A STUDY OF THE HEAT BUDGET COMPONENTS FOR THE BRITISH COLUMBIA AND S.E. ALASKA COAST

by

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ABSTRACT

Knowledge of the surface heat transfer in coastal inlets would permit studies of their thermal structure and circulation. An assessment is made of data available for calculating the surface heat transfer for the coastal regions of British Columbia and S.E. Alaska.

Monthly means of meteorological and oceanographic observations for the years 1961 and 1963 are critically examined for their representativness of conditions that exist over the open water. The location of the observation point is found to be important in choosing values for dew point and wind speed.

Formulae for calculating surface heat transfer are examined for their potential applicability to a coastal climate.

The calculated net annual surface heat transfer is found to be highest in the southern regions, approximately 90 langleys/day in the Strait of Georgia, and to decrease for more northerly regions, to an approximate balance with no net input in northern Chatham Strait.

The annual cycle is found to be strongly modified by fine structure, the radiation balance dominating in summer, the convective losses in winter. Comparison of the calculated surface heat transfer with heat storage indicates that the calculations may be accurate to within 20% of the peak values.

The range and shape of the surface temperature cycle was found to reflect the influence of advection, and deep water temperature, as well as the surface heat transfer.

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INTRODUCTION

The physical oceanography of the deep inlets along the coast of British Columbia and S.E. Alaska is concerned with water structure and circulation. The structure (temperature, salinity and oxygen) of the whole inlet from surface to bottom can be determined relatively easily in a few hours, but to determine the circulation requires many days to obtain an idea of flow at one location, and the entire circulation would require weeks.

Pickard and Trites (1957) showed how the transport in an inlet could be estimated knowing the surface heat exchange and the temperature structure ; therefore there has been a growing interest in heat budget studies to permit at least some estimate of the in and out transport.

There is good evidence that estuarine circulation occurs in most inlets, there being a net outward transport of the upper layer driven by the river runoff and a subsurface net inward transport to replace the salt water removed by entrainment into the upper layer. Since the heat supply or loss of an inlet is essentially (1) by transfer across the air - sea interface, and (2) by horizontal advection, a knowledge of the surface transfer and the temperature structure of the inflow and outflow would give some information about the advection itself. Therefore knowledge of the heat budget would yield directly an explanation of the thermal structure and indirectly information of the circulation.

Waldichuk (1957) in his study of the Strait of Georgia concluded that the Strait was a source of heat to the Pacific Ocean. Tabata's (1958) work in Dixon Entrance gave a calculated net loss of about 25% or 24 langleys/day from the sea surface.

The purpose of this study is to assess the quality of existing meteorological and oceanographic observations available for the coastal area in terms of using them to calculate the surface heat transfer, and to determine the broad latitudinal features of the surface exchange that may be evaluated from these land based observations.

Shown in Fig. 1 A and B is the section of the coast studied. It extends from Juan de Fuca Strait in southern British Columbia to Skagway in the northern part of S.E. Alaska. The coastal climate of this area is dominated by marine weather systems from the N.E. Pacific, though occasionally continental arctic air flows down the many inlets that cut into the coast range bringing a cold relatively dry air to the coast. Since the rugged coastal topography presents a wide range of exposures, an estimate of the degree of modification to passing weather systems is required in order to evaluate the importance of local effects, and thus how far the existing data may be extrapolated.

Nine regions were chosen along the coast, Fig. 1 A and B, to represent the major environmental and climatic conditions for which the existing data appeared applicable. Due to the limited time available, the study was restricted to two years.

After a preliminary assessment of the data, 1961 and 1963 were chosen as having typical climatic conditions. Since published material is normally available as monthly means, the averaging error could be minimized by carrying out the calculations for each month of the year. As the meteorological variables required to calculate the sea surface heat transfer are influenced by the land surface over which the observations were taken, the data for these two years were considered as a function of exposure. What appeared to be the most representative data were used to evaluate the surface heat transfer for each region.

The calculated surface heat transfer was used to interpret some of the measured temperature structure of the coastal region.

HEAT BUDGET THEORY

Net Balance

The simplified equation for balance of heat energy in a column of unit plan area, extending from the sea surface to the sea floor, shown schematically in Fig. 2, can be written as,

$$Q_s (1-r) + Q_b + Q_e + Q_h + Q_v = Q_\theta \qquad \ldots\ldots(1)$$

where Q_s = incident short wave (solar) radiation received on a horizontal surface at sea level, both direct and diffuse,

r = fraction of short wave radiation reflected by the sea surface,

Q_b = net heat transfer by long wave radiation from the sea and the atmosphere,

Q_e = heat transfer by evaporation and condensation of water vapour,

Q_h = net transfer of heat by sensible heat conduction across the air - sea interface,

Q_v = net gain of heat in the water column by advection, and

Q_θ = heat used for local heating of the water column.

In numerical calculations, the unit used for Q will be 1 langley/day (1 gm. cal./sq. cm. day), while a positive sign will indicate flow into the column, and vice versa.

This neglects changes by 1) conduction from the earth's interior, 2) heating by chemical or radioactive processes, 3) dissipation of kinetic energy, or 4) precipitation.

The net exchange of heat by radiation may be lumped as:

$$R = Q_s (1-r) + Q_b \qquad \text{...........(2)}$$

and the heat transfer by convective processes above the sea surface as

$$C = Q_e + Q_h \qquad \text{...........(3)}$$

Then

$$Q_D = R + C \qquad \text{...........(4)}$$

is the heat energy available for transfer downward in the column, from surface to deeper layers.

Short Wave Radiation

After entering the atmosphere, extraterrestrial radiation is subjected to scattering by the air molecules and contaminants, is absorbed by air molecules, mainly water vapour, and is reflected and absorbed by clouds. Upon reaching the sea surface there is a partial loss by reflection.

To simplify evaluation of these effects, the atmospheric turbidity and cloud cover can be considered separately. The value of incident solar radiation that is received on a horizontal surface on a given clear day is taken as a constant.

This assumes that the air mass, and total precipitable water vapour and contaminants along the radiation path for that day are constant. The assumption should be a good approximation for the coastal region, for only occasionally is there any large amount of dust or smoke in the atmosphere.

Kimball (1928), Mateer (1955a), and Budyko (1956) have calculated values of Q_o for areas including the coastal region. Mateer (1955a) presented, contoured for each month, the mid-monthly value of Q_o for Canada as could be derived from the then existing data. The only radiation measurements on the Pacific Coast available for his work were from Friday Harbour, Washington. For the present study, data was available from 1) Climatological Station, University of British Columbia, Vancouver, and 2) Annette, Alaska. Daily values of solar radiation measured at these two stations were plotted for all available years. The larger values in a month may usually be taken to represent the solar radiation received on a totally clear day. Occasionally there may be reflection from clouds that locally increases the measured value above a clear day value, and a smooth curve drawn just below the highest points is a good estimate of cloudless day radiation Q_o (Fritz, 1949). An example is shown in Fig. 3.

The calibration of the pyrheliometer at the University of British Columbia between January 1959 and April 1960 was approximately 15% lower than in more recent years, the change occurring at the installation of the present Eppley pyrheliometer. The recent results are taken as correct.

Mid-monthly values of Q_o at the University of British Columbia obtained by this method were sufficiently consistent from year to year to suggest an accuracy of approximately $\pm$ 1%.

The mid-monthly values of Q_o from Annette were more scattered, and the mean value could have an error of $\pm$ 5%.

The mid-monthly means for the University of British Columbia and Annette are compared in Table I with values given by Mateer (1955a) and Kimball (1928).

Subsequent to Mateer's (1955a) work a change of calibration standard to the new Pyrheliometer Scale of 1956 makes Mateer's values 2% higher than present day measurements. The values given by Mateer and those calculated for the University of British Columbia are different by this magnitude over most of the year. Kimball's values are approximately 20% higher.

The value of Q_o for the coast was taken as the distribution given by Mateer, corrected to fit the values estimated for the University of British Columbia. As the values used as typical for the region of Annette are generally lower than the cloudless day radiation calculated for Annette, it is likely that the Q_o for northern regions is low.

Table II gives the values of Q_o used for this study. They are for the 15th day of the month, at a position centered on the respective regions. The values are estimated to be accurate to $\pm$ 5%.

The climatological station at the University of British Columbia records simultaneously hours of sunshine and solar radiation. From this data, a linear relationship between relative global radiation (the ratio of incident solar radiation to cloudless day radiation) and the monthly mean of percent possible sunshine was obtained by a least squares fit. The curve, as shown in Fig. 4, is:

$$\frac{Q_s}{Q_o} = 0.35 + 0.0070S \qquad \text{...........(5)}$$

where Q_s = average incident solar radiation for the month,

Q_o = cloudless day solar radiation on 15th of the month,

and S = percent possible sunshine for the month,

with radiation in langleys/day.

The intercept at $S = 0$ and the slope of curves drawn through values for a given month were noticed to vary with season, the intercept lower and slope higher in winter. Because of inadequate data the effect was not evaluated. The variation is small, as evidenced by the scatter in points at large cloud cover.

The equation is compared with similar formulae by Fritz and MacDonald (1949), Mateer (1955b), and Kimball (1927) in Fig. 5. The Fritz and MacDonald equation was corrected to allow for the difference in observation method. The close fit to results from other areas suggests its validity for the coastal range under study. The sunshine measurements in Canada are taken with a Campbell-Stokes sunshine recorder which has a threshold of 0.1 to 0.4 langleys/min. therefore at high latitudes it may be necessary to correct for the low sun angle (Mateer, 1955b).

Observations of hours of sunshine and amount of cloud cover at the Vancouver and Victoria airports were used to obtain a relationship between percent possible sunshine and cloud cover. One theoretical point was used at zero cloud cover. The data, fitted to a quadratic, yielded:

$$S = 97 - 2.6C - 0.79\,C^2 \qquad \text{...........(6)}$$

where S = percent possible sunshine for a month, and

C = monthly mean cloud cover on a scale 1 to 10, in which 10 represents complete overcast.

A combination of the formulae (5) and (6) above yields:

$$\frac{Q_s}{Q_o} = 1.03 - 0.018C - 0.0055\,C^2 \qquad \text{...........(7)}$$

with the same definitions. This formula is compared with similar relationships in Fig. 6.

Recent studies have shown that the type of cloud is important in evaluating insolation. Fritz (1954) reports a variation in absorption of radiation with cloud density and thickness. Haurwitz (1948) graded clouds into classes from low stratus and fog of small transmission, 20% - 35%, to high cirrus of large transmission, 80%.

Cloud albedo is thought to vary with solar zenith angle, and with cloud thickness and type, increasing with an increase in thickness of the same type of cloud. The variation with zenith angle is probably the most important.

These variations are suggested reasons for the wide range of cloud cover correction formulae developed. Shown in Fig. 6 is a comparison of cloud cover formula.

Ashburn (1963) and Tabata (1964) used data from Ocean Station 'P', a weather ship located in the north Pacific at nominal latitude 50°N, longitude 143°W. It is expected thatregions such as Station 'P', with a greater than average amount of low thick cloud, would have a lower insolation for a given cloud cover due to higher reflection and absorption. This condition appears to exist between the formula developed at Ocean Station 'P' and that developed for the lower mainland of British Columbia.

Linear formulaedo not appear to be accurate for the entire range of cloud cover.

The values of water albedo used were taken from a linear interpolation of a table given by Budyko (1956). Burt (1954) considered the variation of water albedo with wind speed and surface conditions to be small compared to the effect of solar zenith angle.

<u>Long Wave Radiation</u>

The radiation emitted by the sea is close to that of a black body, and described by the equation:

$$Q_r = S\sigma\Theta_s^4 \qquad \text{...........(8)}$$

where Q_r = the energy emitted in langleys/day,

S = the emissivity, 0.970 after Anderson (1954),

σ = the Stefan-Boltzmann constant, 1.171×10^{-7} langleys/day $^{o}K^{4}$,

and

Θ_s = the absolute temperature of the surface.

The counter radiation by the atmosphere is a function of total water vapour content, air temperature, and cloud cover and type.

In Fig. 7 four different methods used for calculating the effective back radiation, Q_b, are compared, using data from region 1, 1963.

They are:

1) Anderson (1954)

$$Q_b = Q_r\left[1 - R^4(0.740 + 0.025\, C\alpha + 0.00490\, e - 0.00054 C\beta e)\right] \qquad \text{(9)}$$

2) Brunt (Sverdrup et al, 1942, p. 112)

$$Q_b = Q_r\left[1 - (0.48 + 0.046C + 0.08\sqrt{e} - 0.0068C\sqrt{e})\right] \qquad \text{(10)}$$

3) Budyko (1956)

$$Q_b = Q_r\left[1 - (0.61 + 0.0039nC^2 + 0.067\sqrt{e} - 0.00067nC^2\sqrt{e}) + 4R^4\left(\frac{I}{R} - 1\right)\right] \qquad \text{(11)}$$

4) and Sverdrup's method given in Sverdrup _et al_ (1942) p. 111 - 112.

where

Q_b = effective back radiation in langleys/day,

Q_r = energy emitted by the sea in langleys/day,

R = ratio of air temperature to sea surface temperature in ^{o}K,

C = cloud cover in scale 1 to 10, and

e = vapour pressure in mb.

The Budyko and the Anderson values follow each other closely. Brunt's equation does not allow for the changing temperature of the air and this has caused it to be relatively constant through the year. Sverdrup's method, almost 40% higher in summer than other results, obtains its general shape from the cloud cover cycle.

As Anderson's formula is the most complete, and considers the back radiation on a proper basis, it is used in this study. The formula was developed by considering the counter radiation of the air as a function of cloud amount, cloud height, vapour pressure, and air temperature. The terms are summarized by the empirical formula:

$$\left.\begin{aligned} \frac{Q_a}{\sigma\Theta_a^4} &= a + b\,e \\ \text{where}\quad a &= 0.740 + 0.025\,C \exp(-0.0584h) \\ b &= 0.00490 - 0.00054\,C \exp(-0.060h) \\ \text{and}\quad & 1600 \le h \le \infty \end{aligned}\right\} \quad \ldots\ldots(12)$$

Q_a = the atmospheric radiation in langleys/day,

Θ_a = the absolute air temperature, e = the vapour pressure at 2 meters height,

σ = the Stefan-Boltzmann's constant,

1.171×10^{-7} langleys/day $^{o}K^{4}$, and

h = the cloud height in thousands of feet.

A curve of cloud cover vs cloud height is given in Fig. 8. Values were used from all stations in coastal British Columbia and S.E. Alaska recording both variables. Cloud height was taken from this curve.

There was a noticeable trend in the curve of increasing cloud height inland. This is not shown in Fig. 8.

As Anderson's formula was developed for region with a continental climate, two assumptions of its applicability to the coastal climate could introduce error. Firstly the vapour pressure measured in a coastal region, such as the one being studied, would not represent the higher total water content normally found in the coastal climate. Secondly, the type of cloud and cloud height are important. The coastal region with a higher frequency of stratus and alto stratus would tend to give higher atmospheric counter radiation for a given cloud cover and cloud height. Both of these differences would reduce the actual Q_b.

The comparison of different formulae in Fig. 7 indicates that the possible overestimation of Q_b from these assumptions is evident only in Budyko's formula which gives lower results during parts of the summer. The difference is not large, and thus such errors are taken to be small.

Evaporation and Sensible Heat Conduction

Semi-empirical equations with bulk transfer coefficients have been developed to describe energy transfer by sensible heat conduction and evaporation upon substitution of mean values of meteorological observations. The basis for the formulae is the concept of an eddy motion, which transports the property in a manner analogous to molecular transfer. A bulk or eddy transfer coefficient together with the gradient of the property is taken to describe quantitatively the transfer of the property by eddy motion. Although this is admitted to be an empirical approach, our poor understanding of the physical details of the transfer processes at the present time leave us little choice but to use it. For most heat budget studies where direct measurement is not possible this form of equation is used; however the accuracy of using climatological means with this type of formula is still doubtful. Due to the present lack of a better method of estimating the evaporation and sensible heat conduction, formulae of this form will be used in this study.

The equation for conservation of a property 'S' in air flowing over a water surface may be written as,

$$\frac{\partial(\rho S)}{\partial t} = \frac{\partial}{\partial x}\left(\frac{Ax}{\rho}\frac{\partial(\rho S)}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{Ay}{\rho}\frac{\partial(\rho S)}{\partial y}\right) + \frac{\partial}{\partial z}\left(\frac{Az}{\rho}\frac{\partial(\rho S)}{\partial z}\right) - \frac{\partial}{\partial x}(\rho S U_x) - \frac{\partial}{\partial y}(\rho S U_y) - \frac{\partial}{\partial z}(\rho S U_z) \qquad \text{.........}(13)$$

where $\frac{Ai}{\rho}$ is the kinematic eddy diffusion coefficient of S, and Ui the velocity component, in the i direction.

Under the assumptions of:

1) steady state,

2) unidirectional mean motion in a horizontal direction,

3) negligible horizontal diffusion,

4) no down stream advection, and

5) ρ , the density, being constant,

the equation may be reduced to:

$$F = Az \frac{d\bar{S}}{dZ} \qquad \ldots\ldots\ldots\ldots(14)$$

where F, independent of height, is the vertical flux of heat, water vapour, or momentum when $\bar{S}$ represents temperature, specific humidity or wind speed respectively.

These assumptions may be an over-simplification for some water bodies in the coastal region where shore effects and advection could be important.

Dimensional analysis of the above equation shows that over a solid boundary in the absence of buoyancy forces, $\bar{S}$ assumes a logarithmic profile. This is supported by observations of wind profiles taken over land (Sheppard, 1958). Some measurements over water indicate that a similar form for the wind may exist under conditions of neutral stability (Deacon and Webb, 1963).

'A', the diffusion coefficient, is not constant but is a function of height, stability of the air column, and the nature of the sea surface. An attempt to evaluate 'A_z' has led to various coefficients for the bulk formula.

Following the method of Montgomery, we can define coefficients:

$$\Gamma_M = \frac{1}{U_a} \frac{\partial u}{\partial \ln Z}$$

$$\Gamma_E = - \frac{1}{q_s - q_a} \frac{\partial q}{\partial \ln Z}$$

$$\text{and } \Gamma_H = - \frac{1}{T_s - T_a} \frac{\partial T}{\partial \ln Z} \qquad \text{...........(15)}$$

where the subscripts M , E, and H for momentum, water vapour and sensible heat respectively. Ua is the wind velocity at height a, q is specific humidity, and T air temperature at (1) sea surface, 's', and (2) height 'a'.

If the logarithmic profile theory holds over water, then

$$A_z = \rho k^2 \Gamma_M \; Z \; Ua \qquad \text{...........(16)}$$

where k is von Karman's constant.

Assuming that A_z, the diffusion coefficient, is the same for all transfers, implies that $\Gamma_M = \Gamma_E = \Gamma_H$. This choice is on uncertain grounds, due to the dependence upon stability effects, and boundary conditions.

Substitution gives:

$$\tau = C_D \rho U_a{}^2 = k^2 \Gamma_M^2 Z \, U_a{}^2 \qquad \text{...........(17)}$$

$$Q_E = k^2 \Gamma_M \Gamma_E \rho L (q_s - q_a) \, U_a \qquad \text{...........(18)}$$

$$\text{and } Q_H = k^2 \Gamma_M \Gamma_H \rho C_p (T_s - T_a) \, U_a \qquad \text{...........(19)}$$

τ is surface shear stress, C_D is drag coefficient, C_p is the specific heat of air = 0.24 cal/g m. C^o, and L the latent heat of evaporation = 590 cal/gm.

It is from measurements of C_D that a value for Γ_M is obtained and thus the coefficients of Q_E and Q_H.

Deacon and Webb (1963) estimated the drag coefficient at 10 metres to be:

$$C_D(10) = (1.00 + 0.07\ U_{10})\ X10^{-3} \quad \ldots\ldots\ldots\ldots(20)$$

Using a mean wind speed of 4.5 metres/sec. and substituting into equations 18 and 19,

$$Q_E = 5.05\ (e_o - e_{10})\ U_{20} \quad \ldots\ldots\ldots\ldots(21)$$

$$Q_H = 3.32\ (T_o - T_{10})\ U_{20} \quad \ldots\ldots\ldots\ldots(22)$$

where

Q_E = heat transfer away from sea surface by evaporation, langleys/day.

Q_H = heat transfer away from sea surface by sensible heat condition, langleys/day.

e_a = vapour pressure in mb at height 'a' metres,

Ta = air temperature in C^o at height 'a' metres, and

U_{20} = wind speed in metres/sec. at 20 metres height.

The heights 'a' refer to elevation above sea surface with 0 at sea surface. The height correction that is necessary to reduce the wind speed from the 20 metre level to the 10 metre height for which C_D has been evaluated has been incorporated into the coefficients given in equations 21 and 22.

Use of these bulk equations assumes that averaging is unimportant. That is,

$$\overline{(X_o - X_{10})\ U_{20}} = (\ \overline{X_o} - \overline{X_{10}}\ )\ \overline{U_{20}} \quad \ldots\ldots\ldots\ldots(23)$$

This could introduce significant errors, since cold dry continental arctic air may enter the coastal region for short periods with a resulting large convective loss hidden by a monthly average of humidity and air temperature.

CHOICE OF DATA FOR ANALYSIS

General

It was decided to carry out the analysis for each month of the two years, 1961 and 1963. The monthly calculation would give the lowest averaging error with the data available. The choice of 1961 and 1963 was made in an attempt to show two typical years.

Where possible, the monthly means of cloud cover, wind speed, air temperature, dew point, and sea surface temperature were taken from published material. Table III lists the sources and data used from each. The use of dew point rather than relative humidity to give water vapour content enables more accurate interpolation since dew point is independent of air temperature.

For comparison, the long term mean of each meteorological parameter was taken from Tables given by Kendrew and Kerr (1955).

Fig. 1 gives the main distribution of meteorological observation points used in the analysis. Most are operated by meteorological departments of Canada and the United States to provide information for weather forecasting. Also given in Fig. 1 are the position of stations where sea water temperature and salinity are observed.

After a preliminary assessment of the data, nine surface regions were chosen to represent the coastal area. These are shown in Fig. 1, marked by the shaded areas. They do not define a specific water body, but rather designate the broad surface area over some weakly defined limit to which the analysis is directed. For example, Region 1 is the area in the Strait of Georgia south of Vancouver, and Region 5 the eastern section of Dixon Entrance. The nine regions are considered as sufficient to represent the geographical distribution of the coast.

In choosing the required data for each region an attempt was made to compile the available land-based observations surrounding each region and extrapolate to "over water", weighting by geographic position and exposure. The importance of vertical and horizontal changes are considered separately.

Variation of the Vertical Profile

The observations at the shore stations are taken at a range of heights as well as exposures. Since it is assumed that the logarithmic profile is typical for the variables over water, some estimate must be made of the modification to this profile over the land surface in order to choose a representative value of wind speed at 20 metres and air temperature and dew point at 10 metres above the water surface. To aid in the assessment, the environment of the measurement can be put into one of two classes judged by the susceptibility of local modification to offshore conditions.

(1) Class 1 Stations

These are observation points which have good exposure to (offshore) air flow. An example is a lighthouse on a small rocky island or peninsula. Air flowing from the sea to the observation point is presumed not to have altered significantly, other than through increased mixing by turbulence generated along the shore.

Observations taken in a Stevenson Screen, approximately one metre above the ground would then be of an air mixture that is representative for some level below the mean sea level height of the screen since the air will tend to flow up over the land surface. The measurements taken at higher levels, typically twice the ground elevation, are thought to be high enough to be above the ground effect, and can be considered representative for that height. From hydro-dynamics, calculation of the modification to a mean flow field by a blunt object shows that increased flow over the object may be important in cases where the object height is comparable with the observation height, such as at Cape St. James.

(2) Class 11 Stations

Most of the observation points put in Class 11 are at airports, often more than a mile from the shore. They are referred to as poor exposure, in that the profile of the air flow has had time to change from its "over water" form. That is, the wind profile assumes a shape typical for that surface, dew point would be expected to drop through condensation and increased mixing, and the temperature will be modified depending on the surface encountered.

An example of the difference due to exposure of the station can be seen in Fig. 9. In this it shows the annual cycle of an average of differences in monthly mean dew point and air temperature between Cape Lazo and Comox airport. At both stations the observations were taken in a Stevenson Screen on top of a plateau approximately 125 feet above mean sea level in the Strait of Georgia. Cape Lazo station is more exposed to direct advection, situated on a point projecting slightly out into the strait, while the Comox observations were taken at approximately one mile from the shore. The difference in mean air temperature is small, Comox having a slightly larger range from winter to summer. The effect on dew point is more pronounced, Comox having lower values all year. The difference increases during the summer months when the frequency of winds from the northwest is highest, and air is flowing over longer stretches of land.

Using this classification the observation points were put into two groups, as given in Table lV. Also in Table lV are the observation heights for each station. The actual range of heights for wind observations at Class 1 stations is between 15 and 40 metres. Required are values for the 20 metre level. To use a wind speed representative for 15 or 40 metres as representative for the 20 metre level would result in an error of 2% and 6% respectively under the assumption of a logarthmic profile in neutral stability.

The "over water" height for which ground level observations of temperature and dew point are representative is uncertain. If the observations are representative for a range from 5 metres to 25 metres, and used for the 10 metre value, errors of 12% and 8% respectively would result.

As will be shown (Fig. 16 and 18), the monthly mean air temperature and dew point can have a consistent trend along the outer coast regardless of height of observation. Because of the apparent lack of variation with height in the climatological means, no height corrections were made to the observations at Class 1 stations. The 10 metre and 20 metre heights used are a compromise of existing observation levels. No simultaneous observations at two different levels have been made at a Class 1 station. Information on the difference in climatological mean of temperature and dew point between say, ground and the top of a lighthouse, where sea surface temperature observations are taken would be a useful check.

Where coverage was adequate, only Class 1 stations were used. Such a choice was possible for all regions except in the Strait of Georgia, where Merry Island is the only example of good exposure for air temperature and dew point.

Horizontal Distribution of Meteorological and Oceanographic Variables

Each variable was plotted on a worksheet to show the geographical distribution. The regional mean was then evaluated on a basis of quality of the surrounding shore data. The geographical chart is represented in this report by a 'scatter diagram' which reduces the data to a two dimensional picture. Solid dots are used for Class 1 stations, open dots for Class 11.

(1) Cloud Cover

A comparison of the average monthly cloud cover, as estimated from 24 observations used by Canadian stations, with the average from the four 'synoptic' observations shows good agreement. These synoptic observations are taken to be representative of mean conditions for one-quarter of the day centred on the observation time. Values at 1000 and 1600 PST are used as representative of cloud cover during daylight hours. In Fig. 10, the monthly averages of the diurnal change of cloud cover at two representative stations along the coast show that there is a marked diurnal variation for the southern stations, there being up to one-tenth more cloud during the daylight period. The diurnal variation decreases progressively through the more northern stations to a typical maximum of two hundredths in S.E. Alaska. For regions south of Alaska the cloud cover was used in two parts. One was a daytime average as required for insolation calculations. The other was an all day average to estimate effective back radiation.

Fig. 11 gives a scatter diagram of daylight cloud cover for stations used in the assessment of each region. The distribution is consistent along the entire coast, in that the trend is evident from more than one station. That is, the fine structure in cloud cover variation is generally confirmed by all stations.

An "approximate observation range" is given in Fig. 11. It is the approximate diameter of the field of view available to the observer who makes an estimate of cloud cover. The length shown has been corrected for the NW - SE trend of the coast.

The extrapolation seaward into the regions under study should give values of mean cloud cover to within $\pm$ 5%.

In the Strait of Georgia cloud cover distribution is confirmed by sunshine recordings. Since a comparison of fog occurrences with sunshine-cloud cover difference did not show a significant correlation, the effect of fog is taken to be included in cloud cover estimates.

Fig. 12 gives a comparison of monthly mean cloud cover with a long term average for three coastal stations. The values for 1961 and 1963 appear typical for all regions except possibly for greater than average cloud cover on the west coast of Vancouver Island.

(2) Wind

An estimate of the monthly mean wind speed was made by considering the shore measurements as functions of exposure and the wind direction. This was done as follows.

