THE PRIMARY PRODUCTION OF A
BRITISH COLUMBIA FJORD

by

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THE PRIMARY PRODUCTION OF A BRITISH COLUMBIA

FJORD

ABSTRACT

The fjord is a characteristic, but biologically little studied marine habitat of British Columbia. The main objective of this study was to determine the annual cycle of primary organic production in Indian Arm, one of the mainland fjords in this area, and to relate this production to the oceanographic factors of the environment.

Data were gathered in Indian Arm at approximately monthly intervals from 1956 through 1959 on 35 cruises. A detailed analysis of the physical oceanography of the fjord was made and a study of primary production in relation to these data was undertaken. The annual patterns of distribution of salinity, temperature, dissolved oxygen, density, and climate were observed. Analyses of these environmental factors, combined with direct current measurements, were used to establish the circulation pattern and replenishment mechanisms of the fjord waters.

During 1958-59, estimates of the annual cycle of primary production were made using three complementary techniques. Two of these were based on the oxygen budget of the fjord. This budget was established from a detailed study of the changes in oxygen distribution which occurred during the period. Changes in the total oxygen content of the fjord were corrected for non-biological processes and the resulting biological oxygen budget was used in the production estimates. In the first technique, the monthly net oxygen changes were considered to represent the amount of photosynthetic material produced in excess of the fjord's total biological requirements. The total for the year was calculated to be 380 grams of carbon per square meter. Secondly, the oxygen utilization budget of sub-euphotic waters provided an estimate of the organic material consumed within the fjord by non-photosynthetic organisms. This was estimated at 290 g.C/m.²/yr. The third method provided a measurement of photosynthetic fixation in natural phytoplankton samples inoculated with C₁₄ (carbon fourteen) and incubated *in situ*. These values were corrected for the observed variations in production potential of various regions within the fjord. This value was estimated to be 460 g.C/m.²/yr., and is considered representative of the total net production of the fjord ecosystem. The difference between the net phytoplankton production and the sub-euphotic utilization provides a measurement of

excess production, and indicates that approximately 25 per cent of the fjord's production was transported out into neighbouring waters.

The total gross production was estimated as 680 g.C/m.²/yr. by the oxygen budget method and 670 g.C/m.²/yr. by the radio-carbon method. These values are in excess of those usually found in continental shelf or oceanic waters, and approach the high values reported for regions of coastal upwelling. The relatively higher production in Indian Arm is primarily related to the seasonal stability of the water column and the effect of this stability in maintaining the phytoplankton population at favourable light intensities and periodically replenishing the depleted euphotic zone with nutrients. The nutrient source appears to be a biological accumulation in the deep basin of the fjord. In this region these nutrients may reach levels higher than those occurring at comparable depths in neighbouring waters.

This study indicates that the estuarial waters of Indian Arm are probably more productive than coastal shelf waters in the same geographic region, and demonstrates that a fjord can be a highly productive ecosystem.

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I. INTRODUCTION

On certain high latitude coastal sections of the northern and southern hemispheres, there are characteristically long, narrow, deep arms of the sea referred to as fjords. Although these fjords are restricted to a small percentage of the total coast line in these latitudes, their unique characteristics and local importance have brought them particular attention. They are usually associated with four regions: the west coast of North America from British Columbia north into Alaska, the west coasts of Norway, Chile, and New Zealand's South Island. In addition they are common along the coasts of Newfoundland, Labrador, Baffin Island, Greenland, Iceland, Scotland, Spitzbergen, Novaya Zemlya, and the Kerguelen Islands.

Although it is not possible to establish a strict definition that applies in all instances, a combination of geological features distinguishes fjords from other types of narrow embayments. A fjord usually exhibits most of the following features: it is located on a glaciated coast which has undergone post-glacial isostatic uplift; it is a basin, usually over 100 meters deep, partially separated from the adjoining sea by a threshold or sill; it is a long narrow waterway with parallel sides or an interconnected system of such waterways; it is U-shaped in profile with steep valley walls.

In the North Pacific "...between the forty-ninth and fifty-ninth parallels the mountainous western margin of North America meets the Pacific Ocean in a deeply indented, island-studded coast ranking with the grandest fjord-coasts of the world in

the character and magnitude of its physical features." (Peacock,² 1935.) Occupying two-thirds of this region is the British Columbia coast, on which are located 39 fjords not including those located on Vancouver Island and the Queen Charlotte Islands. One of these mainland fjords, Indian Arm (Fig. 1), was selected for this study.

Prior to 1948 (Tully, 1936, 1949) fjord oceanography in British Columbia was virtually unknown. Even today many of the fjords remain incompletely sounded and inaccurately charted. Since 1948 several research projects have provided much information on the physical oceanography of the fjords, although the field investigations of most of these were restricted to summer months. Only a few biological projects have been conducted up to this time in the fjords.

The fjords present a unique environment because their physiography makes possible the analysis of many marine problems which would be impracticable to attempt in the open sea. One of the most pressing problems in biological oceanography is the need for the accurate assessment of the potential organic production of the sea. The local assessment of primary production presented an opportunity to establish in detail one aspect of fjord biology and to assess primary production in relation to oceanographic factors.

The primary objectives of this study were: (i) to establish the yearly cycle of primary production in Indian Arm, (ii) to ascertain the yearly cycle of changes in the physical oceanography, and (iii) to relate these changes to concurrent changes in primary production. In order to achieve these objectives, a

three year program was conducted consisting of 35 cruises at approximately monthly intervals from 1956 to 1959. The research was undertaken in three phases.

The first phase comprised a general oceanographic investigation in 1956 and 1957. Standard oceanographic techniques were used to determine the horizontal, vertical and seasonal distribution of chemical and physical properties of the fjord waters.

The second phase consisted of testing and evaluating the equipment and techniques to be used for the measurement of primary production, and was undertaken in 1957-1958. General physical and chemical oceanographic observations were continued during this second phase of the investigation, on a smaller scale, and were supplemented by a study of the bottom sediments, the plankton, and the benthonic communities present.

The last phase was concerned with primary production measurements combined with a continuation of routine oceanographic measurements and plankton collections. Analysis of data from the first two years together with a comparison with some earlier data focused attention upon certain oceanographic features which were examined in detail during comparable seasons in 1958-1959.

The data gathered during this research are presented, analyzed and discussed in two groups. First, the fjord environment and primary production are considered. Secondly, the annual cycle of primary production in relation to physical oceanography is discussed.

A detailed presentation, analysis, and discussion of the physical oceanography of the fjord, the oceanographic equipment, the routine analyses employed, and the treatment of data are presented in the section on General Oceanography. The study of primary production is based upon the fundamental analysis of the oceanographic data presented in this section.

A. Physiography

Indian Arm extends almost due north from a point at $49^{\circ} 18' 10''$ N. Lat. and $122^{\circ} 56' 20''$ W. Long. (Fig. 1). The total length is approximately 22 kilometers; the average width is slightly over one kilometer. In common with most fjords it has a deep basin occupying three-fourths of its length. This shoals rapidly in the south to a sill 26 meters deep at the mouth, and shoals more gradually to the north into the mouth of the Mesliloet River. The fjord sides rise precipitously from a flat basin floor to mountainous heights of up to 1400 meters, with an average slope of approximately 40° .

B. Circulation

The driving force for circulation in Indian Arm is the influx of fresh water from the river and peripheral streams. Accumulation of this less dense water at the head induces a seaward movement of the surface layer. As a result of this seaward movement at the surface the layer entrains portions of the underlying water mass. Continuity considerations thus require that there be a compensating movement of water into the fjord at some sub-surface depth. The circulation pattern therefore is resolved into a two layer system, with net outflow at the surface and net inflow at an intermediate depth.

C. Distribution of Oceanographic Properties

The oceanography of the fjord is dominated by one basic factor: interchange of water between the fjord and the adjoining water mass is restricted to depths at or above the sill. If the conditions in adjoining water masses were constant, the

distribution of conservative properties would be essentially ⁶ uniform and constant from sill depth to the bottom of the basin, and would be similar to those existing at sill depth in the adjoining water mass. Above sill depth they would be continuous with those of adjoining waters.

Departure from this basic pattern is primarily due to two kinematic features: fluctuating fresh water inflow producing variations in the seaward transport of surface water, and the great variability of properties in the adjoining water mass. These variations occur on many different time scales, ranging from diurnal periods to periods up to several years. The inflow of adjoining sea water having different properties and its subsequent mixing produces the major changes in the conservative properties of the water mass.

1. Salinity

The salinity distribution in Indian Arm is characterized by a distinct two layer system: a thin brackish surface layer (0.5 to 5 meters) showing extreme variations and a relatively stable deep basin. Temporal and spatial variations in the surface layer result from fluctuations in fresh water discharge and entrainment of saline basin water. The deep basin water is influenced primarily by the occasional intrusions of high salinity water from the Strait of Georgia and secondarily by the downward mixing of fresh water.

The fjord is a positive estuary where precipitation and fresh water inflow exceed evaporation. The most important sources of fresh water are the Meslilloet River which discharges at the head and numerous streams which enter the fjord along

the sides. An estimate of the total volume of fresh water 7 entering the fjord can be derived from estimates of peripheral stream discharge and river runoff. The total fresh water inflow presents a characteristic annual cycle with a maximum in May-June, a minimum in August, a second maximum in November-December followed by a second minimum in February-March. The first of the two maxima results from the melting of snow and ice that have accumulated over the winter; the second maximum results from the heavy autumn rains characteristic of the region.

While the sill at the mouth restricts deep water communication with Burrard Inlet and Strait of Georgia, free communication above the sill ensures that the water in the fjord is predominantly saline. Thus the salinity distribution of Indian Arm is primarily a function of the mixing of this sea water with the inflowing fresh water of the Mesliloet River and peripheral streams; and the horizontal distribution of salinities in the upper brackish layer usually shows a gradual increase in salinity with distance from the head.

The main mass of water ranges in salinity from 25 to 27%, values 2 to 3% lower than in similar waters in the Strait of Georgia. The upper brackish water layer averages 4 to 5 meters in thickness and ranges in salinity from values approaching 1% just off the mouth of Mesliloet River to 25% at the mouth of the fjord.

2. Temperature

The vertical temperature distribution of Indian Arm also suggests a two layer system similar to that previously noted

for salinity: a thin surface layer with widely fluctuating 8
temperatures, and a deep thick layer with substantially constant temperatures. The thin surface layer is influenced primarily by insolation and secondarily by the inflow of cold fresh water. The temperature of the deep water mass is influenced by intrusion of water from the Strait of Georgia and by the vertical transfer of heat. This transfer produces seasonal effects discernable at depths from 100 to 125 meters.

The temperature distribution presents three distinct relationships between the surface layer and the deeper waters. In late summer and early fall (between the periods of maximum fresh water discharge and during the period of maximum insolation) a characteristic vertical distribution reveals a surface layer of warm water 10 to 16° C., which usually reaches a depth of 30 meters. The lower boundary of this warm water layer is marked by a thermocline, with temperatures of 6 to 8° C. characterizing the underlying water mass. Within this mass, temperatures decrease gradually with depth, the total change being about 2 C.°

During late fall and winter, runoff is combined with a marked decrease in insolation. The resulting vertical temperature distribution is characterized by cold surface waters with temperatures of 5 to 7° C., separated by a thermocline from warmer deep water with temperatures of 6 to 8° C. As with the previous temperature distribution, these temperatures decrease gradually with depth. However, the coldest bottom water is always warmer than the surface water.

During spring, with increasing insolation and low fresh water discharge, the cold surface layer gradually warms until the vertical temperature distribution throughout the water column is temporarily isothermal. As a result of continued warming, the late-summer and early-fall temperature distribution again forms. The large spring runoff temporarily alters this change, particularly at the head of the fjord. At this time a shallow surface layer, colder than the underlying water mass, develops and persists for several weeks.

3. Oxygen

The distribution of oxygen, which is a non-conservative property, differs from the conservative properties previously discussed since oxygen is strongly influenced by biological activity. Thus both biological (photosynthesis and respiration) and non-biological (water movements and diffusion across the water surface) factors combine to establish the seasonal cycle.

The oxygen distribution in Indian Arm is characterized by high surface values, which at times reach 150 per cent saturation and exceptionally as high as 250 per cent. These decrease rapidly with depth to the approximate region of the halocline, and then more slowly to the bottom of the fjord. The relatively stable bottom waters have saturation values ranging from 10 to 30 per cent. In this region, the oxygen concentration is determined chiefly by the non-seasonal replacement pattern with intervening reduction by respiratory oxidation. The surface layers show one marked maximum during April-May-June and a second during September-October.

Seasonally, these coincide with phytoplankton concentrations reported from other temperate coastal regions.

From an examination of the physiography and distribution of oceanographic properties, Indian Arm is found to be comparable to other mainland British Columbia fjords (with the minor exceptions indicated in the appendices).

III. FJORD PRIMARY PRODUCTION

A. Experimental Methods

The greatest problem in productivity research is to obtain an accurate measurement of production. This problem is complicated by the confusion and lack of uniformity in terminology encountered in the literature. It is generally accepted that primary production is ...the amount of organic material which is synthesized in a unit volume of water by the activity of both photosynthetic and chemosynthetic organisms in unit time... Primary production may be considered synonymous with gross plant production, since the amount of material that is contributed by autotrophic bacteria is probably negligible. Net plant production is the amount of material synthesized by the plant, less the amount utilized, stored, or released as extracellular metabolites by the plant.

When an ecosystem such as Indian Arm is considered, the terms gross and net production are used with a slightly modified meaning. Throughout this thesis gross production refers to the total amount of organic material synthesized within the boundaries of the fjord in a unit time; net production is the amount of organic material produced in excess of the total direct and respiratory oxidative requirements.

The physiography of Indian Arm limits the area of substrata suitable for the growth of benthonic algae to less than 5 per cent of the euphotic zone (Fig.). The phytoplankton occupy virtually all the euphotic zone. For this reason it is apparent that primary production originates almost entirely from the activities of the phytoplankton.

The estimate of the rate of carbon fixation in a marine¹² environment has been approached in several ways. Estimates of production (expressed as grams of carbon fixed per unit time) have been made from measurements of the following:

(1) changes in the size of the phytoplankton population; (2) changes in the concentration of non-conservative water properties which are correlated with the activities of the phytoplankton; (3) temperature and available radiant energy (correlated with their effects at a given photosynthetic rate); (4) the rate of photosynthesis of a sample of the phytoplankton population. Each of these methods has certain advantages.

Estimates of the population size are usually based upon a measure of the standing crop. The standing crop is an expression of the density of the population resulting from a balance between the population's growth and depletion. Accurate estimates of the standing crop are possible; but, even if successive measurements are made at short intervals over a period of time, it is difficult to relate standing crop measurements to production. A small population under favourable conditions may have a high rate of production, whereas a large population under unfavourable conditions may have a low rate of production.

Changes in the concentration of non-conservative water properties yield a more accurate estimate of production. Those properties most commonly used are the concentration of nutrients, the environmental pH (correlated with the utilization of CO_2), and changes in oxygen concentration. However, the observed changes within an area are dependent not only upon the biological activity, but are also related to the amount

3

transported to and from the area (in some instances to the¹³ amounts regenerated by organisms). If an assessment of these factors is not made, the estimate of production will be incorrect. An additional difficulty arises in that changes in concentration are often extremely small and require very sensitive methods of analysis in order to establish seasonal cycles accurately.

If the photosynthetic potential of the population can be estimated, it is possible to calculate roughly the primary production from an evaluation of fluctuations in the temperature and available radiant energy. This approach is based on the concept that in a stable marine environment (where nutrients are not limiting) three factors establish the amount of primary production: the size and efficiency of the phytoplankton population; the amount of available radiant energy throughout the euphotic zone; and the temperature.

The measurement of the photosynthetic rate of a sample of the population is often assumed to be the most accurate method of determining production. Basically this method consists of placing a sample of the natural population in a culture bottle and measuring changes in the concentration of non-conservative water properties. The production of oxygen and the uptake of labeled carbon (C^{14}) are the two processes most often studied to obtain an index of the photosynthetic rate.

The oxygen production method has been used extensively by biological oceanographers (Putter, 1924; Gaarder and Gran, 1927; Gran and Thompson, 1930; Riley, 1939, 1941; Riley and Gorgy, 1948; Smayda, 1957). The method consists of placing

samples of the population in clear (light) and opaque (dark) bottles containing water of measured oxygen content. The samples are then replaced in the natural environment at the original point of collection and left for a period of time, commonly some multiple of 24 hours. The light bottle permits photosynthesis to occur at the levels of light and temperature at which the population existed. The dark bottle provides a measure of respiration under the same conditions. The difference in oxygen content between the light and dark bottle provides a measure of the amount of oxygen produced. By assuming that a given amount of carbon is fixed for each mole of oxygen evolved, it is possible to estimate the quantity of carbon fixed. The main disadvantages of this method are the following: the amount of oxygen produced is obscured by the respiration of zooplankton and bacteria present in the natural water samples, and inaccuracies may be introduced by attempting to interpret the observed photosynthetic rate of a small sample of the population in an artificial environment (the bottle), in relation to a large population in the natural environment.

A commonly used modification of this technique involves incubating the light and dark bottles in an incubator aboard ship rather than in situ. While this provides a relative measure of production, it is very difficult to interpret the observed photosynthetic rates in relation to those which occur in the natural environment.

One drawback to the use of the light and dark bottle technique for oxygen is that within such bottles the high surface

to volume ratio often permits a rapid increase in the bacterial population. This increase influences production measurements in two ways. First, since it is usually assumed that all respiration in the bottle is due to the phytoplankton, the bacterial respiration introduces an error in the net production calculations. Secondly, the bacterial population produces an accelerated rate of nutrient regeneration which results in atypical phytoplankton growth rates (Pratt and Berkson, 1959). Both of these effects increase exponentially with time, and thus greatly reduce the accuracy of the results of long term incubations (24 to 72 hours). This necessitates restricting the incubation time to short periods (6 to 12 hours). On the other hand, a significant increase in dissolved oxygen does not usually occur in 6 to 12 hour periods. Consequently, except during periods of very high photosynthetic rates, the oxygen method cannot be used for in situ incubations in fjord waters.

A more sensitive method for determining production is available in the C_{14} light and dark bottle technique. An estimate using the C_{14} method has been described and discussed in detail (Doty, 1954, 1955, 1956, 1957, 1958, 1959; Ryther, 1956b; Steemann Neilsen, 1957; Strickland, 1959), and is similar to the light and dark bottle method for determining oxygen production in that both methods analyze samples of the population. However, it has not yet been developed to the extent desirable, since the exact physiological interpretation of the results is difficult. Both methods measure the rate of photosynthesis, one by measuring the amount of oxygen evolved, the

other by measuring the uptake of C^{14} . A known quantity of C^{14} in a sodium carbonate solution is added to samples of the phytoplankton population contained in light-dark bottles. After incubation (either in situ or in an incubator) the population is filtered out of the culture solution, and the population's specific activity is measured by means of a gas-flow Geiger counter. From the known activity of the quantity of C^{14} which has been introduced, and an estimate of the amount of CO_2 in the water, the amount of carbon fixed by the phytoplankton can be calculated. The method assumes: that the presence of radioactive carbon does not affect the metabolism of the cell; that no C^{14} is incorporated into organic compounds except through photosynthesis; that the rate of assimilation is the same for C^{14} as for C^{12} ; and that no fixed C^{14} is lost by respiration.

Only the first assumption can be used without qualification; it has been demonstrated that the use of radioactive carbon has no adverse short-term effects on the cell's metabolism (Holm-Hansen et al, 1958). The three remaining assumptions require qualification.

The rate of primary carboxylation is somewhat slower for the heavier carbon isotopes than for the normal C^{12} isotope. This introduces a discrimination against the C^{14} isotope which must be taken into account when relating the uptake of C^{14} to the C^{12} fixation in the natural environment. This discrimination is reported to range from 5 to 6 per cent, and a correction factor of 1.05 is normally applied (Strickland, 1959).

An isotopic exchange (dark assimilation) has been reported ranging from 1 to 2 per cent of that assimilated in the light. This can be corrected for by using a dark bottle blank. The isotopic exchange so measured is then subtracted from the amount fixed under similar temperature conditions (Steemann Nielsen, 1952; Weigl et al, 1951). The problem of respiratory loss of photosynthetically fixed C^{14} has not been solved, and is partially responsible for the reported discrepancy between estimates of primary production by the oxygen method and the C^{14} method. This problem is fully discussed by Strickland (1959). A respiratory correction factor of 4 to 6 per cent is customarily assumed. Despite its faults, the C^{14} method offers a means of measuring photosynthesis 50 to 100 times more sensitive than the oxygen method. This permits a measurement of primary production during periods when production is too low to be estimated by the oxygen method.

There are advantages to be gained from both the in situ and the instantaneous approach. The in situ approach is based on environmental changes in non-conservative properties over some appreciable period of time (weeks to months), and provides an integrated estimate of production over that period. On the other hand, the instantaneous method provides a measure of the production of a small sample of the population in an artificial environment at a particular time. In the latter, it is only by means of interpretation of the data in relation to the total natural population that an estimate of production can be obtained. Theoretically, the in situ

estimate of production would be the most accurate. However, its use is usually restricted because of the difficulties in evaluating the physical and chemical processes occurring simultaneously with production in the natural environment.

In Indian Arm, production has been estimated by three methods: instantaneous measurements of the uptake of C^{14} ; the in situ changes in the oxygen budget of the fjord; the effects of temperature and the available radiant energy upon the phytoplankton population. A comparison of these three methods is discussed, and the methods are evaluated separately in the following pages.

