months, is shown in figure 5. In all profiles the extent of the top layer is exaggerated relative to the thickness of the deep layer. This is intended only as a schematic representation with actual conditions varying over a fairly wide range. Ideally all the fresh water enters at or near the head, but in practice a significant percentage enters from the side. This modification will be fully discussed later. (Section VI). During summer months the surface water at the head is substantially fresh (salinity less than 1‰), forming a layer 2 to 9 metres in depth, depending on the inlet and the stage of the runoff. At 4 to 12 metres the salinity increases to 25‰ or more and at 18-20 metres it has increased to more than 0.9 of its value at the deepest point (referred to as the base salinity). The salinity in the top layer (upper 4-12 metres) increases down the inlet, reaching values at the mouth of the order of 0.7 to 0.9 of the base salinity value. These values correspond approximately to the surface salinities of the surrounding coastal waters. Thus at the head of the inlet the system approximates a two layer one with a very sharp transition zone (halocline). Towards the mouth the halocline is less apparent and may even be non-existent, so that on the basis of salinity no distinct layers can be identified.

The value of the base salinity in inlets south of Knight is 30.5-31.0‰ and north of Knight the value is approximately 32.5‰. The value for Knight is intermediate between these.

With the exception of Bute, the mainland inlets have been surveyed only during the late spring or summer months. During this period the fresh water discharge reaches a maximum value and consequently most of the salinity distributions available show a greater than average dilution due to river runoff. In the winter months the discharge of fresh water decreases, reaching a minimum during late winter or early spring. The fresh water concentration in the inlets likewise decreases and the distinct two layer system evident in the spring and summer may be almost absent. A comparison of figures 6a and 6b indicates the difference that results in Bute Inlet between February and May. In February the salinity at the head is greater than 20‰ and increases rapidly down the inlet until the water becomes nearly vertically homogeneous, whereas in May the water is substantially fresh at the head and even near the mouth it is still less saline than in winter. However in both cases the vertical depth of penetration of fresh water is about equal.

Since for the most part, only observations taken during summer months are available, the rest of this thesis refers to summer conditions, unless otherwise stated.

As well as the seasonal variations in dilution in a particular inlet there are also relatively large variations from inlet to inlet. Salinity profiles taken in Jervis Inlet indicate a different pattern from those in Bute or Knight. (Compare figures 7a and 7b). While the depth to which the fresh water penetrates is about equal in both cases, the general pattern is quite dissimilar. The salinity range in Jervis is markedly less than in Bute as well, with the profiles at the head of Jervis corresponding roughly to those observed at the middle of Bute Inlet. In other inlets the general pattern of salinity curves and range of values may be similar to those of Bute or Knight, but the depth of the brackish layer may be very different. The similarities and differences between the salinity curves for different inlets can usually be expressed in terms of the two features--thickness of the brackish layer and the longitudinal salinity range from head to mouth.

2) Temperature

During the summer months the surface water is usually the warmest part of the column. The brackish surface layer is nearly isothermal. A rapid decrease in temperature occurs below this (thermocline), followed by a fairly steady decrease until a minimum is reached at about 60-90 metres depth. This minimum, which is most conspicuous in Bute Inlet, is most apparent during the early summer months and gradually dissipates during the later summer and autumn until it becomes imperceptible in the winter. This behavior indicates that the seasonal warming and cooling penetrates to about 100 metres.

During the winter months, the temperature increases with depth to a maximum at a depth greater than 100 metres and then decreases gradually to the bottom. The temperature of the deep water remains constant throughout the year at approximately $8^{\circ}C$.

During the summer months the temperature in the top layer usually increases from head to mouth (fig. 5). The increase ranges from less than $0.5^{\circ}C$. to as much as $5^{\circ}C$. It is due to absorption of heat from solar radiation. For a particular case the temperature increase will depend upon the combined effect of absorption of heat from solar radiation and the heat gained or lost in the mixing process. In winter months the temperature usually decreases toward the mouth, indicative of a net loss of heat to the atmosphere.

The density of the water, which is a function of both salinity and temperature, increases with depth in all cases. Typical curves are illustrated at the bottom of figure 5. The shape of the curves is very similar to that of salinity, indicating that, in contrast to the open ocean, temperature plays only a minor role in determining the density distribution in the inlets. Consequently, during most seasons of the year the density may be considered to be directly proportional to the salinity to a good approximation.

Circulation

Fresh water entering an inlet moves seaward because the resultant pressure forces are in that direction. As it moves along it mixes with (or entrains) salt water from below (3). Observations indicate that the bulk of the fresh water remains near the surface, so that the brackish layer remains substantially constant in thickness along the inlet, and therefore the velocities in this layer must increase to seaward. If the amount of salt in the inlet is to remain constant, then there must be a compensating up-inlet flow at deeper levels to balance the net outflow of salt in the surface layer (fig. 5).

Current measurements in Bute and Knight Inlets indicate that the surface layer is moving seaward with a net velocity which is nearly uniform with depth. This extends down to about the top of the halocline from which point the current decreases very rapidly with depth to zero. Below this the velocity changes sign and the magnitudes are smaller than for the surface layer.

Superimposed in this net circulation are the tidal motions. Since these are oscillatory and considerable in magnitude the resultant currents are generally oscillatory in character also, though during large run-off periods the estuarial flow to seaward at the surface may be greater than the tidal velocities and result in a undirectional flow throughout the tidal cycle.

Most of the current observations have been made in the upper 20 metres, but enough reliable measurements in the deeper waters have been made to reveal that large currents do occur at times at the greater depths. However, these high speeds must be in the form of surges and are possibly jet-like in character, extending only over part of the width of the inlet. Application of the principle of continuity of volume indicates that these large velocities cannot exist over the whole tidal cycle.

The volume of water which is supplied to the inlet between low and high tide must all pass through the mouth of the inlet, and since the inlets are relatively constant in width, these tidal velocities are a maximum at the mouth and decrease to zero at the head of the inlet.

Tidal velocities vary in magnitude with depth and in general appear to be somewhat larger below the brackish layer than at the surface. This vertical variation gives rise to variations in the vertical velocity shear over a tidal cycle which reaches a maximum during the flooding tide.

The foregoing constitutes a general description of the circulation and distribution of properties in the inlets. Relatively large departures do occur from the 'typical' structure and circulation as presented in figure 5, but these will be dealt with in subsequent sections.

The factors which influence the structure and circulation will be enumerated in more detail in the following section (II).

This thesis therefore, presents the results of a critical examination and study of the vertical-longitudinal distribution of properties and the factors which may control or relate to it. An attempt will be made to represent the salinity profiles by theoretical curves derived from relatively simple assumptions about the mixing process. This is desirable both from the standpoint of discovering which are the important processes occurring, and as a guide in 'smoothing' the field data. In addition an effort will be directed toward an evaluation of some of the internal and external factors influencing the distribution of mass and circulation within the inlet.

II. FACTORS INFLUENCING THE CIRCULATION AND DISTRIBUTION OF PROPERTIES

The circulation and distribution of properties are influenced by a number of internal and external factors which must be considered and evaluated if an understanding of the important processes prevailing is to be achieved. The following comments will serve to introduce some of these factors whose influence is to be discussed in detail later.

Physiography

The general character and shape of the inlets has been stated in the previous section, but no mention was made of the possible part that these might play in determining the circulation within the inlet. The effect of a shallow sill at the mouth is to impede the flow of deeper water into and out of the inlet. The presence of a sill is a common feature of the Norwegian fiords (6, pp. 1024-1028) and results in stagnation as indicated by the low concentration of dissolved oxygen in the deep water. However it is evident from the absence of stagnation in larger British Columbia inlets (2, No. 99) that the presence of a sill at 25 to 30 fathoms depth has no effect in impeding the deeper circulation.

The second feature of the physiography of the inlets important oceanographically are the lateral boundaries and irregularities. The sinuous character of many of the inlets undoubtedly gives rise to extensive horizontal eddies similar to those which can be demonstrated in hydraulic flumes. The resultant effect of such irregularities is to produce a more nearly homogeneous medium laterally although large variations in lateral circulation will result. The detailed effect of these variations is possibly important but it has not been examined in this study because of the small number of available observations.

The relatively steep sides of the inlets permit one to assume them vertical in so far as the width of the upper layer is concerned. Although there is some variation in width along the inlets, the effect of this in determining the circulation and mixing process is neglected in the subsequent discussion.

Tides

The tides in the inlets and channels of the British Columbia coast are important oceanographically because of their considerable vertical range. The tides are of the semi-diurnal mixed type, there being two high and two low waters each day, no two of which are equal in height.

On the southern half of the coast the inequality in heights of the high waters or low waters is very marked, whereas in the more northern region the inequality is less pronounced. The most rapid change in the character of the tide occurs in the network of channels between Bute and Knight inlets. Times of high and low water are very different between the northern and southern regions, being approximately 180° out of phase.

The general character of the tides is illustrated in figure 8. 8a and b illustrate the maximum variation in types of curves for the southern inlets, with all intermediate combinations possible, while figures 8c and d indicate the large and small range for the northern inlets.

The time of high or low waters occur nearly simultaneously at all points in the inlet. Even in the very long inlets, there is evidently a phase lag between mouth and head of not more than a few minutes (7). To satisfy this condition requires that on the average tidal velocities be largest at the mouth and decrease to zero at the head where the only tidal component is in the vertical direction.

Detailed information on tidal heights is published by the Canadian Hydrographic Service (8). The normal range of the tides, as given in the tide tables, varies from about 6 feet up to 24 feet. Current measurements made by the Hydrographic Service, are limited to the channels of importance to navigation, and are confined to surface observation. The only observations available for the inlets were made during the oceanographic surveys. These indicated a considerable range in value, with the smallest being less than 5 cm./sec. and the maximum exceeding 100-150 cm./sec. (Section V).

Meteorology

Twice daily observations of cloud coverage, wind speed and direction, temperature, and precipitation are recorded at a number of points along the coast. However, the prevailing conditions in the shelter of the inlets are often quite different from those in exposed locations. On all oceanographic surveys observations of cloud coverage, wind speed and direction, air temperature and relative humidity were made. These observations, limited mostly to summer months, indicate that on the average the wind is directed up-inlet in the daytime with a speed of 5-10 miles per hour, decreasing to nearly zero at nightime. Cloud is usually fairly extensive, the average being approximately 7/10 coverage. The relative humidity is usually greater than 70%. Daily totals of solar radiation, direct and diffuse are given for a number of positions along or near the coast, in Kimball's table (9). A more detailed analysis of the influence of meteorological conditions on mixing and heat absorbed from solar radiation will be considered in section VII.

River Discharge and Precipitation

The rivers which flow into the inlets have drainage basins which lie almost entirely in the Coast Range mountains of British Columbia. Average annual precipitation in this region varies over a wide range being less than

35 inches at some points and as much as 275 inches at others. At sea level and at the lower elevations most of the precipitation is in the form of rain whereas at the higher altitudes a fairly high proportion of it falls as snow. Usually the delay between the time precipitation falls on a basin and the time it flows from the rivers into the inlets is less for rain than for snow. The snow may remain frozen and accumulate for several months before it melts, whereas water in the liquid form begins to move immediately towards lower elevations. In general two yearly peaks in river discharge occur. One is in the late autumn months and is associated with the heavy autumn rains. The second, and usually the larger of the two, occurs in late spring or early summer months and is associated with the melting of snow and ice. In many cases the rivers are glacier fed and the flow may continue at relatively high values during the entire summer.

A detailed analysis of fresh water discharge, its distribution within the inlet, and the variation from inlet to inlet will be given in a following section (VI).

Internal Waves

Whenever there is a boundary between two fluids such that there is a density difference across the boundary, then it is possible for waves to exist at the interface. These may be either standing or progressive waves. The air-sea boundary is probably the most familiar example. However, it is not necessary that there be a density discontinuity but merely a vertical density gradient in order for waves to exist. In several of the inlets internal waves have been observed. These are of two types:

i) Shallow internal waves: these are observed at the depth of the halocline (3 to 15 metres) in some localities and have periods of one to three minutes, wavelengths of 60-200 metres, and heights up to 10 metres. The period of oscillation is comparable to the value derived from the stability of the water column (i.e. the natural period of vibration of an element of fluid within the column). These waves are often observed to occur shortly after the tide has turned to flood, when the vertical velocity shear is usually greatest. The waves may progress slowly up the inlet or remain nearly stationary, depending on the seaward velocity of the surface layer.

ii) Mid-depth internal waves: These are observed at depths of 30-120 metres, with a period corresponding to that of the tide, and an amplitude of 25-30 metres. Observations taken simultaneously in Bute Inlet at two locations suggest that these are standing waves, with possibly two nodes, the mouth being an antinodal point.

Internal waves then, produce a distortion on the mass field which complicates the problem of obtaining a reliable mean distribution of properties. In addition the shallow internal waves may have an important effect on the mixing process.

The various factors mentioned in this section all play a role in determining the instantaneous and net circulation and distribution of variables within the inlet. In the subsequent dicussion these components will be considered and expanded, so that for a particular situation a better understanding may be obtained of the relationship between the kinematics, dynamics, and distribution of properties.

III. MIXING PROCESSES IN ESTUARIES

The distribution of mass, which is a function of salinity and temperature, and the motion in the sea must be interdependent. The motion can often be determined from a knowledge of the mass distribution, and conversely the mass distribution is at least in part a consequence of the motion, since moving water may be expected to carry with it its properties. This is usually called 'advective' transfer of a property. In addition to the transport by the mean motion of the fluid, transfer of properties also occurs because of the random of turbulent motion. This 'non-advective' process is referred to as 'diffusion'. In the past there have been many attempts to determine the field of motion from the field of mass. Some have been successful but sometimes erroneous conclusions have been drawn because the role of the non-advective transport has either been neglected or inadequately understood.

The majority of the studies of estuaries so far have been mostly concerned with relating the field of mass to the field of motion. They have been less concerned about the actual processes which cause the redistribution of properties, such as salinity, along the length of an inlet. In fact very little is known for certain about the mixing processes which actually occur in estuaries.

The present study is essentially an attempt to explain the observed distribution of a property such as salt, in an estuary, on the basis of certain assumptions about the mixing process itself. Ketchum has had considerable success, adopting this approach, in his use of the tidal prism method (10), where the essential assumption is that mixing is complete within each tidal prism. Arons and Stommel (11) have refined this method by introducing a 'mixing length' and incorporating the tidal prism concept into a differential equation. The results obtained by these researchers seem to be most applicable to the shallow coastal plain type of estuary, where vertical stability is much less than in the British Columbia inlets, and the distinct two layer system at the estuary's head is less pronounced.

In discussing the hydrodynamics of estuaries oceanographers often speak in terms of superposed layers of water of different density, each layer being separated from its neighbour by an abrupt discontinuity in density. It is possible to imagine systems whose density stratification cannot be destroyed by mixing, such as layers of mutually immiscible fluids like water and benzene. In some estuaries this simplified concept of density structure is nearly realized. However in natural estuaries where stratification of density is due to differences of salt and heat content of a single fluid, water, some degree of mixing invariably occurs to destroy the ideal abrupt density discontinuity of the simple model. This mixing may be related directly to the field of flow in the stratified medium, or it may be due to external influence quite independent of the field of flow, such as winds. The field of flow can be thought of as composed of two parts, the estuarial circulation and the tidal circulation. The velocities of flow due to the tidal wave are generally greater than flows set up by the river water passing through the estuary. These tidal velocities and movements must in many cases result in much more effective mixing than would occur if there were no tides. In turbulent flows there is some evidence that the intensity of mixing is related nonlinearly to the velocity and velocity gradients, (e.g. to the square of the velocity). Irregular lateral boundaries and constrictions in width likewise undoubtedly play a role in controlling the degree of mixing that occurs.

In estuaries such as Burrard Inlet the ratio of width to length is nearly unity and lateral variations are very important. Campbell (12) has investigated the circulation in this inlet and found that lateral stresses play an important part in determining the horizontal circulation. In this report, however, only relatively long narrow inlets will be considered. It will be assumed that they are laterally homogeneous, or nearly so, and for the most part only processes occurring in the vertical-longitudinal plane will be discussed.

To return to the relation between the motion and the distribution of properties: redistribution of mass may take place as a result of the mean motion of the fluid (advection), and also by random motion (diffusion). In the classical case of Laminar or streamline motion the velocity is presumed to be a slowly or smoothly varying function of position and time and the only random motion is that of the molecules (molecular diffusion). In the ocean (or atmosphere) however, evidence suggests that in addition to the random molecular motion there is also a random macroscopic motion. This type of flow is referred to as 'turbulent' and the macroscopic random motion gives rise to a transfer of properties by a process referred to as 'eddy diffusion' by analogy with the molecular case. The flow may be considered to consist of two parts: a relatively simple mean motion, and superimposed on this a complicated random or eddy motion which fluctuates but is not obviously periodic. Evidence suggests that in turbulence where the random motions are on a much larger scale than the molecular agitations, that diffusion is related in some way to the bulk motion and to the boundaries. Mathematically turbulence implies a state in which the instantaneous velocities exhibit irregular and apparently random fluctuations so that in practice only statistical properties can be recognized and subjected to analysis. In part, the situation is analgous to that accepted in molecular physics, i.e. that the exact knowledge of the motion of individual molecules cannot be obtained, but given an assembly of molecules it is possible to deduce the laws of their behavior in statistical terms. But the analogy is only partly true, because the turbulence is usually not completely random, being related in some way to the stability and field of motion within the fluid.

Equation for Distribution of Property

Consider the distribution in a body of water of a property such as salinity (grams of dissolved material per kilogram of water). Assume a rectangular co-ordinate system x, y, z, so that the salinity is given by $S = S(x,y,z,t)$ where t is time.

At any instant, the time rate of change of salinity will be determined by the sum of two processes, the transfer by the macroscopic motion and that by molecular diffusion. The rate of change of salinity can be described by the differential equation:

$$\frac{dS}{dt} = \frac{\partial S}{\partial t} + u\frac{\partial S}{\partial x} + v\frac{\partial S}{\partial y} + w\frac{\partial S}{\partial z} = K\left\{\frac{\partial^2 S}{\partial x^2} + \frac{\partial^2 S}{\partial y^2} + \frac{\partial^2 S}{\partial z^2}\right\} \quad (1)$$

where u, v, and w are the components of velocity in the x, y, and z directions respectively, and K is the molecular diffusivity. In natural turbulent flows, it is assumed that the instantaneous values of salinity and velocity can be represented as the sum of a mean and a fluctuating component, i.e.

$$S = \bar{S} + S', \quad U = \bar{U} + U', \quad V = \bar{V} + V', \text{ and } W = \bar{W} + W' \qquad (2a)$$

where the mean values are defined to be:

$$\bar{S} = \frac{1}{T}\int_{t-\frac{1}{2}T}^{t+\frac{1}{2}T} S\,dt \text{ , etc.} \qquad (2b)$$

The interval T is to be taken over a long enough period so that $\overline{s'} = \overline{u'} = \overline{v'} = \overline{w'} = 0$.

Substituting (2a) in equation (1) and taking the average over a time T, the mean value equation becomes:

$$\frac{\partial \bar{S}}{\partial t} + \bar{U}\frac{\partial \bar{S}}{\partial x} + \bar{V}\frac{\partial \bar{S}}{\partial y} + \bar{W}\frac{\partial \bar{S}}{\partial z}$$
$$= \frac{\partial}{\partial x}\left(K\frac{\partial S}{\partial x} - \overline{U'S'}\right) + \frac{\partial}{\partial y}\left(K\frac{\partial S}{\partial y} - \overline{V'S'}\right) + \frac{\partial}{\partial z}\left(K\frac{\partial S}{\partial z} - \overline{W'S'}\right) \qquad (3)$$

The product terms $\overline{u's'}$ etc. will in general not be zero (equal zero only if there is no correlation between the two quantities). In practice they turn out to be several orders of magnitude larger than the molecular diffusion terms so that for most purposes the latter can be neglected. The terms $\overline{u's'}$ etc., are commonly referred to as the random flux or eddy flux terms. The derivative of these terms is referred to as eddy diffusion. Similar terms arise in the equation of motion where momentum is substituted for salinity. In this case the random flux terms describe the effect of the fluctuations on the transport of momentum across a surface in the fluid and represent frictional stresses (referred to as 'eddy' or 'Reynolds' stresses).

To evaluate these eddy terms the most direct method would be to measure u', s', etc., directly and determine the mean product. Instruments to carry out this process are only in the early development state and are not available for routine use. It is therefore necessary to adopt some other approach. It is noted that equation (3) cannot be solved analytically unless the form of the random flux terms is known. In an attempt to overcome this difficulty for the frictional stress Boussinesq (13, p. 126) postulated the mixing movement is simply equivalent to an increase in the viscosity coefficient i.e., that a term such as $\overline{u'v'}$ could be written as $\eta \frac{\partial \bar{U}}{\partial x}$ by analogy with the molecular case. This idea was extended by Schmidt (6, p. 91), to the transfer of other properties than momentum by writing for example, $\overline{u's'}$ as $A_x \frac{\partial \bar{S}}{\partial x}$ where the quantity A_x the eddy or exchange ('Austauch') coefficient represents the rate of transfer by the random eddy motion. In principle it is probable that this eddy coefficient may be a function of the motion, i.e. of the position in the fluid. With this hypothesis, and neglecting molecular diffusion, equation (3) becomes:

$$\frac{\partial \bar{S}}{\partial t} + \bar{U}\frac{\partial \bar{S}}{\partial x} + \bar{V}\frac{\partial \bar{S}}{\partial y} + \bar{W}\frac{\partial \bar{S}}{\partial z} - \frac{\partial}{\partial x}\left(A_x \frac{\partial \bar{S}}{\partial x}\right) + \frac{\partial}{\partial y}\left(A_y \frac{\partial \bar{S}}{\partial y}\right) + \frac{\partial}{\partial z}\left(A_z \frac{\partial \bar{S}}{\partial z}\right) \quad (4)$$

where the quantities A_x, A_y, A_z, are usually called the eddy diffusivities in the x, y, and z directions respectively. Similarly the terms eddy conductivity and eddy viscosity are used for heat and momentum respectively.

The eddy coefficients are not a property of the fluid but are functions of position, time, and the scale of turbulence and in general the vertical coefficients will be quite different from the horizontal ones.