Mean wind speed and frequency of occurrence by direction were obtained from the Canadian and United States meteorological records. The data was plotted in the form of wind roses for each month of the two years. From the dominant wind directions as shown by the plotted miles of wind per month, and comparison of instantaneous wind vectors, a broad picture of wind pattern was developed.

There are two basic wind patterns which, for convenience, have been defined by reference to the direction of the wind at Estevan Point. With the wind from the south east direction (S.E'lies) at Estevan Point, Fig. 13, A and B, the wind over most of the coast follows the same general direction. There are two possibilities at some stations in S.E. Alaska, but the S.E'ly form is more common.

When the wind is from the north west direction (N.W'lies) at Estevan, Fig. 13 C and D, the coast south of Dixon Entrance will generally have winds from the same direction, while regions of S.E. Alaska will experience S.E'lies. The alternate flow of N.W'lies for the entire coastal area is also common.

Fig. 15 shows the monthly ratio between S.E'lies and N.W'lies for Estevan Point and Lincoln Rock. The total miles from a S.E. (N.W.) direction was taken as a sum of wind miles with a S.E. (N.W.) component, halving south westerlies and north easterlies; south easterlies are the most common wind direction in winter and less so in summer, although either wind direction occurs throughout the year.

Waldichuck (1957) showed a counterclock wise gyre in the southern Strait of Georgia during summer. No indication was found of this. The monthly total of wind for summer periods is complicated by the strong sea breezes which develop, giving a component perpendicular to the shore, and when combined with the gradient wind make flow directions difficult to interpret.

Not all listed Class 1 stations give a representative open water wind speed due to the modification by the local topography. Table 1V shows the Class 1 stations as either good or poor for exposure to the two major wind directions. By definition, Class 11 stations are necessarily poor.

Fig. 15 is a scatter diagram of mean wind speed for one month when north westerlies and for one when south easterlies were dominant. The curve is drawn through stations listed as good exposure in Table lV, except for the most northern group. It is evident that the mean wind speed is low at stations of poor exposure.

For each station the mean wind speed was obtained by increasing the wind mileage in the poor exposure direction until the ratio of S.E. component to N.W. component was the same as at the closest station of good exposure in both directions. This method may have limitations when the sea breeze is important. The mean wind speed for each region was estimated using the corrected values.

The exception is in the Strait of Georgia region where Tsawassen Ferry Terminal is the only station of good exposure in all directions. A mean between Tsawassen and Merry Island was used for Region 1. In Region 2, values were the Comox airport mean increased by the ratio of Region 1 winds to Vancouver airport winds. Simply using the same mean speed as Region 1 may be a better choice. Region 8 winds are an average of Eldred Rock and Sisters Island values. There is some channeling near Eldred Rock that would give means higher than those typical further south. Sisters Island means appeared unrealistically low.

Shown in Fig. 16 is a comparison of monthly mean wind speeds with a long term means for selected stations. The 1961 and 1963 values are near average, except possibly for higher winds at Estevan Point.

(3) Air Temperature and Dew Point

The mean monthly air temperature derived from daily readings of max - min thermometers was found to be a good representation of the integrated mean temperature for the hourly observations. Use of the former makes the measurements of the more numerous climatological stations available. Because of a diurnal variation in cloud cover and wind, choosing a daily mean for air temperature could introduce averaging error as shown by equation 23, page 16. The resulting increased heat loss at night by effective back radiation over the reduced daytime loss will tend to cancel to the first approximation, since cloud cover and air temperature cycles are almost symmetric and in phase. The diurnal variation of sea breeze could be important, particularly in inlets. The dew point has some diurnal cycle, up to 1 or 2 C^o. The mean of the four synoptic observations is considered representative for a monthly average, since the important variation in dew point occurs with passing weather systems rather than as a diurnal effect.

Fig. 17, a scatter diagram for air temperature, gives typical summer and winter distributions. The difference in monthly mean air temperatures between Class 1 and Class 11 stations is generally small. The more numerous climatological stations confirm the regional values that may be interpreted from the points given. A mean temperature, to the nearest 0.5 C^o was chosen for each region. This method should provide a climatological mean for the 10 metre height to $\pm$ 1 C^o.

Fig. 18 is a comparison of monthly mean air temperatures of 1961 and 1963 with a long term average.

A scatter diagram of dew point temperatures is shown in Fig. 19. Class 11 observations are typically lower than those from the nearby Class 1 stations. The values interpreted should be within $\pm$ 1 C^o in all regions except for the Strait of Georgia. Dew point for Region 2 was obtained from a mean between Comox and Merry Island. Values for Region 1 came from Vancouver airport. It is suspected that these are underestimates, but they are used for lack of better data.

Fig. 20 compares the dew point at two stations with an average value; it indicates that the two years are typical.

(4) Sea Surface Temperature

Observations of sea surface temperature taken at shore stations along the coast are used to give an estimate of sea surface temperature in each region. The monthly means used are those compiled by the Pacific Oceanographic Group (1962, 1964) and the U.S. Coast and Geodetic Survey (1964).

Pickard and McLeod (1953) analyzed the sea water observations taken along the British Columbia coast, and recognized three groups of similar property cycles. Using available cruise data an attempt was made in the present study to find which stations gave the most representative temperature for each region. In Fig. 21 is shown a comparison of measurements taken in Queen Charlotte Sound by the Pacific Oceanographic Group (P.O.G.)(1955), with the seven day running mean of sea water temperatures at Pine Island and McInnes Island. During the winter, observations at both lighthouses are close to surface temperatures in Queen Charlotte Sound. In summer McInnes Island follows the increasing offshore temperatures, while Pine Island remains relatively cool. The lower Pine Island values are thought to be due to tidal mixing (Pickard and McLeod, 1953). A similar comparison was made for other stations where cruises gave sufficient coverage. Winter sea surface temperatures tend to be uniform in any one region, with little gradient perpendicular to the coast. Shore observations taken in this season are considered as representative. The summer distribution is more variable, is affected by tidal mixing, and often has large horizontal gradients. This may be seen in most contoured summer sea surface temperature charts for the coastal waters. Waldichuck (1957, page 358) shows an example of such a distribution in the Strait of Georgia for June 1950. In Fig. 22 is shown a typical summer and winter distribution of monthly mean values of sea surface temperature measured at the coastal shore stations. Only three separate stations, all with poor location in restricted channels, are available in S.E. Alaska. How well they represent the actual surface temperatures is uncertain. For the winter season it should be possible to obtain a mean sea surface temperature to $\pm$ 1 C^{o} for each region. In summer it is doubtful if the lighthouse observations can be used to give mean conditions to the same accuracy.

HEAT BUDGET COMPONENTS OF THE COASTAL WATERS

Seasonal Cycle of Calculated Surface Heat Transfer

Using the method outlined in the previous chapters, monthly surface heat transfer was calculated for each month of 1961 and 1963. The results are shown in Fig. 23 and 24. In both years the broad annual cycle is modified by the non-uniform climatic conditions. The fine structure in the cycle is not local and generally occurs along the entire coast with regional trends. For example, the high loss in November of 1963, though evident in all regions, is more pronounced to the north. A comparison between 1961 and 1963 shows that from year to year the surface heat transfer for a given month may vary considerably. Except for February, most monthly values are within 20% of the mean curve for 1961 and 1963. Because of the close coupling between temperature structure in the upper layers of the sea and rate of surface heat transfer, these calculated deviations from a mean are sufficiently large that they should be observed as deviations in a mean advection or mean heat storage in a study with a time scale of months.

The low loss in February 1963 was due to a small vapour pressure and temperature difference between the air and the sea surface. This may be typical of particularly cloudy winter months. The high loss that occurred in November 1963 may be common in the winter cooling cycle.

The heating season varies in length, with a general trend to increased duration in the south. It is approximately six months long in the northern regions, and seven to the south, but local variations exist. For example, in regions 6 and 7 during 1963 the fine structure is similar, but the heating season differs by approximately two months. This is the result of the near balance between gains and losses during February of that year.

Components of the monthly heat transfer for Regions 1 and 7, 1963, are shown in Fig. 25 and 26. They are typical of the seasonal character of the components in the southern and northern regions. The largest input of heat comes from the short wave radiation, with values in Region 1 ranging from approximately 500 langleys/day in summer to 100 langleys/day in winter. The solar input, most variable in summer, is strongly influenced by cloud cover. For all regions long wave radiation is relatively constant through the year, varying between 50 langleys/day and 100 langleys/day, lowest during the summer due to increased counter radiation by the warmed air column. The radiation balance is generally negative for part of the winter.

The convective transfers, evaporation and sensible heat conduction, are highest in winter, lowest in summer. Values during the cooling season are typically between 100 langleys/day to 200 langleys/day. Evaporation losses are normally greater than those by sensible heat conduction, but occasionally the large temperature difference between air and water during the winter months can result in larger transfer by sensible heat conduction, as shown in Fig. 26. In summer Q_h normally heats the water surface, though the input is small, typically less than 20 langleys/day. Q_e is low most of the summer generally less than 50 langleys/day. (Values shown for Region 1 are suspected to be high because of unrepresentative data.)

The peaks of the net input cycle during 1961 and 1963 were found to be approximately 400 langleys/day to 300 langleys/day in the southern regions and between about 400 langleys/day to 500 langleys/day in northern regions.

Latitudinal Change of Surface Heat Transfer

In Fig. 27 is given the yearly average of surface heat transfer, Q_D, for each region studied.

There is a general decrease northward, with large local variations superimposed. The same relative distribution occurred in both 1961 and 1963, although the absolute values differ considerably for some regions. Since most changes in exposure are represented by the regions chosen, the gradient is probably representative for intermediate points along the entire coast.

Also in Fig. 27 are shown the average radiation balance and average convective transfers. R has an approximately unidirectional drop with increasing latitude, except for the dip in Region 3. This dip is due to the noticeable peak in cloud cover at North Vancouver Island. The difference in calculated values of R for 1961 and 1963 is within a few percent of the total, suggesting small yearly variation. The mean slope of R is about 30% higher than values calculated by Mosby and McEwen (Sverdrup 1951) for the mid-North Pacific.

Budyko's (1956) results are almost 10% higher than the present calculated R. The radiative input in the coastal area is expected to be above the mid-Pacific values at the same latitude because of lower cloud cover.

The mean of the convective losses is approximately constant along the coast, increasing slightly to the north. There are wide fluctuations with region and year, the higher losses being found in the less protected regions on the open coast. The regional variation is due largely to the variation in sea surface temperature.

In Fig. 28 the average surface transfer is shown separately as the heating and cooling season. The similarity to the regional trends of the radiative and convective losses in Fig. 27 shows how the former dominates the heating season, while the latter is responsible for most of the variation in the cooling season.

Shown in Fig. 29 are the averages of components in the transfer for 1961. The latitudinal trend in R resulted from decreasing solar radiation input toward higher latitudes. This decrease in short wave radiation is counter-balanced somewhat by lower loss through effective back radiation in the northern section, where water temperatures are lower as well as cloud cover higher. The mean annual short wave radiation input, $Q_s(1-r)$, varies from approximately 280 langleys/day in Region 1 to 180 langleys/day in Region 8. Q_b gives a yearly average loss of about 100 langleys/day to 75 langleys/day between the same regions.

Of the two convective transfers, evaporation is dominant in a yearly average. It is reasonably constant over the regions considered, varying between 50 langleys/day to 80 langleys/day. There is no definite trend and it is more a function of locality. The average sensible heat conduction is lower than the average evaporation, with mean values of approximately 25 langleys/day over most of the coast, increasing to 40 langleys/day in Regions 7 and 8.

The mean of the annual input for the two years decreases from approximately 70 langleys/day to 30 langleys/day between west Vancouver Island and Dixon Entrance. The decrease between these points is due largely to the drop in net gain from the radiation balance, though some increase in evaporation loss also occurs.

Net input for Regions 1 and 2 in the Strait of Georgia is about 90 langleys/day. Though the convective loss is slightly lower than on the open coast, the higher gain is largely due to the higher radiative input. The Strait of Georgia has noticeably less cloud cover than west Vancouver Island during the summer period.

A transition in annual input occurs at the southern channels of S.E. Alaska between Region 6 and Region 5. This results from the much lower convective losses in Region 6 due to lower sea surface temperature. In S.E. Alaska there is a decrease of Q_D from approximately 50 langleys/day to near zero between Regions 6 and 8 respectively.

The large difference between the 1961 and 1963 values of net surface input in Regions 2a and 4 is the result of lower input with higher losses one year, and the opposite the other.

Heat Storage

The annual balance between surface heat transfer, heat storage and advection can be estimated by combining the calculated surface heat transfer with measurements of temperature structure. The rate of change in heat storage may be evaluated from the changing temperature structure. The difference between surface exchange and local storage would be due to advective transfer.

Measurements of temperature structure are available for Saanich Inlet, British Columbia for 1961 and 1963. The rate of heating of the water column in Saanich Inlet was estimated from temperature profiles (bathythermographs) and oceanographic measurements taken during 1961 and 1963 at a single, central station in the inlet. Integration of the heat content in the water column was done by a summation approximation.

Saanich Inlet is an example of a basin with small exchange of the deep waters, the result of little or no estuarine circulation. However Herlinveaux (1962) reports a continuous advection of the surface layers in the inlet by tidal action suggesting that the inlet has a direct connection with events in the southern Strait of Georgia through advection. No advective exchange of bottom water was noted to have occurred in either year studied.

A comparison of the rate of heating calculated for Saanich Inlet from the temperature structure and the surface heat transfer of Region 1 is given in Fig. 30. The calculated surface heat transfer accounts for the temperature structure, to within 20% of the peak values through the winter and spring of the year. A higher rate of heating in Saanich Inlet in the spring of 1963 is expected due to lower cloud cover in that area, compared with values used for Region 1. Beginning in July of both years and continuing until late in the year, October for 1961, the change of heat content in the inlet was well below the calculated net input at the surface in Region 1. The difference is thought to be the result of a loss by advection. The temperature profiles indicate the exchange to be in the surface layers. The onset of this large advection loss could be associated with either

(1) estuarine circulation, or

(2) wind transport.

The first possibility would occur if the estuarine circulation in the Strait of Georgia resulted in an advective loss from the Strait, and the net lower heating transmitted to Saanich Inlet by a tidal exchange.

The runoff cycle of the Fraser River for 1961 and 1963, Fig. 31, has a peak in June for both years, with approximately half the total summer discharge past at this time. Waldichuk (1957) showed that the outflow of water from the Strait of Georgia was delayed approximately one month past the peak in the runoff cycle, close to the observed onset of advection loss in Saanich Inlet. Also in Fig. 31 are the estimated wind directions at Comox. There is no definite increase in N.W. component for July or August, suggesting little direct dependence on wind direction.

Ignorance of the magnitude of the advective exchange in more detail than that shown makes an accurate estimation of error in the calculations impossible. The close correspondence obtained during the cooling season, and first part of the heating season, and good response to the low loss in February 1963, suggests that the absolute as well as the relative cycle calculated is close to actual conditions, possibly to within 20% of the peak values.

It is also interesting to compare the yearly cycle in heat storage in the Strait of Georgia calculated from P.O.G. (1954) with the surface heat transfer for 1961 and 1963. A summation approximation was used to obtain a volume integral of the total heat content in the top 200 metres of the entire basin. The change in heat content was then used to calculate the rate of flux through the sea surface. Where data was available, the full depth range was used. Values obtained from the total integration were not significantly different from the results from the top 200 metres. The results are compared in Fig. 32 with the average of calculated surface heat transfer in Regions 1 and 2 for 1961 and 1963.

The seasonal cycle in heat content of the Strait of Georgia is shown to be similar to that of Saanich Inlet, with the same possible large advection loss in the late summer and fall months. The calculated heating of the water column for 1950 shows a higher loss in January, and higher gain in April than calculated surface heat transfer for 1961 and 1963. In January 1950 the Strait of Georgia region was abnormally cold due to the influx of continental arctic air, a possible explanation for the higher loss. April and July 1950 had lower cloud cover than the mean for 1961 and 1963. This is the probable reason for the increased heat storage in April 1950. The remainder of 1950 was similar to 1961 and 1963.

Waldichuk (1957) calculated the net advective exchange in the Strait of Georgia during 1950. This Q_v term is added to the heat storage, Q_θ, also calculated for 1950. The result shows good agreement with the average surface heating in 1961 and 1963 over most of the year. The high value in $Q_v + Q_\theta$ for May 1950 over the surface heat transfer, Q_D, is probably due to an over estimation of the advective loss during the first part of the year. Again no estimate of accuracy in the calculations is possible,but on the assumption that 1950, 1961 and 1963 had similar meteorological conditions, it appears that the various components of the heat budget in the Strait of Georgia can be calculated to within 20% of the peak value.

This same pattern of possible large advective losses through the runoff period was found in some of the coastal inlets for which a time series of measurements was available.

Shown in Fig. 33 is a comparison of the surface heat transfer calculated in this study with the heat storage in Dixon Entrance for 1954 - 1955. The heat storage terms are those evaluated by Tabata (1958). Because of yearly variation in climatic conditions and the large advection influence, a direct comparison of magnitudes is not possible. The difference does indicate that an advective loss occurred during July - August of 1954, and March - April of 1955. There also appears to be a net advective gain of heat through the winter months.

Effect of Surface Heat Transfer on Surface and Deep Water Temperature

(1) Surface Water Temperature

The surface temperature can potentially be modified by tidal mixing along the shore, upwelling, wind mixing, advection, and surface heat transfer. It appears that all these factors play some role in the surface temperature distribution observed along the coast. There is a noticeable variation in the amplitude and phase relations of the surface temperature cycle for the different coastal regions.

A. Season Cycle

In all regions studied, for 1961 and 1963, the surface water temperature was observed to be at a minimum close to the end of the cooling season, as expected. In some regions there was a slight increase of approximately 1 C^{o} before the onset of surface heating. This is considered to be due to the upward mixing of warmer deep water in areas such as the Strait of Georgia, or warm inflow on the open coast, in combination with a decreasing surface loss.

The occurrence of the temperature maximum is variable and may be more than a month before the end of the heating season, Fig. 34. The earliest observed peaks were in July or August in the protected regions, such as the Strait of Georgia and S.E. Alaska. The possible reason for this premature maximum can be seen in Fig. 35 from the comparison of surface temperature and calculated rate of heating in the Strait of Georgia, 1950. The temperatures at Entrance Island and Cape Mudge begin to drop while the water column is still heating at the rate of 100 langleys/day. The low net input for this time of year results from the advection seaward of the warmed surface layers, which are replaced by colder deep water. Wind mixing is sufficient to overcome the net input, by mixing colder deep water to the surface. In regions less affected by advection loss from the water column, the surface temperature will continue to rise until near the end of the heating season.

In Fig. 36 is a comparison of the root-mean square values for rate of surface heat transfer and monthly change in surface temperature in corresponding regions. The relative distribution in amplitude differences for the coastal regions appears to be strongly influenced by factors other than the surface heat transfer. The possible effect of advection and wind mixing on amplitude, Fig. 37, can be seen in a comparison of temperature structure in the Strait of Georgia with that off west Vancouver Island. In the Strait of Georgia much of the summer heating has been removed by the estuarine circulation. With the subsequent equivalent cooling calculated for the two regions, there would be a greater chance of low temperature at the surface in the Strait of Georgia, provided the stability of the water column is the same. The stability is greater in the Strait of Georgia, thus wind mixing will not dissipate the surface cooling as readily. Advection of warm water into Region 2a is also evident from the curves in Fig.37. Using values calculated for October 9 to December 31 of 1961 as equivalent to the same period in 1960, the calculated drop by surface heat transfer in the Strait of Georgia over this period has a mean of 114 langleys/day and the calculated value from the temperature profile is 101 langleys/day. Off west Vancouver Island the calculated mean loss at the surface is 88 langleys/day while the calculated change in heat content is only 10 langleys/day, indicating an influx of warm water. As seen from the temperature profile, the advection appears to be throughout the upper water column. For the summer, higher insolation and greater stability in the Strait of Georgia are sufficient to yield higher surface temperatures than those off the west coast of Vancouver Island. Thus a greater range of surface temperatures in the Strait of Georgia is possible from the effect of

1) estuarine circulation
2) stability, and
3) no warm advective influx.

Similar conditions to those off west Vancouver Island are probably influencing surface temperatures in the Regions 3, 4 and 5. The low range of values near Pine Island is thought to be partially due to the low surface heating in summer, and an advection influence of heating in winter, cooling in summer.

The transition that occurs between Dixon Entrance and S.E. Alaska is probably due to the same factors which resulted in the differences across Vancouver Island. In Dixon Entrance the yearly cycle of surface temperature is kept low by efficient wind mixing and advection. The comparison of calculated transfer to heat storage estimated by Tabata (1958) showed an influx of warm water during winter and of cold during summer. For S.E. Alaska the summer heat input is kept near the surface by the greater stability of the water column. The removal of most surface heat by the estuarine circulation then makes winter cooling effective.

The greater range of temperatures in S.E. Alaska over that in the Strait of Georgia is probably the result of the colder bottom temperatures found in S.E. Alaska. The difference is approximately 4 C°.

B. Latitudinal Variation

Shown in Fig. 38 is the mean surface temperature for the coastal regions in 1961 and 1963. There is a general decrease northward, from approximately 11°C to 7°C between the Strait of Georgia and northern S.E. Alaska. This gradient is influenced by the same factors which vary the amplitude between regions. For a given surface heating in summer S.E. Alaska will attain the highest temperature change, followed by the Strait of Georgia and then by the more open coast stations. For a given cooling at the surface, the order of decreasing effect is the same. Due to this varying effect for a given surface heat transfer, the mean surface temperature does not reflect the mean annual input calculated for the coastal region.

(2) Deep Water Temperature

A. Seasonal Cycle

The cyclic action of surface heating and cooling has been observed to depths of 100 metres in inlets

(Pickard, 1961) and 100 metres or more in the North Pacific (Tabata, 1961). In the deep channels of S.E. Alaska, temperature sections suggest a seasonal influence to between 100 metres and 200 metres (Institute of Oceanography, University of British Columbia, 1964).

B. Latitudinal Variation

For coastal regions and inlets with a deep sill below 200 metres, connecting across the continental shelf to the deep waters of the Pacific, the bottom water will have a gradient northward similar to that for the nearshore Pacific at a similar depth. Typical temperature values are approximately 7.5°C to 5°C between Juan de Fuca Strait and Chatham Strait. For inlets with sills shallower than 150 metres, there is probably some influence on bottom water from the surface heating, either from penetration by wind mixing or through the local formation of bottom water. Waldichuk (1957) found the bottom water for the Strait of Georgia to be formed by tidal mixing at the sill. A decrease northward of bottom temperatures is reported by Pickard (1961) for the inlets of British Columbia. Bottom temperatures in inlets and channels of S.E. Alaska are at the temperatures of the deep water inflow or lower, (Institute of Oceanography, University of British Columbia, 1964).

SUMMARY AND CONCLUSIONS

In this study an attempt was made to evaluate the heat transfer across the air - sea interface for regions along the coast of British Columbia and S.E. Alaska using empirical formulae and existing data. The two years chosen for the analysis, 1961 and 1963, are typical of climatic conditions found along the coast.

An empirical relation was developed to relate relative global radiation to cloud cover and cloudless day radiation. This equation has an accuracy of approximately $\pm$ 10%. When compared with formulae obtained from data at Station 'P' it is evident that in overcast conditions, insolation in the coastal region is above that in the open ocean, the result of changing cloud properties. A non-linear equation is required in the coastal region since the range of cloud cover is beyond the limits of linear forms.

From a comparison of formulae for calculating effective back radiation, equations of Budyko (1956) and Anderson (1954) appear to show equal response to varying conditions found in a coastal climate.

The convective transfer of heat by evaporation and sensible heat conduction was evaluated using semi-empirical equations that employ a bulk transfer coefficient. Choice of a value for the coefficient requires assumptions about the profile of properties above the sea surface. The logarithmic profile possibly valid for neutral stability of the air column is assumed to be a suitable representation of average conditions. In the winter months when the convective transfers are a large fraction of the surface heat exchange, significant error may be introduced by this assumption. The losses will potentially be underestimated because of the unevaluated increase in turbulence resulting from an unstable air column.

It was found that the location of the observation station is important in obtaining "over water" values of wind or dew point, and possibly air temperature. The coastal stations were considered in two classes of exposure. The data at the more representative stations gave higher mean dew point, higher mean wind speed, and a smaller yearly cycle of mean temperature. There is some uncertainty as to what "over water" height is represented by shore measurements taken close to the ground. At adjacent stations of similar exposure but different heights, there was no noticeable variation in monthly means, that could be attributed to the height difference.

A plot of horizontal distribution of the meteorological and oceanographic data was used to interpolate to the regions of interest.

There are a sufficient number of stations along the coast to show the distribution of cloud cover in the regions studied, with overlapping in some cases. The cloud cover has consistent variation along the coast, is more uniform in winter, and normally larger in the more northerly regions. A 10% difference in monthly mean cloud cover between adjacent stations is uncommon regardless of exposure.

Cloud cover was found to be the most important variable in controlling the heat transfer during the heating season. A further study is required to show how well the existing material may be extrapolated if these observations are to be used for calculations in the more protected inlets. There was no definite trend in cloud cover variation perpendicular to the coast line. In all areas, cloud height increases inland. During the summer the Strait of Georgia typically has lower cloud cover that the west coast of Vancouver Island. Through most of the year there was a peak in cloud cover at the north end of Vancouver Island. The channels of S.E. Alaska were generally found to have more cloud cover than the open coast.

Air temperature is well represented by the stations along the coast. Temperatures from the climatological stations only add detail. A mean gradient to lower air temperature in northern latitudes is found in all months, though in summer the change may be less than 0.5 C^o. The amplitude of the annual cycle of air temperature increases as one progresses from the mouth to the head of coastal inlets.

Dew point was observed to be a strong function of exposure, particularly in the summer months. The stations of good exposure show a consistent variation of dew point along the coast. Dew point generally decreases northward, sometimes with a peak in southern S.E. Alaska, a region of high precipitation.

There is a noticeable correlation between dew point and water temperature, suggesting that the observed dew point values are largely the result of local evaporation. It may be possible to use such a variation to improve the extrapolation of data.

Modification of wind systems by the mountainous coastal topography produces two dominant wind directions, called south-easterlies and north-westerlies, that tend to lie parallel to the coast. The winds are divided into the two typical mid channel flow patterns, by using the wind direction at Estevan Point as a reference. Because of the effect of topography, there is considerable local variation from this general "mid channel" flow.

Good exposure requires that the winds be blowing directly off the water toward the station with little effect from the small scale topography. To apply a first order correction for varying exposure, the mean wind speed at a station with one direction of poor exposure was adjusted by equalizing the ratio of wind mileage between north-westerlies and south-easterlies to that of a nearby station with good exposure in both directions. The mean wind speed does not have a definite latitudinal variation but tends to be highest at the more exposed section of the coast.

Extrapolation into more local regions requires an extension of the flow pattern and some evaluation of effects of channelling. The sea breeze may introduce a significant increase in evaporation loss during the summer because of induced advection. Squamishes, another localized effect, are more important to the calculation of convective losses, since they occur in the winter months. Though a squamish may be of relatively short duration, the cold dry air and high wind velocities could produce significant increase in surface heat loss in the inlets where they tend to be localized. The full intensity of this inflow of continental air is not normally recorded at the present coastal observation points.

The sea surface temperature along the coast was obtained from observations taken at shore based stations. Water surface temperatures are lowest to the north throughout the year. Cruise data show that this method of observation provides a reasonably good estimate of offshore conditions during the winter months, since the surface temperatures are relatively uniform perpendicular to the coast. In summer, due to tidal mixing and horizontal gradients, the representativeness is doubtful. Fortunately it is the winter period when convective losses are large and an accurate sea surface temperature is more important.

From the calculation of the monthly surface heat transfer there was found to be a strong annual cycle with a superimposed fine structure. The deviations from a mean curve may be as large as 40% of the peak values, and are normally present along the entire coast. The observed peak to peak amplitude for the annual cycle is approximately 750 langleys/day.

