AN ABSTRACT OF THE THESIS OF

<u>   Kuo-heng Lee   </u> for the <u>Master of Science </u>
      (Name of student)   (Degree)

in   <u>Oceanography   </u> presented on <u>April 25, 1967</u>
        (Major)   (Date)

Title: <u>GEOPOTENTIAL ANOMALY AND GEOSTROPHIC FLOW OFF</u>

<u>NEWPORT, OREGON   </u>

Redacted for Privacy

Abstract approved: <u>   </u>
                        Dr. June G. Pattullo

The geopotential anomalies and the average meridional geostrophic flow off Newport, Oregon, were computed from the data taken during twenty-one hydrographic cruises.

The annual average of geopotential anomaly was 1.31 dynamic meters with a seasonal variation of the order of 14 dynamic centimeters. Highest values of geopotential anomaly occurred in September; lowest values occurred in April. Both the highest and lowest value of geopotential anomaly occurred at 105 nautical miles offshore.

The currents found by dynamic computation were weak and irregular, generally 5 cm/sec or less. The direction of the average meridional geostrophic flow varied with season. Within 105 nautical miles of the coast, flow was southward in summer, northward in winter. Beyond 105 nautical miles from the coast, flow directions reversed, that is, flow was northward in summer and southward in winter.

Geopotential Anomaly and Geostrophic
Flow Off Newport, Oregon

by

Kuo-heng Lee

A THESIS
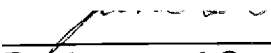
submitted to

Oregon State University

in partial fulfillment of
the requirements for the
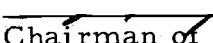degree of

Master of Science

June 1967

APPROVED:
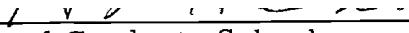
Professor of Oceanography
                in charge of major

Chairman of Department of Oceanography

Dean of Graduate School

Date thesis is presented April 25, 1967

Typed by Donna Olson for   Kuo-heng Lee

ACKNOWLEDGEMENT

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

GEOPOTENTIAL ANOMALY AND GEOSTROPHIC
FLOW OFF NEWPORT, OREGON

I. INTRODUCTION

Scope of Study

This paper is primarily a study of the geopotential anomaly and geostrophic flow off the Oregon Coast, covering the area from 65 to 165 nautical miles from Newport, Oregon ($44^\circ 39'$ N latitude, $125^\circ 35'$ to $128^\circ 55'$ W longitude). For the sake of simplicity, the Newport Hydrographic Coordinates will be used in this paper to indicate the location of study. For example, stations 65 and 85 nautical miles off Newport are referred to as NH 65 and NH 85 respectively.

The average geopotential anomaly relative to 1,000 decibars and its seasonal and spatial variations are plotted and discussed. The relationship between geopotential anomaly and oceanic heat content is examined.

The monthly average meridional component of the geostrophic flow and its seasonal variation at the depths of 0, 100, 200, 500, 800 meters are computed. In order to verify the dynamic computation, a comparison with the results obtained by direct drogue measurement is made. The reliability of geostrophic computation and the factors which affect this computation are intensively discussed. The errors involved in computing the geostrophic flow in this paper are also estimated.

Previous Studies of the Currents Off the Oregon Coast

According to Sverdrup, Johnson and Fleming (1942), the major current that effects the Oregon Coast is the California current which is a portion of the eastern boundary current system in the North Pacific Ocean.

Wooster and Reid (1960) have studied eastern boundary currents and have concluded that these currents are characterized as slow, broad, and shallow, with relatively small transport. One of the striking features off the Oregon Coast is the upwelling phenomenon. During the spring and early summer months, wind from the north to north-west prevails along the coast of Oregon. These winds give rise to coastal upwelling that begins in May or June and continues until early fall. Sverdrup et al. (1942) stated:

> During the entire season of upwelling, a counter-current that contains considerable quantities of equatorial water flows close to the coast at depths below 200 meters. In the fall the upwelling ceases and in the surface layers a current opposite to the direction of the California current develops, the Davidson Current which in December and January runs north along the coast to at least latitude 48° North.

These statements pertain specifically to the California Coast, and to average or prevailing conditions. Details of the circulation off Oregon, and the variability from year to year, or even month to month still need study.

Pattullo and Stevenson (1966) have intensively studied the currents off the Oregon Coast by direct measurement with drogues, and they concluded that subsurface water off the Oregon Coast is influenced by the California Current. The mean flow of surface and subsurface water is southward through the year. The pycnocline, present at 100-200 m, has no apparent effect on north-south flow. Above the pycnocline, the transport is eastward. East-west flow is evidently linked to the density structure. A northward water flow of surface water was noted from several measurements. This flow is probably associated with the Davidson Current.

Burt and Wyatt (1964) studied the surface current by drift bottles and concluded that the northward flowing Davidson Current was observed from October through March. Varying surface currents from both north and south were present during the spring (April and May) and fall (September) transition periods. During the summer (June, July and August), the surface current was toward the south. The Davidson Current varied in width to over 165 miles and was observed to flow as far north as 50° North latitude. The Davidson Current off Oregon appeared to be the direct result of local wind stress.

The upwelling along the Oregon Coast was studied by Smith, Pattullo and Lane (1966). The offshore transport during an early stage of coastal upwelling off the southern coast of Oregon was

computed to be 4.4×10^9 g/cm for the 76-hour period. The vertical

velocities computed from the change in depth of the $25.5 \quad \sigma_t$

surface, which is representative of the upper part of the pycnocine

off Oregon, decrease linearly from 7.0×10^{-3} cm/sec at the inner

station to 0.2×10^{-3} cm/sec at the outer station.

Maughan (1963) studied currents off the Oregon Coast at

$44^{\circ}39'$ N., $125^{\circ}20'$ W. He made comparisons of the data obtained by

drogue measurement and by drift bottles. In his discussion, he

said:

> A great variability appears in the speed of the
> average surface currents for any given month.
> For example, from observations of ship-drift
> during May, the average speed of the surface
> currents is shown to be about 19 cm/sec,
> whereas, from direct measurements with
> drogues, the average current was about 2 cm/sec.
> Thus, the absolute speed of any given surface
> current measurement depends not only on the
> month of observation, but also is highly de-
> pendent on other factors, probably the parti-
> cular weather system that is present during
> the period of observation.

In this paper the author has computed the average monthly

current speed from four years of hydrographic data in order to give

some general idea of the mean currents off the Oregon Coast.

II. HYDROGRAPHIC DATA

The hydrographic data used in this paper are listed in Table 1.
These data were taken from February, 1962 to December, 1965.
There were a total of twenty-one hydrographic cruises which in-
cluded four cruises each in February and July; three cruises in
December; and two cruises each in April, May, June, September
and November. In January, March, August and October no Oregon
State University data were deep enough for computation relative to
the selected reference level which is 1,000 meters. The hydrograph-
ic data used were taken from NH 65 to NH 165 with a spacing interval
of twenty nautical miles, namely, NH 65, NH 85, NH 105, NH 125,
NH 145 and NH 165. From the temperature and salinity observations,
the dynamic heights were computed; in turn the geopotential anomaly,
and geostrophic flow, were computed. The monthly average of dyna-
mic height is the arithmetic mean value for the same month of dif-
ferent years.

The temperature and salinity distributions are illustrated in
the appendix. Pattullo and Denner (1965) have studied the tempera-
ture and salinity distribution off the Oregon Coast. Using the data
from 1961 to 1962, at stations between 5 and 25 nautical miles off-
shore, they found that the temperature ranged from 6 to 17°C with a
mode between 10 and 11°C. Salinity varied from 18.0 to 33.5‰ with

Table I. Hydrographic cruises used for geostrophic computations off Newport, Oregon.

Month	Year 1962	1963	1964	1965	Monthly Total
January					
February	X	X	X	X	4
March					
April	X		X		2
May	X	X			2
June	X			X	2
July	X	X	X	X	4
August					
September	X	X			2
October					
November		X		X	2
December	X	X		X	3
Total					21

a mode between 32. 5 and 33.0‰.

During 1963 to 1965, between the coast and 165 nautical miles offshore, temperatures ranged from 18.5 to 3.4°C. Salinities varied from 33.60 to 28.50‰. The modes of temperature and salinity in the upper 200 m were very close to the values calculated by Pattullo and Denner.

Temperature decreases and salinity increases rapidly in the upper layer (about upper 200 meters). Below 200 meters all gradients are small. At 500 meters the temperature decreases to 5 to 5.5°C and the salinity increases to 34.1 to 34.2‰. At 1,000 meters, the temperature is about 3.5°C and salinity is about 34.4‰. The variation below 1,000 meters is practically negligible.

The spatial variations of temperature and salinity from 5 to 165 nautical miles offshore are primarily due to local processes-- upwelling, rainfall, and river run-off. From May to September water upwells from a depth of about 200 meters which makes shelf water cold and more saline near the coast. During February and March the river run-off reduces salinity near the coast.

III. GEOPOTENTIAL ANOMALY

Definition

In oceanography, a surface which is at right angle to the direction of gravity is called a level surface. This surface is also called a geopotential surface or gravitational equipotential surface.