If the coefficients are independent of position, equation (4) becomes:

$$\frac{\partial \bar{S}}{\partial t} + \bar{u}\frac{\partial \bar{S}}{\partial x} + \bar{v}\frac{\partial \bar{S}}{\partial y} + \bar{w}\frac{\partial \bar{S}}{\partial z} = A_x \frac{\partial^2 \bar{S}}{\partial x^2} + A_y \frac{\partial^2 \bar{S}}{\partial y^2} + A_z \frac{\partial^2 \bar{S}}{\partial z^2} \quad (5a)$$

and the diffusion is termed 'Fickian'.

If the coefficients are independent of direction, i.e., $A_x = A_y = A_z = A$, then (5a) becomes:

$$\frac{\partial \bar{S}}{\partial t} + \bar{u}\frac{\partial \bar{S}}{\partial x} + \bar{v}\frac{\partial \bar{S}}{\partial y} + \bar{w}\frac{\partial \bar{S}}{\partial z} = A\left(\frac{\partial^2 \bar{S}}{\partial x^2} + \frac{\partial^2 \bar{S}}{\partial y^2} + \frac{\partial^2 \bar{S}}{\partial z^2}\right) \quad (5b)$$

There is of course some question as to the physical reality of introducing the concept of eddy diffusivity. Some researchers have felt that there is no basis for it, arguing that one is merely replacing one group of terms about which we know very little with a second group of terms about which we also know very little. Pritchard (1) has taken this point of view and has proceeded to evaluate the mean random flux terms ($\overline{u's'}$, etc.) indirectly from the mean values of velocity and salinity. The method is still empirical and the information obtained about the actual mixing process is rather limited. It involves a very large number of individual observations which are, in some places, from the practical standpoint, impossible to obtain. As well, the information obtained is valid only for the particular place where the observations were made, and at that particular time. Presumably the form obtained for the random flux terms and their variation in time and space would suggest the possible factors to which they are related. In other words, if the results are to have a wider application, it is necessary to relate these random flux terms to some physical entity or its derivatives.

Since large numbers of observations are not available for the inlets, the Schmidt hypothesis will be used in this study and equation (4) will be assumed to apply.

IV. SALINITY DISTRIBUTION AND EDDY DIFFUSIVITY

The general features of the salinity distribution were outlined in the Introduction. Typical summer distributions were presented, and as well, an indication given of the seasonal variation. A more complete analysis of these distributions, together with the vertical eddy diffusion coefficients required to maintain these distributions, will be presented and discussed in this section.

As noted previously, most of the inlets have a relatively large quantity of fresh water discharged into them during the summer months. Consequently the family of vertical salinity curves from head to mouth for each inlet bears a degree of similarity from inlet to inlet, although both the thickness of the brackish layer and its increase in salinity from head to mouth may vary appreciably. (See figs. 6, 7, and 9-14 incl.). Despite these variations, the general agreement in 'shape' of the curves, suggests that similar mixing processes prevail in all the inlets.

To examine a specific family of salinity curves, consider those observed in Bute Inlet on May 29, 1952 (fig. 7b). The depth of the halocline at the head (station 8) is at 6 metres. This depth has changed to very little at station 7, while the mean salinity in the brackish layer has increased from about 1‰ to 6‰. Between stations 7 and 5 however, there has been a marked change in character. The halocline, although not so clearly defined, is now at about 4.5 metres depth. At station 3, the salinity increases almost uniformly with depth down to 5 metres, after which it increases very slowly. While there is little evidence of a halocline at station 3, the profiles indicate on the whole, a decrease in the thickness of the brackish layer to seaward.

Observations taken in Bute Inlet at about the same time, but a year earlier, indicate the same general features as outlined above, (fig. 6b). The brackish layer is slightly thinner at the head (5 metres) than in the previous sample, but the same general decrease in thickness to seaward is evident, with a value of only about 3 metres at station 2. Another feature of the salinity profiles, which is evident here and in numerous other cases is the reversal in direction of the horizontal salinity gradient with depth. (See also figs. 7b, 9a,b, 10a, 11a,c, 12a, 13a, 14a,b.). This reversal occurs just below the halocline and is generally evident only over part of the length of the inlet. It is more prevalent in the upper reaches of the inlet. The magnitude of the reversed gradient is very small relative to that occurring in the top layer. In all cases the horizontal salinity gradient decreased to practically zero at 20 metres depth.

In an attempt to gain some knowledge of the mixing processes occurring in the inlets, application of equation (4) (Section III) in a simplified form will be investigated.

The inlets are relatively long, deep and narrow with little variations in salinity laterally. Assuming a steady state, and that the important processes maintaining the salinity distribution are horizontal advection balancing vertical advection and diffusion, equation (4) reduces to:

$$U \frac{\partial S}{\partial x} + W \frac{\partial S}{\partial z} = \frac{\partial}{\partial z}\left(A_z \frac{\partial S}{\partial z}\right) \tag{6}$$

(where all quantities refer to the mean values as defined by eqn. 2b).

If horizontal and vertical velocity observations were available, then equation (6) could be integrated vertically by numerical methods and the magnitude and the vertical and longitudinal distribution of A_z determined. However observations are not sufficient to do this since velocity observations are limited to one position in the inlet (about 2/3 of the way down from the head). Therefore if A_z is to be evaluated using the available data, further simplification of equation (6) is necessary.

If it is assumed that the vertical advection of salt is negligible or that it can be included in the diffusion term, then equation (6) may be written as:

$$u \frac{\partial S}{\partial x} = \frac{\partial}{\partial z}\left(A_z \frac{\partial S}{\partial z}\right) \tag{7}$$

The equation as written, formally implies that $w = 0$. But is the actual data from the inlets are applied to the equation, then the calculated values of A_z incorporate the diffusion plus the effect of vertical advection.

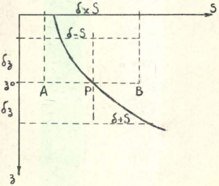
The lack of velocity observations requires one further simplification to equation (7), namely that the term $\frac{\partial A_z}{\partial z} \frac{\partial S}{\partial z}$ be assumed to be much smaller than $A_z \frac{\partial^2 S}{\partial z^2}$

Thus equation (7) becomes:

$$u \frac{\partial S}{\partial x} = A_z \frac{\partial^2 S}{\partial z^2} \quad \text{or} \quad \frac{A_z}{u} = \frac{\partial S/\partial x}{\partial^2 S/\partial z^2} \tag{8}$$

A_z/u can be calculated numerically using the method described by Proudman (14, pp. 111-115). The method is the following:

Consider the vertical distribution of salinity at two stations A and B.



Construct the mean salinity profile of salinity between stations A and B (solid line). The expression for A_z/u will be derived for the depth z_0. Draw two lines parallel to z_0, a distance δz above and below it. Let $\delta_x S$ denote the mean increment of salinity between A and B, $\delta_- S$ the increment of average salinity from $z-\delta z$ to z, and $\delta_+ S$ the increment of average salinity from depth z to $z+\delta z$. Let δx denote the distance between the two stations and let all derivatives be evaluated at P.

Then $$\delta_x S = \frac{\partial S}{\partial x}\,\delta x \qquad (9)$$

$$\delta_+ S = \frac{\partial S}{\partial z}\delta z + \frac{1}{2}\frac{\partial^2 S}{\partial z^2}(\delta z)^2 + \frac{1}{6}\frac{\partial^3 S}{\partial z^3}(\delta z) + \cdots \qquad (10)$$

$$\delta_- S = \frac{\partial S}{\partial z}\delta z - \frac{1}{2}\frac{\partial^2 S}{\partial z^2}(\delta z)^2 + \frac{1}{6}\frac{\partial^3 S}{\partial z^3}(\delta z)^3 + \cdots \qquad (11)$$

Subtracting (11) from (10) gives:

$$\delta_+ S - \delta_- S = \frac{\partial^2 S}{\partial z^2}(\delta z)^2 \qquad (12)$$

to the third order.

Combining (9) and (12) gives:

$$\frac{A_z}{u} = \frac{\delta_x S}{\delta_+ S - \delta_- S}\;\frac{(\delta z)^2}{\delta x} \qquad (13)$$

Equation (13) was applied for most of the inlets using the available data. Some of the results of these calculations are given in Tables I-XIV. In Table I the calculations for Bute Inlet, May 17-25, 1951 are recorded. The profiles (fig. 6b) from which these were calculated represent averages of approximately 8 at each station. The values of A_z/u in the top 3 to 4 metres are generally positive, while those below this are mostly negative. Since in a simple diffusion process it is physically unacceptable to have a negative diffusion coefficient, then this is taken to indicate that u changes sign between these two layers. The results indicate as well that the seaward moving top layer decreases in thickness towards the mouth. The value of A_z/u in the surface layer is largest at the head and decreases toward the mouth. This decrease is associated with increasing values of u rather than with decreasing values of A_z, although in this instance the amount of the decrease is more than can be accounted for by the velocity increase.

In Table II, the results of the calculation for Bute Inlet, May 29, 1952, indicate more regularity than those in Table I. Like that of Table I, it indicates a decrease in the depth of no motion towards the mouth. The decrease is from about 7 metres at the head to less than 4 at the mouth. The reversal in sign of A_z/u, occurs at the inflection point of the salinity curves. The magnitude of A_z/u decreases only slightly towards the mouth. Since the velocity at the head is of the order of 2 cm./sec. (results to be presented in Section V) and increases towards the mouth to about 25 cm/sec., this gives a value of A_z at the

2 metre depth of 0.16 cm.^{2}/sec. and 1.25 cm.^{2}/sec. at the head and mouth respectively. The generally larger values of A_z/u in the lower layer are associated with the decrease in velocity. Values just below the depth of no motion are very large because the velocities are nearly zero, but a decrease in value below this and finally increasing again is associated with u increasing to a maximum up-inlet value below the depth of no net motion and then decreasing slowly to zero at greater depths.

Application of a reasonable velocity field to the results of Table II indicates that for the surface seaward moving layer, A_z varies from about 0.1 to 1.5 cm.^{2}/sec. from head to mouth respectively while those in the deeper up-inlet moving water fluctuate from less than 0.05 to 0.5 c.^{2}/sec.

The values of A_z/u calculated for Jervis Inlet (Table III) from observations taken on May 27, 1952, lead, at first sight to a very interesting conclusion. The values are all negative, which indicates an up-inlet movement at all depths. The suggested explanation of this is the following: the points of observations near the surface are at 0 and 2 metres. Smooth curves plotted through these points give results as shown in figure (7b). The curves contain no inflection point and therefore the results indicate a uni-directional flow. However, in actual fact a very shallow surface layer (2 metres thick) probably existed whose salinity increased more slowly with depth than in the water just below it. If this was the case then the calculations would have given positive values for the surface layer.

The volume of fresh water leaving Jervis Inlet is quite small and consequently the velocities are small. The generally smaller values of the quantity A_z/u, compared to those of Bute Inlet indicate that A_z is smaller in Jervis Inlet than in Bute.

Similar calculations were carried out for most of the other large inlets and also for different seasons of the year. The results for Gardner Canal, July, 1951 are given in Table IV. These indicate all positive values above the 6 metre level but below this the sign alternates irregularly. The amount of information which can be deduced from this table is limited.

Calculations for Portland Canal (Table V), indicate a moderate degree of regularity between station 6 and 10, but above this, the results fluctuate widely. The 'break' in the results correlates with a break in the observations. The observations below station 10 were made on July 22nd, while those above it were made on July 23rd. The results below station 10 indicate that the seaward moving upper layer decreases in thickness from about 8 to 6 metres between stations 10 and 6 respectively. The application of a reasonable velocity field between these stations indicates that A_z is nearly constant for the brackish layer. Observations below station 6 were not used, since the influence of the large outflow from the Nass is very marked in this region.

The results for Dean Channel (Tables VI and VII) likewise show a wide variation in magnitude and a fluctuating sign. Here too the modifying effect of side arms and interconnecting channels give rise to a more complicated salinity distribution than might be expected in a simple, two dimensional inlet, with all the fresh water entering at the head.

Results of calculations for Knight Inlet are shown in Tables VIII and IX. Once again, both the sign and magnitude of A_z/u fluctuates, although the average value of A_z/u in Knight in the upper layer is somewhat higher than for Dean Channel, Gardner, or Portland Canals. Since velocities should be of comparable magnitudes (see Section VII) then one concludes that A_z is larger in Knight Inlet than in those mentioned.

Effort was directed as well, towards an evaluation of the seasonal variation of A_z/u in Bute Inlet, using data taken in October, February, March, and July. (See figs. 6a and 9c). The results of the calculation are presented in Tables X, XI, XII, and XIII. In general, no seasonal pattern can be detected. The values alternate in sign with apparent randomness and as well, no significance could be placed in the magnitude of A_z/u. Since little is known about the velocity field during autumn and winter months then, no conclusions about the seasonal variations of A_z can be drawn.

The apparent randomness in magnitude and sign of the quantity A_z/u in many instances, suggests a number of points.

1) It was assumed in the calculations that a steady state existed. If this condition prevails, and if A_z is required to be positive, then a very irregular and unrealistic velocity field exists. On the other hand, it is doubtful if the steady state assumption is valid. Actually, large inhomogeneous eddies were probably moving through the area, resulting in temporary reversals in horizontal salinity gradients. These inhomogeneties could be caused in a number of ways. Fresh water entering the inlets may vary over a wide range of volume in a short period of time (discussed in Section VI). Alternately, the fresh water does not in most cases, have a single entrance point in the inlet. Significant proportions may at times be entering at a number of points between inlet head and mouth. This undoubtedly accounts for some of the irregularities in A_z/u, in complicated networks such as Dean Channel and to a lesser extent in Gardner Canal. Alteration of the salinity profiles by amounts known commonly to occur locally results in a removal of many of the irregularities of A_z/u both in sign and magnitude.

The tides also constitute another factor in giving rise to local variations in salinity. However investigation of the effect of this factor resulted in only a slight improvement in smoothing the quantity A_z/u.

2) The method of calculation assumes that the vertical mean motion is zero, but the observed values of salinity of necessity include the effect of vertical advection. Therefore the calculated values of A_z/u actually incorporate the effect of vertical advection. The effect of vertical advection acts in the opposite direction to horizontal advection in maintaining the salinity at a point. Consequently a change in the magnitude of the mean vertical velocity, even without a sign change, could be sufficient to make A_z negative. A consideration of this effect in the calculations results in some simplification of the horizontal circulation pattern.

3) In Section II the general features of internal waves, observed in some of the inlets, were discussed. If observations were made during a time when internal waves were present, the salinity distribution obtained would not be representative of the steady state. Since in most cases, it is not known definitely whether internal waves were present, then an evaluation of this effect is impossible.

For data which were reasonably smooth, the calculations of A_z/u gave reasonably consistent results. But the relatively large number of irregularities in the calculations indicate the desirability of having some systematic method for 'smoothing' the data. Since in most cases only one, or possibly two, profiles are taken at each station, statistical methods cannot be applied. If smoothing is employed without any systematic method employed, a certain amount of 'personal' interpretation is involved. The possible results which this can lead to are indicated in Table XIV (see also fig. 9b). The curves were smoothed by eye only and were placed so that the observation points were not farther off the curves than was estimated as possible deviations in positioning the sample bottles. The calculations reveal at least a consistency in the circulation. The depth of no net motion is indicated to be at 4.5 metres at the head and decreases to about 3.5 metres at the mouth. A plot of u/A_z is illustrated in figure 15. (u/A_z is plotted rather than A_z/u to avoid infinities). A_z/u is found to be larger for the seaward moving surface layer than for the water moving up-inlet. The infinite values of A_z/u for some points in the surface layer is not necessarily indicative of zero velocity, but rather as an infinite value for A_z. This has resulted because the observations indicate no vertical salinity gradient, and since salt was observed to have moved vertically, then A_z is required to be infinite. However, since salinities were observed to only the nearest tenth of a part per thousand, a more accurate determination undoubtedly would have revealed a vertical gradient different from zero and consequently a finite A_z.

Type Distributions

The results of the analysis of unsmoothed data, indicate the unsatisfactory situation which prevails in attempting to use observations which are not representative of those of the 'steady state'.

It is often extremely useful and instructive to define a distribution of salinity which approximates to the observed one, and then to examine the velocity field and diffusion coefficient, assuming certain processes to be predominant in maintaining the salinity distribution. As an example, suppose the salinity in a vertical-longitudinal section is described by:

$$S = S_0 \, e^{-\frac{1}{ax + bz^3}} \tag{14}$$

where S_o is the salinity of the deep water

This is plotted in figure 16, for $a = 0.3$, $b = 10^{-3}$ where x is in nautical miles and z in metres. The close similarity in these curves to many of those observed in the inlets is evident.

Suppose that the distribution of salinity is determined by equation (8). Substituting for S and evaluating the expression u/A_z indicates that u changes direction with depth (provided $A_z > 0$), but it has the disadvantage that u/A_z approaches infinity as z goes to infinity. Since in this equation, A_z is constant, this requires that u increase with depth and approach infinity as z goes to infinity. Even if A_z is allowed to vary with depth, but remains finite the same conclusion is reached. Consequently if one is to represent the distribution by equation (14) and assume the distribution of salt to be determined by equation (8) then the function can only be used for a finite depth.

Another salinity function which describes a distribution of salinity that bears some similarity to that observed in the inlets is given by:

$$S = S_o - (S_o - S_s)e^{-Kz}(1 - Kz) \tag{15}$$

where S_s is the surface salinity and $K = K(x)$

This is plotted in figure 17, for $S_o = 30‰$

$S_s = \frac{S_o x}{L}$, $K = \frac{1}{K_o}(1 + \frac{x}{L})$, $L = 40$, $K_o = 6$.

Compared with figure 16, the layer of brackish water is less extensive, and to improve this would require increasing the number of terms in the polynomial part of equation (15).

Assuming as before that the distribution of salt is determined by equation (8), yields the result that $u = o$ at $z = \frac{1}{K}$ $\left\{ \text{or} \quad \frac{1}{K + \frac{1}{A_z}\frac{\partial A_z}{\partial z}} \right.$ if $A_z = A_z(z)$

This function has the added advantage over equation (14) in that the velocity approaches zero as the depth approaches infinity (instead of infinite velocities as in the case of equation 14).

Another interesting difference between equations (14) and (15) is the variation in depth of no motion along the inlet when equation (8) is assumed valid. Equation (14) indicates an increase in depth of the level of no net motion whereas equation (15) indicates that the depth decreases to seaward. The fact that most of the observations indicate a decrease in thickness of the brackish layer to seaward, favours the use of an equation of the form of (15), although a greater number of terms in the polynomial part of the expression is required, if a more realistic representation of the observations is to be obtained.

Relationship between A_s, $K(x)$, $S_s(x)$, and R, the river discharge:

For an inlet of constant width, the volume continuity requirement is:

$$\frac{\partial}{\partial x}\int_0^d u\,dz = 0 \qquad (16)$$

where d is the depth below which there is no motion. (Origin at the sea surface, z positive downward).

Substituting for u from equation (8), and for S from equation (15), yields:

$$A_s K^2 \int_0^d \frac{1-Kz}{az^2+bz+c}\,dz = R \qquad (17)$$

where

$$a = K\frac{\partial K}{\partial x}, \quad b = \frac{K}{S_0 - S_s}\ \frac{\partial S_s}{\partial x}, \text{ and } c = \frac{b}{K}$$

The integral has three different values depending on the relationship between b^2 and 4 ac.

For $b^2 = 4$ ac, one obtains:

$$\frac{2}{2az+b} - \frac{K}{2a} \log |az^2 + bz + c| - \frac{Kb}{2a}\left(\frac{2}{2az+b}\right)\Big|_0^d = \frac{R}{A_z K^2} \qquad (18)$$

The condition $b^2 = 4$ ac yields, upon integration and application of the boundary condition that $S_s = K = 0$ at $x = 0$:

$$S_s = S_0\left(1 - e^{-4K}\right) \qquad (19)$$

Substitution of (19) into (18) yields an equation which, for given values of R, gives a relationship between A_z and $K(x)$. The expression which results from these equations is logarithmic in form and difficulties were involved in obtaining an acceptable solution. Since the probable usefulness of the anticipated results were considered to be of only limited value, no further effort was made to solve the equations.

Other type distributions in addition to the above ones were considered, but no improvement in form and simplicity over the two considered here were found. This approach to the problems of the mixing process is in part, arbitrary, in that it does not consider first a process and then the distribution of salinity which results. Therefore it was considered to be of more value to postulate certain advection and mixing processes to be dominant and then to solve this equation to see how it determines the distribution of property.

Analytic Solutions of the Differential Equation

So far in this study the differential equation (4) which determines the distribution of salinity has only been examined in the differential form. A much more satisfactory and complete understanding of the factors involved in the mixing process would result if the equation could be solved analytically for given boundary conditions. In practice however the equation is too complicated to solve in its most general form. Consequently, several simplifying assumptions are necessary if the equation is to be solved analytically.

Previously calculations of the vertical eddy diffusivity were carried out on the assumption that horizontal advection of salt balances vertical diffusion. To solve this equation explicitly for a general case, with certain boundary conditions is impossible, unless the variation of velocity with position is known. However if the horizontal velocity is assumed to be independent of depth, then the solution can be expressed formally.

An examination of the families of salinity curves for the inlets suggests that formally the profiles have the same appearance as would result if two layers of motionless liquids in which initially the top layer contains no salt, and the lower one a quantity say S_0, were allowed to diffuse and the distribution examined at various times. As time approached infinity the solution would become homogeneous throughout.

1) Diffusion of salt into a semi-infinite layer of fresh water.

In order that the general form of the solution can be seen, a simple diffusion problem will be considered first.

Let the origin of the co-ordinate system be at the boundary between the fresh and salt water at the head of the inlet with x positive to seaward and z positive upward. It will be assumed that the diffusion of salt is determined by only two terms in equation (4) so that:

$$u \frac{\partial S}{\partial x} = \frac{\partial}{\partial z}\left(A_s \frac{\partial S}{\partial z}\right)$$

The subscript "s" replacing "z" will be used in all subsequent equations so as to distinguish the vertical eddy viscosity (A_v) and vertical eddy diffusivity (A_s).