B. Primary Production

1. In situ oxygen changes as related to production

The estimate of fjord production, using in situ changes in dissolved oxygen, is derived from the detailed analysis presented in section V. This analysis indicates that the major factors influencing oxygen distribution are as follows:

a. Processes which increase oxygen concentration

- i. diffusion from air to water
- ii. photosynthesis by algae
- iii. introduction with fresh water runoff
- iv. introduction with intruding water masses from Burrard Inlet

b. Processes which decrease oxygen concentration

- i. diffusion from water to air
- ii. respiration by living organisms
- iii. chemical oxidative processes
- iv. seaward transport in outflowing water

The most effective processes which raise the oxygen level tend to occur in surface waters, whereas those which lower the oxygen level occur in deep water.

By evaluating the oxygen budget of Indian Arm and correcting for all non-biological terms, it is possible to calculate the biological oxygen budget, and thus compute the annual cycle of net fjord production.

The change in oxygen content for a small volume of water can be expressed as:

$$\frac{\partial O}{\partial t} = p - r + \left[\frac{\partial}{\partial x} \left(\frac{A_x}{\rho} \frac{\partial O}{\partial x} \right) + \frac{\partial}{\partial y} \left(\frac{A_y}{\rho} \frac{\partial O}{\partial y} \right) + \frac{\partial}{\partial z} \left(\frac{A_z}{\rho} \frac{\partial O}{\partial z} \right) \right] \quad (1)$$

$$- \left[U_x \frac{\partial O}{\partial x} + U_y \frac{\partial O}{\partial y} + U_z \frac{\partial O}{\partial z} \right]$$

where:

A = eddy diffusion

U = advection

p = rate of oxygen production by
photosynthetic organisms

r = rate of oxygen utilization by
direct and respiratory oxidations

ρ = density of sea water

t = time

x, y, z = axes of the coordinate system

This equation (1) is generally difficult to evaluate using data from most oceanic environments, since the data necessary to correct for vertical and lateral transport are usually lacking. In some cases, lateral homogeneity may minimize the effects of horizontal movements, and vertical movements may be analyzed by observing changes in the vertical distribution of conservative oceanographic properties. In most neritic environments, the complex oceanography and rapid advective-convective processes make the equation even more difficult to evaluate. However, an exception to this is the fjord environment. The boundary conditions established by fjord physiography normally restrict communication with the sea to one point. Hence when

the changes in oxygen within the entire fjord are considered, the evaluation of advective-convective processes is primarily concerned with this one opening. Consequently, the annual oxygen budget can be determined with the limited oceanographic data gathered on monthly, one-ship oceanographic cruises.

In Indian Arm, with a right-handed coordinate system in which x is directed along the length of the fjord, y across, and z downward, equation (1) is modified in accordance with the following considerations. The sides of the fjord allow no movement out of the fjord along the y axis, and upon integration throughout the entire volume of the fjord the A_y and U_y terms are eliminated. The boundaries established by the water surface and fjord bottom permit no movement of oxygen along the z axis except by diffusion across the air-sea interface. Upon integration through the whole volume of the fjord, the U_z term becomes zero and the A_z term takes on the value established by the upper boundary. Equation (1) is therefore reduced to:

$$\frac{\Delta O}{\Delta t} = \int \left\{ p - r + \left[\frac{\partial}{\partial x} \left(\frac{A_x}{\rho} \frac{\partial O}{\partial x} \right) \right] - \left[U_x \frac{\partial O}{\partial x} \right] + \left[\frac{\partial}{\partial x} \left(\frac{A_x}{\rho} \frac{\partial O}{\partial z} \right) \right] \right\} dv \quad (2)$$

When changes in the total amount of oxygen within the fjord are considered, the non-biological addition or subtraction of oxygen will be due to the combined A_x and U_x terms in equation (2), plus the diffusion represented by the A_z term. The changes in oxygen concentration, as will be discussed later, can be converted to changes in fixed carbon. Since the analysis of the bottom sediments indicates that little if any organic material is deposited on the bottom, the biological changes in the total oxygen of Indian Arm represent the production and

utilization of organic material. These observed changes can therefore be converted to net fjord production by the equation:

$$\frac{\Delta C}{\Delta T} = \frac{1}{k} (P-R) + A + D$$

$$\text{i.e. } \Delta C = \Delta t \left[\frac{1}{k} (P-R) + A + D \right]$$

where:

C = changes in the amount of fixed carbon

k = constant to convert oxygen to equivalent fixed carbon

P = amount of photosynthetically produced oxygen

R = amount of oxygen consumed by all direct and respiratory oxidative processes

A = net transport of oxygen into or out of the fjord

D = net diffusion of oxygen across the air-sea interface

The oxygen budget of Indian Arm is based on monthly measurements at seven stations along the fjord. At each station the dissolved oxygen content was determined at a series of standard depths (p. 120-121). It is assumed that the integrated oxygen value for the water column is representative of the entire volume of water at the station (i.e. including half the distance to each adjoining station). Summation of these values provides a measure of the entire amount of dissolved oxygen within the fjord at a given time.

The problem of estimating oxygen diffusion across the air-sea interface has been investigated by Redfield (1948). This diffusion is primarily dependent on a pressure head at the

interface, established by the partial pressures of oxygen in the atmosphere and in the water. The pressure head can be considered as the oxygen saturation deficit or saturation surplus of the surface water. If this water is supersaturated, oxygen moves out into the atmosphere; if undersaturated, oxygen moves into the water. The rate of diffusion can be expressed as:

$$P - p = .21 (1-w) \frac{C_o - C_x}{C_o} \quad (4)$$

where:

P = partial pressure of oxygen in the atmosphere

p = partial pressure of oxygen in the water

w = water vapour tension of the atmosphere

C_o = solubility of oxygen in the water

C_x = oxygen content of the water

The total diffusion of oxygen across a given surface area can then be calculated as:

$$D = (k)(E)(P-p) \quad (5)$$

where:

D = diffusion

k = a gas diffusion constant

E = surface area

The exact value to assign for the gas diffusion constant k in equation (5) is open to interpretation. The only comprehensive determinations in marine environments are based on data from the northwestern Atlantic (Redfield, 1948). Two empirically derived constants were calculated: a summer value of

2.8×10^6 ml. $O_2/m.^2/month/atmos.$, and a winter value of 13×10^6 ml. $O_2/m.^2/month/atmos.$ This seasonal difference was attributed to winter conditions at the sea surface. At this time, the actual area of air-water interface present at the surface of the water column was increased, presumably as the result of winter storms with their associated waves and turbulence. Redfield calculated these values from oxygen saturation values reported by Fox (1909). When they are corrected using the saturation values reported by Truesdale et al (1955), and converted to mg. $O_2/m.^2/month/atmos.$, the summer value is 3.4×10^6 . Since the fjord physiography provides a sheltered body of water, it is probable that conditions in the fjord more closely approximate summer than winter surface conditions in the northwestern Atlantic. Therefore the value 3.4×10^6 has been used in diffusion calculations throughout this paper.

The diffusion quantities for Indian Arm, presented in figure 3, have been calculated from a summation of the quantities determined at each of stations A to F.

The net advective movement of oxygen into or out of the fjord results from the oxygen introduced in fresh water runoff and inflowing deep water, less the amount transported out of the fjord in the surface layers. The total volume of runoff has been estimated as indicated on page . The volume and depth of transport into and out of the fjord can be calculated on the basis of the observed distribution of conservative properties. By relating the observed concentrations of dissolved oxygen at these depths with the volume of transport, the total net movement of oxygen into or out of the fjord can be

calculated. These net transport volumes are presented in table I and figure 4.

The observed changes in the total oxygen content of the fjord, corrected for advective and diffusive processes, provide an estimate of the biological oxygen budget of the fjord. The oxygen data for 1956-1957 and 1957-1958 are insufficient to justify this approach during these years. However, adequate data permit an estimate of the budget for 1958-1959 and are presented in table II.

The factor for converting observed oxygen changes to fixed carbon (k in equation 3) is calculated as follows:

$$k = \frac{0.375}{PQ}$$

where:

k = mg. C/unit time/unit volume

$$PQ = \frac{\text{molecules of } O_2 \text{ released in photosynthesis}}{\text{molecules of } CO_2 \text{ assimilated}}$$

The exact value of PQ has been experimentally determined over a wide range of conditions and with a variety of photosynthetic organisms (Rabinowitch, 1945, 1951, 1956; Ryther, 1956b; Strickland, 1959). The reported values range from approximately 1.0 to 1.5. These values depend on several factors, including the type of material synthesized, the immediate source of available nitrogen, the light intensity and the composition of the phytoplankton population. In this study, the value of 1.2 suggested by Strickland (1959) for marine environments has been used.

When the changes in the 1958-1959 biological oxygen budget are converted to fixed carbon, the net fjord production is 380 g. C/m.²/yr.

An analysis of the bottom sediments indicated that little if any organic material was deposited on the bottom. This suggests that most of the organic material produced within the euphotic zone is either transported out of the fjord or utilized by living organisms within the fjord.

A second method of applying oxygen changes to estimate production is based on the oxygen utilized during the respiratory oxidation of organic material, rather than on that produced during the photosynthetic production of organic material. If all the organic material oxidized originated within the fjord, such utilization must provide a minimal measure of the production.

.During this study, 16 periods occurred when the advective processes were restricted primarily to the upper 100 meters. (When occurring below 100 meters they were weakly developed and influenced only the most southern end of the basin.) During these periods the deep basin oxygen concentration decreased steadily with time as a result of the respiratory utilization of dissolved oxygen.

Possible errors in this approach would be introduced by any advective movements of water from outside the fjord during these periods or by vertical mixing within the fjord. Examination of the waters at sill depth outside the fjord shows that if these waters moved into the deep basin they would have oxygen concentrations higher than those present within the basin. The vertical oxygen gradient is such that any downward mixing would also introduce water with higher oxygen concentrations. Therefore both processes (if they occurred) would tend to reduce

the observed rate of oxygen depletion and make any calculation of production values minimal.

In order to relate the observed oxygen depletion to production, it is necessary to assume that the observed depletion is typical of the entire fjord. But this is not the case. While the vertical migrations of the zooplankton suggest that throughout a 24-hour period zooplankton respiration is roughly the same for the entire water column (McHardy, personal communication), the same cannot be said for bacterial and phytoplankton respiration. A priori, as a consequence of the density distribution, one would expect the rate of bacterial decomposition of organic material to be at a maximum somewhere in the upper 100 meters of the water column. This would most likely occur in the region of a pycnocline (a region of rapid density change) slowing or preventing the sinking of organic material. The heaviest concentrations of phytoplankton are found in the upper 50 meters. Therefore phytoplankton respiration is probably highest in this depth range. Both of these processes suggest a higher rate of oxygen utilization in the upper 100 meters of the water column than in the basin waters. Consequently, errors introduced by assuming that the observed oxygen depletion in the deep basin is characteristic of the entire fjord make any calculated production values minimal. Thus the oxygen-utilization approach provides an estimate of the lower limit for fjord production.

The curve representing the annual cycle of oxygen utilization varies in magnitude from year to year, but it should have about the same shape each year. Thus the 16 calculated

depletion values have been plotted on a single figure (fig. 5) to establish the shape of this curve. Integration of the area under the mean curve provides an estimate of mean oxygen utilization for the three years. Integration of the curves representing the highest and lowest values provides an indication of the year to year variations. These values converted to fixed carbon yield a mean of 290 g. C/m.²/yr. with extremes of 160 and 430 g. C/m.²/yr.

Since the oxygen-utilization and the oxygen-production approaches measure two different fractions of gross fjord production, their summation provides an estimate of the total production. The estimate of mean yearly utilization of organic material produced within the fjord was 290 g. C/m.²/yr. The estimated net fjord production for 1959 was 380 g. C/m.²/yr. Thus, if these values are representative, the gross fjord production for 1959 was 670 g. C/ m.² / yr.

2. Measurement of production by the C₁₄ method

The radiocarbon method of estimated production used in this study is based on the incubation of natural phytoplankton samples at the depth from which they were gathered. This provides a measure of the production occurring under in situ light and temperature conditions. It is in direct contrast to the method usually employed where samples are placed in an incubator under standard light and temperature conditions. While the latter method provides an accurate measure of the relative production of different areas, it provides little information about the actual production of a specific area.

The techniques of sample handling and counting are outlined in detail in section V, and, in general, follow the procedure developed by Doty (1956). In this study, samples of water with their contained phytoplankton were collected from a series of standard depths (0, 5, 8, 12, 19, 30 meters) using a Van Dorn type plastic water bottle (Van Dorn, 1956). The samples were brought to the surface, inoculated with C^{14} and then suspended at the depths from which they were collected. In order to minimize the effects of ship shadow, the point of suspension was the end of a 6 to 8 meter boom attached to the side of the ship. Usually the in situ incubations were carried out from sunrise to sunset. On all but one occasion, full day incubations were carried out at station C. On this one occasion, and in some of the supplementary incubations when time was limiting, the samples were incubated for half a day. In these cases, the incubations were from sunrise to local apparent noon or from local apparent noon to sunset. The results of the half-day incubations were doubled to obtain an estimate of the production that occurred during the entire day.

At the completion of the incubation the samples were filtered first through silk net discs (least aperture dimension 56μ .) to determine the relative proportion of production attributable to net plankton; and secondly, through Millipore filters (HA, pore size 0.45μ) to determine the relative proportion attributable to nanoplankton. The silk discs and Millipore filters were subsequently counted using a gas-flow Geiger counter. The counts, corrected for isotope effect, were converted to mg. C/m.³ using carbonate-carbon values calculated

from salinity, temperature, and pH values by the method described by Harvey(1957). These fixed carbon values were plotted on a log-linear scale as a function of depth, and interpolated at one-meter depth intervals. (See Fig. 6.) A summation of these values was made to represent the total carbon fixed under 1 m.² of the sea surface per day. These values are presented in table III.

Depending on the particular distribution of the phytoplankton population within the fjord, the production may vary with position. Since the object was to determine the mean production of the entire estuary it was necessary to gain some measure of the relative production of the various sections of the fjord. During most cruises it was impracticable to make in situ C¹⁴ production measurements at more than one station in the fjord. However, the use of a shipboard incubator provided a source of supplementary data. During 1959 a slightly modified version of the "Salpa tank" incubator (Doty, 1957) was used. A series of samples was collected at four stations along the fjord and incubated under uniform light and temperature conditions for a period of 24 hours. Relative production values thus determined by the light and dark bottle O₂ technique were used to weight the in situ C¹⁴ measurements of production at a particular station in order to estimate the production of the entire fjord. When time permitted more than one in situ incubation, each series was corrected to provide production values for the entire fjord. Subsequently the values were averaged to provide a mean production estimate for the month.

Unfortunately there is still some difference of opinion as to exactly what the measured fixation of C^{14} represents. Steemann Nielsen (1952, 1958a), the originator of the technique, felt that it yielded a value between net and gross primary production. However, Ryther (1954) has suggested that the technique measures net production alone. An excellent review of this question is given by Strickland (1959), who concludes that the method measures a value between gross and net production, but probably closer to the latter. There is no reason to assume that the value obtained is always at the same point between net and gross primary production. It is likely that it varies with the taxonomic composition of the phytoplankton population as well as its physiological state. In practical field measurements, the question of whether the technique measures net production or a value slightly above it is not serious.

The proportion of respiration to photosynthesis in marine environments ranges from approximately 10 to 50 per cent. If respiration was 50 per cent of photosynthesis, the technique would be measuring 1.0 times net production, or 50 per cent of gross primary production. Conversely, using Steemann Nielsen's corrections, such a value would represent 70 per cent of gross primary production (Strickland, 1959). Therefore the two extreme interpretations vary by only 20 per cent. Actually, respiration usually ranges from 10 to 25 per cent of photosynthesis, and the difference between the two possible interpretations is proportionally smaller. Considering the degree of

accuracy of most field determinations, it is sufficient to assume that the C^{14} data measures net primary production and that any error introduced can be ignored (Doty, 1959; Ketchum et al, 1957; Menzel and Ryther, 1959). The C^{14} data gathered during this study are assumed to represent net production.

A more serious limitation in the C^{14} method is that it does not provide information on the amount of material respired, and the net production measured is only daylight net production. Therefore it does not yield information on the gross production or on the ecologically more significant 24-hour production.

With a phytoplankton population that is essentially homogeneous throughout the euphotic zone, it is possible to estimate respiration by an extrapolation of in situ C^{14} data (Steemann Neilsen, 1959). Unfortunately, this condition has not been observed in Indian Arm and respiration must be estimated in some other way.

During 1959 the shipboard incubator used for determining the relative photosynthetic rates of different regions of the fjord at the same time provided information on the ratio of respiration to photosynthesis. Since dark and light bottles were used during the incubations (with a measurement of the initial oxygen concentration prior to incubation), estimates of the rate of photosynthesis and respiration were calculated. The light level in the incubator was 7.0×10^{-2} ly./min., which is the reported optimum light intensity for diatoms, and just below the optimum light intensity for dinoflagellates (Ryther, 1956a). This is approximately the light level at which the highest production rates occur in situ. While the rates

obtained in the incubator vary in magnitude from those occurring naturally within the fjord, the ratio of respiration to photosynthesis will probably be about the same. This assumption is partially justified when the calculated values are compared with the in situ seasonal cycle of a Danish fjord with a similar annual production cycle (Fig. 7). However, even if an error of 10 per cent were introduced in estimating per cent respiration, the shortness of the night during the periods of high production, coupled with the size of the respiration-correction, combine to reduce the error in the calculated 24-hour net production to a few per cent. Therefore, having discarded the exceptional value in which respiration exceeded photosynthesis (a situation attributable to the enclosure of large zooplankton in the incubation samples), and all data in which the observed photosynthesis or respiration was less than .05 mg./l. O_2 (a value approaching the limit of accuracy of the chemical analysis), the percentage of respiration to photosynthesis was calculated. These values are presented in figure 12.

If the day and night lengths were equal, these respiration values (when subtracted from the observed net primary production) would provide an estimate of the 24 hour net production. However, at 50° N. latitude, the ratio of night-hours to day-hours varies from approximately 0.5 during June to 2 during December. Therefore any correction for respiration at these latitudes must take into account the changes in the length of day.

Thus the 24 hour net production can be calculated from the following equation:

$$P_1 = P_2 - \frac{P}{R} (P_2)(T)$$

where:

P_1 = 24 hour net production

P_2 = observed daylight net production

$\frac{P}{R}$ = observed ratio of net photosynthesis to respiration

T = ratio of hours of daylight to dark

Using the available data, the 24 hour net production has been calculated for 1958-1959. By adding the estimated amount of respired material to the observed C^{14} production, a value for gross production can be calculated. The total net and gross primary production for the year is 450 and 680 g. C/m.², respectively.

The chief error in this method of converting the observed C^{14} net production to 24 hour net and gross production arises in regarding the observed respiration in the incubated dark bottle, as representing phytoplankton respiration alone. Actually the respiration of the bacteria and any zooplankton contained in the sample is included in the calculated "phytoplankton respiration." This error would render the calculated phytoplankton respiration too high, and the calculated net production too low. Therefore the net production, while possibly in error, still provides a minimal estimate of net primary production.

As previously discussed, this method of measuring production suffers from the disadvantage that the observations are usually made on only one day at approximately monthly intervals. In order to gain an estimate of the total annual production, it is necessary to consider these values as representative of the time interval between observations, or to adjust the values to allow for changes in the environmental factors which may influence production. The primary factors determining production are the size and efficiency of the phytoplankton population, the relationship to euphotic zone nutrients, and the availability of radiant energy. Over short term periods, light and population efficiency are assumed to be the most important; primary production is usually approximately proportional to light energy up to certain limiting values, which are characteristic for a particular population and its physiological state. Data were available on both the photosynthetic efficiency of the population and on variations in the daily total incident radiation.

The daily radiation pattern was determined from pyroheliometer measurements continually recorded in Vancouver at a point 24 kilometers due west of the fjord mouth. If one assumes that primary production is proportional to surface radiation, and that the radiation measured is representative of Indian Arm, then it is possible to correct the observed instantaneous C^{14} measurements of production. These corrections make allowance for variations occurring in the radiation pattern between monthly measurements of production. Such corrections do not fully take into account the effects of other

factors such as nutrients and temperature. These will be partially reflected in the observed production efficiency determined at the beginning and end of a particular period of time. Since such corrections alter the total annual production by less than 15 per cent, detailed consideration of changes in other factors is considered to be an unnecessary refinement at the present stage of development of the C^{14} method.

The curve for gross primary production presented in figure 8 has been adjusted for variation in total light input to the fjord, and is considered representative of the year. The yearly cycle of production follows a bimodal curve, rising steeply from a mid-winter low during December and January to a spring maximum during April. It then tapers off slowly to a mid-summer minimum in July, rising steeply again to a fall maximum in September. In general, the net primary production cycle follows the same cycle, except that the spring maximum occurs in May rather than in April. Daily gross production ranges from 0.02 g. to 4.5 g. $C/m.^2/24$ hrs. with an annual mean of 1.9, and net production ranges from 0.01 to 3.2 g. $C/m.^2/24$ hrs. with an annual mean of 1.3. When the production of the entire fjord is considered, net primary production ranges from a low of 35 per cent of gross primary production during the beginning of the spring maximum in April, to a high of 87 per cent during the latter portion of the fall maximum, and has an average of 65 per cent for the entire year. Thus there is a tendency for the proportion of net to gross production to be lowest during periods of rapid growth (e.g. at the time of the development

of the spring or fall rapid increase in population--the so-called "bloom"), and to be the highest during the periods of decline. This may be associated with the phenomenon discussed by Rabinowitch (1959), who reported that respiration is markedly stimulated by the products of photosynthesis. Presumably phytoplankton with high photosynthetic rates are also in the process of converting photosynthetic products to new cellular material, and thus have a high rate of respiration. Phytoplankton with lower photosynthetic rates are not in an active growth phase, and therefore there is a closer relationship between net and gross production. This relationship between a net-gross production ratio and the growth phase of a population is not unique. Similar trends have been reported for the Sargasso Sea (Menzel and Ryther, 1959) and continental-shelf waters off the northeastern United States (Ryther and Yentsch, 1958).