The anomaly of the geopotential distance between two isobaric surfaces P_1 and P_2, or abbreviated as geopotential anomaly, is computed by the following equation (Sverdrup et al., 1942, p. 409)

$$\Delta D = \int_{P_1}^{P_2} \delta \, dp$$

Where:

ΔD = Geopotential anomaly in dyn meters between two isobaric surfaces P_1 and P_2 expressed in decibars

δ = Specific volume anomaly in m^3/kg

$\quad = \alpha_{S,T,P} - \alpha_{35,O,P}$

$\alpha_{S,T,P}$ = Specific volume at given salinity, temperature, and pressure

$\alpha_{35,O,P}$ = Specific volume at 35 parts per mille salinity, $0^\circ C$ temperature, and pressure P

Computational Results

The geopotential anomaly off Newport, Oregon from NH 65 to NH 165 is computed using the 1,000 decibars surface as reference level. The annual, seasonal and monthly average of geopotential anomaly as shown in Figure 1 and Figure 2.

The annual average geopotential anomaly was 1.31 dyn meters, for the entire section, the spatial variation was less than 1 dyn cm.

In summer, which was selected to include data during June, July and September, the geopotential anomaly ranged from 1.37 to 1.32 dyn meters. The maximum value was at NH 105 and the minimum at NH 65.

In winter, which includes data during December and February, the geopotential anomaly was almost uniform at 1.30 except at NH 65 where a maximum value of 1.32 dyn meters occurred.

In spring, including data during April and May, the geopotential anomaly varied from 1.29 to 1.26 dyn meters. Maximum values occurred at NH 65 and NH 165, minimum values at NH 105.

The seasonal variation varied from 1.39 to 1.25 dyn meters. As a whole, the highest geopotential anomaly occurred in summer, especially September, and the lowest anomaly in spring, usually April. In general, the summer anomaly is three dyn cm higher than the annual average, and eight dyn cm higher than the spring average.

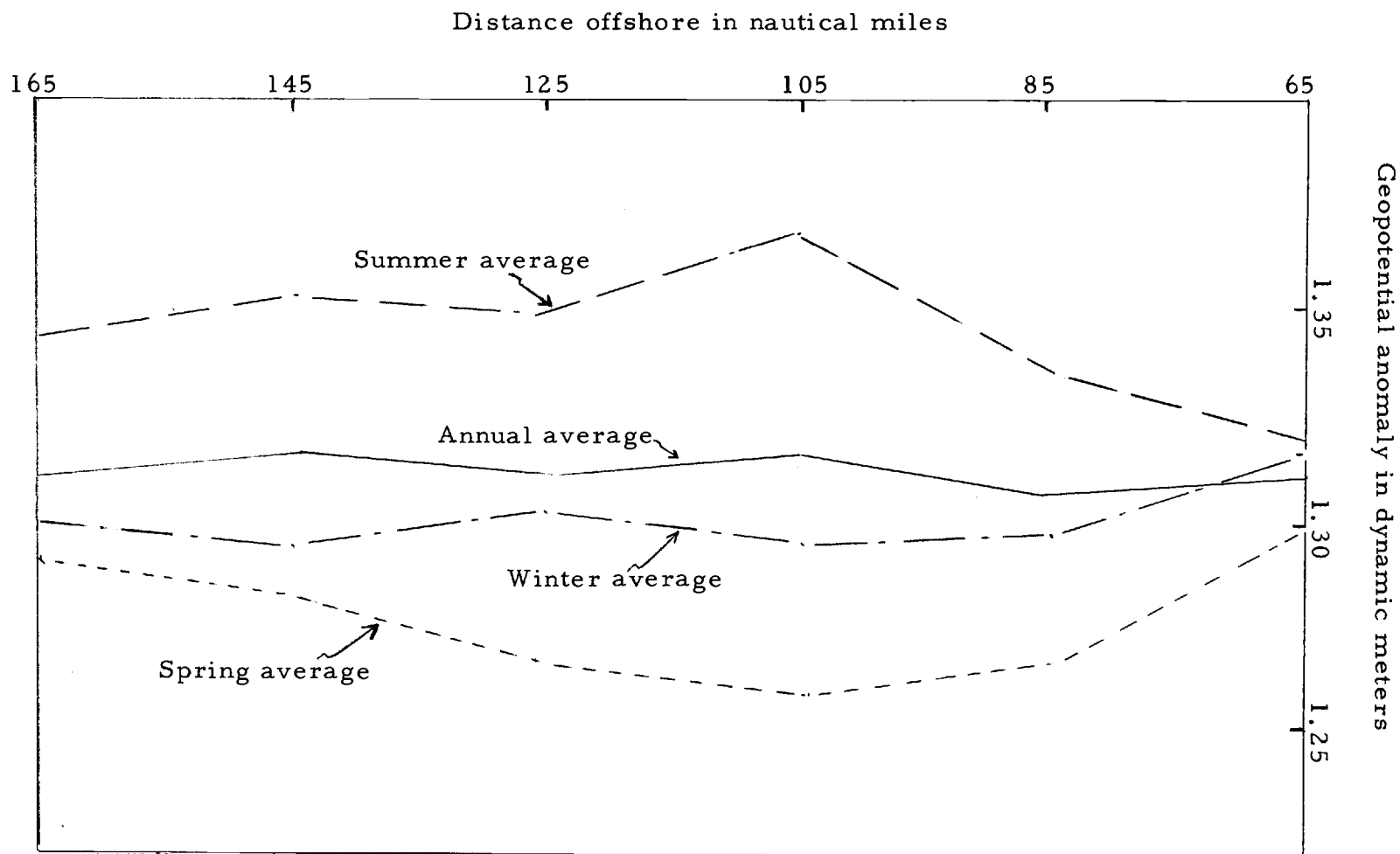Distance offshore in nautical miles

Figure 1. Annual and seasonal average of geopotential anomaly (surface over 1,000 decibars).
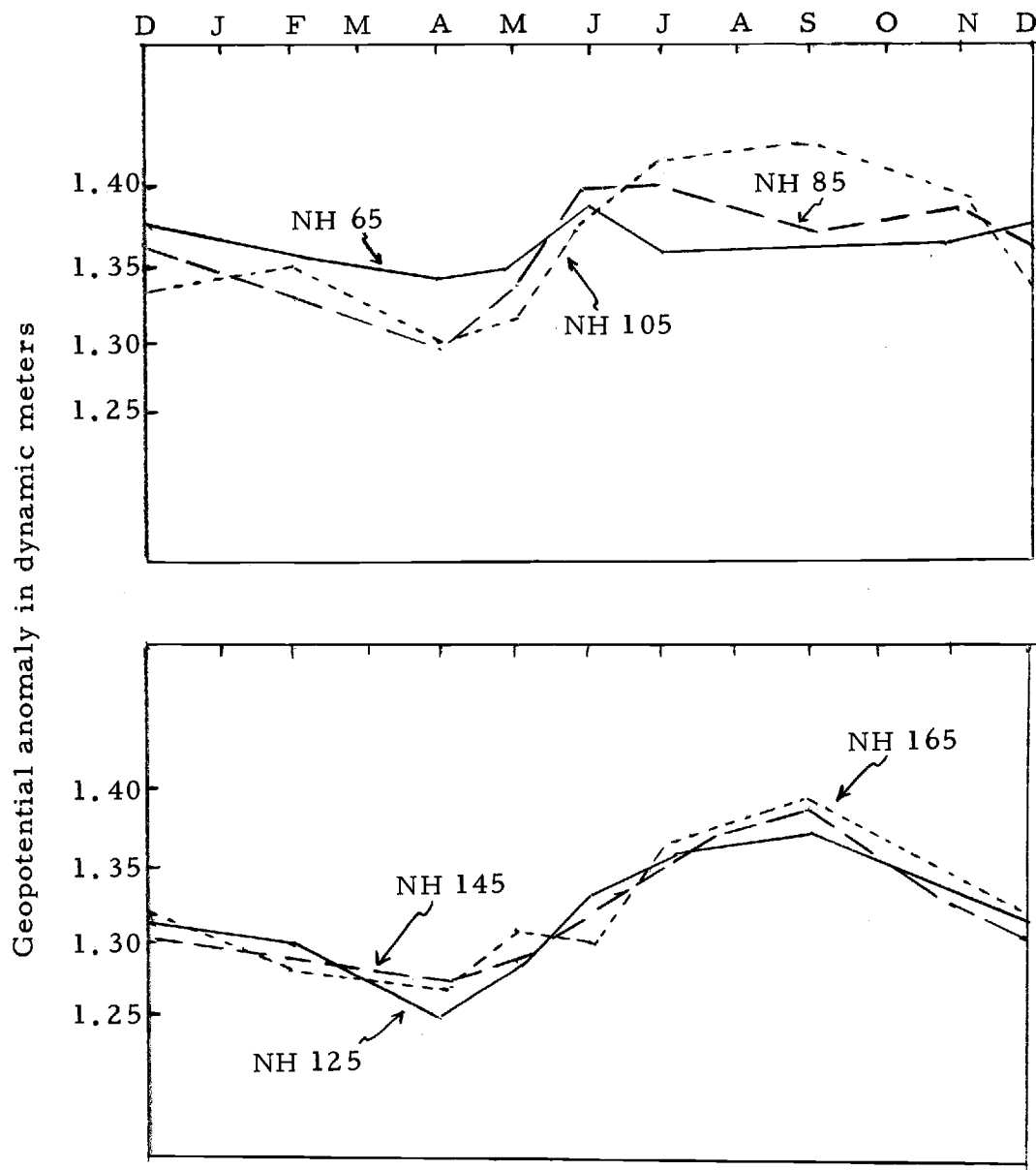
Figure 2. Seasonal variation of geopotential anomaly for each station (surface over 1,000-dbs).