As a first step consider $A_s = A_0$, a constant, and $u = u_0$, a constant. Then:

$$\frac{\partial S}{\partial x} = \frac{A_0}{u_0} \frac{\partial^2 S}{\partial z^2} \qquad (20)$$

Equation (20) is identical in form to the one-dimensional equation for conduction of heat in solids, and the methods employed in solving equation (20) are the same as those given for the solution of the heat equation (15 and

16). These equations have been modified to satisfy the particular boundary conditions appropriate to an inlet.

A solution to equation (20) which satisfies the conditions:

$$S(x,o) = S_o , \quad S(o,z) = o \text{ for } z \quad o$$

is given by:

$$S(x,z) - S_o \left[1 - \operatorname{erf} \left\{ \frac{z}{2\sqrt{\frac{A_o x}{u_o}}} \right\} \right] \qquad (21)$$

where

$$\operatorname{erf} z = \frac{2}{\sqrt{\pi}} \int_o^z e^{-f^2} df$$

Equation (21) is plotted in figure 18, for values of $\frac{Ax}{u}$ ranging from zero to infinity. In the diagram, the depth of 6 metres corresponds to z = o in equation (21). The salinity at 6 metres is maintained at 30‰. Calculations of the eddy coefficient in the previous section indicate that values of A_o of 1 cm.^{2}/sec. are common. Assuming this value, and examining the salinity at a depth of 1 metre, for various values of $\frac{x}{u_o}$, equation (21) gives the following values:

S(‰)	$\frac{x}{u}$ (days)
5	0.75
10	1.55
15	3.20
20	8.00
25	32.0

The solution given by (21) is unrealistic because it represents the case where the thickness of the fresh water layer is infinite. However, the general increase in the time increment for a unit increase in salinity would still be of the same sign even for a bounded medium. Work done by Tully (3) and Cameron (17) indicates that the velocity in the upper layer increases to seaward. This is required because the thickness of the brackish layer does not increase to seaward, and consequently to transport both the fresh water and the entrained salt water requires an increase in velocity. This, then, indicates that it is unrealistic to assume a constant velocity at all positions in the inlet. But even if the velocity is permitted to vary with position, the fitting of observed salinity curves to the family given by equation (21) would require that the velocity decrease to seaward for a constant eddy coefficient. The assumption of a constant eddy diffusivity is likewise unjustified and unrealistic and therefore it is necessary to have both the diffusion coefficient and the velocity a function of position in the inlet.

2) Fresh water bounded, eddy diffusivity and velocity a function of horizontal position.

The treatment in section 1) was seen to be unsatisfactory from a number of aspects.

If one assumes $u = u(x)$ and $A_s = A_0 f(x)$, then equation (7) becomes:

$$\frac{u(x)}{f(x)} \frac{\partial S}{\partial x} = A_o \frac{\partial^2 S}{\partial z^2} \tag{22}$$

Defining a new variable ρ such that

$$\frac{u(x)}{f(x)} \frac{\partial}{\partial x} = \frac{\partial}{\partial \rho} \tag{23a}$$

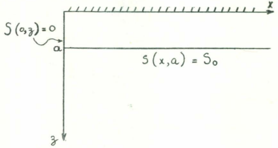
or

$$\rho = \int \frac{f(x)}{u(x)} dx + C \tag{23b}$$

Then equation (22) becomes:

$$\frac{\partial S}{\partial \rho} = A_o \frac{\partial^2 S}{\partial z^2} \tag{24}$$

For the following solutions in this section, the origin of the co-ordinate system will be taken to be at the surface at the head of the inlet with z positive downward.



Solving equation (24) using Laplace transform methods and the following conditions.

$$S(0,z) = 0 \qquad 0 \leq z \leq a$$
$$S(x,a) = S_0 \qquad \frac{\partial S(x,0)}{\partial z} = 0 \qquad (25)$$

gives:

$$S(x,z) = S_0 \sum_{n=0}^{\infty} (-1)^n \left\{ \operatorname{erfc}\left[\frac{(2n+1)a - z}{2\sqrt{A_0 \beta}}\right] - \operatorname{erfc}\left[\frac{(2n+1)a + z}{2\sqrt{A_0 \beta}}\right] \right\} \qquad (26)$$

This converges most rapidly for small values of β. (β has the dimensions of time). One which coverges most rapidly for large values of β is given by:

$$S(x,z) = S_0 \left\{ 1 - \frac{4}{\pi} \sum_{n=0}^{\infty} \frac{(-1)^n}{2n+1} \cos \frac{(2n+1)\pi z}{2a} \, e^{\frac{-(2n+1)^2 \pi^2 A_0 \beta}{4a^2}} \right\} \qquad (27)$$

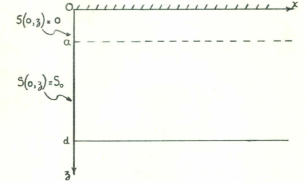
Since in the inlets, one is concerned mostly with relatively large values of β, only equation (27) has been evaluated (fig. 19). This is evaluated for values of the parameter $A_0 \beta$ ranging from zero to infinity, a = 6 metres, and S_0 = 30‰.

The solution is a marked improvement over equation (21) in that for the same value of eddy diffusivity, the salinity increases much faster with x at a particular depth. This is especially noticeable for large values of time (β).

3) Solution for inlet of finite depth.

The solution discussed in part 2) can be criticized because of the maintenance of salinity at a fixed value at the depth a.

Consider an inlet of depth d and the following conditions:



$$S(0,z) = \begin{cases} 0 & 0 \le z \le a \\ S_0 & a < z \le d \end{cases} \qquad \frac{\partial S(x,0)}{\partial z} = \frac{\partial S(x,d)}{\partial z} = 0 \tag{28}$$

A solution which satisfies equations (24) and (28) and converges most rapidly for large values of ξ is given by

$$S(x,z) = \frac{1}{d}\int_0^d S(\zeta,0)\,d\zeta + \frac{2}{d}\sum_{n=1} \cos\frac{n\pi z}{d}\int_0^d S(\zeta,0)\cos\frac{n\pi\zeta}{d}\,d\zeta\, e^{-\frac{n^2\pi^2 A_0 \xi}{d^2}} \tag{29}$$

Substituting for $S(\zeta,0)$ and integrating yields:

$$S(x,z) = S_0\left\{\frac{d-a}{d} - \frac{2}{\pi}\sum_{n=1}^{\infty}\frac{1}{n}\sin\frac{n\pi a}{d}\cos\frac{n\pi z}{d}\, e^{-\frac{n^2\pi^2 A_0 \xi}{d^2}}\right\} \tag{30}$$

This result is evaluated in figure 20, for $a = 6$, $d = 30$, $S_0 = 30‰$, and various values of $A_0 \rho$.

4) Bottom layer unbounded:

The results of part 3) are unrealistic in that as $z \to \infty$, $S \to 0.8 S_0$ instead of S_0. To remove this difficulty, and still not require any level to be maintained at a constant salinity requires extending the bottom layer to infinity.

The boundary conditions for the problem are now:

$$\left. \begin{aligned} S(0,z) &= \begin{cases} 0 & 0 < z \leq a \\ S_0 & a < z < \infty \end{cases} \\ \frac{\partial S(x,0)}{\partial z} &= 0 \end{aligned} \right\} \tag{31}$$

A solution which satisfies this is given by

$$\left. \begin{aligned} S(x,z) &= S_0 \left\{ 1 - \frac{1}{2} \left[\operatorname{erfc} \frac{z-a}{2\sqrt{A_0 \rho}} - \operatorname{erfc} \frac{a+z}{2\sqrt{A_0 \rho}} \right] \right\} \quad z > a \\ &= \frac{S_0}{2} \left\{ \operatorname{erfc} \frac{a-z}{2\sqrt{A_0 \rho}} + \operatorname{erfc} \frac{a+z}{2\sqrt{A_0 \rho}} \right\} \quad 0 < z \leq a \end{aligned} \right\} \tag{32}$$

where $\operatorname{erfc} z = 1 - \operatorname{erf} z$.

Equation (32) is evaluated in figure 21. It shows the desirable feature that $S \to S_0$ as $z \to \infty$, but the depth to which 'dilution' occurs for values of $A_0 \rho$ between 5 and 50 appears too great.

5) Salinity maintained constant at finite depth.

Comparing the results in parts 1) to 4) suggests that to obtain realistic results for the depth of penetration of the dilution, a level at some finite depth below z = a should be maintained at a constant salinity. To see the effect of this, consider equation (24) and the following boundary conditions

$$\left.\begin{aligned} S(0,z) &= \begin{cases} 0 & 0 \leq z \leq a \\ S_0 & a < z \end{cases} \\ S(x,d) &= S_0 \qquad \frac{\partial S(x,0)}{\partial z} = 0 \end{aligned}\right\} \tag{33}$$

Using separation of variable procedure, a solution which satisfies the conditions in (33) and converges most rapidly for large values of β, is given by:

$$S(x,z) = S_0\left\{1 - \frac{4}{\pi}\sum_{n=1}^{\infty} \frac{1}{2n-1} \sin\frac{(2n-1)\pi a}{2d} \cos\frac{(2n-1)\pi z}{2d}\, e^{-\frac{(2n-1)^2\pi^2 A_0 \beta}{4d^2}}\right\} \tag{34}$$

Figure 22 indicates the form of equation (34) for a = 6, and d = 18, and S_0 = 30‰. In figure 23 the results are illustrated for a = 6 and d = 12.

The results obtained for these idealized situations indicate a close agreement in many cases with the observed profiles. The rather close agreement in 'shape' suggests that the vertical variation of the eddy diffusivity may be of second order importance relative to its variation in the longitudinal direction.

Before any practical use can be made of the results of the above idealized considerations a knowledge of the velocity distribution is required so that $\int$ can be evaluated for each of the observed profiles. Once this can be done, then the form of the diffusion coefficient can be determined.

Evaluation of A_s from field data and application of equation (34)

Bute Inlet, May 29, 1952.

From a consideration of velocities calculated by heat budget (Section VII) together with those observed directly (Section V), and the salinity distribution, the time differences between an observed profile and that at the head were calculated for each station in the inlet. From figure 23, the values of $A_o \int$ which yielded profiles corresponding to those observed in the inlets were noted. A graph with $A_o \int$ plotted against t ($t = \frac{x}{u}$) was constructed (see fig. 24). An examination of equation (23b) indicates that a change in $A_o \int$ must equal the corresponding change in $A_s t$. Therefore the slope of the graph represents the values of A_s. Since $\int = o$ when $x = o$, the method can be applied to evaluate A_o (the value at the head of the inlet). The values of A_s are tabulated in Table XV. A_s ranges from 0.17 cm.^{2}/sec. at the head to 4.7 cm.^{2}/sec. at station 3. This compares favourably with the results of

Table II, where for similar velocities A_s has a similar range. (Explicit figures cannot be given since the vertical variation of velocity is not known).

Similar calculations were made for the observations taken in Knight Inlet in June, 1951 and in July, 1953. The graphs from which A_s is determined are illustrated in figure 24. The general form of the curves is seen to be similar to that calculated for Bute Inlet. The results of the calculation of eddy diffusivity, given in Table XVI, indicate that in June 1951, A_s was smaller at the head, than for July 1953, but that it increased faster towards the mouth. In both cases the range was of the same order of magnitude.

Although the close agreement between the theoretical and observed salinity curves suggests that the vertical variation of the diffusion coefficient may be small, especially in the brackish layer, the departure of the curves from reality in some cases, in the layer immediately below the halocline, may be due to this restriction of an A_s constant vertically. In addition to this, the variation of A_s longitudinally suggests that diffusion is related in some way to the vertical structure, for example the stability or the velocity gradient. The following section deals with the part these parameters may play in controlling vertical diffusion.

Variation of Vertical Diffusion Coefficients with Depth

Since it is well established that vertical mixing in a fluid is decreased with increased stability and increased with increased velocity shear, it would appear that the treatment of diffusion considering only constant eddy coefficients will be of limited value.

Fjeldstad (18) has demonstrated that the turbulence as expressed by the coefficient of eddy viscosity (A_V) decreases with increasing stability. He found that he could obtain satisfactory agreement between observed and computed tidal currents in shallow water by assuming that the eddy viscosity was a function not only of the distance from the bottom, but of the stability as well. He introduced the expression $A = f(z)/1 + aE$ where E is the stability of the water column defined as $E = \frac{g}{\rho} \frac{\partial \rho}{\partial z}$ (where ρ = density), and the factor a was determined empirically. The relationship between eddy viscosity and eddy diffusion has been examined by Taylor (19). He treated the kinetic energy of a turbulent medium as composed of two parts, one associated with the mean motion and the other with the turbulent components of the motion. He showed that the rate at which turbulent energy enters a unit volume is $A_V\left(\frac{\partial u}{\partial z}\right)^2$ when u is the mean velocity. This energy used in increasing the potential energy of the column is given by $\frac{A_S g}{\rho} \frac{\partial \rho}{\partial z}$. If turbulence is to continue then the rate which turbulent energy (Reynolds stresses) is supplied to a region must be greater than the rate at which the potential energy of the region increases. This condition for the maintenance of turbulence may be written as:

$$A_v\left(\frac{\partial u}{\partial z}\right)^2 \geq \frac{A_s\, g}{\rho}\; \frac{\partial \rho}{\partial z}$$

$$R_i = \frac{g\, \partial\rho/\partial z}{\rho\left(\frac{\partial u}{\partial z}\right)^2} \leq \frac{A_v}{A_s} \tag{35}$$

This non-dimensional number is referred to as the 'Richardson number'. Richardson (20), who was the first to employ the above expression, derived the result from consideration of the supply of energy to and from atmospheric eddies. He assumed that the coefficient of eddy viscosity and diffusivity were equal in magnitude for the atmosphere. Consequently, the criterion for the maintenance of turbulence was taken to be $R_i \leq 1$. (above this value, turbulence will die out). More recent investigations (21) have indicated that the criterion should take the form that turbulence will decay when the Richardson number is greater than 1 - a where a is a small positive quantity whose value has not been determined in general.

In the ocean however, the magnitude of A_v to A_s may be considerably different from unity, and normally A_v is considerably larger than A_s. The physical explanation usually given is that the elements in turbulent motion can give off their momentum rapidly to the surroundings, whereas the other properties such as salt or heat, cannot be transferred so rapidly in bulk, and ultimately must be transmitted by the slower molecular processes. Observations in the sea indicate that the medium is generally turbulent, because the observed changes cannot be accounted for by molecular processes alone. Consequently the critical value of the Richardson number for determining whether turbulence will increase or be maintained will be considerably larger than unity (only unity if the medium is homogeneous in all respects). Alternately, assuming the medium is turbulent at all times, the Richardson number gives a measure of the ratio of $A_v : A_s$. Taylor (19) calculated the Richardson number, from Jacobsens data (22) for Schultz's Grund, and found A_v to be from 10 to 50 times larger than A_s.

A summary of the values of the Richardson number calculated for Bute and Knight Inlets will be presented in a later section, (VIII) after the velocity observations have been presented.

Mixing and stability near a boundary

In a search for ideas and methods of attacking the problem of mixing, considerable time was devoted to meteorological literature, since in certain respects the problems are similar to those in oceanography.

Meteorologists have devoted considerable attention to the nature of the flow of air near the surface of the earth. The region is a relatively complicated one and turbulence will depend on a number of factors such as the roughness of the boundary, the vertical stability of the air column, and the vertical velocity gradient.

Using the Prandtl mixing length concept (12), Rossby and Montgomery (23) considered that the effect of a stable density gradient is to reduce the mixing length from its value in neutral equilibrium ℓ_o, to a value $\ell_s < \ell_o$. If $\frac{\partial \bar{u}}{\partial z}$ is the mean velocity gradient for neutral equilibrium and $\left(\frac{\partial u}{\partial z}\right)_s$ that for stable equilibrium, the corresponding turbulent kinetic energies are proportional to $\ell_o^2\left(\frac{\partial u}{\partial z}\right)^2$ and $\ell_s^2\left(\frac{\partial u}{\partial z}\right)_s^2$ respectively. The decrease in the kinetic energy of the fluctuations is proportional to the work done against gravity in the vertical movement of eddies in the stable state.

Rossby and Montgomery express this in the equation:

$$\ell_o^2\left(\frac{\partial u}{\partial z}\right)^2 - \ell_s^2\left(\frac{\partial u}{\partial z}\right)^2 = \beta \frac{g}{\rho} \ell_s^2 \frac{\partial \rho}{\partial z} \tag{36}$$

where β is considered to be an absolute constant. Hence $\ell_s = \frac{\ell_o}{\sqrt{1+\beta Ri}}$, or expressing this in terms of diffusion coefficients:

$$A_s = \frac{A_o}{\sqrt{1 + \beta Ri}}$$

Sverdrup (24) made an extensive study of the velocity profiles and eddy conductivity over a snow field, and deduced a value for β of 11. He found that in all values of stability observed, β was very nearly a constant, and suggested the possibility that β was a universal constant.

Munk and Anderson (25) applied the expression derived by Rossby and Montgomery to describe the vertical variation of eddy viscosity and eddy conductivity. They substitute these into differential equations governing the distribution of temperature and current, and solve the equations simultaneously, to obtain an expression for the depth of the thermocline. Comparison of the theoretical results with observations gave a thermocline depth which, although too shallow, was of the right order of magnitude. In this development Munk and Anderson used a value for β_V (viscosity) of 10 and β_T (conductivity) of 3.33. Their results then, support partially at least, the value of 10 for β which Sverdrup reported. However more recently Deacon (26) used profiles of velocity and density over short grass and in a desert, and found that β increases systematically as the stability increases.

The conclusion which one may reasonably draw from the extensive examination of β is that it is really only an empirical constant which allows, for a particular situation, a relationship between the kinematics and the distribution of property to be ascertained. The incorporation of the Richardson number into the eddy coefficients has shown in part upon what the coefficients depend, but the fact that is not constant, indicates that a more complicated function is required to satisfactorily describe the eddy diffusion processes.

The Richardson number, together with β will be considered further in Section VIII, and applied to determine the eddy diffusivity as expressed by Rossby and Montgomery.

Solution of diffusion equation for coefficient varying as a power of depth.

The number of equations which can be solved analytically for a variable diffusion coefficient is very limited. For most choices of A_s, the equation is non-linear and the usual difficulties ensue. However, there are a limited number of cases which can be solved. One such type is for a diffusion coefficient varying as a power of the depth. In the simplest cases, the resulting equation can be transformed into Bessel's equation. If the eddy diffusivity is chosen to be of the form $A_s = A_o z^n$, then in general one will be led to the 'non-homogeneous' Bessel equation (i.e. right hand side not zero). The method of solution for the homogeneous part of the equation is given in Carslaw and Jaeger (13). The particular integral, which satisfies the non-homogeneous equation can be found by applying 'variation of parameter' technique.

This approach was applied using small powers of n, and very simple boundary conditions, but the results indicated very little improvement over the results assuming A_s constant, vertically.

If the method can be applied for a more realistic choice of A_s together with boundary conditions, then the solution obtained would be a considerable advance over those discussed for A_s constant vertically. However, no solutions were obtained by this method which had a general enough form of A_s together with suitable boundary conditions.

V. CURRENT MEASUREMENTS

In the introductory section a general description of the circulation was given. In this section a more complete analysis will be given of the direct observations made in the mainland inlets.

Only a limited number of direct current observations are available and these are for only three of the inlets -- Bute, Knight, and Loughborough. The measurements were made with Ekman current meters for the deep casts and C.B.I. (Chesapeake Bay Institute) type current drags for the shallow depths.

Knight Inlet

The most complete set of measurements which were made was in Knight Inlet between July 11th and July 15th, 1953. More than 70 profiles were taken at station 4 during the period. They were distributed approximately equally over the tidal cycle. During this period the water was moving seaward at the surface at all times (see fig. 25a,b, c). Typical profiles for various phases of the tide are shown in figures 25a and 25b. Two for each stage of the tide are represented. Mixed tides existed during this period, consequently one ebb tidal range was larger than the other. The profiles are numbered from '(1)' to '(8)' inclusively and cover two tidal cycles (i.e. two ebbs and two floods). One of the interesting features of these curves is the minimum velocity layer centred at about 6 metres depth. This characteristic persists during most of the time that the tide is ebbing and is evident as well during part of the time during which the tide is flooding. On the average much larger velocity gradients exist during the flood tide period in the 3-6 metre layer. The variation in velocity shear over a tidal cycle indicates that the tidal velocities are not uniform throughout the column. In figure 25c mean velocity profiles are presented. Although the minimum velocity at 6 metres depth in the individual profiles suggested the possibility of an up-inlet movement, the net velocity profile (average of the individual profiles) does not indicate a resultant up-inlet velocity at this depth. However it does decrease to zero and then increases below this reaching a secondary maximum at 6-9 metres. Below 9 metres it decreases to zero and is negative below 11 metres. Measurements were made using a current drag and consequently the velocity at the minimum would be recorded as larger (positive seaward) than in reality due to the drag exerted on the wire which passes through the swift seaward moving surface layer. The low up-inlet recorded velocities in the layer below 11 metres is likewise probably less than the real value, due to the same effect. (An error of 1° in wire-angle reading gives a variation in velocity calculations of from 10 cm./sec. to 2.5 cm./sec. for an angle reading varying from 1° to 10° respectively).

The mean flood and ebb 'tidal' velocities (fig. 25c) show considerable vertical variation. (By 'tidal' is meant the mean velocity during the ebb or flood tide minus the net velocity). During both the flooding and ebbing tide the velocities are smaller at the surface than at depths above the 20 metre level at least. The reason for this is not readily apparent.

A calculation based on the volume of the tidal prism above station 4 and the time required to fill it, yields a relationship between the mean velocity (v_i) and the depth (d_i) of the inflowing water given by:

$$v_i d_i = 18.4 \ M^2/sec.$$

In the top 55 metres, the average flood tidal velocity is approximately 40 cm./sec. Therefore to supply the required volume, d_i = 46 metres. This indicates that the velocity is not uniform across the inlet. A mean value of 25 cm./sec. gives a d_i = 75 metres, which is probably a realistic value.

The mean salinity profile (fig. 25c) indicates that the halocline depth corresponds approximately to the depth of maximum velocity shear. If the layer at 6 metres is considered to be moving up-inlet, then it suggests that at least some of the water which dilutes the seaward moving surface water has a salinity of less than 30‰. The up-inlet movement suggests that an up-inlet pressure gradient probably exists at this level. This is substantiated in part by the salinity observations which indicate, in a considerable number of cases, an increase in salinity towards the head at this intermediate depth.

