SEA LEVEL FLUCTUATIONS

IN THE

GULF OF ST. LAWRENCE

MEAN SEA LEVEL FLUCTUATIONS
IN THE
GULF OF ST. LAWRENCE

by

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ABSTRACT

Sea level variations in the frequency range between zero and one cycle per day (cpd) were analyzed for several tide gauges within, and near, the Gulf of St. Lawrence.

On the annual scale standard regression and correlation analysis showed that sea level and atmospheric pressure variations were 180° out of phase - leading to high sea levels in December and low sea levels in March and April. The direct action of atmospheric pressure can explain 30% of the sea level variance with the winds being able to account for an extra 15%.

Spectrum analysis of daily sea level, atmospheric pressure and winds at Grindstone Island showed that sea level and atmospheric pressure exhibited a close relationship in the frequency band from 0 to 0.55 cpd. However, the response of sea level to pressure was less than that to be expected from the pure hydrostatic relationship. The response also exhibited a change from summer to winter seasons - being less in the winter. This effect is probably due to the shift towards higher frequencies of all fluctuating quantities in the winter and, an increased influence of the winds although their direct effect was difficult to assess because of high sampling variability in the records.

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CHAPTER I

INTRODUCTION

1.1 Scope of the Present Study

Although considerable effort has been expended in simulating, through various mathematical models, the surface circulation of the Gulf of St. Lawrence (Jarlan,1961; Blackford, 1965,1966) their solutions, based on many simplifying assumptions, have failed to give a realistic picture of the upper circulation.

Because of its relatively complicated physiographic and topographic shape, any improvements in the above noted steady state models must be based on experimental evidence which, especially during the winter season, is lacking in quantity. In order to study the time dependence of the circulation or, to predict the distribution of conservative properties, a knowledge of the following must be sought (Welander,1961): (1) distribution of pressure and shear forces at the free surface and other boundaries, (2) net heat flux through the surface layer (vertical and lateral), (3) net mass flux through the surface layer(including evaporation, precipitation, freezing, melting and estuary inflow)

Studies dealing with item (2) have been made by Matheson (1967) and Coombs (1962). Some aspects of (3) were studied by Forrester (1964, 1967) and Sandstrom (1919) among others. Nontidal fluctuations in sea level, representing the result of the direct interaction of ocean and atmosphere, are thus amenable to analysing items (1) and (3). The success of this approach for other areas has been aptly shown by Palmén (1932, 1936), Lisitzin (1958, 1962), Thiel (1953).

Using tide gauge recordings from several locations within and outside the Gulf of St. Lawrence, the purpose of this thesis is to examine, on different time scales, the effect of certain meteorological and oceanographical perturbations on mean sea level. To avoid confusion, mean sea level will be defined as the time averaged height of the sea surface referred to some local datum which is fixed.

Once the meteorologically disturbing effects have been eliminated from the sea level records, the variations of the monthly mean sea level might elucidate on the nature of the annual water balance in the Gulf. Due to its partially enclosed nature, the above variations might then be used as possible indicators of currents through its openings to the Atlantic Ocean.

Daily variations in mean sea level together with variations in atmospheric pressure and winds have been investigated for one location only. Grindstone Island, Quebec was chosen as the most suitable location because of the following considerations:

(1) Its central location in the Gulf insured us that the winds as measured at this location, were fairly representative of the wind field over a wide portion of the Gulf (Ingram, personal communication).

(2) Homogeneous, hourly values of sea level, recordings of atmospheric variables at 3 hourly intervals, existed in sufficient quantities to make a statistical analysis.

(3) Its proximity to the amphidromic point located to the north-west insured us that the high frequency tidal noise would have a smaller effect on the records as compared to any other location. In addition, the shallow depths around the island would magnify the atmospherically induced sea level variations.

1.2 Area of Investigation

The Gulf of St. Lawrence (Fig. 1) is a shallow, marginal sea situated on Canada's east coast between latitudes 46^{o}N and 52^{o}N. Its surface area of approximately 1.77×10^{11} $meters^{2}$ is distributed in a roughly triangular shape of which 52% by area is deeper than 100 meters but less than 20% is deeper than 300 meters (Banks,1966; Lauzier,1957).

The most important topographic features are formed by the Laurentian channel and its branches, Esquiman and Mingan channels. Extending from the edge of the continental shelf to the mouth of the Saguanay river, the Laurentian channel permits substantial amounts of Slope and Labrador current waters to penetrate through the 500 meter deep Cabot Strait some 900 Km into the interior. The resulting interaction between freshwater outflow and deeper saline waters permit the formation of the required solenoidal field of mass which drives part of the Gaspe Current (Hachey,1961; Sandstrom,1919).

In the southwestern portion of the Gulf there exists a circular, shallow water body known as the Magdalen Shallows. Including Northumberland Strait, the average depths over this platform are less than 75 meters although it is cut by many troughs whose depths exceed this value (Loring and Nota,1966). To the northeast, lies the Strait of Belle Isle. Its shallow sill depth of 60 meters restricts the exchange with Atlantic waters to the surface layers only . Hence, it does not play an important part in the mass balance of the Gulf even though

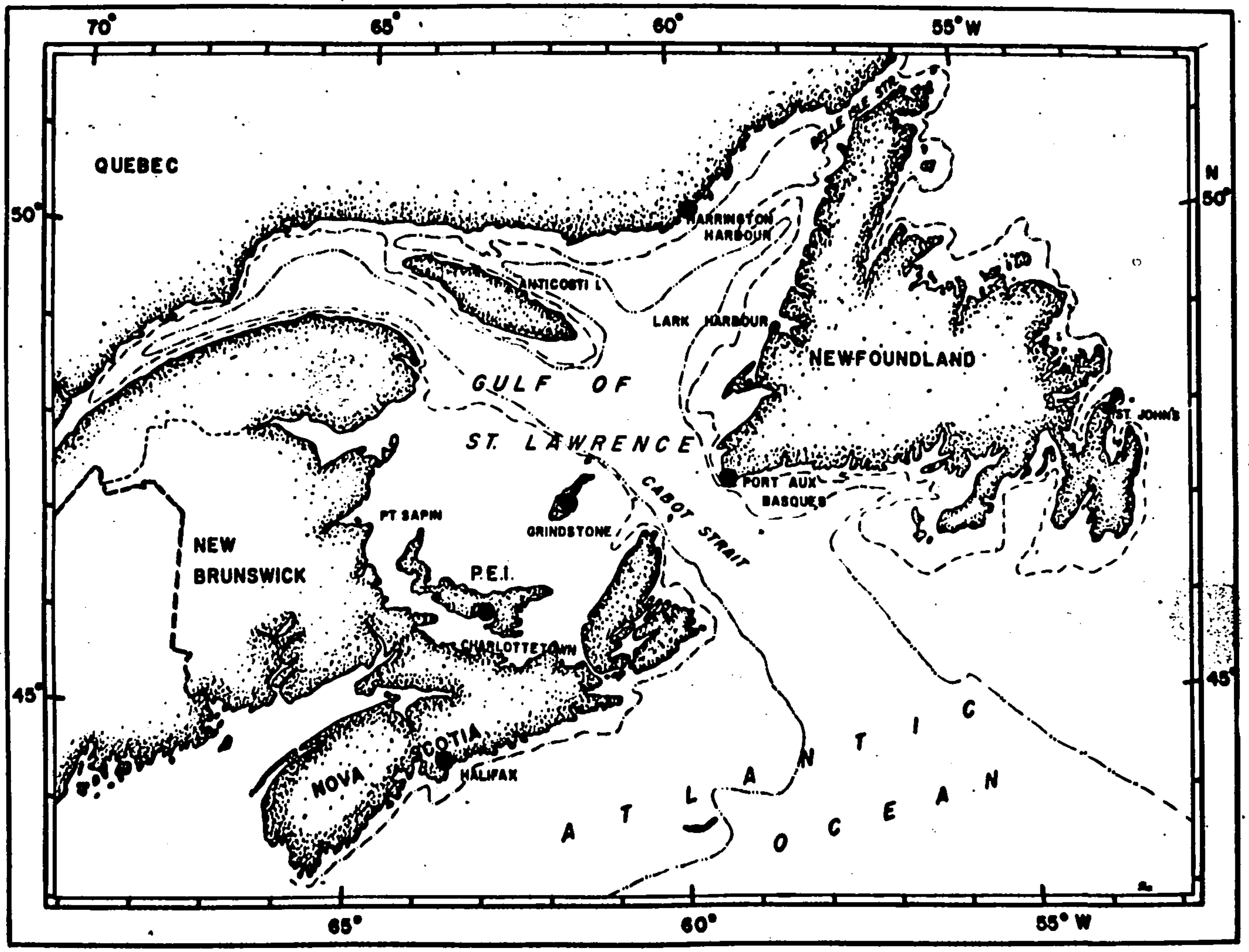

FIG. 1 GULF OF ST. LAWRENCE

it does figure in its oceanography.

1.3 General Climatology

The seasonal variation in the prevailing winds and the subsequent weather systems over the Gulf are, to a large extent, dominated by the two semipermanent pressure cells lying over the North Atlantic namely, the Icelandic low and the Bermuda-Azores high.

For the average winter conditions, here taken to be in January, the core of the Icelandic low is spread along latitude 62° N with its western edge usually over the southern tip of Greenland. The associated cyclonic circulation results in prevailing winds from the northwest generally also bringing cold continental (cA) or maritime (mA) air into the region (Hare,1951). Considerable variations in the wind velocity field are brought about by travelling cyclones which have their origin along the American arctic or Atlantic Polar front (Pettersen,1956). Frequently, these migratory lows become stationary east of Newfoundland resulting in a northeasterly or southwesterly flow over the Gulf.

With the advent of summer, the Bermuda-Azores high extends more over the continents causing winds to flow predominantly from the southwest at speeds from 15 Km/hr to 25 Km/hr (Thomas,1953). At this time also, the Polar front is displaced northward causing extra tropical cyclones to take a more northerly route than in winter. Wind variability is high, yet the seasonal change in the wind direction accounts for only 90° .

1.4 Tides

Comprehensive accounts of tidal phenomena in the Gulf have been given by Dawson (1898, 1920) and Farquharson (1957, 1962, 1966).

Figure 2 shows the semidiurnal wave propagated through Cabot Strait. Through Coriolis force, this wave is transformed into an amphidromic system having its centre near $47^{\circ}20'$ N and $62^{\circ}10'$ W. Because of bottom topography, coastal boundaries and meteorological effects, the type of tide at any one locality varies. Table 1 summarizes the fundamental tidal information for the various gauges used in this study.

1.5 Data Used

The tide and meteorological stations utilized are shown in Fig. 1. Although other data were available, these stations were selected because all contained fairly homogeneous and continuous records. As pointed out by Stommel (1963), in order to obtain any significant and conclusive results regarding the annual trends, the sampling duration used must be greater than two years. Most of the stations rejected either had extensive records gaps or, atmospheric information was not recorded.

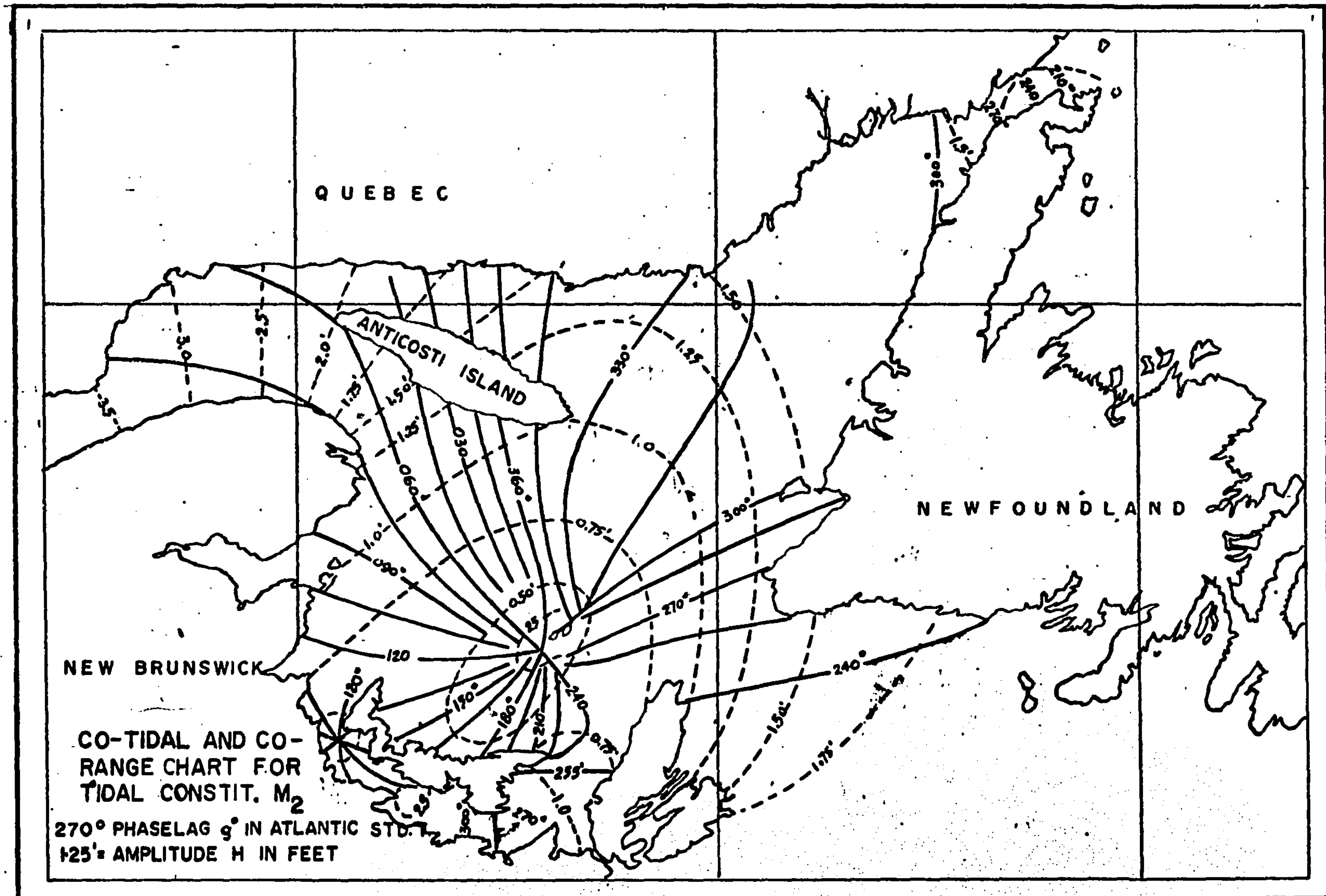

FIG. 2 THE M_2 TIDE (after Farquharson, 1962)

TABLE 1

Harmonic Constants for Tide Gauges

Place	Harrington Hbr.		Charlotte-town		Halifax		Port aux Basques		St. John's		Grindstone Island	
ϕ (lat. N)	50° 30'		46° 13'		45° 37'		47° 35'		47° 34'		47° 23'	
λ (long.W)	59° 29'		63° 08'		61° 22'		59° 09'		52° 42'		61° 52'	
Component	A	θ	A	θ	A	θ	A	θ	A	θ	A	θ
M_2	53.9	313.8	72.1	305.7	63.1	234.2	44.4	270.9	35.1	213.9	21.0	258.2
S_2	17.7	337.7	17.9	8.4	14.2	261.7	13.3	303.7	14.6	254.3	8.1	297.1
K_1	15.1	202.9	25.4	265.4	10.0	60.5	7.9	198.7	7.7	106.6	13.3	236.4
O_1	14.5	183.7	23.2	237.8	5.0	37.0	9.3	181.3	6.8	80.6	13.0	210.1
MF	2.8	190.7	2.3	208.9	0.7	219.6	0.6	283.9	3.1	209.3	1.7	211.2
Sa	5.3	291.7	2.9	3.1	4.4	15.1	3.5	349.9	7.8	351.1	2.0	255.9
Ssa	4.8	259.8	6.8	273.8	4.9	266.2	2.1	219.3	7.4	262.9	6.7	281.2
MSf	2.5	256.4	2.7	254.2	2.0	286.3	2.7	240.3	2.3	272.4	-	-

where A = amplitude in cm. θ = phase in degrees

Hourly heights, daily and monthly mean sea levels were supplied by the Canadian Hydrographic Service. The water level heights are automatically digitized from continuous traces made by Ott, float operated gauges recording to an accuracy of ± 0.61 cm (Mackenzie, personal communication). Time is considered accurate to ± 5 minutes.

The daily and monthly mean sea levels represent consecutive nonoverlapping arithmetic averages over their respective periods. Short period tidal waves are practically eliminated in the monthly means (Patullo et al., 1955). They appear with less than 1% of their amplitude. The annual and semiannual waves are very little affected. Daily mean sea level as derived from consecutive 24 or 25 hourly arithmetic averages is not sufficiently free from short period tidal effects. Several numerical filters which were more efficient in suppressing tidal variations were applied to the hourly sea level heights. The filter characteristics are discussed in Chapter 3.

Monthly mean barometric pressure, monthly totals of precipitation, mean monthly temperatures were abstracted for the years 1960 to 1966 inclusive from the Monthly Record. The pressure represents the mean of the daily synoptic hours (0000,0600, 1200,1800) adjusted to sea level. Winds, obtained from the Climatology Branch, were tabulated as monthly wind milage and their frequency in hours partioned over the eight principal compass directions.

Recorded values of wind speed and direction, atmospheric pressure at 3 hourly intervals for Grindstone Island were abstracted from microfilm records.

Temperature and salinity data to a depth of 175 m were supplied by Dr. Templeman for Station #27. As no systematic sampling program is carried out at this station, the data introduce large errors in showing variability. In addition, the shallow sampling depth of 175 m precludes the determination of dynamic heights for steric calculations. These data could,however, give indications of the variability of that part of the Labrador Current flowing through Avalon Passage; thus, its relation to sea level as recorded at St. John's. Steric computations were attempted by the author for the Halifax and Cabot Strait sections, using the data presented by Gibson and Mann, 1966; CODC reference numbers 267 to 274, without much success. That the information content of these data may be very doubtful has been shown by Mann and Needler (1967).

CHAPTER II

SEASONAL VARIATIONS IN SEA LEVEL

2.1 Characteristics of the Sea Level Records

Monthly mean sea levels at several stations along the shores of the Gulf of St.Lawrence are shown in Fig. 3 . For comparative purposes, mean monthly sea levels at Halifax,N.S. and St.John's, Nfld. have also been included.

Although the individual monthly mean sea levels differ considerably, not only from month to month but also from station to station, it is interesting to note that their mean annual variation do show some similarity. On the basis of Fig. 3, it may be noticed that the seasonal variation at Harrington and Charlottetown are quite similar; low sea levels occur in late winter or early spring, rise rapidly to a secondary maximum sometime in the summer, and peak in late November or December.

In contrast to the variations in the Gulf are those from Halifax and St.John's. Again, the mean annual variation between the two stations is similar; however, they differ in some important ways from those in the Gulf. Most notable among

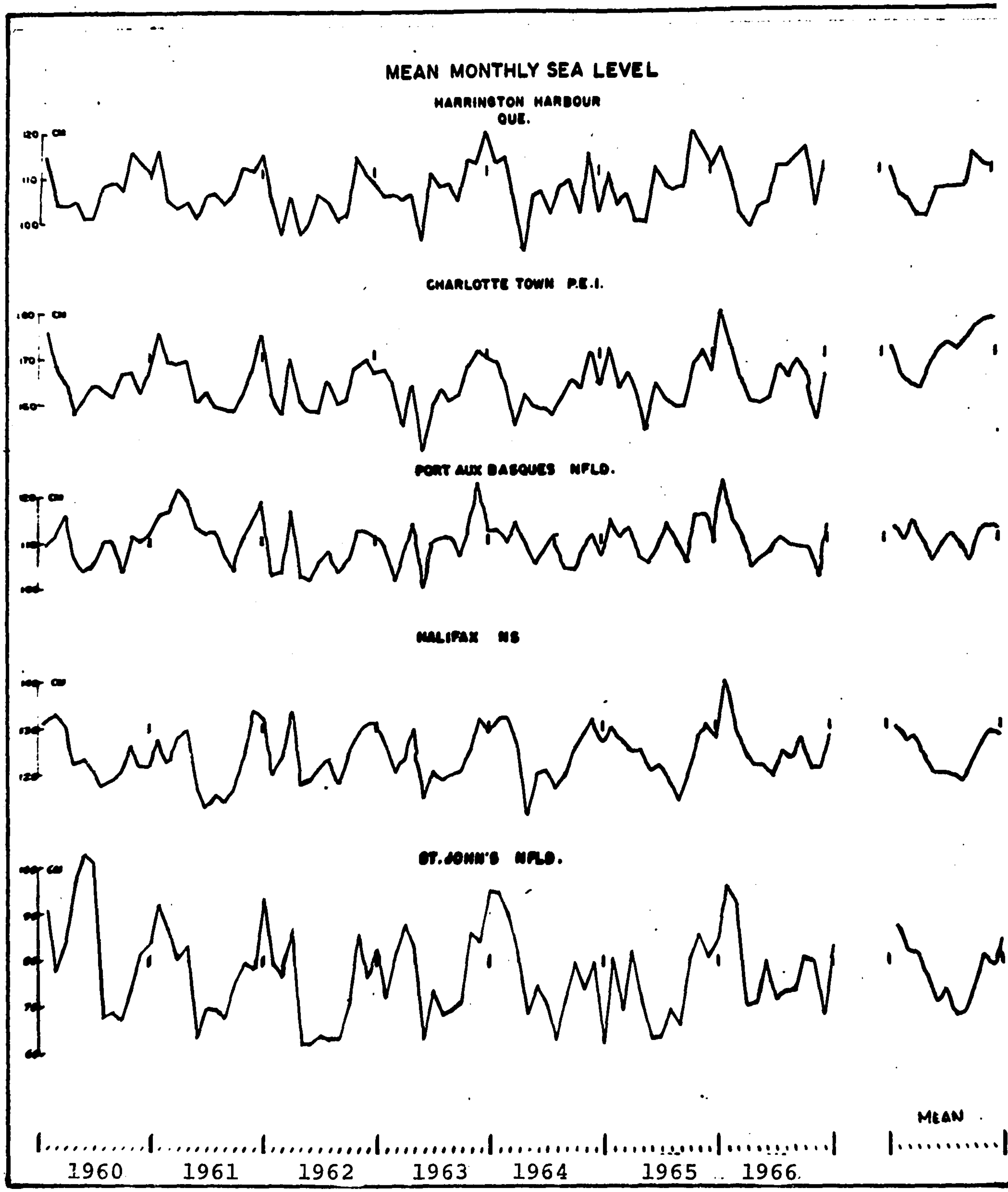

Figure 3 - Monthly Mean Sea Level

these is the fact that no appreciable secondary maximum occurs in the summer, and the phase of the sea level minimum is shifted by several months. The mean annual variation of sea level at Port aux Basques does not seem to follow any regular pattern. Between the two minimums located in May and September, is a maximum in July which attains a level nearly equal to the primary maximums in September or October. Whether this summer peak is real is difficult to state in view of the short period used in the analysis- the peaks for individual years show differences but these differences are usually of the same size and phase suggesting a basic phenomenon. It should be noted however that Port aux Basques is located such that it would experience effects which are dominant in changing sea levels both outside and inside the Gulf. Most important is the fact that the prime exchange of waters takes place through Cabot Strait and hence, it would seem likely that these variations should exhibit themselves in the sea level records.