The seasonal variation was a maximum at NH 105 with a magnitude in the order of 14 dyn cm.

The most striking feature discovered from the computations was that the lowest surface geopotential anomaly occurred in spring rather than in the winter. This is primarily related to the oceanic heat content. It seems that the upwelling occurring in late spring plays an important role in affecting the geopotential anomaly.

The geopotential anomaly with minimum values in April, increased as time went on, and reached its peak value in September, then decreased again until April. This annual oscillation appeared to be related to, as aforementioned, heat content and thermosteric effects which will be discussed in detail in a later section.

Thermohaline Effects on the Geopotential Anomaly

The thermohaline effects have been studied by Pattullo et al. (1955). The relationship they found between geopotential anomaly and thermohaline effects is expressed approximately by the formulas:

$$Z_T = g^{-1} \int_{P_a}^{P_o} (\partial \alpha / \partial T) \; \Delta T \; dp \qquad \text{and}$$

$$Z_S = g^{-1} \int_{P_a}^{P_o} (\partial \alpha / \partial S) \; \Delta S \; dp$$

where

Z_T = thermal departure from annual mean geopotential anomaly

Z_S = haline departure from annual mean geopotential anomaly

P_a = atmospheric pressure

P_o = the pressure to which the integration carried

For any depth let $\Delta T = T - \bar{T}$, and $\Delta S = S - \bar{S}$ designate the monthly departures in temperature and salinity from their annual means, $\bar{T}, \bar{S}$. For small values of ΔT and ΔS the corresponding departure in specific volume is given by

$\Delta \alpha = \alpha (T, S, P) - \alpha (\bar{T}, \bar{S}, P).$

$= (\partial \alpha / \partial T) \Delta T + (\partial \alpha / \partial S) \Delta S + \cdots$

Where $\partial \alpha / \partial T$ and $\partial \alpha / \partial S$ are to be evaluated at T, S, P. But the explicit dependence of $\partial \alpha / \partial T$, $\partial \alpha / \partial S$ on pressure can be neglected, and accordingly in all numerical work $\partial \alpha / \partial T$ and $\partial \alpha / \partial S$ may be read as functions of $\bar{T}(p)$, $\bar{S}(p)$, O.

Comparison of Geopotential Anomaly with Oceanic Heat Content

A comparison of geopotential anomaly and heat content has been made and shown in Figure 3 and Figure 4.

The oceanic heat content in the upper 100 m off Newport, Oregon has been studied by Burt, Pattullo and Kulm (in preparation) with the hydrographic data from 1962 to 1965 (the same data used

in this paper for computing the geopotential anomaly).

Figure 3 shows that the two results are very closely correlated. The seasonal variation of upper layer heat content and geopotential anomaly followed the same trend, and in both cases, the lowest values occurred in April and the highest values in September.

Geopotential anomaly versus surface heat content (NH 105 - 165) is shown in Figure 4. The straight line is obtained by regressional analysis. The simple positive correlation between heat content and geopotential anomaly has been calculated as 3.1×10^{-1} dyn cm/kg cal/cm^2. The result showed that the relationship between geopotential anomaly and surface heat content is indeed linear. It also suggests that the specific volume anomaly off the Oregon Coast is primarily influenced by the heat content of the upper layer.
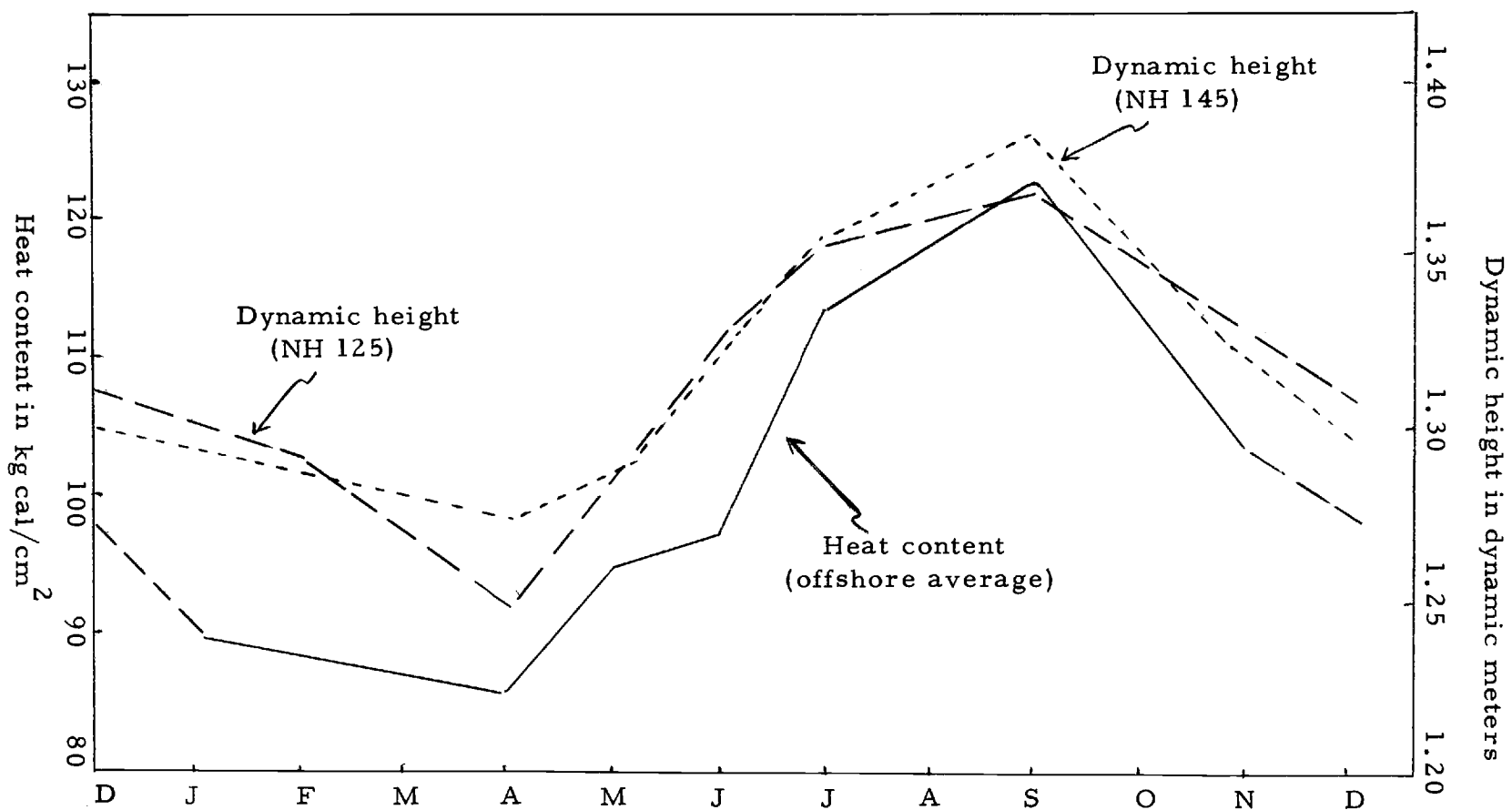
Figure 3. Geopotential anomaly (surface over 1,000-dbs) and heat content (offshore average).
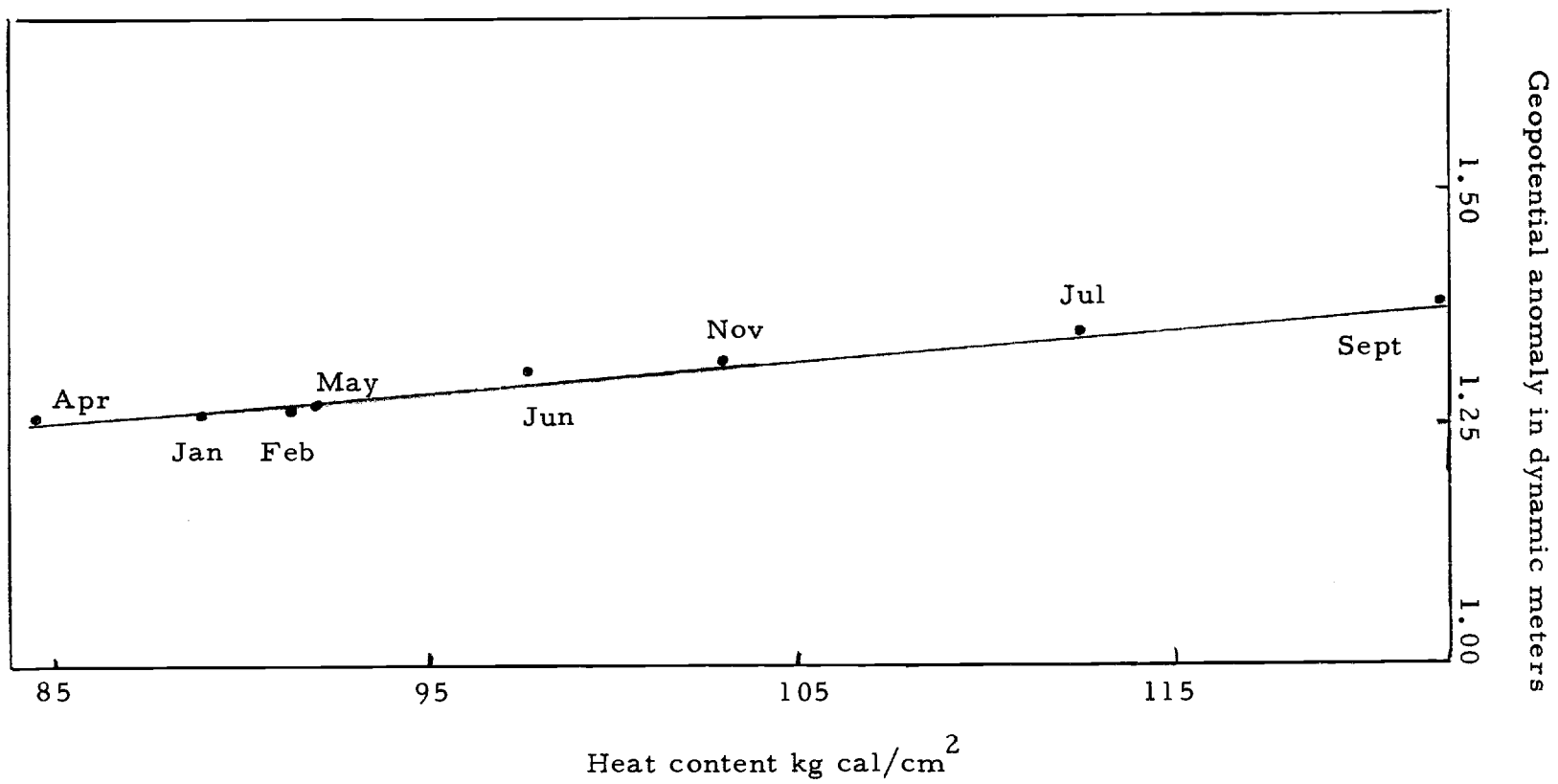
Figure 4. Relationship between geopotential anomaly and heat content (see Figure 3).