Between August 6-8, 1952 a number of velocity measurements were made at the same position in Knight Inlet (station 4). The mean ebb and flood tidal velocities together with the net velocity and salinity profiles are illustrated in figure 26. These are considerably different from the observations presented previously, although the existence of smaller tidal velocities at the surface than just below it is noted in both cases. The net velocity in the brackish layer (average of 12 cm./sec.) is smaller than for the previous observation (average of 27 cm./sec.), but the thickness of the seaward moving layer is greater, being 9 metres deep. (The thickness for the July observations is taken as 6 metres). The net up-inlet velocity between 9 and 20 metres is 8 cm./sec.

A comparison of the mean salinity profile (fig. 26) with the net velocity profile indicates the halocline to be at a slightly smaller depth than that of maximum shear. The water moving up-inlet at this time evidently has a salinity greater than 28‰.

Bute Inlet

Current measurements were made in Bute Inlet on several occasions, but considerable difficulty was experienced both in obtaining and in averaging the data. The measurements made both in 1952 and 1953 are similar in that wide fluctuations in magnitude and direction were observed. Many of the measurements indicate cross-inlet flow, and the majority have at least a cross-inlet component. The individual profiles of velocity did not lie in a plane (i.e. the horizontal flow all in the same or exactly opposite direction), but very often the points of the vectors traced out an irregular spiral with depth. Hence to average the observations required breaking the individual observations at each depth into two components at right angles to each other, and then averaging the components. Finally the averaged components were combined again to give the mean velocity at each depth.

Between August 10-12, 1952, station 4 in Bute Inlet was occupied. The net velocity in the surface layer is seaward, with a magnitude of 24 cm./sec. while the up-inlet velocity between 4 and 20 metres has a net value of 5 cm./sec. (see fig. 27). The depth of no net motion is at 4 metres. This depth coincides with the depth of maximum shear and with the halocline. Mean tidal velocities, which have maximum at the surface of 22 cm./sec., decrease to 8 cm./sec. at 4 metres and remain nearly constant with depth down to at least 20 metres.

The maximum tidal range at this time was 3 metres, with the mean range slightly less. A calculation based on the tidal prism volume yields a relationship between the velocity and thickness of the inflowing water, $v_1 d_1 = 6.4\ M^2/sec.$ Choosing $v_1 = 8$ cm./sec. gives a depth of tidal inflow of 80 metres.

Velocities of 50 cm./sec. or more at depths greater than 100 metres are not uncommon. However they cannot possibly continue during the entire flooding or ebbing tide, and consequently must have the character of a jet. These jets probably extend only over a portion of the width of the inlet.

Between July 1st and 3rd, inclusive, 1953, station 4 was occupied again. The net circulation (fig. 28) indicates a mean seaward velocity in the top 4 metres of 20 cm./sec. The depth of no net motion is at 4 metres and between 4 and 20 metres the mean net velocity is 7 cm./sec. As in the previous case, the depth of maximum velocity shear coincides approximately with the halocline. Tidal velocities are nearly uniform in the top 20 metres, but the mean ebb tidal velocity is slightly larger than for that of the flood tide. It is doubtful if this excess would be compensated for at deeper levels since this would require a strong vertical circulation and it is difficult to see how the observed salinity structure could be maintained for this situation. It is more likely that the number of observations used was inadequate to give a reliable mean value. That this is probably the case is further indicated by the fact that the net up-inlet transport of the deeper water slightly exceeds the seaward transport in the brackish layer. The reverse situation must actually exist, in order to account for the net movement of fresh water seaward.

Miscellaneous Observations

A few current measurements were also made both in Bute and Knight Inlets in June 1952. Although insufficient data are available to determine the net circulation accurately, the few profiles taken indicate that the depth of no net motion at station 4 in Bute on June 7-8 was at 4-5 metres. The direction of flow in the surface layer was seaward at all times, with an average of approximately 30 cm./sec. The measurements in Knight at station 4, June 3-5, likewise indicate a depth of no net motion at 4-5 metres, and a surface layer moving seaward at all phases of the tide.

Current measurements were undertaken at two locations in Loughborough Inlet between July 20-23, 1953. At station 3, which is near the head, the currents were weak and variable and ship movements during measurements were usually large enough so that the observations indicated predominantly the ship's motion. The ship was anchored at station 1, where tidal velocities are larger, but difficulties in keeping the ship steady made it impossible to obtain many reliable current measurements.

During 1953 observations of the surface currents were made over the full width of both Knight and Loughborough Inlets. The method consisted in photographing at regular intervals the position of fluorescein dye patches laid in a line across the inlet. These observations showed that the main current was often limited to only a portion of the width of the inlet. Marked jet effects were observed even in straight reaches with substantially parallel sides. Often the bulk of the flow was in mid-channel, so that the effective width of the inlet was much less than the geographic width. These observations indicate that it is not surprising that the velocities measured at one position near mid-channel are found to be considerably higher than the average for the entire cross-section. It is very probable that similar effects occur below the surface to account for the fast currents which have been observed at times in the deep water.

VI. PRECIPITATION AND RIVER DISCHARGE

The volume of fresh water discharged into an inlet must play an important role in determining the distribution of salt in the upper layer, the thickness of the brackish layer, and possibly the nature and intensity of the mixing process. It is the object of this section to evaluate quantitatively the effect of this factor and to compare it from inlet to inlet.

As mentioned previously, the inlets vary in length from 5 to 90 miles. A qualitative inspection of the salinity data for the inlets indicates that the surface layer salinity in summer is approximately zero at the head and increases to seaward, reaching a value at the mouth which is roughly the same for all inlets. If this is actually the case, then it suggests either one of two possibilities:

1) An inverse relationship exists between length of inlet and intensity of mixing, so that for a given volume of fresh water, the longer the inlet, the less intense is the mixing process. This implies that the mouth of the inlet exercises a control on the dynamics and circulation within the entire inlet. The concept of a control exercised on an estuarial circulation by a certain section in the estuary has been emphasized in theoretical papers on the dynamics of estuarial circulation by Cameron (17) and Stommel (27). Although the method of approach is quite

different in the two cases, both lead to the concept of a limiting or critical velocity in the surface layer. It is possible that the section where this critical velocity is reached is at or near the mouth, where the inlet suddenly widens as it connects with the open coastal waters.

or ii) The fresh water discharges are by coincidence related in some way with the lengths of the inlets. For instance, the total discharge might be directly proportional to length, or the drainage basins of the rivers flowing into the longer inlets might be more extensive, or have a higher precipitation than those of the shorter inlets. Alternatively the arrangement geographically of the rivers surrounding the inlets may be peculiarly spaced so that the distribution of fresh water in the inlets appears similar in all cases.

To investigate this point it was necessary to obtain information on the discharge of fresh water into the inlets. Since only very few of the numerous rivers flowing into the inlets are or have been metered, a method had to be devised for evaluating the total run-off. The Water Resources Division of the Department of Resources and Development, who undertake the responsibility of collecting all fundamental hydrometric data in Canada, have discharge measurements for the larger rivers draining into the coast, but only very few of the numerous small rivers have been metered.

To evaluate the discharge for the unmetered rivers recourse was made to direct precipitation observations. From a knowledge of the average annual precipitation over an area one can estimate the average discharge of water from the rivers draining the area. The precipitation observations (from approximately 35 stations) were taken from the ' Climatic Summaries ', Meteorological Division, Department of Transport (28) and ' Climate of British Columbia ' tables, Provincial Department of Agriculture (29).

Since the coastal drainage basin area is largely uninhabited, the number of stations making precipitation observations is limited. One is faced with the problem of determining the precipitation distribution from very widely spaced observations. To supplement the direct measurements the river discharge measurements, where available (approximately 25 in number), were converted to the equivalent average annual precipitation on their drainage basin. This figure represents the average precipitation over the entire drainage basin, and wide variations in precipitation within the area are possible. However, it does have the effect of smoothing the data which is to be desired since it is well known that precipitation is often very localized.

Evaporation of water from a drainage basin may be an important factor in an area of small rainfall in reducing the amount of water reaching the rivers. However, in the Coast Range the precipitation on the average is so high that, on a relative basis, the quantity of water lost in evaporation is considered to be negligible.

Very little is known about the variation of precipitation with altitude, or the effect on precipitation of the configuration and orientation of the land with respect to the general movement of the air masses. River discharge measurements aided in removing this difficulty in a number of cases where the precipitation was known near a metered river from direct observations.

The presence of glaciers at certain elevations in some areas and not in others with the same or higher elevation was taken as an indication of a higher precipitation over the former. Since one would expect that equivalent elevations in about the same latitude and degree of shelter should have approximately the same temperature variation, the fact that snow and ice persist at all times of the year in the one and not the other, suggests that the rate at which it is being deposited in the glacier area is the greater of the two.

A simplified diagram of the average annual precipitation in inches for the British Columbia Coast Range, based on the available observations and the methods outlined above, is illustrated in figure 29. In general the precipitation at mainland coastal points is less than 100 inches per year. Inland from the coast the precipitation usually increases and reaches a maximum which varies from 100 to nearly 300 inches, and then decreases quite abruptly east of the coast range to values of less than 20 inches per year in the central interior of the province.

The maximum precipitation reached between the coast and the interior is usually about 150 inches per year. The regions with 150 inches per year or greater are usually in the form of isolated cells and may in some cases cover an extensive area. The largest area with the heaviest precipitation is in the central coastal region between Dean Channel and Gardner Canal. The precipitation is greater than 150 inches over most of this area and the maximum, calculated from discharge measurements of the Nascall River, located on Dean Channel, is approximately 290 inches per year.

From the average annual precipitation chart, the average yearly discharge of water from each drainage basin was calculated. Thus the total average discharge was determined for each inlet. In most cases it was possible to go a step further by splitting the annual discharge into monthly figures. This was done by using the month by month discharge figures for such rivers as are or had been metered in the particular region. If the drainage basin covered a similar area to that of the metered river, then it was assumed that the monthly variations in river discharge were similar for the two areas. The appropriate ratio factor then enabled one to express the monthly mean discharge for the drainage basin in question. A number of small rivers along the coast are metered, or were for a number of years, so that in most cases discharge measurements for a river near to the particular drainage basin, and draining a similar type of terrain, was available. In this way tables were built up for all the major mainland inlets in the coast from Howe Sound to Portland Canal, giving monthly and yearly mean discharge figures.

The drainage basin above the head of the inlets, drainage basin at the sides (where appropriate), total drainage area, and the corresponding average discharges are given in Table XVII. All the records for the metered rivers are published in British units. For case of comparison with future data and with rainfall figures, the discharges in this report are given in the same units, cubic feet per second. The mean discharge varies over a wide range. The maximum for the areas considered is for the Portland Inlet -Portland Canal system, where a mean discharge of 56,000 cu. ft./sec. was calculated.

Although this table is incomplete, it does indicate a number of interesting features. First to be noted is the proportion of fresh water which enters the inlet below the head. In most cases the relative quantity is significant. Of those tabulated, the percentage of side runoff is usually 20% or more of the total discharge. In certain cases however, it exceeds 50%, and in exceptional cases such as Call Creek (not tabulated) it may exceed 90%.

Another feature indirectly brought out in this table is the relatively greater importance of the coastal region (Coast Range Mountains) compared to interior localities in supplying water to the inlets. In a number of cases the rivers flowing into the head drain large areas which have relatively low precipitation, whereas those emptying in from the sides drain high precipitation areas. An extreme example of this situation is found in Dean Channel. The drainage basin of the Nascall River is only about 3% of that of the Dean River, but the discharge is nearly 50% of that of the Dean. The contribution of local discharge into the sides of Dean Channel, above station 6, is approximately 55% of the total. Of the total drainage area, the Dean River drains about 75% of it, but contributes only 25% of the water.

In Table XVIII the mean monthly discharge of fresh water into the inlets is given. In general two yearly peaks occur. The larger is in late spring or early summer and the smaller in late autumn. The first is associated with the melting snow and ice in the interior and higher elevations which has been stored during the winter months. The second maximum is associated with the heavy autumn rains. Rainfall observations indicate a maximum in the autumn or early winter months (the time of maximum rainfall gets progressively earlier from south to north). This means that in most instances the local side drainage into the inlets usually reaches a maximum in autumn or winter because there is little time delay between time of rainfall and its discharge into the inlets. Inlets which receive fresh water from drainage areas predominantly in this region will have a minimum fresh water discharge during the summer months, and a broad maximum during autumn and winter months. Those whose drainage basins store snow or ice will have a monthly minimum fresh water discharge in March or April and a sharp maximum during late spring or early summer. For most inlets the resulting discharge is a combination of these two factors with the spring maximum usually the greater of the two.

If a relationship exists between the mixing process, the length of the inlet, and the fresh water discharge, the total discharge figures for the inlets need to be expressed as a function of the size of the inlet. In Table XIX the mean monthly and yearly discharge for each inlet is given for the total drainage per square mile of inlet surface. The mean yearly values vary by a factor of nearly 10. The largest is for Portland Inlet with a value of 557 cu. ft./sec./sq. mile of water surface, and the smallest is for the Jervis-Sechelt system with a value of 61.4 cu.ft./sec./sq. mile. The maximum monthly variation occurs in August when one is nearly 65 times the other. The lengths and areas of the inlets used in these calculations are given in

Table XVII. In some inlets it is impossible to determine the location of the mouth and therefore the lengths and areas used may not be the proper ones.

In Table XX discharges per unit width are recorded. The range of values is slightly less than for discharge per unit area, varying by a factor for the mean yearly discharge of about 8 or 9, and the variation for individual months by about 40. The variations from inlet to inlet indicate that it is unlikely that a coincidental relationship exists between total discharge per unit width and the length of the inlet. On the other hand it cannot be definitely established that the mouth exerts a control on the mixing process. Even if a relationship between length and mixing did exist, only a very limited correlation could be expected from these tables since no attempt was made to evaluate the variation in side run-off along the inlet. In this study the mouth of an inlet was chosen at the position where the inlet widens abruptly. It is conceivable that some other criterion, such as salinity might be more appropriate in certain cases. As an extreme example, the mouth of North Bentinck Arm was chosen as the junction point with South Bentinck Arm. The salinity of the brackish layer at this position (see fig. 11a) was very little different from that at the head, and if salinity were the basis for determining the mouth, then some position further down the inlet system would have been chosen.

Despite this however, the figures do suggest certain possible relationships between fresh water discharge and salinity distribution. For example salinity profiles taken in Loughborough Inlet in July 1953 (fig. 9b) indicate a general agreement in longitudinal range from head to mouth similar to those taken in Bute in May, 1952 (fig. 7a), but the thickness of the brackish layer is much smaller in Loughborough than in Bute. A comparison of discharge/unit area figures (Table XIX) indicates values of about the same order of magnitude, whereas comparison of discharge/unit width figures (Table XX) indicates that Bute has nearly double the value for Loughborough. On the other hand, when Bute and Knight are compared, both the salinity profiles and the discharge figures per unit width and unit area are similar. It is possible then that the range of salinity from head to mouth within the brackish layer is related to the discharge per unit area, whereas the thickness of the layer is related to the discharge per unit width.

In correlating the figures in the tables with the observed salinity distributions it should be remembered that the tables represent averages over a number of years, whereas the salinity profiles represent the distribution at a particular time. An examination of river discharge measurements indicates that inflow varies widely from year to year for corresponding times, and also in any one year, a wide fluctuation can occur in a few days. It is difficult then to check the results of Tables XIX and XX with observed salinity distributions in any very precise manner. However a more careful scrutiny of the possible relationships between these quantities should be made. A further breakdown of discharge calculations so that variation along the inlet is expressed, together with a more careful consideration of the location of the mouth of the inlet or system would in all probability throw more light on the interrelationship between discharge and mixing mechanism.

To summarize the results of this aspect of the study it is emphasized that for most inlets a significant proportion of the fresh water enters from the sides. This is in contradiction to the assumption made in theoretical investigations to date which have considered the fresh water all to be entering at the head. In future theoretical discussions this point should be borne in mind when considering the continuity of fresh water transport in an inlet.

Total fresh water discharge, and likewise the discharge per unit area and per unit width, vary from inlet to inlet. No coincidental relationship between discharge and length is evident. On the other hand the suggestion of a control placed on the mixing process by the mouth or some section of the inlet cannot be definitely proved or disproved from the results. Further salinity data are necessary, as well as a more critical study of what location in the inlet might exert the control.

Further use of the results of this section will be made for comparison purposes in the following sections.

VII. TEMPERATURE DISTRIBUTION AND HEAT BUDGET

The distribution of temperature in the upper 20 metres, mentioned briefly in Section I, will be considered in some detail in this section.

The processes tending to modify the distribution of heat in the brackish layer can be divided into two groups: external processes, active only at the boundary surface of a fluid, and internal processes active anywhere in the fluid. Away from boundaries the processes tending to alter temperature and salinity are considered to produce the same effect on each property.

Most studies of estuaries have considered only the conservation of salt and conservation of volume in determining the circulation and mixing process. Usually no use has been made of the principle of conservation of heat because of the complication that heat can flow through the sea surface and therefore within the water it is not a conservative property. However, if the quantity of heat passing through the sea surface can be determined, the principle can be applied and valuable information gained from it.

The distribution of temperature in a number of inlets is illustrated in parts (b) and (d) of figures 10 to 14. In parts (a) and (c) of these figures the corresponding salinity profiles are illustrated. In general the thermocline and halocline correspond in depth, although often the thermocline is less pronounced than the halocline. As a rule in summer months the temperatures of the brackish layer increases to seaward, although in certain cases a decrease to seaward over part of the inlet does occur.

In spring and summer months the heat absorbed by the water from solar radiation and conduction exceeds that lost to the atmosphere. Consequently the water column is heated. On the other hand, the salt water entrained into the surface layer from below is cooler than the brackish layer and the column is cooled. The decrease in temperature to seaward in certain instances indicates that the cooling brought about by the entrained water has exceeded the heating due to insolation.

A study of the heat budget for a particular body of water, provides valuable information about its mean motion and therefore the average currents in a particular area. Information about mixing and the depths from which the entrained water is drawn can also be obtained. In addition a knowledge of the heat flux through the sea surface together with the vertical temperature gradient permits the eddy conductivity to be determined for the brackish layer.

In the heat budget analysis presented here three principles are involved--the conservation of volume, salt, and heat. The complete equation expressing a heat balance in any part of the sea in a particular time interval is given by (6, p. 101):

$$Q_s - Q_b - Q_e - Q_h + Q_\theta + Q_v = 0 \qquad (37)$$

where:

Q_s = heat received by insolation

Q_b = back radiation from the sea surface

Q_e = heat required for evaporation

Q_h = convection of sensible heat to atmosphere

Q_θ = heat required locally to change the temperature of the water

Q_v = heat brought into or out of the region by currents or processes of mixing.

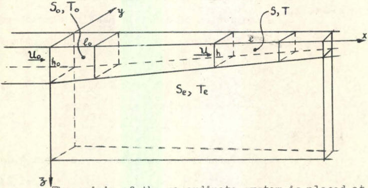
(Q's given in cal./cm.^{2}/min.)

If all the quantities in equation (37) but Q_v are known, then the rate at which heat is removed or supplied to an area can be determined from the equation.

In order to apply equation (37) the following observations are required: air temperature, relative humidity, wind speed, cloud cover, and the vertical temperature profile in the sea. In addition it is necessary to know the daily total solar radiation striking the earth's atmosphere.

Using the equations given in <u>The Oceans</u> (6, Ch. IV), the quantities Q_b, Q_e, etc. can be calculated, and therefore the net amount of heat passing through the sea surface can be evaluated. But to estimate the currents from this result requires some information about the depth to which the heat is penetrating beneath the surface, and in addition a knowledge of the quantity of heat brought into the surface layer by entrainment.

In order to see how these quantities are related consider the following model:



The origin of the co-ordinate system is placed at the head of the inlet, with x directed down the inlet, y across and z positive downward.

Consider a column of water at $x = 0$, with unit width, length ℓ_o, and a depth h_o. Suppose it to have a velocity u_o in the x direction, a salinity s_o and temperature T_o. For any position x, the quantities u, s, T, h, and ℓ describe the column of water. The quantities in all instances refer to mean values for the entire column. Assume that the quantity of heat per unit time passing through a unit area of sea surface and remaining is constant with time and position and is equal to Q cal./cm.^{2}/sec.

For any position x, the quantity of heat, δ H passing down through a surface of unit width and length ℓ in a time interval δ t is given by:

$$\delta H = Q \ell \delta t \tag{38}$$

Integrating gives:

$$H = Q \int_0^t \ell \, dt \tag{39}$$

At position x, the original volume $h_o \ell_o$ will have been increased by entrainment to $h\ell$. The quantity of water entrained may be represented as $h_1 \ell$ so that the following continuity of fresh water conditions must hold for a constant width:

Volume:

$$h_o \ell_o = (h - h_1)\ell \tag{40a}$$

Transport:

$$u_o h_o = u(h - h_1) \tag{40b}$$

Dividing equation (40a) by (40b) yields:

$$\frac{\ell}{u} = \frac{\ell_o}{u_o} \tag{41}$$

Changing variable from t to x in equation (39) by making use of the relation, $u = \frac{dx}{dt}$ gives:

$$H = Q\int_{o}^{x} \frac{\ell}{u}\, dx \tag{42}$$

Substituting (41) into (42) and integrating, gives:

$$H = \frac{Q \ell_o x}{u_o} \tag{43}$$

Equation (43) indicates that for a given initial velocity u_o and length ℓ_o, the quantity of heat entering the water mass is directly proportional to the distance traversed by the volume of water that has passed through the position $x = o$, and is <u>independent</u> of any change in thickness of the layer, or of entrainment.

The equation expressing the conservation of heat for the brackish layer can be represented as:

$$h_o \ell_o (T - T_o) + h_1 \ell (T - T_e) = \frac{Q \ell_o x}{u_o} \tag{44}$$

where T_o is the temperature of the entrained water.

For a given value of h_o, x, T_o, and u_o, equation (44) can be satisfied by all combinations of $h_1 \ell (T-T_e)$ = constant. However, the entrained water has carried salt as well as heat with it, and hence the volume of water of salinity S_e, entrained will also be $h_1 \ell$.

The equation for the continuity of salt is given by:

$$S_o h_o \ell_o + S_e h_1 \ell = S(h_o \ell_o + h_1 \ell) \tag{45}$$

where S_e = salinity of entrained water.