In the summer period there is a domination by the radiative transfer. The net transfer by evaporation and sensible heat conduction remains small and nearly constant at about 15% of the net radiative transfer. During the winter the convective transfers are the largest terms, and together comprise about 75% of the total loss.

Effective back radiation is approximately constant throughout the year.

A comparison of the surface heat transfer with the estimated heat storage in Saanich Inlet and the Strait of Georgia shows that for these water bodies, net advection of heat energy appears to be important only during the summer estuarine circulation period. In more exposed areas, such as Dixon Entrance and west Vancouver Island, advection may be important throughout the year. In these exposed regions, advection is observed to remove heat from the water column in summer, and add heat in the winter. From the correspondence between surface heat input and heat storage over the period of apparently small advection, it appears possible to calculate the surface heat transfer for the regions studied to within 20% (of the peak values), using existing data and formulae.

The latitudinal variation in annual surface heat transfer was calculated to decrease from approximately 90 langleys/day in the Strait of Georgia to near zero in northern S.E. Alaska. Transfers in the cooling season, which depend largely on the convective losses, are responsible for much of the regional variation. The convective losses reflect the varying sea surface temperature. The net input during the heating season shows a more uniform gradient along the coast, and varies relatively little from year to year.

The minimum in sea surface temperature occurs near the end of the cooling season, while the surface maximum tends to occur before the end of the heating season, earliest in the region influenced by the estuarine circulation system.

The observed variation in the annual range of surface temperature appears to be controlled by a combination of heat loss by the estuarine circulation, stability of the water column, offshore advection of upwelling, and bottom temperatures, as well as surface heat transfer. The greatest range of amplitudes occur in S.E. Alaska, followed by the Strait of Georgia and then the more open coastal stations in Dixon Entrance south to Vancouver Island.

Since the seasonal cycle of temperatures has not been observed at depths below 200 metres, the deep water of the coastal area will probably be influenced by the surface heat transfer only if deep water is formed locally, or a sill forces deep inflow to be above this depth.

Future studies to show how well the surface heat transfer may be calculated by extrapolation of the existing data would best be evaluated in a water body such as Powell Lake or a shallow silled inlet for which the advective exchange is small. Most of the surface transfer will then be reflected in the heat storage term.

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TABLE I

Comparison of Mid-monthly Cloudless Day Solar Radiation (langleys/day)

	J	F	M	A	M	J	J	A	S	O	N	D
49° 16' N. 123° 15' W.												
U.B.C.	155	268	413	582	696	758	729	623	465	306	181	126
Mateer	159	278	428	601	720	780	740	630	478	316	189	138
Kimball	191	306	468	633	749	803	746	649	502	343	215	160
55° 02' N. 131° 34' W.												
Annette	107	228	390	580	710	790	755	615	440	270	145	70
Mateer	110	210	375	575	710	775	730	600	425	250	135	80
Kimball	150	260	430	610	760	810	780	650	470	300	180	100

TABLE II

Mid-monthly Values of Cloudless Day Solar Radiation langleys/day

Reg-ion	Lat. °N.	Long. °W.	J	F	M	A	M	J	J	A	S	O	N	D
1	48° 55'	123° 05'	154	268	413	580	698	756	733	624	465	308	182	126
2	49° 50'	124° 50'	148	254	400	586	698	756	733	624	453	289	177	118
2a	49° 10'	126° 40'	148	254	400	586	698	756	733	624	453	289	177	118
3	51° 10'	128° 25'	140	242	388	586	698	756	733	622	441	270	168	107
4	52° 25'	129° 35'	121	223	372	572	697	756	730	614	428	252	146	87
5	54° 30'	131° 20'	112	208	363	560	690	754	726	604	415	244	136	78
6	56° 0'	132° 45'	100	192	355	550	687	750	722	595	404	229	123	64
7	57° 20'	133° 35'	82	178	344	538	684	746	718	584	396	214	109	55
8	58° 10'	135° 10'	68	164	327	524	680	746	712	574	378	204	96	45

TABLE III

Sources of Data

	Cloud Cover	Wind Speed and Direction	Dew Point	Air Temperature	Water Temperature
British Columbia Monthly Record Meteorological Branch, (1962, 1964a)	x	x	x	x	
General Summaries of Hourly Weather Observations in Canada. Meteorological Branch (1963)	x			x	
Project No. 06064 Meteorological Branch (1964b)		x			
Observation of Sea Water Temperature and Salinity Pacific Oceanographic Group (1962, 1964)					x
S.E. Alaska Climatological Data, Alaska Weather Bureau (1962a, 1964a)				x	
Local Climatological Data Weather Bureau (1962b, 1964b)	x	x	x		
Local Climatological Data (Supplement) Weather Bureau (1962c,1964c)		x	x		
Surface Water Temperature of the Pacific Coast. Coast and Geodetic Survey (1964.)					x
Monthly Means of Meteorological Data of S.E. Alaska. Weather Bureau, (1964d.)	x		x		

TABLE IV

STATION CHARACTERISTICS

Stations		Metres Approximate Observation Height				Good Exposure to Winds	
Class I	Class II	Wind		Air Temperature and Dew Point			
		Above MSL*	Above Ground	Above MSL*	Above Ground	SE'lies	NW'lies
British Columbia							
Alert Bay		est-imated		55			
Bull Harbor		est-imated		6			x
Cape St. James		104		92		x	x
Chatham Pt.		33		22			
	Comox		16		1		
	Ethelda Bay		16		1		
Estevan Pt.		20		7		x	x
Langara		est-imated		44			x
McInnes I.		35		25		x	x
Merry I.		25		9		x	
	Nanaimo		5		1		
	Port Hardy		10		1		
	Sandspit		20		1		
Spring I.		37		12		x	
	Tofino	19			1		
Triple I.		est-imated		23		x	x
	Vancouver		19		1		
	Victoria		22		1		
	Victoria (Gonz.)		15		1		
S.E. Alaska							
	Annette				1		
Cape Decision		36		15		x	x
Eldred Rock		25		16		x	x
Five Fingers		38		11		x	x
Guard I.		18		7			x
	Gustavas		10		2		
Lincoln Rock		21		14		x	x
Sisters I.		19		12		x	x

* Mean Sea Level

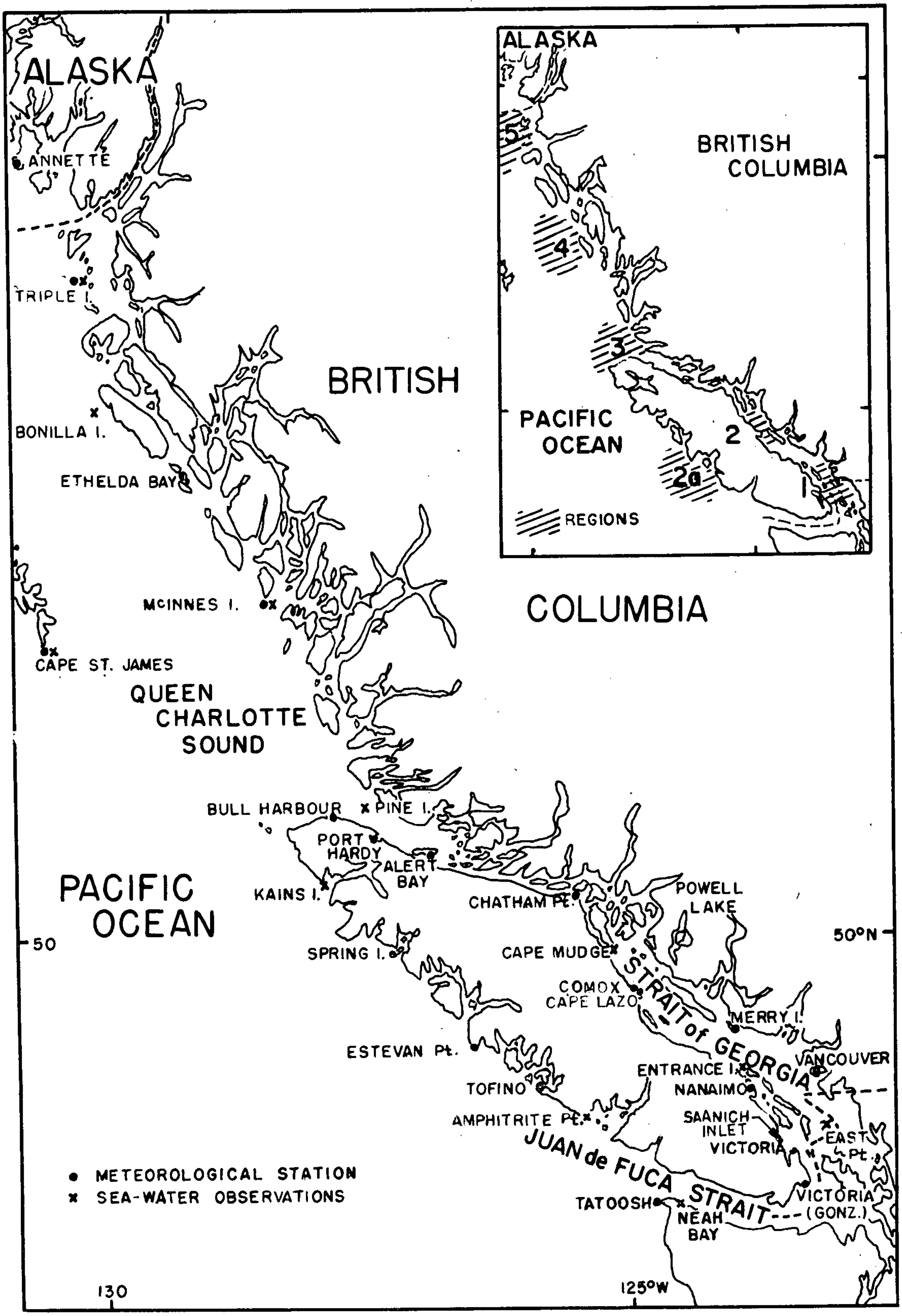


Fig. 1A: Observation Stations for Coastal British Columbia.

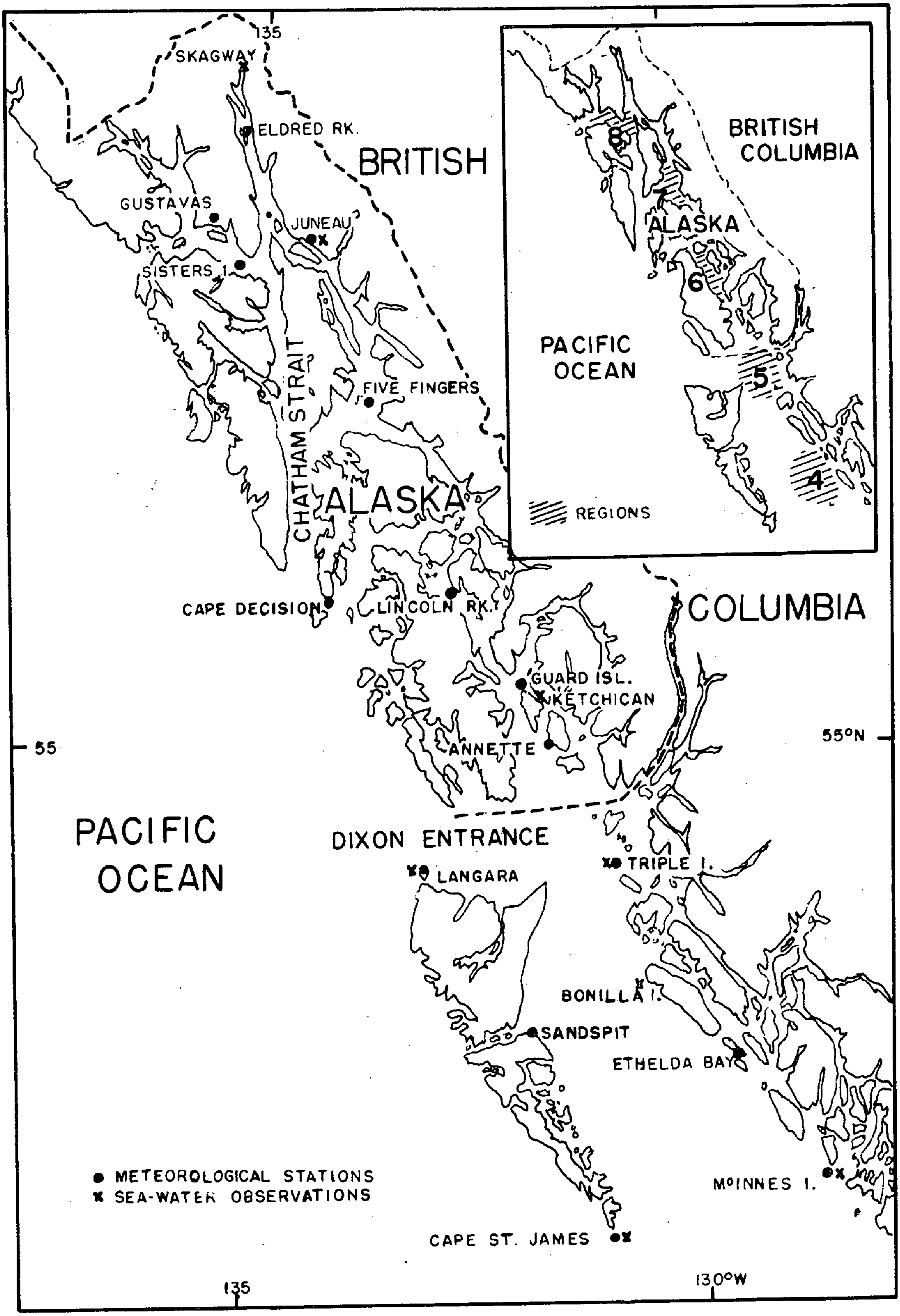
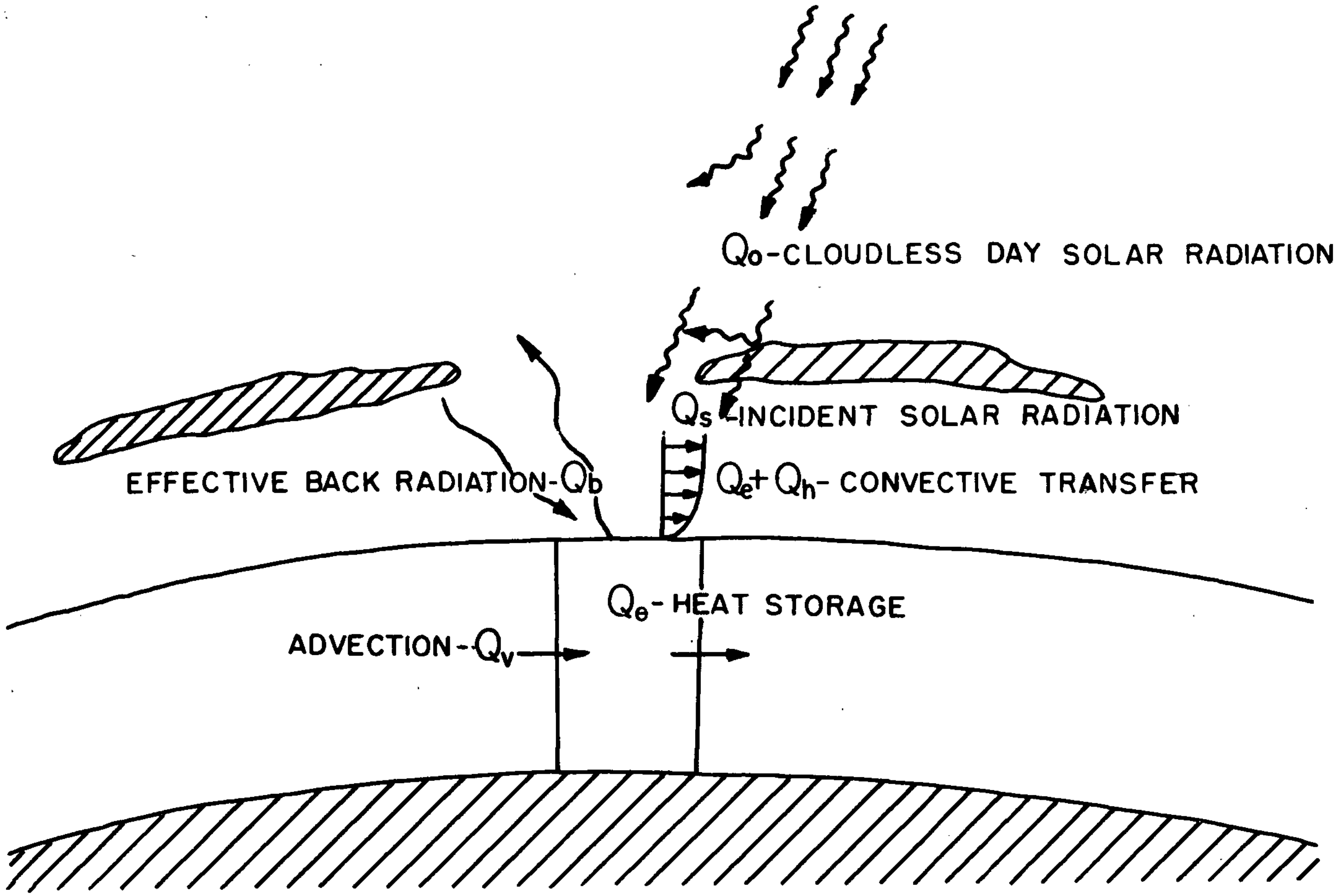


Fig. 1B: Observation Stations for S.E. Alaska.



$$Q_b + Q_s(1-r) + Q_e + Q_h + Q_v = Q_\theta$$

Fig. 2: Balance of Heat Energy in a Column of Water.

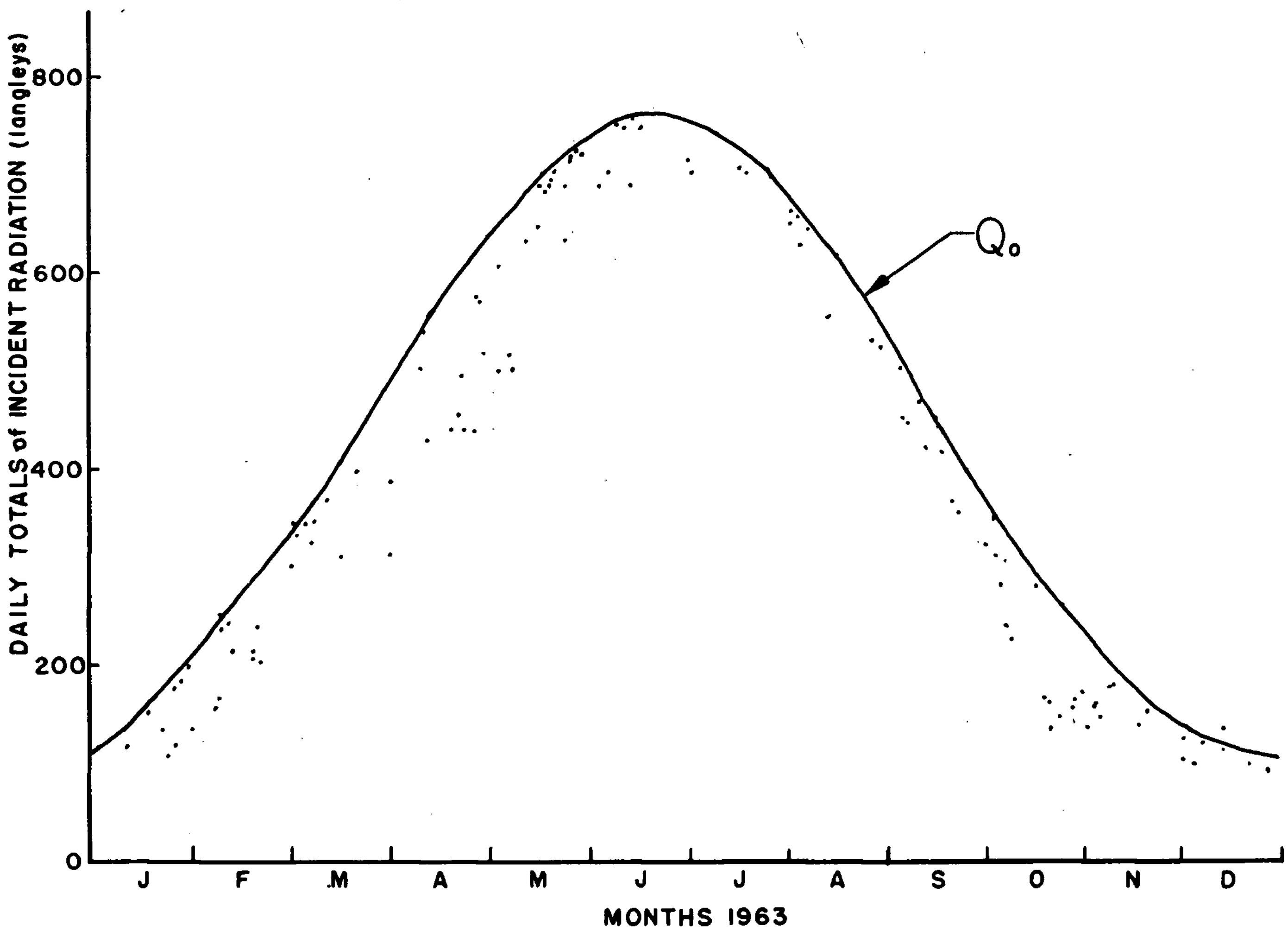


Fig. 3: Daily Values of Solar Radiation at Climatological Station, University of British Columbia, Vancouver, British Columbia. (49°16'N, 123° 15'W).

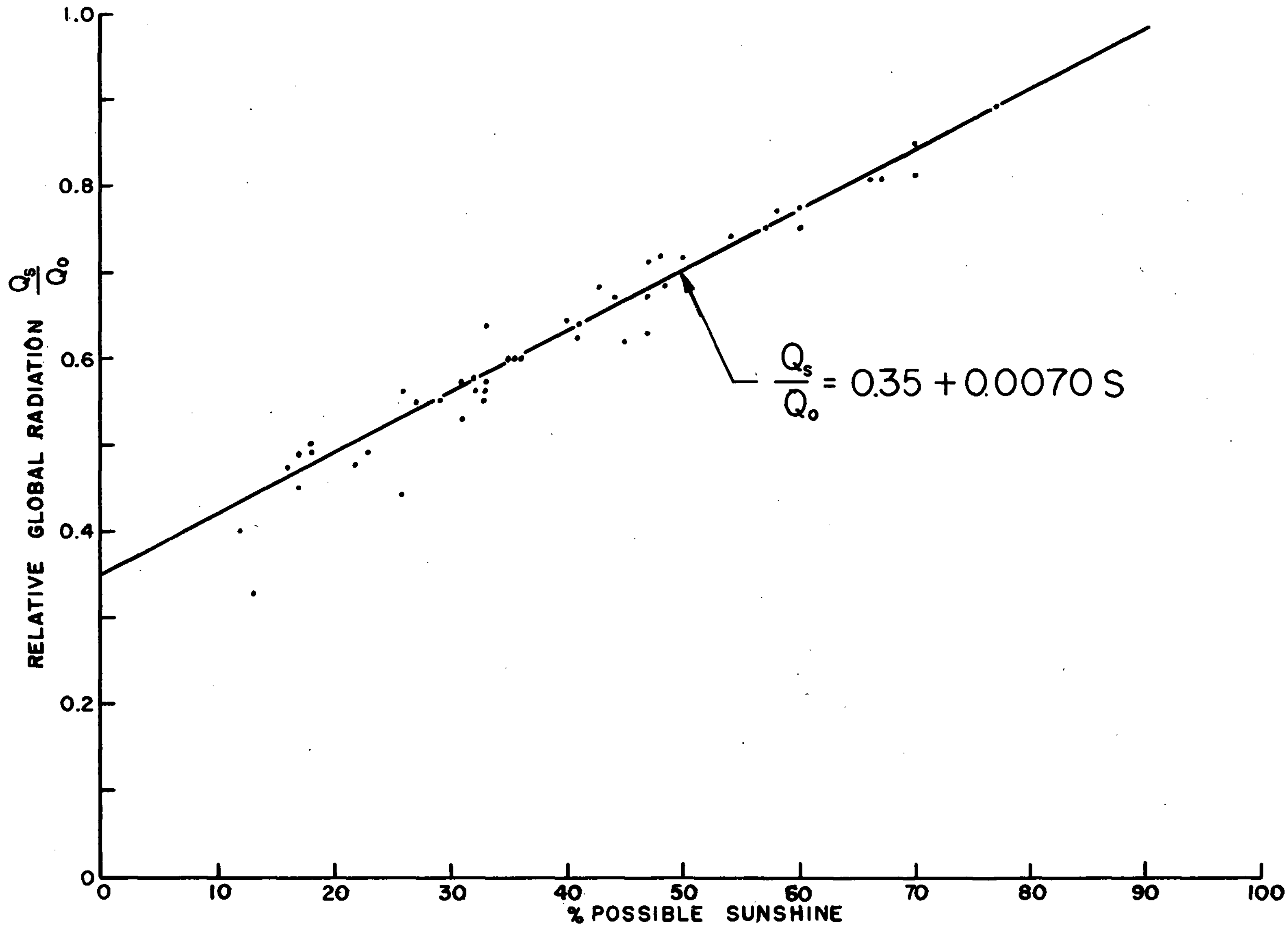


Fig. 4: Comparison of Relative Global Radiation with % Possible Sunshine.

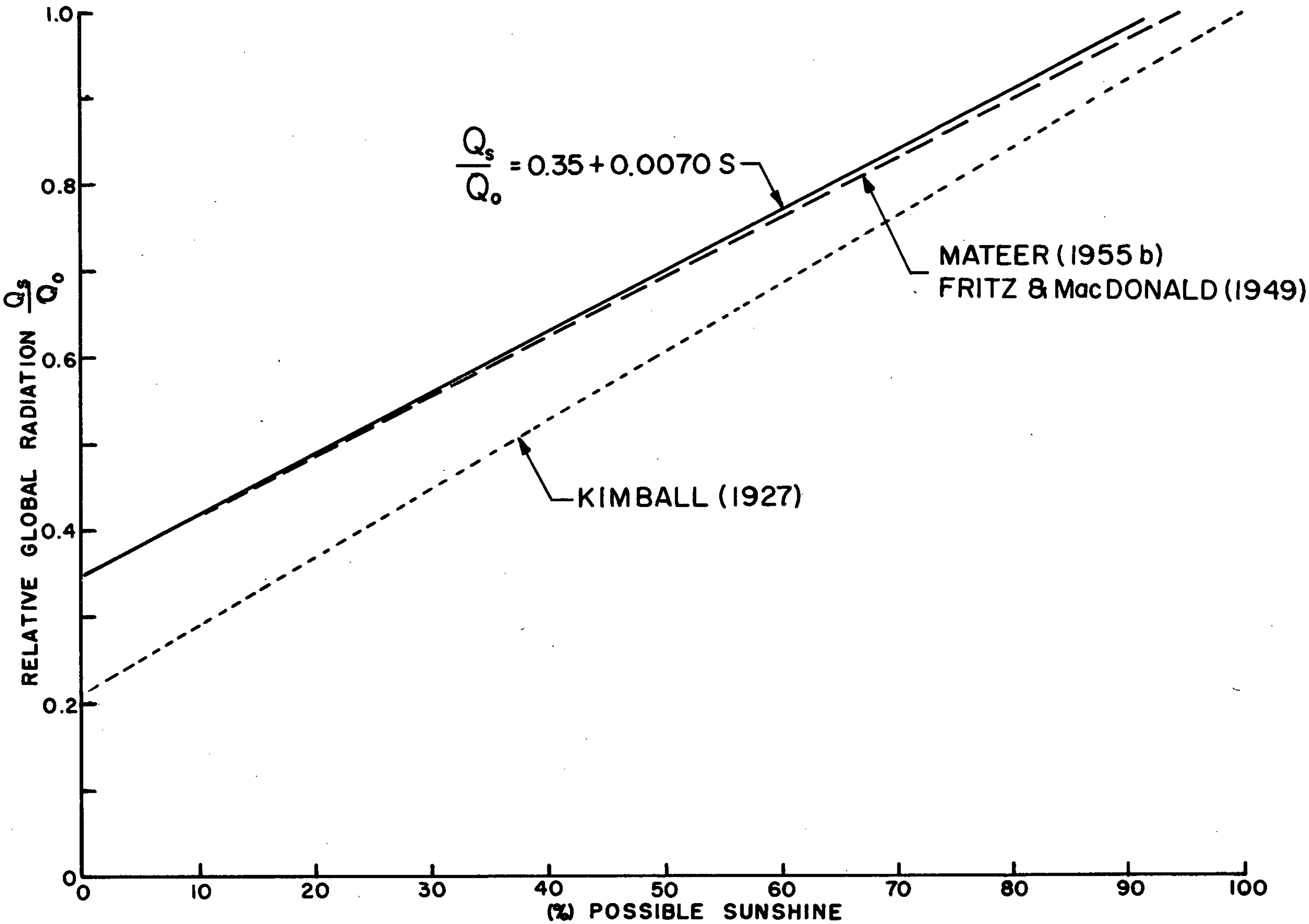


Fig. 5: Comparison of Sunshine Formulae.