The adjusted production values for the fjord are 450 and 680 g. C/m.²/yr. These are based on a summation of production values for both the nanno- and net-phytoplankton fractions of the incubated population sample. Comparisons of the fractions indicate that the nannoplankton were responsible for all fjord primary production during 1958/59. Only on two occasions, in February (90 per cent) and in September (89 per cent), was their contribution observed to fall below 92 per cent; and during the period from May through July their average contribution was over 99 per cent.

The period of time between observations makes it difficult to relate the changes in nannoplankton contribution to

bloom development. However, it is suggestive that the series prior to the spring bloom shows a steadily decreasing contribution from the net phytoplankton; 10 per cent in March, 5 per cent in April, and 0.5 per cent in May. The converse trend is present during the fall bloom. This suggests that the nanoplankton are relatively most important during the spring bloom and least important during the fall bloom, and is partially supported by standing crop measurements which show a marked peak in September of 1959.

IV. ENVIRONMENTAL CONTROL OF PRIMARY PRODUCTION

The annual cycle of production in Indian Arm is established through the influence upon the photosynthetic population of environmental factors which influence its metabolism and affect its spatial distribution. These factors are not necessarily mutually exclusive, but their interaction and combination determine the production characteristics of the ecosystem.

A. Factors affecting metabolism

The phytoplankton population of the fjord is in a state of flux as a result of the constantly changing oceanographic factors. Also, with each new set of factors, the species best adapted to a particular environment develop and dominate the phytoplankton population. Under these conditions, factors which may influence the production rate of a particular species under laboratory conditions decrease in importance in the fjord ecosystem. As conditions become unfavourable for one species, another better adapted one replaces it. It therefore follows that in a broad sense, the factors which can be expected to influence the production of the population are: the amount of available light, concentration of essential nutrients, and temperature.

1. Light.

The energy necessary for the photosynthetic production of organic material is provided by light. It is now known that some species can adapt to changing light intensities, particularly in raising the photosynthetic efficiency under conditions of low light intensity (Kok, 1951). The major groups of algae

making up the phytoplankton have different light requirements (Strain, 1951). The concept of seasonal succession and adaptation of species under different conditions has been pointed out by Rodhe et al (1958). Steemann Nielsen and Jensen (1957) conclude that in the sea as a whole (excluding the Arctic and Antarctic), light is not a limiting factor in production. Sufficient light energy is usually available for the production that can occur with the nutrients available in the euphotic zone. However, physical movements of the water can carry the phytoplankton to depths lacking sufficient light for maximum production.

The comparison of the annual production cycle in Indian Arm (Fig. 9) with the seasonal availability of radiant energy points out the significance of the above conclusions. The year's production minimum occurred in December and January, with a daily production of 34 and 23 mg. C/m.²/respectively. By February the production rate had increased to 732 mg. C/m.² and it continued to increase until the spring maximum of 3.68 g. C/m.² was attained in May. This increase in production during the spring months was not associated with an increase in available radiant energy. Quite the contrary, the production increase preceded the radiant energy increase by approximately two months. At the time of the February upswing in production, the available radiant energy was not appreciably different from the preceding two months, which had the lowest production values of the year. This suggests that factors other than radiation affected both the time and magnitude of the mid-winter low of production. This midwinter low is probably

related to the maximum surface transport occurring at this time.

2. Nutrients

The problem of the nutrient-production relationship is much more complex than was originally suggested by Brandt (1899), who hypothesized that phytoplankton production was dependent on the availability of only nitrates and phosphates. Subsequent work tended to support this concept. Consequently, because the ionic ratio of the sea is quite constant, there was a tendency to evaluate the effects of nutrients on phytoplankton growth by the consideration of one element alone. The ease of analysis for phosphate-phosphorus concentrations produced a large amount of data concerning the relationship between phosphate and production, with the implication that such relationships held for the other nutrients as well. However, Cooper (1937) showed that the ratio of nutrients departed significantly from the reported "normal" ratio, particularly in enclosed seas (and by analogy, in estuaries). This concept was supported by Riley and Conover (1956) in their studies in Long Island Sound, where they found nitrogen:phosphate ratios of 8:1, about half the 15:1 reported for the open sea (Sverdrup, 1942). When the possible variation in primary nutrient ratio is considered, combined with the fact that silicate, iron, manganese (Harvey, 1957), organic growth factors (Provasoli and Pitner, 1953), and ectocrines (Lucas, 1949, 1955) can also be limiting, the impossibility of determining a relationship between nutrient levels and production without a complete series of nutrient analyses is apparent. Lacking

such analyses, an estimate can be gained by considering the effect on nutrient level of the various known physical processes.

In Indian Arm the nutrient level is established by a combination of biogenic processes (including bacterial regeneration and algal utilization), and various transport mechanisms. The surface brackish layer occupies a large portion of the euphotic zone. As shown in figure 10, 90 per cent of total production of the water column usually occurs at depths less than 5 meters. The level of nutrients within this layer depends upon the amounts of the nutrients contributed by fresh water discharge and the amounts entrained with the underlying saline water. Since the river appears to bring in few nutrients (p. 113), the level of nutrients in the euphotic zone results primarily from the concentration of nutrients in the underlying waters and the mechanisms which move them vertically.

The close correspondence between stability and production (Fig. 11) suggests that this factor plays an important role in establishing the type of production cycle in Indian Arm. Two mechanisms relate stability to production. High stability favours the maintenance of the phytoplankton population in a zone of favourable light intensity but restricts the vertical transfer of nutrients. Low stability favours the vertical replenishment of nutrients in the euphotic zone. However, it allows also the phytoplankton population to be moved vertically to less favourable mean-light intensities. Thus, the vertical movement of the phytoplankton may take them through regions of

both favourable and unfavourable light conditions, and their production results from the mean light energy received. Prior to the initiation of each production peak, a period of low stability occurred, whereas, during each production peak the stability was increasing. This particular cycle of stability then can be visualized as acting in the following manner. The period of low stability permits a replenishment of nutrients in the euphotic zone, followed by a period of high stability favourable to the maintenance of production by the population at favourable depths. However, continued high stability results in the depletion of nutrients which may then become limiting factors.

3. Temperature

A comparison of the annual cycles of temperature and production in Indian Arm (Fig. 9) shows little if any direct correlation. A possible exception to this occurs in July, when the maximum temperature coincides with the midsummer production minimum. This general lack of correlation might be expected for two reasons. First, the phytoplankton population is in a constant state of flux as the species complex changes in adaptation to fluctuating environmental conditions. As a result, there is a strong tendency for species to exist under their optimal temperature requirements. In addition, great adaptation to changing temperatures is possible (Steemann Nielsen, 1955). Secondly, temperature would be expected to act metabolically, either through an effect on photosynthesis or on respiration. It has been reported that under conditions where light is not the limiting factor (the condition existing

throughout most of the euphotic zone), there is little or no direct temperature effect on the photosynthesis of phytoplankton (Baly, 1935; Barker, 1935; Riley, Stommel and Bumpus, 1949).

The same cannot be said for respiration. A priori, one expects the rate of respiration to increase with temperature. This appears to be supported by the few laboratory studies that have been conducted with phytoplankton (Riley, 1949). If the same population were present under all temperature conditions, an increase in respiration would result in a low value of observed net primary production. However, microscopic studies of the phytoplankton show that this is not the case. With the ever-changing species composition of the phytoplankton, the direct effect of temperature on primary production is difficult to establish. In investigations conducted in Scandinavian fjords, no direct relation was noted between primary production and temperature (Stemann Nielsen, 1940, 1951). This conclusion is partially supported by my data from light and dark bottle determinations of photosynthesis and respiration in Indian Arm.

During 1959, the determinations of relative production in the fjord provided 7 sets of light and dark bottle measurements of photosynthesis and respiration. In all, a total of 49 light and dark bottle-pairs (gathered throughout the euphotic zone) were incubated under standard light conditions and at temperatures within 2°C of in situ temperatures.

A comparison of respiration with gross photosynthesis should show any gross relationship between respiration and temperatures. A graphic presentation of this relationship, as shown in figure 12, indicates no apparent direct relationship.

An increased temperature has a secondary effect of some importance upon nutrient regeneration. Under conditions restricting vertical transfer, nutrients may become limiting in the shallow surface layer which makes up most of the euphotic zone. At this time bacterial regeneration of nutrients may occur and prolong the period of production. However, at such times the high temperatures also contribute to the increase in water column stability. It is doubtful whether such regeneration would compensate for the resulting lower nutrient input to the euphotic zone. However, it could prolong the spring production bloom.

B. Factors affecting population distribution

The horizontal and vertical circulation pattern of the fjord controls the recruitment, depletion, and distribution of the phytoplankton population. At times the fjord waters and their contained phytoplankton are completely replaced. One such replacement took place in the spring of 1957, and a similar flushing appears to have taken place in the spring of 1960. During periods between flushing, the population is continually depleted by two mechanisms. First, during each tidal exchange, a portion of the population near the mouth is transported out of the fjord. Secondly, the continual surface outflow removes portions of the surface layer (0 to 10 m.). This depth range includes the upper portion of the euphotic zone. The rate of such outward movement is associated with the annual cycle of freshwater discharge into the fjord. At times, for example during the 1959 spring maximum discharge, transport values in excess of 400 m.³/sec. occurred. At this rate, a volume equal

to the upper 10 meters of the entire fjord water mass (and its contained phytoplankton) would be transported out of the fjord every 51 days. At the other extreme, during the 1959 summer minimum, the observed transport was $40 \text{ m}^3/\text{sec.}$, a rate which would take 52 days to remove an equal volume of water.

This transport of water out of the fjord removes a portion of the photosynthetically effective population. However, it also brings about an increased entrainment from underlying waters. These vertical movements tend to replace with "seed stock" the phytoplankton removed from the euphotic zone of the fjord. The relative effectiveness of these two mechanisms is difficult to establish. It is noteworthy that a comparison of the seasonal pattern of the runoff with the seasonal distribution of production shows a nearly inverse relationship (Fig. 11). The spring-production bloom begins at the same time that the spring runoff slackens, and the fall-production bloom diminishes at the beginning of the winter runoff.

The two chief inflow mechanisms of fjord circulation (tidal action and compensating inflow) recruit phytoplankton, but on a smaller scale. During each tidal cycle a volume of water equal to the tidal lens of the fjord is introduced. The extra-fjord phytoplankton contained in this water may or may not join the effective phytoplankton population of the fjord. There is a strong tendency for the inflowing tidal water and compensating water to move in at depths well below the euphotic zone. Under these circumstances, the recruited phytoplankton may not survive to add to the effective phytoplankton population. An exception occurs during late summer, when appreciable

quantities of the inflowing waters move in at depths just below the surface brackish layer. These depths vary from 10 to 20 meters. Phytoplankton introduced in this way are added to the effective phytoplankton population of the fjord.

The vertical circulation plays an important role, both in relation to euphotic zone nutrients and in the maintenance of the population under conditions of favourable light. The relative effects of salinity and temperature gradients on stability, and the time lag that can develop between unusual seasonal fluctuations in temperature and mid-depth stabilities, are discussed in section V.

During most of the year, the vertical salinity gradient plays the major role in determining water column stability. However, this is not always the case. During July 1959 (Fig. 34), the temperature was equally effective in establishing the stability at the 0 - 25 meter level. With increasing depth the temperature became more important in establishing stability. Because of the vertical transfer of nutrients, the stability of the deeper waters is important in establishing the nutrient levels in the euphotic zone. The temperature conditions during the previous summer can establish the stability structure in the deep portion of the water column during the following spring. Conceivably, as a result of high surface temperature during the preceding summer, higher than normal stability conditions may develop the following spring. These could restrict the vertical transfer of nutrients, and might result in lower production during the following spring and summer.

V. PHYSICAL OCEANOGRAPHY

A. Physical Features1. Physiography and geology

Indian Arm extends almost due north into the coast range from a point at $49^{\circ} 18' 10''$ north latitude and $122^{\circ} 56' 20''$ west longitude, thus placing it near the southern end of the British Columbia fjord system. An aerial view (Fig. 1) shows a typical fjord appearance--a long, narrow embayment bounded by a precipitous mountain wall. The total length of Indian Arm is 22.1 kilometers and average width is 1.3 kilometers, yielding a length/width ratio of 17. The major portion of the inlet is characterized by a deep basin with an average depth of 200 meters and a maximum depth of 224 meters. Figure 2 shows station positions in Indian Arm and longitudinal and transverse bottom profiles. Indian Arm follows a somewhat sinuous course from the mud delta of the Mesliloet River, which marks the northernmost end of the fjord, to the wide, shallow sill at the mouth. Two features deserve particular comment: the maximum depth of 26 meters at the sill; and the constriction which exists $1\frac{1}{2}$ kilometers north of the mouth of the fjord. As will subsequently be discussed, these features play an important role in the dynamic oceanography of this fjord.

The formation of the British Columbia fjord system has been investigated in some detail (Bancroft, 1913; Carter, 1933; Peacock, 1935) with the conclusion that both tectonic movements and glacial action have been instrumental in the formation of the present system. Tectonic forces and movements produced a fractured and deeply dissected plateau before the beginning of

the latest glacial epoch. A combination of longitudinal and horizontal fault patterns guided the erosive action forming an immature valley system which was subsequently subjected to glacial excavation during the descent of glaciers from the major ice sheets. The thickness of the glaciers prevented their floating when they reached sea level and thus their scouring action extended well below sea level. The characteristic fjord structure subsequently resulted from a combination of local subsidence and glacial retreat, and the glacial-scoured, drowned-valley system formed two patterns of fjords in British Columbia, running primarily NE-SW and NW-SE.

The basic fjord pattern of two systems intersecting at 90° divides the majority of fjords into two groups. Most of the mainland fjords, which are in the first group, lie transversely to the mainland coastline. These fjords normally receive the drainage of a river at the eastern end and are open to the sea at the western end. The second group tend to lie parallel to the coastline and often are smaller in size. These may connect two or more of the mainland fjords.

The first group, represented by 33 fjords in British Columbia, is considered to be the most typical of the B. C. coast. General physiographic features are similar within this group. They range in length from 6.5 to 114 kilometers and in average width from 0.6 to 6.0 kilometers. The width of fjords appears to be correlated with length. With only four exceptions, the ratio of fjord length to width is between 12 and 50. The average ratio is 20. The mid-inlet depth of the fjords ranges from 76 to 510 meters, with an average depth of 270

meters; the maximum depth ranges between 144 and 744 meters, with an average maximum depth of 388 meters. The depth of the sill is between 6 and 363 meters, with more than 50 per cent being less than 75 meters. The greatest physiographic differences between fjords are found in the wide variations of sill depths at the entrance and in the presence or absence of inner sills.

Physiographic features and dimensions of Indian Arm place it within the main group of British Columbia fjords. Its length and width are slightly less than average, but its length/width ratio is typical. Its average and maximum depths are also slightly less than average. It differs in having a shallower sill than most B. C. fjords. The 26-meter deep sill is appreciably shallower than the average sill depth of 106 meters found in other fjords, yet it is not the shallowest sill in the British Columbia fjords. Therefore, with the possible exception of the sill depth, Indian Arm can be considered physiographically typical of British Columbia fjords.

2. Bottom sediments

The characteristics of the bottom sediments of Indian Arm are shown in figure 14. This analysis is based upon 18 samples collected with a Dietz-Lafond snapper along 6 cross-fjord transects. On each transect samples were collected from both eastern and western flanks and from the center of the fjord. All samples were light grey to grey in colour, indicating non-stagnant conditions (Twenhofel, 1939). With but three exceptions (A west, B east, E east), samples contained very little organic material. The organic material consisted of leaves

from deciduous trees, pine needles, small twigs and wood chips. No appreciable H_2S odour or varves were noted in any of the samples.

The particle composition of the samples collected along the deep center line of Indian Arm show certain consistent trends. The relative proportion of clay particles increases with distance from station A toward station D. The slight departure from this trend shown by the presence of clay in the sample from station B may be attributed to the increased current velocities at this station. This would result from the small cross-sectional area of the channel adjacent to Croaker Island. Conversely, the proportion of silt decreases with distance from station A toward station E, and the proportion of sand also decreases but at a much faster rate. It is believed that progressive sorting has occurred with distance from the head of the fjord. Heavier particles are deposited closer to their original source. This is consistent with that reported for the distribution of sediments in Bute Inlet (Toombs, 1956).

Samples collected at three widely-separated points across the mouth of the fjord (station H) show a similar distinctive pattern of particle size distribution, even though they were collected at different depths. This strongly suggests that the same current pattern and water mass is present at all points across the mouth of the fjord. In addition to being consistent, the distribution pattern here is quite different from that observed within the fjord. The high proportion of sand is particularly noticeable. This difference suggests that most of the water mass in this region has a different immediate

origin than that of the water from within the fjord. When the distribution pattern of the stations just inside the "narrows" (the lateral restriction between stations G and H) is examined, another point is apparent. The distribution of particle sizes found in the center sample of station F is intermediate between that found at the mouth of the fjord at station H and that found within the fjord at station E, indicating that there has been intrusion of the outside waters and progressive sorting of the suspended particles.

Except in two sets of samples, the different depths of the east and west flank samples preclude comparison. However, two sets do show cross-fjord differences, as, for example, in the east and west set collected at station F, which is just inside the narrows north of the sill. The west flank sample contains a relatively higher proportion of clay than the east flank sample, and conversely, the east flank sample contains a relatively higher proportion of sand and silt. From this it is assumed that the waters entering the fjord from Burrard Inlet contain a higher proportion of large-sized particles than corresponding waters within the fjord. This assumption is consistent with the water's history of tidal passage through a region of extreme turbulence at the second narrows of Burrard Inlet. This assumption would indicate that waters entering the fjord tend toward the eastern side of the channel, and that waters moving out of the fjord tend toward the western side of the channel. Such a current distribution would result in the deposition of a higher proportion of larger particles toward

the western side and the deposition of a higher proportion of⁵³
clay-size particles along the eastern side of the channel.
This type of estuarian current pattern results from the effects
of Coriolis force (Pritchard, 1952).

The sediment samples taken across the fjord at station A
show an increase in the proportion of sand from east to west.
This can be attributed to the Mesliloet River, which enters at
a point toward the western side of the fjord.

In summary, the bottom sediments of Indian Arm suggest
that:

- a. progressive sorting occurs along the fjord length in
two directions, both north and south, indicating two
primary sources of suspended material: namely, the
Mesliloet River and intruding outside waters from
Burrard Inlet.
- b. waters entering the fjord tend to flow inward along
the eastern side of the channel, and outward toward
Burrard Inlet along the western side of the channel.
- c. non-stagnant conditions prevail.

B. Distribution of Conservative Oceanographic Properties

1. Introduction

Much of the oceanographic research on British Columbia fjords (Carter, 1933; Giovando, 1959; Hutchinson, 1931; Pickard, 1955; Tabata and Pickard, 1957; Trites, 1955; Tully, 1936, 1949) has dealt with the distribution of conservative oceanographic properties. The bulk of this research has been conducted during the summer months and little has been known of the seasonal distribution of these properties. Tabata and Pickard (1957) worked with data gathered on 11 cruises during a two-year period, with gaps between successive cruises ranging up to five months. During 1931, the Pacific Oceanographic Group (P.O.G.) made limited observations on 9 cruises over a period of one year in Indian Arm. With these exceptions, all previous work has been conducted primarily during the summer months. Since 1951 and prior to this work, only three of the 19 other cruises conducted in British Columbia fjords were made between October and May. The data upon which this work is based were collected on 35 cruises at approximately monthly intervals over a period of three years, and is the most complete set of oceanographic data available on a British Columbia fjord. It provides a unique opportunity to establish in some detail the characteristic annual cycle of fjord oceanographic properties.

2. Salinity

Two features control the salinity distribution in Indian Arm: connection with adjoining straits and fresh water entering the fjord. The former results in the bulk of fjord water

being saline. The latter subsequently reduces the salinity of this saline water. Therefore the fjord is a positive estuary, since it contains sea water measurably diluted with fresh water (Pritchard, 1952).

The fjord's water mass tends to exist as a two-layer system because of the quantity of fresh water which flows into the fjord. A thin layer of brackish water at the surface overlays a more saline water mass. The principal variations in time and space which occur within the fjord are differences in thickness of this brackish water layer and differences in the surface salinity gradient from the head to the mouth of the fjord.

The annual changes in the salinity distribution of Indian Arm are clearly discernable and consistent. The same basic pattern was observed during each of the three years of this investigation. Figure 15 presents this repeating pattern for nine depths of a single station representative of the fjord water mass. The annual changes can be divided into four periods, each of which approximately corresponds to the variations in the seasonal input of fresh water. These periods are as follows: the late winter minimum from January to March, the spring maximum from April to June, the mid-summer minimum from July to September, and the early winter maximum from October to December. For convenience, the longitudinal and vertical salinity gradients will be discussed in detail for the period of highest runoff (the spring maximum), followed by a discussion of the seasonal deviations from this basic pattern. Finally, long-term deviations from the basic annual cycle will be discussed.