Solving for $h_1 \ell$, gives:

$$h_1 \ell = \frac{h_o \ell_o (S - S_o)}{S_e - S} \tag{46}$$

Substituting (46) into (44) gives:

$$h_o(T - T_o) + \frac{h_o(S - S_o)(T - T_o)}{S_e - S} = \frac{Qx}{u_o} \tag{47a}$$

or:

$$\frac{T - T_o}{T - T_e} + \frac{S - S_o}{S_e - S_o} = \frac{Qx}{h_o u_o (T - T_e)} \tag{47b}$$

This equation can be applied to the field data in two ways. First, if the salinity and temperature of the entrained water are known, the velocity u_o of the water entering the section can be calculated. In principle this can be applied all the way down the inlet to determine the velocities at each point. Secondly, if the velocity of the top layer is known, a relationship between the temperature and salinity of the entrained water is given. If all quantities but T_e and S_e in equation (47b) are known, then the equation is of the form:

$$T_e = C_1 + C_2 S_e \tag{48}$$

which indicates a linear relationship between the temperature and salinity of entrained water.

Since the entrained water carries both its temperature and salinity with it, the depth from which it is drawn can be determined by plotting equation (48) on the same graph as the T-S curves for the stations of the inlet in question. The point where the straight line crosses the T-S curves represents an acceptable depth. The range of depths where the line lies between the T-S curves also represents possible values.

If the range of possible depths from which the entrained water could be drawn was extensive, a further elimination of the possibilities could conceivably be obtained by considering the densities of the mixtures involved.

Penetration Downward

The previous development considered only entrainment upward. If there is downward penetration of fresh water into the lower layer then equation (44) will be modified.

Let the subscript "d" denote the lower layer, $h_{11}\ell$ the volume of water penetrating into the lower layer, and T_p and S_p the temperature and salinity respectively of the penetrating water.

The equation for the conservation of heat in the brackish layer becomes:

$$h_1\ell(T-T_e) + h_o\ell_o(T-T_o) - h_{11}\ell(T-T_o) - h_{11}\ell(T_d-T_p) = \frac{Q\ell_o x}{u_o} \quad (49)$$

Equations for conservation of salt:

For upper layer:

$$h_o \ell_o S_o + h_I \ell S_e - h_{II} \ell S_p = h_o \ell_o S + (h_I - h_{II}) S \tag{50}$$

For lower layer:

$$h_{d_o} \ell_o S_{d_o} - h_I \ell S_e + h_{II} \ell S_p = h_{d_o} \ell_o S_d - (h_I - h_{II}) S_d \tag{51}$$

Rearranging equations (50) and (51) and solving for $h_I \ell$ and $h_{II} \ell$, one obtains:

$$h_I \ell = \frac{h_o \ell_o (S_o - S)(S_p - S_d) + h_{d_o} \ell_o (S_{d_o} - S_d)(S_p - S)}{(S_e - S)(S_p - S_d) - (S_e - S_d)(S_p - S)} \tag{52a,b}$$

and

$$h_{II} \ell = \frac{h_o \ell_o (S_o - S)(S_e - S_d) + h_{d_o} \ell_o (S_{d_o} - S_d)(S_e - S)}{(S_p - S)(S_e - S_d) - (S_p - S_d)(S_e - S)}$$

Substitution of (52a,b) into (49) gives the equation which must be satisfied.

For a given value of the variables u, x, S, T, etc., the expression gives a combination of values for S_e, S_p, T_e, and T_p which will satisfy the equation. However, as in the previous case, the only allowable solution will be one where the depth from which the entrained heat and salt is drawn coincide; similarly for the penetration of heat and fresh water.

In this study, the downward penetration of fresh water was found to be small and hence the equation for entrainment only was applied (47b).

There are a few points which should be noted when applying the data to the equations derived for the heat budget. First it requires a knowledge of the depth of no net motion. The direct current measurements (Section V) reveal that it corresponds approximately with the halocline and thermocline. This is further substantiated by the calculations of the expression A_z/u which reveal the depth of no net motion to be at the inflection point in the salinity curve (Section IV). It appears then, that the halocline or thermocline depth approximates closely the depth of no net motion in most cases. Secondly, near the head of the inlets the salinity of the surface layer is very low, and the increase (in absolute value) in salinity between stations in the upper reaches of the inlets is small. Consequently, when applying equation (47) only a small error is introduced if incorrect values are chosen for the salinity and temperature of the entrained water (S_e and T_e). However, as one examines the profiles to seaward, the importance of entrainment becomes very apparent, and the temperature and salinity of entrained water needs to be known more accurately. Therefore, since the velocities calculated for the upper reaches of an inlet appear to be quite reliable, these values should be applied to the equation (47) for regions where entrainment is important, to determine the depth from which the water is drawn. This information is likely to be of considerable value in studying the process of mixing, both from the physical point of view and also in connection with the maintenance of plankton populations, etc.

Application of the Method

1) Discharge

Bute Inlet

In Bute Inlet in July, 1953, simultaneous observations of the vertical temperature distribution were made at half-hour intervals for two positions 11 miles apart (stations 4 and 6). These observations were continued for a period of 48 hours. During this time meteorological observations were made and vertical salinity profiles observed. Calculations for the heat budget indicate that 0.112 cal./ $cm.^2$/min. were passing through the surface and remaining. Temperature and salinity observations revealed the thermocline and halocline respectively to be at a depth of 3 metres. The mean temperature increase from station 6 to station 4, in the top 3 metres was 1.8C°. Below this the increase was insignificant. Salinity observations indicate that the increase in salinity was very small between the two stations. If the entire increase was due to entrainment only, it would amount to less than 5% of the total volume in unit column of brackish water considered. Hence in this calculation, this effect will be neglected.

The time necessary to heat the water column $1.8C^{o}$ equals 4900 minutes. This indicates a mean velocity between the two stations of 7.6 cm./sec. Assuming that the velocity equals zero at 3 metres and varies linearly to a maximum at the surface yields a 15.2 cm./sec. net current at station 6. Direct current observations at station 4 indicate a net surface current of 28 cm./sec. (Section V). Since this is 11 miles down-inlet from station 6, and as well, in mid-channel where the velocity would be expected to be higher than the mean value for the entire cross-section, the results in the two cases appear compatible.

The mean value of 7.6 cm./sec. compares very favourably with the values 6.8 - 8.7 cm./sec. (average values for June and July respectively) calculated from the mean monthly discharge (Table XVIII) and a mean salinity of entrained water of 28‰.

Portland Canal

Portland Canal was surveyed between June 22nd and June 25th, 1951. Some of the observed profiles of salinity and temperature are shown in figures 10a and 10b.

Equation (47b) was applied at station 12 with u_o etc., at station 14. The value of u_o was calculated to be 3.0 cm./sec. with a fresh water velocity of 2.8 cm./sec. (S_e = 26‰).

At station 12, (using data at 12 and 9) the calculations indicated a velocity of 3.2 cm./sec. This checks favourably with the value at station 14. Assuming that this represents the mean velocity for the cross-section, one arrives at a value for fresh water discharge of 19,300 cu.ft./sec. The mean monthly discharge for June in Portland Canal is estimated (Table XVIII) to be 14,460 cu.ft./sec.

Other Inlets

Temperature and salinity profiles for Gardner Canal, July 2-3, 1951, are illustrated in figures 10c and 10d. The values used in the calculations of velocities and discharge are given in Table XX. The agreement between the discharge calculations computed by the two methods (i.e. heat budget and precipitation) is remarkably good. The figures for discharge based on precipitation are however, only monthly averages, and consequently only partial agreement could be expected between the two calculations.

Similarly calculations were carried out for a number of the other inlets and in all cases the results compared favourably with the discharges computed from precipitation and river discharge measurements. The pertinent values used in these computations are presented in Table XX.

2) Depth of entrainment

In order to demonstrate the method of applying the T-S diagram to reveal the depth from which the entrained water is drawn, observations in Knight and Bute Inlets will be considered.

Knight Inlet, June 3-4, 1951.

The salinity and temperature profiles for the inlet are given in figures 13a and 13b. In figure 30, T-S curves are plotted for stations 7, 8, and 9.

A calculation for the velocity of the fresh water entering Knight, based on observations at stations 9 and 11, gives a value of 2.4 cm./sec. Applying this result together with observations at stations 8 and 9, for various values of h and Q, to equation (47b) give relationships between S_e and T_e :

$$(h = 7,\ Q = 0.224) \quad T_e = 20.1 - 0.88\ S_e \tag{53a}$$

$$(h = 6,\ Q = 0.224) \quad T_e = 19.3 - 0.72\ S_e \tag{53b}$$

$$(h = 6,\ Q = 0.112) \quad T_e = 15.6 - 0.28\ S_e \tag{53c}$$

These are plotted on the T-S diagram (fig. 30). From them the following conclusions can be drawn: it is impossible to have a 7 metre upper layer, if the value of u_o and Q are assumed correct, because the required T-S characteristics of the entrained water does not exist below the 7 metre level. (The line of equation (53a) crosses the T-S curves above 7 metres). If the thickness of the upper layer is chosen as 6 metres (53b), and the other variables the same as before, then the water can be drawn between the depths of 6-8 metres only. The effect of halving the rate of heating (which is equivalent to doubling the fresh water transport), is shown by equation (53c) and this too is plotted in figure 30. This reveals a wider range of values from which the entrained water can be drawn. Values as deep as 10 metres would satisfy the requirement imposed by equation (47b).

Other Inlets

Plots similar to those illustrated in figure 30 were constructed from observations taken in Bute Inlet, May 17-24, 1951 (see fig. 7b). The equation derived from (47b) was:

$$(h = 3.4) \qquad T_e = 17.4 - 0.27\, S_e$$

This was found to intersect the T-S curves at a depth of about 3.8 metres. Assuming the calculated values of u_o and Q to be correct, the results indicate that the diluting water was drawn from a level immediately below the depth of no net motion.

Similar plots were construced for Bute, Knight, and other inlets, for observations taken at other times and other positions and in all cases the results indicated that the entrained water was drawn from a very thin layer immediately below the level of no net motion.

The results of this analysis also indicated that generally the depth of no net motion decreases to seaward. (This was required if the conditions on T_e and S_e in equation (47b) were to be met). The salinity distributions indicate that the salinity of the entrained water usually decreases to seaward. They suggest also that the major part of the up-inlet motion is confined to a fairly narrow range. The increase in salinity towards the head of an inlet results in a pressure gradient in that direction, which supports the suggestion of the proposed circulation. However in order for this salinity distribution in the intermediate layer to persist, requires that a quantity of salt must be entering the layer from deeper levels. This in turn requires a net up-inlet flow of salt water at the deeper layers, although the quantity necessary is small relative to that which would be supplied in the thin layer immediately below the depth of no net motion.

It appears that the predominant up-inlet motion near the mouth transports water whose salinity is somewhat less than that of the deeper water. It implies that in some region beyond the stations considered, mixing downward must become important. It seems conceivable that a critical velocity exists, as suggested by Cameron and Stommel, and that this occurs near the mouth. Up-inlet from this position the predominant vertical process across the level of no net motion is entrainment (i.e. a unindirectional flow) of salt water from below, with only a limited downward penetration of fresh water. However, to seaward of the position where the critical velocity exists, exchange processes (i.e. both upward and downward penetration) may be predominant.

Eddy Conductivity at Surface

Since the flux of heat through the sea surface is known, as well as the vertical temperature gradient of the water, the vertical eddy conductivity can be calculated from the relation: $Q = -A_t \frac{\partial T}{\partial z}$, where Q = flux of heat through the surface.

Calculations for Bute Inlet, July 1-2, 1953, indicate a value for A_t of 12 cm.^{2}/sec., whereas in Knight Inlet on July 17, 1953, A_t was only 0.8 cm.^{2}/sec. It is interesting that such a large difference is found for the two cases. It is quite possible that the wind has been the important factor in mixing the water in the surface layer in Bute and thus reduced the vertical temperature gradient.

Calculations of the eddy conductivity for other inlets gave values usually intermediate between those given for Bute and Knight but no systematic relationship to position in the inlet was evident. This is not surprising since the rate of heating and the penetration is strongly dependent on meteorological conditions.

VIII. RESULTS AND DISCUSSION

The analysis of the data presented up to this point, has been segregated into specific categories, and in certain instances results and pertinent points have been omitted because they depended in part upon the contents of a following section. This section will be devoted to bringing the various aspects of the study into one composite picture, and to a general discussion of the results.

Richardson Number

The theoretical results (Section IV) indicate that the Richardson number is related to the turbulence of the fluid. The conclusion was that turbulence would not decay if $Ri \leq A_v/A_s$. Since the medium is normally turbulent, determination of the Richardson number should give an idea of the value for the ratio of $A_v : A_s$.

Two velocity profiles which were observed simultaneously with the salinity at station 4 in Bute Inlet, August 11, 1952 are illustrated in figures 31a and 31b. One set of observations was taken during the flooding tide and the other during the ebbing tide. In figure 31a the velocity profile indicates very little vertical shear. The halocline, located at a depth of 5 metres, coincides with the depth of minimum velocity. The Richardson number lies between 1 and 5 for the upper 2.5 metres and then increases rapidly reaching an infinite value at 5 metres depth. The infinity arises because of zero velocity gradient at this point, with a finite stability. Below 5 metres it decreases abruptly, remaining at values less than 10 over a 5 metre layer, and finally increases to a value of several hundred at 20 metres depth.

During the ebb tide much wider variation of the velocity gradient is evident but as for the flood tide, a minimum in the velocity coincides with the halocline depth. The velocity gradient is very high in the top 4 metres while the density gradient is relatively small. Consequently the Richardson number is much less than unity. At 5 metres depth it is infinite and then decreases to less than 5 at 7.5 metres. It remains less than 5 down to 16 metres before finally increasing to a value in excess of 100 at 20 metres.

The observations presented in figure 31c were also made at station 4 in Bute Inlet on July 2, 1953. These were taken during the flood tide, but are markedly different from the observations shown in figure 31a, in that larger velocity gradients are present. Two minima and two maxima also occur in the velocity and give rise to a very irregular Richardson number. The high velocity shear and small density gradient in the top 3 metres give a Richardson number much less than unity, but below this the value varies widely.

The fairly extensive current measurements made at station 4 in Knight Inlet between July 11-15, 1953, (discussed in Section V) allow an estimate to be made of the variation in Richardson number over a tidal cycle. The lack of simultaneous salinity and current measurements prevents an accurate evaluation of the Richardson number, but since the variation in the salinity gradient is small relative to that of the square of the velocity gradient over a tidal cycle, the error introduced in using a mean value for the density gradient is probably small.

The velocity profiles presented in figures 32a and 32b represent typical mean values for the ebb and flood tide respectively. The maximum velocity gradient occurs during the flood tide and therefore the Richardson numbers are smaller in this layer during the flooding tide. For both flood and ebb tide maxima and minima occur in the velocity profile and consequently infinite values are obtained for the Richardson number at these points.

A summary of the variations of the Richardson number over two tidal cycles is given in Table XXII. Values range from less than 0.3 to over 3000. However the average value is probably less than 50 (if the infinite ones are excluded). In general the smaller numbers occur during the flooding tide. If it is assumed that the medium is turbulent at all times, except possibly at points where the velocity gradient vanishes, then the mixing is more intense during the flooding tide. Ignoring the very large values of the Richardson number, it would appear that during most periods A_v is from 2 to 50 times larger than A_s (i.e. the momentum transfer is much larger than the salt transfer). This is in general agreement with the values calculated by Taylor (19) using Jacobsen's data (22) for Schultz's Grund.

The relationship derived by Rossby and Montgomery (see Section IV) expressing the eddy diffusivity in terms of its value at neutral stability was given as:

$$A_s = \frac{A_o}{\sqrt{1 + \beta R_i}}$$

where A_o = eddy diffusivity at neutral stability.

The function A_s/A_o was plotted with depth for the results presented in figures 32a and 32b, for $\beta = 10$. These curves are shown in figure 32c.

The ratio A_s/A_o in the upper 20 metres is generally less than 0.10 for the ebb tide and generally greater than this value for the flood tide. The coefficients in both cases are minimal just below the level of maximum stability, and maximal in the top brackish layer. The curves are more regular than those for the Richardson number. Because of the manner in which this number is introduced in this expression, infinities are removed. If the Richardson number is to be incorporated into the differential equation determining the salinity, and the equation solved analytically, then the introduction in the form as expressed by Rossby and Montgomery appears desirable, as it eliminates any infinities.

Dynamics and Distribution of Mass

There can be no doubt that the dynamics and the distribution of property in the inlets are interrelated. Because, if mixing occurs so as to change the salinity distribution, then the horizontal distribution of mass is altered and in turn the horizontal pressure gradient. To see how the two factors are related, consider the following simple dynamic situation coupled with a simple equation for the distribution of mass.

Only the x and z components of the equation of motion will be considered. Assume that the only two important forces in the x direction are the pressure gradient and the vertical friction, and that the hydrostatic equation holds for the vertical direction. Therefore,

$$\frac{\partial p}{\partial x} = \frac{\partial}{\partial z}\left(A_v \frac{\partial u}{\partial z}\right) \tag{56}$$

and

$$\frac{\partial p}{\partial z} = \rho g \tag{57}$$

where p = pressure and ρ = density.

Consider also that the horizontal advection of salt is balanced by vertical diffusion (eqn. (7)). Since the density is approximately proportional to salinity only, S can be replaced by ρ. Therefore,

$$u \frac{\partial \rho}{\partial x} = \frac{\partial}{\partial z}\left(A_s \frac{\partial \rho}{\partial z}\right) \tag{58}$$

The flow must satisfy equations (56), (57), and (58).

Differentiating (56) with respect to z and (57) with respect to x, and combining, yields,

$$g\,\frac{\partial \rho}{\partial x} = \frac{\partial^2}{\partial z^2}\left(A_v\,\frac{\partial u}{\partial z}\right) \tag{59}$$

Substituting for u from equation (58) gives,

$$g\,\frac{\partial \rho}{\partial x} = \frac{\partial^2}{\partial z^2}\left[A_v\,\frac{\partial}{\partial z}\left\{\frac{\frac{\partial}{\partial z}\left(A_s\,\frac{\partial \rho}{\partial z}\right)}{\frac{\partial \rho}{\partial x}}\right\}\right] \tag{60}$$

Equation (60) contains only terms involving the density and the two eddy coefficients. But even for A_v and A_s constant, the equation appears insoluble by analytic methods. However with certain simplifying assumptions, it may be solved numerically using field observations to gain some knowledge of the magnitude of A_v and A_s.

To carry out the integration it is necessary to assume A_v independent of depth.

Integration from the depth d where the horizontal and vertical density gradients vanish to some level z, one obtains:

$$A_s A_v\Big|_{z_1} = \frac{g\int_d^{z_1}\left\{\left(\frac{\partial \rho}{\partial x}\right)_\zeta \int_d^{\zeta}\int_d^{\eta}\int_d^{\xi} \frac{\partial \rho}{\partial x}\,d\xi\,d\eta\,d\zeta\right\}dz}{\left.\frac{\partial \rho}{\partial z}\right|_{z_1}} \tag{61}$$

A calculation of this quantity was carried out using the observations taken in Bute Inlet in August, 1953 (see figs. 14c and d). Anchor stations were taken at stations 2, 4, and 7. The densities were averaged at each station and the integration performed between stations 2 and 4, and, 4 and 7. The product $A_s A_v$ is tabulated (Table XXIII) for the two positions in the inlet (between the stations). The results indicate that $A_v A_s$ varies in magnitude from 0 to 30 with negative values at some depths. The negative values imply that the mixing process may not be as simple as assumed and that possibly vertical advection cannot be neglected. The expression A_s/u was also determined from the salinity distribution using the methods outlined in Section IV. If reasonable values for the velocity are chosen, A_s for the brackish layer lies between 0.05 and 0.5 cm.^{2}/sec. Substituting this range of values in the expression $A_s A_v$ indicates that A_v/A_s ranges from 1 to 100, which is of the same order of magnitude as the Richardson number calculations indicated.

There are two features in particular with regard to the water structure which deserve particular attention. Firstly, the decrease in thickness of the surface layer to seaward appears to be a common feature. Secondly, the vertical advection process seems to be more important than vertical diffusion in introducing salt into the brackish layer. If the observed salt in the brackish layer entered this layer by a diffusion (i.e. exchange) process, then, if there were no change in thickness in the seaward moving layer, the velocities would remain constant at all positions along the inlet. However, both direct velocity measurements and calculations from heat budget considerations indicate that the increase in velocity down the inlet is only partly due to the decrease in thickness.

The reason for entrainment of salt water into the upper layer (advection), predominating over diffusion processes does not appear obvious. However there is considerable evidence to suggest that with sharp density discontinuities this is the case. For instance Keulegan (30) has carried out a number of experiments using flumes, to study the structure and nature of the interface when a light liquid flows over a pool of motionless heavier liquid. In these tests he found that at certain velocities waves appeared at the interface of the two layers. If the velocity was increased, the crests of the waves broke off and were ejected into the lighter liquid. The reason for there being no corresponding ejection downward does not appear evident, but would seem to be due to gravitational force being unidirectional and therefore asymetrical as far as instability of waves is concerned. Alternatively, it may depend in some way upon the fact that the heavier liquid was effectively bounded longitudinally whereas the lighter fluid was not.

Near the head of the inlets the boundary between the fresh and salt water is very abrupt, although the salinity is not discontinuous. The mean velocity of the fresh water to seaward is small in this part of the inlet, but the velocity gradient through the halocline could conceivably be very high. It is plausible that this could give rise to small unstable waves at the halocline and consequently a transport or entrainment of salt water into the upper layer. But once the salt water is ejected into the upper layer the usual diffusion processes would distribute the salt throughout the brackish layer. The entrainment type of process is possibly only important in the upper reaches of the inlet where the density gradient between the layers is high. But, on the other hand, the higher velocities in both layers, to seaward, may be sufficient to keep a similar type of process in existence over most of the length of the inlet. Certainly internal waves at the halocline are present at times in the lower reaches of some of the inlets, but whether they are ever unstable is not known.