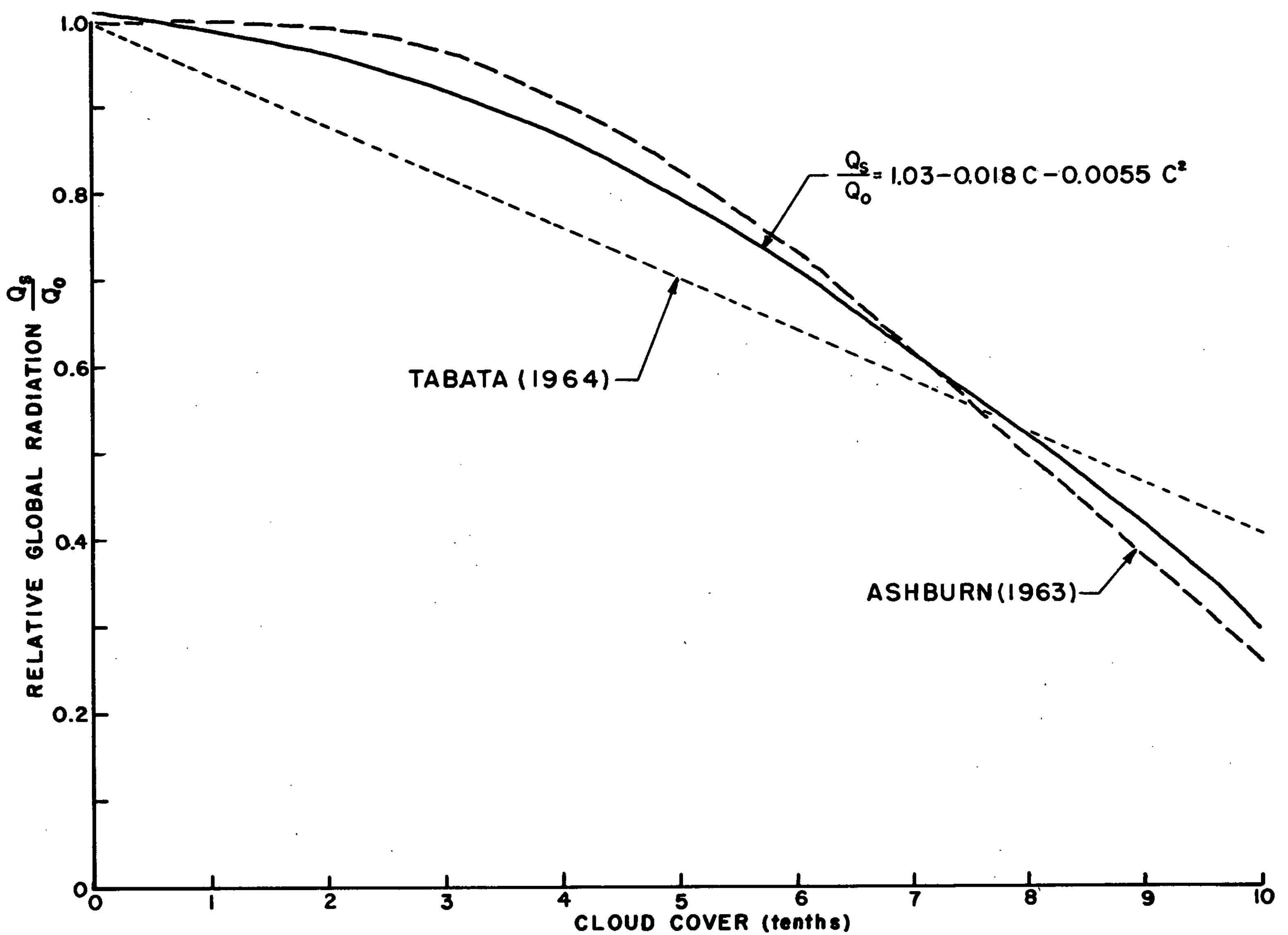


Fig. 6: Comparison of Cloud Cover Formulae.

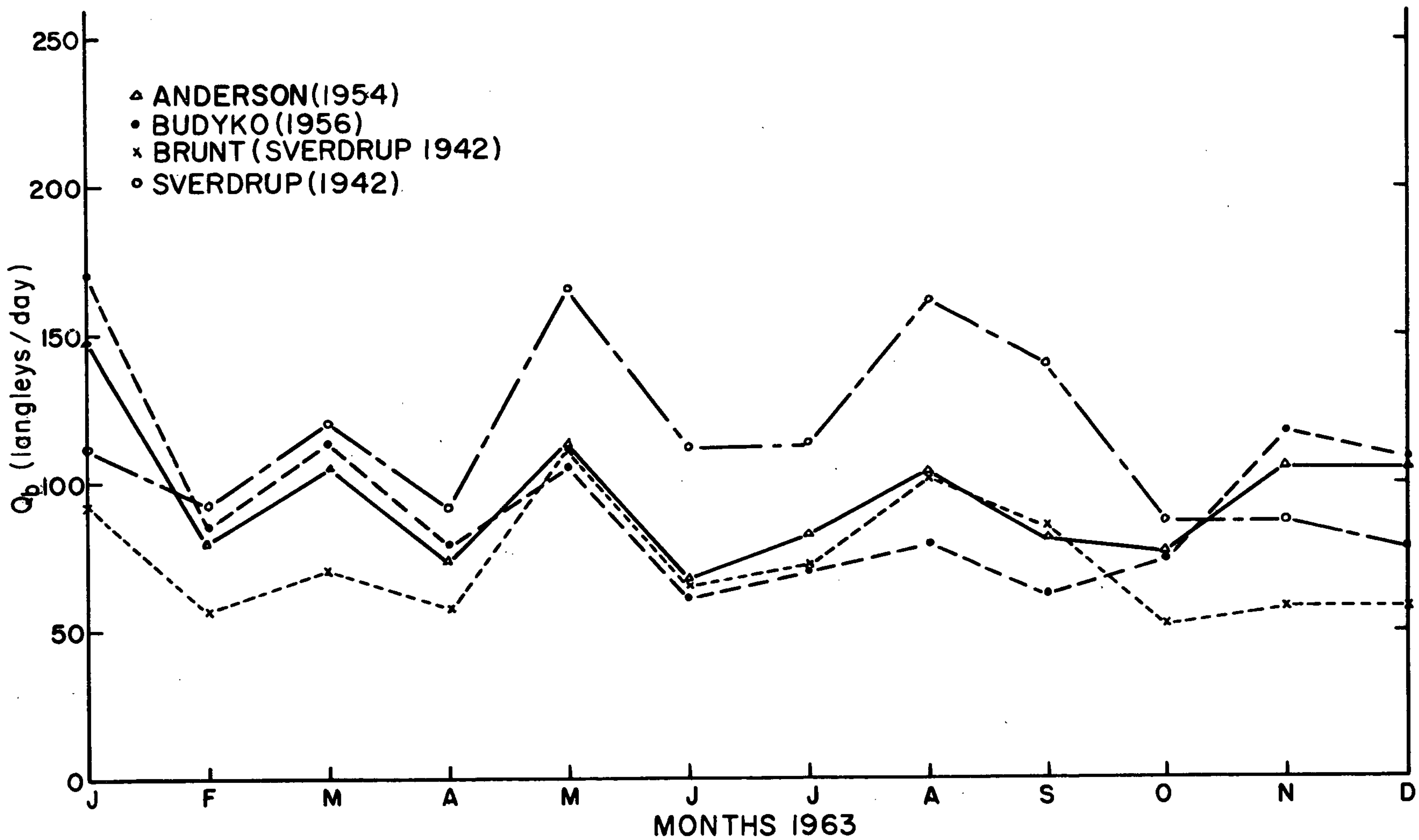


Fig. 7: Comparison of Long Wave Radiation Formulae. Data from Southern Strait of Georgia.

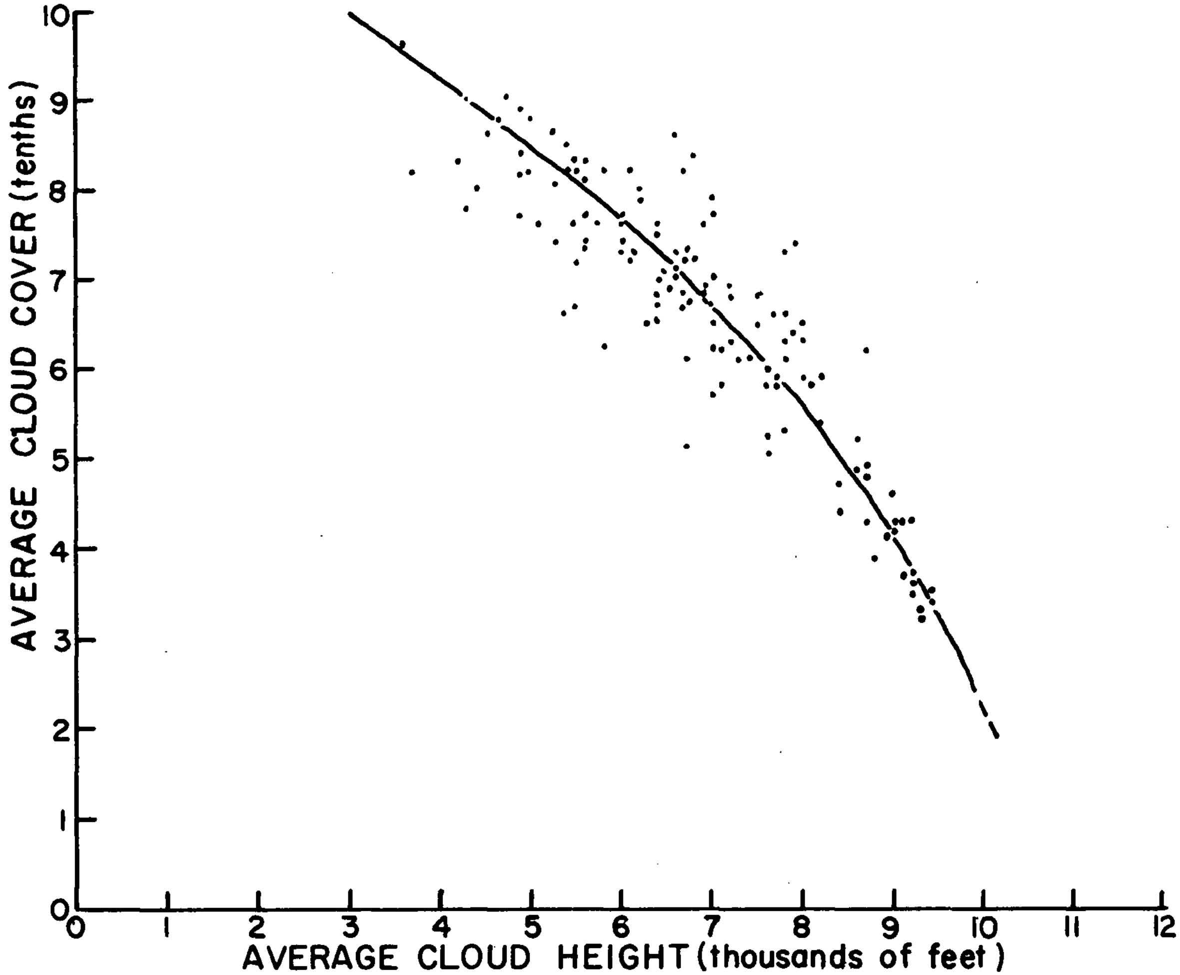


Fig. 8: Average Cloud Height for a given Cloud Cover, for Coastal British Columbia and S.E. Alaska.

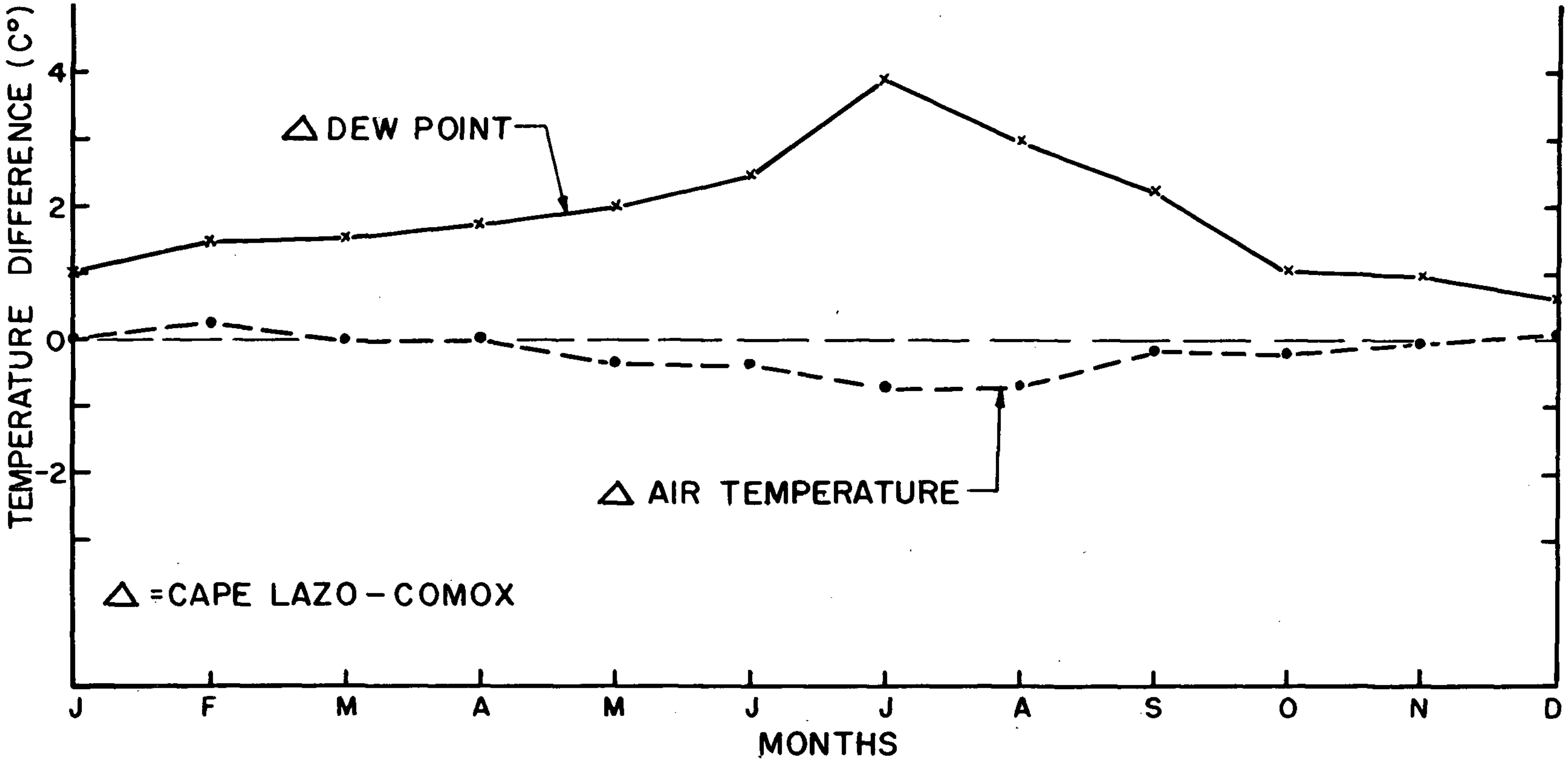


Fig. 9: Average Air Temperature and Dew Point Difference between Comox and Cape Lazo, British Columbia.

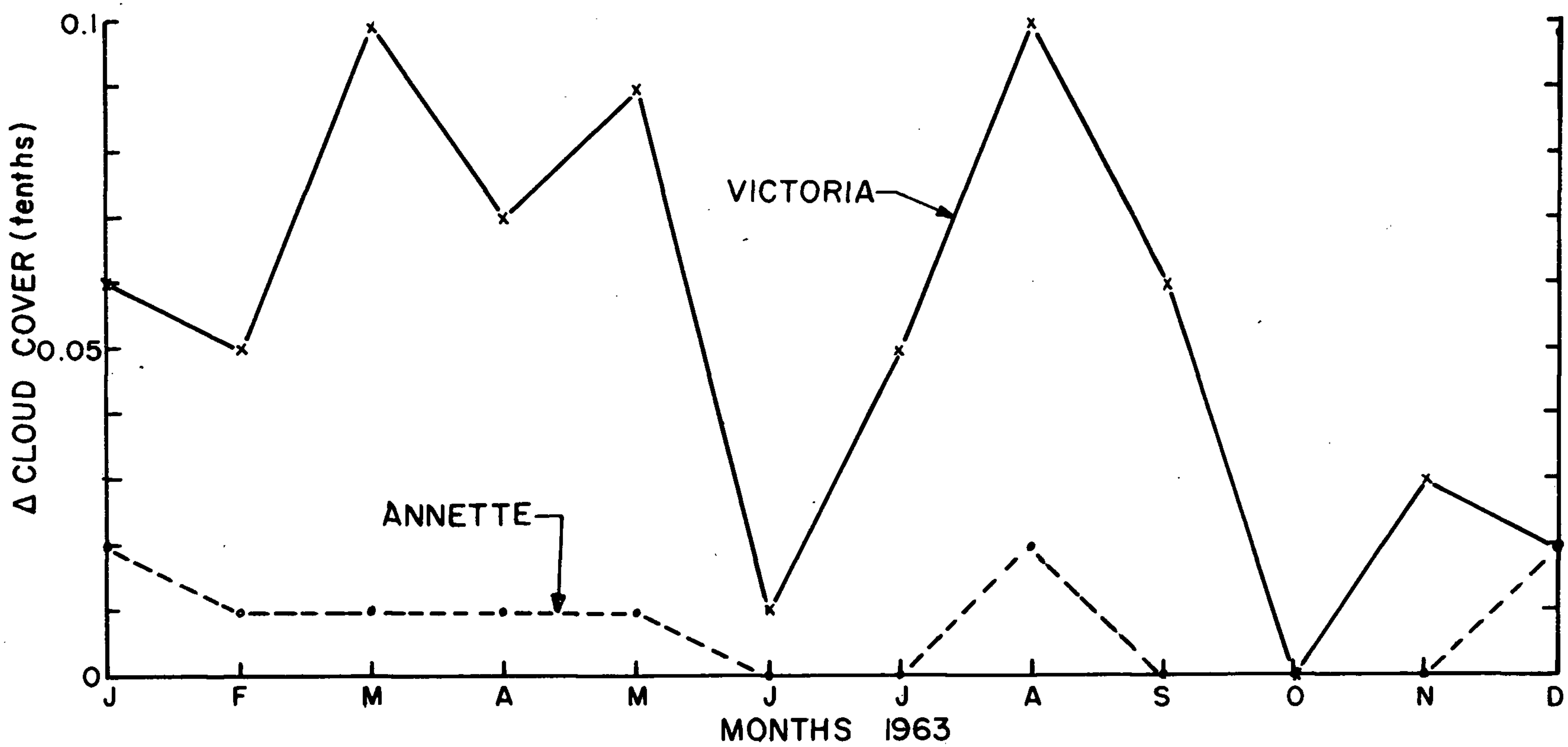


Fig. 10: Average Diurnal Change in Cloud Cover, Daytime - All Day.

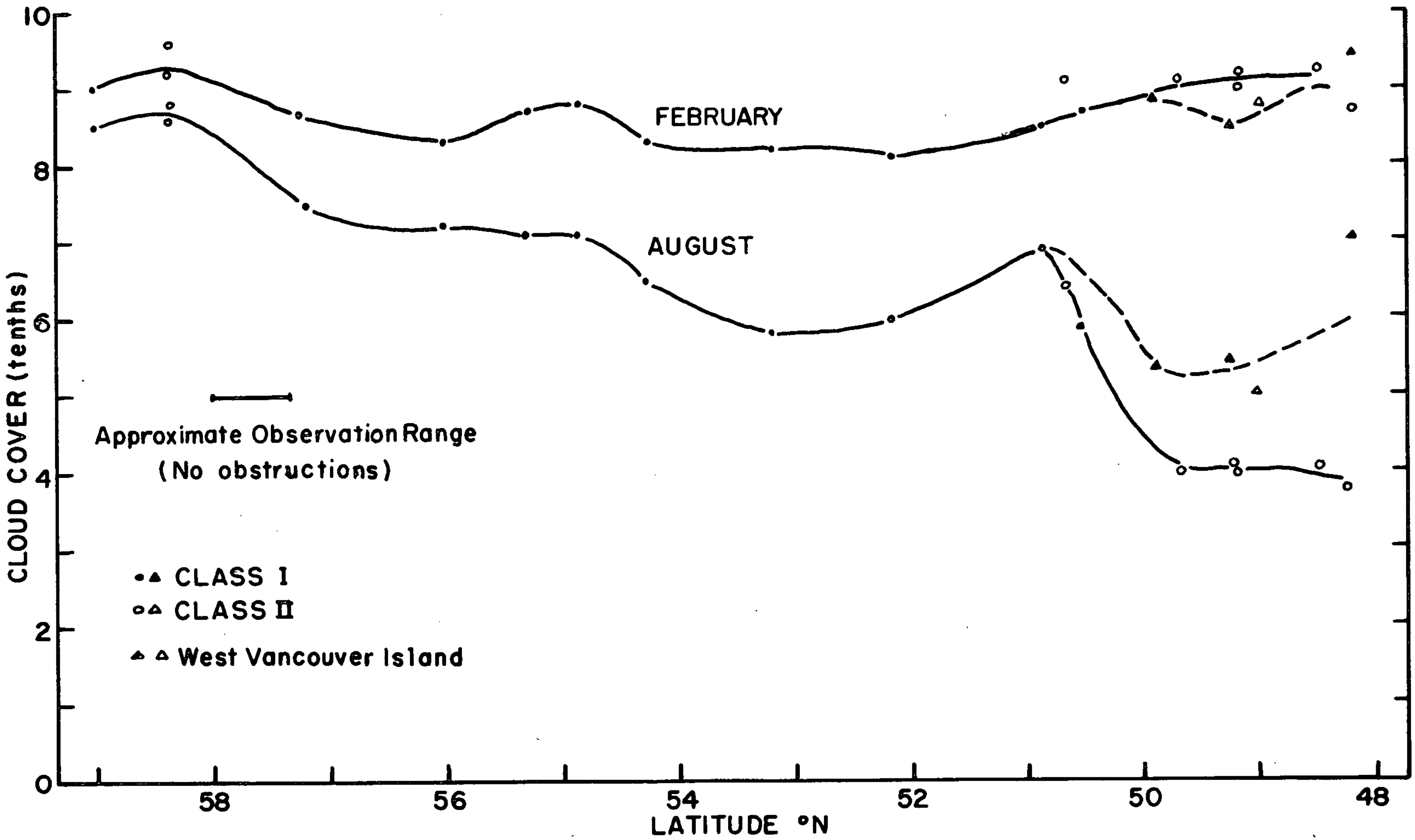


Fig. 11: Scatter Diagram of Mean Monthly Cloud Cover at Coastal Stations, 1961.

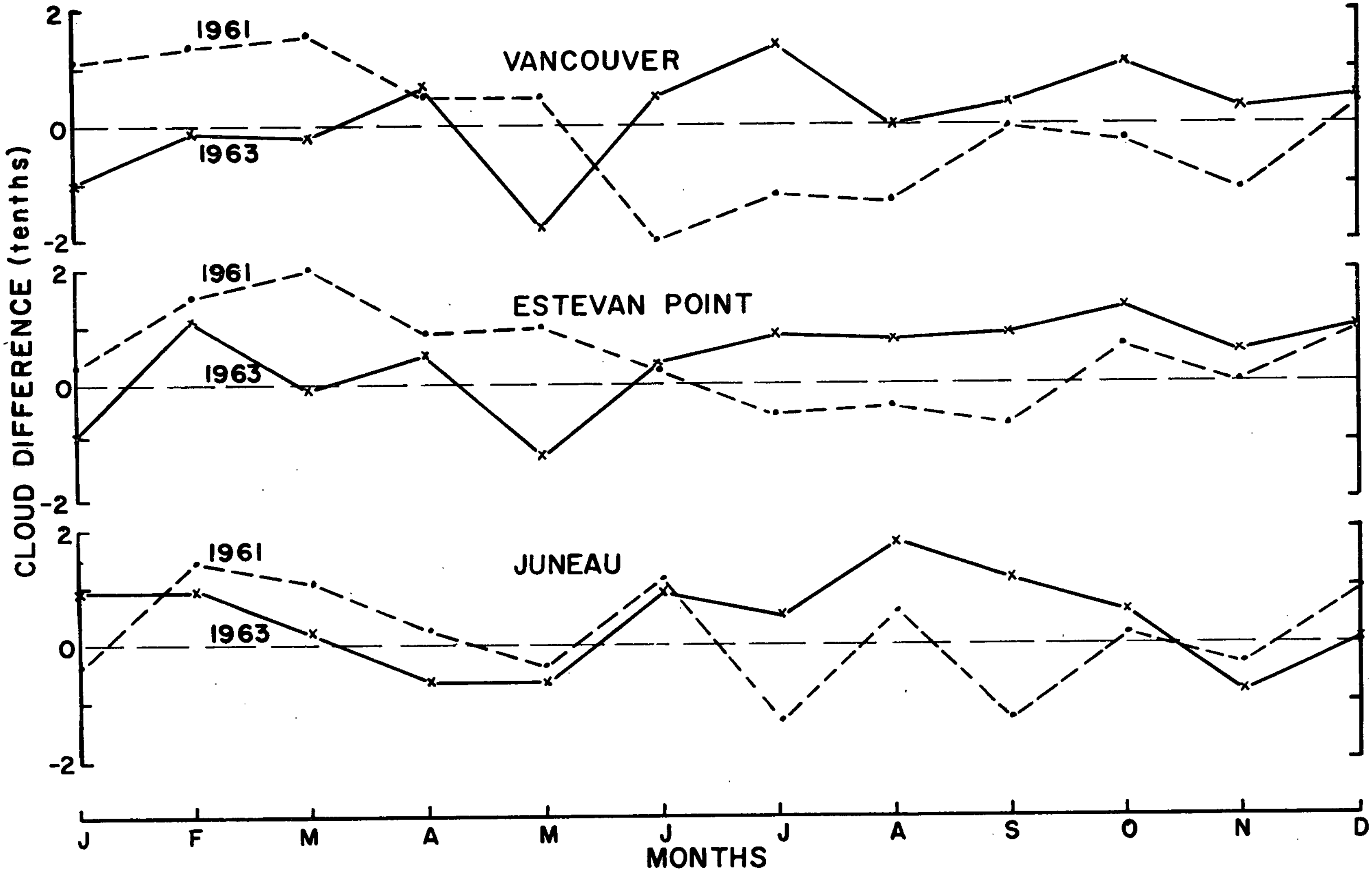


Fig. 12: Comparison of Mean Cloud Cover for 1961 and 1963 with a long term mean.