The phase difference for the summer minimum between the Gulf stations and those on the Atlantic coast could be viewed as a difference in sea level, existing between these two areas. Such a difference in sea level must involve a mass transport. In what direction this is to occur is difficult to state as the levels given in Fig. 3 are with respect to local chart datum. Nevertheless, it has been suggested (Farquharson,1962) that this phase inequality might affect, or be a cause of, the seasonal variation in the direction of the dominant flow through the Strait of Belle Isle and, in this respect, also through Cabot Strait (Bailey, 1958; Farquharson and Bailey, 1966). The range of the monthly mean sea level variations are all about the same, approximately 16 cm. At St. John's the variations are close to 28 cm.

Tables 2 and 3 show the mean monthly values of sea level, atmospheric pressure, wind speed together with their RMS amplitudes of the annual and non-annual oscillations. Included are the standard deviations for each variable in order to indicate the seasonal variability of these elements. The RMS amplitudes of the annual A and non-annual ΔA oscillations are defined by:

$$A = \left(1/12 \sum_{i=1}^{12} (\bar{X} - \bar{X}_i)^2 \right)^{\frac{1}{2}} \qquad (1)$$

$$\Delta A = \left(1/N' \sum_{i=1}^{N'} (\bar{X}_i - X_i)^2 \right)^{\frac{1}{2}} \qquad (2)$$

where $\bar{X}$ = long term mean of any element

$\bar{X}_i$ = monthly values for the normal mean year

X_i = mean monthly value of any element

N' = total number of observations

	1	2	3	4	5	6	7	8	9	10	11	12	A	ΔA
HARRINGTON, QUEBEC														
Z,cm	110.3	104.7	103.1	99.7	99.9	105.7	105.9	105.9	106.1	113.8	110.7	110.6	4.16	3.71
σ,cm	4.5	4.7	1.4	4.5	2.8	3.3	3.0	3.3	3.4	2.6	4.0	5.1		
P,mb	7.2	8.7	8.9	10.3	13.0	10.3	10.7	11.1	13.1	9.6	13.0	8.4	1.86	3.04
σ,mb	2.3	6.2	3.2	2.9	1.9	1.5	1.9	1.8	2.3	2.0	4.0	3.3		
V,Km/h	29.7	18.3	24.5	21.1	15.1	15.1	16.4	12.7	17.9	15.3	21.7	22.3	4.64	9.63
σ,Km/h	12.0	8.0	12.2	9.0	8.0	5.6	9.4	6.8	8.8	9.4	10.5	13.0		
CHARLOTTETOWN, PRINCE EDWARD ISLAND														
Z,cm	171.0	165.0	163.1	161.8	157.1	160.6	161.8	160.3	162.4	165.5	166.4	167.1	3.56	4.02
σ,cm	5.6	4.6	5.8	3.5	4.0	2.5	3.2	2.1	3.6	2.2	5.3	3.7		
P,mb	10.2	11.1	10.5	12.0	14.5	12.3	12.4	13.6	16.3	12.8	14.9	12.6	1.74	2.63
σ,mb	2.5	5.2	2.1	3.7	1.4	1.3	1.4	1.6	2.2	1.4	2.9	2.8		
V,Km/h	11.2	7.6	10.3	8.9	5.2	8.4	6.9	7.6	6.6	7.5	8.7	8.6	1.56	3.51
σ,Km/h	2.6	3.6	4.3	2.9	2.2	5.6	3.2	2.3	2.5	3.7	3.0	4.5		

Z = monthly sea level
V = wind speed
σ = standard deviation

P = (sea level atmospheric Pressure -1000)mb
ΔA = RMS amplitude of non annual variation
A = RMS amplitude of annual variation

TABLE 2 - Mean Annual Variation of Variables, 1960-1966

	1	2	3	4	5	6	7	8	9	10	11	12	A	ΔA
PORT AUX BASQUES, NEWFOUNDLAND														
Z,cm	112.3	109.4	113.9	108.6	104.8	107.9	110.7	107.4	104.9	111.0	112.3	111.8	2.83	4.03
σ,cm	5.8	5.1	4.3	5.5	3.9	1.9	1.6	2.7	1.7	2.6	6.1	3.4		
P,mb	7.4	8.9	8.2	10.2	14.2	12.1	12.7	13.3	15.4	11.9	13.7	10.6	2.42	2.95
σ,mb	2.7	5.3	3.4	3.6	1.6	1.6	1.2	1.6	2.0	1.4	3.6	4.1		
V,Km/h	8.9	6.3	6.8	4.5	4.3	4.9	6.8	3.0	3.1	3.4	5.7	1.7	2.82	5.15
σ,Km/h	3.4	2.6	3.2	2.0	2.8	1.7	0.9	6.0	0.9	0.9	2.1	2.9		
HALIFAX, NOVA SCOTIA														
Z,cm	129.4	126.7	127.2	122.4	119.4	119.4	119.1	117.9	121.9	125.7	128.6	127.7	4.02	3.89
σ,cm	5.7	4.5	3.4	6.1	2.5	2.8	3.1	2.8	3.2	2.9	4.6	3.2		
P,mb	10.8	11.4	10.6	11.6	15.1	13.4	13.4	14.7	17.0	13.6	15.2	13.1	1.89	2.56
σ,mb	2.5	5.0	2.1	4.1	1.4	1.5	1.4	1.5	2.2	1.0	2.6	2.5		
V,Km/h	7.8	7.1	8.9	6.0	5.6	7.1	7.3	7.8	4.6	5.7	5.4	6.2	1.20	2.45

Z = monthly sea level
P = sea level atmospheric pressure, 1000 mb
V = wind speed

σ = Standard deviation
A = RMS amplitude of annual oscillation
ΔA = RMS amplitude of non annual oscillation

TABLE 3 - Mean Annual Variation of Variables, 1960-1966

These departures become more significant as the data series become longer. Except for a few differences, the monthly deviations σ follow the annual pattern of sea level as depicted in Fig.3.

The ratio of the RMS non-annual oscillation to RMS annual oscillation for sea level ranges from 0.87 at Harrington to 1.4 at St.John's. Although St.John's and Halifax are both exposed to meteorological elements of about the same magnitude, the RMS amplitude of non-annual sea level variations at St.John's, exceeds the one at Halifax by a factor greater than two.

2.2 Meteorological Influences on Sea Level

The most important forces which cause "short period" fluctuations (for a definition of this term see Donn et al., 1964) in mean sea level may be summarized into internal and external forces (Rossiter, 1962; Montgomery, 1938):

(a) external:

- (i) atmospheric pressure variation
- (ii) effect of the wind stress
- (iii) formation of an ice cover

(b) internal:

- (i) steric oscillation as caused by changes in density due to variation in temperature and salinity. These are termed isostatic as it is implied that the total pressure on the sea bottom does not change with time, i.e., it is a volumetric change.

(ii) dynamical effects due to fluctuations in gradient currents.

Although no investigations have been made as to the relative importance of each of these factors in causing the seasonal fluctuations in the Gulf, the similarity of the mean variations shown in Fig. 3 would tend to indicate a common origin.

The comprehensive studies of Patullo et al. (1955), Lisitzin (1961, 1963) and Patullo (1962), among others, have shown that the seasonal variation in sea level is primarily determined by the variations in the combined effect of water density and atmospheric pressure. These works have also shown that north of latitude 40°N the pressure factor is the dominant - leading to low sea levels in summer and high levels in the late autumn.

Typical time variations (month to month) of sea level and atmospheric variables are depicted in Figs. 4 and 5. Their mean annual variations have also been included. To determine amplitudes and phases the variables were subjected to a harmonic analysis. The series of N equally spaced data points can be described by the following (Hamming, 1962):

$$X(t) = [X_i]_N + \sum_{N=1}^{N/2} C_n \sin(2\pi nj/P + \phi_n) \qquad (3)$$

where $X(t)$ = time dependent value at any time t.

where j = time factor varying from 0 to N-1

P = period of the fundamental cycle

C_n = amplitude of the n^{th} harmonic

ϕ_n = phase angle

2.2.1 Atmospheric Pressure

The results of the harmonic analysis of sea level and pressure are presented in Tables 4 and 5. Here N=12=P and x varies from 0 (January) to 11 (December), corresponding to mid-monthly values. Statistically, it is also interesting to evaluate the number of harmonics required to reduce the variance below a certain value. The variance of the 0^{th} harmonic is the variance of the data about the average of X_i. The n^{th} harmonic can account for $C_n^2/2$ except for the last which contributes C_n^2. The results on the other variables are presented in Tables 22 to 26 in Appendix B.

Table 4 - Harmonic Constants of Normal Sea Level

Location	Annual Mean (cm)	n=1		n=2		σ_1^2	%
	$\langle X_i \rangle_{12}$	C_n*	ϕ_n**	C_n*	ϕ_n**		
Harrington	106.37	5.21	158	1.45	134	3.07	83
Charlottetown	163.49	4.49	117	1.27	93	1.79	86
Port aux Basques	109.58	2.53	99	0.62	106	3.39	43
Halifax	123.78	5.49	127	0.43	244	1.22	93
St. John's	77.09	7.62	241	1.22	193	3.18	91

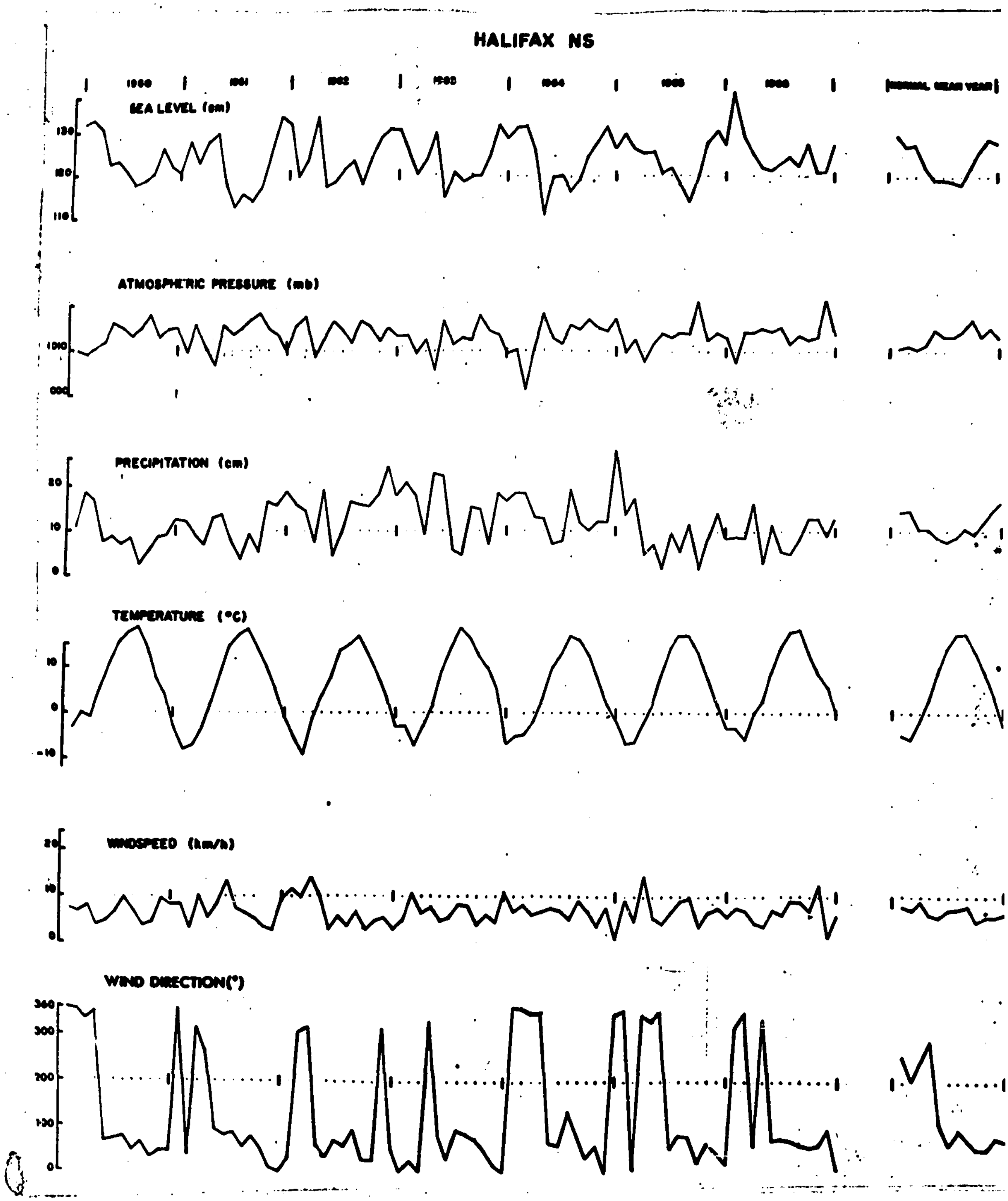

Figure 4 - Annual Variation for Halifax, Nova Scotia, 1960-1966.

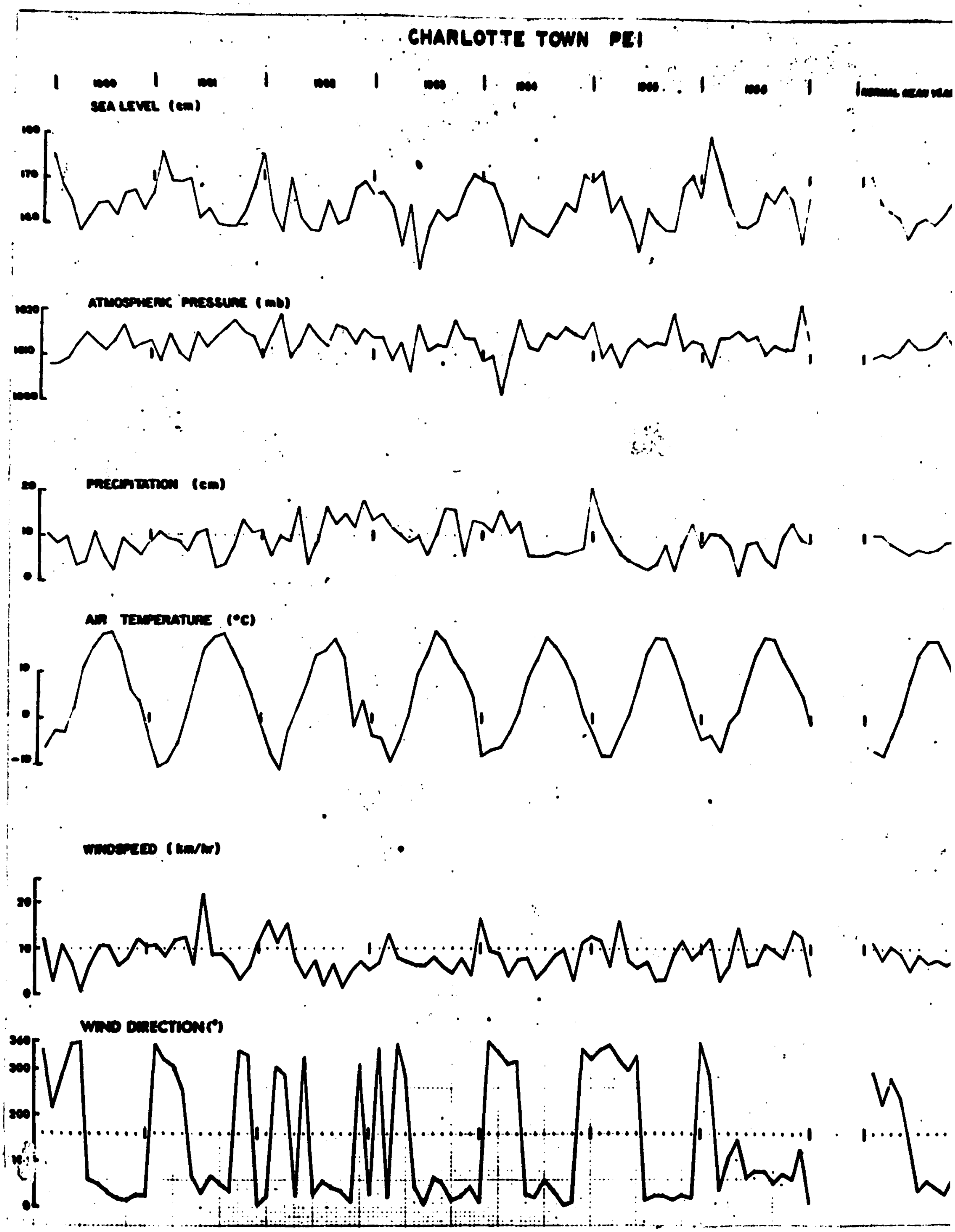

Figure 5 - Annual variation for Charlottetown, Prince Edward Island, 1960-1966.

* values in cm

** values rounded off to nearest degree

$\sigma_1^2 = \sigma^2 - 1/2\ \Sigma\ C_n^2$ is the amount of unexplained variance $(cm)^2$

% = percent of variance explained by 1^{st} and 2^{nd} harmonic

Table 5 - Harmonic Constants of Atmospheric Pressure at Sea Level

	Annual Mean (mb - 1000)	n=1		n=2	
Location	$X(t_i)$ 12	C_n*	$\emptyset_n$**	C_n*	$\emptyset_n$**
Harrington	10.35	1.62*	251**	1.22	250
Charlottetown	12.77	1.71	222	0.99	245
Port aux Basques	11.56	2.83	238	1.14	233
Halifax	13.33	2.12	224	0.76	233
St. John's	11.14	4.48	241	1.22	193

* values in mb

** values rounded off to nearest degree

Inspection of Table 4 and 5 immediately shows that the annual component of sea level and pressure is everywhere dominant with, however, the semi-annual wave also contributing substantially. The magnitude of this component does not differ greatly between stations except for St. John's, where both magnitude and phase relationships are different. The average annual Sa wave for sea level is 5.1 cm. At Port aux Basques, the amplitude of the Sa component is only 44% of the magnitude of the average. In fact, while the sum of the annual and semi-annual waves in the sea level records can usually explain more than 80% of the variance, these two components can only account for about 43% of the variance at Port aux Basques.

The harmonic analysis of the normal mean values of sea level and pressure indicated that the pressure variation ran parallel to, but with a small time lag, to changes in sea level. In addition, the amplitudes of the pressure variations were significantly smaller than the sea level variations.

The response of the sea surface to changes in atmospheric pressure is usually expressed by the hydrostatic equation. Using this, however, implies that there is no motion in the atmosphere and ocean and, no secular change in the total volume of the ocean basin. If the hydrostatic hypothesis is correct, and indications given by Groves (1954) show that for periodic changes in pressure with periods greater than one day it is, then the change in sea level due to a change in pressure is given by Proudman (1953).

$$\frac{\partial Z}{\partial t} = \frac{1}{\rho g} \left[\frac{\partial P_a}{\partial t} - \frac{1}{\sigma} \iint_{\sigma} P_a \, d\sigma \right] \qquad (4)$$

where Z = sea level in cm

P_a = sea level atmospheric pressure

ρ = density of the water

σ = area of the basin considered

Substituting the appropriate values for the constants, the relation of 1 cm change in water level yields for each mb change in pressure. To evaluate the constant for the particular area, two different methods were used. To find the effect of any given pressure change on sea level, the mean annual cycle plus any long period tides were first eliminated from the records by forming anomalies from the normal mean values. These deviations were then plotted against each other on a scatter diagram and linear regression lines, giving the most probable sea levél deviation for any pressure change, were fitted. Assuming these deviations are distributed normally about their mean, a t-test was applied and the significance of the correlation coefficients at the 5 % level determined.

The other method did not eliminate the seasonal cycles. Linear regression and correlation were also carried out on these two variables. Comparing the coefficients may then indicate the effects of the mean annual variation or trends. The results are summarized in Table 6 and 7 .

The highest correlation between a change in pressure and the corresponding change in sea level was found to be at Harrington and the lowest at Port aux Basques. The negative coefficients indicate the inverse relationship between pressure and sea level. Although all coefficients are significant at the 5% level, their magnitudes nevertheless are low. It is suspected that pressure and sea level are both related to a third (or more) variable(s), in most likelihood the wind.