Fresh Water Inflow at Side

Stommel and Farmer (27) constructed a mathematical model of a deep estuary in which they consider entrainment of salt water from the deep motionless layer into the upper layer, but allow no penetration downward. They consider each of the layers to be vertically homogeneous with an abrupt change of density across the interface. The density of the deep layer is assumed constant over the entire estuary, but the density in the upper one increases from head to mouth and as x approaches infinity, the density of the top layer approaches that of the bottom layer. The mixing is considered to be fixed, independent of the mean flow and the cause of the mixing is not specified beyond the statement that it may be due to wind, tidal currents, or the shear developed at the interface. One of the important results of the paper is an expression relating the thickness of the upper layer to the transport, and density distribution. The equation derived is:

$$\frac{dD}{dT} = \frac{1}{\sqrt[3]{bK}} \left[\frac{2b-1}{b-2} \right] \tag{62}$$

where $b = \frac{v_u^3}{k}$ and $k = \frac{gT}{2}\left(\rho_e - \frac{\rho_u}{\rho_e}\right)_0$

and where D = thickness of upper layer, T = transport, v_u = velocity of upper layer, ρ_u and ρ_e the density of the upper and lower layers respectively. The subscript "o" indicates the value of the quantities at the head of the estuary.

The equation expresses two important results: Firstly, for values of b between 0.5 and 2.0, the thickness of the top layer decreases with increasing transport. Since the value of b increases very rapidly after a certain value of the transport, Stommel concludes that the model breaks down as b approaches 2.0. Secondly, he notes that at b = 2.0, the velocity of the upper layer is equal to the velocity of an internal wave at the interface.

Stommel also derives a relationship between the change in thickness of the brackish layer to seaward and the fresh water discharge. The results indicate that for 'sub-critical' flow (i.e. less than the velocity of an internal wave at the interface) the layer decreases in thickness to seaward. Observations in the inlets indicate that the brackish layer usually does decrease to seaward, and therefore according to the Stommel criterion the flow is sub-critical.

Stommel and other investigators have all considered the fresh water inflow to be confined to the head of an estuary. However in a number of inlets, and at certain times of the year a large percentage of the fresh water enters from the sides. An attempt was made to see what effect this external influence might have in altering the circulation derived for an inlet with inflow only at the head. The modification to the simpler model, where Stommel considered the effect of entrainment and friction separately for a one layer system indicated that although no radical change in the form of the results occurred, the particular effect of entrainment or friction was enhanced. From this preliminary analysis it was presumed that the condition as expressed in equation (62) would be altered in such a way as to make the surface layer even thinner for sub-critical flow.

Comparison of Measured with Calculated Velocities

The transport values derived from heat budget considerations and those from fresh water discharge calculations in general agree closely with each other. But these, usually indicate smaller values for the net seaward movement in the brackish layer, than those measured directly. It appears, that the net currents obtained from direct measurements made in mid-channel, are considerably higher than the average for the complete cross-section. Evidently the mean velocity for the entire cross-section is of the order of 0.5 to 0.7 of the value at the centre of the cross-section. Visual observations of dye patches placed in the surface water supports the suggestion of a considerable lateral variation in velocity (Section V). These observations indicate that it is not uncommon for relatively high velocities to occur over part of the width, with the remainder practically motionless.

Although there is evidence for a wide variation in velocity laterally, there is no indication of a comparable variation in salinity or temperature. If the flow were laminar and the velocity decreased to zero at the sides one might expect that in summer months the water would be considerably warmer along the sides of the inlet than in the central portion. Since no significant lateral variation in temperature is evident, one concludes that lateral mixing is sufficient to keep the water nearly homogeneous across the full width of the inlet.

The heat budget method offers an advantage over direct current measurements in that it gives a mean value for the entire cross-section, whereas to achieve the same result by direct methods would require that current measurements be made at several positions across the inlet. It offers a practical advantage as well in that more extensive areas can be surveyed in a shorter period of time and with less expense.

Irregular Boundaries

No attempt has been made to evaluate the effect of irregular lateral boundaries on the circulation and distribution of properties longitudinally. However, there is some evidence to suggest that these irregularities may play an important role in certain areas. A conspicuous example of this was observed in Knight Inlet in July, 1953. The salinity increased very slowly down the inlet in the surface layer to station 6 where its value was approximately 6‰. Between stations 6 and 5 the salinity increased to more than 12‰. Between stations 5 and 2 the salinity again increased very slowly reaching a value of 16‰ at station 2. Beyond this the rate again increased. A glance at the chart indicates that the inlet configuration changes abruptly between the two stations, 6 and 5. The inlet below station 5 is relatively straight, lying approximately east-west, whereas above that it lies nearly north-south. Whether the abrupt transition in salinity is due directly to the change in inlet orientation or indirectly to the associated meteorological conditions is not known. The wind is usually up-inlet during the daytime and the effect of a long fetch would probably enhance the wind mixing process in this part of the inlet.

Current measurements in Bute Inlet indicated that cross-inlet flows of considerable magnitude were common. In view of the very irregular lateral boundaries it is thought that these give rise to lateral eddies in certain regions, which may appear for only a portion of the tidal cycle. The location of the anchor station was quite possibly in an area influenced by such eddies. If this was the case then it explains the irregular nature of the measured currents and the consequent difficulty in determining the net circulation.

IX. SUMMARY AND CONCLUSIONS

The emphasis in this thesis has been directed along two paths: first, an effort has been made to evaluate the diffusion coefficients for salt in British Columbia inlets, and in addition to obtain theoretical salinity profiles, which with suitable coefficients would agree closely with actual observations. Second, an examination of some of the factors involved in or controlling the circulation and distribution of properties has been undertaken. This included a study of the fresh water discharge into the inlets, and the circulation and entrainment based on heat budget requirements.

On the assumption that horizontal advection balanced vertical diffusion, eddy diffusion coefficients were calculated numerically by applying the data to the difference equation. For data which were reasonably smooth the expression A_z/u was found to be positive in the brackish layer and negative below it. The depth of no net motion, as indicated by the sign change of A_z/u, was generally found to decrease towards the mouth of the inlet. Application of reasonable velocity fields indicated that A_z usually increased in value from head to mouth in the brackish layer. No simple pattern was detected in the deeper up-inlet moving water.

In many cases however, large and inconsistent variations in the expression A_z/u were found when the unsmoothed data were applied. It was concluded that this inconsistency resulted because, in the first place, the observations were not always representative of the 'steady state', and secondly, the vertical advection was probably under some circumstances significant. Since the calculated coefficients incorporated the effect of this factor, negative values of the diffusion coefficient were possible, and could alternate in sign merely by variation in magnitude of the vertical velocity, even without a change in sign.

The fluctuations in the quantity A_z/u made it desirable to have some systematic method for smoothing the data. As a first attempt, empirical functions, usually exponential in form, were constructed which were reasonably simple in form and resembled the observations in general character. A number of expressions were found, each having certain desirable features. But these curves did not contain all the desired features, possibly because they were not based on any particular physical mixing process, and therefore on this latter account were considered unsatisfactory.

The differential equation expressing a balance between horizontal advection and vertical diffusion together with various boundary conditions was solved analytically for a diffusion coefficient which was a function of longitudinal position but constant vertically. Appropriate values for the velocity were applied to the solutions so that the salinity profiles observed at different positions in the inlet could be fitted to the theoretical profiles. . Calculations indicated that the eddy coefficient was nearly constant in the upper reaches of the inlets, with a value of the order of 0.1 cm.^{2}/sec., but increased very rapidly down the inlet reaching values of 2 - 10 cm.^{2}/sec. at or near the mouth. It is probable that this increase is associated with decreasing stability and increasing velocity shear toward the mouth, but the lack of direct velocity measurements along the length of the inlets did not permit more than a qualitative relationship to be expressed between eddy diffusion, and the circulation and distribution of mass.

The reasonably close agreement between the theoretical curves of salinity against depth and the observations suggested that the variation in eddy diffusion with depth was possibly of second order importance, although in the region below the halocline, where the departure of the theoretical curves from the observed ones was greatest it suggested that in this region the vertical variation probably was significant.

The theoretical curves which most closely approximated the observations required that the salinity be maintained constant at a depth about twice that of the halocline. The results also indicated a reversal in longitudinal salinity gradient between the upper and lower layer, in the upper reaches of the inlet. The fact that this condition is evident in many of the field observations supports the theoretical assumptions that have been made. Since this layer was observed to be moving up-inlet, the increase in salinity towards the head can only be maintained by diffusion of salt vertically from the deeper water. This suggests that the assumption that the salinity is maintained constant at some depth (and therefore a flow of salt across this level) in the theoretical considerations is probably realistic.

An evaluation of the Richardson number, indicated that the value of the eddy viscosity is generally from 10 to 50 times larger than the eddy diffusivity. The Richardson number was usually found to be smaller during the flood tide than during the ebb, indicating that it is probable that the bulk of the mixing occurs during the flooding tide. The variation in the Richardson number over a tidal cycle is controlled principally by the velocity shear.

A method has been developed for determining total fresh water discharge into the inlets. The method utilized direct precipitation observations on the drainage basins together with the available river flow measurements.

The calculations indicated that a significant proportion of fresh water inflow was entering from the sides of the inlets, and in certain instances the amount exceeded that entering at the head. Theoretical models of estuarine circulation to date have assumed that all the fresh water enters at the head. It is conceivable that inflow from the sides may have an important influence on the estuarine circulation.

No specific relationship was found between length of inlet and total discharge. Alternatively it could not be definitely proved or disproved that an inverse relationship existed between length or area of inlet and intensity of mixing. A more critical analysis of the salinity observations indicated that at the positions chosen for the mouths of the inlets the salinity varied considerably from inlet to inlet. The major difficulty in drawing an explicit conclusion from the data in regard to intensity of mixing, is the choice of the mouth of the inlet. In this study, the mouth was chosen as the point where the inlet widened quite abruptly, and in systems such as Dean and Burke channels, a moderate degree of personnel selection was involved. A more thorough analysis of the possible relationship between mixing intensity and the length or area of inlets above positions of lateral constrictions is recommended. These studies might reveal a proper criterion for determining the location of the inlet mouth.

A new method, based on the heat budget has been developed to determine the mean seaward movement of the brackish layer. This method was also applied to gain information on the mixing process, and to determine the depth from which salt water was entrained into the surface layer.

The application of the heat budget method for determination of net currents in the surface layer gave results that were in good agreement with those derived from fresh water discharge analysis. The method offers some very real advantages over direct methods for determining fresh water inflow and circulation. One advantage is that the accumulation of fresh water in the inlets effectively act as a smoothing device in determining the mean fresh water discharge over a period of a few days, the period depending on the inlet size. The method permits the determination of net currents without the necessity for anchoring a vessel. Although a number of temperature, salinity, and meteorological observations are necessary, the effort required to give a reliable value for the net current is considerably smaller than is required by direct current measurements.

The usefulness of the method to determine the average depth from which the entrained water is drawn has also been demonstrated. It revealed that the entrained water is drawn from a layer only a few metres in thickness and directly below the level of no net motion. The salinity of the water being entrained was found to decrease towards the mouth, and also the depth of no net motion decreased in this direction as well.

Although considerable success was obtained by theoretically considering the vertical transfer to be described by a diffusion equation it must still be emphasized that the actual vertical transfer of salt across the level of no net motion may be predominately an advective one. Only a small penetration of fresh water downward is evident in most parts of the inlet, although near the mouth and beyond, the exchange process may predominate. Why water is entrained upward more readily than downward is not at once evident. It appears that the entraining process is dominant in transferring salt water across the level of no net motion, but that above this, the usual diffusion process predominates.

Since the relative importance of the entraining process is probably greatest where the most abrupt halocline is found, it is recommended that observations be made on the behavior of the halocline and the velocity shear near the head of an inlet. A critical analysis of the interfacial region would give a sounder basis for making hypothesis about entrainment processes.

It is recommended also that further direct current measurements be made, so that a criterion for mixing based on velocity shear, may be investigated more thoroughly. Observations of salinity should be made simultaneously with velocity. Even though the variation in the mass field may be small relative to the velocity field, the small variations may be extremely significant.

There can be no doubt that the mass and velocity fields are interrelated. Alteration in the mass field produces important changes in the dynamics, which in turn alter the circulation. The mixing process and the dynamic may be thought of as controlling each other, and consequently for a complete understanding of the fundamental nature of the mixing process, the dynamics must be considered also. This does not mean that to obtain a relationship between the kinematics and the distribution of properties requires a knowledge of the dynamics. On the contrary, the results of this study have indicated that some success in relating these factors is possible without directly considering the forces involved.

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REFERENCES

(1) Pritchard, D. W. (1952). Estuarine hydrography Advances in Geophysics, vol. 1, pp. 243-280. Academic Press Inc, New York, N. Y.

(2) Pickard, G. L. (1953,1954). Oceanography of British Columbia mainland inlets I, II, III, IV. Fish. Res. Board of Can. Progress Reports of the Pacific Coast Stations Nos. 96, 97, 98, 99.

(3) Tully, J. P. (1949). Oceanography and prediction of pulp mill pollution in Alberni Inlet. Fish. Res. Board of Can. Bull. 83, 169 pp.

(4) Le Brasseur, R. J. (1954). The physical oceanographic factors governing the distribution of plankton in the British Columbia inlets. M.A. thesis, Dept. of Zoology, Univ. of B. C.

(5a,b) Tabata, S. (1954). a) Physical oceanography of Bute Inlet. M.A. Thesis, Dept. of Physics, Univ. of B.C. b) Oceanography of British Columbia mainland inlets, V. Fish. Res. Board of Can. Progress Reports of the Pacific Coast Station No. 100.

(6) Sverdrup, H.U., M.W. Johnson, and R. H. Fleming. The Oceans. Prentice-Hall, Inc. New York, 1942.

(7) Dawson, W. Bell (1920). The tides and tidal streams, Dept. of Naval Service, Ottawa, Can. King's Printer, Ottawa.

(8) Pacific Coast tide and current tables (1951-54). Canadian Hydrographic Service.

(9) Kimball, H.H. (1928). Daily totals of solar radiation, Monthly Weather Review. Vol. 56, No. 10, W.B. No. 969, Oct.

(10) Ketchum, B.H. (1951). The exchanges of fresh and salt waters in tidal estuaries. Journal of Marine Research, Vol. 10, No. 1, pp. 18-38.

(11) Arons, A.B. and H. Stommel (1951). A mixing length theory of tidal flushing. Trans. American Geophysical Union, Vol. 32, No. 3, pp. 419-421.

(12) Campbell, N. J. (1954). A study of lateral circulation in an inlet. Doctoral dissertation, Univ. of B.C.

(13) Durand, W.F. (1943) Aerodynamic theory. Vol. III, Divisions F-I, Cal. Inst. of Technology, California.

(14) Proudman, J. (1953). Dynamical Oceanography. Nethuen and Co., Ltd., London.

(15) Carslaw, H.S. and J.C. Jaeger (1950). Conduction of Heat in solids. Clarendon Press, Oxford.

(16) Churchill, R.V. (1944). Modern Operational Mathematics in engineering. McGraw-Hill, Inc. New York, N.Y.

(17) Cameron, W.M. (1951). On the dynamics of inlet circulations. Doctoral dissertation, Univ. of Cal., Los Angeles, California.

(18) Fjeldstad, J.E. (1936). Results of tidal observations. Norwegian North Polar Exped. with the Maud 1918-1925. Sci. Results, Vol. 4, No. 4, 38 pp.

(19) Taylor, G. I. (1931). Internal waves and turbulence in a fluid of variable density. Conseil Perm. Intern. p. l'Explor. de la Mer. Rapp. et Proc.-Verb. Vol. 76, pp. 35-42.

(20) Richardson, L.F. (1920). The supply of energy from and to atmospheric eddies. P.R.S. London, Ser. A, Vol. 97, pp. 354-373.

(21) Calder, K.L. (1949). The criterion of turbulence in a fluid of variable density, with particular reference to conditions in the atmosphere. Royal Meteor. Soc. Jour. Vol. 75, pp. 71-88.

(22) Jacobsen, J.P. (1913). Beitrag zur Hydrographie der danischen Gewasser. Komm. f. Havunders. Medd., Ser. Hydr. Vol. 2, No. 2, 94 pp. Copenhagen.

(23) Rossby, C.G. and R.B. Montgomery (1935). The layer of frictional influence in wind and ocean currents. Papers in Physical Oceanography and Meteorology, Vol. III, No. 3.

(26) Sverdrup, H.U. (1936). The eddy conductivity of air over a smooth snow field. Geof. Publ. Vol. XI, No. 7.

(27) Munk, W.H. and E.R. Anderson (1948). Notes on a theory of the thermocline. Journal of Marine Research. Vol. 7, No. 3, pp. 276-295.

(26) Deacon, E. L. (1949). Vertical diffusion in the lowest layers of the atmosphere. Quart. Jour. Roy. Meteor. Soc. Vol. 75.

(27) Stommel, H. and H.G. Farmer (1952). On the nature of estuarine circulation Part I. Ref. no. 52-88, Woods Hole Oceanographic Institution, Woods Hole, Mass. U. S. A.

(28) Anonymous. Climatic Summaries for selected meteorological stations in the Dominion of Canada. Vol. I, Meteorological Division, Dept. of Transport, Toronto, Canada.

(29) Anonymous. (1910-1953). Climate of British Columbia. Tables of temp., prec. and sunshine. Dept. of Agriculture, Prov. of B. C.

(30) Keulegan, G.H. (1949). Interfacial instability and mixing in stratified flows. Jour. of Res. National Bureau of Standards, U.S. Dept. of Commerce, Vol. 43, pp. 487-500.

(31) Anonymous. Surface water supply of Canada. Pacific Drainage, Dept. of Resources and Development Water Resources Division. Water Resources paper nos. 65-114.

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TABLE I

A_z/u (cm.x10^2)* Bute Inlet, May 17-25, 1951

Stn. Depth(M)	2	4	7	8
1	0.3	0.3	21.9	
2	0.5	-0.6	21.6	
3	0.1	1.4	2.5	
4	-0.5	8.5	0.3	
5	-0.5	3.3	-1.7	
6	-0.2	-8.5	0.9	
7	-0.4	-0.3	0.5	
10	∞	-0.4	-2.2	

TABLE II

A_z/u (cm.x10^2) Bute Inlet, May 29, 1952

Stn. Depth(M)	3	4	5	6	7	8
2	5.2	2.6	4.3	10.7	7.9	
4	-31.5	3.3	1.4	6.9	2.9	
6	-2.9	-2.0	-1.7	32.8	4.3	
9	-9.8	-6.5	-5.8	-1.0	-26.9	
18	-150	-38.7	-61.8	-10.7	-2.1	

TABLE III

A_z/u (cm.x10^2) Jervis Inlet, May 27, 1952

Stn. Depth (M)	2	4	6	10
1	-4.2	-5.0	-1.4	
2	-1.1	-0.7	-0.3	
3	-0.6	-1.8	-0.4	
4	-0.3	-0.9	-0.4	
5	-1.7	-0.8	-0.3	
8	0.9	-2.4	-5.3	

*Example: 2.5 in table equals 0.025 cm.

TABLE IV

A_z/u (cm.x10^2) Gardner Canal, July 1-3, 1951

Depth (M) \ Stn.	1a	2	3a	4	5	6	7a	8	9a	10	11	12	13	14
3	4.4	3.6	0.0	11.8	2.6	0.0	0.5	0.6	1.5	5.9	1.4	2.8	9.0	
6	-2.9	0.6	0.0	-1.9	-2.4	-0.7	0.0	-0.7	-5.4	2.6	0.0	0.7	0.9	
8	-2.4	13.5	-21.6	-1.8	-2.1	0.0	2.8	-2.5	-2.3	-2.0	0.0	-0.7	-1.3	
11	1.2	15.6	-59.4	-14.0	11.1	-4.6	2.0	-1.9	1.1	-3.4	0.0	-0.9	-5.2	
14	2.7	10.5	-20.3	-2.0	37.8	-2.6	-2.1	-5.6	9.0	-1.4	-2.0	-1.3	-1.5	
18	4.8	42.4	-98.3	-74.5	125	-5.2	0.0	-8.2	12.0	-10.2	-4.5	0.0	-6.0	
28	8.5	123	-16.3	-62.2	101	-15.0	21.6	-2.2	-5.2	-26.1	-13.5	-9.0	-44.0	

TABLE V

A_z/u (cm.x10^2) Portland Canal, July 22-23, 1951

Depth (M) \ Stn.	6	7	8	9	10	11	12	13	14
4	0.7	0.9	3.1	2.6	2.0	6.2	-5.1	7.0	
6	1.5	1.4	2.3	1.0	-4.2	9.5	-10.9	14.0	
7	-0.7	-3.7	-2.7	2.9	-7.4	4.6	-11.6	-80.0	
9	-2.4	7.6	-3.2	-1.1	10.3	-1.7	3.2	-18.8	
14	-6.3	-2.5	-1.5	-6.6	-7.8	3.4	3.4	11.3	
18	-12.4	-2.7	0.4	-11.2	-2.6	-1.8	11.1	-34.2	
28	-49.6	-4.0	-6.6	-31.9	-1.0	-12.0	43.5	-116	

TABLE VI

A_z/u (cm.x10^2) Dean Channel, June 17-19, 1951

Stn. Depth(M)	1		2		3		4		5
2		8.9		2.8		3.7		27.1	
4		-13.0		17.4		2.2		3.3	
6		-10.0		54.7		0.9		3.4	
9		-6.2		-5.7		16.6		-10.7	
14		95.5		-92.2		97.4		-25.6	
18		13.0		-25.7		20.3		-24.2	
28		42.9		-100		99.0		-51.5	

TABLE VII

A_z/u (cm.x10^2) Dean Channel, June 19-20, 1951

Stn. Depth(M)	5		6		7		8		9		10
3		-0.8		0.7		2.1		2.8		2.0	
6		2.8		-6.8						0.0	
8		4.4		-20.4						-0.1	
11		-13.6		30.8						-14.8	
14		-3.9		6.5		-6.3		8.1		-2.9	
18		-16.4		24.5		73.9		29.4		-24.1	
28		-65.2		93.0		-81.4		104		-21.2	

TABLE VIII

A_z/u (cm.x10^2) Knight Inlet, June 1-2, 1951

Stn. Depth(M)	2		3		4		5		6		7		8		9		10		11
4		9.0		31.0		20.0		24.0		6.0		6.0		4.0		-0.6		2.0	
9		56.0		-25.0		2.0		-12.0		4.0		-6.0		-3.0		0.6		-3.0	
14		165		322		171		208		-6.0		-5.0		-1.0		7.0		-19.0	
18		72.0		-523		-103		-35.0		-13.0		-13.0		2.0		16.0		-27.0	
28		159		-263		36.0		-81.0		-36.0		-36.0		1.0		-54.0		-68.0	