IV. GEOSTROPHIC FLOW

Method for Computing Geostrophic Flow

Permanent ocean currents can be computed from the observed distribution of density on the assumptions that: (1) the horizontal pressure gradient is balanced by the coriolis force (2) the horizontal velocity and the horizontal gradient vanish at a moderate depth below sea surface (3) the acceleration and frictional force are neglected. Those assumptions simplify the equation of motion into the so called geostrophic equation (Sverdrup, 1947)

$$\frac{1}{\rho} \frac{\partial p}{\partial x} = fu \tag{1}$$

$$\frac{1}{\rho} \frac{\partial p}{\partial y} = -fv \tag{2}$$

The relative current between two isobaric surfaces is given by the equation

$$V_1 - V_o = \frac{10 \, (\Delta D_A - \Delta D_B)}{f \cdot L} \tag{3}$$

Where

V_o = Current velocity at the reference level

V_1 = Current velocity at a given level

ΔD_A = Geopotential anomaly at hydro-station A

ΔD_B = Geopotential anomaly at hydro-station B

f = Coriolis parameter

L = Distance between the two hydrographic stations

The monthly average value of current speed is obtained by using the monthly average of ΔD, substituted into equation (3).

Reference Level

The 1,000 decibar surface was chosen as the reference level for the computations in this paper. This is partly due to the availability of hydrographic data, which were mostly taken down to 1,000 meters. However, using Defant's method, the 1,000 decibar surface seems close to the depth of no motion off the Oregon Coast.

Fomin (1964) studied various methods in determining the depth of no motion and concluded that of the existing methods for determining the depth of no motion, the best are that of A. Defant and H. Sverdrup's method. The Defant's method was summarized by Fomin as follows:

> From analysis of the differences in the dynamic heights of isobaric surfaces at a great number of pairs of neighboring stations, A. Defant (1941) discovered a layer of relatively great thickness in which these differences vary considerably along the vertical. The thickness of this layer in the Atlantic Ocean ranges from 300 to 800 meters and its depth varies rather uniformly in the horizontal direction, while the change in differences in the dynamic heights of isobaric

surfaces only amounts to several dynamic milli-
meters. The constancy of differences in dynamic
heights indicates that the gradient component of
current velocity is constant along the vertical in
this layer. Defant assumes that this water layer is
almost motionless and considers it to be the layer
directly adjoining the 'zero surface'.

Using Defant's method, the depth of no motion off Newport,

Oregon was found to be at 600 to 800 meters depth approximately.

Since the hydrographic data used to compute the depth of no motion

were mostly taken down only to 1,000 meters, there is no way to

detect the depth of no motion below 1,000 meters.

According to Wooster and Reid (1960) the principal equatorward

flow of the Pacific eastern boundary currents is above 1,000 meters

and in most places is above 500 meters. Furthermore, their trans-

port calculations relative to the 1,000-db or 2,000-db surfaces show

little difference. Wooster and Reid also computed the transport of

various Pacific eastern boundary currents. The results show that

the choice of reference level below 1,000-db seems to make little

difference to the computed values with respect to 1,000-dbs.

McAlister (1962) has examined the depth of no motion in the

North Pacific with the data down to 5,000 meters. The depth of no

meridional motion that McAlister computed was at about 1,000 meters

in the area along 48°N. and between 130°W. to 150°W. That is close

to the area studied in this paper.

Therefore, it seems reasonable to believe that the depth of no

motion off Newport, Oregon is at some depth around 1,000 meters.
In other words, the depth of 1,000-dbs chosen for the dynamic compu-
tations in this paper seems to be an adequate approximation.

Computational Results

The results of the meridional geostrophic currents relative to
1,000 decibars off Newport, Oregon are shown in Table 2 and illu-
strated in Figure 5. The computed geostrophic currents generally
were weak and irregular.

The surface currents mostly ranged from 2 to 6 cm/sec. The
maximum speed computed was 21 cm/sec between stations NH 85
and NH 105 in September. Several times speeds over 10 cm/sec
appeared in April and July.

For sake of convenience in generalizing the currents in the
whole area, stations NH 65 - NH 105 ($125^{\circ}35'$ to $126^{\circ}31'$ W. longi-
tude) are classified as the inshore area, and NH 105 - NH 165 (126°
$31'$ to $128^{\circ}55'$ W. longitude) as the offshore area.

Through the year, inshore average currents were stronger
than those offshore. Inshore current speeds ranged from 3 to 12
cm/sec except for a couple of very weak currents in April and May.
Offshore current speeds ranged from 1 to 5 cm/sec. Generally,
the current speed tended to decrease with depth.

The current direction was irregular. Sometimes the flow

reversed direction at every station. This result is primarily due

to the fact that the difference of geopotential anomaly between two

adjacent hydrographic stations was so small that the differences

are uncertain, since the geopotential anomaly difference, from

which the geostrophic currents are computed, are uncertain to 0.01

dyn meters (Smith, 1964). The geostrophic current will be uncer-

tain, if the magnitude of currents are less than the order of 2.6

cm/sec. (Note: In the latitude $44^{\circ}39'$ N. with station spacing of

20 nautical mile, the geostrophic current is in the order 2.6 cm/sec

when $\Delta D_A - \Delta D_B = 1$ dyn cm.)

More than 50% of the geostrophic currents were less than

2.6 cm/sec. Therefore, it is very difficult to draw any generaliza-

tions about the current system in this area.

However, if we ignore currents of small magnitude (less than

2.6 cm/sec), the data show some characteristics in common.

The meridional flow of geostrophic current through the New-

port hydrographic section may be summarized as follows:

From June through November inshore average flows were pre-

dominantly southward, while offshore average flows were northward.