Table 6 - Linear Regression and Correlation Coefficients for Atmospheric Pressure and Sea Level Anomalies

Station	r	SE	Regression Equation	Sign
Harrington	-0.69	2.73	$X_1 = 0.0002 - 0.84X_2$	s
Charlottetown	-0.50	3.51	$X_1 = 0.004 - 0.77X_2$	s
Port aux Basques	-0.38	4.05	$X_1 = 0.003 - 0.49X_2$	s
Halifax	-0.65	3.50	$X_1 = 0.006 - 0.98X_2$	s
St. John's	-0.46	7.25	$X_1 = 0.003 - 1.04X_2$	s

Correlation and regression analysis was also carried out for the other variables. In general, precipitation and temperature were not significantly correlated to sea level. From Table 6 it can be seen that the correlation coefficient is low at all stations. This is especially evident at those locations where currents predominate. The isostatic factor is significantly different from 1 at Charlottetown and Port aux Basques only.

r = correlation coefficient

SE = standard error

X_1 = sea level anomaly in cm

X_2 = pressure anomaly in mb

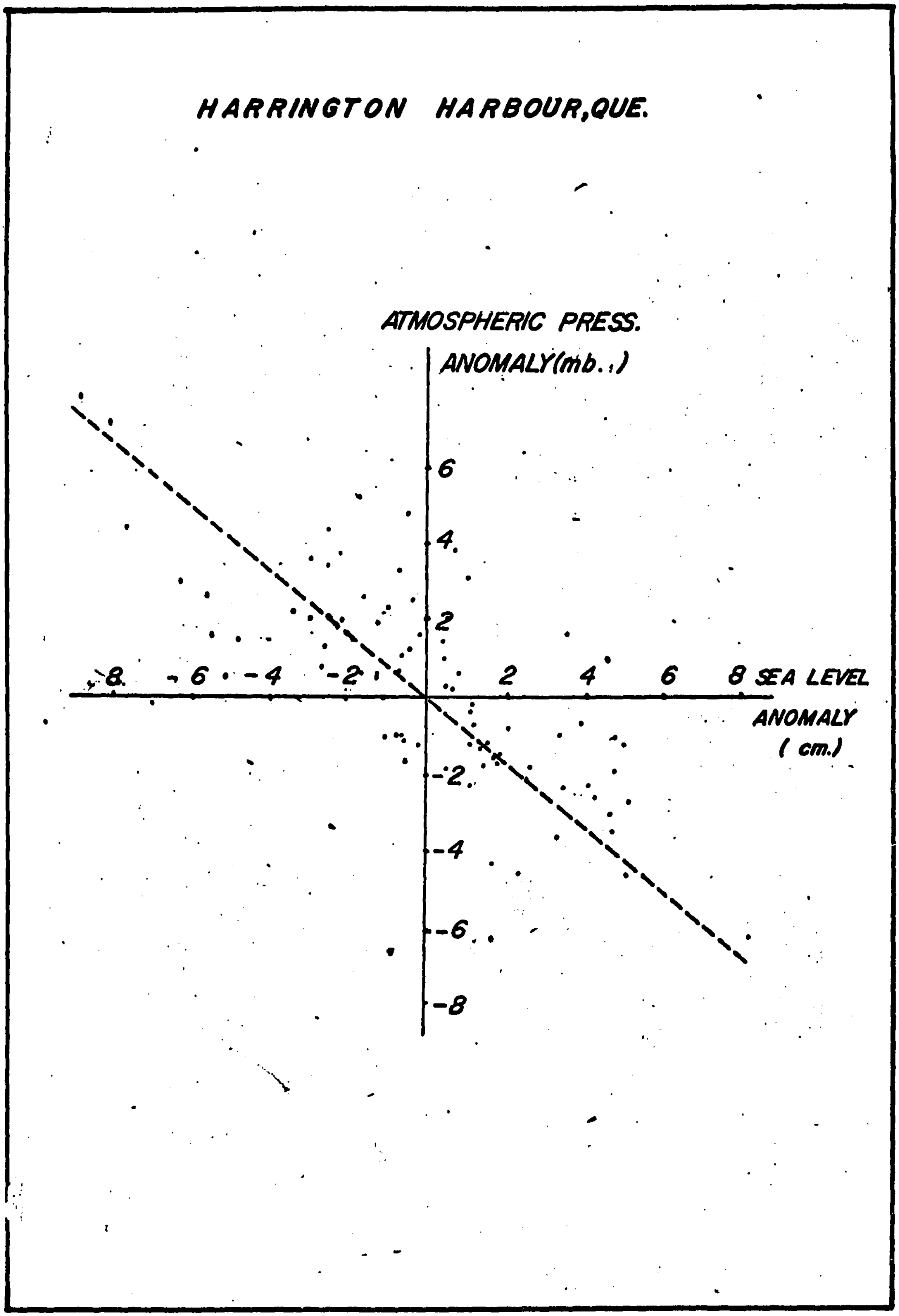

FIG. 6 Relation between Sea Level and Atmospheric Pressure at Harrington, Quebec

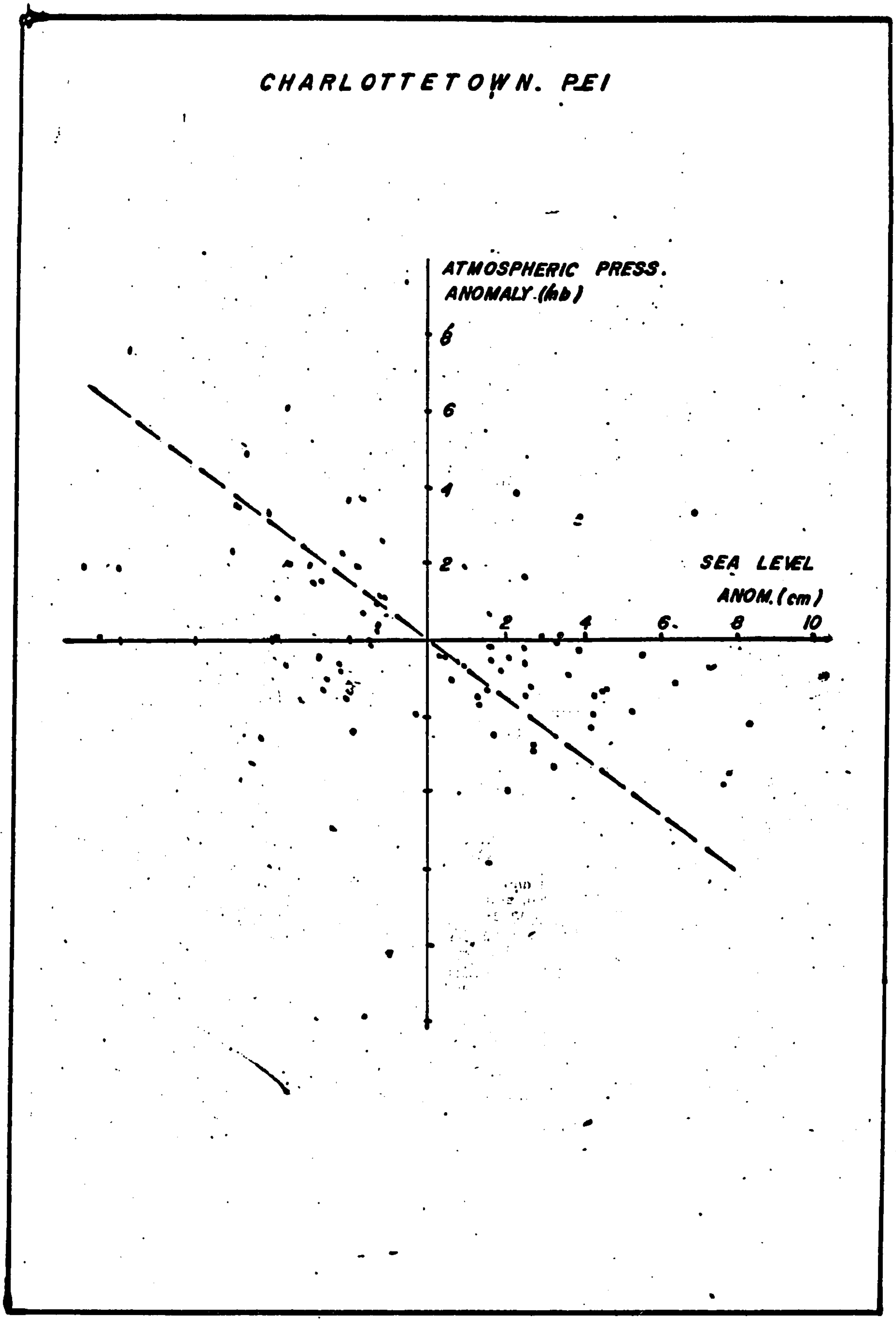

FIG. 7 Relation between Sea Level and Atmospheric Pressure at Charlottetown, P.E.I.

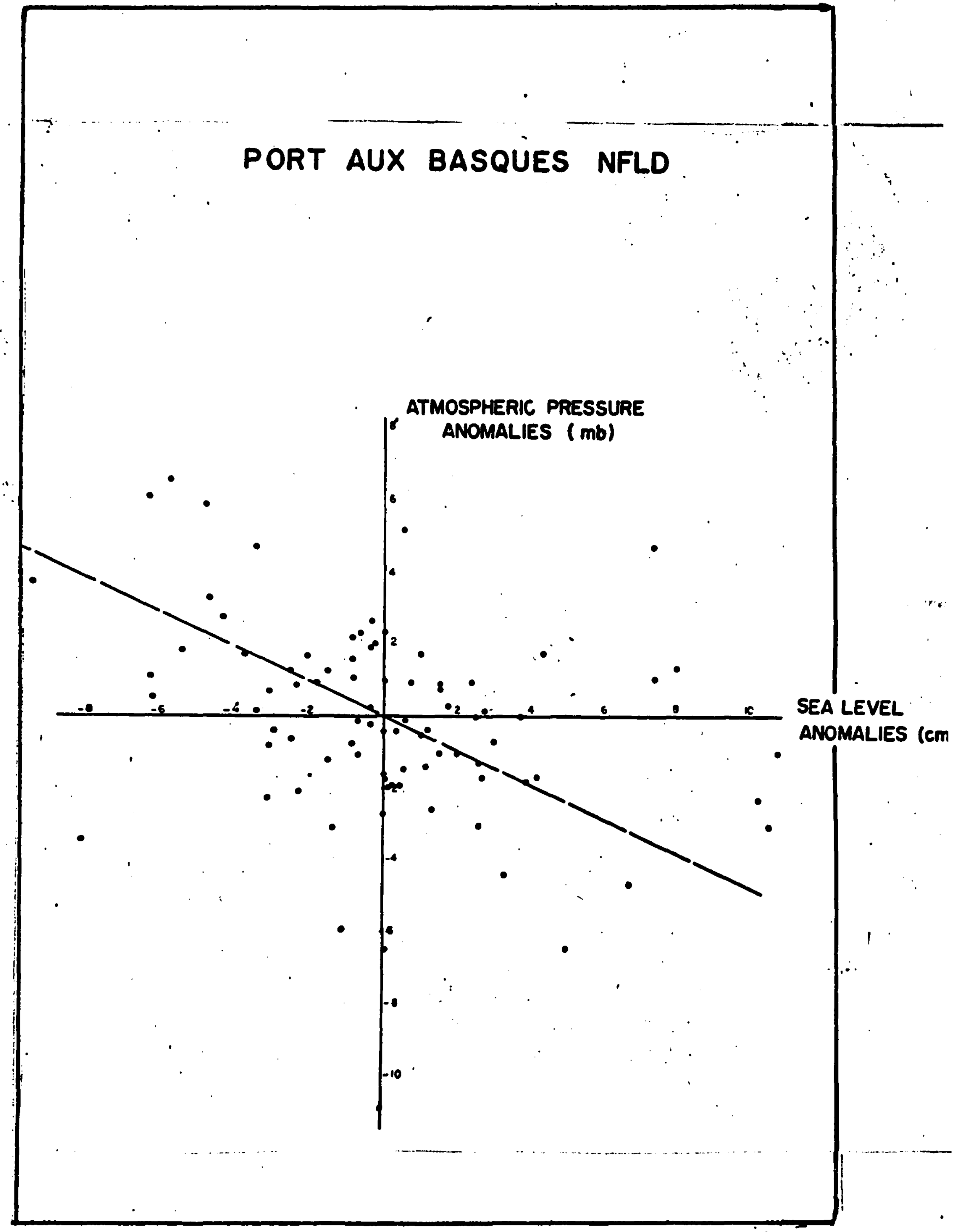

FIG. 8 Relation between Sea level and Atmospheric Pressure at Port aux Basques, Nfld.

Figures 6, 7 and 8 have been included to show the scatter of the points for the stations showing best and least correlation. It can be seen that the greatest deviation from the pure hydrostatic 1:1 relationship occurs at either very high pressures or at very low. The addition of the seasonal variation in pressure and sea level has, in general, reduced the correspondence in the variables. Changes in atmospheric pressure can only account for about 28% of the sea level variation (given by the square of the correlation coefficient, $r^2 \times 100\%$), seasonal effects eliminated.

Table 7 - Correlation and Regression between Atmospheric Pressure and Sea Level

Station	r	SE	Regression Equation	Sign
Harrington	-0.49	4.91	$X_1 = 881.5 - 0.77X_2$	S
Charlottetown	-0.45	4.85	$X_1 = 938.2 - 0.77X_2$	S
Port aux Basques	-0.47	4.38	$X_1 = 725.9 - 0.61X_2$	S
Halifax	-0.55	4.53	$X_1 = 781.5 - 0.65X_2$	S
St. John's	-0.59	8.08	$X_1 = 1262.5 - 1.17X_2$	S

X_1 = sea level in cm

X_2 = atmospheric pressure in mb

2.2.2 Wind Stress

As was seen in the previous section, regression of sea level on pressure yielded coefficients which differed from the pure hydrostatic law. Since wind and pressure are also correlated over distances of a few hundred kilometers, the wind effect was determined using a step-wise multiple regression procedure as outlined by Efroymson (1960).

The atmospheric pressure effect was first eliminated from the records using the calculated regression coefficients of Table 6. To give a rough approximation of the wind at any particular locality, the monthly mean values of the "run of the wind " decomposed into the eight principal compass directions was used to compute a net north and east component wind by the following:

$$C_N = S - N - (SW - SE - NW - NE) \cos 45^{\circ} \tag{5}$$

$$C_E = W - E - (NW - SW - NE - SE) \cos 45^{\circ} \tag{6}$$

with the resultant given by:

$$R = (C_N^2 + C_E^2)^{\frac{1}{2}} \tag{7}$$

in the direction:

$$\Theta = \tan^{-1} C_N / C_E \tag{8}$$

where C_N and C_E are the component winds from south to north and from west to east respectively. Direction is measured positive anticlockwise from the east. From equations (7) and (8) on-offshore and along-shore components of the wind were calculated.

The above method of calculating the wind effect is far from satisfactory. Implied is the belief that the local wind field is homogeneous to some distance offshore, and the cal-

calculated value is the one acting on the water surface. The conditions are naturally not perfect. For nearly all stations bottom topography will exert some degree of channelling to the wind. The use of only eight compass points also reduces the accuracy of the results, especially where the winds are strong in the directions not recorded.

For the open coast a more suitable method would probably have been the calculation of geostrophic winds from monthly mean pressure maps. However, even this method is not free from errors- especially in enclosed seas.

Correlation and regression analysis as carried out between corrected sea level deviations and on/off-shore and alongshore component of wind deviations squared (with retention in sign) are presented in Tables 8 and 12 to 31 in Appendix A.

Table 8

Correlation and Regression between corrected Sea Level deviations and Wind deviations squared

Station	r_o	r^1	Regression Eqn.	Positive Dir.
Harrington	0.17	-0.30	$Y = -0.003\ L^2$	↗
Charlottetown	0.10	0.34	$Y = 0.016\ L^2$	↖
Port aux Basques	0.47	-0.08	$Y = 0.058\ O^2$	↑
Halifax	0.40	-0.25	$Y = 0.028\ O^2$	↗
St. John's	No significant correlation found.			

where r_o = correlation coefficient between the on-off-shore component of the wind and sea level.

r^1 = correlation coefficient between the alongshore component of the wind and sea level. Positive direction of the wind indicated by arrows.

O = onshore - offshore component of the wind in Km/hr.

L = alongshore component of the wind in Km/hr.

The correlation coefficients between wind and sea level deviations shown in Table 8, are much smaller than those given in Table 6 for atmospheric pressure deviations. The greatest correlation between the onshore - offshore component of the wind was at Port aux Basques and is greater than that found for atmospheric pressure. In this case, the onshore winds will give high sea levels because of the direction chosen - from the southeast. Ekman (1905) has shown that for stationary winds acting on a water surface, a corresponding surface drift results whose net transport is directed to the right of the wind stress at an angle of 45°. This holds for northern latitudes only. Hence, a wind blowing with the coast on its left (looking downstream), would create low sea levels. This relation is borne out in the signs of the correlation coefficients. From Table 8, it can be seen that a positive longshore component in the direction NW would create high sea levels at Charlottetown. The onshore correlation is small. This could be due to the fact that the fetch is limited, or that the winds are not truly representative.

For those areas not fetch-limited and bordering deep waters, a higher wind speed would have to occur for the same sea level deviation as recorded in shallower water. This follows from the fact that for any column of water the horizontal force, due to the wind stress, is independent of the water depth. However, horizontal forces caused by pressure gradients, resulting from sea surface slopes, are proportional to the depth, hence a greater wind stress is needed in deep water to balance a given slope.

2.2.3 Thermohaline Effects

It is well known that changes in the specific volume of sea water brought about by changes in temperature and salinity result in a well marked seasonal oscillation of sea level (Nomitsu and Okamoto, 1927; LaFond, 1939; LaFond and Rao, 1954). By analysing temperature and salinity values at depth for many areas, Patullo et al (1955) were able to show that the seasonal change in sea level, as caused by changes in the water volume, was the dominant factor in tropical and middle latitudes. Latitude 40°N was indicated as being a boundary north of which, the steric effect could not be calculated with any reliability due to a lack of deep salinity observations.

To calculate this steric effect, a long and regular series of observations of temperature and salinity at depth are necessary. Continuous sampling to a depth of 175 m , at the average rate of once a month, is carried out by the Fisheries Research Board in St. John's at station 27. Other sections of subsurface measurements are taken irregularly by the Atlantic Oceanographic Group located in Halifax. Because of this irregularity, any calculation based on these values will have large errors.

The steric sea level is defined by (Rossiter,1962):

$$z = \frac{1}{g} \int_{P_o}^{P_a} \Delta\alpha \, dp \qquad (9)$$

where z = steric level of the sea surface with respect to a level which shows no seasonal variation,

$\Delta\alpha$ = specific volume anomaly,

P_a = atmospheric pressure,

P_o = isobaric level where all seasonal effects cease,

g = acceleration of gravity.

Using values of σ_t, we can calculate the "order of magnitude" of the steric oscillations as follows (Nomitsu and Okamoto, 1927):

Let $\sigma_t = (\rho_t - 1) \times 10^3$ (10)

and γ_t the specific volume and ρ_t = density

then $\Delta\alpha = \gamma_t'' - \gamma_t = (\sigma_t'' - \sigma_t) /1000$ (11)

where σ_t'' is some reference density.

Using the 300 db surface as a reference layer in Cabot Strait, the values of density were integrated from the surface using Simpson's rule. The results of the calculations for Port aux Basques are given in Table 9. Fig. 9 shows the effect of the "isostatic"*correction on the seasonal variation. Similar calculations for a station in the Halifax section did not give any consistent results.

Fig. 10 shows the seasonal variation of temperature, salinity and computed density for station #27 according to Bailey (1961).

Judging from the isopleths of σ_t, no level exists where seasonal effects die out. The variations of σ_t would indicate that density changes could produce low sea levels in late winter, followed by high levels in summer and early autumn.

2.3 Spatial Correlation of Sea Level

To give a statistical estimate of the coherence of sea level fluctuations, a correlation analysis was made on the monthly mean uncorrected sea level (uncorrected for atmospheric pressure), using St. John's as the dependent variable. The results are presented in Table 10.

* "isostatic compensation" is a term applied by Patullo et al. (1955), to changes in sea level which did not cause a corresponding change in pressure at the bottom of the sea. That is, it is a volumetric expan sion or contraction.

Table 9 - Effect of Density on Sea Level (Port aux Basques, Newfoundland)

Month		1	2	3	4	5	6	7	8	9	10	11	12	Range
DEPTH (m)	0	25.50	25.67	25.58	25.50	25.30	24.66	23.11	23.24	23.50	24.62	24.99	25.25	2.56
	50	25.51	25.80	25.75	25.75	25.68	25.72	25.81	25.66	25.84	25.93	25.41	25.46	0.52
	100	26.20	26.32	25.98	26.05	26.07	26.19	26.30	26.25	26.34	26.45	26.36	26.28	0.47
	150	26.65	26.70	26.56	26.55	26.62	26.61	26.81	26.64	26.54	26.77	26.75	26.70	0.27
	200	27.03	26.88	26.89	26.70	27.03	26.98	27.11	27.09	27.01	27.13	27.03	27.02	0.43
	250	27.16	27.07	27.23	27.05	27.24	27.14	27.26	27.24	27.21	27.27	27.20	27.18	0.22
	300	27.37	27.23	27.30	27.15	27.38	27.29	27.37	27.57	27.40	27.43	27.36	27.35	0.42
Effect of Density (cm)		-1.50	-2.15	-0.81	1.23	-1.21	0.20	-0.88	1.07	0.55	-2.90	0.40	-0.45	4.13

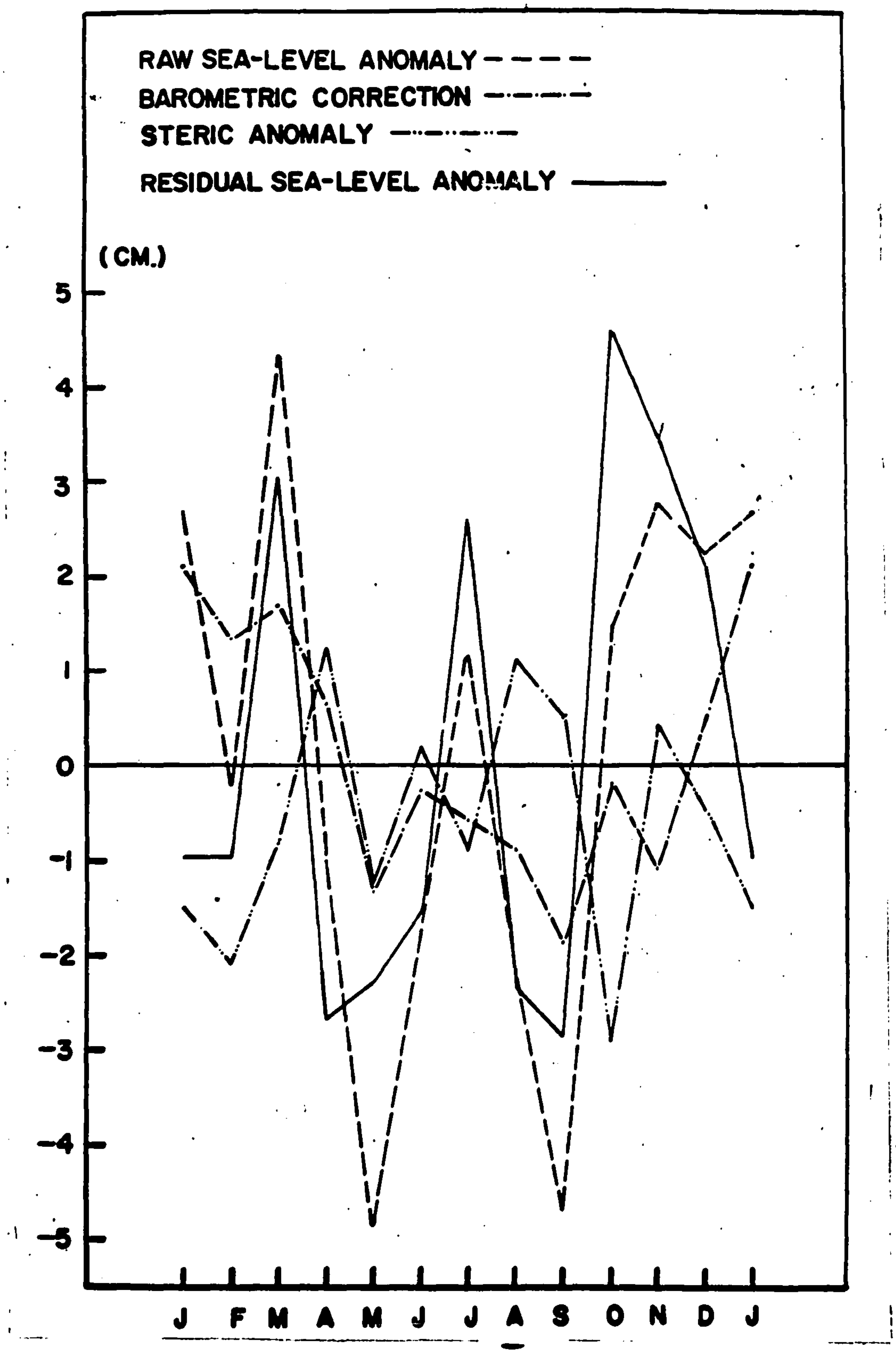

Figure 9 - Time series of raw sea level, isostatic corrections, and residual sea level anomaly (Port aux Basques)

Table 10

Correlation of Uncorrected Sea Level Between Stations

Tide gauge	1	2	3	4	5
1	1.00	0.68	0.49	0.59	0.43
2		1.00	0.65	0.52	0.48
3			1.00	0.73	0.56
4				1.00	0.60
5					1.00

where 1 Harrington Harbour, Qué.
2 Charlottetown, P.E.I.
3 Port aux Basques, Nfld.
4 Halifax, N.S.
5 St.John's, Nfld.