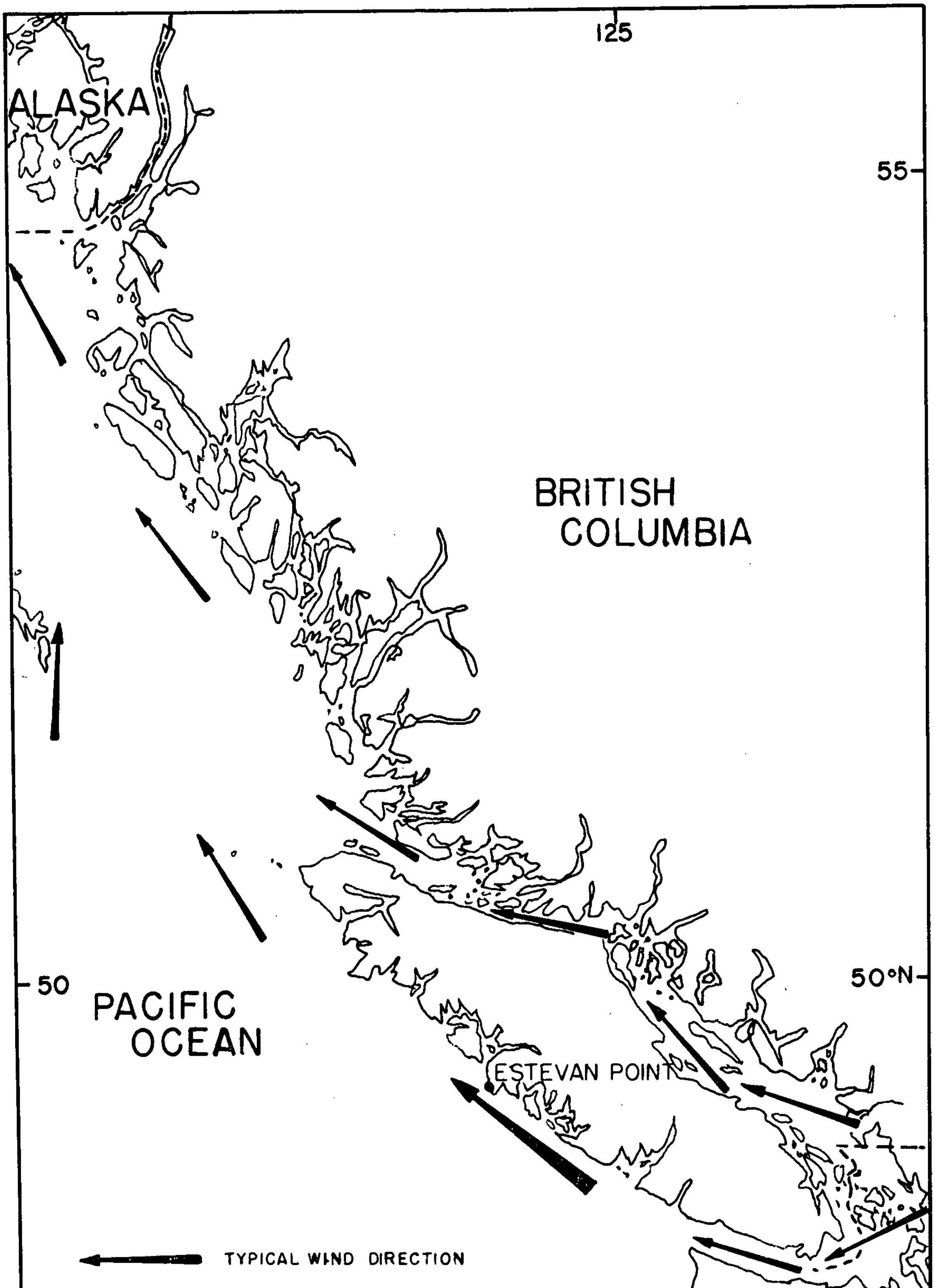


Fig.13A: Coastal Wind Pattern for SE'lies at Estevan Point, south coast.

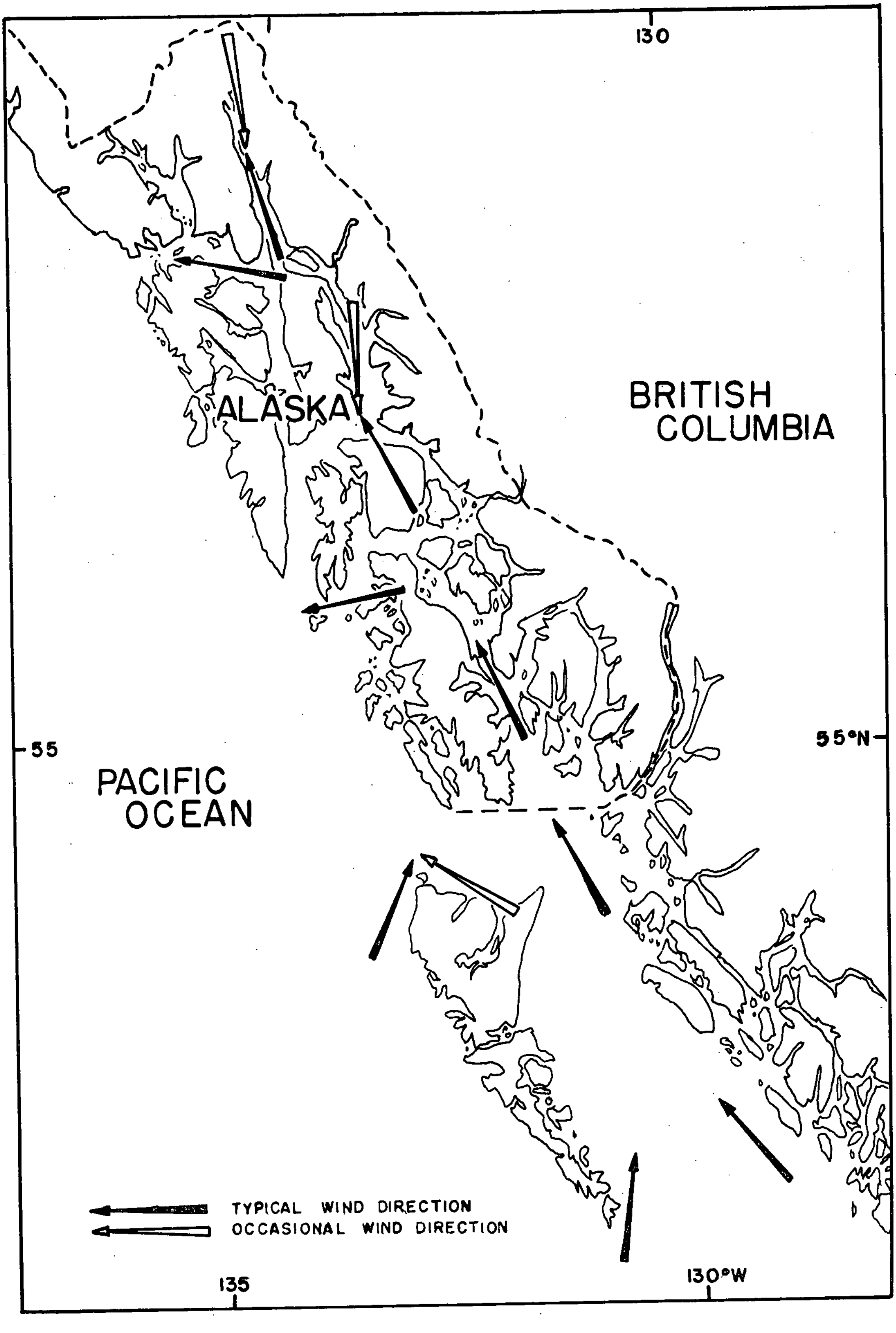


Fig. 13B: Coastal Wind Pattern for SE'lies at Estevan Point, north coast.

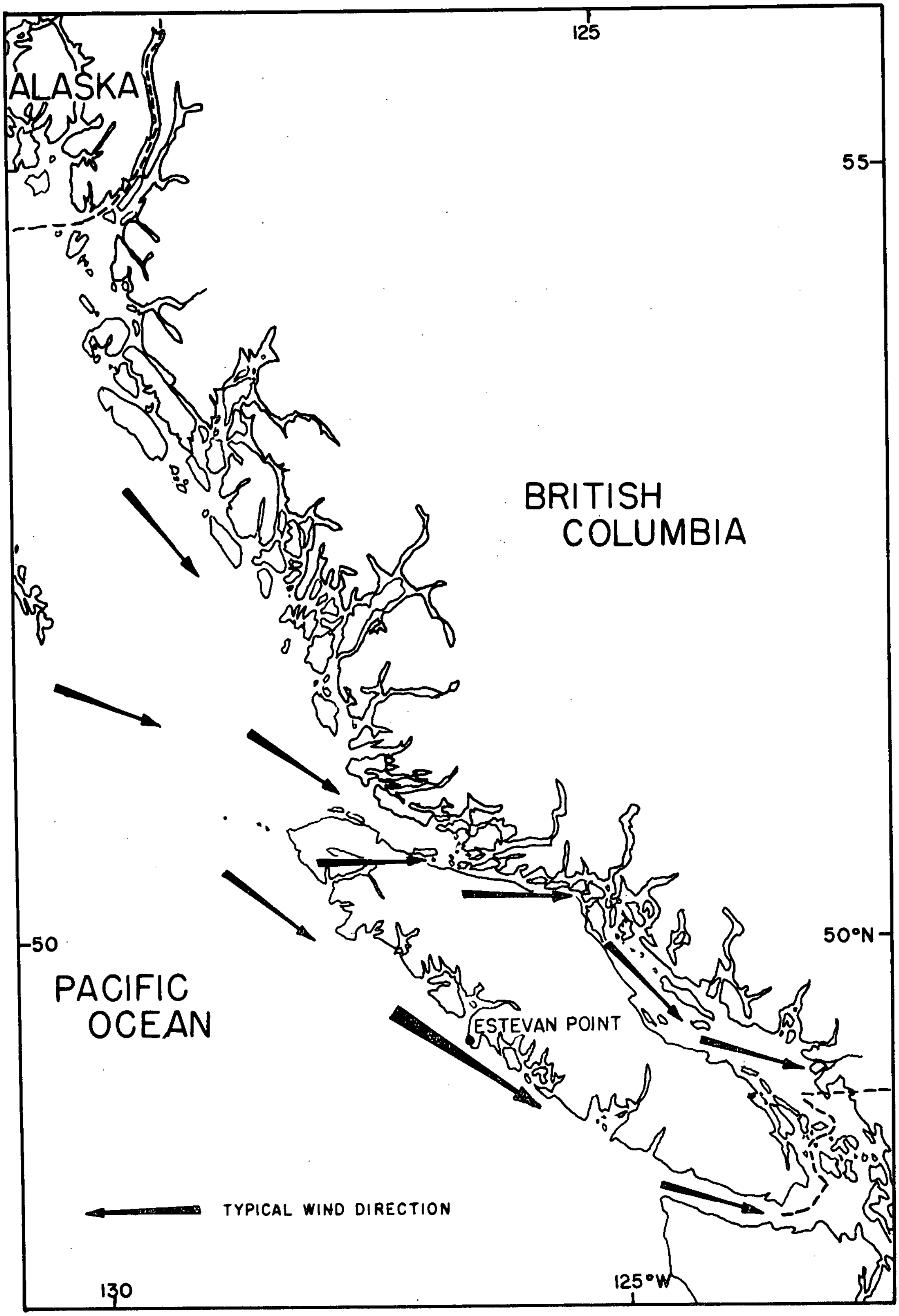


Fig. 13C: Coastal Wind Pattern for NW'lies at Estevan Point, south coast.

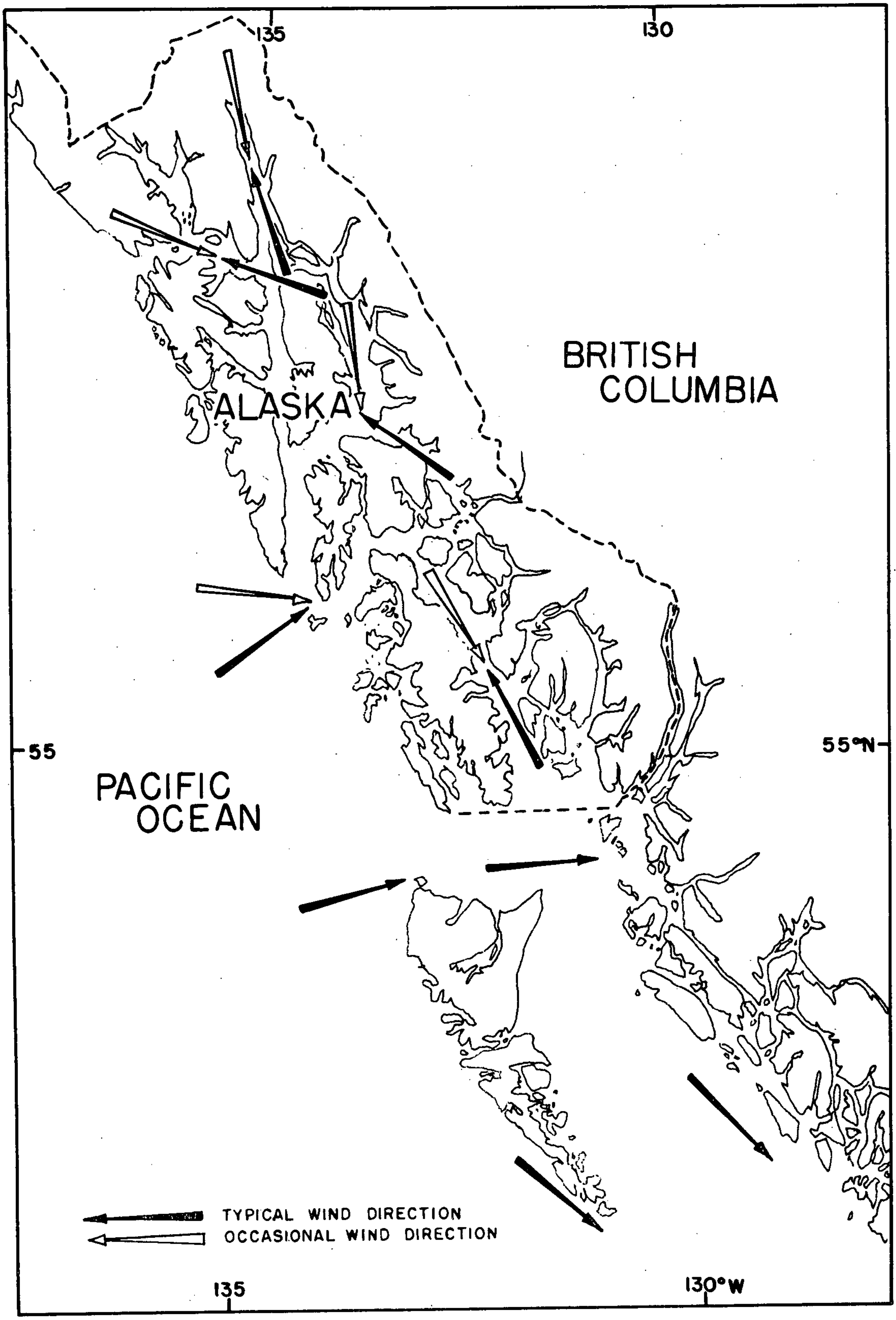


Fig. 13D: Coastal Wind Pattern for NW'lies at Estevan Point, north coast.

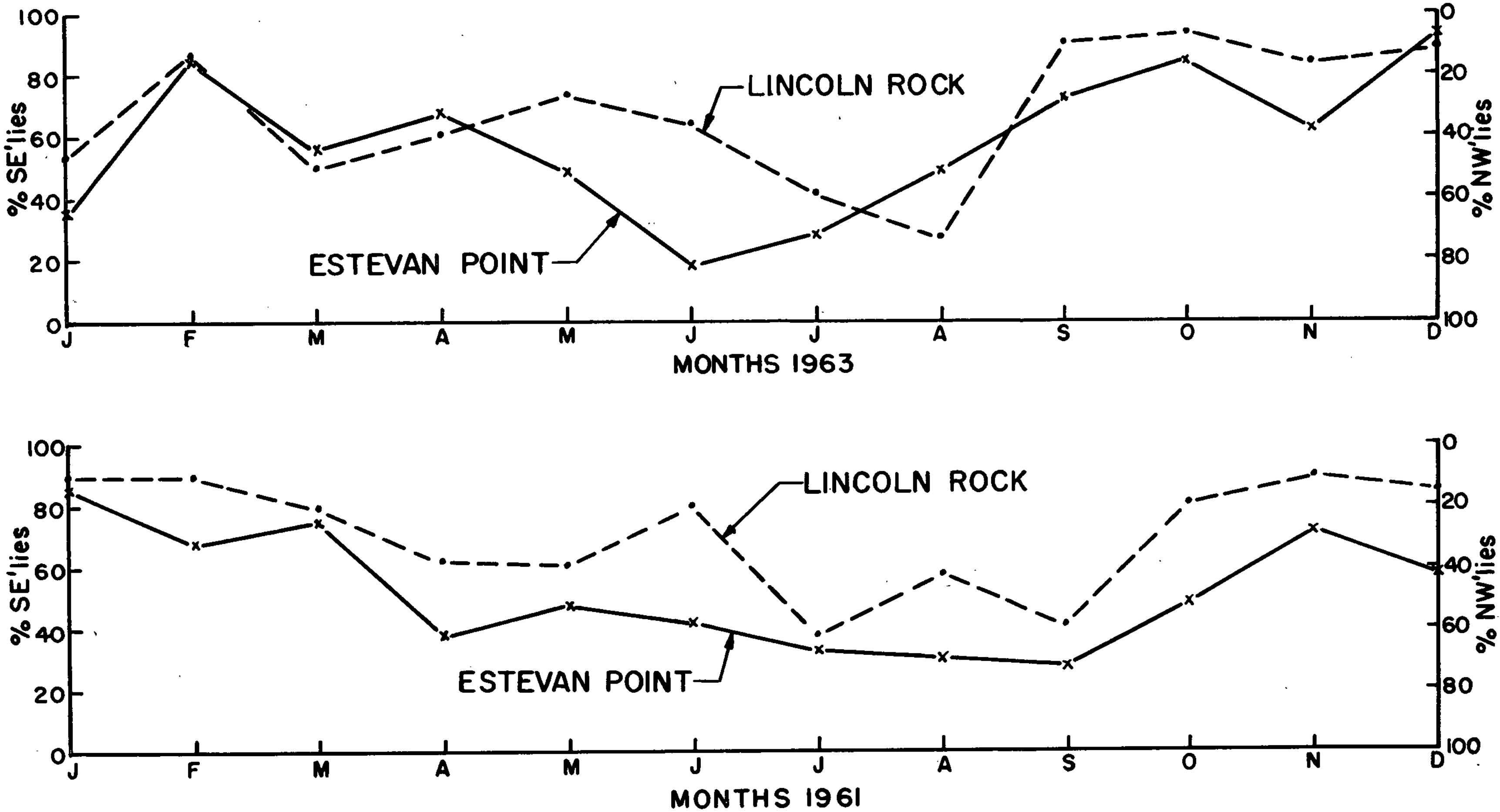


Fig. 14: Monthly percentage of miles of wind by direction at Estevan Point and Lincoln Rock for 1961 and 1963.

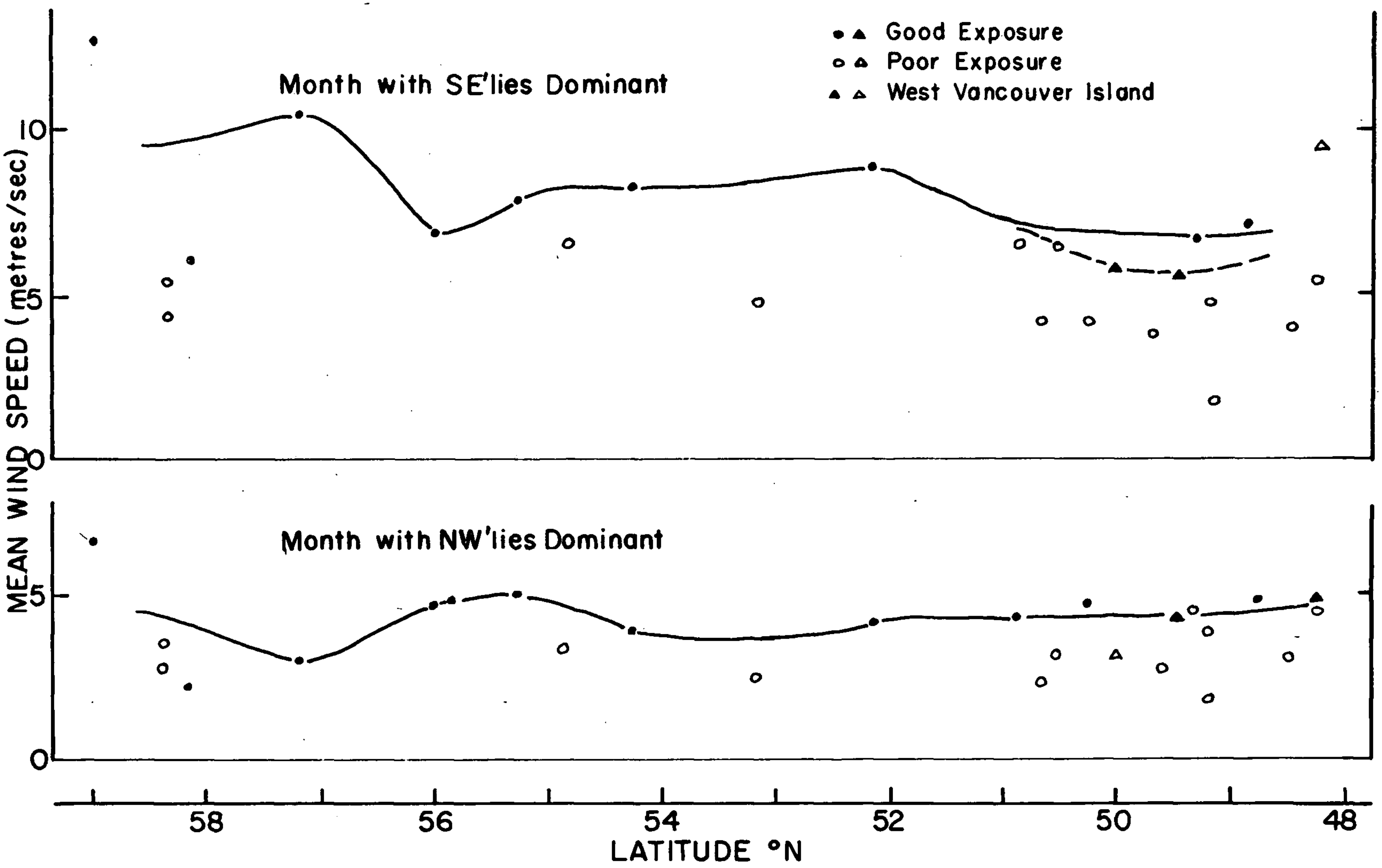


Fig. 15: Scatter Diagram of Monthly Mean Wind Speed at Coastal Stations.

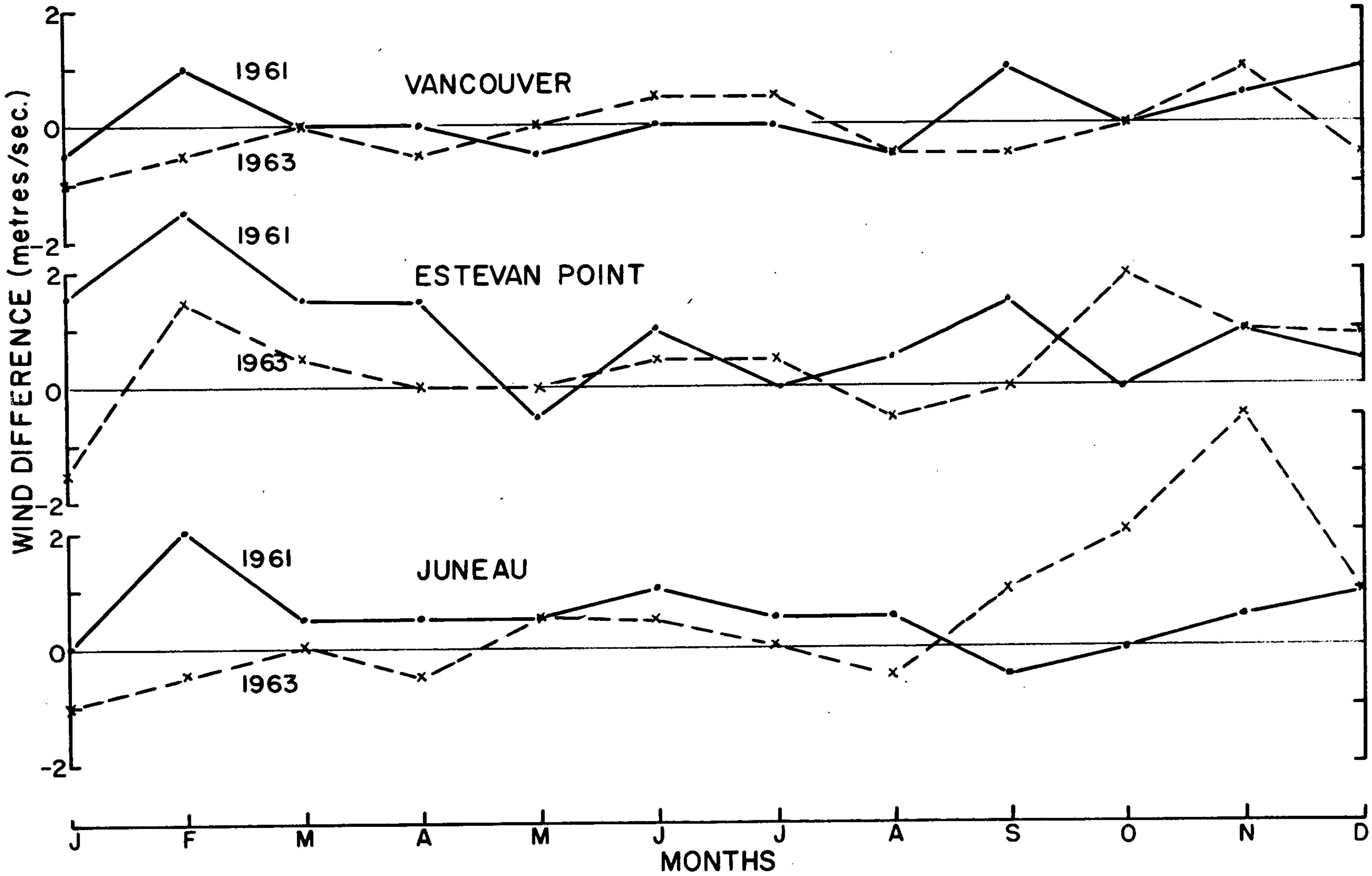


Fig. 16: Comparison of Mean Wind Speed for 1961 and 1963 with a long term mean.