From December through May, inshore average flows were pre-
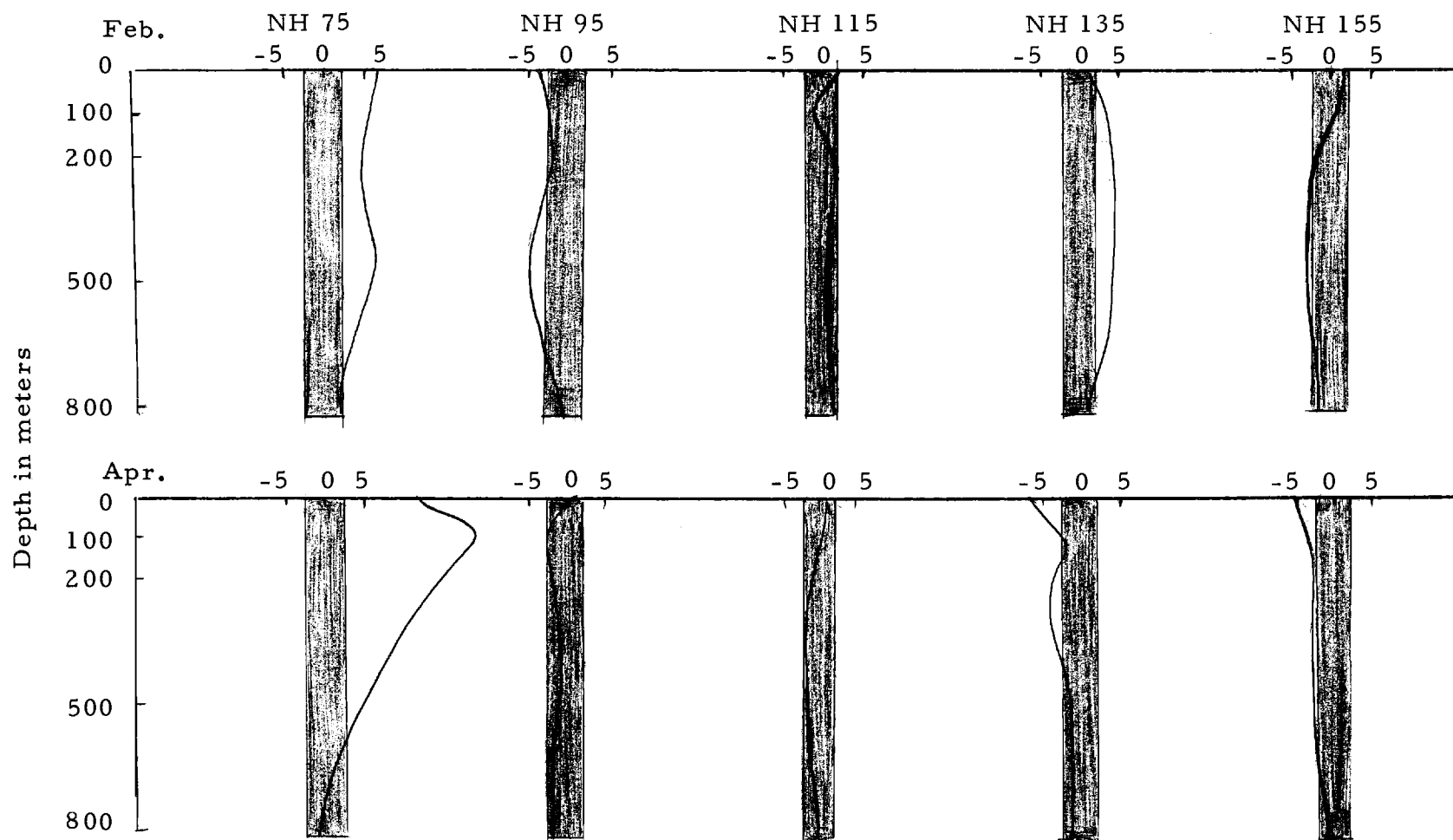
dominantly northward while offshore flows were southward.

Table II. Meridional geostrophic current relative to 1,000 decibars
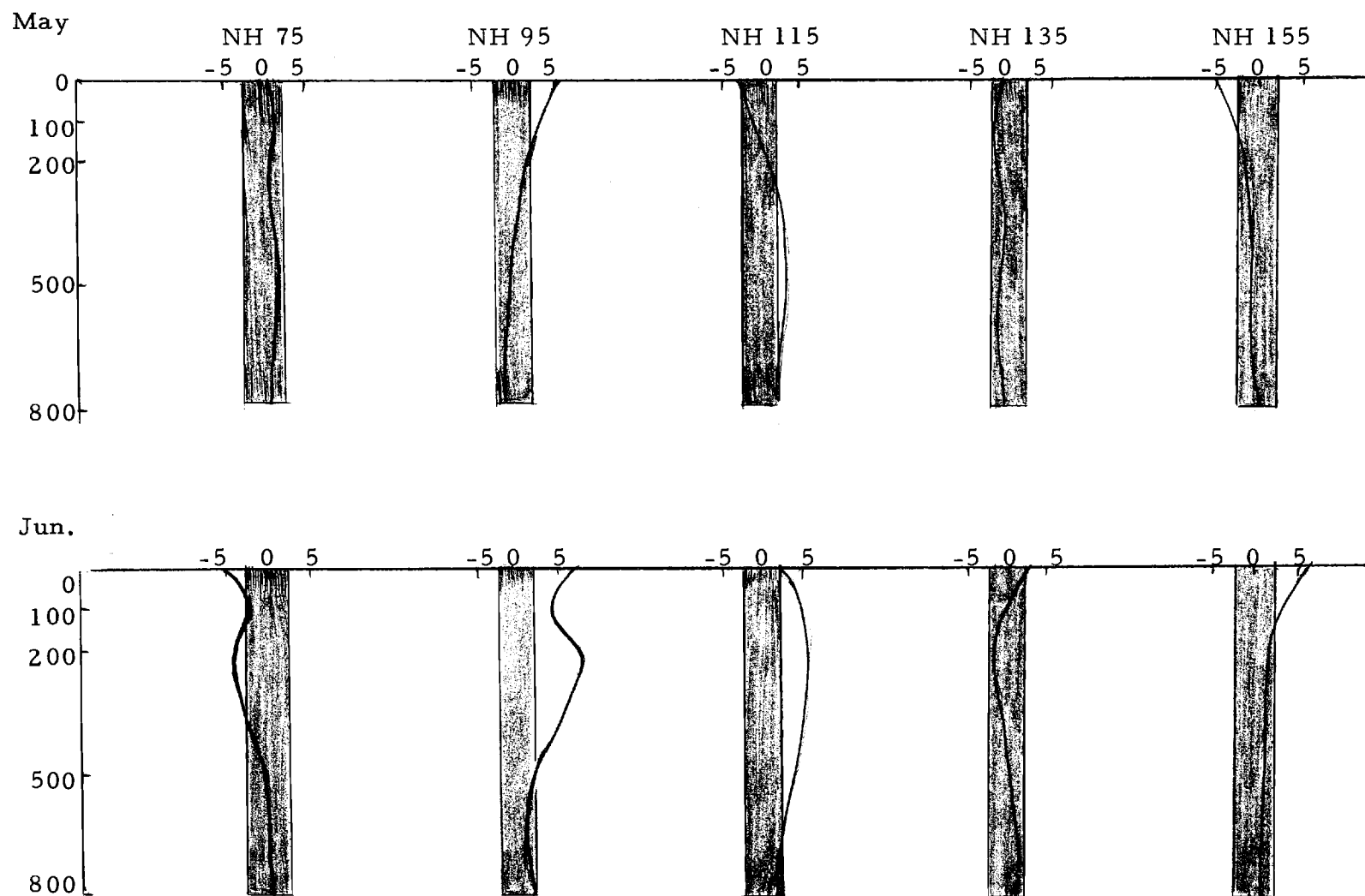in cm/sec. (positive values indicate northward flow)

February

Depth (meters)	NH 65-85	NH 85-105	NH 105-125	NH 125-145	NH 145-165
0	6	-4	1	2	1
100	6	-3	1	3	0
200	5	-3	1	4	2
500	6	-5	1	4	2
800	1	-1	1	1	1

April

0	14	-1	0	-5	-4
100	18	-1	1	-2	-3
200	14	-1	2	-4	-2
500	5	0	2	-2	-2
800	2	0	1	-1	-1

May

0	1	4	-3	-1	-5
100	1	3	0	-2	-4
200	0	1	2	-1	-3
500	1	0	3	-2	-1
800	1	-1	2	-1	-1

Table II. cont.

June

Depth (meters)	NH 65-85	NH 85-105	NH 105-125	NH 125-145	NH 145-165
0	-5	7	2	2	5
100	-3	5	5	-1	2
200	-5	8	5	-2	1
500	0	2	3	-1	1
800	0	3	1	0	0

July

0	-12	-4	5	-1	-2
100	-4	-3	4	-2	0
200	0	-3	2	-1	0
500	0	-2	0	2	-1
800	0	0	1	1	0

September

0	4	-21	3	-5	4
100	6	-7	7	-4	-2
200	5	2	7	-4	-1
500	6	-1	4	-2	0
800	2	0	1	0	-1

Table II. cont.

November

Depth (meters)	NH 65-85	NH 85-105	NH 105-125	NH 125-145	NH 145-165
0	-6	-3	5	3	-
100	-6	1	4	2	-
200	-4	0	5	2	-
500	-2	1	2	3	-
800	1	1	1	0	-

December

Depth (meters)	NH 65-85	NH 85-105	NH 105-125	NH 125-145	NH 145-165
0	4	5	-5	3	-4
100	4	3	-4	2	-2
200	5	1	-4	3	-1
500	2	0	-3	3	1
800	1	0	0	0	1

Figure 5. Meridional geostrophic currents relative to 1,000 decibars in cm/sec (positive values indicate northward flow, shading means current of uncertainty).

May

NH 75   NH 95   NH 115   NH 135   NH 155

Figure 5. cont.