The time of highest runoff of the spring maximum varied from year to year. These maxima occurred during May in two years and during June in one year. It should be noted that two factors suggest that these means may not include the lowest values present in the fjord during the year. It is known that the fresh water discharge may fluctuate over one hundred per cent within a few days (Anon. 1956a, 1957a, 1958a, 1959a). These fluctuations, combined with the fixed monthly scheduling of the oceanographic cruises, make it impossible to be certain that the observations were always made precisely at the time of maximum fresh water discharge. Although the mixing and seaward movement of the fresh water tend to smooth these short term fluctuations, it is likely that surface salinity distribution patterns fresher than the mean presented in figure 9 occurred during the spring maximum.

The seaward salinity gradient previously reported for British Columbia fjords (Pickard, 1955) is present in Indian Arm. During the three years of this investigation, spring maximum salinities were $0.95^{\circ}/\text{oo}$ to $4.38^{\circ}/\text{oo}$ at the head, increasing seaward to values between $13.16^{\circ}/\text{oo}$ and $19.40^{\circ}/\text{oo}$ at the mouth. During this period, the mean surface salinity change from head to mouth is nearly linear, as shown in figure 10. The greatest departure from a linear relationship occurs in the region from station F to station H, and apparently is associated with passage of the surface waters through the "narrows". Here the mean rate of change of salinity is $1.10^{\circ}/\text{oo}$ per kilometer, a rate of change over twice value $0.47^{\circ}/\text{oo}$, observed within the fjord. If the base salinity were

considered to be the highest salinity observed in the fjord, these salinities at the mouth would be approximately 60 per cent of base salinity.

At the head, the vertical salinity distribution exhibits a strongly stratified system composed of an essentially homogeneous layer 1.5 to 2.5 meters thick, subtended by a thick bottom layer of much more saline water. These two layers are separated by a transition zone 2.5 to 5 meters thick, which has a high vertical salinity gradient. Thus at depths less than 10 meters, salinities reach over 90 per cent of base salinity. With distance seaward, as shown in figure 17, the deep water retains essentially the same salinity character.

Of the four seasonal salinity distributions, the pattern present during the spring is the most consistent from year to year. This is probably in part due to the nature of the fresh water discharge. During this period, the chief source of fresh water entering via the Mesliloet River and peripheral streams is melt water. Examination of river discharge data shows that discharge during the spring melt period is much more consistent from year to year than discharge during the fall, when there is direct runoff.

The most clearly defined and most consistent pattern is present during the maximum runoff periods in late spring and early summer. The distribution during this period is shown in figure 19. At this time there is a consistent increase in surface salinity, with distance toward the mouth of the inlet. At about 2 meters depth, the salinity varies little along the length of the inlet. From about 2 to 18 meters, the trend

found at the surface is reversed. The salinities decrease with distance toward the fjord mouth. From approximately 20 meters to the bottom, the salinities are essentially uniform, except that the salinities in the southern portion of the deep basin are consistently higher than the salinities at corresponding depths in the northern portion of the deep basin. This tendency is strongest during the spring maximum. During other seasons, the mean salinity gradient is small, and the deep water of 75 to 200 meters can be considered longitudinally homogeneous.

The mean salinity structure at station H is noteworthy. This station, located at the sill south of the "narrows" has a much lower mean salinity between 2 and 18 meters than at corresponding depths north of the "narrows". This pattern can be seen clearly when the mean salinity profile for the spring maximum is compared with the profile for the following summer minimum (Fig. 19). This increase in salinity of surface water, combined with a decrease in salinity at intermediate depths, suggests that intense mixing has occurred when the water moves through the "narrows". The higher salinities apparent in the bottom waters suggest that mixing took place down to about 18 meters.

Figure 19 shows the mean salinity pattern during the mid-summer minimum, and presents a distinctly different pattern than that present during the spring maximum. During the summer minimum, there is again an increase in surface salinity with distance toward the fjord mouth. The mean salinity increase is from 13⁰/00 to 22⁰/00. The seaward increase in salinity is

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apparent down to a depth of 50 meters, with constant salinities extending from approximately 75 meters to the bottom. This is in contrast to the salinity pattern present during the spring maximum, when constant salinities exist much nearer to the surface at approximately 20 meters.

The range of salinities during the early winter maximum (Fig. 19) is greater than during the spring maximum. At this time there is again an increase in surface salinity toward the mouth, which increases the main salinity from $8.2^{\circ}/\text{oo}$ to $19.8^{\circ}/\text{oo}$. However, the pattern differs in that this longitudinal gradient is present at lower depths than it was in the spring. The water approaches longitudinal homogeneity at about 5 meters rather than at 2 meters. Below 5 meters, the vertical salinity gradient is only about $1/3$ that present in the spring. The rate of change of salinity with depth is $0.018^{\circ}/\text{oo}$ per meter compared with $0.067^{\circ}/\text{oo}$ per meter in spring. The same apparent effect of mixing at the "narrows" can be seen in the salinity profile above the sill (Fig. 19). Here the effects of mixing extend to a depth of approximately 12 meters, compared with 18 meters during the spring. Water from 0 to 3 meters is higher in salinity, and water from 4 to 12 meters is lower in salinity than corresponding waters north of the sill.

The pattern present during the late winter minimum, as shown in figure 22, is not nearly as consistent as the other three periods. In 1956 and 1957 there was a decrease in surface salinity seaward; in 1958, a decrease in salinity to station C and then an increase in salinity to station H. The

resulting mean pattern therefore is not considered as representative. This confused picture probably results from the high discharge of fresh water from the Buntzen power plant during low runoff at this time. This confused pattern does not extend to the deeper water masses; here, both longitudinally and vertically the salinity gradients are similar to those found during the other three periods.

The annual salinity distribution discerned in three years is shown in figure 15. The depths selected are considered representative. Several characteristics are significant. The surface salinity follows very closely the seasonal pattern of runoff, maxima occurring both in May and October, and minima in August and February. However, the salinities at 2.5 meters are often significantly out of phase with the surface salinities. This is particularly evident during the last half of the year. Such a characteristic inverse relationship appears to be associated with the mechanism of mixing and tidal feedback at the "narrows". From 5 to 50 meters the annual cycle is characterized by a single maximum and a single minimum which occur progressively later with increasing depth.

For convenience, the characteristic annual salinity cycle is considered in three zones: 0 to 5 meters, 10 to 50 meters, and 100 to 200 meters.

The surface and inversely related sub-surface pattern is controlled by two features: the seasonal discharge of fresh water into the fjord, and the action of the "narrows" upon the seaward transport of the surface brackish layer. The

restriction in cross-channel width caused by the "narrows" forces the velocities of tidal currents and the seaward-moving surface layer to increase. If the relative velocity exceeds a critical value, the interface between the surface layer and the next lower layer breaks down and a new, thicker surface layer is formed. This layer continues to move seaward, but at a much lower velocity, which is below the critical value for surface layer breakdown. During periods of high runoff, the resulting mixed layer still has a high enough salinity to form a large density gradient between it and the underlying bottom waters. This large density gradient combined with the seaward-moving surface currents result in the tidal water entering the fjord having a relatively high salinity.

During periods of low runoff, the resulting water mixture has an appreciably higher salinity than during high runoff. The density gradient between the surface layer and deep water is quite low. This, combined with the low velocity of the surface water, results in the tidal inflow extending from the top to the bottom of the water column. Such an inflow, subsequently mixed during passage through the "narrows", has a lower salinity than the inflow during periods of high runoff.

The above concept is supported by the inverse relationship observed in the main fjord basin between the salinities at the surface and sub-surface layers. Examination of longitudinal salinity profiles of individual cruises suggests that the thickness of the newly-formed surface layer is approximately 18 meters during times of high runoff and 12 meters during

maximum winter runoff. The results of upward mixing can be seen as a marked increase in surface salinity seaward of the "narrows", as shown in figure 18.

The zone between 10 and 50 meters has salinity minima at progressively greater depths as the year advances. The initiation of salinity minima appears to be associated with minima in the sub-surface waters during June and July. Its progression downward is probably related to downward mixing of fresher water from this sub-surface layer.

In the deepest zone, 100 to 200 meters, the distinctive feature is the rather sharp increase in salinity during February and March of some years. Yet even with these annual fluctuations in deep basin waters, the salinity remains quite constant compared with the surface waters. The average salinity remains close to $27.5^{\circ}/\text{oo}$.

When the deep-water salinities of all British Columbia fjords are considered, a natural division is evident. Fjords of the northern group have the waters of Hecate Strait and Queen Charlotte Sound as their salt water source; those of the southern group are connected to the Strait of Georgia. As a consequence of the diverse oceanographic characteristics of these bodies of water, the base salinities of the northern and southern groups are significantly different. The northern group average $32.5^{\circ}/\text{oo}$ and the southern group $30.0^{\circ}/\text{oo}$.

Indian Arm is connected to the Strait of Georgia, and consequently is a member of the southern group of British Columbia fjords; yet its deep-water salinities vary significantly from the other fjords of the group. The average deep

basin salinity of $27.5^{\circ}/\text{oo}$ varies as much from the average of the southern group as the southern group does from the northern group. This is probably associated with the shallowness of the sill.

Within the southern group there is a further tendency toward separation on the basis of differences in quantity of fresh water discharge. As discussed later in the section on fresh water inflow, the fjords can be divided into high and low runoff subgroups. During maximum runoff, the high runoff fjords have surface salinities approaching $0^{\circ}/\text{oo}$ at the head, and at the mouth are between seven and nine-tenths of the base salinity for the fjord. In the low runoff fjords, the salinity gradient from head to mouth is much less steep. While similar salinities are obtained at the mouth, the salinities at the head may range from 7 to $28^{\circ}/\text{oo}$. Correlated with this horizontal salinity gradient is a vertical gradient which is largest during periods of maximum runoff. At these times, salinities may be as much as 90 per cent of the base salinity at depths of 10 meters. During low runoff periods and in the low runoff fjords, the salinity gradients are much smaller. The characteristics of Indian Arm agree well with the reported characteristics of the low runoff group, and it is therefore considered to be a typical member of this group.

However, in Indian Arm, in spite of the annual variations in salinity, the total amount of salt contained in the fjord waters is nearly constant. The mean salinity of the water column can be considered an approximate measure of the salt in

the fjord. The annual range in this cycle is $0.7^{\circ}/\text{oo}$, and shows a single maximum and minimum. The maximum occurs in March and presumably reflects the annual influx of more saline waters from outside the fjord. The minimum occurs in July and probably reflects the downward mixing of fresh water. The mixing process tends to decrease the salinity of the deep water and increase the salinity of the surface layers.

The most outstanding point in the long-term salinity picture is the existence of a sharp increase in deep water salinity between January and March of 1957. Unfortunately, the exact time cannot be established because of a gap in the cruise schedule during this period. At this time, the salinity at all depths from 50 meters to the bottom of the fjord increased by nearly $0.5^{\circ}/\text{oo}$, a change in salinity commonly associated with shallower waters. The only source of such highly saline water is the Strait of Georgia. It is believed that a massive intrusion of highly saline water occurred at this time.

In examining this picture several other points become evident. During November 1957, a similar increase in salinity is apparent at all depths from 50 meters to the bottom. In September of 1959, a salinity inversion was present with the salinity at 150 meters being higher than that at 200 meters, suggesting a short-term influx of outside waters. During March of 1958 and 1959, the waters from 150 to 200 meters were essentially homogeneous, again suggesting an influx during these periods.

The fresh water discharge from the Buntzen power plant has an important bearing on the distribution of salinity in Indian Arm. In order to establish the extent of this influence, a grid of hydrographic stations was occupied near the power plant during October 1957. At this time, the fresh water discharge from the Mesliloet River and from peripheral streams was low, and the discharge from the power plant was high. Figure 20 shows the effects of the fresh water influx at 0, 2.5, 5, and 10 meters.

Adjacent to the plant outfall, the surface inflow results in a reduction in salinity by about $2.5^{\circ}/\text{oo}$ relative to upfjord values. Over a wider area, the salinities are reduced by about $1.5^{\circ}/\text{oo}$. At 2.5 meters, the affected area is smaller, with an average salinity reduction of about $1.5^{\circ}/\text{oo}$. The intrusion of higher-salinity water from upfjord is apparent as an indentation in the isohalines opposite the power plant on the west side of the fjord. At 5 meters, the fresh water influx lowers the salinity by about $0.2^{\circ}/\text{oo}$. At 10 meters, a slight indication is apparent in a salinity reduction of about $0.05^{\circ}/\text{oo}$ in a small water mass immediately opposite the power station.

These data indicate that the effect of fresh water discharge from the power plant significantly reduces the salinity in the immediate area of the plant and primarily influences the surface water along the eastern half of the fjord. The surface water outflow apparently maintains its identity along the western side of the inlet. During periods of low runoff and high discharge from the power plant, the adjacent water mass may

have its salinity reduced 2 to 3‰. The influence becomes weaker with increasing depth, and has almost disappeared at 10 meters depth.

During periods of low discharge, the plant is operated intermittently, at times being closed completely. Such an operation pattern would result in the formation of clouds of fresh water which might retain their identity as they are moved about by tidal currents, wind, and the net seaward motion of the surface brackish water layer. Such clouds could introduce errors in the interpretation of surface salinity measurements; however, the effect would be restricted to the surface water, and would not appreciably influence the deeper water mass.

The surface brackish layer can be considered as consisting of two zones: an upper zone in which the salinity remains substantially constant from the surface to 2 to 5 meters, and a transition zone, the halocline, from 4 to 10 meters, where the salinity increases rapidly with depth. Observations relating to the increase in volume and velocity with entrainment of underlying waters suggest that the brackish layer decreases in thickness with distance from the head of the fjord, and that a point may be reached where the sharp transition zone becomes blurred (Tully, 1949). Such a breakdown of the surface layer appears to have been observed in the course of the oceanographic investigations of Alberni Inlet, a British Columbia fjord (Tully, 1949), but is not strongly supported by the data from Indian Arm. However, it should be noted that the salinity samples are collected using 50 cm. long water bottles, which would

prevent noting any decrease less than this distance. Such a decrease should be strongest during periods of high runoff, as represented in figure 18. In this figure there is a slight suggestion in the 21⁰/oo isopleth of a decrease in thickness from station C to station F. The noticeable increase in thickness seaward of this station is attributed to tidal mixing in the region of the "narrows" rather than the observed breakdown in the surface layer interface.

The breakdown in the surface layer is assumed to be due to an increase in velocity, attributable to an increase in volume resulting from entrainment. It can be speculated that an analogous situation exists in Indian Arm, but is attributable to a different cause. A reduction in the salinities of the upper 10 meters can often be seen at station B. This reduction is associated with periods of high runoff and correspondingly high velocity of the surface water. Because of the restriction in channel width in the vicinity of Croaker Island, the critical velocity for surface layer breakdown is apparently reached north of station B, with the associated breakdown in surface layer caused by massive turbulence and the subsequent downward mixing of fresh water.

3. Fresh water

Indian Arm has three main sources of fresh water; the Mesliloet River, which discharges at the head of the fjord, numerous peripheral streams which cascade down the fjord walls, and discharge from the Buntzen power plant. The character and proportion of these primary sources combine to establish the annual cyclic pattern of fresh water input into the fjord.

Most British Columbia fjords also have one or several rivers which discharge near the head. The two main sources of river water are precipitation, runoff and snow, and glacial melt waters. Examination of river discharge cycles shows that although wide variations do occur from year to year, a consistent pattern is usually apparent for a fjord. In fjords in which the main river is large and primarily taps glaciated regions, the river runoff has a maximum in June and July and a minimum in December and January. Fjords which receive rivers draining regions with little or no stored precipitation have maximum runoff from August through December, followed by a minimum runoff from April to July. Many gradations between these two types are noted (Trites, 1955). However, for convenience previous considerations of runoff pattern and amounts have divided fjord rivers into groups: those with relatively high runoff, and those with relatively low runoff (Pickard, 1953).

The Mesliloet River, draining into Indian Arm, was unmetered during the period of this study, but data are available for the period from 1914 to 1922. On the basis of these data, the relationship between the Mesliloet River and currently metered rivers of similar runoff pattern has been calculated. The present runoff of these rivers was utilized to provide an estimate of the Mesliloet River runoff during the time of this study. An independent check of this method showed that the predicted monthly values varied on an average by 15 per cent from the actual values. The greatest difference occurred in December, during the latter part of the period of winter rains.

In most cases the rivers contribute the major portion of fresh water input, but in some fjords peripheral streams play an important role. In the fjords that have been considered from this aspect, peripheral streams contributed from 20 to 65 per cent of the total fresh water which enters the fjord (Trites, 1955). Except in local regions--that is, cells of high precipitation--the amount of fresh water contributed by direct precipitation is small relative to the amounts contributed by rivers and peripheral drainage.

The Mesliloet River drains an area of 121 square kilometers; the peripheral streams to the west and east of the fjord drain a total area of 44 square kilometers; and the fjord itself has a surface area of 13 square kilometers. The mountainous terrain comprising the drainage area of the Mesliloet River is similar in topography to a major part of the drainage area to the east and west of Indian Arm. A comparison of the annual precipitation data from the Britannia Beach meteorological station with the data from the Ioco meteorological station shows good agreement (Anon., 1956b, 1957b, 1958b, 1959b). The annual precipitation cycle is similar, and the total precipitation for the two stations differed by 10 per cent during the time of this study. It is assumed, therefore, that precipitation is the same over the drainage area of the Mesliloet River and the drainage area of the peripheral streams. On this assumption, peripheral stream discharge has been calculated from Mesliloet River runoff on a proportional drainage area basis.

Ignoring the stored precipitation, and assuming that 50 per cent of the total precipitation drains into the fjord (Tabata and Pickard, 1957), it can be calculated on the basis of rainfall area that 66 per cent of the precipitation which reaches fjord waters arrives as river runoff, 25 per cent arrives as peripheral stream runoff, and 9 per cent as direct precipitation. Thus, over an annual cycle, both peripheral streams and direct precipitation do contribute a significant proportion of the fjord's fresh water.

If little or no storage of precipitation occurred in the drainage area of the fjord, the Mesliloet River outflow would be expected to follow closely the annual precipitation cycle. As shown in figure 21, this is not so. From December through March, river discharge lags behind precipitation, indicating that an appreciable portion of the precipitation is being stored in the drainage area. During spring and early summer, although the rate of precipitation decreases, the rate of river discharge increases. This is effected by the melting and release of the snow and ice which accumulated during the late winter months. During May and June, river discharge is at its maximum and precipitation is near the year's minimum. With the onset of winter rains, the precipitation and discharge curves fall into phase, indicating that the primary source of discharge during this period is direct runoff. Thus the bimodal cycle of river discharge results from the combination of two basic cycles: a spring peak primarily caused by melt waters and a fall peak attributable to direct runoff.

In the 26 fjords which have been analyzed to date (Trites, 1955), the ratio between mean yearly fresh water discharge and unit area of fjord surface varies by a factor of about 10 from 6.8 to 62.3 cubic meters per second per square kilometer. The mean is 19.1 cubic meters per second per square kilometer. Indian Arm has a mean yearly ratio of 9.5, placing it within the range previously reported but toward the lower limit of the range. This is primarily attributable to the local climate of this fjord's drainage area. This region of the British Columbia Coast Range receives an average annual precipitation of 250 centimeters, while many of the drainage areas which are tapped by other fjord rivers and streams have figures in excess of the 380 centimeters per year.

The fresh water sources of Indian Arm differ from those of other British Columbia fjords in one respect. Situated on the east side of the fjord opposite station D is the Buntzen power plant. The water which is used by this plant is primarily collected from outside the drainage area of Indian Arm. In most respects the fresh water discharge from this plant can be treated as a large peripheral stream. There is one distinct advantage: data are available for the exact quantity of fresh water flowing out from this source. On an average, during the year this plant contributes approximately 50 per cent--well within the range of 20 to 65 per cent reported for other British Columbia fjords.

Under unusual meteorological conditions, low temperatures can freeze almost all the peripheral streams and drastically

reduce the outflow of the Mesliloet River. At these times the primary sources of fjord fresh water are direct precipitation and Buntzen power plant discharge. At such times a high proportion of the fresh water entering the fjord is attributable to the discharge of the generating plant.

The location of the Buntzen power plant and the distribution of the major peripheral streams suggest that nearly 50 per cent of the total fresh water which enters the fjord during the heavy winter rains enters at points well below the head. It should be noted that this is in contrast to most of the theoretical considerations of fjord circulation in which it is assumed that all fresh water entering the fjords is contributed by the river at the head (Pritchard, 1952).

Figure 22 presents the annual cycle of total fresh water contributions to the fjord, and indicates the relative proportions attributable to each of the three primary sources.

4. Temperature

The primary processes which effect the changes in the heat content of the British Columbia fjord water masses are radiation-back radiation, condensation-evaporation, convection of sensible heat to and from the atmosphere, and advection. These combine to produce an annual cycle of temperatures with a single maximum and minimum. The cycle of surface temperature is in phase with the air temperature cycle, although of slightly lower amplitude. The annual temperature cycle in Indian Arm agrees in this respect. The three-year-mean cycle is presented in figure 23, in which the mean air temperature cycle is plotted on the same scale as the surface water temperature.