Table 10 indicates that the linear correlation of sea level between adjacent stations is fair with the correlation usually decreasing in magnitude with distance from any particular station. It should also be noted that all the coefficients were positive, implying that at zero time lag the variations usually occur "in phase". This is borne out in Fig.3. The results might, however, be changed if the lags are considered in correlation.

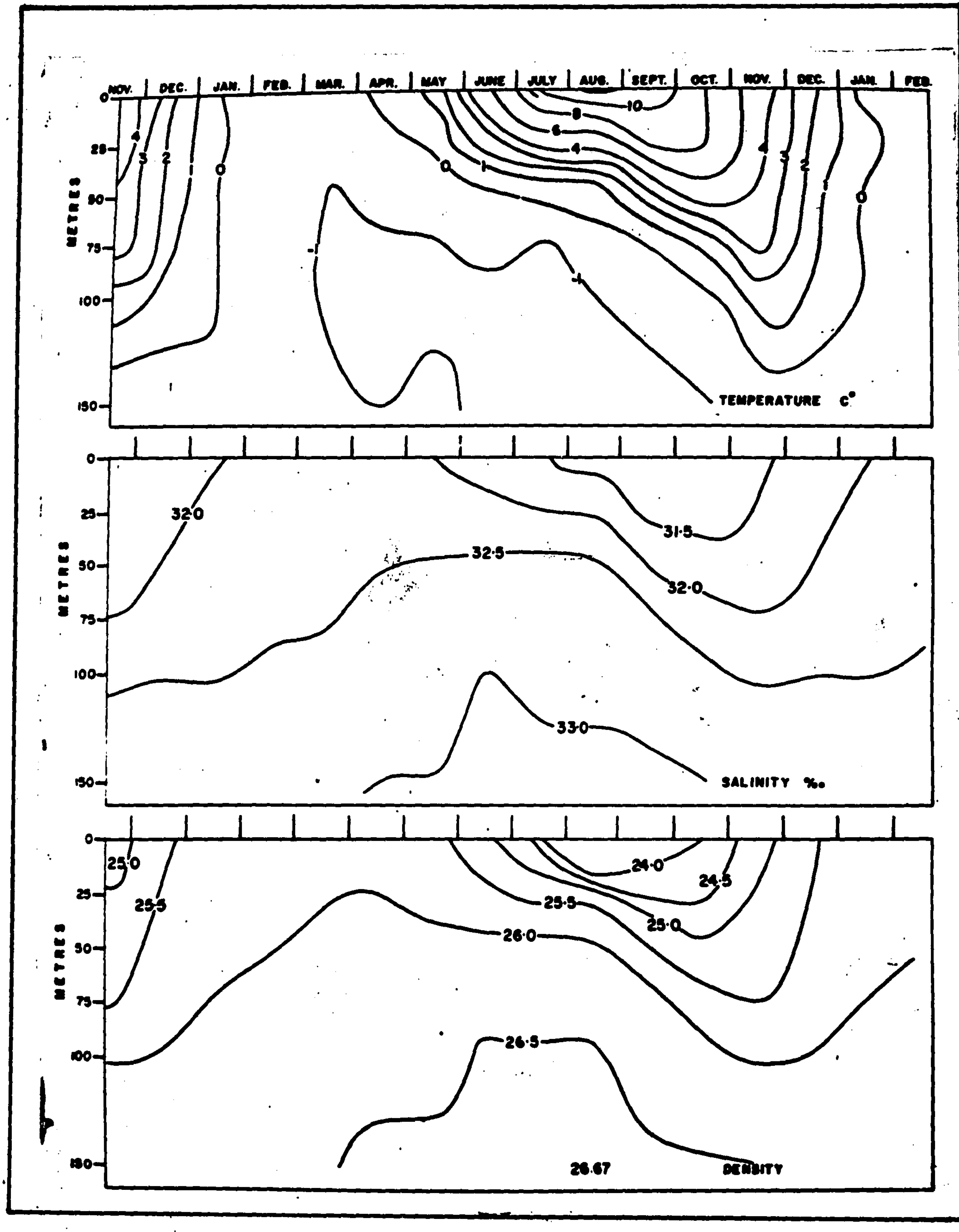

Figure 10 - Seasonal variation of Temperature, Salinity, and Density, Station 27. (after Bailey, 1961).

2.4 Discussion

Although showing large variability from month to month, sea levels in the Gulf of St.Lawrence and on the Atlantic coast follow the same pattern of high sea level in the fall and winter, followed by low levels in the summer. Atmospheric pressure and prevailing on- and offshore winds were equally responsible for explaining the variation.

Port aux Basques showed a relative maximum in the summer, which could not be explained by any isostatic factor and hence, must be due to some other variable not considered. Because the time of maximum fell in the summer, this peak might be explained by the effect of fresh water runoff.

Waldichuck (1963), analyzing sea level variation in the Strait of Georgia on the west coast of Canada, found a similar spike in the sea level curves. He was able to show that this was a direct effect of the fresh water volume in the Strait.

The only other station felt to be able to exhibit such a spike was Charlottetown. Sea level rises in the summer,but not as pronounced as at Port aux Basques. In addition, the water structure in these shallow areas (Magdalen shallows) is frequently changed by the winds (Lauzier, 1952). This, combined with tidal mixing in Northumberland Strait, makes any inference of the effect of runoff difficult, although runoff does affect the salinity values (Lauzier, 1957).

The effects of the winds were significant in all cases, usually being as important as the pressure effect. From the regression analysis, it is felt that the wind-driven currents are of greater importance than the "windstau", in causing a rise or fall in sea level.

The importance of an ice cover in reducing sea surface slopes has been studied by Lisitzin (1957). Applying the steady state formula of Ekman (1905), relating slope to surface stress, she was able to show that substantial reductions of sea surface slope occurred during the winter months. The success of her method of analysis should, however, be discussed in order to show why the same procedures cannot be applied to the Gulf of St. Lawrence.

Most importantly, the greatest part of the water level fluctuations are caused by the wind. This is due to the fact that the Bothnian Bay is orientated in a NNE-SSW direction which favours the action of the prevailing winds. In addition, the Bay is very shallow and of a narrow but uniform width. Numerous tide gauges, placed more or less uniformly around the Bay, makes the determination of the sea surface slope quite feasible.

Using 15 years of data (sea level and wind observations) 40 cases were selected where uniform and high winds blew along the longitudinal axis of the Gulf of Finland. Similar reasoning cannot be applied to the Gulf of St. Lawrence. On a monthly basis, the correlation of sea level between stations decreased rapidly with distance. This fact, combined with the non-uniform distribution of the tide gauges (a total of 8) makes the determination of sea surface slopes unreliable. Lastly, the use of the local wind creates certain errors in the determination of the slope which, at high winds, can be appreciable.

The sea surface slope , under steady state conditions, can be determined by the following (Neuman and Pierson, 1966):

$$\beta = \frac{\tau}{2\rho g d}\left[3-n\right] \qquad (12)$$

$$0<n<1$$

d = depth of the sea

τ = wind stress

β = slope

ρ = density of the water column

We can thus see that if equal weight is given to the action of onshore or offshore winds in creating sea surface slopes, it could happen that for a zero net wind, a certain sea surface slope would remain.

With the limited data available, a somewhat crude attempt to determine the qualitative effect of an ice cover was made by studying the variability of mean sea level and other meteorological variables on a month to month basis. This would then indicate which factors control the variation in sea level the most. Table 2 and 3 indicated that the variation of variability, expressed by the standard deviation, ran parallel to the mean annual variation.

The question arises as to whether the differences in the standard deviations over the season are real, or, whether they can be attributed to random effects. Considering Harrington, the average variance for the summer months (June, July and August) is 10.24 cm^2, while the average variance for winter (January, February, March) is 12.46 cm^2. In order to test for significance at the 5 % level a chi-squared test was applied. The 95% confidence limits for six degrees of freedom are 12.59. That is, 95 % of the time $1.635 < \chi^2 < 12.59$. Assuming the distribution to be normal, we set $\chi^2 = \frac{Ns^2}{\sigma^2}$. This gives us a value of $\chi^2 = 8.51$. Since this value lies within the 95% confidence limit for χ^2, it may be assumed that the difference in the variance is not significant.

A similar analysis as the above was made on the regression and correlation coefficients for two periods covering maximum and minimum ice concentration periods. Results are given in Appendix B, Tables 27 to 31. Taking Harrington as an example again, we find that correlation between pressure and sea level does not vary over the two periods. The onshore component of wind correlated significantly with sea level in summer, but not at all during winter. Although it might be concluded that the presence of an ice cover might be responsible for this reduction, it should be noted that the standard error in the dependent variable is still quite high. In addition, the low number of data points (summer, 49 and 35 for winter) does not really make this difference significant.

CHAPTER III

DAILY VARIATIONS IN SEA LEVEL

3.1 Introductory Remarks

The results of the foregoing chapter have indicated the complexity of the ocean-atmosphere interaction leading to variations in sea level. Unfortunately, these results may not be extrapolated to variations occurring with periods much shorter than one month. Day to day fluctuations in sea level were analyzed with the view in determining the presence of any statistically significant periodicities, the nature of the response of the sea surface to atmospheric inputs and, the cause of any differences in the above calculated response factors for the winter and summer periods; notable among such possible causes could be the formation of an ice cover in winter.

Treating the recorded sea level as the output of a linear system (for the scale of motion involved, this is a good first approximation), the mechanism(s) which couple random inputs of atmospheric pressure and wind to sea level can be studied using correlation and spectral functions. An analysis in the frequency domain is preferable as high and low frequency variations are usually related differently.

Thirteen months (March 1964 to March 1965) of hourly sea level, three-hourly values of atmospheric pressure and north-south, east-west component winds were analyzed for Grindstone, Magdalen Islands, P.Q. After preliminary filtering to reduce or eliminate tidal noise, auto- and cross spectral information was computed using the methods as outlined by Tukey (1949), Blackman and Tukey (1959) and Goodman (1957). Variations in the response of sea level to atmospheric inputs were studied by splitting the records into summer (April to September) and winter (October to March) periods. No attempts were made to ascribe the variations in response found for the two different periods to any specific ice concentrations or thickness. Basically, this restriction was forced on the study as reliable and accurate information on these parameters were, and are not, available.

One weakness in applying the so-called Tuckey methods to the estimation of the spectrums is that they require the time series of observations to remain stationary. Whereas this can be approximated during the summer months by careful filtering, this assumption will not in general hold during the winter months due to the melting-freezing process, drifting of ice. Although the stationary assumption might not strictly hold, useful information can still be derived under weaker assumptions.

Before presenting the principal results , the design of suitable filter functions which smooth the time series from unwanted high frequency variations are discussed. This is followed by a section outlining the theoretical basis for the spectral analysis computations.

3.2 Design of Preliminary Filter Functions

Because tidal variations in the sea level usually overshadow all other variations, it is most important that any tide remover be so designed as not to affect the other variaitions to any extent. Several low pass filters were applied to the original data. The filtered records were then sampled at half-day intervals giving a Nyquist frequency of 1 cycle per day.

A statistical filter is a type of moving average consisting of a sequence of weights (usually fractional values) which are convoluted to a sequence of observations, leading to a filtered value. The linear algebraic operation in the time domain can be represented by

$$Y_k = \sum_{j=-n}^{m} W_j \cdot X_{k+j} \tag{1}$$

where $Y_k = k^{th}$ member of the filtered series

$W_j = j^{th}$ weighting coefficient

$X_k = k^{th}$ observation in the original record

For the general case of smoothing where past and future data are available, it is usual to consider symmetrical filters. That is, $W_j = W_{-j}$ and n=m=N. With this restriction, it can be shown (Holloway, 1958) that phase relations between original and filtered records are preserved. Moreover, for a given number of weights, the symmetrical filter is better than any other. The application of the (2N + 1) weight factors to the original data can now be represented by

$$Y_k = W_o X_k + \sum_{j=1}^{N} W_j \left(X_{k+j} + X_{k-j}\right) \tag{2}$$

The measure of a filter's effectiveness in removing or modifying the amplitudes of the various frequency bands in the original data is given by the filter's frequency response or transfer function R(f). It is derived from the inverse Fourier transform of the weighting array W_j (Holloway, 1958; Panofsky and Brier, 1958):

$$\begin{aligned} R(f) &= \sum_{j=-N}^{N} W_j \cos 2\pi f j \\ &= W_o + 2 \sum_{j=1}^{N} W_j \cos 2\pi f j \end{aligned} \tag{3}$$

where f = frequency in cycles per data interval.

The actual synthesis of the sequence W_j corresponding to a given R(f) can be accomplished in several ways. In our case, we have determined the weighting sequence by specifying the Transfer function at selected frequencies. Then, by transforming R(f) we obtain:

$$W(t) = 2 \int_0^\infty R(f) \cos 2\pi ft \, df \qquad (4)$$

The integral in equation (4) can be approximated by some quadrature rule. The corresponding frequency limits in the summation will run from 0 to 1 cycle per 2 data intervals.

The atmospheric pressure and wind data were tapered by a weighting array generated such as to give a Transfer function as shown in Fig. 11. The curve is not exactly reproduced as an infinite number of weights would be needed. The effect of truncating the weights at some finite number is to introduce substantial ripples outside the main lobe. These could be reduced by increasing the number of weights. Unfortunately, this procedure is not always possible as a loss of 2N data points occurs. The required number of weights finally chosen usually represents a compromise between data length requirements and side lobe heights.

The above procedure for generating the sequence W_j subject to the constraints set by R(f) was programmed on a digital computer following the suggestions made by McCulloch (1965). Filters which adequately suppress the high frequency tides were

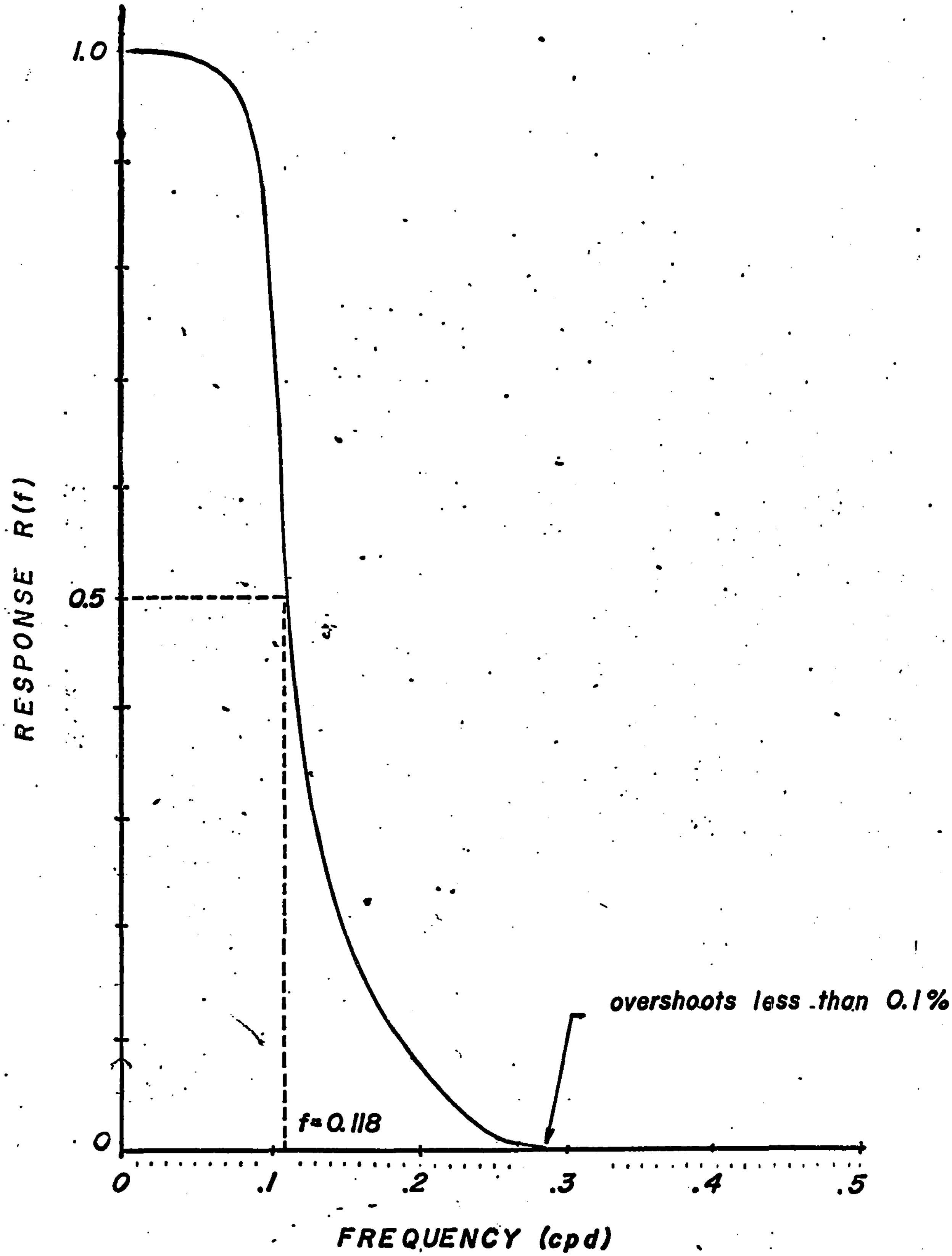

FIG.11 Atmospheric Pressure and Wind filter

already available in the literature (e.g. Doodson and Warburg, 1941; Groves, 1955; Godin, 1966,1967) and thus it only remained to chose the most suitable. A composite filter function constructed from a diffraction function and smoothed with a Hanning window (see Blackman and Tukey , 1959, pg.98) is shown in Fig. 12 together with the response characteristics of the other filters considered. Frequency units in Fig. 11 and Fig. 12 are in terms of cycles per data interval.

The arithmetic average, although using the least number of weights is, in fact, a very poor filter. Its large overshoots at the higher frequencies could lead to large sampling variability and erroneous spectral estimates. It also permits substantial leakage from tidal phenomena whose periods are not integral multiples of 25 hours. Continuing in this vein, it can be seen that the Doodson filter is a more effective smoother even though it uses a larger number of weights. It eliminates the lines of the tidal spectrum however, its large side lobes also exhibit positive and negative responses which would lead to errors. The final choice thus rests on using either the Godin filters or the Hanning-Lanczos filter.