July

NH 75   NH 95   NH 115   NH 135   NH 155
-5 0 5   -5 0 5   -5 0   5   -5 0 5   -5 0 5

0
100
200

500

800

Sept.

-5 0 5   -5 0 5   -5 0 5   -5 0 5   -5 0 5

0
100
200

500

800

Figure 5. cont.

27

Nov.

NH 75　　　　NH 95　　　　NH 115　　　　NH 135　　　　NH 155
-5　0　5　　　-5　0　5　　　-5　0　5　　　-5　0　5　　　-5　0　5

0
100
200

500

800

Dec.

-5　0　5　　　-5　0　5　　　-5　0　5　　　-5　0　5　　　-5　0　5

0
100
200

500

800
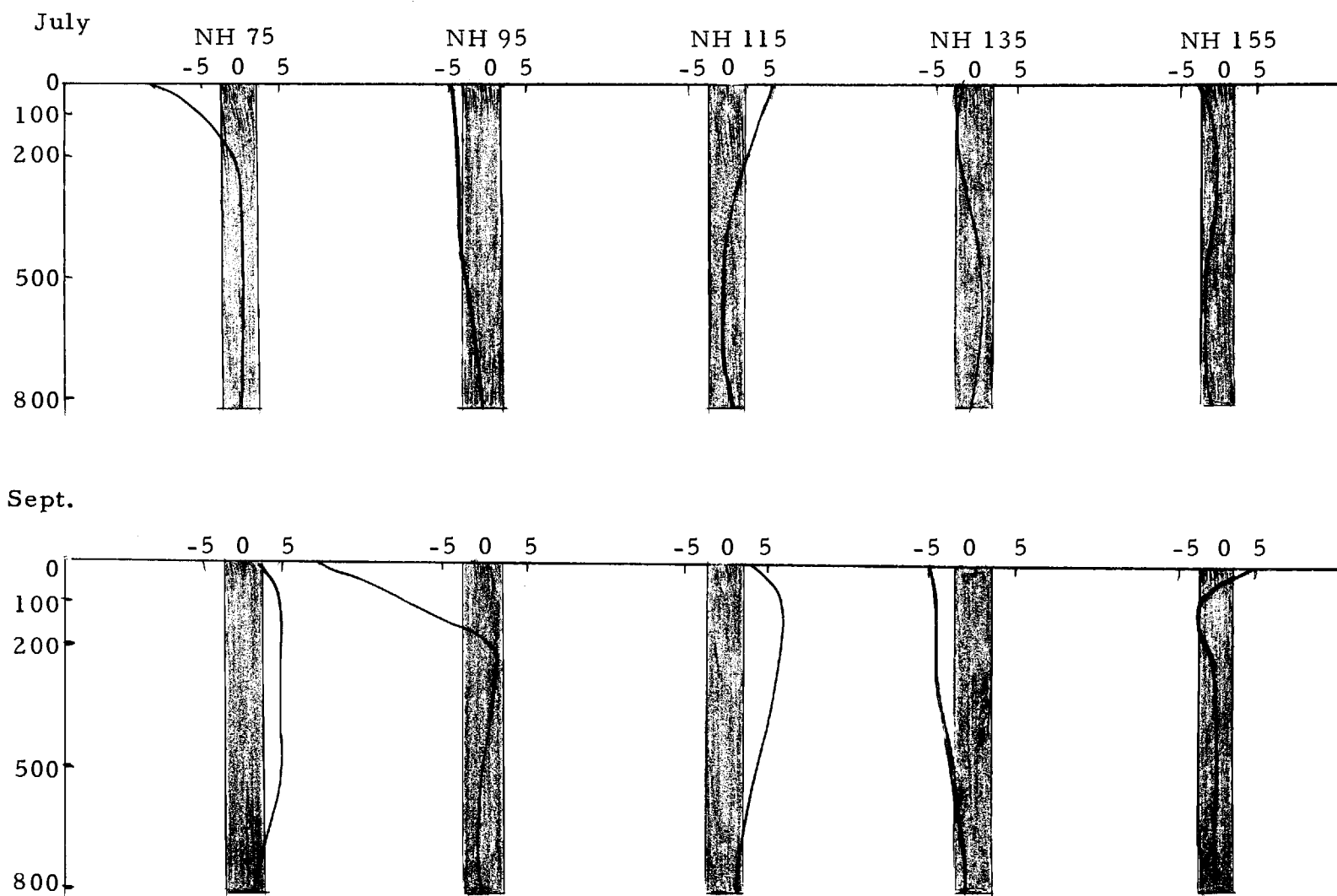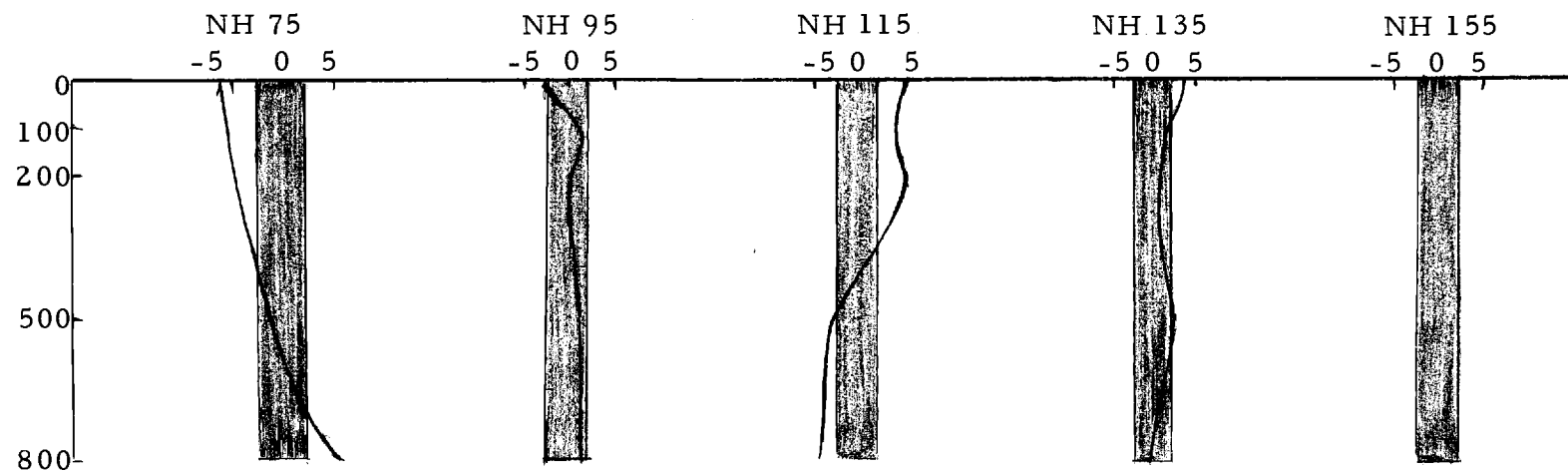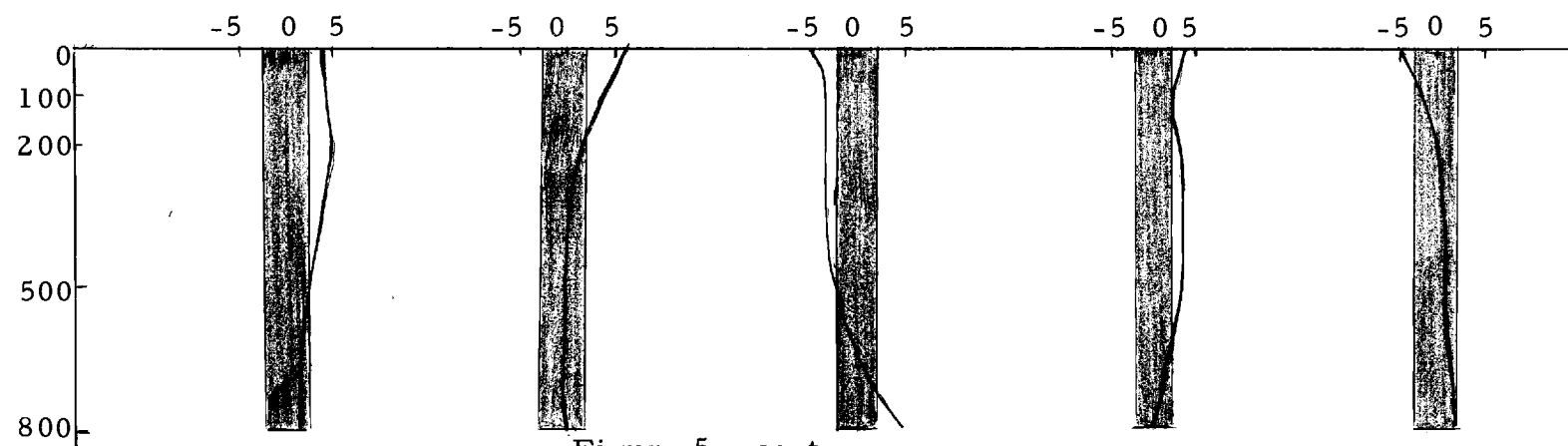
Figure 5.　cont.