The correlation between the air and surface water temperatures is quite evident, both minima occurring in January and both maxima in August. These are similar to those of the British Columbia coastal waters where a minimum of $7.2^{\circ}\text{C} \pm 0.5$ is reached in January or February, and a maximum of 10° to 18°C . in August (Pickard and McLeod, 1953). Throughout five months of the year, the air temperature exceeds the surface water temperature. During the remaining seven months (fall and winter) the air temperature is less than the surface water temperature. At times when air temperature is higher than the temperature of the surface water, the difference between the two is small. When the maximum surface water temperature is reached, the air temperature exceeds it by less than one degree. In months when the air temperature is less than the surface water temperature, there is greater divergence. When the minimum is reached, the temperatures differ by more than three degrees. This pattern of convergence during warming periods and divergence during cooling periods is probably associated with the stability of the water column, which in turn is associated with density. In general, the effects of radiation and sensible heat transfer are only important at the surface; sub-surface heat transfer is accomplished chiefly by turbulent action. Quite apart from the effects of salinity variations, the warming of the surface results in lower densities and a more stable water column. Under these conditions, transfer of sensible heat from the atmosphere tends to produce a surface water temperature comparable within 0.5 degrees of

the air temperature. Under cooling conditions, the surface waters become denser, which decreases the stability and brings about a movement of water away from the surface.

Seasonal fluctuations in temperature are evident down to a depth of 75 meters, and are less noticeable, though still apparent, down to 100 meters. Figure 23 shows the typical recurring annual cycle of the seasonal fluctuations. Below 100 meters the waters are relatively uninfluenced by seasonal fluctuations. Their character is determined by an occasional large intrusion of outside water. This occurs at intervals of some years. Between such intrusions, a slow, steady increase in temperature is evident. A layer with characteristics between these two regions occurs at approximately 100 meters.

The curves for the mean temperature at 10, 50, 75, and 100 meters show a gradual increase in the time of maximum relative to the surface waters. In addition, there is a suggestion of an increasing difference in time with increasing depth between the temperature minimum and the temperature maximum. The time of temperature minimum varies from February at the surface to October at 100 meters. Thus the difference in time between the minimum and maximum at the surface is six months, and at 100 meters it is nine months. This error seems to indicate that the rate of downward cooling is 50 per cent faster than the rate of downward warming.

This difference in the rates of heat transfer appears to be associated with changes in the intensity of vertical turbulent transport at different seasons. The stability of the

water column is lowest during spring and early summer and is greatest during late summer and fall. If one assumes that the rate of turbulent transport is inversely proportional to the stability of the water column, the downward transfer of the spring minimum would be speeded up and the downward transfer of the summer maximum would be retarded. Although the assumption is often made that the source of energy for such turbulence is derived from weak horizontal currents and tidal movements, it does not appear to be so in this instance. Since the horizontal temperature gradients are small relative to the vertical gradients, horizontal currents probably do not greatly influence the vertical temperature gradients.

The three-year cycle of temperatures is shown in figure 24. In addition to distinct annual fluctuations, several other features become apparent which were obscured by the mean values shown in figure 23. Some of these features are:

- a. the varying depths to which seasonal fluctuations occur during different years.
- b. the long-term increase in deep water temperatures.
- c. the consistency of time of temperature inversion from year to year.

When the temperature fluctuations at representative depths are examined, it can be seen that a seasonal pattern extends to different depths in different years. During the spring of 1957 and 1959, sharp decreases occurred in the temperatures of the bottom waters. Two mechanisms could account for these decreases: the cooling of surface waters, followed by an overturn of

the water column, or an intrusion of colder water into the fjord. As a consequence of the high vertical salinity gradient, the stability in Indian Arm is such that overturn of the water column does not occur. Inasmuch as an increase in salinity occurred concomitantly with the temperature decrease and since there is no source of higher salinity water in the water column, an intrusion of outside water must have occurred at this time. The 1959 intrusion did not affect the established trend of an increase in bottom temperature, either because it did not reach the bottom waters or because the temperatures of the intruding waters were within 0.1° C. of the temperature of the bottom waters.

During 1958, such an intrusion either did not occur, or it was small in extent and effect, as evidenced by a small decrease in temperature at 75 and 100 meters during February. Since a massive intrusion of colder water during early spring obscures the downward transfer of the surface temperature minimum, it is difficult to ascertain the maximum depth which this minimum attains. If one assumes that little or no cold water was introduced horizontally during the spring of 1958, the observed changes in temperature structure would be due primarily to vertical mixing. During this period, a clearly defined temperature minimum occurred at the surface in February, at 20 meters in March, and at 50 meters in April. At this time the temperature maximum was at approximately 75 meters, with a decrease in temperature both toward the surface and toward the bottom of the water column. This suggests that under seasonal

conditions (with little or no inflow of outside water) the surface minimum does not extend deeper than about 75 meters.

The dates of the spring temperature inversion are nearly the same in each of the three years. By linear interpolation, the date on which the spring inversion occurred was April 10 in 1957, April 3 in 1958, and April 11 in 1959. The fall inversion is just as consistent, with the interpolated dates being October 9 in 1957, October 4 in 1958, and October 17 in 1959. Both the spring and fall inversions for the three years occur within a range of eight days. Such regularity suggests a remarkably consistent spring and fall surface radiation pattern. This is supported by the air temperatures; the mean monthly air temperatures (primarily influenced by the surface radiation pattern), from 1956 to 1960 were 9.6° for April and 9.7° C. for October.

During the time of spring temperature inversion, the water column is virtually isothermal during a "normal" year. In 1957 and 1959 the mean temperature gradient from the surface to 200 meters was 0.0035° C. and 0.0032° C. per meter respectively. During 1958 a similar low vertical gradient of 0.0030° C. per meter was present in the 0 to 75 meter portion of the water column, but a much higher gradient of 0.01° C. existed between 75 and 200 meters.

The fall temperature inversion is not accompanied by as strongly developed an isothermal pattern as the spring inversion. In all three years, isothermal conditions extended to a depth of about 30 meters, with the underlying portion of the water column presenting a much higher temperature gradient.

The time of the summer maximum of the surface water is close to the time of maximum air temperature. The maximum surface temperature for a particular year occurred during the three years varied by about six degrees. In 1957 the maximum surface temperature was between 15 and 16°C, in the following year the maximum was 21°C, and in 1959 the maximum temperature was about 19°C. This follows the trend of the air temperatures. The year 1958, in which the surface temperatures were highest, was also the year in which the highest mean air temperatures (22°) were recorded. In the preceeding year, when the surface temperatures were appreciably lower, the monthly mean was 17°C during the same month. In 1959 when the surface temperatures were intermediate between the 1957 and 1958 temperatures, the air temperature was 18°C.

A similar, though less well defined relationship between the minimal temperatures of the air and surface water is evident. The temperature minimum of the surface water occurred during January in three years. The highest minimum was recorded in 1958. This month also showed the highest air temperature minimum. The lowest surface minimum was observed during 1959. During January 1957 the mean air temperature was 1.1°C and the surface temperature was 5.60°C. Calm conditions during the cruise of January 1959 are probably responsible for the low minimum recorded at this time.

Surface ice was observed twice - once in January 1957 and once in January 1959. On both occasions the ice was thin and did not persist; it formed at the northern end of the

fjord during the night and broke up during the following day under the influence of light winds. Residents at Buntzen Bay report that during particularly severe winters the fjord may be frozen over for several days from the region near station D north to the head.

The monthly changes in vertical temperature structure are clearly shown when the temperatures are plotted as isotherms with depth, as indicated in figures 25, 26 and 27. The disruptive effects of the massive intrusion occurring in January through March 1957 are clearly evident in figure 25. In 1959 a similar effect is apparent. Some concept of the relative extent of these two intrusions can be gained by comparing the vertical extent of the isotherms of these periods. In 1957 the intrusions appears to have virtually flushed the fjord, whereas in 1959 the more restricted vertical development suggests a more limited flushing. The effects of vertical mixing can be seen in the downward sloping isotherms plotted for spring and early summer. A high temperature core is present between 10 and 35 meters during August of each year. When the summer maximum (Fig. 24) is examined in detail, the top of this core is delimited by a temperature inversion at 10 to 20 meters. This inversion was of short duration in 1957 and 1959, but lasted for two months in 1958. An examination of the longitudinal distribution of temperatures prior to and during the summer, suggests there is an influx of outside water at 20 to 30 meters depth. During July 1958 the vertical temperature gradient at all stations north of station F shows a decrease in temperature with depth. At station F,

which is just north of the "narrows", there is a temperature inversion of 0.33°C at 30 meters. During August the temperature inversion is evident from station F north to station B near the head. The greatest inversion occurs at station D where it is 1.5°C . At station C the inversion is 0.99°C and at station B it is 0.92°C . No trace of the inversion can be seen at station A. The temperature inversion commenced at 5 meters at station F, at 7.5 meters at the head, and had its center lying between 20 and 30 meters.

During 1959, when the inversion was not as persistent nor as strongly developed, it could still be traced from the fjord mouth north to station B. The depth at which the inversion began was 10 meters, but the maximum development occurred at 20 to 30 meters. No trace of this inversion could be seen in the preceeding July or the following September. It was either residual when observed in 1957 or very weakly developed and it was barely detectable in the central basin at stations E, D, and C.

An analysis of the data shows that this inversion pattern, beginning at the mouth of the fjord and developing northward at intermediate depths, is the result of a very slow inflow of outside water at 20 to 30 meters. This slow inflow is a result of the pattern of tidal intrusion occurring during this season.

When the deep water is considered, a gradual increase in temperature is found from April 1957 to October 1959. Throughout this period of 31 months, the temperature increases in every succeeding month except November 1958. The total increase

is nearly 1.5°C . Such a slow steady increase is attributable to downward mixing of warmer water which occurs in the absence of large intrusions of colder outside water. Under these conditions the deep basin water does not appear to be affected by conditions outside the basin. This provides an opportunity to determine accurately the rate of two processes: the eddy diffusion, and the rate of oxidative and respiratory processes.

The pattern of horizontal temperature distributions is not as clear as that for vertical distribution. The gradient in the fjord basin averaged less than $.003^{\circ}\text{C}$ per kilometer at 200 meters, $.01^{\circ}\text{C}$ per kilometer at 150 meters, and $.04^{\circ}\text{C}$ per kilometer at 100 meters. The maximum gradient observed below 100 meters never exceeded $.06^{\circ}\text{C}$ per kilometer. In December 1957 a gradient of $.12^{\circ}\text{C}$ per kilometer was observed, apparently associated with a small intrusion of warmer water at 100 meters.

The highest gradients were observed in the surface layer during the mid-summer maximum in association with periods of low runoff. At this time they were as high as $.21^{\circ}\text{C}$ per kilometer. During the remainder of the year the horizontal gradients rarely exceeded $.06^{\circ}\text{C}$ per kilometer. At intermediate depths they were proportionally lower.

The direction of the gradients shows some consistency. Table VI depicts the gradients at representative depths and indicates that the yearly trends fall into four groups: from January through March, the general tendency is for a decrease in temperature from fjord head to mouth; from July through September, for an increase in temperature from head

to mouth; and in intervening periods of April-June and October-December there are less consistent patterns that vary from year to year.

A strongly developed minimum in the vertical temperature distribution as reported in other fjords, (Pickard, 1953) was not observed in Indian Arm. However, during the spring months of 1958 and 1959 well defined maxima were observed at 50 to 75 meters. The occurrence of these maxima might be attributed to the same basic mechanism which produces the minima noted in other fjords. In order to explain these features under ideal-conditions the following model is presented.

At a time during late spring, when the water column is essentially isothermal and horizontal currents are negligible, the input of heat to the water column exceeds the loss, and gradual warming of the surface layer takes place. With continued warming and some vertical mixing, a characteristic summer pattern with high surface temperatures forms. These temperatures decrease with depth until the lowest temperatures are reached in the bottom waters. In early fall the loss of heat from the water column exceeds the heat input, resulting in a decrease in temperature of the surface layer. If the cooling process during the winter is of the same intensity and duration as the warming process during the summer, further cooling with associated convective and turbulent transfer of heat will again produce isothermal conditions from surface to bottom. Continued cooling produces typical winter temperature gradients, in which there is a minimum at the surface and an increase in temperature toward a maximum in deep water.

Since both cooling and warming processes of the water column originate at the surface, a point will be reached during the process of cooling when there is a temperature maximum at an intermediate depth; and conversely, during the warming process a temperature minimum will be reached at an intermediate depth. These maxima and minima will persist if they are strongly developed as a result of a particularly cold winter or warm summer; or if the warming process is longer and more intense relative to the bottom waters than the cooling process and vice versa; or if the advective process is weak during one season.

In Indian Arm two factors combine to prevent a well-defined persistent temperature minimum. The warming trend is much stronger relative to the bottom waters than is the cooling trend. This probably results from the warming period occurring chiefly during low runoff and high stability. Secondly, intrusions of outside water occur in February, which prevents the maximum development of the cooling trend.

The formation of the most clearly defined maximum is depicted in figure 28. The vertical temperature gradients are plotted for station C at the northern end of the fjord. This station is less affected by horizontal currents and tidal intrusions than more southerly stations. Isothermal conditions were present in April 1957, probably resulting from the thorough flushing of the fjord which occurred during the preceding months. In April, the first slight warming of the surface waters could be detected. Throughout succeeding months the warming trend intensified until the maximum was

reached in October. In November, the surface layers had begun to cool and the cooling trend continued until March 1958. In March, the surface layer was again beginning to warm, and by April isothermal conditions were approached. However, the maximum which was attained the preceding fall persisted throughout the cooling period, and could still be seen as late as May 1958. This same pattern was present during 1959, although it was partially obscured by an intrusion of water during February.

When this temperature maximum is considered as a longitudinal feature, its dissipation in the southern end of the basin can be seen clearly. During the summer of 1957, the warming of the water column was apparent at all stations within the fjord. During the fall, the temperature maximum persisted as a high temperature layer extending the length of the fjord. By November and December, the seaward end of this layer had been eroded, probably by horizontal currents resulting from tidal intrusions. The maximum thus developed the appearance of a high temperature tongue extending seaward from the head of the fjord at intermediate depths. With succeeding months, the continued erosion of the seaward end of the tongue progressively reduced its length. In January, it extended only to station D; in February, only to station C; and by April the tongue remained strongly developed only at station A near the head of the fjord. A strongly developed maximum can be considered as a conservative feature of the intermediate depths, and the rate of its erosion can be used to provide a measure of the water replacement pattern near the fjord mouth. However, since it is

a seasonal pattern, such a measure would only hold true for a few months during the year.

Two local conditions may temporarily alter the temperature distribution in the fjord: conditions which produce a great deal of mixing, and the influx of large volumes of low temperature water. Meteorological conditions can produce strong down-fjord winds which mix the upper layers of water. This condition reduces the temperature gradient in the upper layers, and may result in an isothermal condition. Cooling of the surface water near the entrance of rivers and streams may occur. This is particularly noticeable off the mouth of the Mesliloet River during the runoff of melt water. Since the runoff occurs during the late spring and early summer, this source of cold water tends to counteract the warming trend normally present at this time of year.

5. Density and stability

The density anomaly (σ_t) and its effect upon stability has been superficially examined by previous workers in British Columbia fjord waters. It has been demonstrated that variations in σ_t generally follow the variations in salinity closely enough to validate the use of salinity distribution patterns as an approximation of densities (Tabata and Pickard, 1957). Values of σ_t previously reported for these waters range from 0 at the head of fjords with high runoff to 26.0 in deep basin waters.

The annual cycle in Indian Arm is presented in figure 29. A comparison between this annual cycle and the salinity cycle

depicted in figure 15 indicates a gross similarity. The surface waters show a bimodal curve generally reaching maxima in May and October. At 5 to 50 meters there is a unimodal curve with the maximum occurring progressively later with increasing depth. At 5 meters the maximum usually occurs in June at 50 meters in September. Below 50 meters no precise annual cycle exists. Figure 29 shows the sharp increase in densities which occurred during January through March 1957, followed by a slow steady decrease in density until 1959. This increase in density resulted primarily from a corresponding increase in salinity which in turn is attributable to an intrusion of saline water at this time. The lowest value (1.98) was encountered in May 1958 in the surface waters of station A and the highest values (21.10 to 21.72) were found in the deep waters of the fjord basin.

Of direct importance to dynamic oceanography is the influence of density distribution upon stability of the fjord water mass. The stability may be considered directly proportional to the vertical density gradient (Sverdrup, 1942) and can be expressed as:

$$E = 10^{-5} \frac{d\sigma_t}{dz}$$

Stability values have been computed for the three years, and are presented as seasonal profiles in figures and 30, 31, and 32. As with temperature and salinity, the distributions for 1957 and 1959 follow the same basic pattern; 1958 was an atypical year. During typical years, and to some extent during 1958, there was a tendency throughout the water column for a decrease in stability to occur during late winter and early

spring, followed by an increase in stability with a maximum in June and July. The values of $10^5 E$ throughout the water column range from about 150,000 to 400,000 m^{-1} at the surface to about 20 to 200 m^{-1} in deep basin water.

The stability shows a roughly exponential decrease with depth during most of the year. However, consistent exceptions occur which are probably associated with the circulation pattern of the fjord. During September, October, and November, regions of low stability occur at depths of 20 to 50 meters. While their intensity varied, these low stabilities were present during all three years. Since fluctuations in the density gradient initiated by a change in surface salinities and temperature produce only a progressive increase or decrease in stability, such inversions must have resulted from lateral intrusions of water with different density characteristics. A similar pattern occurred in February to May of 1959 at 60 to 125 meters depth. These isolated regions of low stability suggest intrusions of outside waters. When the longitudinal profiles of the fjord are examined during these periods, such intrusions often appear as an upfjord tongue of density isolines. The strongest intrusion which occurred in the spring of 1959 is illustrated in figure 33.

In order to apportion the relative effectiveness of temperature and salinity in establishing stability, temperature and salinity gradients, based on values linearly interpolated to the fifteenth of each month, were computed for three layers of water at representative depths. The gradients were then

weighted according to their relative effect in establishing the stability. The results of these calculations are presented graphically in figure 34.

In the surface layer, a salinity gradient is much more important in determining stability. Throughout the three years, the average contribution of the temperature gradient was only 10 per cent; however, in July 1959 the temperature gradient was more important. At intermediate depths the effects of the salinity gradient diminish. Between 75 and 100 meters the average effect from temperature is 47 per cent; during winter and early spring the temperature effect may be over 90 per cent. A similar situation exists in the deep water, where average temperature contribution is slightly less than 40 per cent. Thus for short periods temperature may be the sole determiner of stability.

An important consequence of the role of temperature in determining stability is related to the delay in the occurrence of the summer maximum temperature at the surface, as shown in figure 24. This delay may produce a situation where the temperatures at the surface during the preceding summer determine the stability pattern of intermediate and deep waters during the following winter and early spring; for example, the unusually warm summer of 1958. By late fall of this year, high surface temperatures had established a sharp gradient in the surface layers, but as the temperature maximum and associated gradient were transmitted downward, this gradient progressively overshadowed the salinity gradient in establishing the stability. During late winter and early spring, the temperature

gradient was 85 per cent responsible for establishing the stability at intermediate depths, and in deep water the temperature gradient was solely responsible for the existing stability.

Thus, while the general pattern of σ_t follows the salinity pattern, it does not follow that the salinity is always the most important factor in establishing the stability pattern. At times, particularly in deep water, the stability is determined predominantly by the temperature gradient, and in surface water during periods of low runoff, the temperature gradient may be as important as the salinity gradient in establishing the stability pattern.

6. Eddy diffusion

Vertical mixing in Indian Arm results from convection, vertical turbulence, vertical advection, or a combination of these mechanisms; and its presence is shown by the observed vertical transfer of such conservative properties as heat or fresh water. Convection is most strongly developed during periods of extreme winter cooling. At this time, surface cooling increases the density of the surface water to the point where it sinks beneath the immediately underlying waters. However, such convection is inhibited by any appreciable stability. Examination of the annual cycle of stabilities (see Fig. 30, 31, and 32) suggests that at all times the stability in the upper layers is large enough to prevent any deep-reaching convection. Consequently, with the possible exception of deep basin waters, it is believed that convective vertical transfer

of properties did not occur to any significant extent during the three years studied. The vertical transfer of properties was therefore most likely effected by eddy diffusion associated with vertical turbulence.

An approximation of the magnitude of eddy diffusion can be gained by considering the rate of vertical transfer of a property. If a time is selected when the advective transfer of a property is negligible, and if one assumes that the observed vertical transfer is due to turbulent transfer alone, then it is possible to calculate an eddy diffusion coefficient. This can be expressed as follows: let A be the eddy coefficient, Q the amount of heat, T the temperature, P the amount of salt, S the salinity, z the depth, ρ the specific heat of the water, and t the time, then:

$$Q = \rho A_t \frac{\partial T}{\partial z} dt \quad (1)$$

$$Q = A_s \frac{\partial S}{\partial z} dt \quad (2)$$

By assuming that A is constant over the time and depth considered, equations (1) and (2) can be approximated for numerical analysis by the difference equations (Tabata and Pickard, 1957).

$$A_t = \rho \frac{T}{t} \frac{(\Delta z)^2}{T_1 T_3 2T_2} \quad (3)$$

$$A_s = \frac{S}{t} \frac{(\Delta z)^2}{S_1 S_3 2S_2} \quad (4)$$

where T_1 , T_2 , T_3 and S_1 , S_2 and S_3 are the temperatures and salinities respectively at equidistant depths, Δz is the distance above or below T_2 and S_2 ; and ρ is taken as 1 gram per cubic centimeter.