The latter filter function was eventually utilized . In a direct comparison, the Godin filter although using less weights and having small side lobe heights actually attenuates part of the wave phenomena in the frequency band of interest. In this respect, the Hanning-Lanczos filter approaches the

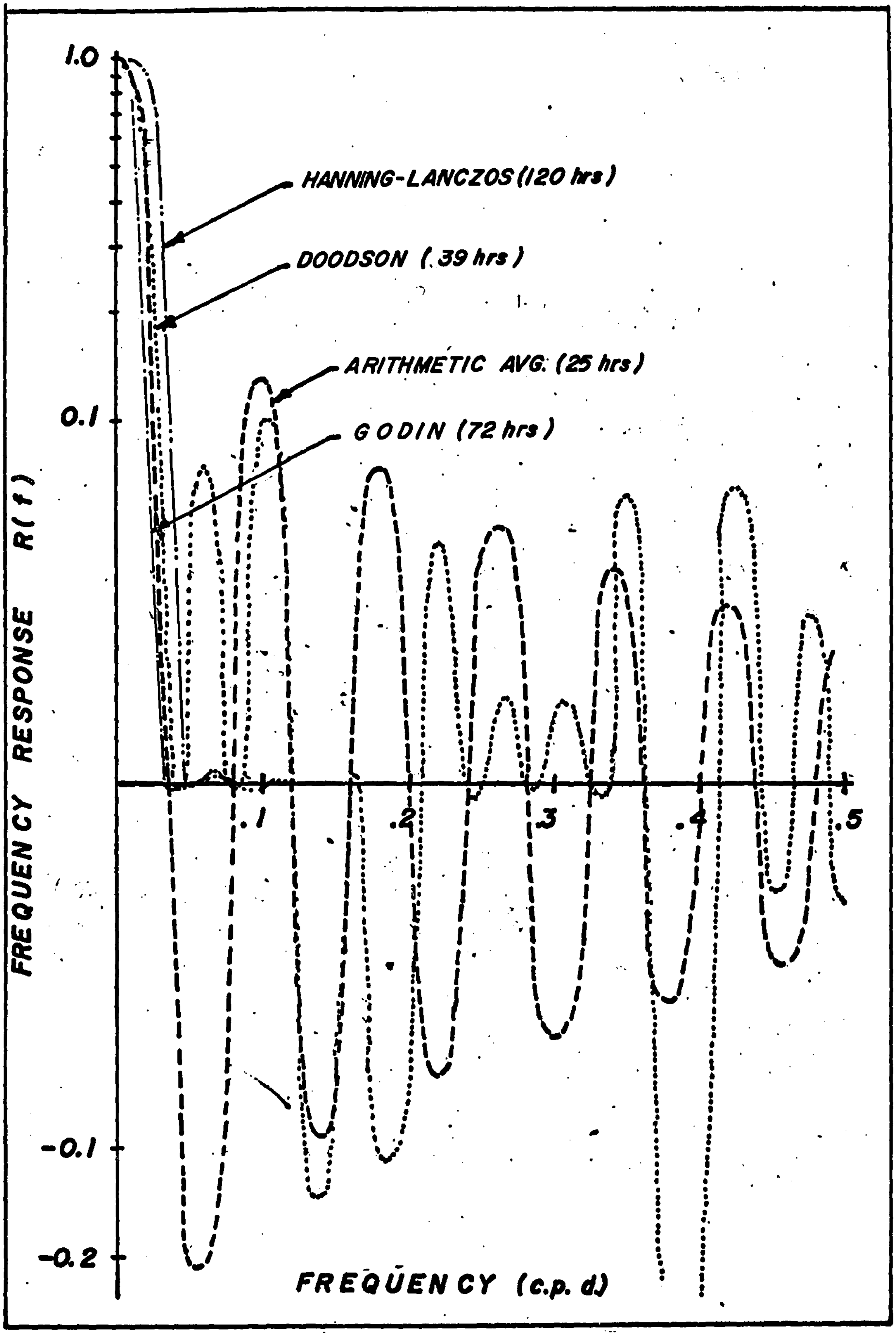

FIG. 12 Frequency Response characteristics of Tide filters

ideal closest by giving a sharp cutoff. This quality more than offsets the increased number of weights and the incom-plete elimination of some diurnal tidal energy. In order to additionally reduce diurnal tides and waves appearing with fortnightly periods, a low pass filter together with Munk's triangular filter with m=8 (Munk et al.,1959, pg.300) was used.

3.3 Spectral Analysis Theory

Most geophysical time series are nondeterministic, consequently their description and prediction are based on probability statements and statistical averages, rather than explicit equations.

We assume that our finite piece of record for analysis is a particular realization drawn from a stationary, ergodic random process. That is, it represents only one of a finite or infinite collection of possible results wich could have occurred. Using the stationary assumption assures us that the probability of any event occurring during the series record is constant. The ergodic hypothesis permits us to replace, with unit probability, the ensemble average of the process (the average of the process over both, time and space, domain from $-\infty$ to ∞) with the time average of a single series in the ensemble (Bendat, 1958; Lee, 1960).

Suitably restricted then, the spectrum of any time series yields the distribution of its variance as a function of frequency. Although the spectrum can be obtained by numerical, electrical filters or classical harmonic analysis, the best estimates of the true "population" spectrum are obtained by a Fourier transform of the sample autocovariance function. For a single stationary series x(t) with zero mean, the exact autocovariance function at lag τ may be defined (Granger and Hatanaka, 1964) as:

$$C_x(\tau) = \lim_{T\to\infty} \frac{1}{T}\int_{-T/2}^{T/2} x(t)\cdot x(t+\tau)\,dt \qquad (1)$$

whose properties are:

$$C_x(\tau) = C_x(-\tau) \quad \text{an even function of } \tau \qquad (1a)$$

$$C_x(0) = \sigma_x^2 = \text{variance of } x(t) \qquad (1b)$$

$$C_x(0) \geq |C_x(\tau)| \quad \text{for all } \tau \qquad (1c)$$

As shown by Blackman and Tukey (1959), $C_x(\tau)$ may also be expressed by the Fourier transform of a distribution function $P_x(f)$. Thus,

$$C_x(\tau) = \int_{-\infty}^{\infty} P_x(f)\exp(2\,i f\pi\tau)\,df \qquad (2)$$

where

$$P_x(f) = \lim_{T\to\infty}\left|\frac{1}{T}\int_{-T/2}^{T/2} x(t)\exp(-2\pi i f t)\,dt\right|^2 \qquad (3)$$

$P_x(f)$ is called the spectral distribution function or spectral density function for the stationary process $x(t)$ and represents the contribution to the variance of $x(t)$ from frequencies between f and $f+df$. We can invert (2) such that $P_x(f)$ is expressed as the Fourier transform of $C_x(\tau)$:

$$P_x(f) = \int_{-\infty}^{\infty} C_x(\tau)\exp(-2\pi i f\tau)\,d\tau \qquad (4)$$

using the symmetry property (1a), we may write (2) and (4) as:

$$C_x(\tau) = 2 \int_0^{\infty} P_x(f) \cos 2\pi f \tau \, df \qquad (5)$$

$$P_x(f) = 2 \int_0^{\infty} C_x(\tau) \cos 2\pi f \tau \, d\tau \qquad (6)$$

Note that

$$C_x(0) = 2 \int_0^{\infty} P_x(f) \, df = \sigma_x^2$$

Thus, $P_x(f)$ is still considered as the two-sided even function containing only half the total variance in the positive frequency range.

The discrete and finite nature of our data does not permit us to evaluate the autocovariance and spectral density function as defined by (5) and (6). We define the sample autocovariance function by

$$C_x'(\tau) = C_x'(k\Delta t) = \frac{1}{N-k} \sum_{n=1}^{N-k} x_n \cdot x_{n+k} \qquad (7)$$

where

x_i = sample data whose mean is zero

N = total number of data points

Δt = digitizing interval

k = lag number = $0,1,2,\ldots\ldots\ldots,m$

m = maximum lag

τ = time lag = $k\Delta t$

The maximum lag $\tau_{max} = m\Delta t$ is usually set such that $m = N/10$ where $N\Delta t$ equals the total time span of the record. This, represents a compromise between the equivalent bandwidth resolution of the spectral density estimates and their stability.

Assuming that the sampling interval Δt is small enough such that no appreciable variance is contributed to the spectrum above the Nyquist or cutoff frequency $f_c = 1/2\Delta t$, we can calculate the "raw" estimates of the spectral density function over the frequency range $0 \leq f \leq f_c$. Applying a finite discrete cosine series transform to $C'_{ox}(\tau)$, where

$$C'_{ox}(\tau) = D_o(\tau)\, C'_x(\tau) \quad \text{for all } \tau \tag{8}$$

and

$$D_o(\tau) = \begin{cases} 1 & 0 \leq \tau < m\Delta t \\ \frac{1}{2} & \tau = m\Delta t \\ 0 & \text{otherwise} \end{cases} \tag{8a}$$

we get:

$$P'_x(k) = \frac{\delta_k}{\pi}\left[C'_x(0) + 2\sum_{r=1}^{m-1} D_o(r)\, C'_x(r)\cos\pi\frac{kr}{m} + (-1)^k C'_x(m)\right]$$

where

$$\delta_k \begin{cases} = \frac{1}{2} & k=0,m \\ = 1 & 0 < k < m \end{cases} \tag{9}$$

and,

$$k = 0,1,2,3,4,\ldots\ldots,m$$

Here k represents the index number of a particular frequency defined by $f_k = k/2m\Delta t = kf_c/m$.

By the convolution theorem, it follows that the "raw" estimates of the spectral density function $P'_x(k)$ actually represent a weighted average of the true spectrum $P_x(f)$ in the vicinity of each frequency k.

The value of $P'_x(k)$ at any k is the sum of the contributions of the whole spectrum weighted according to the functional form of the spectral window $Q_o(f)$ shown in Fig. 13. $Q_o(f)$ and $D_o(\tau)$ form a Fourier transform pair.

$Q_o(f)$ has undesirable large side lobes, these being about 1/5 the height of the main lobe. In order to obtain the best estimates $P'_x(f)$ near the nominal frequency, the main lobe of any spectral window $Q_i(f)$ should be concentrated near the nominal frequency with the side lobes as low as possible. It is easily shown (Kinsman, 1965; Jones, 1964) that these are contrary requirements. Widening $Q_i(f)$ means narrowing its transform $D_i(\tau)$ and concentration on values of $C'_x(\tau)$ near $\tau = 0$. A compromise is made by smoothing the raw estimates by a sequence of weights $a_{k,i}$ as follows.

$$\hat{P}_x(k) = \sum_{i=0}^{m} a_{k,i} P'_x(k) \qquad (10)$$

where

$a_{k,i} = 0$ for $i \neq k, k-1, k+1$

	Hanning	Hamming
$a_{k,k-1}$	0.25	0.23
$a_{k,k}$	0.50	0.54
$a_{k,k+1}$	0.25	0.23

and k assumes the values from 1 to m-1 .

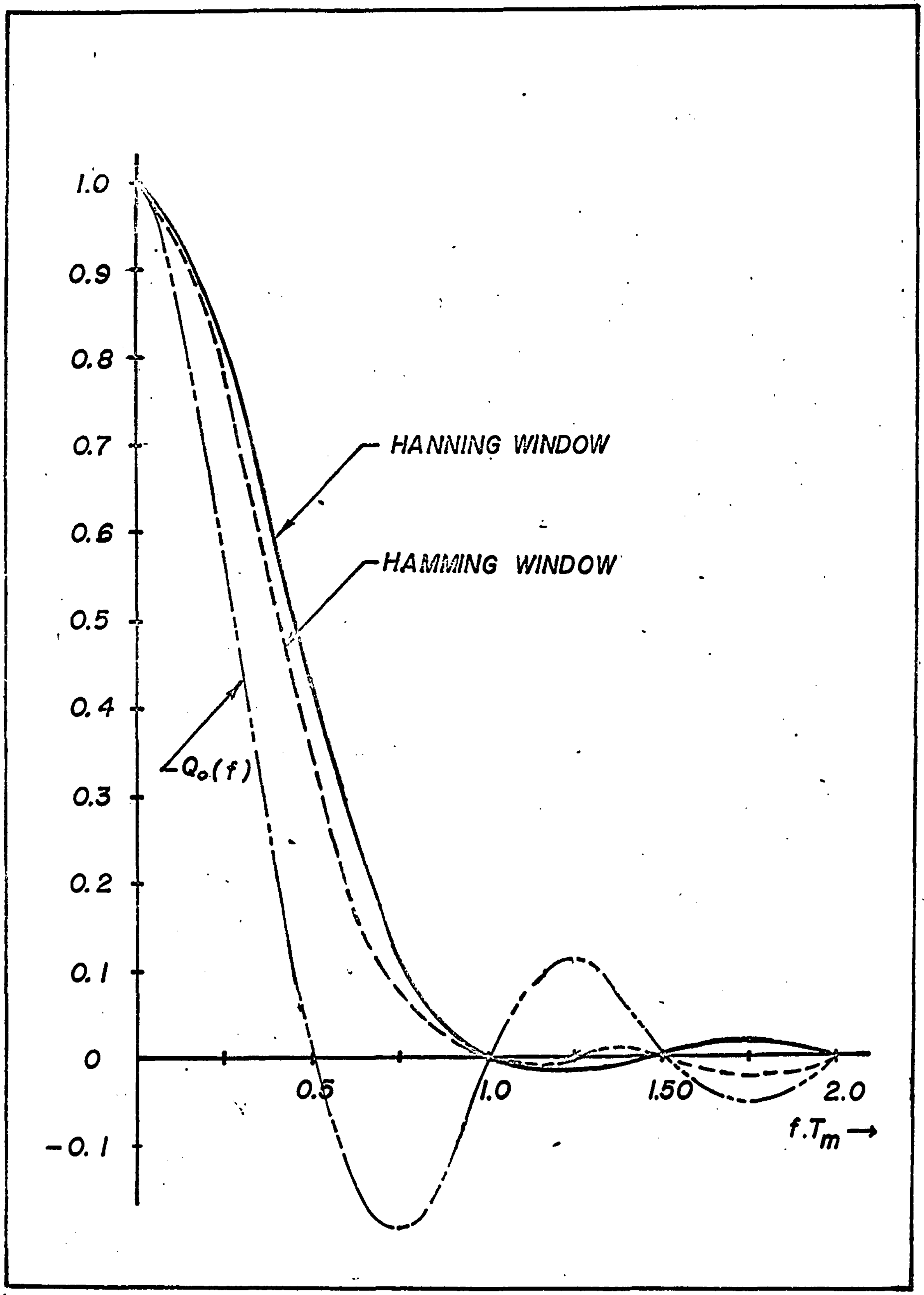

FIG. 13- Spectral Windows

For the end points, the estimates are adjusted as follows:

The Hamming window $Q_3(f)$

$$\hat{P}_x(0) = 0.54\ P'_x + 0.46\ P'_x(1)$$

$$\hat{P}_x(m) = 0.46\ P'_x(m-1) + 0.54\ P'_x(m)$$

Using the Hanning window $Q_2(f)$

$$\hat{P}_x(0) = 0.50\ P'_x(0) + 0.50\ P'_x(1)$$

$$\hat{P}_x(m) = 0.50\ P'_x(m-1) + 0.50\ P'_x(m)$$

Both these windows are also shown in Fig. 13 in order to reveal their relative merits. Several other smoothing filters have been described. A discussion of their properties is given by Jenkins (1961).

Cross covariance functions characterize the degree of coupling between x(t) and y(t).

As before, we consider that both time series belong to a stationary, ergodic Gaussian random process. The unbiased estimates of the sample cross covariance function for lags $k\Delta t$, where $k = 0,1,2,\ldots\ldots,m$ is given by:

$$C'_{xy}(k\Delta t) = 1/N-k \sum_{n=1}^{N-k} x_n \cdot y_{n-k} \qquad (11)$$

$$C'_{yx}(k\Delta t) = 1/N-k \sum_{n=1}^{N-k} y_n \cdot x_{n-k} \qquad (12)$$

where $C'_{xy}(k\Delta t) = C'_{yx}(-k\Delta t) \qquad (13)$

$C'_{xy}(k\Delta t)$ is always a real valued function which may be either positive or negative. It is not necessarily an even function as contrasted with the autocovariance. In general, it is neither even nor odd. However, it may be considered to be the sum of an even function (covariance) and an odd function (quadrature variance) as given below for $k = 0,1,\ldots\ldots,m$. The covariance function is defined as:

$$E'(k\Delta t) = 1/2 \left[C'_{xy}(k\Delta t) + C'_{yx}(k\Delta t) \right] \qquad (14)$$

and the quadrature variance as:

$$Q'(k\Delta t) = 1/2 \left[C'_{xy}(k\Delta t) - C'_{yx}(k\Delta t) \right] \qquad (15)$$

As with the case for a single time series, each of (14) and (15) are subjected to a Fourier analysis which, for the case of a covariance function, yields "raw" estimates of the cospectral density function. Similar analysis on the estimates of $Q'(k\Delta t)$ will yield the "raw" estimates of the quad-spectral density function $Q'_{xy}(k)$. Each of the $m+1$ "raw" estimates are then smoothed, using the same sequence of weights as given by (10).

In order to study the time-lag relationships between the functions $x(t)$ and $y(t)$ we form the phase diagram by plotting the phase differences $\theta_{xy}(k)$ against k. Thus we can write:

$$\theta_{xy}(k) = \tan^{-1} \left[\frac{\hat{Q}_{xy}(k)}{\hat{CO}_{xy}(k)} \right] \qquad (16)$$

for $\hat{Q}_{xy}(k) > 0 \qquad 0 < \theta < \pi$

and $\hat{Q}_{xy}(k) < 0 \qquad \pi < \theta < 2\pi$

The coherence function $CH^2_{xy}(k)$ can be written as:

$$CH^2_{xy}(k) = \left[\frac{\hat{CO}^2_{xy}(k) + \hat{Q}^2_{xy}(k)}{\hat{P}_x(k)\ \hat{P}_y(k)} \right] \qquad (17)$$

where $0 \leq CH^2_{xy}(k) \leq 1$.

It may be thought of in analogy to the square of the correlation coefficient between two variables x and y. It measures the degree of association of the two series for various frequencies. The absence of coherence does not necessarily mean that x(t) and y(t) are independent. It suggests, however, that no linear relation exists between the variables, or that extraneous errors are present (Cartwright et al., 1962).

3.4 Results

The energy density distribution of sea level, atmospheric pressure, south to north and west to east component winds are shown in Figs. 14 to 18. The summer and winter spectra are based on 364 and 366 data points respectively and were evaluated for $m = 30$. In this case, the degrees of freedom for estimating the precision of the computed energy density values are given by $DF = (2N - m/2)/m = 24$ (Panofsky and Brier, 1958). The 95% confidence limits are shown by the vertical arrows.

The spectra have been plotted on a log-ordinate vs. linear frequency scale in order to give equal resolution at all ranges of the spectrum. Unfortunately, this has the disadvantage of distorting the relative magnitudes of the energy distribution versus frequency.

Confidence limits for coherence and phase depend not only upon the coherence that would be obtained from an infinite record (R_∞) but also, on the degrees of freedom (Goodman, 1957; Munk et al., 1959). For $DF = 24$ and $R_\infty = 0.50$, the 95% limit for coherence is 0.40 and the limit for phase is $\pm 31°$.

3.4.1 Spectrum of Sea Level, Atmospheric Pressure and Wind

Fig. 14 shows the energy density distribution of daily mean sea level and atmospheric pressure for the period April, 1964 to March, 1965.

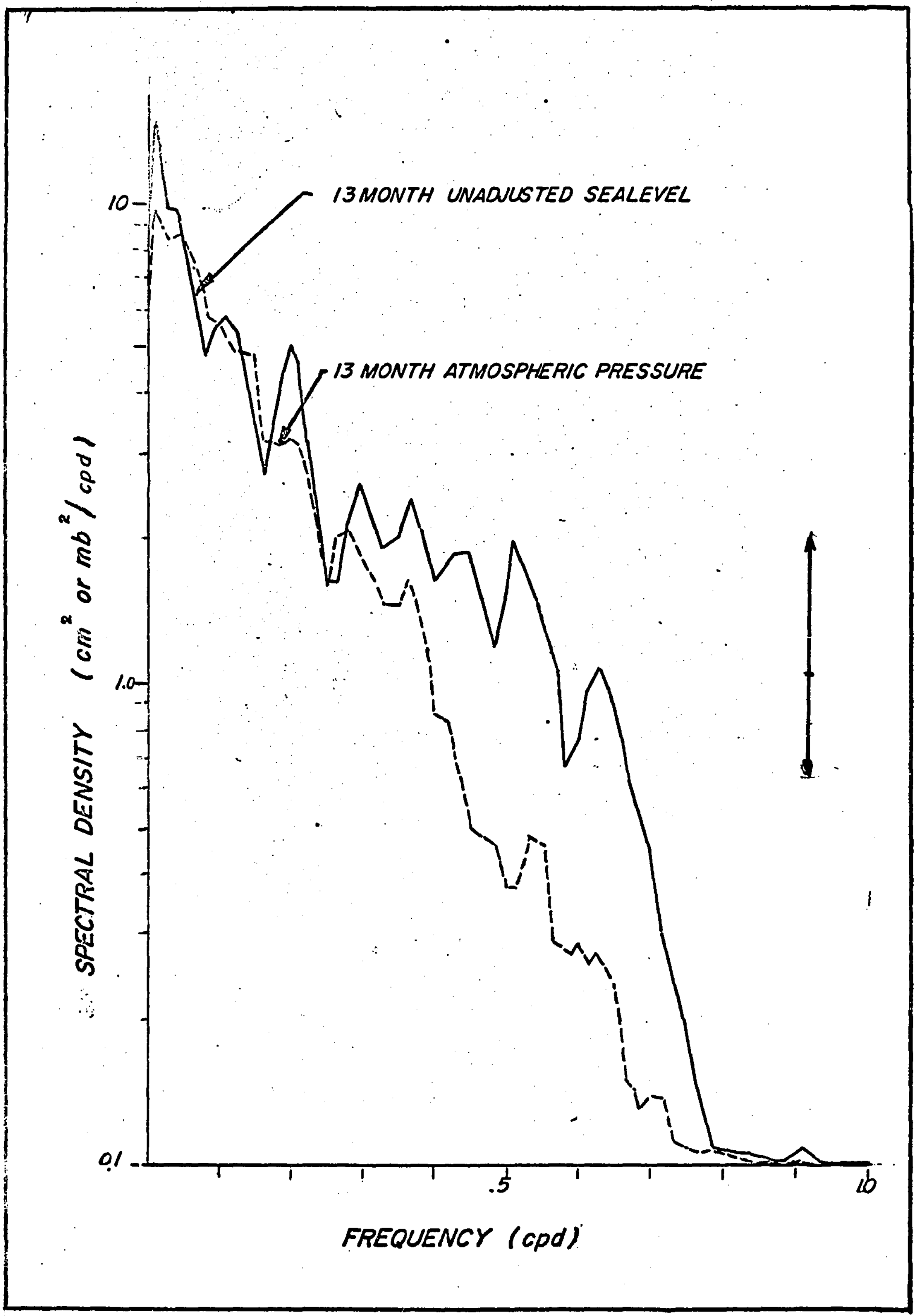

FIG. 14 Spectrum of Sea Level and Atmospheric Pressure

A weak peak in the sea level spectrum is apparent at 0.2 cpd and also at 0.55 cpd. No significant periodicities can be ascertained in the atmospheric pressure spectrum. It can be seen that most of the variance in both spectra are associated with fluctuations between the frequencies of o.o5 and 0.5 cpd (corressponding to periods of 20 and 2 days). Below 0.4 cpd atmospheric pressure and sea level spectra are very similar. It is conceivable that this part of the spectrum is associated with the movement of major weather systems from west to east.

Above 0.3 cpd, the atmospheric pressure spectrum decays more rapidly than does the sea level spectrum. In conjunction with their respective autocorrelation functions, it would seem that the noise level in the sea level spectrum is higher in this frequency band or, sea level variations with periods greater than two days are more presistent than corressponding changes in the pressure variations.