Comparison of Geostrophic Computation with Drogue Measurements

The currents obtained by geostrophic computation were compared with direct measurements using drogues (Pattullo and Stevenson, in preparation). These are shown in Table 3 and Figure 6.

Table 3 shows that the currents determined by the two methods were generally in agreement with a few exceptions in September and November. However, the great descrepancies occurred during weak currents which are indeterminant by the geostrophic computation. Figure 6 shows the correlation between currents determined by geostrophic computation and by drogue measurements. The correlation coefficient is 0.96. For a one-to-one plot, the slope of correlation line is 43.8° which demonstrates a very good correlation between two methods.

It is noteworthy that the drogue measurements were taken at NH 45, while the hydrocasts, from which the geostrophic currents were computed, were taken between NH 65 and NH 85. The distance between two measurements was twenty or thirty nautical miles. Furthermore, the drogue measurements and hydrocasts were taken in the same month, but the dates were different. This leads to the postulation that in the open ocean (beyond NH 45) the steady currents are somewhat persistent. However, the data are limited, there were only six cruises for which data could be compared.

Table III.   Comparison of geostrophic currents with currents
measured by drogues.

	Meridional Component of Drogue Measurement (at NH 45)	Computed Meridional Geostrophic Current (At NH 65-NH 85)
	May 31, 1962	May 2, 1962
0 m	-1.39	-0.26
100 m	+0.79	+0.79
200 m	+1.39	+2.37
	July 5, 1962	July 27, 1962
0 m	-13.92	-19.01
100 m	- 6.70	-10.56
200 m	-10.86	- 5.81
	July 11, 1965	July 15, 1965
0 m	-20.2	-18.00
100 m	- 7.6	- 8.70
200 m	- 6.1	- 2.64
500 m	- 3.6	- 1.58
	Sept 24, 1962	Sept 5, 1962
0 m	- 2.04	+ 0.26
100 m	+ 0.60	- 1.05
200 m	+ 3.59	- 0.52
	Nov 17, 1962	Nov, Average
0 m	-13.87	- 5.81
100 m	- 3.28	- 5.54
200 m	- 4.52	- 3.69
	Dec 17, 1964	Dec, Average
0 m	+ 5.80	+ 3.96
100 m	+ 0.30	+ 3.69
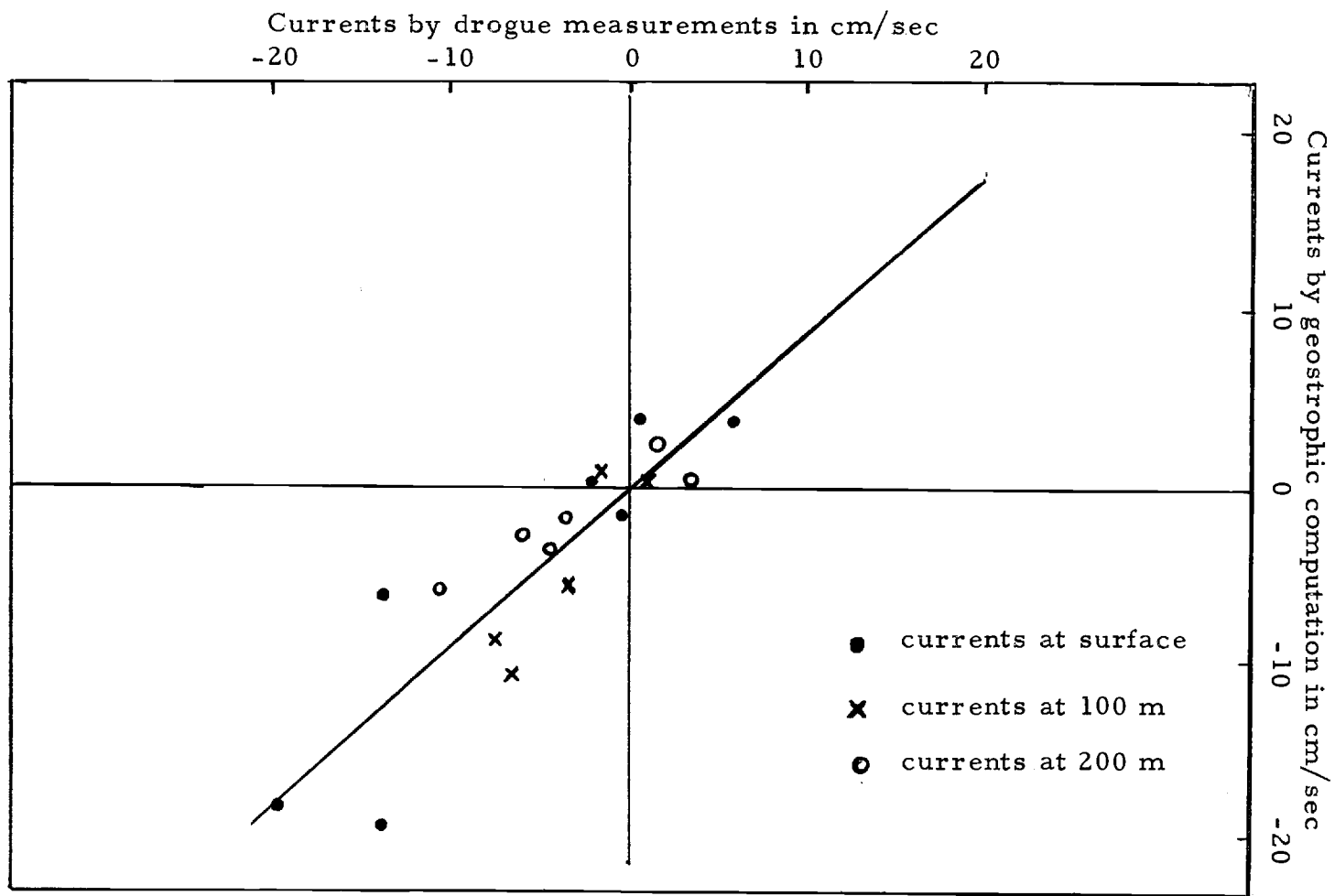
Figure 6. Correlation between geostrophic computation and drogue measurements.

V. DISCUSSION

Reliability of Geostrophic Currents

The reliability of geostrophic computation had been studied by

Van Arx (1962) as follows:

> Comparison of direct measurements of flow observed
> by Pillsbury (1890) in the Straits of Florida, with
> dynamic sections computed by Wüst (1924), for
> the same area show that the relative field of motion
> obtained from dynamic sections, using Pillsbury's
> observed surface of no motion, lead to substantially
> the same results. Further comparisons of the
> field of motion inferred from the distribution of
> density by means of the geostrophic approximation
> have been made by the International Ice Patrol and
> by the Meteor expedition. In these cases, it found
> that the current observed by direct methods may
> depart from the geostrophic field of motion by
> amounts which range from 5% to 25%, with the
> mean departure in the neighborhood of 15%. Stu-
> dies of the effects of curvature on the transport
> of the Gulf Stream, using data from the 1950
> multiple-ship survey of that current off the coast of
> New England. These studies show that the difference
> between the computation based on geostrophic flow
> and that for meander flow can amount to 15%. How-
> ever, Parr (1938), in a comparison of isopycnic and
> geostrophic analysis of the flows of the Labrador Cur-
> rent and the Gulf Stream system south of the Grand
> Banks of Newfoundland, showed that agreement "would
> be utterly unobtainable" in some portions of the area.
> (Von Arx, 1962, p. 254)

Possible Factors and Errors Affecting the
Geostrophic Computation

In this paper, the results of dynamic computation showed some

variability in current direction that led to the examination of possible

factors which may affect the dynamic computation of geostrophic

flow.

Defant (1950) has shown that the errors produced by long-period

internal waves are not negligible, and has suggested observational

precautions which tend to suppress their influence.

Wooster and Taft (1958) examined the measurement error in

the geopotential anomaly of the upper 1,000 m (0/1,000 db) and con-

cluded that the error in anomaly difference between two stations is

± 1.1 dyn cm (two standard deviations). They were aware that the

large error due to salinity titration constituted the greater part of

the anomaly error. With the modern salinometers, the measurement

error still exists ± 0.3 dyn cm as compared to their value of ± 1.1

dyn cm (Reed and Laird, 1966).

Internal waves of near tidal period could produce variations in

anomaly greater than those caused by measurement error. Reid

(1956) reported 12-hr variations as great as 5 dyn cm at single sites

off the California coast but later suggested (1961) that such large

fluctuations are rare over most of the Pacific Ocean. According to

Reed et al., the measurement error and fluctuations due to waves

combine to produce a value very close to the measurement error

reported by Wooster and Taft; consequently, ± 1.1 dyn cm will be

used as the total error estimate in the anomaly difference

(0/1,000 db) between the two stations.

The Department of Oceanography, Oregon State University, has occupied anchor stations to study the internal wave fluctuation off the Oregon Coast. Data from one such cruise showed that the dynamic height anomaly due to internal wave fluctuation at station NH 65 for a 25-hr period was in the order of ± 1.0 dyn cm (personal communication with Mr. Mooers). If we consider the difference of geopotential anomaly between two stations, we have to take the root mean square value. Then, the anomaly difference between two hydrographic stations will be 1.4 dyn cm (i.e. $\sqrt{2} = 1.4$). That is close to the value 1.1 dyn cm given by Reed (1966) for Pacific area.