During most of 1958, the distribution of properties within the fjord suggests that the water mass below 100 meters was relatively undisturbed. Eddy coefficients have been calculated for these periods and are presented in table VIII with A_t values from Bute Inlet (Tabata and Pickard, 1957).

The approximation is commonly made that A_t equals A_s under conditions where eddy diffusion greatly exceeds molecular diffusion (Salen, 1950). In Indian Arm the calculated values for A_t and A_s are several hundred times as large as would be expected if the observed mixing resulted from molecular diffusion alone. Therefore, it is assumed that if the observed mixing results from eddy diffusion alone, A_t will approximately equal A_s in the deep basin waters of Indian Arm. This is not the case; during April, for example, the difference was quite large. Some departure from the basic assumption has occurred, and the most likely explanation is that there was an occasional weak horizontal inflow of water (having temperature and salinity characteristics different from the fjord waters) from outside. These weak inflows would not be readily identifiable by the distribution of properties alone. Conversely, periods with close correspondence of A_t and A_s suggest that the deep water was virtually undisturbed between the times of observations. During these periods, determinations of changes in non-conservative properties within the fjord can be made on the assumption that the observed changes were not influenced by advective processes.

The somewhat higher A_t values in Indian Arm compared with those under similar stability conditions in Bute Inlet suggest

that turbulence may be a more important feature in the vertical mixing of deep water in the former than in larger British Columbia fjords. This is most likely associated with the shallow sill and the "narrows". Such features contribute to a more intense tidal pulse which supplies the necessary energy for the higher levels of turbulence.

7. Circulation

The circulation pattern of Indian Arm is dominated by one basic factor: the interchange of water between the basin and the adjoining water mass is restricted to depths at or above the sill. If the conditions in adjoining water masses were constant, the distribution pattern of conservative properties would be uniform and constant from sill depth to the bottom of the basin, and similar to those existing at sill depth in the adjoining water mass. Above sill depth, these would be continuous with those of adjoining waters.

Departure from this basic pattern is primarily due to two features: the fluctuating fresh water inflow and the wide variations of properties occurring in adjoining water masses. These variations occur with regular and irregular time scales, with the regular cycles ranging from diurnal to several years. The intrusions in and mixing of these waters with the fjord water mass produce the observable changes in the properties.

The main driving force for fjord circulation is the inflow of fresh water from the river and from peripheral streams. This accumulation of less dense water toward the head of the fjord results in a seaward slope of the water surface which induces a seaward movement of brackish water. The seaward movement of

this surface brackish layer entrains more saline water from the underlying water mass. Continuity considerations require that there be a compensating movement of water into the fjord at some sub-surface depth. The basic pattern that results is a two-layer system with net outflow on the surface and net inflow at intermediate depths. This pattern is characteristic of most fjords. Various modifications of this basic pattern occur in both British Columbia and Norwegian fjords. Most of these modifications are associated with the relationship between the depth of the sill and the thickness of the brackish surface layer. If the depth of the sill is equal to or less than the thickness of the brackish layer, stagnant conditions result, with a marked restriction in flushing of the fjord (Strom, 1939). None of the investigations conducted in mainland British Columbia fjords to date have shown this type of stagnant condition (Anon., 1959-1955, 1956c-1959c).

Superimposed upon the basic circulation pattern are the effects of tidal action. While the net tidal flood must equal the net tidal ebb, a particular tidal cycle usually shows some asymmetry. Thus, all the ebbing waters may not be returned by the following flood and vice versa. This temporary retention of a portion of the total tidal inflow for approximately six hours permits a small amount of mixing of the adjoining waters with fjord basin waters, with the result that a portion of the fjord water mass is renewed during each tidal cycle. The amount of flushing produced by this mechanism is usually small compared to the flushing action produced by entrainment.

While the circulation pattern of Indian Arm is the result of many different processes, the total circulation can be most clearly seen if it is resolved into two basic components, horizontal movements and vertical movements.

In common with almost all fjords, the main horizontal movement consists of a surface outflow of brackish water, driven by the accumulation of fresh water within the fjord. The net amount of fresh water entering the fjord is the sum of the river runoff, peripheral stream runoff, discharge from the Buntzen power plant and direct precipitation, less that lost by evaporation.

The mean annual precipitation at the Buntzen power plant opposite station D, which is halfway up the fjord, is 275 cm. per year. If this precipitation rate is assumed to be representative of the entire fjord, the mean annual contribution by precipitation over the entire fjord north of the narrows would be $49.6 \times 10^6 \text{ m}^3$ per year. Thus the excess of precipitation over evaporation, converted to mean transport out of the fjord, would be approximately 1.1 m^3 per second. Therefore, with the excess of precipitation over evaporation averaging less than 5 per cent of the total runoff, the fresh water inflow into the fjord can be considered due to runoff alone.

The estimated mean monthly runoff of the Mesliloet River is shown in figure 22, with estimates of the additional fresh water contributed by peripheral streams and the Buntzen power plant. This inflow is dominated by the biyearly peaks previously discussed. Apart from short-term tidal fluctuations, the

amount of water in the fjord remains relatively constant.

Thus, stability factors require that most of this fresh water be removed by the seaward flow of the surface brackish layer. A small portion of the fresh water mixes downward and temporarily remains in the fjord. Therefore, the total transport out of the fjord equals the total fresh water inflow plus any entrained water.

There is a general increase in salinity from the head toward the mouth of the fjord, indicating that the seaward movement of fresh water has entrained portions of the underlying salt water. This increase in salinity ranges from 3 to 5% during the fresh water minima to 15% during the maxima. If the salinity of the outflowing water at a particular point is known, and if the salinity of the entrained water is known, it is possible to calculate the total transport to this point, knowing the fresh water inflow. This can be expressed as follows: let X_0 be the head of the fjord and T_0 the total transport at X_1 . If S_0 is the observed salinity of the outflowing surface layer at X_1 , S_e the mean salinity at the depth of the entrainment of X_0 to X_1 , and D the total amount of runoff entering between X_0 and X_1 , the total transport at X_1 will be:

$$T_0 = \frac{D S_e}{S_e - S_0}$$

The difference between the total transport and the fresh water runoff will represent the compensating volume of water which enters the fjord at some surface depth.

Direct current measurements indicate that at the depth where no net motion occurs salinities average approximately 90

per cent of base salinity. The monthly rates of total transport out of the fjord at station F and at the narrows have been calculated using this depth as an approximation of the depth of entrainment.

From these net transport volumes it is possible to calculate the theoretical net currents at station F and at the narrows. Using the assumed depth of entrainment as the depth of no net motion, these values are calculated to range from 0.3 cm. per second during the years of inflow minimum to 8 cm. per second during the inflow maximum. Corresponding values at the "narrows" are 1 and 27 cm. per second respectively.

The horizontal water movements, which are net currents resulting from the outflow of fresh and entrained waters on the surface, vary according to the state of the tidal ebb or flood. On a flooding tide, the outflowing surface current is retarded, causing an accumulation of surface waters within the fjord. During the ebbing tide, the accumulated surface waters contribute to the velocity of the outflowing surface waters. Direct observations, and one 40-hour series of current measurements during a month having 27 cubic meters per second fresh water inflow, suggest that the surface outflow is completely stemmed only during periods of low runoff. As a consequence, the tidal exchange normally takes place at depths below the surface brackish layer.

The tides in Indian Arm are of the declinational type. In common with most deep estuaries, the high and low tides occur virtually simultaneously throughout the length of the fjord, suggesting that the tides act primarily as a pulse at the mouth

of the fjord (Dawson, 1920). The tidal amplitude ranges to 4.6 meters during spring tides, with a mean amplitude of approximately 3 meters. If one assumes that the tidal current is uniform across a particular section of the fjord and varies sinusoidally, then an approximation of the tidal currents at various tidal stages can be calculated. This can be expressed as follows: let V_s be the tidal velocity at section X_1 during tidal stage s , A be the surface area upfjord from X_1 , a the cross-sectional area of the channel, h the height of the tide at stage s , and t the time from the preceding slack water.

Then:

$$V_s = \frac{A}{a} \cdot \frac{dh}{dt}$$

Calculations for tidal velocities for various tidal amplitudes and tidal stages are presented in Table V. These approximations show tidal currents that exceed 50 cm./sec. at the "narrows". The frictional effects of the sides and bottom of a fjord probably insure that the flow through the fjord is not laterally uniform. These effects are most pronounced in extremely restricted sections such as the "narrows". Thus the calculated tidal velocities are considered minimal.

Tidal velocities at the "narrows" play an important role in the dynamics of fjord circulation. Both as a consequence of the reversing actions of the tidal currents and of the velocity gradient existing between the two layers, intense turbulence can develop which effectively mixes portions of the surface and deeper layers of water. This results in two different types of water entering the fjord. When high velocity gradients between

the surface and underlying water exist, appreciable volumes of intermediate depth water (Fig. 35, layer C) are formed and enter the fjord. Because of its intermediate density it will probably enter just beneath the surface layer. The results of such inflow can be seen clearly in the annual cycle of salinities at 2.5 meters, as previously discussed. The remainder of the underlying layer (Fig. 35, layer B) flows into the fjord in the form of a tongue along an isopycnal surface. During periods of low velocity gradient, either at the beginning and end of the flooding tide, or during low fresh water runoff, all the inflow moves into the fjord basin and flows upfjord at a depth appropriate to its density. Such intrusions are often apparent in the distribution of various properties, as shown in figure 36.

The development and maintenance of such tongues can be explained by considering the mechanism of tidal exchange in Indian Arm. The physical restrictions of the "narrows", the density structure within the basin, and the associated high current velocities all combine to force the water flowing in during a flooding tide to take the form of a jet, as shown in figure 37a. During the ebbing tide there is no comparable restriction and the outflowing water is drawn from a comparatively wide depth range, as shown in figure 37b. Consequently, there is a net addition of outside water during each tidal cycle throughout the depth represented by the tidal tongue (Fig. 37c). These form and maintain the characteristic tongue which demarks the inflowing waters.

In order to demonstrate the tidal formation of this tongue, a series of hydrographic stations was occupied on the sill north and into the basin during different periods. In each series, temperature, salinity, and oxygen were determined at standard depths at four stations. At this time the difference in oxygen content between outside waters and the basin was larger than any of the other properties measured. Oxygen isopleths have therefore been plotted in order to show the tidal inflow (Fig. 38). The distribution of isolines during the middle of the ebb is shown in figure 38a. The nearly vertical position of the 5.75 mg./l isopleth and the slope of the 5.50 mg./l isopleth suggest that most of the water between 5 and 50 meters moves seaward en masse. In figures 38b and 38c the tongue of inflowing tidal water during early and late flood stages is evident. Figure 38d shows the condition at the beginning of the following ebb. The resulting upfjord tongue is presented in figure 39, which is a longitudinal plot of the entire fjord basin.

The net effect of this mechanism is the addition of outside water to the fjord basin, usually at an intermediate depth. The volume of this addition is somewhat in excess of that calculated for entrainment alone. At times this movement is clearly marked by the distributions of temperature, salinity and oxygen within the fjord. Examination of the longitudinal plots during each month have been made in order to estimate the depth of inflow. These depths, and the northernmost extent of these inflows, are presented in table VII. During months when there was no clear

evidence of inflow at an intermediate depth, even though the characteristics of the water at the sill suggested that such an inflow might be detected, it is assumed that the inflow occurred chiefly just beneath the brackish surface layer. The combination of a rough quantitative estimate of the amount of entrained water, and a qualitative estimate of the extent and depth of inflow provides an approximation of the horizontal circulation of the fjord.

C. Distribution of Non-conservative Oceanographic Properties

1. Oxygen

The pattern of dissolved oxygen distribution in Indian Arm is much more variable than that of temperature and salinity, since the concentration is influenced by biological activity. Biological processes may increase as well as decrease oxygen levels. The major factors which bring about an increase in oxygen are diffusion from air to water; production by photosynthetic organisms; introduction in water from rivers and peripheral streams; introduction from intruding water masses of higher oxygen content than corresponding fjord waters. Those processes which decrease oxygen concentrations are respiration by living organisms; utilization in chemical oxidative processes; diffusion from water to air; seaward transport in the surface layer.

The most effective processes which increase the oxygen level tend to occur in surface waters; most of the processes tending to lower the oxygen level occur in deep water. Consequently, the vertical oxygen distribution usually ranges from approximately 100 per cent saturation at the surface to quite low values in deep water.

In common with other British Columbia fjords having low runoff, Indian Arm shows great variation from year to year in the distribution of dissolved oxygen. The measured values at selected depths at station D are shown in figure 41. Two features of these distributions are conspicuous: a nearly unimodal cycle evident at 2.5 to 50 meters and a long-term decreasing trend in deeper water.

The curves depicted in figure 41 have two components, a maximum which occurs in the spring of each year, and a much smaller overlapping maximum which occurs in the fall. During 1957 and 1958, the spring maximum developed in April and May, and was the larger of the two maxima. In 1959, the fall maximum appeared to be the larger of the two. While this may be an artifact introduced by the time of observation, it is considered unlikely when the duration and intensity of the preceding spring maxima in 1957 and 1958 are compared with the cruise schedule of spring 1959. It is unlikely that a well-developed spring maximum was missed. The mean annual cycle for the surface and intermediate depths is illustrated in figure 42.

The deep-water oxygen concentrations show a long-term tendency to decrease between periodic complete flushings of the fjord. This decreasing trend is broken by short-term seasonal increases, which usually occur during the spring of the year. The unusually high oxygen concentrations present in September 1956 are difficult to interpret because earlier data are lacking. These values decreased to a low of 4.5 mg./l in January of 1957. Between January and April, a large sub-surface inflow raised oxygen values in the deep waters to the highest levels (values exceeding 7.00 mg./l).noted. These decreased consistently to 2.1 mg./l during March 1958. At this time, another inflow increased the deep-water oxygen, which reached 3.2 mg./l by May 1958. This was followed by another consistent decrease lasting until March 1959, when a low of 1.1 mg./l was observed. This latter value is equivalent to 10 per cent saturation, and is among the lowest noted in any mainland British Columbia

fjord, since very few have oxygen concentrations less than 103
30 per cent saturation in deep water. The deep-water oxygen
concentration is one of the primary differences between British
Columbia and Norwegian fjords. In the latter, low oxygen values
and associated stagnant conditions are frequently present. In
general, oxygen concentrations in British Columbia fjords de-
crease with depth down to 200 meters, where there is 40 to 60
per cent saturation. From this depth to the bottom there is
little change (Anon., 1951-1955, 1956c-1959c).

The oxygen distribution pattern in Indian Arm suggests that
the deep basin water is relatively undisturbed from March
through December of each year. This view is supported by relat-
ed temperature, salinity and density data. It thus provides an
opportunity to establish the rate of oxygen consumption by re-
spiratory and oxidative processes.

The longitudinal pattern of oxygen distribution in surface
and sub-surface layers is difficult to establish. There is a
noticeable tendency for values to increase from head to mouth
during winter and to decrease from head to mouth during summer.
Moreover, the highest oxygen values which were observed during
the spring and fall were at the head of the fjord. These
usually ranged from 100 to 150 per cent saturation, but during
1957 they were as high as 250 per cent saturation. This value,
observed on April 4, greatly exceeds the highest values previ-
ously reported for other British Columbia fjords (Anon., 1951-
1959). Values in excess of 200 per cent saturation are unu-
sual. In April 1957, the mass of high-oxygen water extended
from the head south to the region between station B and C, and

was confined to the upper 2 meters. By May 23, oxygen concentrations had decreased to 94 per cent saturation. Excessively high values are probably associated with extremely intense and short-term phytoplankton bloom.

Both surface and sub-surface maxima and minima of oxygen concentrations may result from several different causes. One can distinguish maxima which are due to biological activity and others which result from an influx of water with high oxygen concentration from outside the fjord. Those due to biological activity usually exist at shallow or intermediate depths and form near the head of the fjord and spread southward. Maxima which result from an influx of high oxygen water may occur at any depth in the fjord and spread northward. A third type of maximum can result from the influx of low oxygen water at intermediate depths and produces the appearance of a high oxygen core at greater depths.

The origin of oxygen minima may also be either kinematic or biological. A low oxygen layer can develop from the decomposition of organic material which is slowly sinking following a bloom at the surface. A low oxygen core can result from an influx of low oxygen water or an influx of high oxygen water near the bottom producing the appearance of a low oxygen core in intermediate depths.

Two features of the upper 30 meters which recur each year are a shallow minimum at the mouth in fall and a shallow maximum at the head in spring. The maximum is apparently biological in origin, whereas the fall minimum is probably associated with circulation.

In about May of each year, an oxygen maximum forms at about 2.5 meters near the head of the fjord. The maximum appears to be related to the spring phytoplankton bloom, and at times can be traced to the mouth of the fjord. The extent of the maximum varies from year to year. In 1957 a maximum of 112 per cent saturation was observed at 2.5 to 5 meters. In 1959 a maximum of 145 per cent saturation occurred in June and had its strongest development at the head of the fjord. This maximum could be traced seaward to station E and was most intensively developed between 2 and 3 meters depth. The most extensively developed maximum occurred in 1958 (Fig. 43). The maximum began in April, reached 150 per cent saturation in May, and gradually tapered off.

The second recurring feature of the shallow waters is a minimum which occurs during August and/or September of each year. The minimum ranges from 10 to 20 meters depth, and during its maximum development extends from the sill north to about station C (Fig. 44). During 1957 and 1959, the minima were similar in intensity and extent, occurring during September and August respectively. In August 1958 the minimum occurred at 20 meters.

The strongly developed minimum near the mouth of the fjord suggests the intrusion of a tongue of outside water with low oxygen concentration. It is a short-term feature, however, and was never observed to persist more than one month. A low-intensity residual minimum, no longer continuous with outside water in the region of the sill, usually can be observed during the following month.

Both of these features support the conclusions reached in the foregoing discussions on other oceanographic properties: that 1957 and 1959 are similar in pattern and represent typical years, while 1958 represents an atypical year. The difference observed in these years also suggests that an indication of factors contributing to the extensive development of the shallow maximum in 1958 may be found by comparison of the general oceanographic features of 1957 and 1959 with those of 1958.

When the distributions of the other oceanographic properties are examined, several features are apparent. The period May through July 1958 had intermediate levels of river runoff, between a high in 1957 and 1959. However, examination of surface temperatures focusses attention upon the unusually early increase in surface temperatures in 1958. During 1958 the highest surface temperature, 21° C., of the three years was recorded (Fig. 24). High temperatures extended to depths of more than 50 meters at this time. The surface light intensities measured during the cruises of 1958 were higher than those measured during corresponding periods in 1957 and 1959. Therefore the development of a phytoplankton population was probably encouraged by the warmer temperatures and high light intensities. Moderate runoff, and concomitant small entrainment of water and contained phytoplankton, would have permitted such a population to develop more fully than was possible during corresponding periods in 1957 and 1959.

Both maxima and minima can be seen at greater depths and it is difficult to determine which is the more important. When a maximum occurs there are two associated minima, unless the

maximum occurs at the bottom. In this discussion, attention will be centered upon the maximum or minimum which seems to have originated at the mouth of the fjord. Formation at the mouth of the fjord implies that the maximum or minimum is associated with intrusions of outside waters, and in most cases this is supported by temperature and salinity data. One exception to this occurred in 1959.

In May 1959, a well-developed maximum was observed which extended from the head of the fjord south to about station E (Fig. 33). This maximum occurred at 25 to 30 meters, with its strongest development at the head of the fjord, and it gradually decreased toward the mouth. The tongue of high oxygen concentration appears to have its source in a center at a depth of 20 to 30 meters near the head. No sign of this tongue was observed on a preceding cruise 5 weeks earlier, and only a trace was evident at station A on the succeeding cruise 5 weeks later. This tongue had dissolved oxygen values as high as 115 per cent saturation, which were probably biological in origin. Three possible mechanisms can be suggested for its development: (1) the tongue could have developed in situ as a result of an unusual distribution of the phytoplankton population; (2) the tongue could be a residual maximum resulting from the development of an intense phytoplankton bloom in the top 30 meters (this region of high oxygen might later have been cut off from the surface by an inflowing tongue of low oxygen water from outside the fjord at depths of 10 to 15 meters); (3) the tongue could also have resulted from the transport of water from a high oxygen source of production at the head of the fjord.

The first suggestion is least tenable, primarily because of the amount of light at 30 meters in May, 0.15 per cent of surface light. This is at or below the compensation depth of most phytoplankton. It is therefore unlikely that an in situ population could produce sufficient oxygen to raise the levels from the observed values of 3.18 to 5.98 mg./l. in April to the values approaching 10.0 mg./l. observed during May.

The second hypothesis is even less tenable. If one assumes that the inflowing tongue of low oxygen water intruded into the high oxygen region just prior to May, the region would probably have had an average oxygen level of about 10.4 mg./l., required throughout a region 14 kilometers long and 30 meters deep. This is a rate of oxygen production exceeding that observed during the bloom in sub-surface water in the spring of 1958. Examination of the T/S structure of the water of this region of low oxygen yields no suggestion that this water is of a different origin. The foregoing, combined with the fact that most of the region under discussion is at a depth well below the optimum light intensity negates the second hypothesis (that the tongue was formed by an influx of low oxygen water cutting off a high oxygen region).

The third hypothesis appears to be the most likely explanation. Such an area of high oxygen can be seen at station A, where values in excess of 10 mg./l. occur at all depths to 30 meters. The internal structure of the tongue itself indicates that the highest values occur at the head and decrease in value toward the mouth. This is contrary to the general trend shown by the oxygen distribution in the preceding months, when at all

depths from the surface to 50 meters there was an increase in oxygen from the head seaward. This decrease was still observed at 50 meters during the May cruise. By postulating that the high oxygen layer flows from the head of the fjord at intermediate depths, the requirement for oxygen production is reduced by about one-third to one-half, placing it just within the possible phytoplankton production during the time available.