Figs. 15 and 16 represent the summer and winter spectrums of sea level, adjusted sea level and atmospheric pressure. Adjusted sea levels, Z_t, were calculated from the expression as given by Hamon (1962):

$$Z_t = Y_t + 0.06\ X_{t-1} + 0.88\ X_t + 0.06\ X_{t+1} \tag{18}$$

where Y_t = the daily mean sea level
X_t = daily mean atmospheric pressure
t = time in ½ days

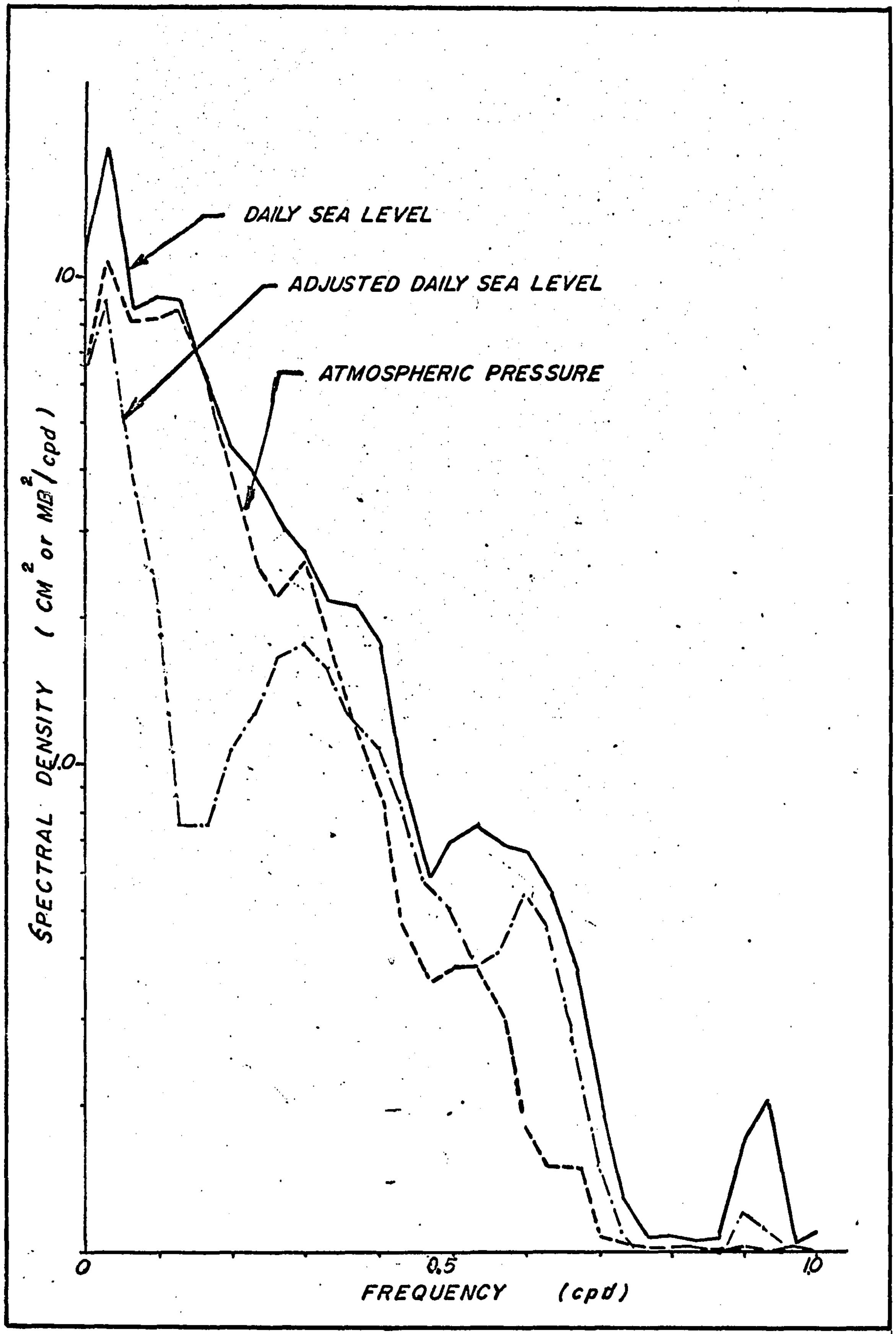

Fig. 15 SUMMER SEA LEVEL AND ATMOSPHERIC PRESSURE SPECTRA

It can be shown that this expression is equivalent to using the usual hydrostatic relation (1 cm decrease in sea level occurs for each pressure rise of 1 mb) at low frequencies, but also makes a small adjustment at high frequencies for the filters used to compute mean sea level and atmospheric pressure.

The summer spectrums of sea level and atmospheric pressure tend to exhibit the same gross features as depicted in Fig. 14. The use of a wider window has reduced some of the variability of the estimates. The weak peak in the sea level spectrum at 0.2 cpd is no longer present although the variance in the band centered about this frequency has increased. A significant peak in the sea level and pressure spectrums occurs at approximately 0.06 cpd. The weak peak in the sea level spectrum of Fig. 14 centered at 0.55 cpd is no longer present and the variance in the band centered on this frequency has been reduced substantially.

The interesting feature in Fig. 15 is however the distribution of energy density of adjusted summer sea levels. This shows a broad peak centered at the frequency of 0.3 cpd. In general, the 1:1 inverse adjustment has decreased the variance of summer sea level in all bands. Notable exceptions occur above 0.5 cpd and could perhaps be due to the different numerical filters used for sea level and atmospheric pressure.

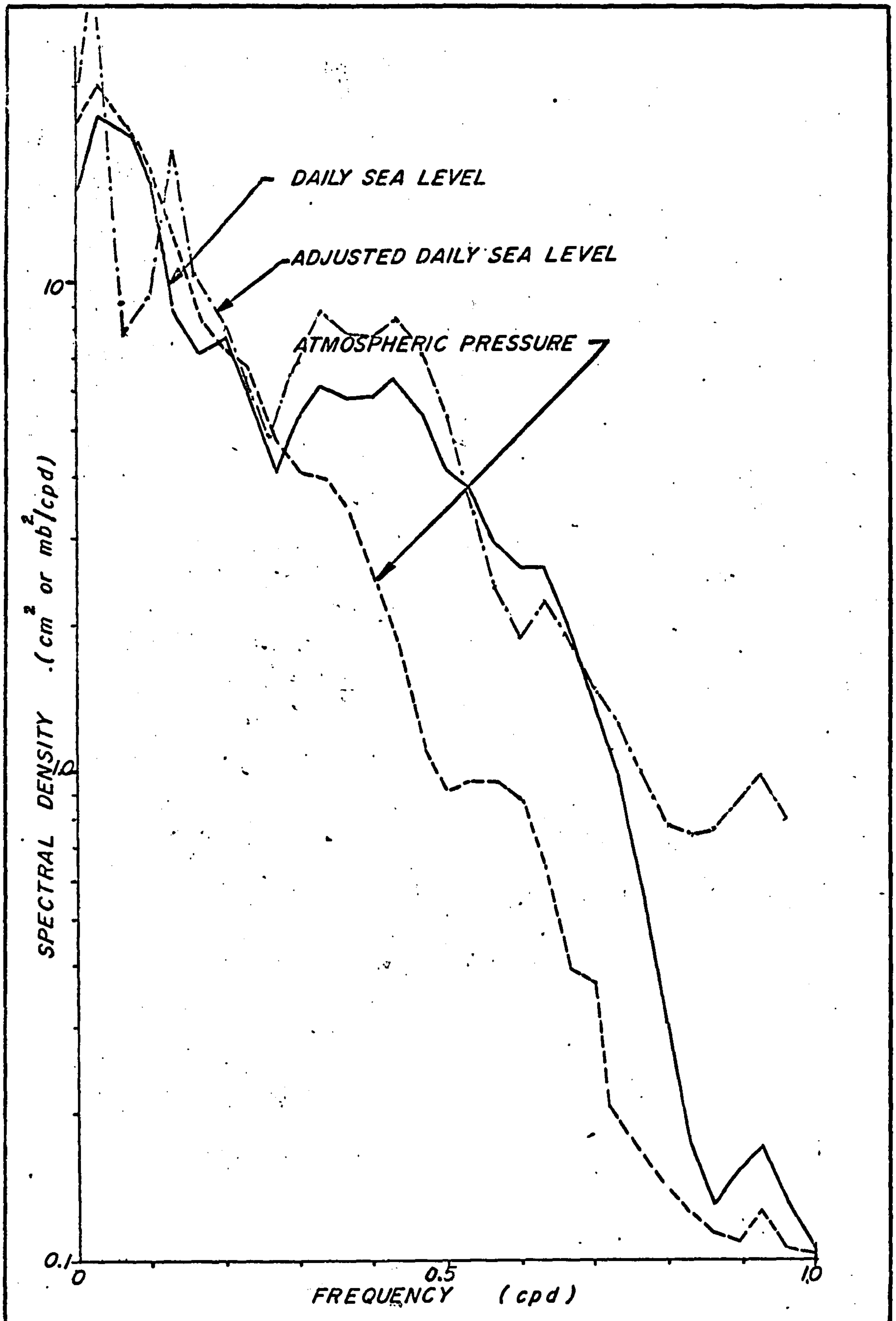

Fig.16 *WINTER SEA LEVEL AND ATMOSPHERIC PRESSURE SPECTRA*

The introduction of a peak at 0.3 cpd in the summer adjusted sea level spectrum leads one to believe that some other factor, perhaps the wind, correlated with pressure has also influenced the sea level. The high energy reduction in the adjusted sea levels in the band from 0.06 to 0.3 cpd indicates that sea level and atmospheric pressure are probably correlated and in phase. This is also borne out by the cross-spectrum analysis shown in Fig. 19.

A comparison of the winter (Fig.16) and summer (Fig.15) spectrum of sea level, atmospheric pressure and adjusted sea levels shows that the winter spectrums have a higher energy content with a significant shift towards higher frequencies in winter. The shift is however not uniform over the whole band. In contrast to the summer period, the isostatic adjustments has increased the spectral estimates of winter sea levels. Particular increase has occurred near the pressure peaks centered at 0.35 cpd and also above 0.65 cpd. A significant peak now occurs in adjusted winter sea levels at 0.15 cpd (period of 6.7 days). It can be seen that during the winter season the hydrostatic hypothesis is a poorer approximation throughout the frequency range. Additional support for the above is also given in Fig. 20.

Summer and winter spectrums of south to north and west to east component winds are presented in Figs. 17 and 18.

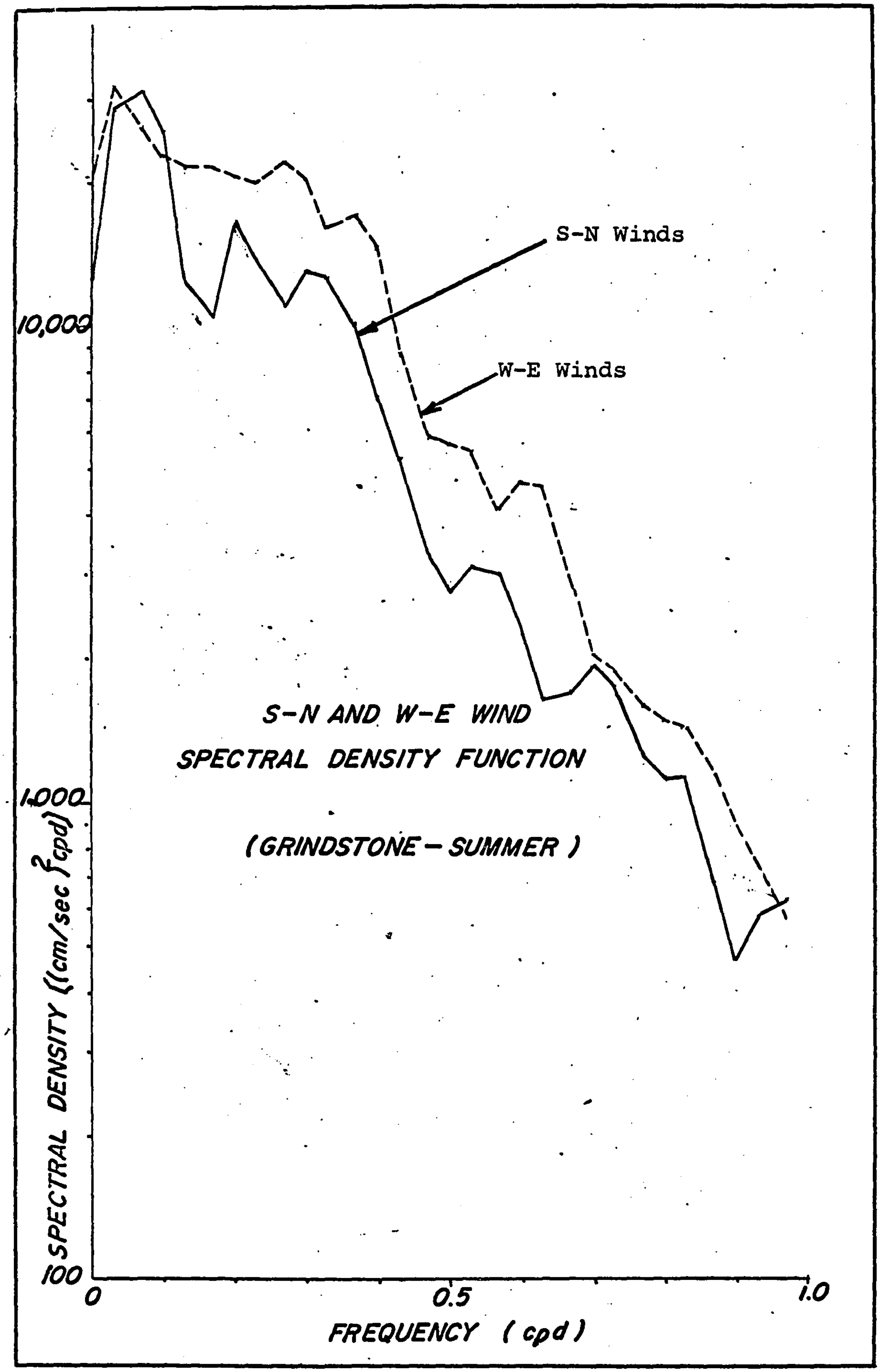

Fig. 17 Summer Component Wind Spectrums

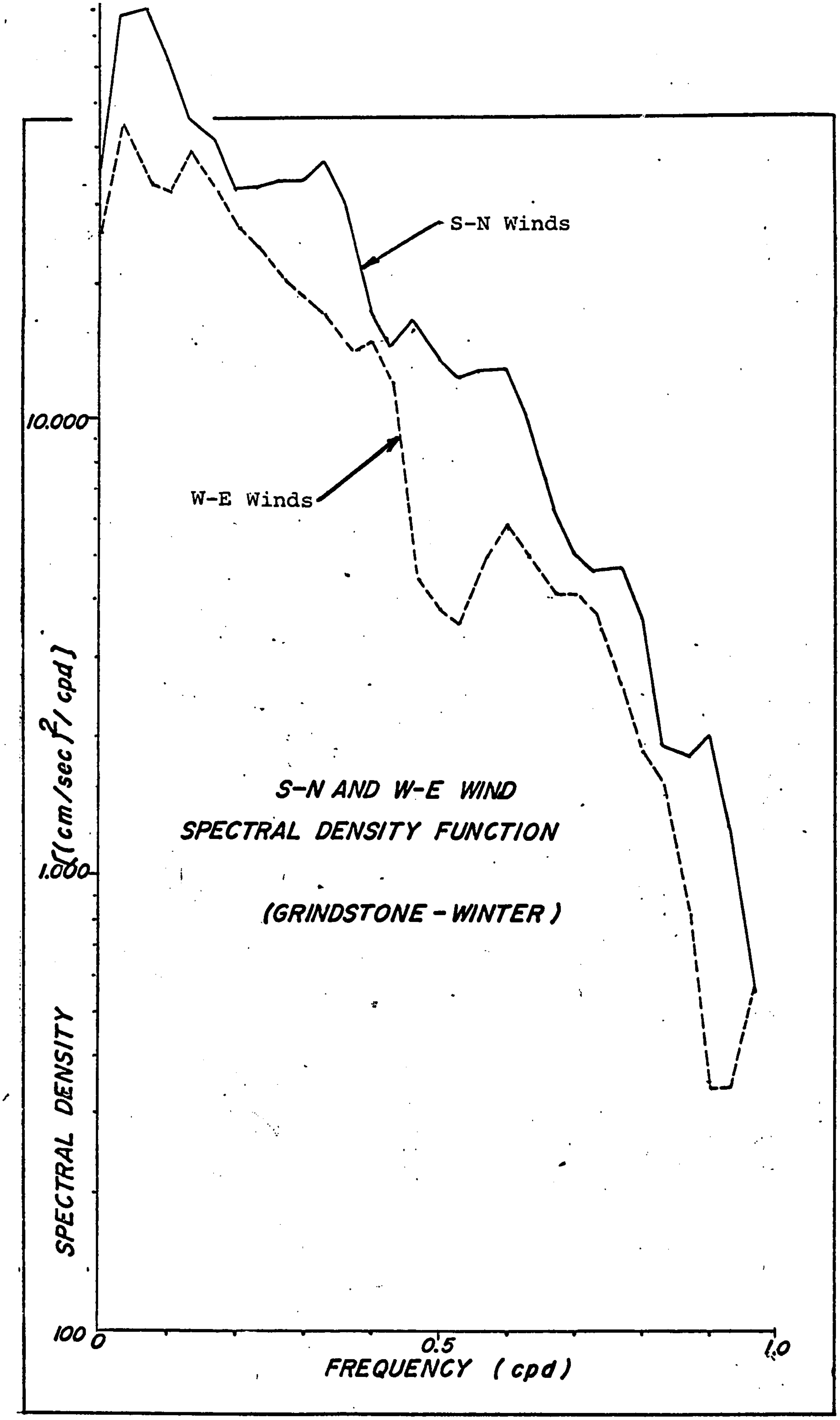

Fig. 18 Winter Component Wind Spectrums

It should be stated in advance that the wind spectrums are less reliable than the estimates for atmospheric pressure and sea level. The winds represent one minute averages centered on the synoptic hour and in all probability suffer from high frequency sampling errors. Panofsky and Brier (1959, pg.146) state that in order to attain stable wind estimates at the surface winds should be averaged for at least 30 minutes.

The summer (S-N) wind spectrum shows a significant peak centered at 0.1 cpd with a slight rise centered at 0.3 cpd. Energy density estimates of (W-E) winds are generally higher than corressponding (S-N) spectrum estimates. Winter spectrums as shown in Fig. 18 do not show any significant peaks. As contrasted to the summer period , the (S-N) winds show higher variability over the whole frequency range as compared to (W-E) winds. This perhaps is an indication of the fact that during winter, the dominant winds are from the north or north-west while summer shows the dominant winds from the south or south-west. Winter wind variability for both components has also increased by an order of magnitude.

3.4.2 Frequency Response of Sea Level to Atmospheric Pressure

To examine the relationship between sea level and atmospheric pressure, the methods of cross spectral analysis are particularly useful. It enables us to evaluate the systems response function and also the time delays as a function of frequency. Figs. 19