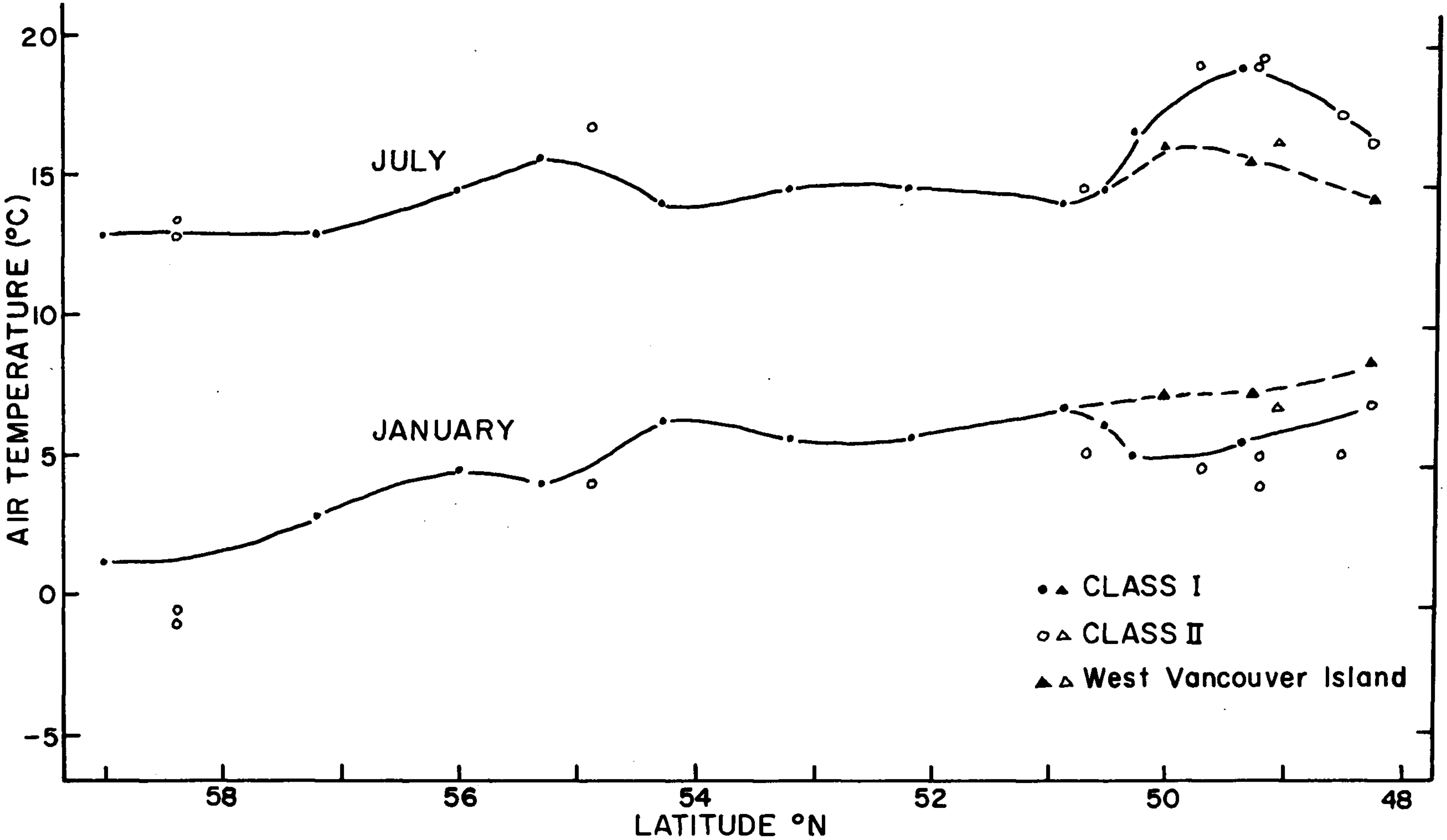


Fig. 17: Scatter Diagram of Monthly Mean Air Temperature at Coastal Stations, 1961.

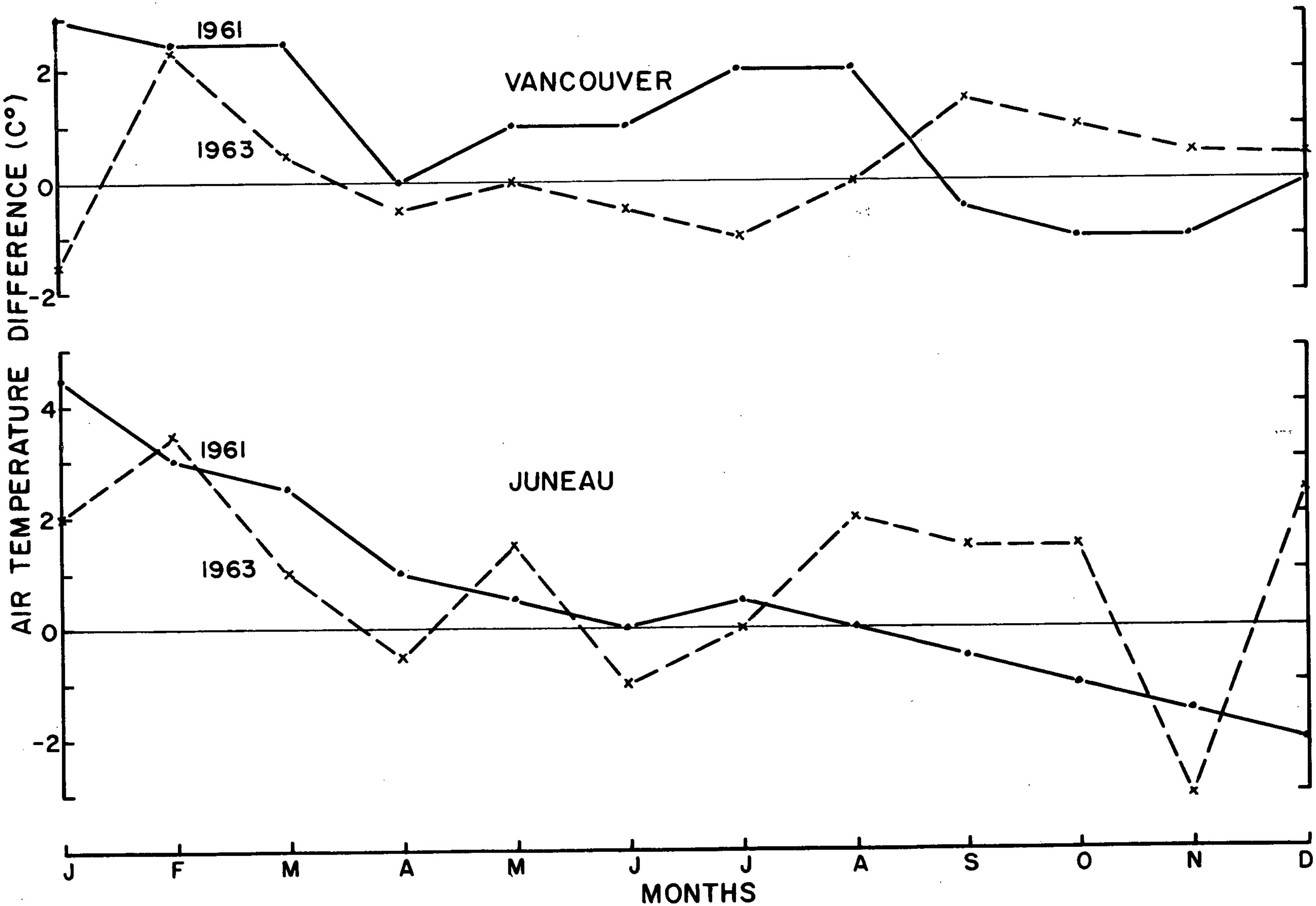


Fig. 18: Comparison of Mean Air Temperature for 1961 and 1963 with a long term mean.

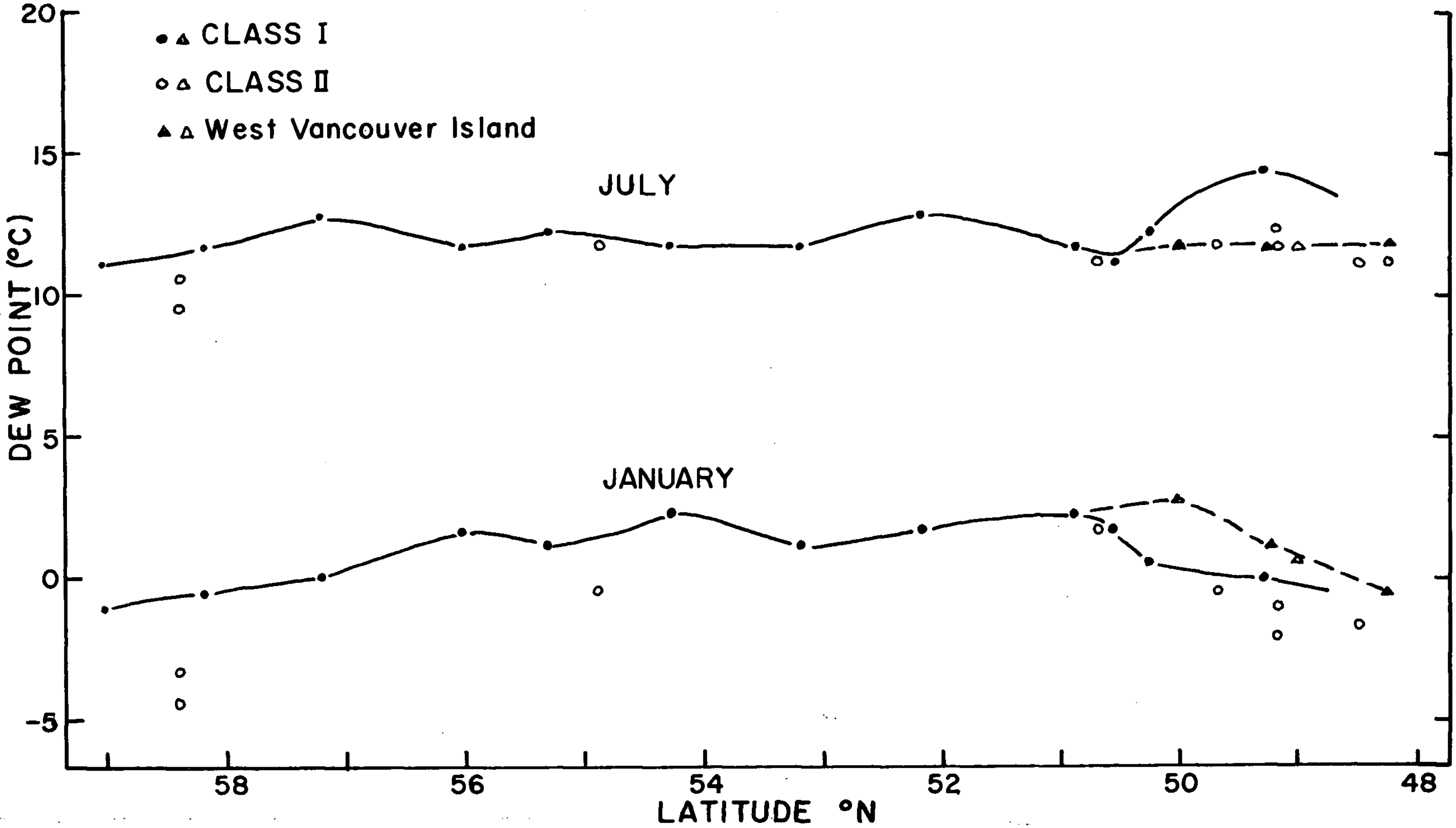


Fig. 19: Scatter Diagram of Monthly Mean Dew Point at Coastal Stations, 1963.

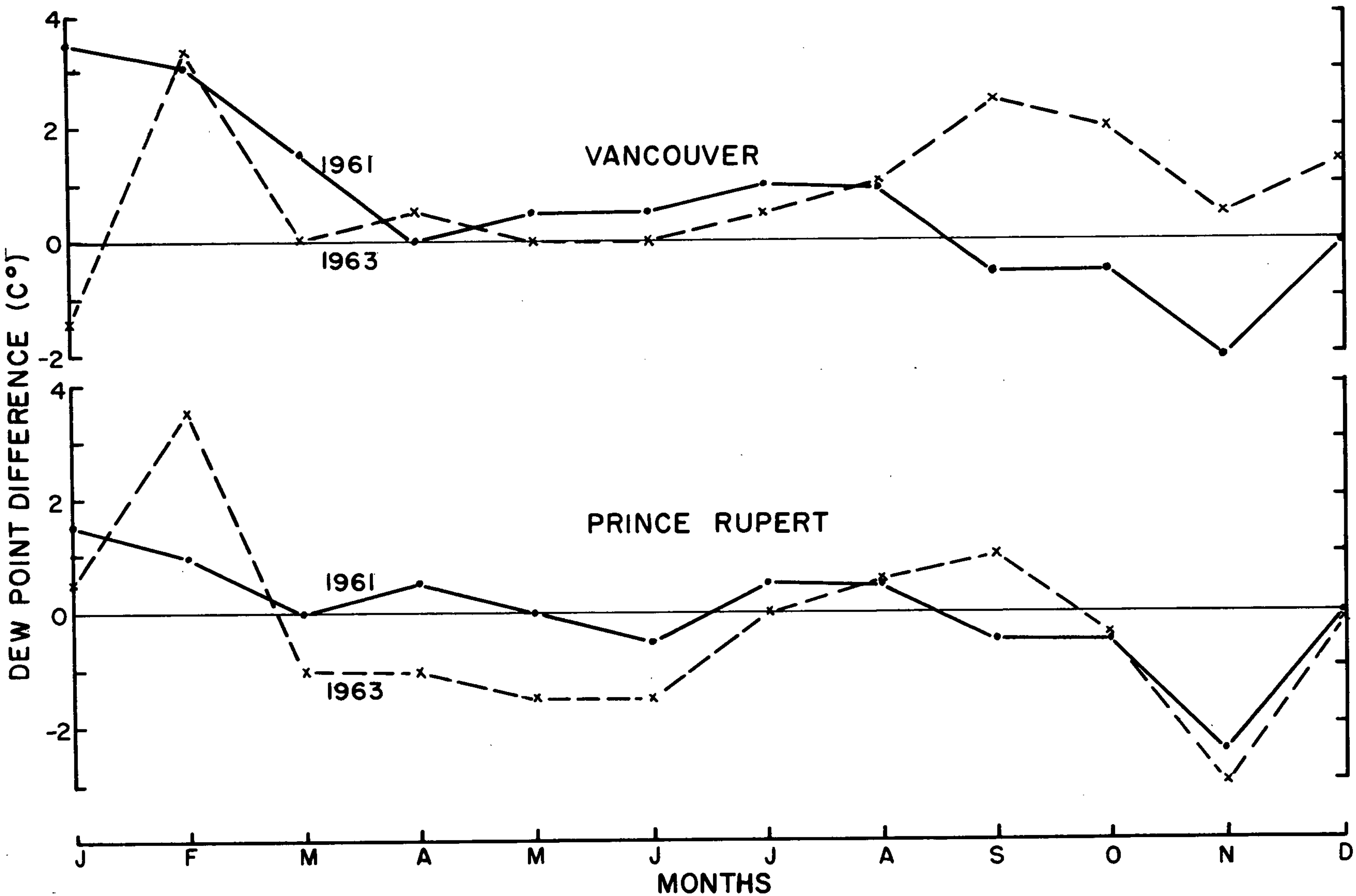


Fig. 20: Comparison of Mean Dew Point for 1961 and 1963 with a Long Term Mean.

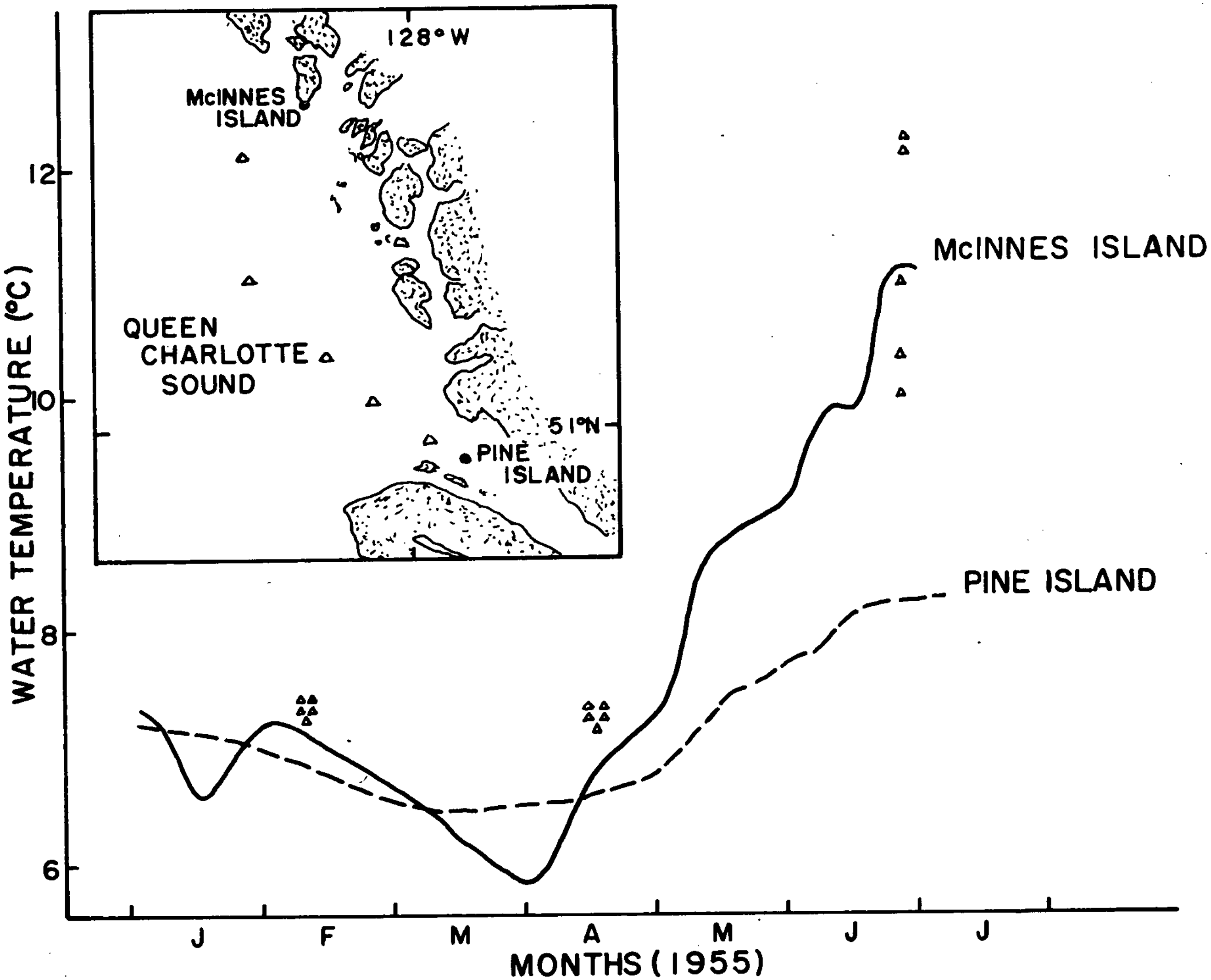


Fig. 21: Comparison of Shore Observations of Sea Surface Temperature with Cruise Data, Queen Charlotte Sound.

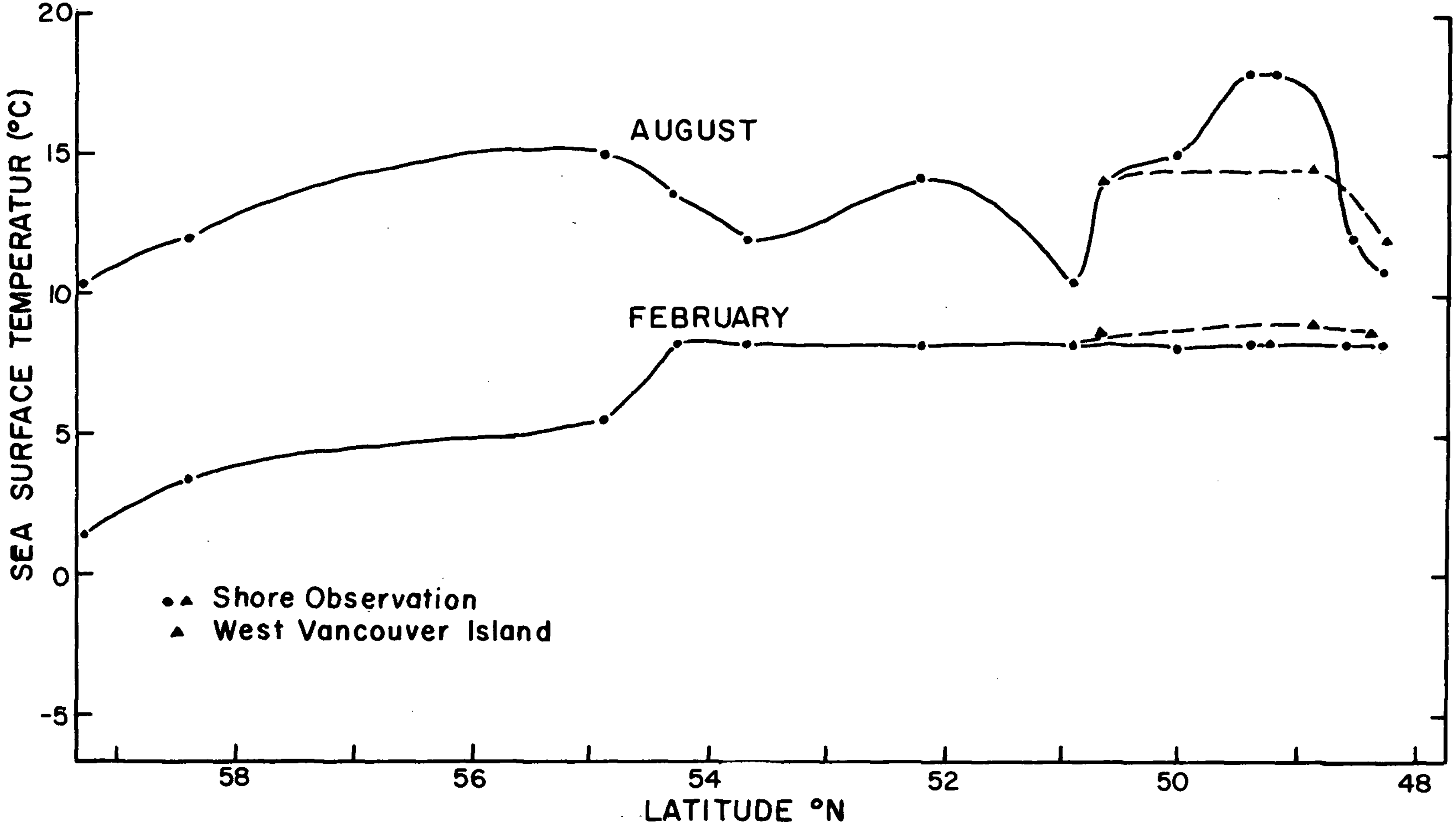


Fig. 22: Scatter Diagram of Monthly Mean Sea Surface Temperature at Coastal Shore Stations.

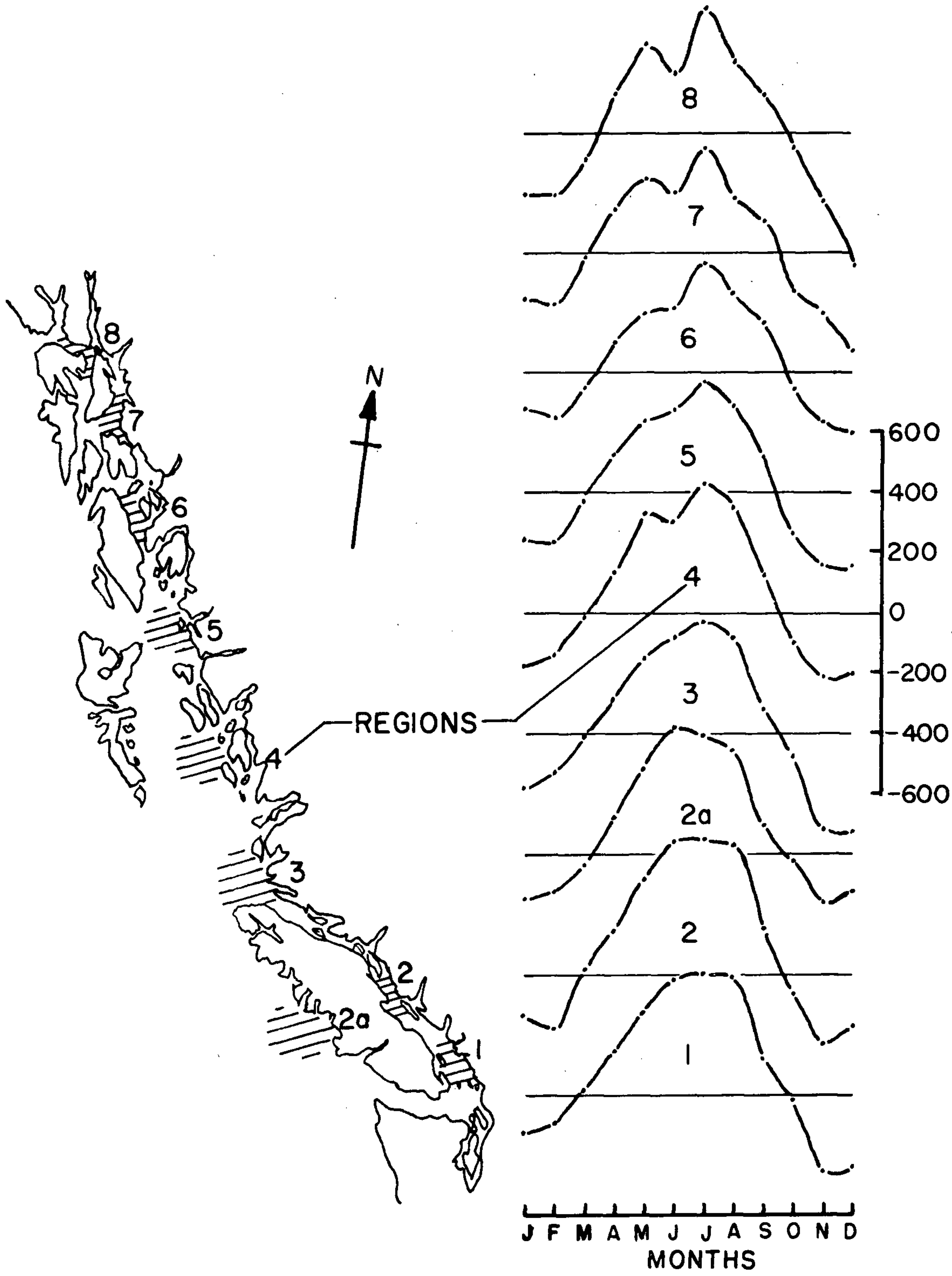


Fig. 23: Yearly Cycle of the Calculated Monthly Mean Surface Heat Flux, 1961.

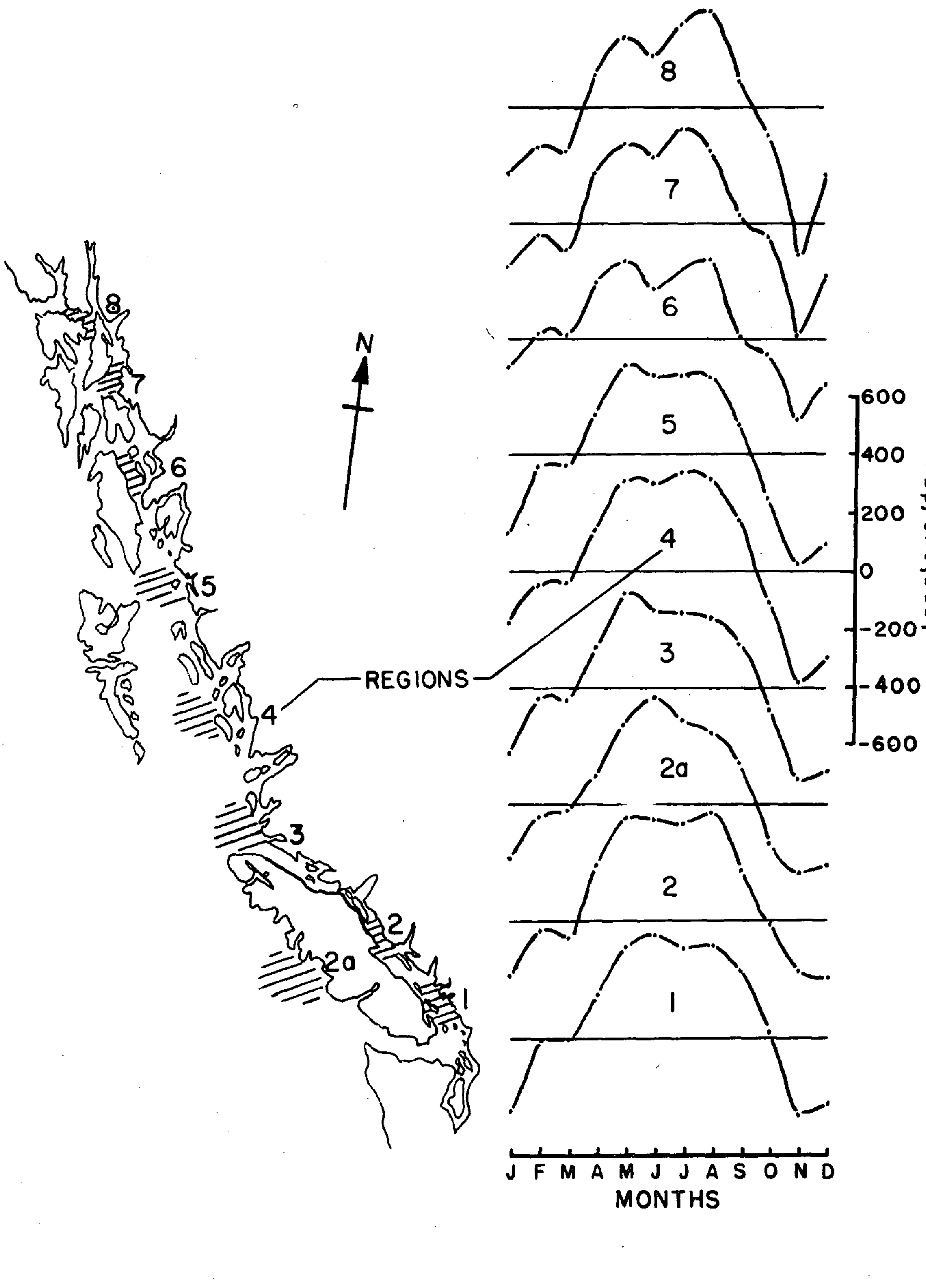


Fig. 24: Yearly Cycle of the Calculated Monthly Mean Surface Heat Flux, 1963.