However, the possible dynamic mechanism providing such a downfjord flow is not obvious. The kinematics of the fjord, as represented by the May data, offer no supporting evidence for such a flow. It is possible that the tongue is the residual effect of a flow occurring between the cruises and that it does not continue during May. Whatever mechanism is responsible for the development of this tongue, one must take into account the fact that it is supersaturated with oxygen and is probably biological in origin.

All of the other observed maxima can be traced to the mouth of the fjord. This implies that they result from the inflow of water having different dissolved oxygen characteristics. These maxima occur most frequently during the winter months--usually between October and February.

The highest maximum observed during the 3 years occurred in January 1957. At this time, the entire southern half of the fjord basin below 50 meters was occupied by a mass of water with 7.2 mg./l. dissolved oxygen. A similar maximum was observed during December of 1957, when a smaller mass was observed intruding as a tongue at 50 meters and extending northward to Station C. Again, during February of 1959 a large tongue

intruded at 75 meters and was traceable northward to station

B. All three of these water masses were of sufficient volume to influence the entire length of the fjord at a particular depth.

At other times of the year--usually during the fall months --smaller tongues were observed. One set of these occurred in September through November 1956 at about 50 meters. During December 1958, a slight tongue of high oxygen water dropped into the basin at station F. Examination of the oxygen distribution during the following month shows little residual influence and suggests that the volume of water which intruded was small.

These oxygen distributional patterns suggest that oxygen, even though it is a non-conservative property, can provide an indication of the circulation pattern of the fjord. This is primarily a consequence of two features. The horizontal distributions of temperature and salinity usually show very small gradients compared with dissolved oxygen. This distribution results in part from the influence that these properties have upon density. Changes in temperature and salinity result in changes in density, and the resulting intrafjord movement of water seeking its own density confounds the picture. On the other hand, oxygen levels are influenced by biological activity. This is particularly evident when the steady decrease in dissolved oxygen of the relatively undisturbed deep basin waters is considered. The reduction in dissolved oxygen at all depths below the euphotic zone proceeds at a rate much greater than the rate of decrease in density resulting from the downward transport of heat and fresh water. Waters entering the fjord come

in at or above the sill depth and therefore usually have a much higher dissolved oxygen content than those of the basin. As a consequence these intrusions show up clearly.

Such intrusions can be noted in two ways. If an oceanographic observation is made during an intrusion, the intrusion is apparent as a tongue of high oxygen water traceable to the mouth of the fjord. If the observation is made shortly after such an intrusion, the intruding tongue is evident as a detached mass of water with different oxygen characteristics at some intermediate depth. If the observation is made after the intruding waters have had a chance to mix with the fjord waters, the relative slope of the isolines reveals the previous intrusion. To illustrate this the following schematic model is presented. Figure 45 shows the effects of the intrusion of a tongue of high oxygen water entering a fjord which has a small horizontal oxygen gradient. In figure 45a are shown the isolines prior to the intrusion when they are approximately parallel to the water surface. An intruding mass of high oxygen water would appear as a tongue replacing the upper isolines (Fig. 45b). Successive vertical mixing of the intruding tongue would develop a pattern of isolines which slopes downward (Fig. 45c). If the intruding tongue were an oxygen minimum, the isolines resulting from vertical mixing would slope upwards from head to mouth. Tidal intrusions would produce a displacement of the isolines roughly proportional to the influence of the mixed intruding waters. A typical pattern resulting from such intrusions is shown in figure 45d.

Thus, examination of the oxygen isolines at a particular time can provide at least a qualitative picture of the circulation pattern of the preceding month provided that: the examination is confined to depths below the euphotic zone, and the outside waters at sill depth show different characteristics in their dissolved oxygen content which distinguish them from waters in the fjord basin.

2. Phosphates

There is little available data concerning the distribution of nutrients in British Columbia fjords. During 1926-1931 some data were gathered at irregular intervals in the Strait of Georgia and several of the southern fjords (Anon., 1931; Anon., 1956; Carter, 1934; Hutchinson et al, 1929).

In March 1959, a ship operation with the University of Washington provided the opportunity for a survey of $\text{PO}_4\text{-P}$ distribution in Indian Arm. The results of this survey, showing both horizontal and vertical gradients, are presented in figure 40.

The phosphate values at the surface were 1.89 gram-atoms of phosphorus per liter of water at the fjord head and 1.49 gram-atoms per liter near the mouth. This difference, which probably existed in spite of the entrainment of waters with higher phosphate values into the surface layer, results from phosphate utilization by the phytoplankton population. Thus the distribution can be used to obtain an approximation of primary production--if seaward transport and the amount of entrainment are known.

The phosphate contribution of the Meslilloet River can be estimated by considering the relationship between the salinity-phosphate values at two depths off the river mouth. In March 1959 the salinity at the surface was 20.26 and the $\text{PO}_4\text{-P}$ was 2.40 gram-atoms per liter. If one assumes that the values observed at the surface result from a mixture of fresh water containing virtually no phosphates with water such as that at 2.5 meters, then the theoretical phosphate level resulting from the mixture would be 1.96 gram-atoms per liter. The agreement between the calculated and the observed surface phosphate values suggests that at this season the river contribution is negligible.

The vertical phosphate distribution shows approximately a threefold increase, from a minimum of 1.49 gram-atoms per liter at the surface to a maximum of 5.29 gram-atoms per liter at 200 meters. Since the values at the depth of the sill are estimated to range from 2 to 3 gram-atoms per liter, it would appear that some concentrating mechanism is increasing the phosphate values in deep water. Such high values in Indian Arm could result from a combination of the following mechanisms:

- a. the fixation of phosphates by phytoplankton in surface waters with their subsequent descent and decomposition of the latter and release of phosphates in deep water.
- b. the consumption of the phytoplankton population by herbivorous zooplankton with subsequent phosphate release by decomposition of the animal bodies and their fecal pellets (the rate of this downward transfer of phosphates would be increased if the particular

zooplankters exhibited extensive vertical migration).

- c. the decomposition of terrestrial organic matter that has sunk to the bottom of the fjord.

These processes would result in a net transport of phosphates from the surface to deep water.

If the water replacement pattern of the fjord results in outside water being introduced at intermediate depths, or if the bottom waters are slowly replaced, the downward transport of phosphates would produce phosphate levels appreciably higher than those at corresponding depths in outside waters. Examination of the (Anon., 1931) data suggest that this may be the case in Indian Arm. The phosphate levels reported for deep basin waters in Indian Arm are approximately three times as large as those at corresponding depths in the Strait of Georgia.

Thus it appears that the organisms within the fjord effectively collect and concentrate the phosphates (and associated inorganic nutrients) of the surface waters in the deep basin. Such a reservoir of nutrients could return to the surface layers during periods of low stability or at times of large inflows into the deep basin of the fjord. The influx of high density water would force the nutrient-laden bottom waters up to the surface. It should be noted that this picture is based on data of known reliability from only one cruise in 1959. Since the reliability of the data from the 1931 period is not known, they are not readily comparable.

VI. PHYTOPLANKTON POPULATIONS

A survey of the phytoplankton populations of the fjord was based on the two different collection techniques, with the emphasis being divided between the net- and the nanoplankton. By definition, the nanoplankton are those organisms that will pass through a No. 20 mesh net (i.e. smaller than 50-75 microns).

A. Netplankton

The netplankton were sampled using the standard Clarke-Bumpus plankton samplers with No. 20 mesh nets. By using the samplers in tandem, three different depths were done simultaneously. Normally the length of the fjord was sampled at 6 standard depths (0, 5, 10, 30, and 50 meters) at least once during each cruise. The samplers were retrieved and recast at stations A, C, E, and H, thus sampling the fjord with three approximately equal tows.

All samples of the netplankton have been examined microscopically (with emphasis on the phytoplankton), but only a 17-month series has been studied in detail.

The net phytoplankton samples studied covered the period extending from June 1957 to October 1958; thus they included a partial and a complete spring phytoplankton bloom and two fall blooms. The spring blooms occurred in May-June and both the fall blooms occurred between September 1 and October 15. Indications of the onset of the fall blooms were noted as early as August 6. While the fall blooms occurred during the expected times for north temperate waters, the spring blooms were slightly later than would have been expected and had lower cell densities than either of the observed fall blooms. On the basis of

the samples examined it is difficult to establish whether this situation is typical for fjord phytoplankton populations, since these data cover only one complete spring bloom.

Analysis of the net phytoplankton usually showed a numerical dominance of some species of armoured dinoflagellates. The two fall blooms observed were composed primarily of the same species of Dinophysis, provisionally identified as D. acuta Ehrenberg. The two spring blooms were dominated by the same species of Ceratium, which has been provisionally identified as C. fusas Ehrenberg.

The species of Dinophysis occurring at times other than during bloom months has not been determined. However, D. acuta was present in all but 3 of the 17 sets of monthly samples examined. Further examination of the samples will probably show that it is present during these three months also. Ceratium fusas, the spring dominant, was present in the phytoplankton throughout the year. Thus it is evident that the species responsible for blooms are present in the fjord throughout the entire year.

Species of Peridinium were noted in all monthly samples, often with two or more species being abundant during the same month. In 13 of the 17 months examined, some species of Peridinium ranked as one of the three most numerous net phytoplankton. Throughout the year, numerous other genera of dinoflagellates were noted in the net phytoplankton. Some of the more common of these were species of the genera Oxyphysis, Gymnodinium, Goniodema, and Phalacroma.

On one occasion, and possibly on two others, diatoms played a significant numerical role in the phytoplankton. The diatoms were also important, primarily because of their larger size, when the relative biomass of different community species is considered. During March 1958, a centric diatom provisionally identified as Coscinodiscus oculus-iridis Ehrenberg, was the dominant species. This species ranked second in the preceding month and third in the succeeding month, thus suggesting that this species is the early spring dominant preceding the spring dinoflagellate bloom.

Many of the dinoflagellates were noted to vary in size and shape throughout the year. A species in which these changes were most marked, Ceratium longipes (Baily) Gran, was analyzed in detail. This species has long attenuated "horns" which varied greatly in length during the course of a year. On examining various factors to which this change might be related, a statistical correlation was found to exist between the viscosity of the water from which the specimens were collected and the "horn" length. The degree of correlation was estimated using the so-called product-moment correlation coefficient technique. A negative correlation coefficient of 0.88 was found. This coefficient is highly significant at the $P = 0.05$ level. It is suggested that projecting "horns" decrease the sinking rate of the organism. An increase in "horn" length, correlated with a decrease in water viscosity, could be considered as having a definite survival value in preventing the organisms from sinking out of the euphotic zone.

B. Nannoplankton

It is impracticable to collect nannoplankton with plankton nets, so a "whole water sample" technique was used during this study. At each station samples of water were collected from 0, 5, 10, 20, 30, 50, 100, 150, and 200 meter depths using Atlas water bottles. The organisms in a given water sample were killed and fixed using Lugol's solution modified by the addition of glacial acetic acid (Rodhe, personal communication), and subsequently removed from the water sample on Millipore filters (HA, pore size 0.45 microns). The filters, with their surface layer of nannoplankton, were cleared with Cedarwood oil and permanently mounted on microscope slides for subsequent examination.

On the basis of visual examination, it is difficult, if not impossible, to separate autotrophic members of the nannoplankton population from the heterotrophic members. Certain flagellates have been reported to exist both as autotrophs and heterotrophs, depending on environmental conditions. This may be significant, as it is quite possible that by utilizing organic food during periods of low light intensity certain flagellates may be able to exist under conditions which would prove fatal to other photosynthetic organisms. Therefore, all flagellated members of the nannoplankton have been considered together in this study.

The nannoplankton samples examined covered the period extending from June 1957 through March 1958; thus, a total fall phytoplankton bloom and the beginning of the following spring bloom were studied. During this period, a wide range of cell

densities was noted. These values ranged from 7×10^6 to 967×10^6 cells per cubic meter in the euphotic zone, and from 5×10^6 to 46×10^6 cells per cubic meter at 150 meter depths. The highest densities occurred during the fall phytoplankton bloom, and were followed by a marked winter minimum.

The vertical distribution showed several noteworthy points. If the euphotic zone is defined as that portion of the water column receiving more than one per cent surface illumination, it is found that the cell density remains quite constant from just below the euphotic zone to the bottom of the water column. The greatest fluctuations occur within the euphotic zone. This would suggest that the population of flagellates can be divided into two parts: namely, a portion composed of flagellates which are existing under heterotrophic nutritional conditions, and an autotrophic portion existing within the euphotic zone. The sub-euphotic portion remains relatively constant in cell density throughout the year, whereas the euphotic portion fluctuates widely. This is most likely associated with the distributional pattern of environmental factors, in which the properties of the shallow surface layer vary widely throughout the year relative to the more stable subeuphotic waters.

VII. OCEANOGRAPHIC EQUIPMENT AND METHODS

The oceanographic data used during this study were collected at a series of stations located along the center line of Indian Arm. The exact station positions and depths are presented in table XI, and the cruise dates in table XII. The physical data have been published as data reports (Anon., 1956d-1959d).

The stations were reoccupied on each cruise; the positioning of the ship on station was based on pelorus bearings to known geographic features, or at night by radar rangings.

On each station the bottle cast was made at a series of standard oceanographic depths. These were selected at particularly shallow intervals in the upper 100 meters to insure as complete a coverage as practical throughout and just below the euphotic zone. The standard depths used were 0, 2.5, 5, 7.5, 10, 20, 30, 50, 75, 100, 150, and 200 meters. On occasion, when data suggested that intrusions might be taking place along the bottom, these standard depths were supplemented by sampling at shorter depth intervals between the deepest standard depth and the bottom.

All samples for chemical analyses were collected using Atlas water bottles; samples for biological culture work were collected using plastic water bottles modeled after the Van Dorn water bottle (Van Dorn, 1956).

Temperatures were determined using Richter and Wiese reversing thermometers mounted on the Atlas water bottles. In addition, two bathythermograph casts were made at each station, one with a 70-foot BT, and a second with a 200, 450, or 900-foot BT, depending on the station depth.

All salinities were stored and subsequently analyzed in the laboratory. The standard Mohr method was used (Strickland, 1957). This method involves the titration of the halides in sea water to a chromate end point using silver nitrate standardized against sea water of known chlorinity (Depot d'Eau Normale, Charlottenlund, Denmark).

The analyses for dissolved oxygen were conducted on board using the Winkler method (Strickland, 1957) and 300 ml. samples. In this method iodide ions are oxidized to iodine by the dissolved oxygen in the water sample, and the amount of liberated iodine is determined iodometrically, using standardized sodium thiosulfate as a reducing agent and starch as the indicator.

All light determinations were made using an instrument constructed around two Westinghouse model 856 photoelectric cells (Clarke Submarine Photometer, Fred Schuler). One of the cells was mounted in a watertight case in a frame so constructed as to minimize the amount of shadow from the frame and cable. The other cell was mounted in gimbels on deck. A two-way switch permitted miliampere reading of both underwater and deck cell current output in rapid succession. The cells were standardized against a pyroheliometer.

The current measurements were made while the ship lay to a single anchor on station. A CBI drag was usually used for observations from the surface to 30 meters. On occasion, when weak surface currents were present, the Ekman current meter was used in the upper 30 meters.

All carbon-fourteen used during the study was received as a NaCO_3 solution from the Atomic Energy of Canada, Ltd.

Commercial Products Division), and was made up by dilution with 30% NaCl solution to 1 ml. portions in 2 ml. glass ampoules. The ampoules were sealed and autoclaved in a dye solution to sterilize and to detect any imperfectly sealed ampoules.

The carbon-fourteen solution was added to the culture bottles with a hypodermic. The solution was withdrawn from the ampoule and released at the bottom of the sample bottle. The pipette was then rinsed with water from the middle of the sample bottle; water was then withdrawn by the hypodermic from the top of the bottle for the first ampoule rinse. This rinse was returned to the middle of the sample bottle and the procedure repeated for a second ampoule rinse.

All glassware was washed with concentrated HCL and rinsed copiously with water. Immediately prior to use, the culture bottles were rinsed twice with sea water of the same temperature and salinity as the sample.

After incubation, the samples were filtered first through silk net discs (least pore dimension $56\ \mu$) and then through Millipore filters (HA, pore size $0.45\ \mu$).

The counting of the C_{14} samples and the activity determination were done by the courtesy of Dr. M. S. Doty at the University of Hawaii. This counting was done by a Tracerlab SC16 windowless gas flow counter and a Tracerlab 1000-scaler or a Nuclear-Chicago 161A-Scaler. The planchettes were mounted on Tracerlab E-8b planchette holders, which had been lightly greased to hold the filter down. Instead of prepared "Q" gas,

a mixture of U.S.P. XII Helium bubbled through ethanol at -15° to -5° C. was used.

The light bottle samples were counted for 3 to 10 minutes for a minimum of 1080 total counts. Dark bottle samples were similarly counted, although a minimum total count of 400 was used. These minimum counts gave a standard error of 5%. Some of the dark bottle counts may have a somewhat greater error because, due to pressure of time, it was not always possible to obtain the desirable minimum count.

Every third planchette counted was a carbon-fourteen standard, which was rated at 25,000 counts per minute. This was counted for five minutes. Each sample was corrected to the standard count immediately preceding or following it.

In cases where more than one light or dark bottle was obtained from the same sample, and these replicates were incubated for the same length of time, the values obtained for each (in counts per minute) were averaged if the difference between the lowest and highest figures obtained was not over 25% of the highest value. In the event the difference exceeded 25 per cent, the lower value was discarded.

VIII. DISCUSSION

With the exception of the low salinities in the deep basin, Indian Arm is typical of the low runoff group of southern British Columbia fjords. The basic distribution pattern is a two-layer system, with a thin brackish surface layer overlying a heavier, more saline water mass. Four features control this salinity distribution: the volume of fresh water runoff, the volume and velocity of surface outflow, the fjord physiography, and the character of the outside water mass. The foregoing suggests a replacement pattern dependent upon several mechanisms. During most of the year, the rate of mixing at the "narrows" determines the salinities of the sub-surface waters that are entering the fjord. The pattern of tidal replacement of the fjord waters is such that a small quantity of water is introduced at sub-surface to intermediate levels during each of the tidal cycles. This form of replenishment results in continuous and gradual changes of the salinity pattern and is related to the volume and velocity of surface runoff. These short-term intrusions of salt water counteract in part the tendency for the deeper waters to become progressively lower in salinity because of downward mixing of fresh water. However, a gradual decrease in salinity eventually increases the density gradient between the fjord and outside waters. During years when the density gradient is sufficiently large, a massive intrusion of more saline water into the fjord takes place. Such an intrusion is most likely to occur during the winter minimum when the salinity of the Strait of Georgia water is at its highest and the

fresh water discharge and the contingent surface layer velocity are at their lowest. Such a massive intrusion probably occurred from January through March of 1957, and to a lesser degree during January through April of 1959.

The annual cycle of fresh water entering Indian Arm is a bimodal one. It is intermediate between the two extreme types of discharge patterns common to most British Columbia fjords: that dominated by stored runoff, and that dominated by direct runoff. The relationship between total fresh water runoff and fjord area is similar to that reported for other British Columbia fjords. Thus the fresh water system of Indian Arm is considered to be typical and more closely related to the fjords with low runoff.

The temperatures previously reported for British Columbia fjords have ranged from summer maxima of about 15° C. in fjords with high runoff to 21° C. in fjords with low runoff. Previous observations of winter temperatures have only been made on two cruises in British Columbia, and these limited data suggest a winter minimum ranging from 6° to 7° C. The temperature range encountered in Indian Arm agrees with these. Thus Indian Arm is typical of the low runoff group of fjords. The annual temperature distribution presents a more consistent pattern than that noted for salinity. This is attributed to the small horizontal temperature gradients. When these are small, effects of such factors as wind-induced surface currents and internal waves do not confuse the seasonal temperature cycle. For this reason, the annual circulation pattern based upon temperature variations

is probably more reliable than one based upon salinity variations.

The annual dissolved oxygen pattern of Indian Arm shows an approximate unimodal curve in the upper 50 meters which is composed of two overlapping elements: spring and fall maxima. The distribution below 50 meters is determined primarily by the replacement pattern of the deep water. Values encountered during the three years of the investigation usually ranged from 150 per cent saturation during spring and fall phytoplankton blooms to 10 per cent saturation in relatively stable bottom waters. The horizontal gradients are variable, but show a tendency to increase toward the mouth in winter and a tendency to decrease toward the mouth in summer.

The radiocarbon and oxygen budget methods used to estimate net and gross production in this study are complementary. The values calculated by the oxygen method provide three estimates: the amount of organic material produced in excess of the fjord requirements; the amount of organic material utilized by respiration in the sub-euphotic waters of the fjord; and the gross fjord production. The mean values for 1958-1959 were 380, 290, and 670 g. C/m.²/yr. respectively. The C¹⁴ method is assumed to provide a measure of daylight net primary production. Adjustment of these values, using the observed ratio of respiration to net primary production, provides an estimate of the 24-hour net and gross primary production. The 1958-1959 mean fjord values were 450 and 680 g. C/m.²/yr. respectively. With the many assumptions made in arriving at these values, the close agreement between the two gross production values, i.e. 670 and

680, is probably fortuitous. However, it is believed that a gross production value of 600-700 g. C/m.²/yr. is a realistic estimate for annual production. The estimated deep-water utilization of organic material was found to be 290 g. C/m.² and the net primary production was 450 g. C/m.². The difference between these two values provides an estimate of the amount of organic material not utilized within the fjord. Since analysis of the bottom sediments indicates that very little organic material is deposited on the bottom, this difference also represents the amount of material transported out of the fjord. These calculations indicate that about 25 per cent of fjord production is exported. This is consistent with the pattern of circulation, which has a constant outflow of the shallow surface layer--a layer where most primary production occurs. A comparison of the utilization and production curves (Fig. 13) provides three interrelated cycles: an annual cycle of primary production, which has two maxima (one in spring and one in fall); a cycle of sub-euphotic utilization with two maxima (the spring maximum larger than the fall); and a net fjord production cycle with two maxima (the fall maximum being larger). The comparison of these curves indicates that the material produced within the fjord during the spring and early summer tends to be used within the fjord, whereas that produced during the fall tends to be exported.