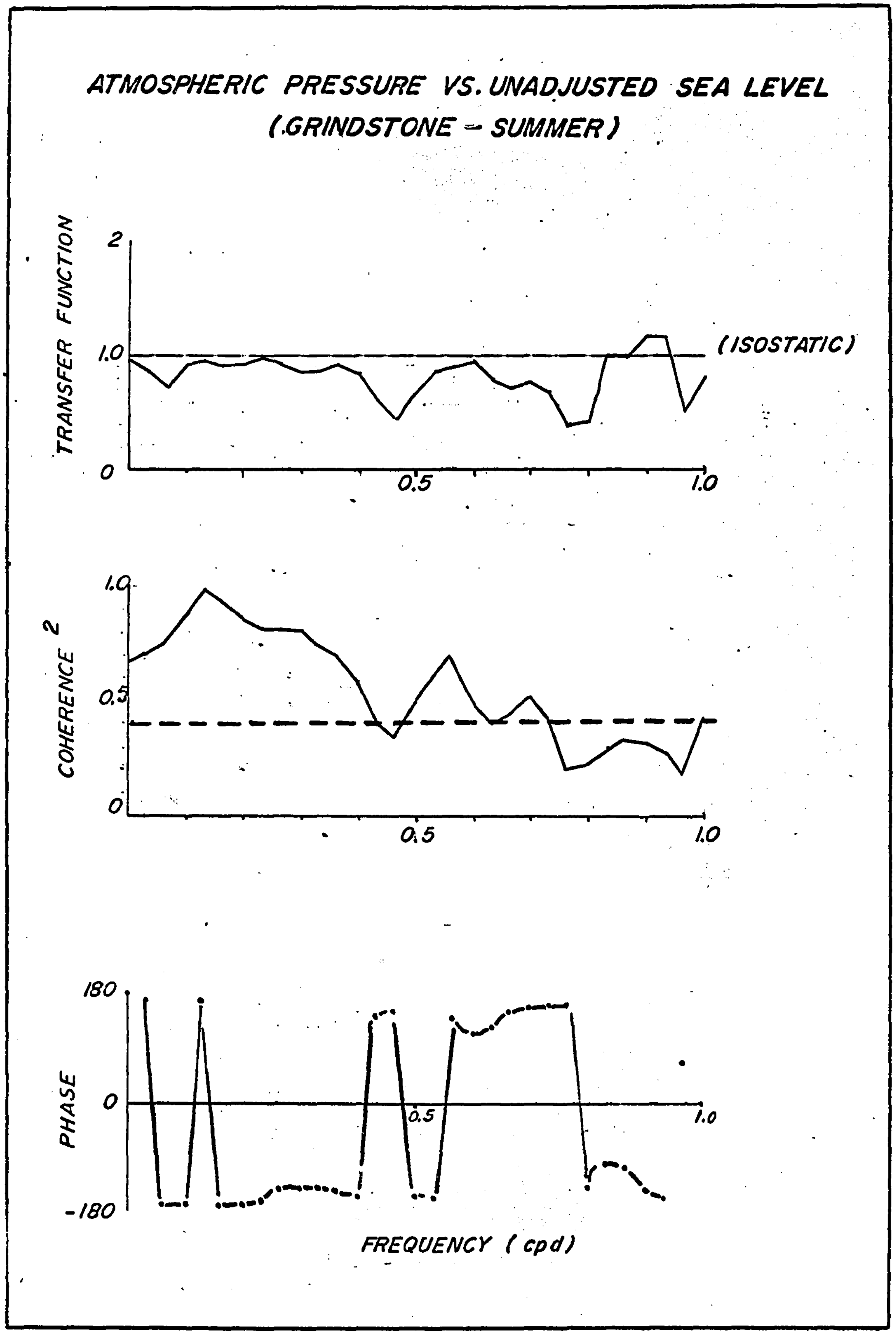

Fig. 19 Cross-spectrum between Sea Level and Atmospheric Pressure

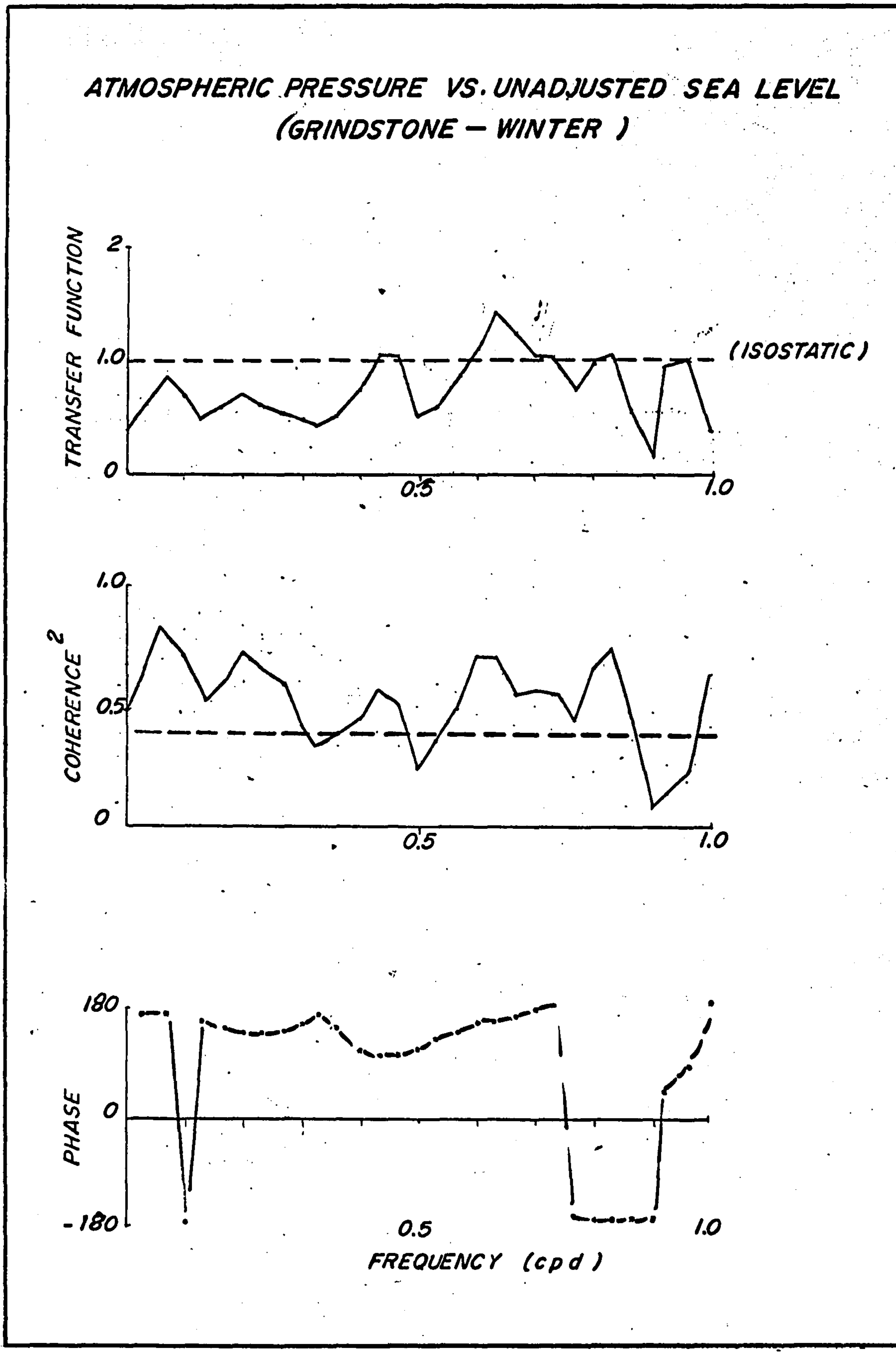

Fig. 20 Cross-spectrum between Sea Level and Atmospheric pressure

and 20 show the results of the cross spectral analysis. The upper graph shows the systems transfer function as a function of frequency. For a physically realizable, stable system such as the ocean/atmosphere boundary the concept of frequency response function and transfer function is interchangeable without loss of any useful information (Bendat and Piersol, 1966).

The summer relationships (Fig.19) shows the transfer function close to the isostatic value of 1 cm/-1 mb for frequencies between 0 and 0.45 cpd. Over this frequency interval the general level of coherence is significantly high lending additional support to the computed values of the transfer function. Above 0.55 cpd the level of coherence drops below significance. It is interesting to note that the peak in the coherence spectrum centered at 0.55 cpd also coincides with the frequency where summer sea levels rise in energy density.

The relationship between unadjusted sea levels and atmospheric pressure during winter (Fig. 20) is considerably different from that which occurs in the summer. It is most evident that the transfer function is below the isostatic value for frequencies less than 0.5 cpd but rises to a peak cntered at 0.62 cpd which is greater than the isostatic value. The coherence spectrum also has a peak at this frequency.

The phase spectrums support, in general, the fact that the input/output relations change from winter to summer. Over the frequency range from 0.12 to 0.7 cpd the winter phase func-

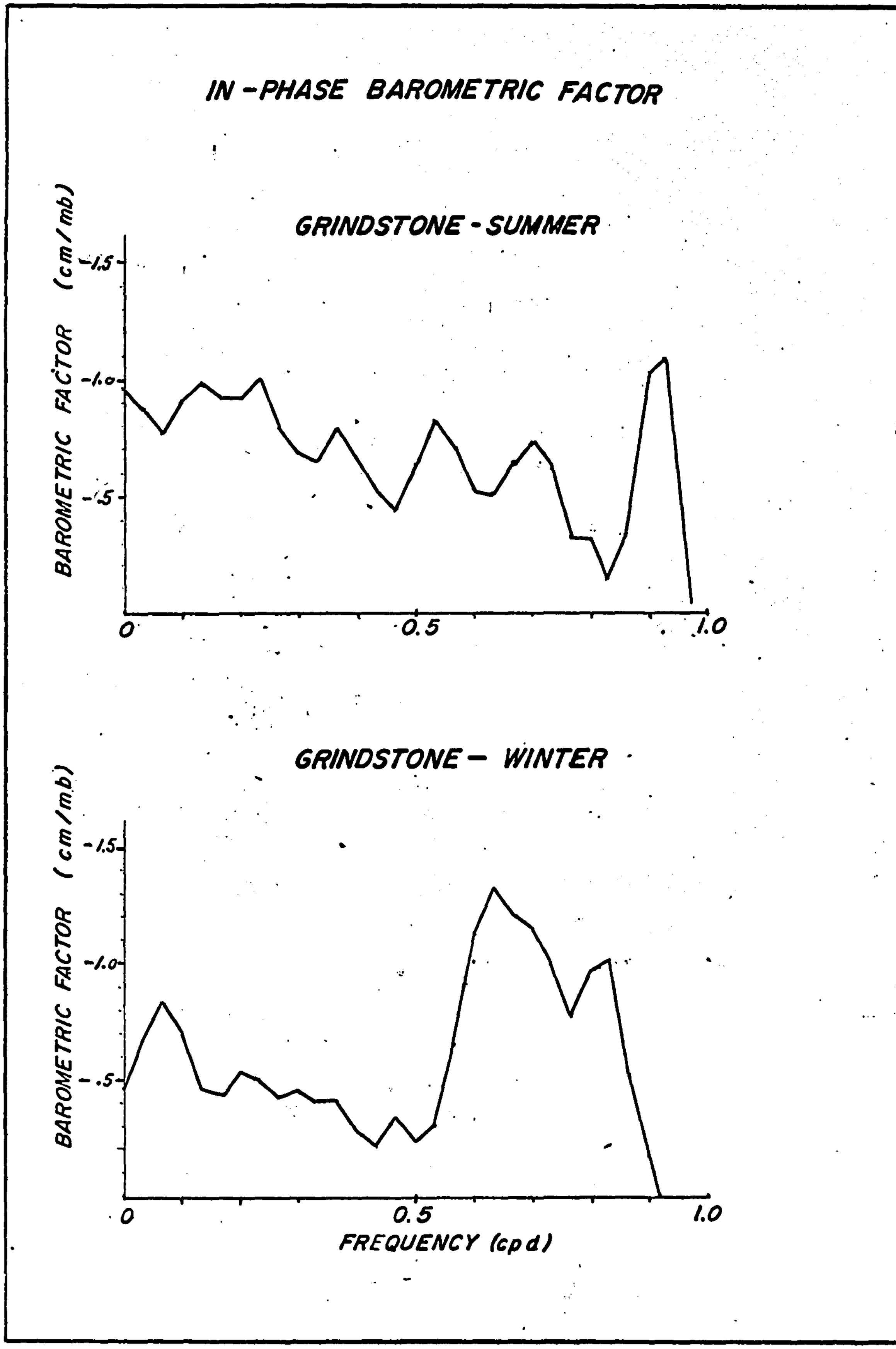

Fig. 21 In-phase barometric factor b(k)

tion shows that sea level lags atmospheric pressure by about 150°. In view of the confidence interval for phase, this is close to the isostatic relationship. The rapid changes in phase in the vicinity of 0.4 to 0.5 cpd cannot really be significant because of the low coherence at these frequencies.

Some measure of the apparent barometric factor, that is the relationship between sea level response to atmospheric pressure (in phase) changes, can be derived by computing the in-phase barometric factor b(k) where k is a non-dimensional frequency. It can be viewed as the transfer function multiplied by the cosine of the phase function as a function of frequency. The isostatic response or hypothesis corresponds to b (k) = -1.01 cm/mb.

Fig. 21 shows the in-phase barometric factor for Grindstone Island for winter and summer seasons. The main features of Fig. 21 are as follows: (1) b(k) is not independent of frequency in either season; (2) summer and winter relationships are nearly opposite. That is, during summer periods the in-phase barometer factor is close to the isostatic value at low frequencies but gradually decreases at a constant average slope with increase in frequency. The rapid rise at high frequencies is probably insignificant in view of the low coherence indicated in Fig. 23. The winter b(k) values are on the average below the isostatic value. The peak of -1.40 cm/mb at 0.65 cpd is, in fact, close to the rise in the winter (W-E) wind autospectrum.

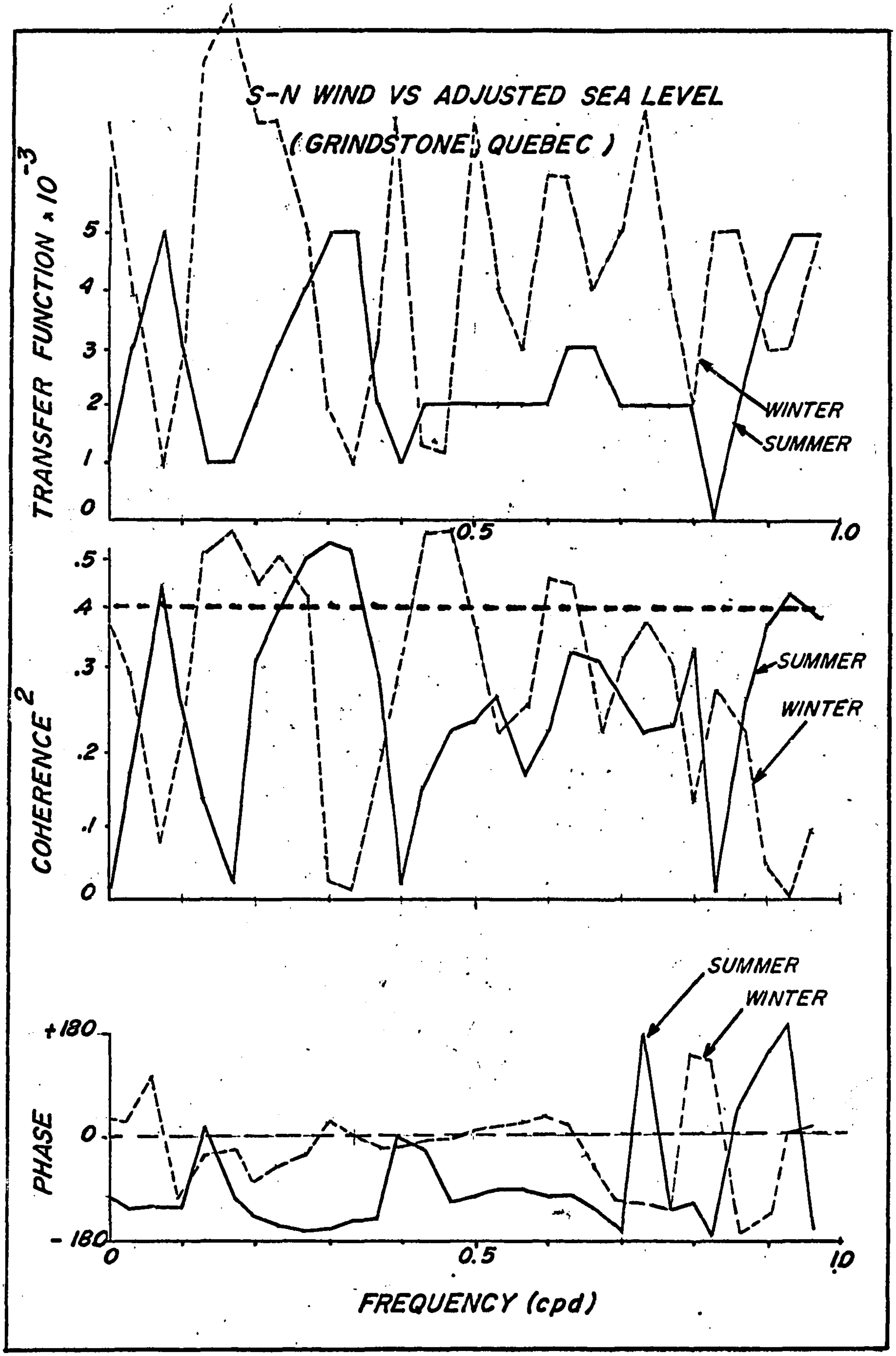

Fig. 22 Relation between adjusted Sea Level and (S-N) Winds

3.4.3. Frequency Response of Adjusted Sea Level to Winds

As stated previously, the non-isostatic values of the barometer factor could have been due to the omission of one or more variables which were also correlated with atmospheric pressure or, due to a resonant response of sea level to atmospheric pressure. It is difficult to conceive how this second effect might be generated in the Gulf. The wind effect (via its transfer function) for (S-N) winds is depicted in Fig. 22. The only significant peaks in the coherence spectrum for the summer period occurs at 0.1 and 0.3 cpd. The transfer function shows a rise at corresponding frequencies. In conjunction with the phase diagram it shows that an increase in (S-N) component winds bring about a decrease in sea level with the change in sea level lagging changes in the wind by approximately 120°.

The coherence spectrum for the winter season is also low. Some weak peaks occur at 0.2 and 0.5 cpd. Above 0.5 cpd the coherence spectrum seems to fluctuate in a random manner. Fig. 23 showing the (W-E) component wind effect can be contrasted to the (S-N) wind effect not only in magnitude but also in its frequency behaviour. The most interesting feature in Fig. 23 is the stability of the estimates. Significant coherence is exhibited in the frequency range 0.05 to 0.7 cpd. Above the latter frequency, the coherence falls below significance and

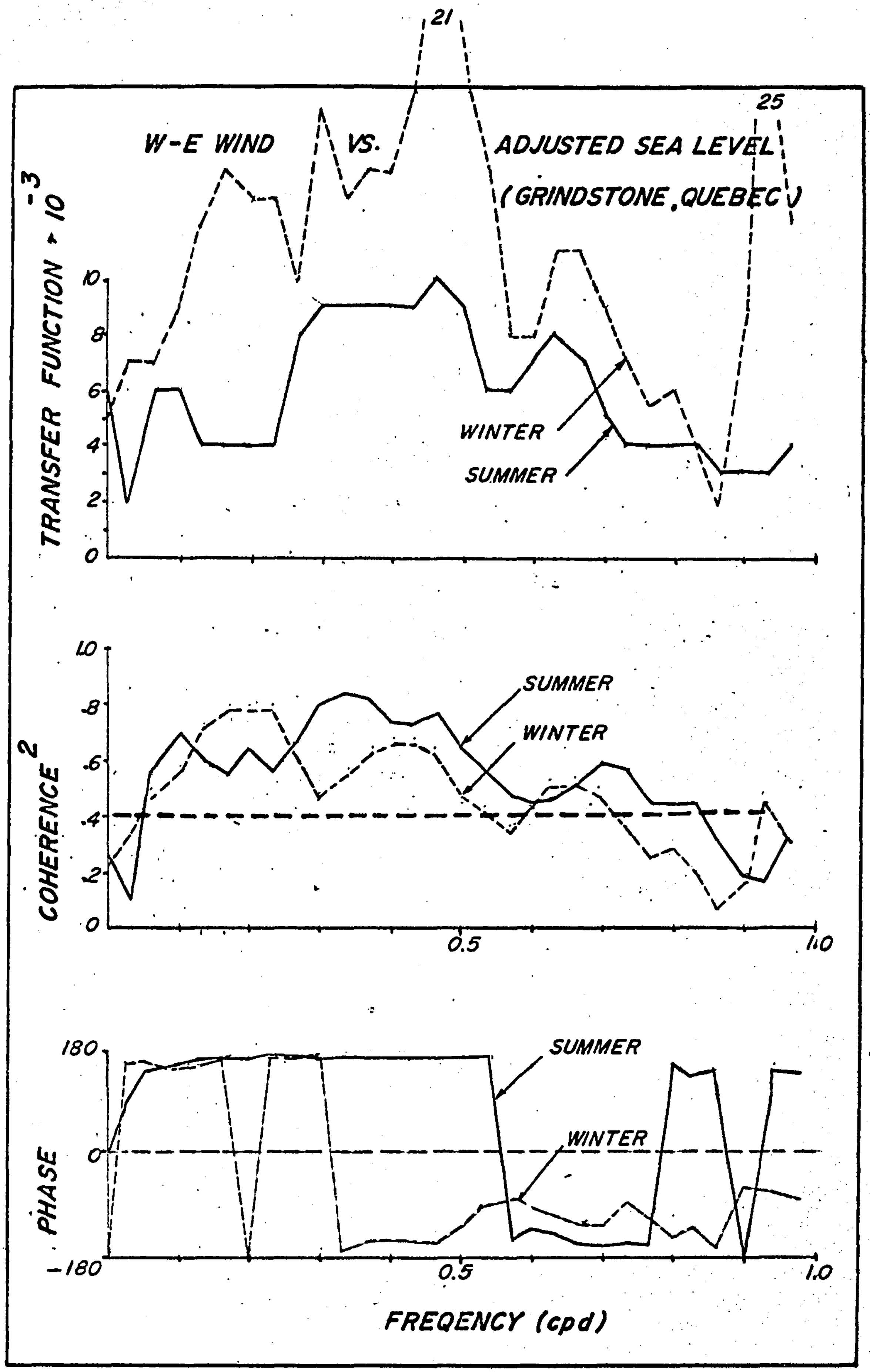

Fig. 23 Relation between adjusted Sea Level and (W-E) winds

the phase relationships are, in all probability, not very reliable. Between 0.1 and 0.55 cpd, summer sea levels lead summer (W-E) winds by about 170°. Above 0.55 cpd a rapid phase shift occurs.

Although the form of the coherence spectrum during summer and winter seasons are similar, the high transfer function would indicate that the actual relationship between summer and winter response of adjusted sea level to (W-E) component winds is different.

3.4.4 A Simple Regression Model

The results of the last two sections, that is the response and phase relationships between sea level , atmospheric pressure and component winds would indicate that the following model could be used to describe the process.

The model we assume is of the form :

$$Y(t) = \alpha + \sum_{i=1}^{N} \beta_i X_i(t) + e(t) \tag{1}$$

where

$Y(t)$ = daily mean sea level

$X_1(t)$ = daily mean atmospheric pressure

$X_2(t)$ = daily (S-N) component wind

$X_3(t)$ = daily (W-E) component wind

Although we can estimate α and β_i from the auto and cross spectrums of the different variables (Hamon and Hannan, 1963), the constants were evaluated by a multiple least squares regression procedure. The results are presented in Table 11 below.

Table 11 - Correlation and Regression Constants for Daily Sea Levels, Atmospheric Pressure and Winds

	SUMMER	WINTER
α	994.4	775.8
β_1	-0.91	-0.69
β_2	-0.002	Not Significant
β_3	-0.005	-0.07

It can be seen that the above results for the summer period reinforce the discussion of the previous sections. The results for the winter seasons are inconclusive. The barometer factor is significantly less in the winter- this is also borne out in Fig.21. One possible improvement in the above procedure could have resulted by weighting the coefficients by the signal to noise ratio in the respective frequency bands.

3.5 Discussion

From the evidence presented in section 3.4, it is apparent that although no mutual peaks occur in the local pressure and sea level spectrums, atmospheric pressure and sea level are strongly correlated over the frequency band of 0 to 0.55 cpd. According to theory (Groves, 1954; Unoki, 1950) , one would expect sea level and pressure to be linearly related by the 1:1 in phase isostatic hypothesis for fluctuations having periods greater than one day. The data in Figs. 19 to 21 do not quite support this hypothesis over the frequency band in question; both the transfer function and the in-phase barometer factor are below the isostatic value.

Spectrum analysis of Australian sea level records by Hamon (1962, 1966) has indicated that the barometer factor is appreciably less than the isostatic value on the east coast of continents and greater on the west coast. Subsequent analysis on the west and east coast of the United States (Mooers and Smith, 1968; Mysak and Hamon, 1968) have, in general, verified this relationship with differences usually to be explained by the local factors acting at the station.

The non-isostatic response of sea level to atmospheric pressure variations usually occurred in the frequency bands which exhibited mutual spectrum peaks and were explained, by the above mentioned references, to be due to travelling continental shelf waves as described by Robinson (1964) and Mysak (1967). In the case of Grindstone Island the importance of shelf waves is not very evident. This is not so much due to the fact that they could not occur. Rather, the lenght of records used in the analysis plus their digitizing interval does not provide us with high enough confidence intervals. The variable most often thought of resulting in non isostatic response of sea level to atmospheric pressure is the wind.