TABLE IX

A_z/u (cm.x10^2) Knight Inlet, June 3-4, 1951

Stn. Depth(M)	1	2	3	4	5	6	7	8	9	10	11
3		-10.8	-9.1	22.5	37.5	10.8	10.9	80.0	-29.2	-120	6.5
6		51.8	30.0	-21.8	6.8	-26.0	57.7	-2.2	2.1	1.0	0.9
8		-21.0	243	25.9	16.3	-16.1	21.8	20.7	-3.2	-0.3	-0.8
11		153	0.6	-46.8	-3350	-9.4	-3.5	-0.7	0.5	2.5	-1.0
14		128	25.8	177	-112	-17.2	-7.8	1.8	2.0	5.0	2.5
18		827	41.5	-85.8	71.4	-36.1	-18.8	5.0	4.5	8.0	7.0
28		-135	-33.0	-145	127	-102	-57.2	16.0	11.0	27.0	23.0

TABLE X

A_z/u (cm.x10^2) Bute Inlet, October 25-26, 1951

Stn. Depth(M)	2	3	4	5	6	8
4		-29.1	0.9	-4.9	2.2	0.0
6		297	6.5	27.9	-0.7	2.7
7		1.7	-0.7	5.8	-0.4	-0.7
9		112	-64.0	4.7	-0.9	-11.8
14		-319	-17.8	-31.9	-6.2	-7.8

TABLE XI

A_z/u (cm.x10^2) Bute Inlet, March 27-28, 1953

Stn. Depth(M)	1	2	3	4	5	6	7	8
5		39.0	4.0	-19.0	-4.0	-1.4	-0.61	-5.0
9		-25.0	-0.7	0.2	0.2	0.0	-3.0	-155
14		-17.0	69.0	0.0	20.0	10.0	-13.0	16.0
18		-36.0	121	0.0	0.0	0.0	-79.0	124
28		-91.0	273	75.0	-83.0	0.0	-332	516

TABLE XII

A_z/u (cm.x10^2) Bute Inlet, February, 1951

Stn. Depth(M)	2	4	7
1		∞	-7.0
2		∞	12.3
3		2.1	1.1
4		2.6	0.9
5		-1.7	-0.4

TABLE XIII

A_z/u (cm.x10^2) Bute Inlet, July 3-4, 1953

Stn. Depth(M)	4	6	8
2	0.6		2.9
4	-1.9		1.7
6	0.1		-4.0
9	-1.0		-1.8
18	-4.3		-6.1
35	-12.5		-25.4

TABLE XIV

A_z/u (cm.x10^2) Bute Inlet, June 8, 1954

Stn. Depth (M)			
9.6	19.1	9.2	∞
1.2	9.8	∞	∞
1.8	10.1	5.3	4.8
2.4	1.2	1.1	3.8
3.0	0.6	0.3	0.3
3.6	-0.3	6.0	0.7
4.3	-0.3	-0.1	0.7
4.9	-0.4	-0.2	-0.2
5.5	-0.3	-0.4	-0.2
7.2	-1.1	-1.0	-1.2
8.5	-0.4	-0.4	-0.7
10.0	-1.3	-0.7	-0.5

TABLE XV

A_s (cm.^{2}/sec.) and Surface Velocity (cm./sec.)

Bute Inlet, May 29, 1952

Stn.	Vel. (cm./sec.)	Time (sec.x10^{-5})	A_o	A_s (cm.^{2}/sec.)
Head		0.0	0.0	0.2
8	3.4	0.6	0.7	0.2
7	4.3	2.3	2.0	0.2
6	6.7	4.1	10	0.6
5	8.9	5.4	22	1.1
4	14.5	6.3	35	2.3
3	25.0	6.9	55	4.7

TABLE XVI

A_s (cm.^{2}/sec.) Knight Inlet, June 1951, and July, 1953

Station	A_s (cm.^{2}/sec.) June, 1951	A_s (cm.^{2}/sec.) July, 1953
11	0.01	0.02
10	0.16	0.03
8	0.30	0.15
6	0.75	0.94
4	4.5	5.0
3	25.0	8.0

TABLE XVII

Drainage basin areas and mean discharge

Inlet	Length Statute Miles	Area (Sq.miles)	Drainage Area (Sq.Mi.Statute Above Head	S-E Side	N-W Side	Total	Mean Discharge (Cu.ft./sec) Above Head	S-E Side	N-W Side	Side Total	Total
Howe	25	111	1,330	140	225	1,690	13,310	1,140	1,810	2,950	16,260
Jervis (above Sechelt)	40	73	189	440	247	876	1,740	3,200	1,460	4,650	6,390
Sechelt	28	48				550					3,670
Jervis + Sechelt	79	164				1,430					10,060
Toba	23	38	847	100	191	1,140	7,810	600	1,120	1,720	9,530
Bute	47	112	2,820	332	275	3,430	12,910	2,150	1,510	3,660	16,570
Loughborough	30	26	249	100	77	426	2,390	586	454	1,040	3,430
Knight to Stn. 2.	58	115	2,360	243	422	3,020	10,700	1,350	2,530	3,870	14,570
Kingcome	21	46				676					5,480
Seymour	40	58				428					3,620
Belize	30	38				342					2,900
Seymour + Belize System		107				827					7,000
Smith	21	21				428					4.100
Draney	14	10				145					1,390
Rivers	28		1,600	154	579	2,330	12,100	1,610	5,370	6,980	19,080
River (Total System)	40	63									
N. Bentinck	11	19	2,130			2,390	7,540			2,130	9,670
S. Bentinck	28	35	384			083	2,120			1,650	3,770
Burke Total	82	157				3,210					14,740
Dean (above Stn. 6)	36	54.3	3,540			4,260	8,060			9,920	17,960
Dean (total)	63	145				4,850					25,560
Gardner	56	77				2,510					20,300
Gardner + Kemano	56	77				8,470					29,050
Douglas	50					2,610					19,200
Work	30	42				303					2,900
Alice	9	9				338					1,870
Hastings	11	16				285					1,680
Observatory	44	86.5				1,020					5,810
Portland Canal (above 5)	65	113	392			1,430	2,450			6,850	9,300
Portland Inlet	25	100	7,510			8,080	31,070				34,870
Portland Inlet + Canal		461				11,800					55,700

Table XVIII

Monthly and yearly mean discharges of fresh water (in thousands of cu.ft./sec.)

Inlet	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Mean
Howe	14.40	8.79	12.90	6.78	8.67	6.05	11.90	23.90	29.20	30.20	24.90	17.70	16.30
Jervis (above Sechelt)	7.62	7.20	10.40	7.83	7.76	5.03	5.93	8.11	6.81	3.03	1.31	2.77	6.15
Sechelt	4.94	3.84	6.27	3.71	4.29	2.85	4.17	6.05	5.12	2.24	8.75	2.52	2.91
Jervis + Sechelt	12.60	11.04	16.70	11.50	12.10	7.94	10.10	14.20	11.90	5.27	2.19	5.29	10.10
Toba	11.00	5.70	5.47	3.17	4.32	3.15	6.07	10.50	16.50	18.10	15.20	15.10	9.52
Bute	18.30	14.60	12.50	10.30	10.40	10.50	15.50	24.00	31.50	24.60	15.30	11.00	16.50
Loughborough	8.80	3.03	2.60	2.15	2.16	2.19	3.21	4.99	6.55	5.12	3.18	2.28	3.43
Knight (2)	16.20	12.90	11.00	9.09	9.14	9.27	13.60	21.20	27.80	21.70	13.50	9.67	14.60
Seymour	4.87	4.70	5.32	3.19	3.38	3.42	3.59	4.65	4.11	2.53	1.51	2.23	3.62
Belize	3.90	3.76	4.25	2.54	2.70	2.73	2.87	3.72	3.29	2.02	1.21	1.79	2.90
Sey. + Bel. System	9.43	9.09	10.30	6.15	6.53	6.60	6.93	9.00	7.94	4.88	2.92	4.31	7.00
Smith	5.53	5.32	6.02	3.61	3.82	3.86	4.06	5.27	4.66	2.86	1.71	2.53	4.10
Draney	1.87	1.30	2.04	1.22	1.29	1.31	1.37	1.78	1.58	0.97	0.58	0.86	1.39
Rivers (Tot.)	24.6	19.4	13.10	11.00	10.20	9.55	12.40	19.80	29.30	29.40	27.30	22.10	19.10
N. Bentinck	10.40	7.16	3.30	1.88	3.09	3.57	5.67	9.69	21.00	19.10	16.60	14.10	9.67
S. Bentinck	4.01	2.78	1.28	0.73	1.20	1.39	2.20	3.77	3.14	7.40	6.43	5.47	3.77
Burke Tot.	16.10	11.60	6.49	3.74	5.51	6.19	8.00	15.10	30.60	27.40	23.60	20.40	14.70
Dean (above 6)	15.30	9.70	7.35	5.77	4.51	3.48	5.23	11.10	17.20	14.80	13.30	11.50	9.92
Dean Total	32.30	22.30	18.40	13.30	10.80	9.12	15.80	28.40	50.50	39.60	31.00	25.20	25.60
Gardner	31.20	19.80	15.00	11.80	9.21	7.10	10.90	22.50	35.10	30.20	27.20	22.60	20.30
Work	5.29	4.51	2.90	1.46	1.79	1.66	2.66	2.95	4.18	3.51	1.68	2.15	2.90
Alice Arm	1.97	1.29	0.64	0.48	0.42	0.29	0.99	2.91	4.71	3.87	2.90	1.98	1.87
Hastings Arm	1.76	1.15	0.58	0.43	0.38	0.26	0.88	2.61	4.21	3.46	2.70	;.77	1.68
Observatory	7.58	6.62	4.16	2.39	2.36	1.97	3.82	7.67	11.70	9.59	6.62	5.31	5.81
Portland Can. (to 5)	14.20	14.40	9.75	5.15	5.29	4.60	7.20	10.30	14.50	11.90	6.89	7.33	9.30
Portland Inlet (Mass)	36.70	24.00	12.00	8.85	7.78	5.35	18.40	54.30	87.60	72.20	54.20	36.90	34.90
Portland Can. + Inlet	68.20	55.50	33.40	20.20	19.50	15.50	34.40	77.70	121.00	99.50	70.30	53.50	55.70

TABLE XIX

Monthly and yearly mean discharges per unit area of inlet surface (cu.ft./sec./sq. mile)

Inlet	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Mean
Howe	130	79	116	61	78	54	108	215	263	273	224	160	147
Jervis (above Sechelt	104	99	143	107	106	70	81	111	93	42	18	38	84
Sechelt	103	80	131	77	90	60	87	126	107	47	18	53	82
Jervis + Sechelt	77	67	102	70	73	49	62	86	73	32	13	32	61
Toba	287	148	142	82	112	82	158	263	427	471	395	393	247
Bute	163	130	112	92	92	94	138	214	281	219	136	97	147
Loughborough	148	118	102	84	84	85	125	195	256	200	124	89	134
Knight	141	112	96	79	80	81	119	184	242	189	118	84	127
Seymour	85	82	93	56	59	59	62	81	71	44	26	39	62
Belize	101	98	111	66	70	71	75	71	86	53	31	47	76
Belize + Seymour	88	85	96	58	61	62	65	84	74	46	27	40	65
Smith	270	259	294	176	181	188	198	257	227	140	84	123	200
Draney	187	180	204	122	129	131	137	178	158	97	58	85	139
Rivers	391	308	207	175	162	152	197	315	465	467	434	351	303
N. Bentinck	540	373	171	98	161	186	295	505	1090	955	965	735	504
S. Bentinck	116	30	37	21	35	40	64	109	235	214	186	158	109
Burke + Bentincks	103	74	41	24	35	39	51	96	195	175	150	130	94
Dean (above 6)	379	273	236	173	135	119	215	367	687	521	384	302	331
Dean (total)	222	154	127	95	74	64	109	195	348	273	224	173	176
Gardner	406	257	195	153	120	92	141	293	457	392	352	294	264
Work	126	117	69	35	43	39	63	70	100	84	40	51	69
Alice	218	142	72	53	47	32	110	324	523	430	323	220	208
Hastings	110	72	36	27	24	16	55	163	263	216	169	111	105
Observatory	88	77	48	28	27	23	44	39	135	111	765	61	67
Portland Canal	126	127	86	46	47	41	64	91	128	106	61	65	82
Portland Inlet	682	555	333	202	195	155	343	777	1210	995	703	535	557

TABLE XX

Monthly and yearly mean discharges per unit width of inlet (in tens of cu.ft./sec./mile width)

	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Mean
Howe	325	198	290	152	195	136	270	538	657	632	560	400	367
Jervis (above Sechelt	416	394	572	428	425	278	325	444	373	166	72	151	338
Sechelt	288	225	367	216	201	167	244	353	300	131	52	147	228
Jervis + Sechelt	613	539	816	564	587	388	493	690	582	257	106	259	492
Toba	660	340	326	190	258	138	363	603	980	1080	908	903	567
Bute	765	610	527	430	434	439	643	1010	1320	1030	639	456	690
Loughborough	444	354	306	251	253	256	375	585	768	600	372	267	402
Knight (above 2)	817	650	555	458	461	467	690	1070	1400	1100	685	437	737
Knight (total)	1040	830	709	585	587	595	380	1360	1790	1400	622	874	940
Seymour	339	327	370	222	235	238	249	324	286	176	105	155	251
Belize	305	294	332	193	211	213	224	291	257	158	94	139	226
Belize + Seymour	634	612	692	414	440	445	467	606	534	328	196	290	471
Smith	535	532	602	361	382	386	406	527	466	286	171	253	410
Draney	262	252	286	171	161	183	192	249	221	135	81	120	195
Rivers	1090	860	580	490	454	426	552	881	1310	1310	1220	982	848
Rivers (plus Arms)	1570	1230	700	648	608	788	1260	1870	1860	1870	1740	1410	1210
N. Bentinck	594	411	138	108	177	205	325	555	1200	1050	952	809	555
S. Bentinck	325	224	104	59	97	112	173	305	659	600	521	442	306
Burke + Bentincks	918	658	368	212	313	351	454	856	1740	1560	1340	1160	834
Dean (above 6)	1370	983	350	623	486	428	774	1320	2470	1880	1380	1090	1190
Dean (total)	1510	1050	864	646	503	434	739	1330	2370	1860	1530	1180	1200
Gardner	2270	1440	1090	858	670	517	793	1640	2560	2200	1980	1640	1470
Work	378	322	207	104	128	118	190	211	298	251	120	152	207
Alice	221	145	73	53	43	32	111	327	530	435	326	222	211
Hastings	165	108	72	40	36	24	83	245	395	324	253	166	157
Observatory	385	336	212	121	120	100	194	390	593	488	337	270	295
Portland Canal	818	825	560	296	304	264	414	593	832	686	396	422	535
Portland Inlet	1710	1390	832	505	488	388	832	1940	3020	2480	1760	1340	1390

TABLE XXI

Fresh water discharge calculated from heat budget analysis (cu.ft./sec.)

Inlet	Date	Data from Stn. Nos.	Q_2 (Cal./cm^2/sec.)	h_o (M)	T_o (^oC)	S_o ($^o/_{oo}$)	u_o (cm/sec)	Discharge (from heat budget) cu.ft./sec.	Discharge (from prec.) cu.ft./sec.
Portland	22-25/vii/51	14-12	0.186	8.0	7.0	26.0	3.0	19,300	14,460
Portland	22-25/vii/51	12-9	0.186	7.0	8.0	25.0	3.2		
Gardner Canal	2-3 /vii/51	14-13	0.200	7.0	6.7	29.5	12.1	27,400	20,000
Gardner Canal	2-3 /vii/51	13-12	0.200	6.5	6.7	29.5	10.3		
Gardner Canal	2-3 /vii/51	12-11	0.200	6.25	6.7	29.5	6.45	20,300	
Gardner Canal	2-3 /vii/51	11-9	0.200	6.0	6.7	29.5	7.0		
Gardner Canal	2-3 /vii/51	9-7	0.200	5.25	6.7	29.5	11.2	24,600	30,200
Gardner Canal	2-3 /vii/51	7-6	0.200	5.0	6.7	29.5	13.1		
N. Bentinck	22/vi/51	11-10	0.149	5.5	9.6	27.5	3.57	12,500	12,100
S. Bentinck	23/vi/51	15-13	0.276	9.0	9.0	28.0	4.4	17,700	8,140
Rivers	16/vi/51	5-4	0.340	3.5	8.1	27.5	4.8	15,000	12,700
Toba	8/vi/54	3-2	0.166	2.0	12.6	26.0	10.0	9,600	16,450
Bute	17-24/v/51	7-4	0.167	4.5	8.6	27.5	4.6	19,800	18,800
Bute	17-24/v/51	4-2	0.167	3.5	11.2	23.5	5.3		
Bute	10-12/viii/52	7-6	0.221	5.3	9.2	25.0	3.1	16,100	15,300
Bute	10-12/viii/52	6-4	0.221	5.3			3.2		
Bute	1-2 /vii/53	6-4	0.112	3.0			7.6	22,000	24,600
Bute	20-23/viii/53	7-4	0.135	6.7	9.5	27.5	5.4		
Bute	20-23/viii/53	4-2	0.135	4.9			9.8	16,600	15,300
Knight	3-4/vi/51	11-10	0.224	7.0	9.0	25.0	1.7	14.500	22,300
Knight	3-4/vi/51	10-9	0.224	7.0	9.0	25.0	5.6	37,500	22,300
Knight	3-4/vi/51	11-9	0.224	7.0	9.0	25.0	2.7	21,700	22,300
Knight	3-4/vi/51	9-8	0.224	7.0			3.0		
Knight	3-4/vi/51	8-7	0.224	7.0			4.4		
Knight	17/vii/53	11-10	0.157	5.5			3.0	19,440	16,200

TABLE XXII

Richardson number variations over a tidal cycle,

Knight Inlet, July 1953

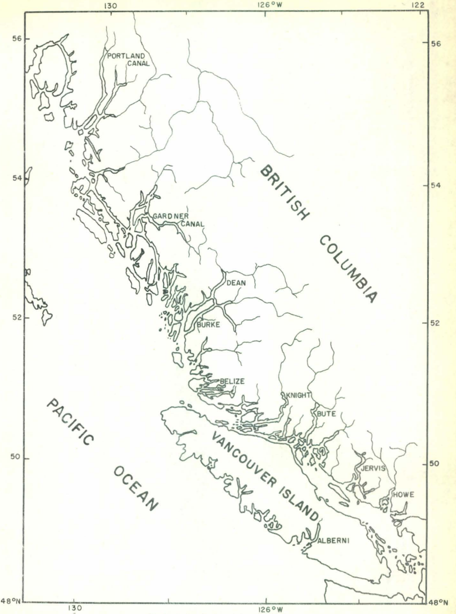
Tide / Depth(M)								
1	4.0	2.1	1.0	0.5	0.8	4.5	2.1	417
3	10.3	72.5	0.5	0.3	5.0	784	0.41	0.43
5	334	730	32.1	8.3	5.5	28.1	19.7	2.4
7.5	6.3	181	37.5	52.0	52.0	20.6	177	17.9
10	280	3040	57.8	3.2	3040	333	189	12.1
12.5	172	16.1	1.2	4.3	11.8	3.9	8.3	714
17.5	30.4	15.1	4.6	175	1400	62.3	1400	1400

TABLE XXIII

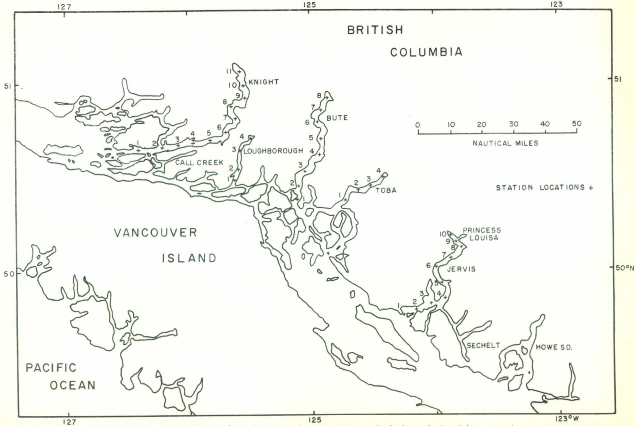
Eddy coefficients of viscosity and diffusion

Bute Inlet, August 1953

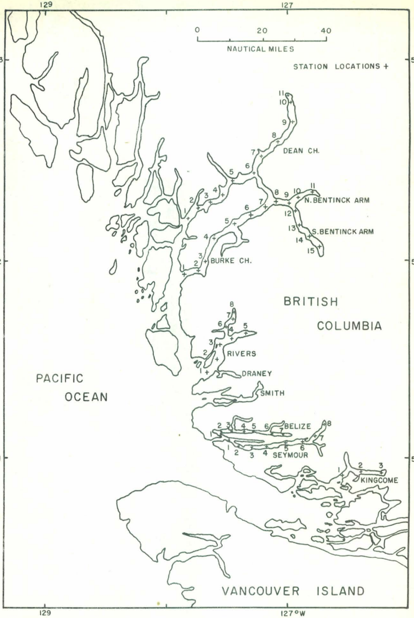
	Between Stations 7 and 4		Between Stations 4 and 2	
	A_sA_v (cm.^{4}/sec.^{2})	A_s/u (cm.x10^2)	A_sA_v (cm.^{4}/sec.^{2})	A_s/u (cm.x10^2)
1	22.4	2.0	30.5	3.7
2	8.6	1.45	7.2	1.2
3	4.2	0.7	-0.2	1.8
4	3.0	-4.8	-2.9	-0.5
5	1.6	9.4	-5.5	-0.4
6	0.2	3.7	-1.1	-0.1
7	0.0	-0.4	-13.9	0.6
8	0.0	-1.2	-15.6	0.8
9	0.0	0.0	-19.2	4.4
10	0.0	-0.1	-1.5	2.0
12	0.1		-1.5	
15	0.2		-2.1	
20	0.1		19.7	
25	0.0		15.5	
30	0.0		12.5	
40	0.0		3.1	



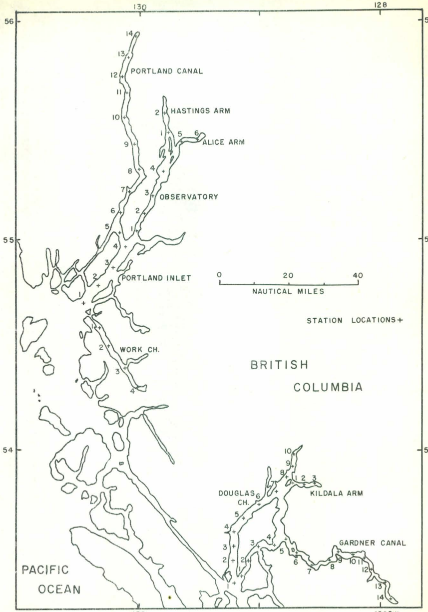
British Columbia Inlets
Figure 1.