Thus the primary factors determining the annual production cycle in Indian Arm appear to be those affecting the physical distribution of the phytoplankton population, as horizontal

movements which deplete the population, or as stability restrictions on vertical movements which establish the total light that the population will receive during a day; and as the effect of stability on the replenishment of nutrients within the euphotic zone. The effects of grazing by herbivores have been neglected in this analysis, but will contribute to the degree of development of the characteristic features of the annual production cycle.

A comparison of the annual production in Indian Arm with other similar habitats is difficult, because little data is available on the annual cycle of organic production in marine estuaries. Representative figures for a series of some estuarian and coastal habitats is presented in table IV. Data from the earlier work (Gaarder and Gran, 1927) is excluded because the long incubation periods used are difficult to interpret. Because annual production estimates are scarce, representative sets of daily production data are compared for certain coastal and estuarian habitats.

The estimated annual production of Indian Arm is higher than that reported for other estuaries. Long Island Sound, the most intensively investigated estuary (and one of the few where the annual production cycle has been established in detail), is reported to have an annual gross production of $470 \text{ g. C/m.}^2/\text{yr.}$ (Riley and Conover, 1956), a value about 30 per cent less than that estimated for Indian Arm. This disparity may result from the presence of a reserve of nutrients in the deep water basin of the fjord, a feature lacking in other types of estuaries.

It should be noted that daily production levels of Indian Arm are appreciably higher than those measured in typical coastal shelf regions off the eastern United States (Ryther and Yentsch, 1958), Denmark (Steemann Nielsen, 1958b), and Sweden (Steemann Nielsen, 1937), and are lower than the values reported from regions off the coasts of Ecuador (Holmes, 1957) and S. W. Africa (Steemann Nielsen, 1937). The latter is the highest value reported for a coastal region. Off Ecuador and S. W. Africa, the high rates of nutrient replenishment were assumed to be responsible for the production levels. This agrees with my conclusion that high nutrient levels in the deep waters of Indian Arm are transported periodically to the euphotic zone during periods of low stability.

The highest annual productions reported for any marine habitat are those of the restricted benthonic communities of the Florida Turtle Grass beds and Pacific Ocean coral reef communities, with values ranging from 2900 to 4650 g. C/m.²/yr. (see table IV).

IX. SUMMARY

1. From data gathered on 35 cruises at monthly intervals from 1956 through 1959, the distributions of conservative and non-conservative oceanographic properties in Indian Arm were determined. From an analysis of the cycles of distribution of these properties, as well as from direct current measurements, the circulation pattern and method of fjord water replacement were determined.
2. The annual cycle of primary organic production was established using three complementary methods: a net oxygen budget method--to determine the amount of organic material produced in excess of the fjord's requirements or utilized in excess of production; an oxygen-utilization method--to estimate the amount of organic material utilized in sub-euphotic waters; and a C^{14} method--to determine primary production in excess of phytoplankton respiratory requirements. The values for 1958-1959 were estimated to be 670, 290, and 460 g. C/m.²/yr. respectively. Approximately 25 per cent of the gross primary production is exported to neighbouring waters.
3. The relative contributions of the nanno- and net-plankton fractions to primary production were established, and it was shown that over the year more than 90 per cent of the primary production is attributable to the nanoplankton (organisms smaller than about 56 microns in diameter). These minute photosynthetic organisms are generally neglected in routine net surveys of the phytoplankton population.
4. The relationships between the physical environmental factors

and primary production are discussed. The primary factor controlling production is believed to be the yearly variations in the stability of the water column. This variable stability influences the mean light intensity received by the phytoplankton population, and effects the replenishment of nutrients in the euphotic zone.

5. The primary production of the ecosystem of Indian Arm is compared with that of other marine regions. It is shown that this fjord can be more productive than most oceanic or continental shelf regions, and that it approaches the level of productivity found in regions of coastal upwelling.

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Fig. 1. Oblique aerial view looking north into the mouth of Indian Arm.

Fig. 2. Plan of Indian Arm showing station positions and longitudinal and transverse bottom profiles.

LONGITUDINAL BOTTOM PROFILE

TRANSVERSE BOTTOM PROFILES

DATA FROM
C.H.O. CHART 3435

Fig. 3. Seasonal variation in the diffusion of oxygen across the air-sea interface.

Fig. 4. Seasonal variation in the net oxygen budget of the fjord.

Fig. 5. Seasonal variation in oxygen utilization within the fjord.

OXYGEN IN MG./L./MO.

Fig. 6. Daylight ^{14}C net primary production plotted as a function of depth.

Fig. 7. Comparison of the relation between gross photosynthesis and respiration
for Isefjord and Indian Arm.

RATIO OF GROSS PRODUCTION TO RESPIRATION

Fig. 8. Seasonal variation in gross and net primary production.

Fig. 9. Seasonal relation between total daily incident radiation and net primary production.

Fig. 10. Seasonal relation between the depth of compensation, the depth of 90 per cent net primary production, and the depth of 1 per cent surface radiation.

Fig. 11. Seasonal relation between runoff, water column stability and net primary production.

Fig. 12. Seasonal variation in the relation between the gross primary production-respiration ratio and temperature.

Fig. 13. Seasonal relation between the net oxygen budget, oxygen utilization, and net primary production.

Fig. 14. The longitudinal distribution of bottom sediments.

Fig. 15. Seasonal variation of salinity at station D for various depths.

Fig. 16. Vertical sections showing mean salinity for representative stations during the 1956 to 1959 spring runoff maxima.

Fig. 17. Mean longitudinal distribution of salinity during the 1956 to 1959
runoff maxima.

Fig. 18. Seasonal variation in the longitudinal distribution of surface salinity.

STATIONS

Fig. 19. Mean vertical salinity profile at selected stations during: a) 1956 to 1959 spring runoff maxima, b) 1956 to 1959 summer runoff minima, c) 1956 to 1959 winter runoff maxima, d) 1956 to 1959 winter runoff minima.

Fig. 20. Influence of the fresh water discharge from the Buntzen Power Plant on salinity: a) 0 meters, b) 2.5 meters, c) 5 meters, d) 10 meters.

OUTFALL
(X) STATION D

A. 0 meters

B. 2.5 meters

C. 5 meters

D. 10 meters

Fig. 21. Relation between mean monthly river discharge and precipitation
(1956 to 1959).

PRECIPITATION (cm./mo.)

Fig. 22. Monthly mean discharge of fresh water into Indian Arm from primary sources
(1956 to 1959)

Fig. 23. Mean seasonal cycle of temperature at various depths
(1956 to 1959).

Fig. 24. Seasonal variations of temperature at station D for various depths.

Fig. 25. Seasonal cycle of temperature at station D (1956/57).

Fig. 26. Seasonal cycle of temperature at station D (1957/1958).

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Fig. 27. Seasonal cycle of temperature at station D (1958/59).

Fig. 28. Vertical temperature profiles at station C (1956 to 1959).

Fig. 29. Seasonal variation in density at station D for various depths.

Fig. 30. Seasonal cycle of stability at station D (1956/57).

STABILITY ($E = 10^5 \frac{d\sigma_T}{dz}$) IN $10^5 M^{-1}$

Fig. 31. Seasonal cycle of stability at station D (1957/58).

STABILITY ($E = 10^5 \frac{d\sigma_r}{dz}$) IN $10^5 M^{-1}$

Fig. 32. Seasonal cycle of stability at station D (1958/1959).

STABILITY ($E=10^5 \frac{d\sigma_z}{dz}$) IN $10^5 M^{-1}$

1958 / 59

Fig. 33. Distribution of dissolved oxygen in longitudinal section through the fjord: a) February, 1959, b) March, 1959, c) May, 1959.

Fig. 34. Relative effects of the temperature gradient and salinity gradient in establishing stability: a) 1956/57, b) 1957/58, c) 1958/59.

Fig. 35. Schematic illustration of mixing between the surface and underlying waters of the narrows

S

N

Fig. 36. Distribution of dissolved oxygen in longitudinal section through the fjord
December 1957.

Fig. 37. Schematic illustration of the effects of tidal exchange between the fjord and neighbouring waters.

Fig. 38. Distribution of dissolved oxygen in longitudinal section across the fjord sill: a) ebbing tide, b) flooding tide, c) flooding tide
d) ebbing tide.

OXYGEN IN MG./L.

Fig. 39. Distribution of dissolved oxygen in longitudinal section through the fjord,
October 1959.

Fig. 40. Distribution of phosphate in longitudinal section through the fjord, March 1959.

Fig. 41. Seasonal variation in dissolved oxygen in station D for various depths.

Fig. 42. Mean seasonal cycle of dissolved oxygen at various depths at station D (1956 to 1959).

Fig. 43. Distribution of dissolved oxygen in longitudinal section through the fjord, May 1958.

Fig. 44. Distribution of dissolved oxygen in longitudinal section through the fjord, August 1959.

Fig. 45. Schematic illustration of the effect on dissolved oxygen distribution in the fjord of intruding water masses with higher dissolved oxygen content.

TABLE I

Net transport of fjord oxygen.

Date	T_o	T_i	T_r	$T_i - T_o + T_r$	T
7 Nov.	148.20	78.37	72.82	+ 2.99	+ .744
15 Dec.	270.35	145.56	126.18	+ 1.39	+ .360
5 Jan.	391.18	292.05	97.66	- 1.47	- .381
23 Feb.	269.20	202.02	51.16	-16.02	-4.15
31 Mar.	152.54	85.14	62.73	- 4.66	-1.21
11 May	104.81	47.08	34.54	-23.19	-6.01
19 June	188.33	113.81	59.70	-14.82	-3.84
20 July	101.49	74.60	27.17	-26.40	-6.84
25 Aug.	57.67	26.84	6.97	-23.87	-6.18
14 Sept.	37.64	12.43	17.09	- 8.12	-2.10

+ = infjord

- = outfjord

 T_o = mg. O_2 /sec. transported out by outflowing waters T_i = mg. O_2 /sec. introduced by inflowing waters T_r = mg. O_2 /sec. introduced by freshwater runoffT = total net transport in mg. O_2 /month $\times 10^6$

1958/59

Net oxygen budget for fjord.

Date	O ₁	t	D ₁	D ₂	T ₁	T ₂	O ₂
7 Nov.	+ 6.379	1.2	-12.54	+15.06	+ .567	+ .680	- 8.00
15 Dec.	+ 7.146	.7	+14.00	+ 9.80	- .0105	- .0075	- 2.64
5 Jan.	+14.698	1.6	+15.50	+24.80	-2.266	-3.626	- 6.47
23 Feb.	+ 7.432	1.2	+ 7.78	+ 9.34	-2.680	-3.610	+ 1.31
31 Mar.	+ 5.109	1.4	+14.51	-20.31	-3.610	-5.054	+29.86
11 May	- 7.448	1.3	-31.80	-41.34	-4.925	-6.403	+40.30
19 June	- 3.340	1.0	-35.68	-35.68	-5.340	-5.340	+37.68
20 July	-14.128	1.2	-60.58	-72.70	-6.510	-7.812	+66.38
25 Aug.	- 3.919	.7	-63.19	-44.23	-4.14	-2.898	+44.45
14 Sept.	- 4.040	1.2	-24.98	-29.98	-1.77	-2.12	+28.06
20 Oct.							

O₁ = change in total fjord oxygen (mg. O₂ x 10¹¹)

t = time in months between observations

D₁ = mean monthly diffusion (mg. O₂ x 10¹¹)D₂ = monthly diffusion correction for time between observationsT₁ = mean monthly net transport (mg. O₂ x 10¹¹)T₂ = monthly transport corrected for time between observationsO₂ = net monthly transport in total oxygen due to biological activity (mg. O₂ x 10¹¹)

TABLE III

In situ measurements of primary production corrected for relative production potential of fjord regions
and adjusted by area to provide mean production value for Indian Arm.

Date	Sta.A	P_o Sta.C	Sta.E	Sta.A	P_e Sta.C	Sta.D	Sta.E	Sta.A	P_s Sta.C	Sta.E	Sta.A	P_a Sta.C	Sta.E	P
5 Nov.						295								
6 Nov.						477								
$\bar{x}$						386								386
16 Dec.						34								34
6 Jan.									29					
7 Jan.									18					
$\bar{x}$									23					23
24 Feb.	.05	.08	.20		597			370	597	1491	105	374	1386	
25 Feb.					265			165	265	663	47	166	616	
$\bar{x}$											76	270	1001	732
1 Apr.	.60	.33	.58		226			412	226	398	117	142	370	
2 Apr.					1211			2203	1211	2131	628	759	1979	
$\bar{x}$											373	451	1174	1086
12 May	.78	4.12	3.88				6179	1236	6550	6179	352	4106	5740	
13 May					2810			531	2810	2644	151	1762	2456	
13 May					3713			702	3713	3494	200	2328	3246	
$\bar{x}$											234	2729	3810	3680

TABLE III (cont'd.)

Date	Sta.A	P_o Sta.C	Sta.E	Sta.A	P_c Sta.C	Sta.D	Sta.E	Sta.A	P_s Sta.C	Sta.E	Sta.A	P_a Sta.C	Sta.E	P
16 June	1.11	1.34	.90		2665			2201	2665	1791	629	1671	1664	
17 June							455	560	678	455	160	425	423	
18 June				1961				1961	2372	1590	559	1488	1477	
$\bar{x}$											449	1193	1189	1538
21 July	.45	.53	.58		883			716	883	963	204	554	894	898
17 Sept.	3.03	2.63	.66		1870			2150	1870	469	613	1172	436	
18 Sept.						2282		10475	9083	2282	2985	5695	2120	
$\bar{x}$											1799	3434	1278	3538
21.Oct.	4.70	2.80	1.35		654			1098	654	315	313	410	293	552

P_o = relative production (mg. O_2 /l./24 hrs.)

P_c = C^{14} in situ measurement of primary production (mg. C/day $\times 10^7$)

P_s = estimated in situ primary production (mg. C/day $\times 10^7$)

P = mean fjord primary production (mg. C/m.²/day)

TABLE IV

Selected examples of productivity measurements

Locality	Time	g.C/m. ² /yr.	mg. C /m. ² /day	Method	Reference
Indian Arm		670		O ₂ budget (gross)	author
Indian Arm		450		C ¹⁴ (net)	author
Long Island Sound		470		O ₂ and PO ₄ budget (gross)	Riley and Conover, 1956
Isefjord		175		O ₂ L-D incubation (gross)	Steemann Neilsen, 1951
Puget Sound		30-111		PO ₄ budget	Waldichuck and Gould
Trevor Channel		52-62		O ₂ and PO ₄ budget	Waldichuck, 1956
Georges Bank		309			Riley <i>et al.</i> , 1949
Gulf of Maine		270		PO ₄ budget (gross)	Riley, 1941
Indian Arm			24-3700	C ¹⁴ (net)	author
English Channel	June	500	500	C ¹⁴ (net)	Steemann Neilsen, 1954
Ecuador coastal	Autumn		500-1000		Holmes <i>et al.</i> , 1957
S. W. Africa coastal	Autumn		500-4000	C ¹⁴ (net)	Steemann Neilsen, 1954
N. Y. Coastal	April		250	Chlor. A and light (gross)	Ryther and Yentsch
Kattegat	Oct.		250	O ₂ L-D incubation (gross)	Steemann Neilsen, 1957
	Dec.		10		
Danish coastal	Mar.		300	C ¹⁴ (net)	Steemann Neilsen, 1957
	Aug.		700		
Florida (turtle grass bed)			4650	O ₂ budget (gross)	Odum, 1956
Marshall Is. (coral reef)			3500	O ₂ budget (gross)	Odum and Odum, 1955
Hawaiian Is. (coral reef)			2900	O ₂ budget (gross)	Kohn and Helfrich, 1957

TABLE V

Calculated tidal current velocities for
various tidal amplitudes and stages.

Tidal Amplitude (meters)	Tidal stage			
	1/8 flood	1/4 flood	3/8 flood	1/2 flood
2	10 cm./sec.	17	23	25
3	14	26	35	37
4	19	35	47	50

TABLE VI

Direction of horizontal temperature gradient at various depths.

Year	Depth (m.)	Month											
		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.
1956	5									-	+	+	≠
1957	5	-			-	+	-		≠	-	≠	≠	=
1958	5	+	+	+	+	≠	+	-	-			≠	≠
1959	5	≠	≠	≠		≠	≠	-	-		≠		
1956	10									-	+	+	≠
1957	10	≠			-	+	-		-	-	≠	≠	≠
1958	10	+	+	+	≠	-	+	-	-			≠	+
1959	10	+	+	+		≠	≠	-	-		≠		
1956	30						≠						
1957	30						≠		-	-	≠	≠	≠
1958	30	+	+	+	+	≠	+	-	-			+	+
1959	30	+	+	+		≠	-	-	-		≠		
1956	50									-	≠	+	≠
1957	50	+			-	-	+		-	-	≠	-	≠
1958	50	+	+	+	+	≠	≠	-	-			≠	≠
1959	50	+	+	+		≠	≠	≠	-		-		
1956	100											+	+
1957	100	-		-	-		-		-	-	≠	-	-
1958	100	-	+	-	≠	≠	≠	-	+			≠	≠
1959	100	≠	+	+		+	+	-	-		-		

- = decreasing temperature trend mouth to head

+ = increasing temperature trend mouth to head

≠ = no consistent temperature trend

TABLE VIII

Numerical values for eddy coefficient derived from
observed variations of temperature and salinity

Indian Arm					Bute Inlet ¹			
Date	Depth (m.)	S	A _t	A _s	Date	Depth (m.)	S	A _s
June-Aug. 1957	150	210	0	3.13	Aug. '50	110	500	.58
Feb.-Mar. 1958	150	220	.71	2.68	Sept. '50	110	300	1.25
Mar.-Apr. 1958	150	350	2.29	0	May '51	90	700	.23
Apr.-May 1958	150	210	1.37	17.15	Aug. '51	90	1000	.63
Aug.-Sept. 1958	150	220	1.04	3.62	Oct. '51	130	500	.55
Sept.-Nov. 1958	150	230	.92	.45	Aug. '52	110	400	.21
Nov.-Dec. 1958	150	300	1.03	1.12				
Dec.-Jan. 1958	150	370	.61	2.13				
Jul.-Aug. 1959	150	180	9.31	6.82				

S = stability ($10^5 \frac{d\delta t}{dz}$)

A_t = eddy coefficient derived from temperature
variations (g./cm./sec.)

A_s = eddy coefficient derived from salinity
variations (g./cm./sec.)

¹ Tabata and Pickard, 1957

TABLE VII

Seasonal variation in the horizontal extent and depth of intrusions of outside waters.

Year	Property		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1956	O ₂	northward extent									sta.D	sta.A	sta.A	sta.F
1956	O ₂	depth range									50	150	50	50
1956	T	northward extent									sta.E	sta.E		sta.F
1956	T	depth range									30-50	150		50
1957	O ₂	northward extent	sta.D			sta.D		sta.C			sta.C sta.F	sta.D	sta.D	sta.C
1957	O ₂	depth range	50-200			100-150		75			10 30-50	30-50	10	50
1957	T	northward extent												
1957	T	depth range												
1958	O ₂	northward extent	sta.E	sta.E	sta.D	sta.D	sta.E			sta.F			sta.E	sta.E
1958	O ₂	depth range	30-75	30-50	30-75	30-75	50-75			30			10-50	10-75
1958	T	northward extent	sta.E	sta.D		sta.D	sta.F	sta.E	sta.E	sta.F			sta.E	sta.E
1958	T	depth range	30-75	30-50		30-50	20-30	30	10-30	10-30			10-50	10-100

TABLE VII (cont'd.)

Seasonal variation in the horizontal extent and depth of intrusions of outside waters

Year	Property		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1959	O ₂	northward extent	sta.D	sta.B	sta.D		sta.D		sta.E	sta.C		sta.C		
1959	O ₂	depth range	10-50	30-100	30-100		75-100		10-30	10		30-50		
1959	T	northward extent	sta.F						sta.F	sta.D	sta.F	sta.E		
1959	T	depth range	10-30						10-50	30-50	10-30	30-50

a. Prior to Intrusion

b. During Intrusion

**c. Complete Vertical
Mixing**

d. Intrusion with Partial Mixing

1956 / 59

Aug 41

S

N

STABILITY ($E = 10^5 \frac{d\sigma_z}{dz}$) IN $10^5 M^{-1}$

STABILITY ($E = 10^5 \frac{d\sigma_T}{dz}$) IN $10^5 M^{-1}$

Fig 17

STATIONS

East Slope

Center

West Slope

 Terrestrial Plant
 Material

1/2/64

19063

Lytle

Fig 1

RATIO OF GROSS PRODUCTION TO RESPIRATION

565
1958