Some weight can be given to such an explaination after viewing Figs. 22 and 23. Over the band from o.05 to 0.55 cpd the coherence between (W-E) winds and adjusted sea levels is significant. From our simple regression model of section 3.4.4 it can also be seen that a pressure rise will usually lead to weaker (W-E) winds. Fig. 23 shows that rising (W-E) winds will in general reduce sea levels. Hence, pressure and winds act contrary. Whether this effect is really significant over the whole frequency band is difficicult to state as the wind records suffer from appreciable sampling errors.

The above situation has dealt with the effect of the local weather only. The apparent barometer factor, frequency response of sea levels to winds can also be changed due to the relatively enclosed nature of the Gulf. Assuming that typical weather systems are of horizontal dimensions smaller or comparable to the Gulf, it is conceivable that significant pressure gradients can exists between Gulf and Atlantic Ocean. This will lead to significant mass transports through Cabot Strait and , to a lesser extent, through the Straits of Belle Isle. Because of these constrictions to free flow, the compensating flow will not take place immediately and will introduce a reduction in the response characteristics of the system to atmospheric imputs.

Significant changes in the system relations occur when summer and winter seasons are compared. This is particularly evident in the in-phase barometric factor b(k) in Fig. 21 and, in the (W-E) wind- adjusted sea level relations in Fig. 23. Some of this change in the system can be explained by the general shift of the fluctuations in pressure, winds and sea level towards higher frequencies in winter.

CHAPTER IV

SUMMARY AND CONCLUSIONS

4.1 Summary and Conclusions

In summary, several techniques for and results from the analysis of sea level variations on the time scales of 1 year to 1 day have been illustrated.

On the annual scale, the sea level variations at individual gauge locations show some similarity - giving high sea levels in the early winter and low levels in the summer. Those differences that do occur can be traced to the greater effects of coastal currents or bottom topography. The direct effect of atmospheric pressure can usually explain about 30 % of the variance in the sea level records with the wind accounting for an additional 15 %. On shore winds are usually more effective in raising sea level along the open coast while the alongshore component of wind is usually more effective where some channeling due to bottom topography can occur.

The annual and semi annual waves of sea level and atmospheric pressure can usually explain about 80% of the variance in the records. Furthermore, the annual wave is the dominant.

The contributions of the thermohaline variations to sea level fluctuations are, in general inconclusive. Variations in temperature and salinity would tend to favour low sea levels in late winter and high levels in summer or early autumn.

Daily variations of sea level at one permanent tide gauge in the Gulf of St. Lawrence exhibited a close relationship to changes in atmospheric pressure in a frequency band from 0 to 0.55 cpd. The general shape of sea level and atmospheric variables spectrums show a rapid rise with diminishing frequency and, a general shift towards higher frequencies from the summer season to winter.

The response of sea level to atmospheric pressure is less that the pure hydrostatic over frequencies from 0 to 0.65 cpd. The response is also altered from summer to winter season; being less during the winter probably due to the general shift of the spectrums to higher frequencies and a greater wind effect. The removal of the direct linear effect of atmospheric pressure has not revealed any new features in the spectrum.

Wind effects on sea level were difficult to asses due to the fact that the wind records used suffered from considerable sampling errors.

4.2 Suggestions for Further Research

The spectral study has led to viewing the sea level variations at Grindstone to be caused by a dynamical system assumed to be linear to a first approximation.

Of interest then is the following question: Can sea level, atmospheric pressure and winds be utilized to predict sea level and finally currents in the Gulf of St. Lawrence ? This study has shown that the systems approach to the problem is valuable. However, in order to answer the question fully, at least two tide gauges and meteorological information is needed for only then can we separate the different types of motions possible.

Some of the inconclusive results of this study were due to the short period of analysis and inaccurate sampling procedures. In order to obtain better confidence to our results, longer and more frequent data from several stations are needed. With the greater record lenghts more sophisticated numerical filters can be designed leading to improved confidence intervals.

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APPENDIX A

Multiple Regression and Correlation Results

Table 12

Multiple Regression - Harrington, Québec

I	J	CORR.	SD(I)	SD(J)
1	2	-0.25	3.0	4.6
1	6	-0.69	3.0	3.7
3	4	-0.30	1.7	350.9
3	5	-0.29	1.7	335.4
5	6	-0.29	335.4	3.7

Standard Error of DEP variable = 2.59

Variable	COEFF.	Standard Error
X - 1	-0.82	0.09
X - 4	-0.001	0.0008
X - 5	-0.002	0.0008

Variables are deviations from normal mean year

X_1 = pressure (mb)

X_2 = precipitation (cm)

X_3 = temperature (C^o)

X_4 = (onshore component of wind)^2 (Km/hr)^2

X_5 = (longshore component of wind)^2 (Km/hr)^2

X_6 = sea level (cm)

Table 13

Multiple Regression - Charlottetown, P.E.I.

I	J	CORR.	SD(I)	SD(J)
1	4	-0.37	2.6	44.3
1	6	-0.50	2.6	4.0
3	4	-0.34	1.6	44.3
4	6	-0.33	44.3	4.0

Standard Error of DEP variable = 3.48

Variable	COEFF.	Standard Error
X - 1	-0.69	0.16
X - 4	0.01	0.009
X - 5	0.006	0.006

Variables are deviations from normal mean year

X_1 = pressure (mb)

X_2 = precipitation (cm)

X_3 = temperature (C^o)

X_4 = $(\text{onshore component of wind})^2$ $(Km/hr)^2$

X_5 = $(\text{longshore component of wind})^2$ $(Km/hr)^2$

X_6 = sea level (cm)

Table 14

Multiple Regression - Port aux Basques, Newfoundland

I	J	CORR.	SD(I)	SD(J)
1	6	-0.38	2.9	4.0
5	6	-0.47	32.6	4.0

Standard Error of DEP variable = 3.23

Variable	COEFF.	Standard Error
X - 1	-0.51	0.12
X - 2	-0.09	0.08
X - 4	-0.02	0.008
X - 5	0.06	0.01

Variables are deviations from normal mean year

X_1 = pressure (mb)

X_2 = precipitation (cm)

X_3 = temperature (C^o)

X_4 = (onshore component of wind)^2 (Km/hr)^2

X_5 = (longshore component of wind)^2 (Km/hr)^2

X_6 = sea level (cm)

Table 15

Multiple Regression - Halifax, N.S.

I	J	CORR.	SD(I)	SD(J)
1	4	-0.30	2.6	21.8
1	6	-0.64	2.6	3.9
4	5	0.28	21.8	29.8
4	6	-0.24	21.8	3.9
5	6	-0.40	29.8	3.9

Standard Error of DEP variable = 2.29

Variable	COEFF.	Standard Error
X - 1	-1.14	0.10
X - 4	-0.07	0.01
X - 5	-0.03	0.008

Variables are deviations from normal mean year

X_1 = pressure (mb)

X_2 = precipitation (cm)

X_3 = temperature (C^0)

X_4 = (onshore component of wind)^2 (Km/hr)^2

X_5 = (longshore component of wind)^2 (Km/hr)^2

X_6 = sea level (cm)

Table 16

Multiple Regression - St. John's, Nfld.

I	J	CORR.	SD(I)	SD(J)
1	5	0.27	3.5	59.6
1	6	-0.46	3.5	8.0
2	4	-0.30	4.2	81.3
4	5	0.33	81.3	59.6

Standard Error of DEP variable = 7.25

Variable	COEFF.	Standard Error
X - 1	-1.04	0.22

Variables are deviations from normal mean year

X_1 = pressure (mb)

X_2 = precipitation (cm)

X_3 = temperature (C^o)

X_4 = (onshore component of wind)^2 (Km/hr)^2

X_5 = (longshore component of wind)^2 (Km/hr)^2

X_6 = sea level (cm)

Table 17

Multiple Regression - Harrington, Quebec

I	J	CORR.	SD(I)	SD(J)
1	2	-0.34	3.6	2.0
1	3	0.33	3.6	13.8
1	6	-0.49	3.6	5.6
2	3	-0.3	2.0	13.8
3	5	-0.28	13.8	9.8
4	5	0.28	9.2	9.8
5	6	0.28	9.8	5.6

Standard Error of DEP variable = 4.57

Variable	COEFF.	Standard Error
X - 1	-0.88	0.15
X - 3	0.10	0.04
X - 5	0.19	0.05

X_1 = monthly pressure (mb)

X_2 = monthly precipitation (cm)

X_3 = monthly temperature (C^0)

X_4 = monthly onshore component of wind - Km/hr

X_5 = monthly longshore component of wind - Km/hr

X_6 = monthly sea level (cm)

Table 18

Multiple Regression - Charlottetown, P.E.I.

I	J	CORR.	SD(I)	SD(J)
1	3	0.33	3.2	16.3
1	4	-0.41	3.2	3.1
1	6	-0.45	3.2	5.4
2	3	-0.30	1.5	16.4
3	4	-0.57	16.4	3.1
3	5	-0.50	16.4	3.8
3	6	-0.43	16.4	5.4
4	5	0.46	3.1	3.8
4	6	0.40	3.1	5.4

Standard Error of DEP variable = 4.54

Variable	COEFF.	Standard Error
X - 1	-0.55	0.17
X - 2	0.59	0.34
X - 3	-0.06	0.04
X - 4	0.23	0.20

X_1 = monthly pressure (mb)

X_2 = monthly precipitation (cm)

X_3 = monthly temperature (C^o)

X_4 = monthly onshore component of wind - Km/hr

X_5 = monthly longshore component of wind - Km/hr

X_6 = monthly sea level (cm)

Table 19

Multiple Regression - Port aux Basques, Nfld.

I	J	CORR.	SD(I)	SD(J)
1	3	0.49	3.9	12.8
1	4	-0.50	3.9	2.8
1	5	-0.26	3.9	2.4
1	6	-0.48	3.9	4.9
3	4	-0.67	12.8	2.9
3	6	-0.27	12.8	4.9
4	5	0.30	2.8	2.4
5	6	0.41	2.4	4.9

Standard Error of DEP variable = 4.14

Variable	COEFF.	Standard Error
X - 1	-0.58	0.14
X - 4	-0.22	0.19
X 5	0.68	0.20

X_1 = monthly pressure (mb)

X_2 = monthly precipitation (cm)

X_3 = monthly temperature (C^o)

X_4 = monthly onshore component of wind - Km/hr

X_5 = monthly longshore component of wind - Km/hr

X_6 = monthly sea level (cm)

Table 20

Multiple Regression - Halifax, N.S.

I	J	CORR.	SD(I)	SD(J)
1	3	0.04	3.2	14.9
1	4	-0.36	3.2	1.4
1	5	0.39	3.2	3.0
1	6	-0.6	3.2	5.6
2	3	-0.4	2.1	14.9
2	6	0.3	2.0	5.6
3	4	-0.3	14.9	1.4
3	5	0.7	14.9	3.0
3	6	-0.6	14.9	5.6
5	6	-0.6	3.1	5.6

Standard Error of DEP variable = 3.87

Variable	COEFF.	Standard Error
X - 1	-0.55	0.15
X - 2	0.53	0.23
X - 3	-0.06	0.04
X - 5	-0.62	0.19

X_1 = monthly pressure (mb)

X_2 = monthly precipitation (cm)

X_3 = monthly temperature (C^0)

X_4 = monthly onshore component of wind - Km/hr

X_5 = monthly longshore component of wind - Km/hr

X_6 = monthly sea level (cm)

Table 21

Multiple Regression - St. John's, Nfld.

I	J	CORR.	SD(I)	SD(J)
1	3	0.64	5.0	12.6
1	4	0.34	5.0	4.0
1	5	0.40	5.0	3.8
1	6	-0.59	5.0	9.9
2	3	-0.36	2.0	12.6
2	4	-0.30	2.0	4.0
3	4	0.40	12.6	40.4
3	5	0.32	12.6	37.6
3	6	-0.50	12.6	9.9
4	5	0.60	4.0	3.8

Standard Error of DEP variable = 7.97

Variable	COEFF.	Standard Error
X - 1	-0.91	0.23
X - 3	-0.16	0.09

X_1 = monthly pressure (mb)

X_2 = monthly precipitation (cm)

X_3 = monthly temperature (C^0)

X_4 = monthly onshore component of wind - Km/hr

X_5 = monthly longshore component of wind -Km/hr

X_6 = monthly sea level (cm)

Table 27

Multiple Regression - Harrington, Québec

Summer

I	J	CORR.	SD(I)	SD(J)
1	6	-0.59	2.7	5.2
3	5	0.37	7.3	7.7
4	6	0.42	7.7	5.2

Standard Error of DEP variable = 3.69

Variable	COEFF.	Standard Error
X - 1	-1.13	0.19
X - 4	0.29	0.07

Winter

I	J	CORR.	SD(I)	SD(J)
1	2	-0.37	3.9	2.4
1	6	-0.59	3.9	5.9
3	4	-0.33	7.6	9.4
3	5	-0.51	7.6	12.2
4	5	0.55	9.4	12.2

Standard Error of DEP variable = 4.99

Variable	COEFF.	Standard Error
X - 1	-0.89	0.21

X_1 = monthly pressure (mb)

X_4 = monthly onshore component of wind Km/hr

Table 28

Multiple Regression - Charlottetown, P.E.I.

Summer

I	J	CORR.	SD(I)	SD(J)
1	6	-0.36	2.3	4.5
2	6	0.30	1.6	4.5
3	4	-0.36	9.0	2.2
3	5	-0.35	9.0	2.6
3	6	-0.33	9.0	4.5

Standard Error of DEP variable = 3.96

Variable	COEFF.	Standard Error
X - 1	-0.82	0.25
X - 3	-0.20	0.06

Winter

I	J	CORR.	SD(I)	SD(J)
1	4	-0.40	3.6	3.3
1	6	-0.39	3.6	5.7
3	4	-0.43	6.8	3.3
4	5	0.31	3.3	3.9
4	6	0.33	3.3	5.7

Standard Error of DEP variable = 5.45

Variable	COEFF.	Standard Error
X - 1	-0.62	0.26

X_1 = pressure (mb) - (monthly)

X_3 = temperature (C^0) - (monthly)

Table 29

Multiple Regression - Port aux Basques, Newfoundland

Summer

I	J	CORR.	SD(I)	SD(J)
1	6	-0.52	2.3	4.3
3	4	-0.46	7.8	2.3
4	5	0.36	2.3	1.0

Standard Error of DEP variable = 3.75

Variable	COEFF.	Standard Error
X - 1	-0.95	0.23

Winter

I	J	CORR.	SD(I)	SD(J)
1	4	-0.50	4.1	2.3
1	6	-0.35	4.1	5.3
3	4	-0.36	5.5	2.3
5	6	0.60	3.3	5.3

Standard Error of DEP variable = 4.19

Variable	COEFF.	Standard Error
X - 1	-0.32	0.18
X - 5	0.88	0.22

X_1 = monthly pressure (mb)

X_5 = monthly longshore component of wind - Km/hr

Table 30

Multiple Regression - Halifax, N.S.

Summer

I	J	CORR.	SD(I)	SD(J)
1	6	-0.24	2.1	4.9
2	6	0.43	1.9	4.9
3	6	-0.66	8.5	4.9
5	6	-0.57	1.9	4.9

Standard Error of DEP variable = 2.94

Variable	COEFF.	Standard Error
X - 1	-0.56	0.23
X - 3	-0.17	0.07
X - 5	-1.29	0.30

Winter

I	J	CORR.	SD(I)	SD(J)
1	6	-0.60	3.5	5.3
2	6	0.03	2.1	5.3
3	6	-0.21	6.3	5.3
5	6	-0.38	2.9	5.3

Standard Error of DEP variable = 4.31

Variable	COEFF.	Standard Error
X - 1	-0.90	0.21

X_1 = monthly pressure (mb)

X_3 = monthly temperature (C^0)

X_5 = monthly longshore component of wind - Km/hr

Table 31

Multiple Regression - St. John's, N.B.

Summer

I	J	CORR.	SD(I)	SD(J)
1	6	-0.28	2.8	8.9
4	5	0.38	2.6	0.2

Standard Error of DEP variable = 8.56

Variable	COEFF.	Standard Error
X - 1	-0.67	0.46
X - 3	-0.26	0.16

Winter

I	J	CORR.	SD(I)	SD(J)
1	6	-0.63	4.9	9.1
3	6	-0.35	4.1	9.1
4	5	0.61	5.0	4.9

Standard Error of DEP variable = 7.26

Variable	COEFF.	Standard Error
X - 1	-1.18	0.25

X_1 = monthly pressure (mb)

X_3 = monthly temperature (C^0)

APPENDIX B

Harmonic Constants for Meteorological Variables

Table 22 - Harmonic Constants for Pressure, Temperature, Precipitation and Wind Speed at Harrington, Quebec

n	Pressure		Temperature		Precipitation		Wind Speed	
	C_n	$\emptyset_n$	C_n	$\emptyset_n$	C_n	$\emptyset_n$	C_n	$\emptyset_n$
1	1.62	251.40	10.72	247.55	2.38	88.70	5.22	80.73
2	1.22	249.90	1.60	281.23	0.80	80.23	0.74	99.79
3	0.23	147.22	0.53	285.40	0.73	255.33	1.22	179.22
4	1.18	309.76	0.54	95.89	0.14	304.13	2.25	139.65
5	0.73	335.59	0.73	155.90	1.20	165.46	1.69	60.48
6	0.63	89.76	0.30	270.00	0.20	89.99	1.70	90.00
σ^2	3.46		55.80		4.20			
Mean $\langle X_i \rangle_{12}$	1010.35		2.19		9.72		19.18	

P_n = atmospheric pressure in mb

T_n = temperature in °C

W_n = wind speed in m/sec

$\emptyset_n$ = phase angle in degrees

σ^2 = variance of the 0^{th} harmonic

PR = precipitation in cm

Table 23 - Harmonic Constants for Pressure, Temperature, Precipitation and Wind Speed at Charlottetown, Prince Edward Island

	Pressure		Temperature		Precipitation		Wind Speed	
n	C_n	$\emptyset_n$	C_n	$\emptyset_n$	C_n	$\emptyset_n$	C_n	$\emptyset_n$
1	1.70	222.19	13.19	247.10	2.30	118.09	1.38	79.73
2	0.99	245.29	1.75	256.34	0.32	112.77	0.46	68.19
3	0.64	122.13	1.65	346.68	0.15	342.99	0.55	243.36
4	1.00	290.44	1.55	83.47	0.36	222.05	0.87	143.88
5	0.72	316.24	1.82	170.80	0.33	348.78	1.30	81.95
6	0.35	89.56	0.59	270.00	0.10	270.02	0.03	89.93
σ^2	3.03		19.10		2.84			
Mean $\langle X_i \rangle$ 12	1012.76		6.05		9.00		8.12	

P_n = atmospheric pressure in mb T_n = temperature in°C   PR = precipitation in cm

W_n = wind speed in m/sec $\emptyset_n$ = phase angle in degrees

σ^2 = variance of the 0^{th} Harmonic

Table 24 - Harmonic Constants for Pressure, Temperature, Precipitation and Wind Speed at Port aux Basques, Newfoundland

	Pressure		Temperature		Precipitation		Wind Speed	
n	C_n	ϕ_n	C_n	ϕ_n	C_n	ϕ_n	C_n	ϕ_n
1	2.83	237.67	9.17	247.62	1.87	147.18	1.30	54.98
2	1.14	233.21	0.97	79.96	0.86	112.56	1.69	62.15
3	0.44	118.04	0.62	160.24	1.26	4.47	0.63	165.82
4	1.14	299.34	0.80	85.77	0.81	46.12	0.26	49.47
5	0.81	303.22	1.00	90.68	0.81	268.63	0.96	129.79
6	0.39	89.61	0.44	90.00	0.40	270.01	0.05	89.97
σ^2	5.86		48.44		3.76			
Mean $\langle x_i \rangle_{12}$	1011.56		5.09		11.17		5.15	

P_n = atmospheric pressure in mb

T_n = temperature in °C

PR = precipitation in cm

W_n = wind speed in m/sec

ϕ_n = phase angle in degrees

σ^2 = variance of the 0^{th} Harmonic

Table 25 - Harmonic Constants for Pressure, Temperature, Precipitation and Wind Speed at Halifax, Nova Scotia

	Pressure		Temperature		Precipitation		Wind Speed	
n	C_n	ϕ_n	C_n	ϕ_n	C_n	ϕ_n	C_n	ϕ_n
1	2.12	224.29	11.57	254.98	3.59	110.99	0.75	14.65
2	0.76	232.74	0.13	202.59	0.64	114.82	1.06	55.98
3	0.65	105.52	0.26	267.10	0.65	201.28	0.60	268.26
4	1.03	288.64	0.17	355.95	0.62	352.71	0.17	148.56
5	0.57	310.85	0.37	92.22	0.32	281.19	0.89	131.64
6	0.36	89.57	0.19	90.00	0.48	270.00	0.03	270.05
σ^2	3.57		67.08		7.34			
Mean $\langle X_i \rangle_{12}$	1013.33		5.91		11.51		6.63	

P_n = atmospheric pressure in mb

T_n = temperature in °C

PR = precipitation in cm

W_n = wind speed in m/sec

ϕ_n = phase angle in degrees

σ^2 = variance of the 0^{th} Harmonic

Table 26 - Harmonic Constants for Pressure, Temperature, Precipitation and Wind Speed at St. John's, Newfoundland

	Pressure		Temperature		Precipitation		Wind Speed	
n	C_n	ϕ_n	C_n	ϕ_n	C_n	ϕ_n	C_n	ϕ_n
1	4.47	240.98	9.64	246.99	4.07	115.74	0.29	103.57
2	1.22	193.31	0.67	62.28	0.51	225.74	0.99	78.96
3	0.34	112.86	0.32	238.28	0.72	232.05	0.25	269.23
4	1.18	311.72	0.05	115.88	0.11	143.29	1.36	105.98
5	1.00	284.63	0.19	314.37	1.25	152.58	0.42	356.48
6	0.38	89.60	0.14	90.01	0.28	89.99	0.29	270.01
σ^2	12.11		46.79		9.55			
Mean $\langle X_i \rangle_{12}$	1011.14		4.75		11.08		10.99	

P_n = atmospheric pressure in mb

T_n = temperature in °C

W_n = wind speed in m/sec

ϕ_n = phase angle in degrees

σ^2 = variance of the 0^{th} harmonic

PR = precipitation in cm