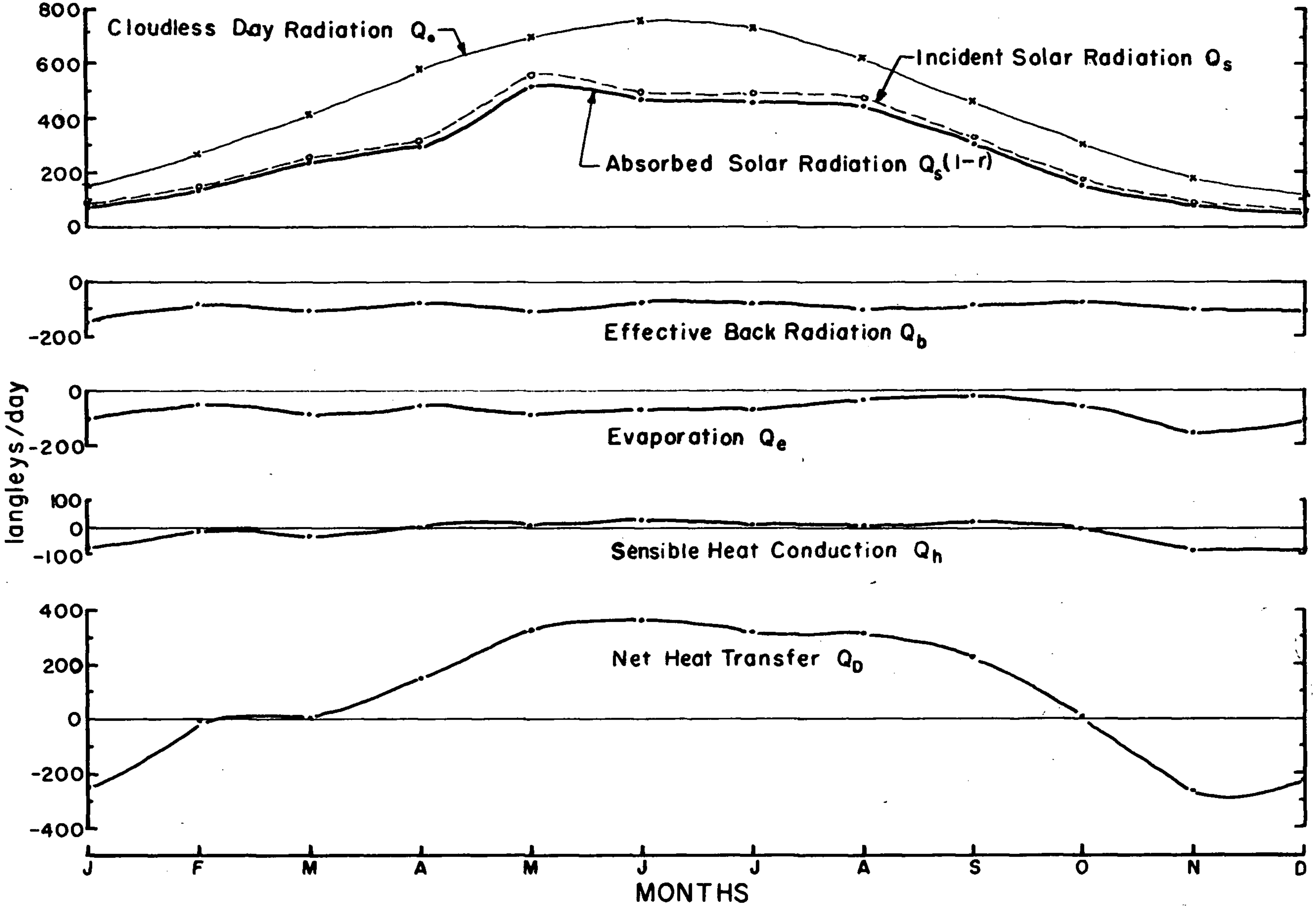


Fig. 25: Annual Cycle of the Monthly Mean for Components of the Surface Heat Transfer, Region 1, 1963.

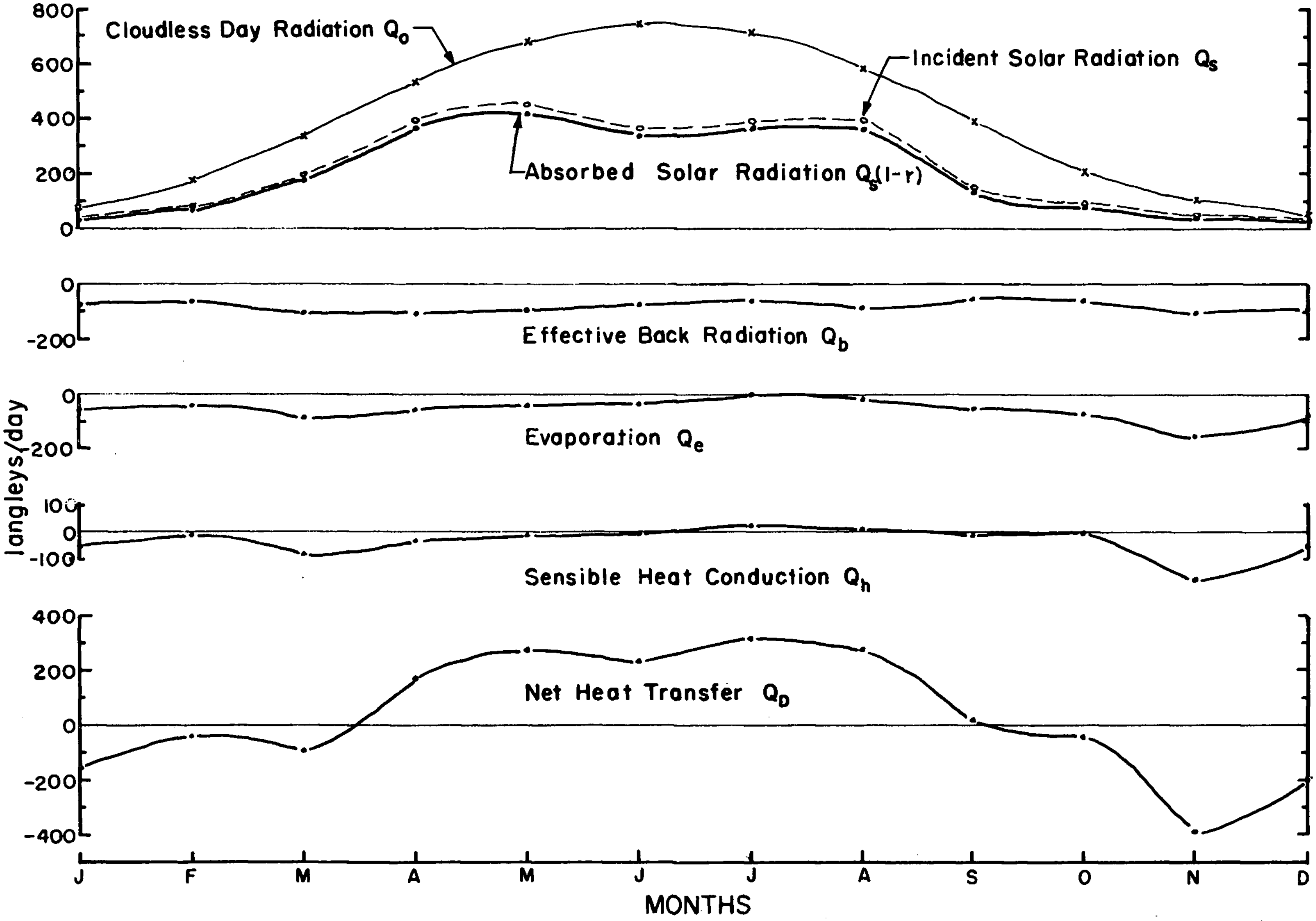


Fig. 26: Annual Cycle of the Monthly Mean for Components of the Surface Heat Transfer, Region 7, 1963.

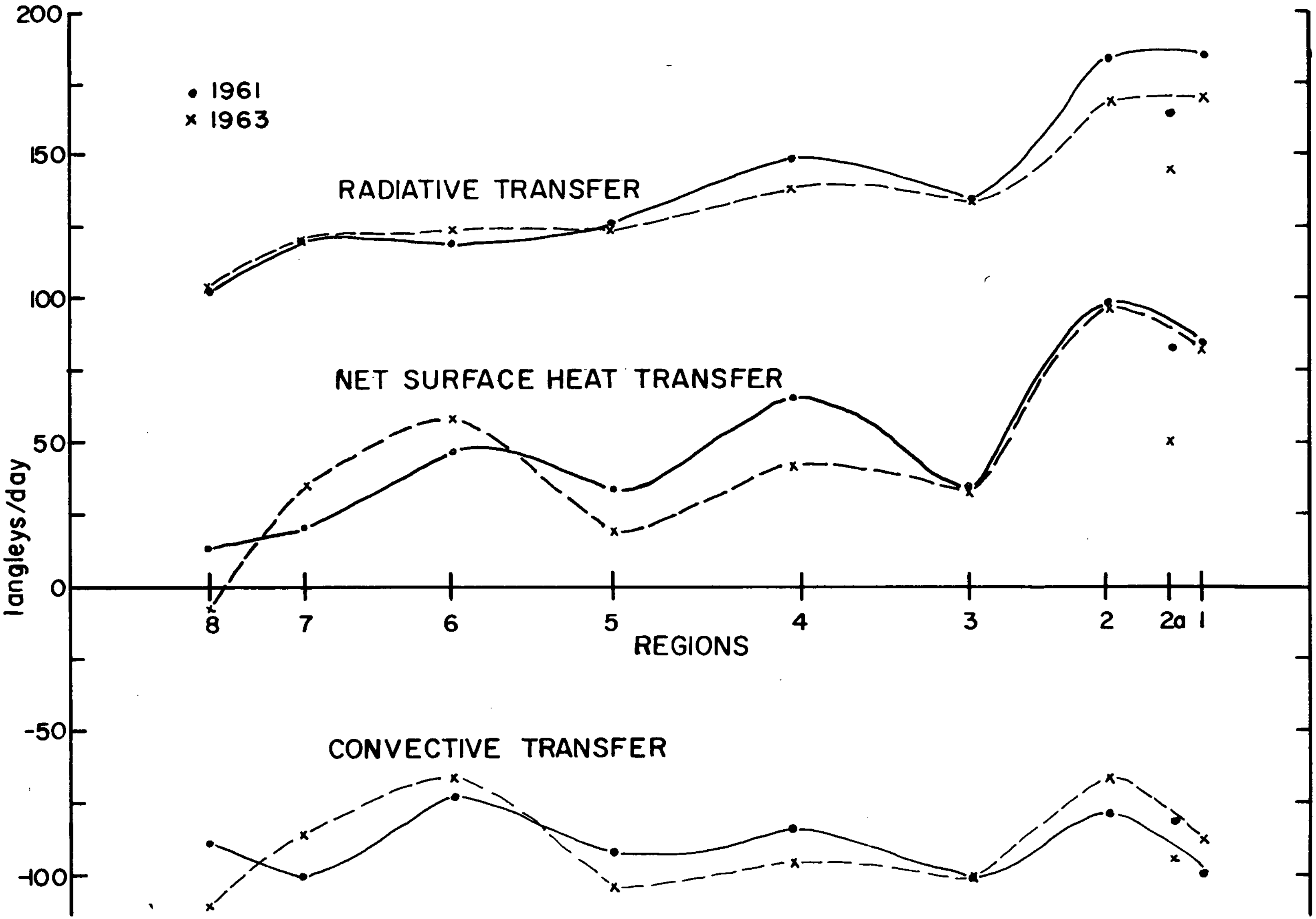


Fig. 27: Latitudinal Change of Mean Annual Surface Heat Transfer, Radiation Transfer, and Convective Transfer, 1961 and 1963.

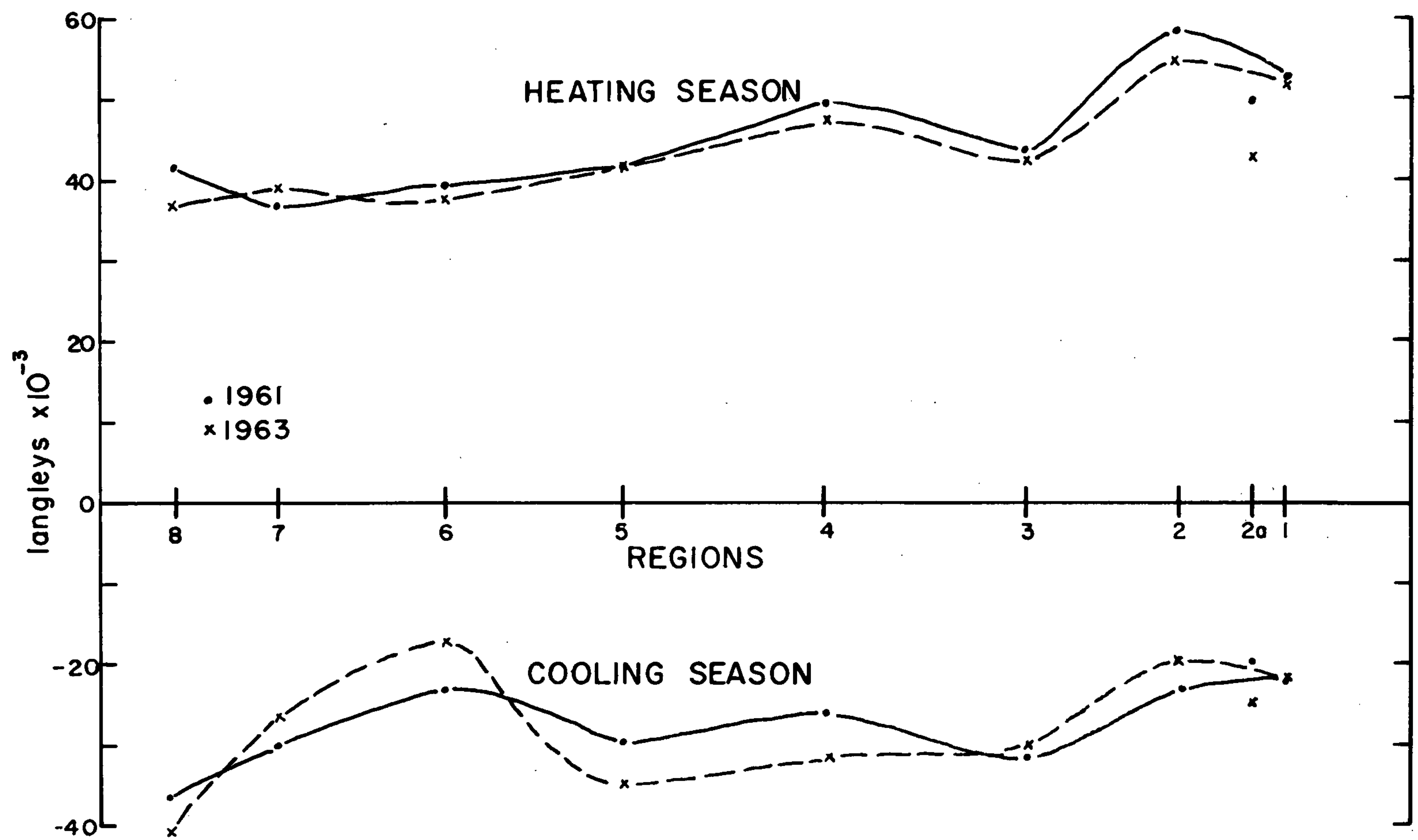


Fig. 28: Total Heat Transfer during the Heating and Cooling Season, 1961 and 1963.

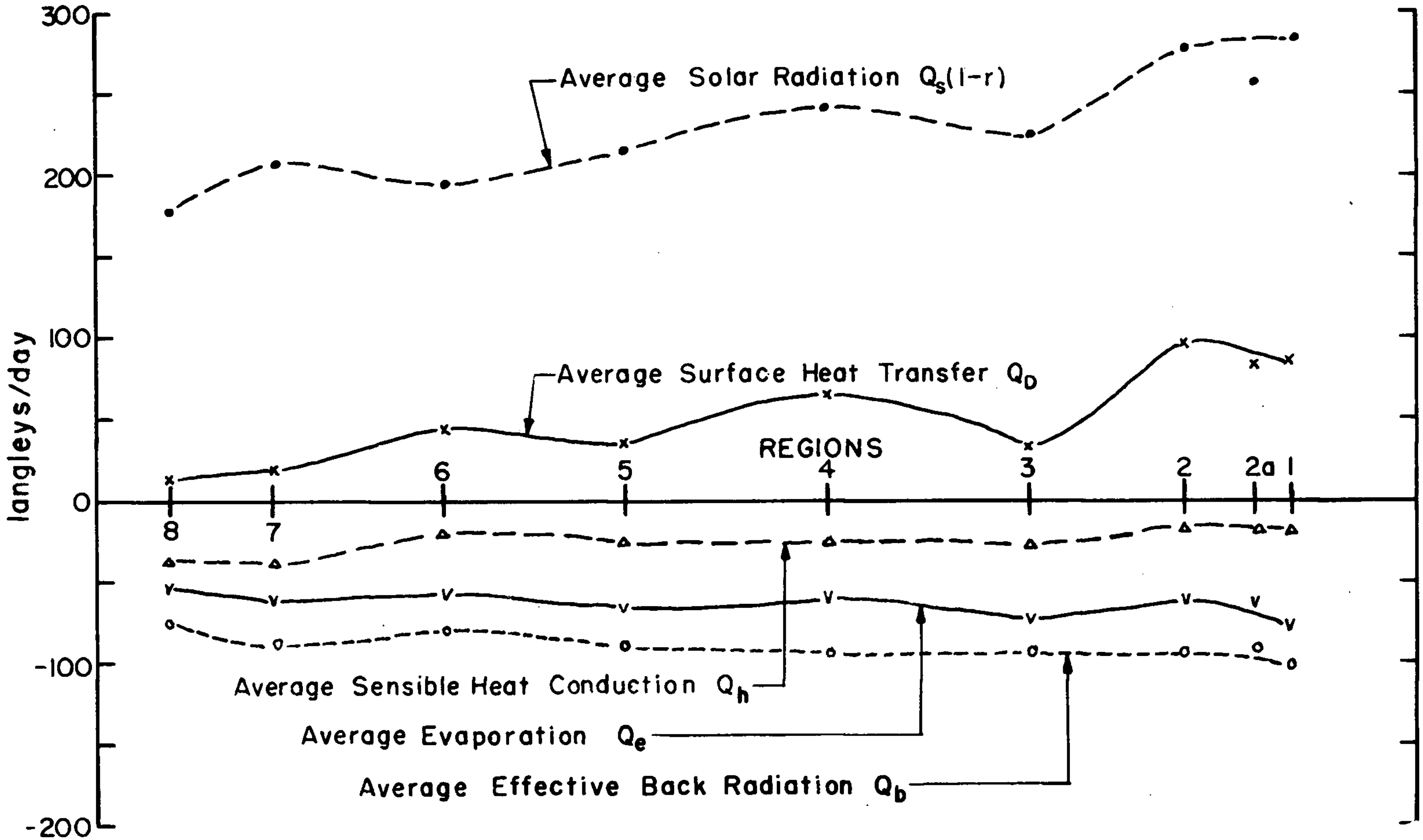


Fig. 29: Latitudinal Change of Annual Mean for Components of the Surface Heat Transfer, 1961.

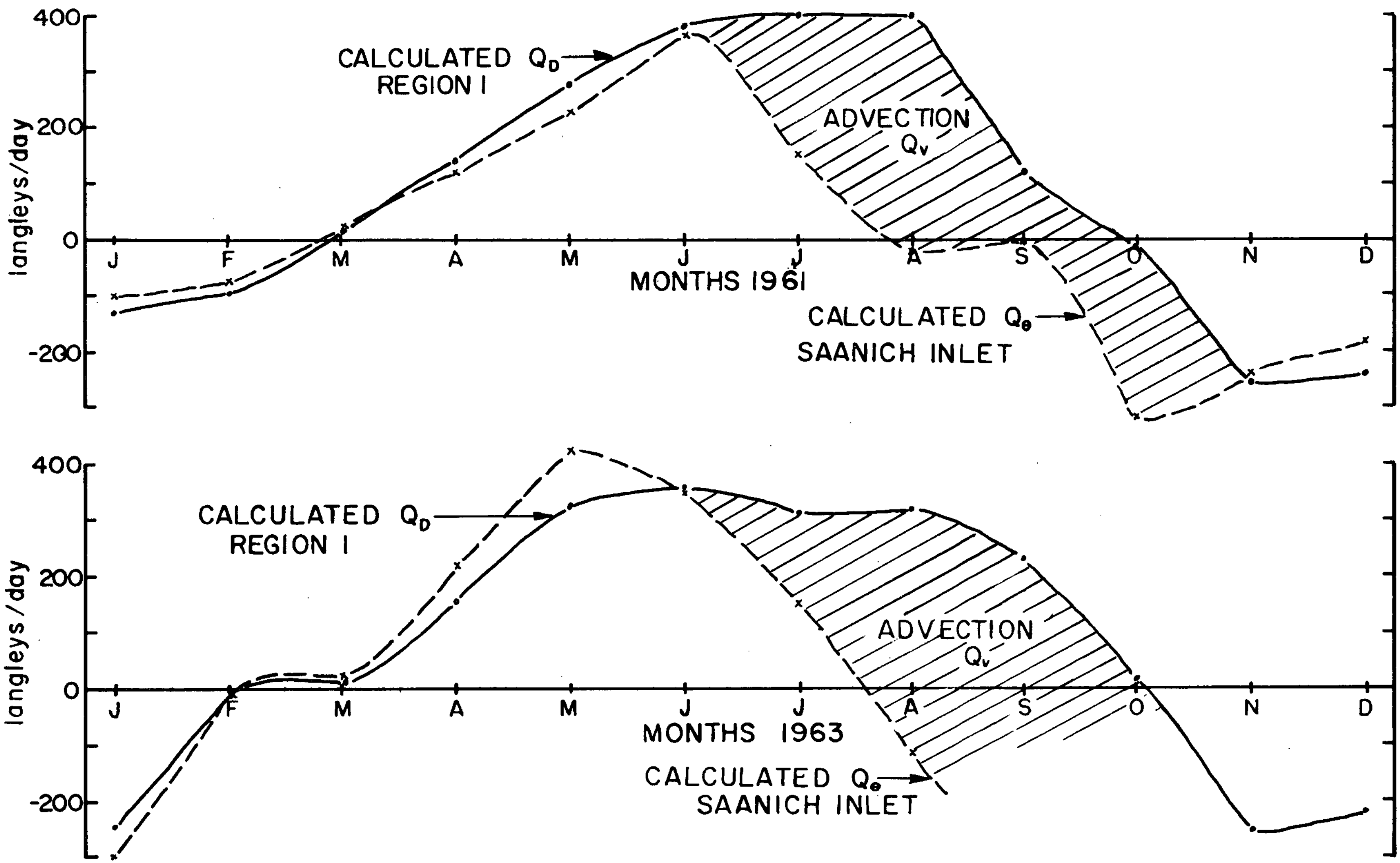


Fig. 30: Balance of Heat Budget Components in Saanich Inlet, 1961 and 1963.

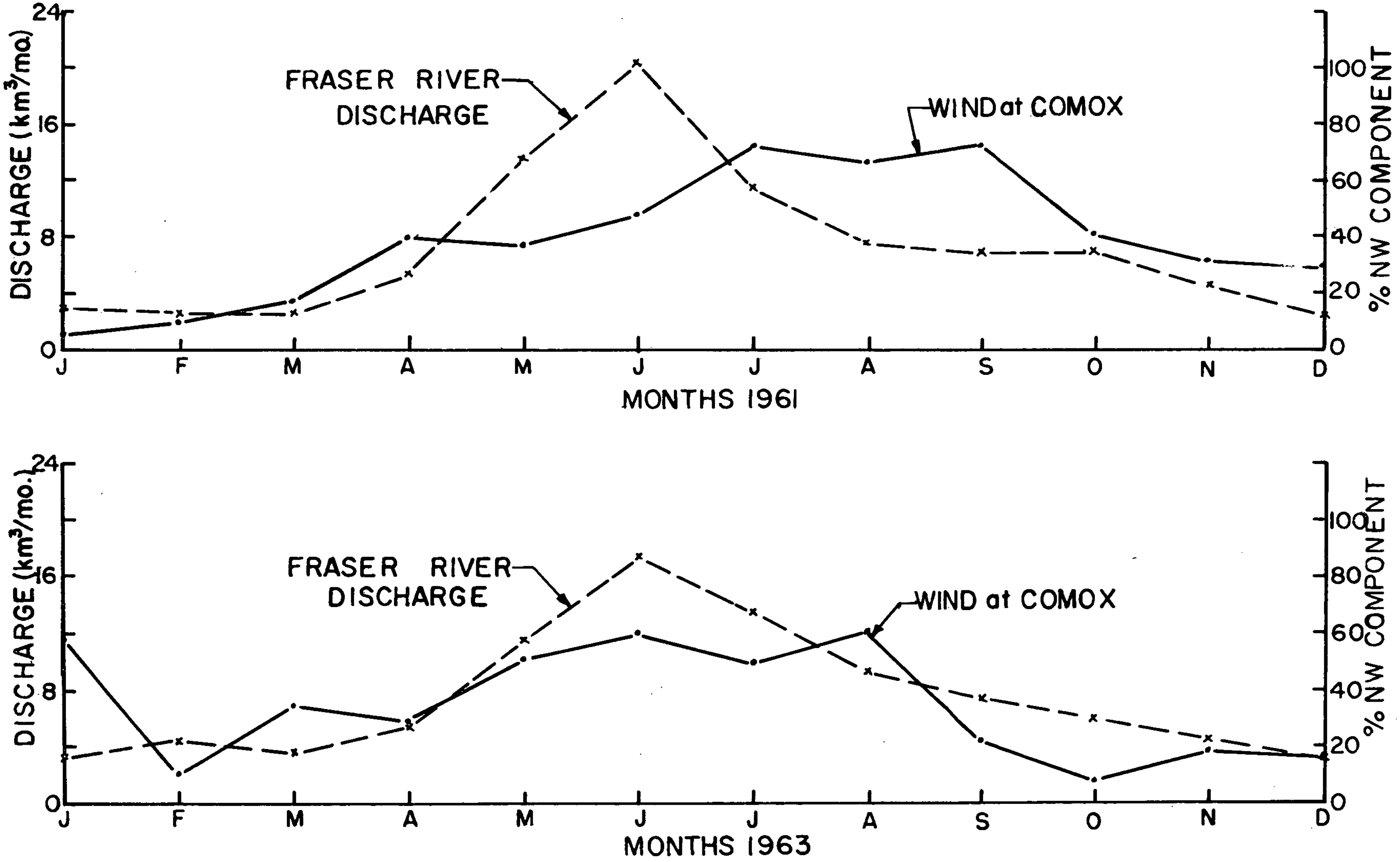


Fig. 31: Fraser River Discharge and Wind Direction at Comox, 1961 and 1963.