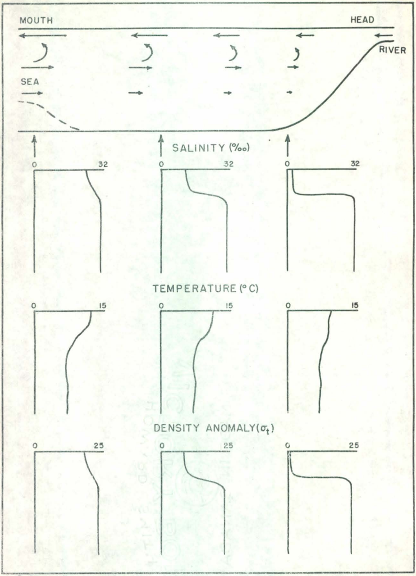
Station locations for inlets in southern British Columbia coast
Figure 2.



Station locations for inlets in Central British Columbia Coast
Figure 3.

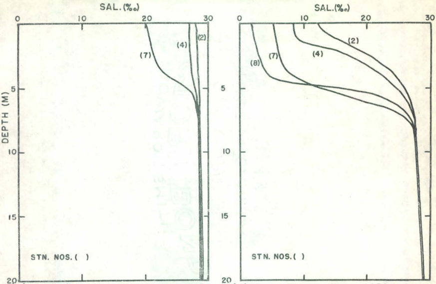


Station locations for inlets in Northern British Columbia coast
Figure 4.



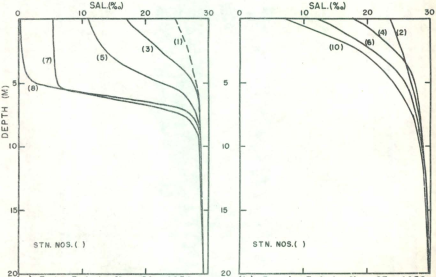
Schematic diagram illustrating structure and circulation in a deep estuary for large fresh water discharge

Figure 5.



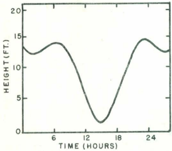
(a) Bute Inlet, February 20,1951 (b) Bute Inlet, May, 1951

Salinity profiles
Figure 6.

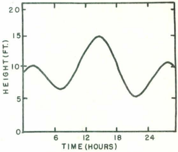


(a) Bute Inlet, May 29, 1952 (b) Jervis Inlet, May 27, 1952

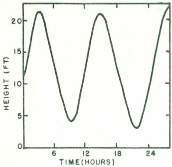
Salinity profiles
Figure 7.



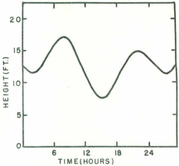
(a) Large range
(southern coast)



(b) Small range
(southern coast)



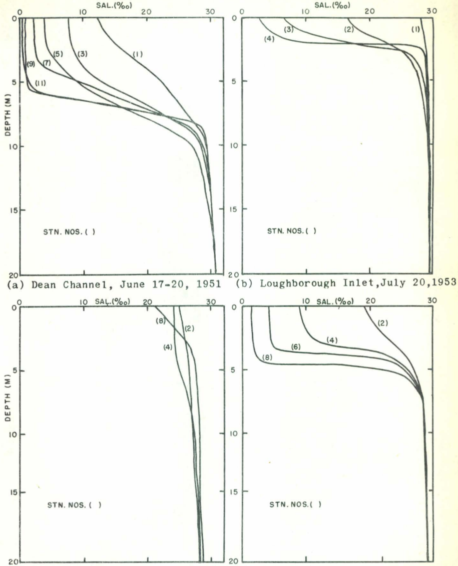
(c) Large range
(northern coast)



(d) Small range
(northern coast)

Typical tide curves for British Columbia inlets

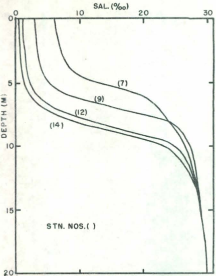
Figure 8.



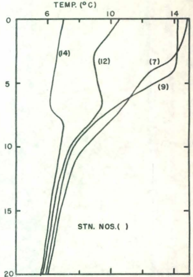
(a) Dean Channel, June 17-20, 1951
(b) Loughborough Inlet, July 20, 1953
(c) Bute Inlet, October 25-26, 1951
(d) Bute Inlet, June 8, 1954

Salinity profiles

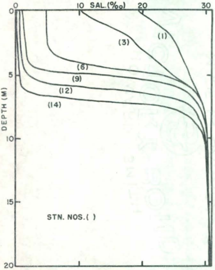
Figure 9.



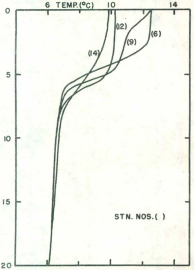
(a) Portland Canal, July 22-25, 1951



(b) Portland Canal, July 22-25, 1951



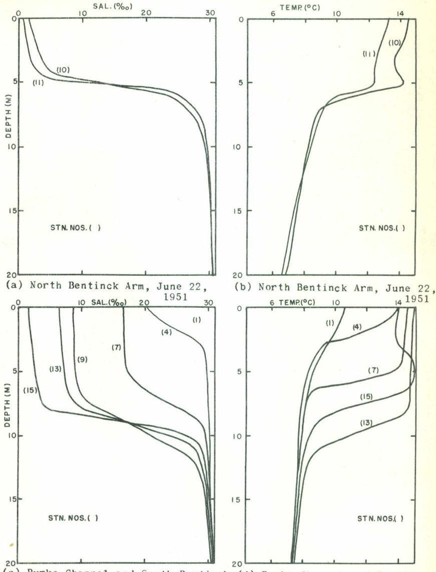
(c) Gardner Canal, July 2-5, 1951



(d) Gardner Canal, July 2-5, 1951

Salinity and temperature profiles

Figure 10.



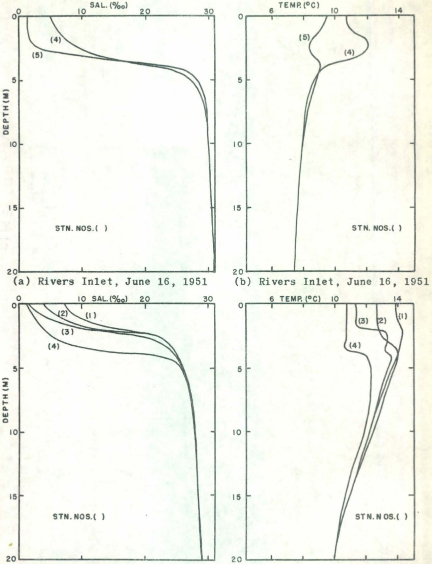
(a) North Bentinck Arm, June 22, 1951

(b) North Bentinck Arm, June 22, 1951

(c) Burke Channel and South Bentinck Arm, June 23-26, 1951

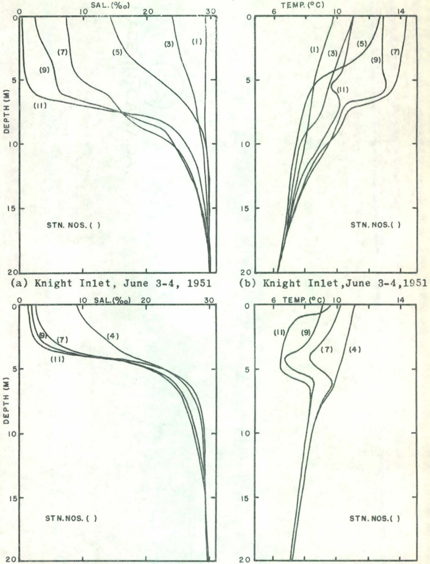
(d) Burke Channel and South Bentinck Arm, June 23-26, 1951

Salinity and temperature profiles
Figure 11.



(a) Rivers Inlet, June 16, 1951
(b) Rivers Inlet, June 16, 1951
(c) Toba Inlet, June 9, 1954
(d) Toba Inlet, June 9, 1954

Salinity and temperature profiles
Figure 12.

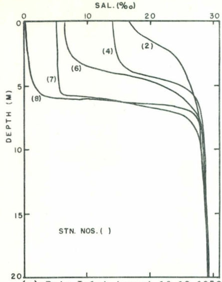


(a) Knight Inlet, June 3-4, 1951 (b) Knight Inlet,June 3-4,1951

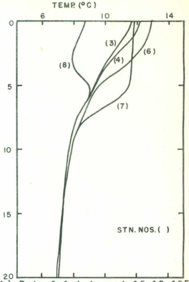
(c) Knight Inlet, July 15, 1953 (d) Knight Inlet,July 15,1953

Salinity and temperature profiles

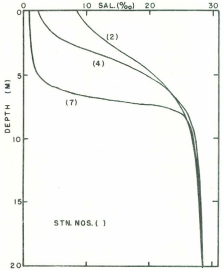
Figure 13.



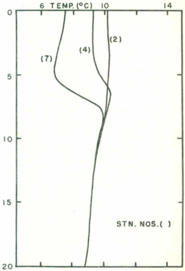
(a) Bute Inlet, August 10-12, 1952



(b) Bute Inlet, August 10-12, 1952



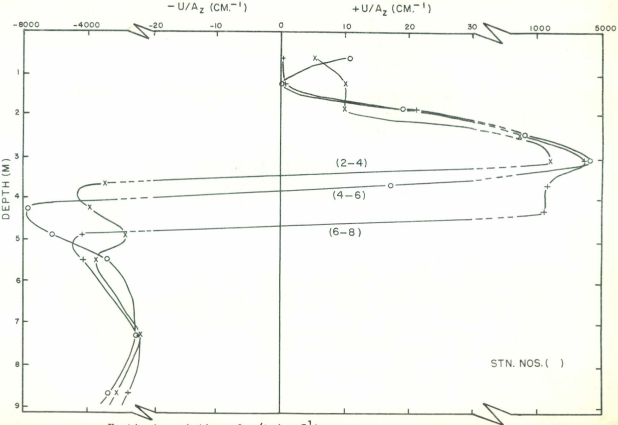
(c) Bute Inlet, August 20-23, 1953



(d) Bute Inlet, August 20-23, 1953

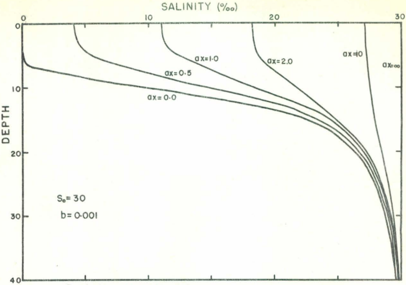
Salinity and temperature profiles

Figure 14.

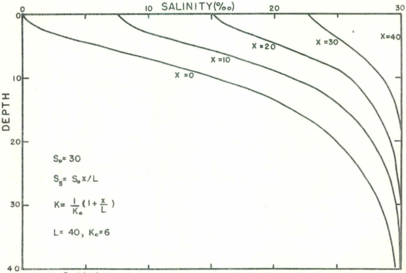


Vertical variation of u/A_z(cm.$^{-1}$), Bute Inlet, June 8, 1954

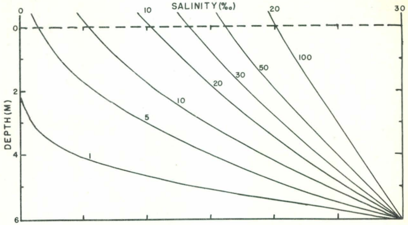
Figure 15.



Salinity distribution given by equation (14)
Figure 16.

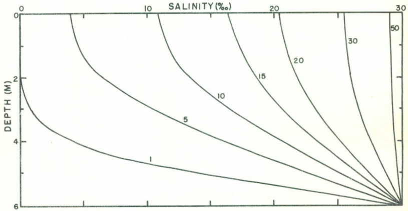


Salinity distribution given by equation (15)
Figure 17.



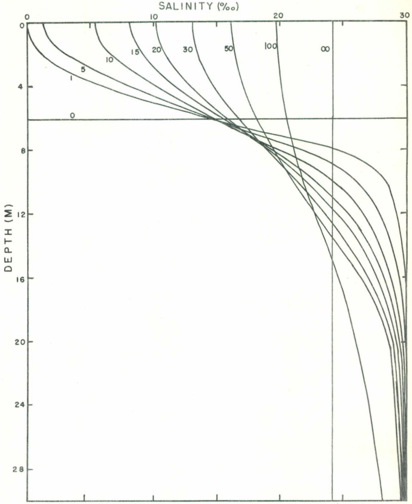
Salinity distribution for various values of A_0x/u (eqn.(21)), with depth $z = 6$ maintained at 30 $^0/oo$ and no upper boundary

Figure 18.



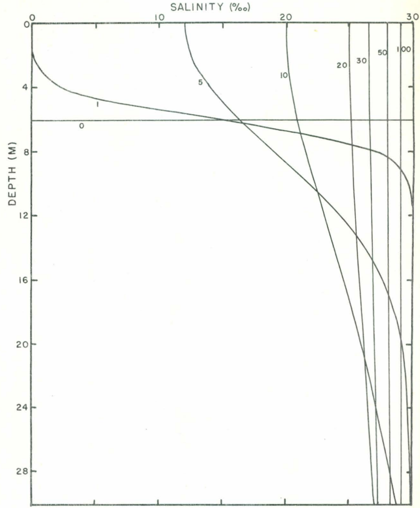
Salinity distribution for various values of $A_0 \xi$ (eqn.(27)), with depth $z = 6$ maintained at 30 $^0/oo$ and no flux across $z = o$

Figure 19.



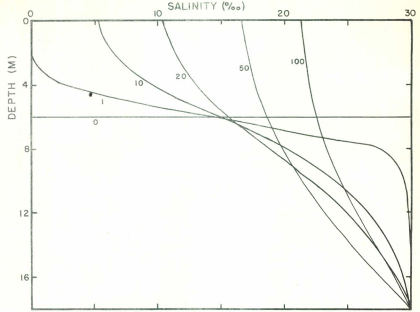
Salinity distribution for various values of $A_0 \beta$ (eqn.(29)), with no flux across z = o and z = 30

Figure 20.



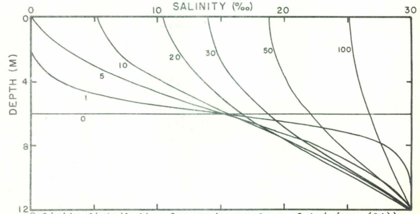
Salinity distribution between z = o
and z = 30 for various values of $A_o \zeta$ (eqn.(32)),
and salinity = 30 ‰ for z = ∞

Figure 21.



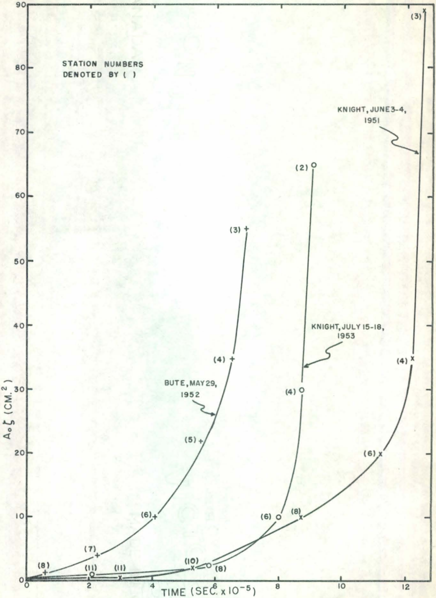
Salinity distribution for various values of $A_0\rho$ (eqn.(34)), with no flux across $z = o$, and $z = 18$ maintained at 30 $^0/oo$.

Figure 22.



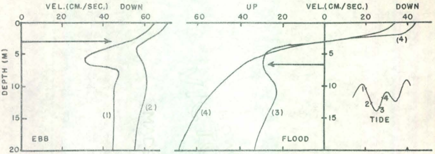
Salinity distribution for various values of $A_0\rho$ (eqn.(34)), with no flux across $z = o$, and $z = 12$ maintained at 30 $^0/oo$

Figure 23.

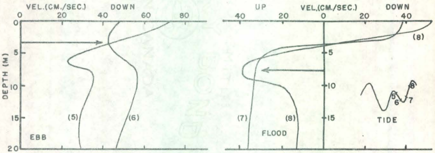


Plot for determining eddy Coefficient of diffusion

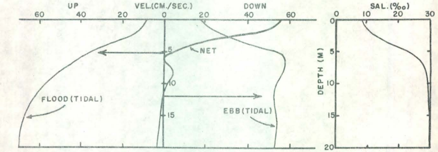
Figure 24.



(a) Flood and large ebb tide velocity profiles



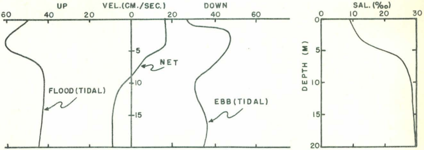
(b) Flood and small ebb tide velocity profiles



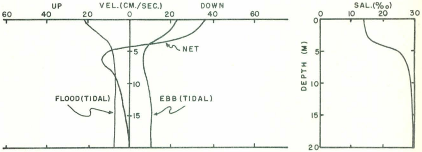
(c) Mean flood and ebb tidal velocities, net velocity, and salinity profile

Velocity profiles, Knight Inlet, Station 4, July 11-15, 1953

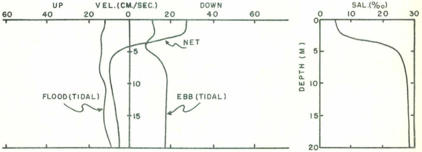
Figure 25.



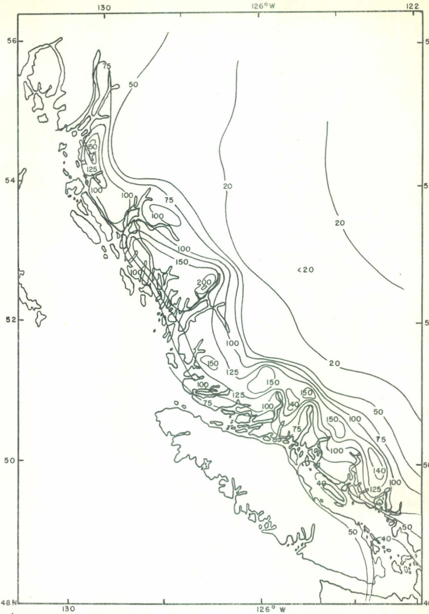
Velocity and salinity profiles, Knight Inlet, Station 4,
August 6-8, 1952
Figure 26.



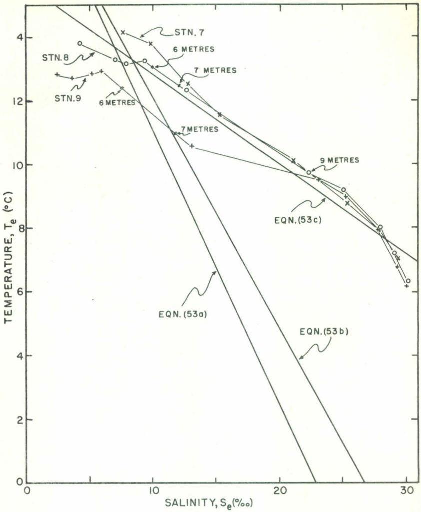
Velocity and salinity profiles, Bute Inlet, Station 4,
August 10-12, 1952
Figure 27.



Velocity and salinity profiles, Bute Inlet, Station 4,
July 1-3, 1953
Figure 28.

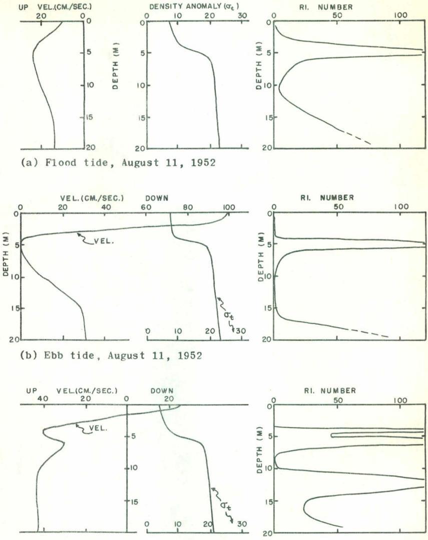


Average annual precipitation (inches) for British Columbia Coast Range
Figure 29.



T-S relations, Knight Inlet, June 3-4, 1951

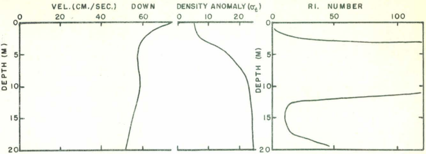
Figure 30.



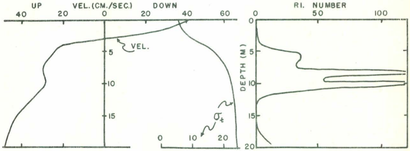
(c) Flood tide, July 2, 1953

Velocity, density anomaly, Richardson number, Bute Inlet, Station 4

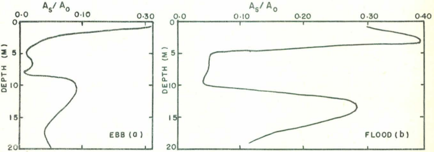
Figure 31.



(a) Ebb tide



(b) Flood tide



(c) Ratio $A_s:A_o$ for ebb and flood tide (parts (a) and (b))

Velocity, density anomaly, Richardson number, and A_s/A_o, for ebb and flood tide, Knight Inlet, Station 4, July 11-15, 1953

Figure 32.