Among those factors which affect the geostrophic computation, internal wave effect is the largest. Thus, the geostrophic flow may be completely indeterminate if the difference in geopotential anomaly between two adjacent stations is less than 1.4 dyn cm since this difference $(\Delta D_A - \Delta D_B)$ by which the geostrophic flow is computed, may be overweighed by the internal wave effect. In other words, a geostrophic current less than the magnitude of 3.6 cm/sec will be uncertain. Therefore, the value of 3.6 cm/sec seems to be the critical test value below which the current from geostrophic computation in this paper might be uncertain. However, the dynamic computation in this paper was based on several cruise data. Thus, the internal wave effect should be eliminated to some extent. In

other words, if the computation is based on an average value of

N-cruises' data, the critical value should be divided by $\sqrt{N}$

(since standard deviation of the error should be divided by $\sqrt{N}$).

In this paper, the data used to compute geostrophic flow were

varied from four to two cruises (see Table 1), therefore, the criti-

cal value should be reduced to 1.8 cm/sec, 2.0 cm/sec, and 2.5

cm/sec respectively. If we round off the decimal point which is

insignificant in geostrophic computation, the overall critical testing

value becomes 2 cm/sec. Namely, any current computed in this

paper less than 2 cm/sec is uncertain.

A second factor affecting the geostrophic compution is the

error of positioning induced by the use of Loran. Von Arx (1962)

states:

> Experience at sea indicates that the standard error
> in Loran, a fix made in range 300 to 500 miles from
> transmitters is approximately 0.7 nautical miles.

With the present equipment aboard the R/V Yaquina, and

R/V Acona, experience has shown that the average error in posi-

tioning with Loran is less than 0.5 nautical mile, which is some-

what better than that stated by Von Arx.

In this paper, the station spacing, L, chosen for geo-

strophic computation is 20 nautical miles, and the positioning

error induced by Loran, dL, is ± 0.5 nautical miles. Then

positioning error ϵ_p will be

$$\epsilon_p \;=\; \frac{dL}{L} \;=\; 0.05 \quad \text{or} \quad 5\%$$

In view of the fact that the geostrophic currents computed in this paper were generally in the order of 5 cm/sec or less, the positioning error induced by Loran can be neglected.

VI. SUMMARY AND CONCLUSIONS

The annual average of geopotential anomaly off Newport,

Oregon was 1.31 dynamic meters. The seasonal variation was of the

order of 14 dyn cm. The highest value occurred in summer, parti-

cularly September, and lowest value occurred in spring, generally

April. This feature was primarily related to oceanic heat content in

the upper layer. By regressional analysis, the geopotential anomaly

and oceanic heat content was found to be linear in character, and the

correlation computed was 3.1×10^{-1} dyn cm/kg cal/cm^2.

The currents off the Oregon Coast were weak and irregular.

The surface currents mostly ranged 2 to 6 cm/sec with a maximum

current speed of 21 cm/sec, and a minimum speed of zero cm/sec.

The average magnitude of meridional flow was in the order of

5 cm/sec or less. The direction of meridional geostrophic flow

varies with season. From June through November , inshore average

flow was southward whereas offshore average flow was northward.

From December through May the average inshore flow was north-

ward whereas offshore average flow was southward.

The geostrophic computation was affected by the internal wave

fluctuation. A current less than 2 cm/sec was uncertain. The geo-

strophic computation due to positioning error can be practically

neglected. The currents determined by geostrophic computation and

drogue measurements were generally in agreement.

BIBLIOGRAPHY

Burt, W., J. Pattullo and S. Kulm. 1967. Unpublished research on oceanic heat content off the Oregon Coast. Corvallis, Oregon. Oregon State University, Department of oceanography.

Burt, W. and B. Wyatt. 1964. Drift bottle observations of the Davidson Current off Oregon. University of Washington, Studies on Oceanography, p. 156-165.

Defant, Albert. 1961. Physical oceanography. New York, Pergamon. 2 vols.

Fomin, L. M. 1964. The dynamic method in oceanography. New York, Elsevier. 212p.

Maughan, Paul M. 1963. Observation and analysis of ocean currents above 250 meters off the Oregon Coast. Master's thesis. Corvallis, Oregon State University. 49 numb. leaves.

McAlister, William. 1962. The general circulation in the North Pacific Ocean referred to a variable reference surface. Ph.D. thesis. Corvallis, Oregon State University. 69 numb. leaves.

Pattullo, J. and W. Denner. 1965. Processes affecting seawater characteristics along the Oregon Coast. Journal of Limnology and Oceanography 10 (3): 443-450.

Pattullo, J., M. Munk, R. Revelle and E. Strong. 1955. The seasonal oscillation in sea level. Journal of Marine Research 14 (1): 88-121.

Reed, R. K. and N. R. Laird. 1966. On the reliability of geostrophic flow determinations across a section of the North Pacific Ocean. Paper presented at the Pacific Northwest Regional Meeting of the American Geophysical Union. Corvallis, Oregon.

Reid, J. L., Jr. 1956. Observation of internal tides in October 1950. Transactions of the American Geophysical Union 37 (3): 278-286.

Reid, J. L., Jr. 1961. On the geostrophic flow at the surface of the Pacific Ocean with respect to the 1,000-decibar surface. Tellus 13 (4):489-502.

Smith, Robert. An investigation of upwelling along the Oregon Coast. Ph. D. thesis. Corvallis, Oregon State University. 83 numb. leaves.

Smith, R., J. Pattullo and R. Lane. 1966. An investigation of the early stage of upwelling along the Oregon Coast. Journal of Geophysical Research 71 (4): 1135-1140.

Stevenson, Merritt. 1966. Subsurface currents off the Oregon Coast. Ph.D. thesis. Corvallis, Oregon State University. 140 numb. leaves.

Sverdrup, H. U. 1947. Wind-driven currents in a baroclinic ocean. Proceedings of the National Academy of Sciences 33: 318-326.

Sverdrup, H. U., M. W. Johnson and R. H. Fleming. 1942. The oceans, their physics, chemistry and general biology. New York, Prentice-Hall. 1087p.

Von Arx, W. S. 1962. An introduction to physical oceanography. Massachusetts, Addison-Wesley Publication Co. 422p.

Wooster, W. S. and J. L. Reid, Jr. 1963. Eastern boundary currents. In: The sea, ed. by M. N. Hill. Vol. 2. New York, Interscience. p. 253-280.

Wooster, W. S. and B. A. Taft. 1958. On the reliability of field measurements of temperature and salinity in the ocean. Journal of Marine Research 17: 552-556.

APPENDIX FIGURES

TEMPERATURE AND SALINITY DISTRIBUTION
OFF NEWPORT, OREGON

DISTANCE OFFSHORE (NAUTICAL MILES)

185 165 145 125 105 85 65 45 35 25 15 5

12
13
12
11
10
9
8
/7

DEPTH (METERS)
0 50 100 150 200

T

DISTANCE OFFSHORE (NAUTICAL MILES)

185 165 145 125 105 85 65 45 35 25 15 5

30
31
32
32.5
33
33.5

DEPTH (METERS)
0 50 100 150 200

S

DISTANCE OFFSHORE (NAUTICAL MILES)

185 165 145 125 105 85 65 45 35 25 15 5

12
11
10
9
8
7
6
5
4

DEPTH (METERS)
0 200 400 600 800 1000 1200

T

DISTANCE OFFSHORE (NAUTICAL MILES)

185 165 145 125 105 85 65 45 35 25 15 5

32
31
33
34

DEPTH (METERS)
0 200 400 600 800 1000 1200

S

6305 NH

41

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

6309 NH

42

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

6311 NH

DISTANCE OFFSHORE (NAUTICAL MILES)

185   165   145   125   105   85   65   45 35 25 15 5

T

DISTANCE OFFSHORE (NAUTICAL MILES)

185   165   145   125   105   85   65   45 35 25 15 5

S

DISTANCE OFFSHORE (NAUTICAL MILES)

185   165   145   125   105   85   65   45 35 25 15 5

T

DISTANCE OFFSHORE (NAUTICAL MILES)

185   165   145   125   105   85   65   45 35 25 15 5

S

DEPTH (METERS)

6402 NH

44

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

6404 NH

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

6502 NH

46

DISTANCE OFFSHORE (NAUTICAL MILES)

185   165   145   125   105   85   65   45 35 25 15 5

DISTANCE OFFSHORE (NAUTICAL MILES)

185   165   145   125   105   85   65   45 35 25 15 5

DEPTH (METERS)

9.5

9

8.5

8

T

6503 NH

32

32.5

33

33.5

S

47

6506 NH

48

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

6507 NH

49

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

DISTANCE OFFSHORE (NAUTICAL MILES)

6511 NH

DISTANCE OFFSHORE (NAUTICAL MILES)

T

DISTANCE OFFSHORE (NAUTICAL MILES)

S

DISTANCE OFFSHORE (NAUTICAL MILES)

T

6512 NH

DISTANCE OFFSHORE (NAUTICAL MILES)

S

51