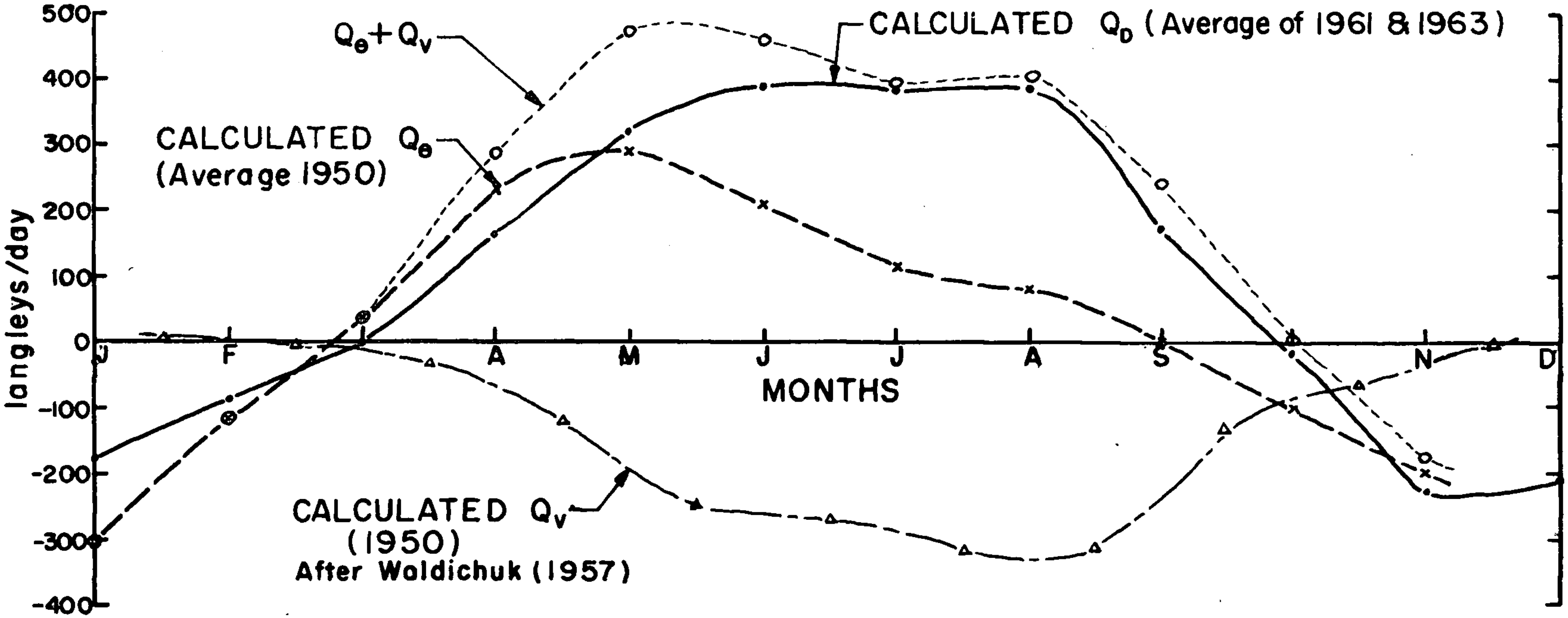


Fig. 32: Comparison of Heat Storage in the Strait of Georgia, 1950; with Average Surface Heat Transfer, 1961 and 1963.

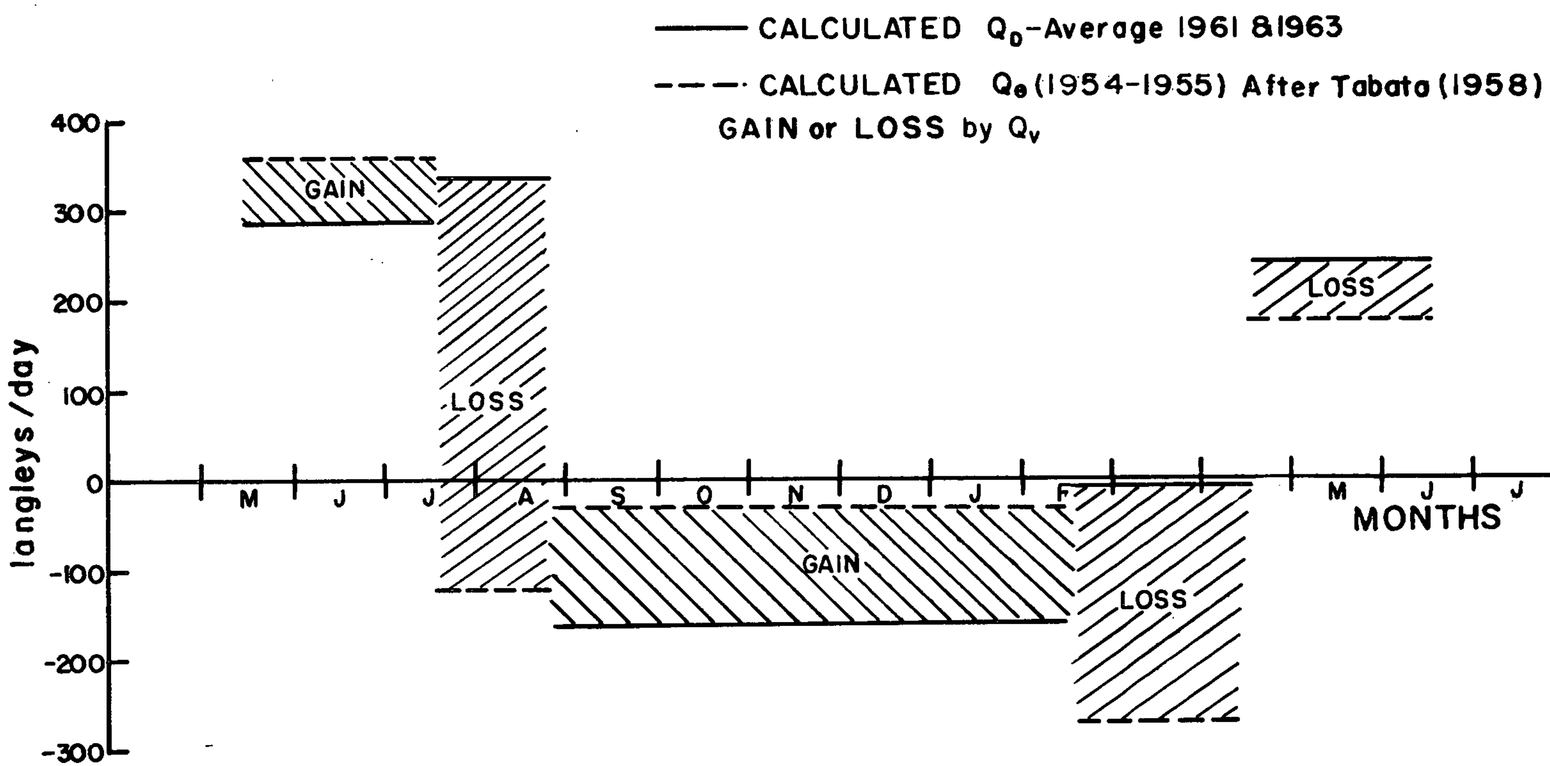
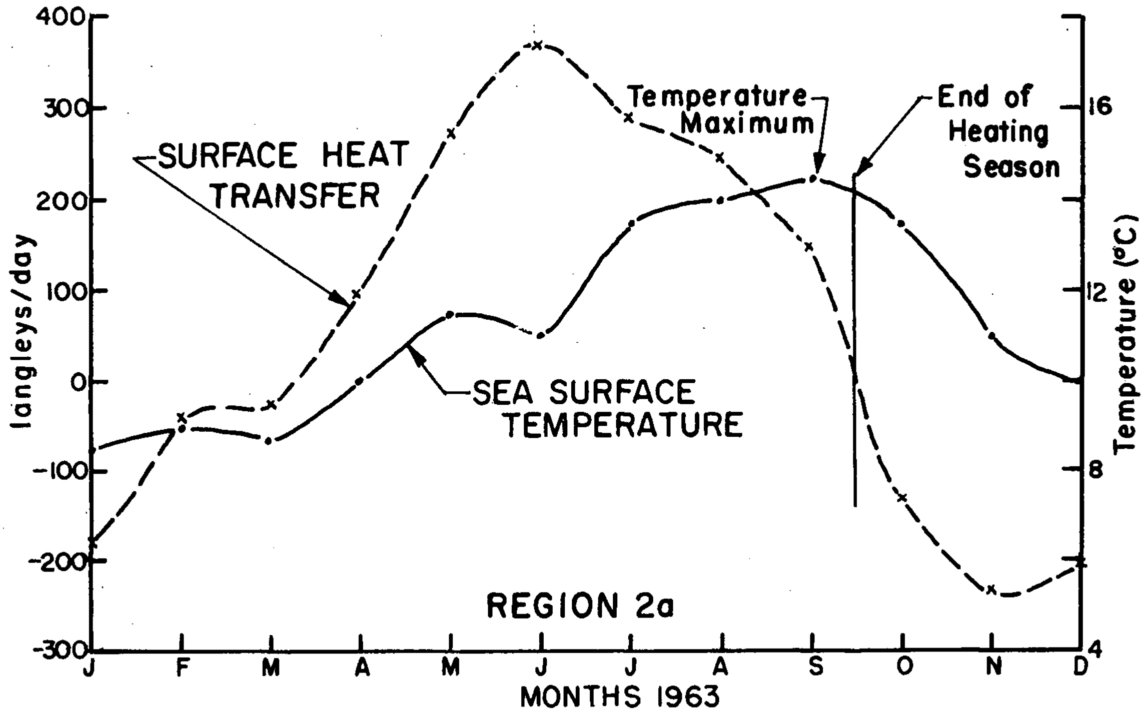


Fig. 33: Comparison of Heat Storage in Dixon Entrance, 1954 and 1955; with the Average Surface Heat Transfer, 1961 and 1963.



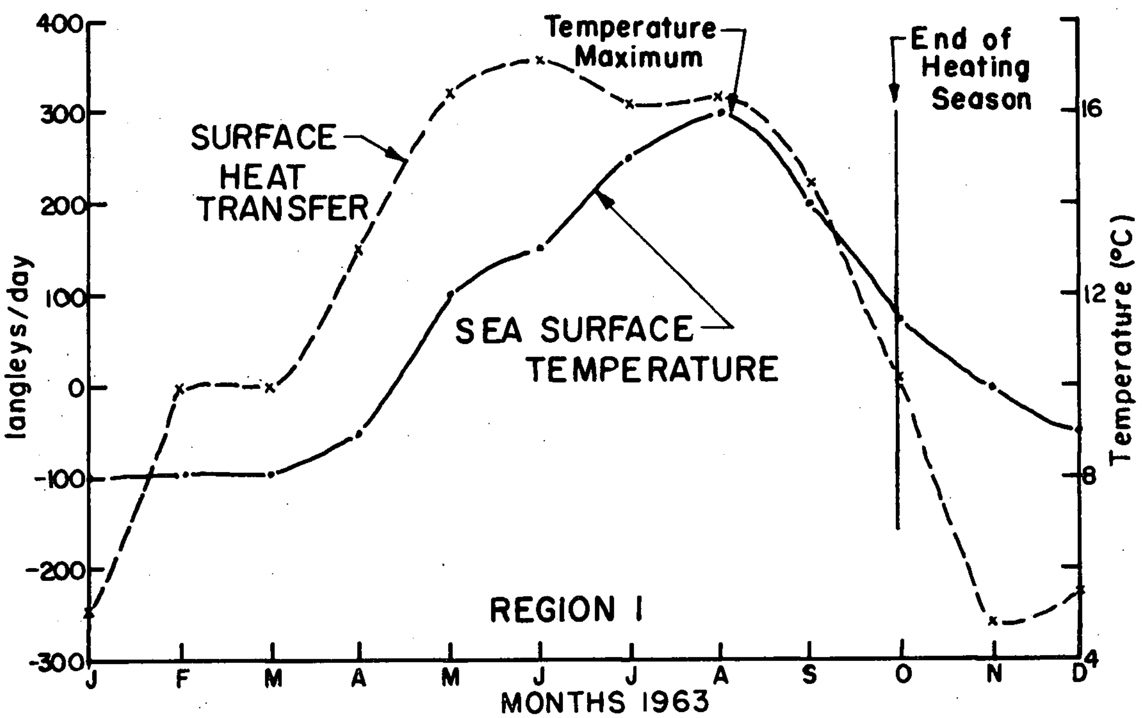


Fig. 34: Phase of Temperature Maximum with End of Heating Season.

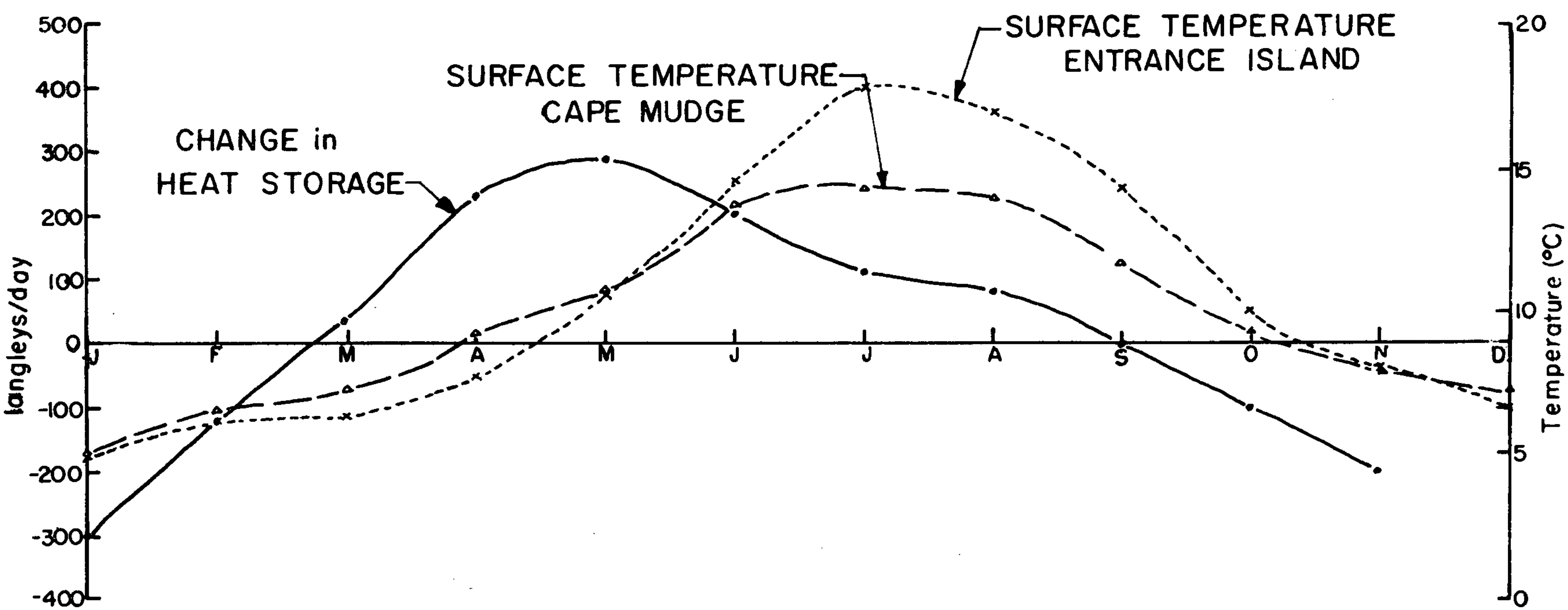


Fig. 35: Influence of Advection on Surface Temperature in the Strait of Georgia, 1950.

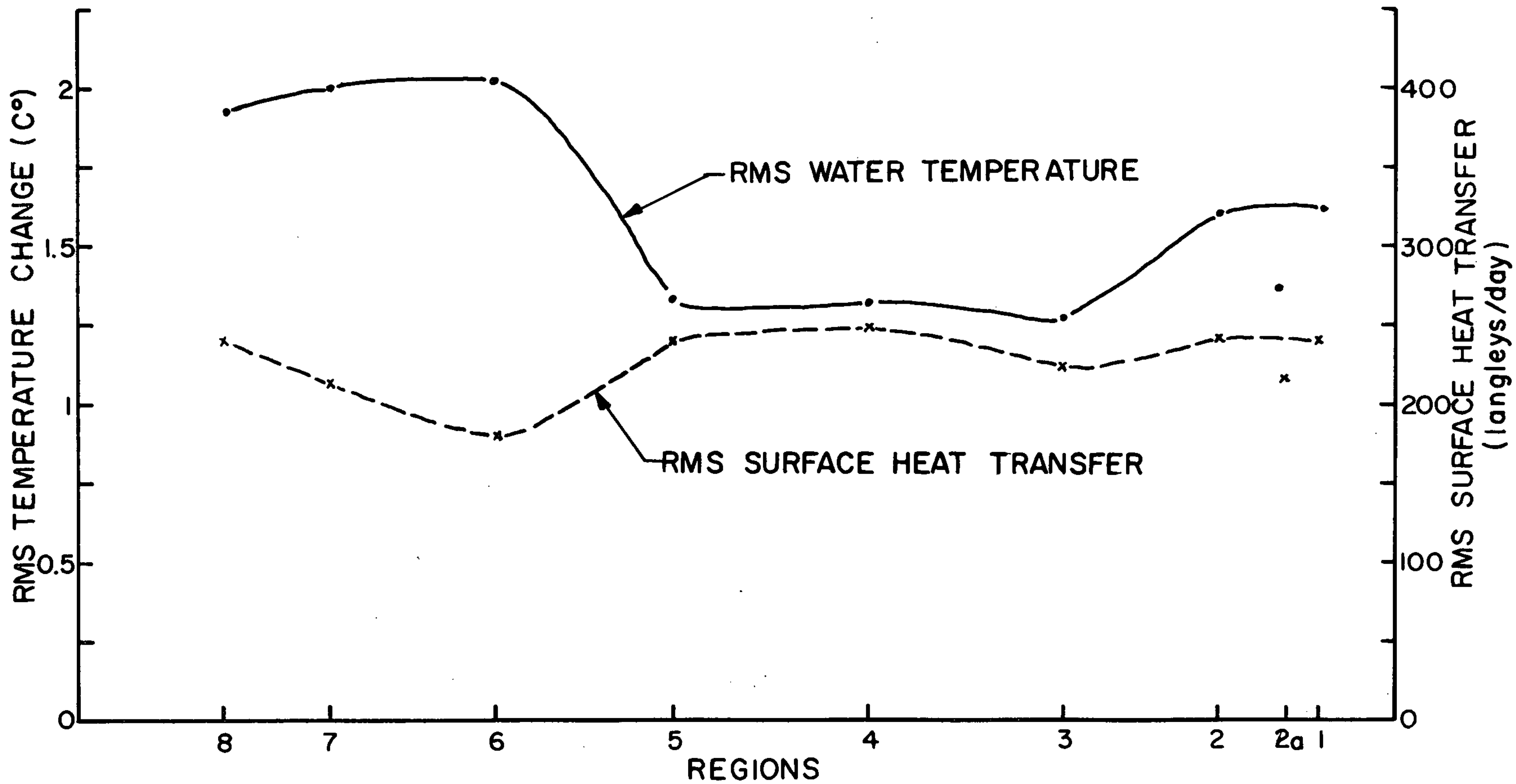


Fig. 36: Root-Mean-Square Values of Rate of Surface Heat Transfer and Surface Temperature Change, 1961 and 1963.

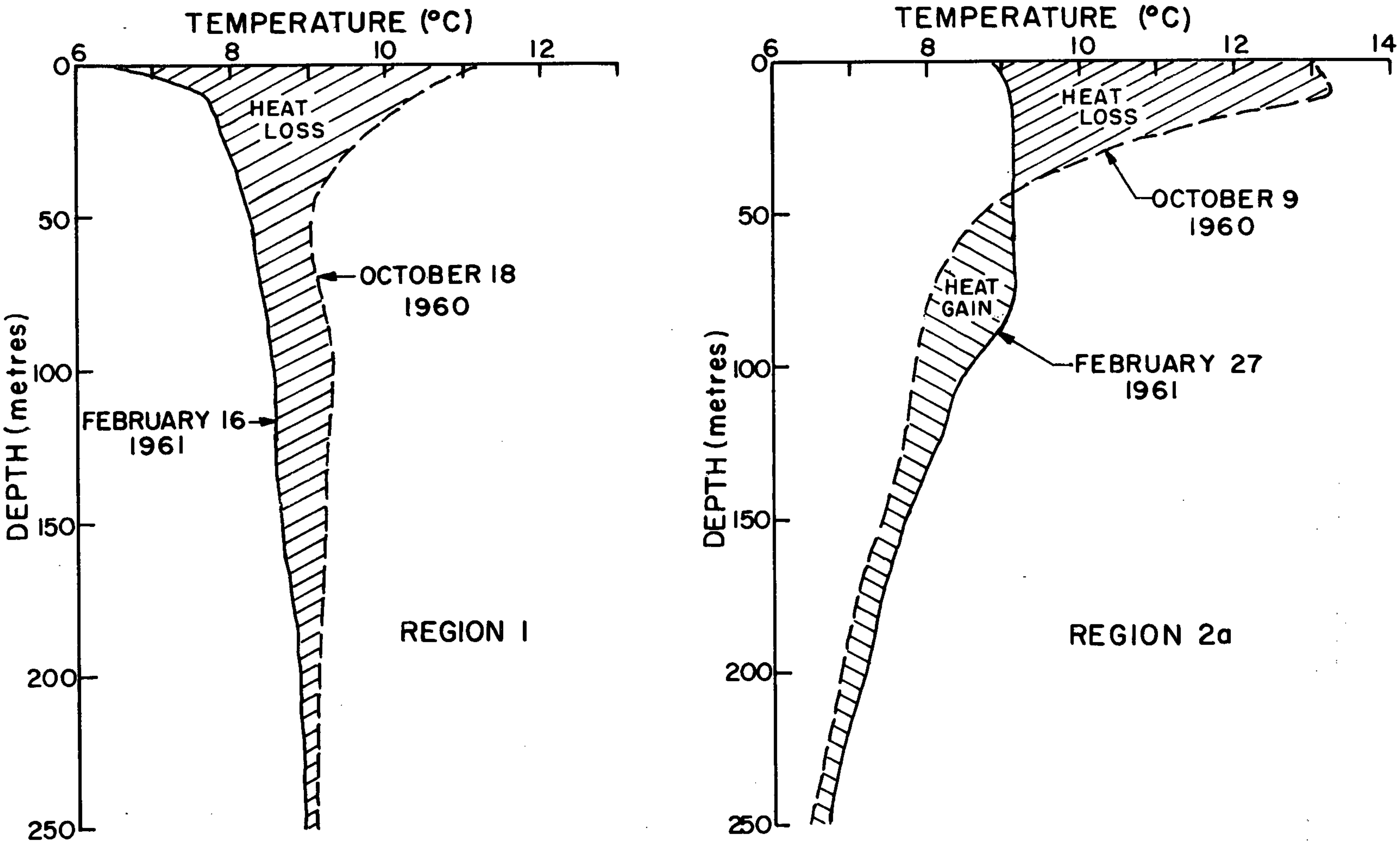


Fig. 37: Change in Heat Content of Water Column over one Cooling Season, Strait of Georgia and West Vancourver Island, 1960-1961.

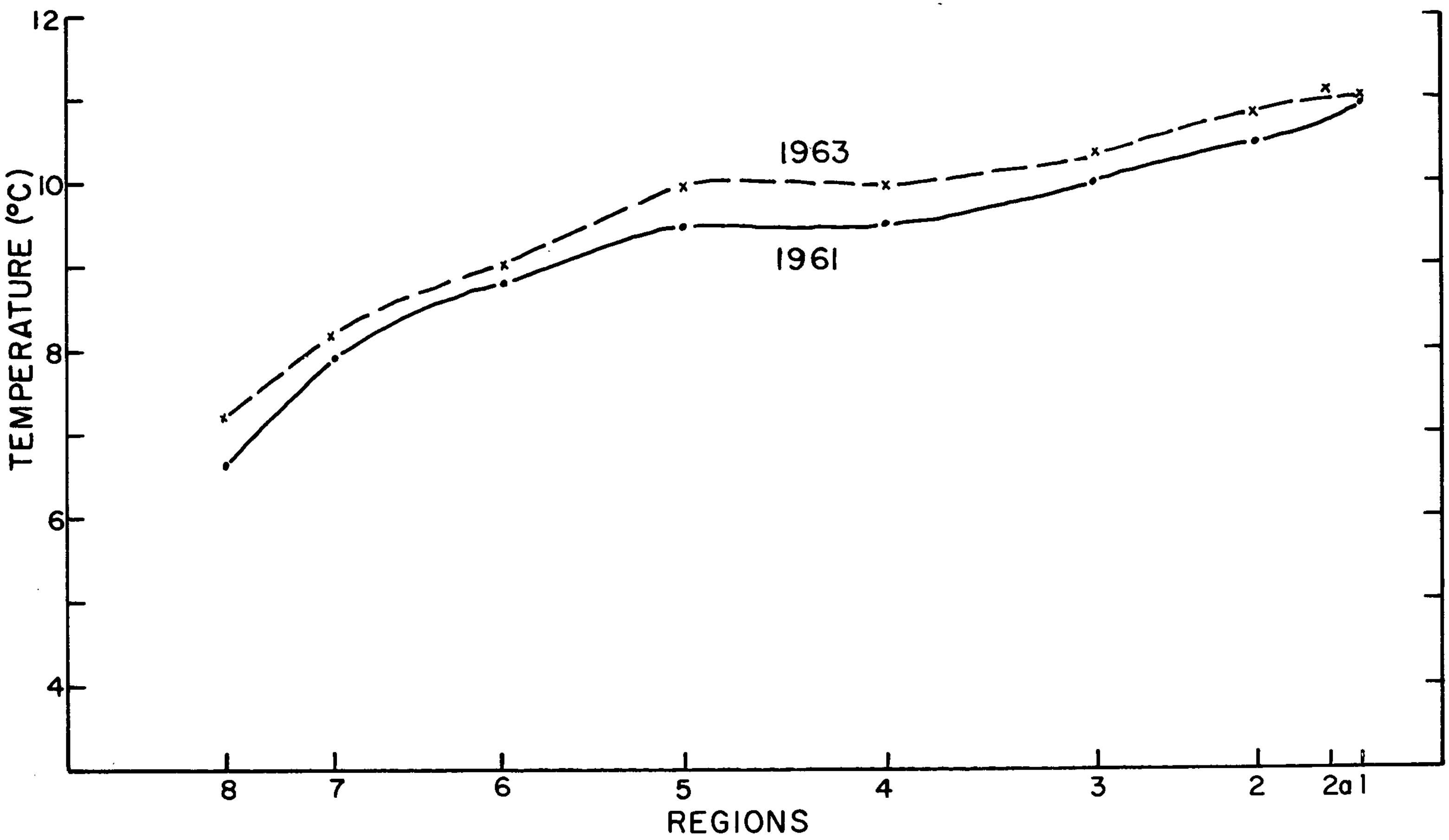


Fig. 38: Mean Surface Temperature of Coastal Waters, 1961 and 1963.