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Title: A STATISTICAL STUDY OF WINDS AND SEA WATER TEMPERATURES DURING OREGON COASTAL UPWELLING

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The effect of the wind on sea temperature during the early stage of upwelling was studied by descriptive and statistical analyses. The temperature was recorded at a 20 meter depth five and seven nautical miles offshore from Depoe Bay, Oregon, in the spring of 1967 and at a 25 meter depth seven nautical miles offshore in the spring of 1968. The temperature records were filtered to suppress frequencies higher than about one cycle per day in order to study the response to meteorological storms.

A comparison was made between geostrophic winds computed for 45° N. and 125° W. and winds observed by the U. S. Coast Guard at Newport, Cregon. The geostrophic winds were considered to be the best estimate of the actual wind over the sea surface. The correlation of temperature with the components of the geostrophic wind at successive intervals prior to the time of the temperature observation indicated that both components were correlated with temperature and the components approximately three days prior to the temperature were the most highly correlated. A model was presented to explain the combined effect of both components on the temperature. The model also indicated an intensification of the southerly longshore surface wind.

A multiple regression analysis was used to determine which geostrophic wind components preceding the time of the temperature observation had the most significant effect on the temperature. In general the north-south component preceding the temperature by two and one-half to three and one-half days had the most significant effect. However, the east-west component for similar periods preceding the temperature observation also had a significant effect. The linear relationship between the temperature and the significant wind components explained from 30% to 62% of the variation in temperature.

A Statistical Study of Winds and Sea Water Temperatures During Oregon Coastal Upwelling

by

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A STATISTICAL STUDY OF WINDS AND SEA WATER TEMPERATURES DURING OREGON COASTAL UPWELLING

INTRODUCTION

Upwelling is a widely used term in oceanography but no definition has yet been universally accepted. Perhaps this is because, according to Defant (1961):

> "A totally satisfying explanation of upwelling at continental coasts has not been given, and is probably not possible at all since the total process is composed of a number of sub-stages each of which is always controlled by other factors".

Smith, Pattullo and Lane (1966) have applied a basic definition, developed by Sverdrup (1938) and Wyrtki (1963), to upwelling along the western coasts of the continents and the Oregon coast in particular. Their definition is that the term upwelling applies only to ascending motion, of some minimum duration, of water from depths not exceeding a few hundred meters which results from horizontal divergence in the surface layer. Along the western coasts of continents this horizontal divergence is usually wind induced.

The effect of the wind on surface currents was first explained by Ekman in 1905. The net transport due to the wind occurs within the upper 100 meters and is referred to as the Ekman transport. The Ekman transport of surface water normal to a coast is determined by the ratio of the longshore component of the wind stress to the Coriolis parameter, and in the northern hemisphere this net transport is directed 90° to the right of the longshore wind. Along the

Oregon coast during the spring and summer predominately northerly winds result in offshore transport of surface waters which are replaced inshore by the colder water upwelled from below the surface layer. These winds are produced by the North Pacific High Pressure Cell which dominates the atmospheric circulation during these seasons. In late fall and winter the North Pacific High weakens and moves south, southerly winds predominate and upwelling is no longer common. However, there are occasional occurrences of upwelling in fall and winter, and there are often reductions of upwelling during the summer, which appear to be closely related to the winds (Reid, 1960).

This study was made during the early stage of upwelling, a period when the isopycnals near the coast are beginning to rise toward the surface. This change in density occurs because subsurface water, which is colder and more saline, rises toward the surface during upwelling. The intent of this study was to observe near surface water temperature during the early stage of upwelling before equilibrium was established in order to determine how the wind affected the water temperature.

The Ekman transport is related to the wind stress which can be estimated by a function of the square of the wind speed and an empirically determined drag coefficient. However, the wind used in this paper was not converted to wind stress but resolved into its velocity components, and a statistical analysis was used to explain the changes in the temperature record. It was considered preferable to use the more directly determined wind velocity rather than an empirical (and non-linear) function of the wind velocity for statistical correlations of temperature with winds.

Wind data were obtained by the calculation of geostrophic winds from six-hourly and monthly mean barometric pressure charts, and from surface winds observed by the U. S. Coast Guard at Newport, Oregon. Monthly means of the wind data acquired by the different methods were compared to determine if they all recorded similar meteorological conditions. The correlation between semi-daily geostrophic winds computed for 45° N., and 125° W. and the winds for the same interval observed by the U. S. Coast Guard was determined. Both types of wind data were used in earlier studies related to upwelling. The corresponding geostrophic wind data were studied in relation to two records of 20 meter sea water temperature in 1967 and one record of 25 meter sea water temperature in 1968 using descriptive and regression analyses.

The descriptive analysis of the sea water temperature and wind data was limited to the periods of marked change in the temperature record to determine their association with major changes in wind components prior to the time that the temperature was recorded. During some of these periods temperature versus depth profiles were available to indicate the relationship of the thermal structure of the water column to the temperature recorded by the thermograph placed in that water column.

The regression analysis indicated which wind components, preceding an observed temperature, were most significant in affecting the temperature. It is important to remember that these wind components apply to temperature observed during the early stage of upwelling. During this period there are also some major meteorological events that are reflected in the temperature record. The differences in relationships of wind components to temperature for 1967 and 1968 were noted, and the descriptive and regression analyses were examined for similarities.

REVIEW OF THE LITERATURE

The literature on upwelling is extensive and wide-spread as shown in the recent review article by Smith (1968). In this chapter only the work on upwelling that has had a direct influence on this thesis is reviewed. In Appendix I a brief review of some of the theories of coastal upwelling is given and serves to indicate that the wind, through the wind stress, is probably the most important single parameter in upwelling.

Coastal upwelling off the Oregon coast has been separated into an early stage and an equilibrium stage (Smith, Pattullo and Lane, 1966). Before the early stage the pycnocline, which is represented by the 25.4 and 26.0 σ_T surfaces (Collins, 1964), is level. In the early stage the average vertical velocity at the pycnocline can be approximated by the rate of rise of the 26.0 σ_T surface. The pycnocline may rise rapidly and intersect the surface as a front during intense upwelling causing a large horizontal gradient of surface temperature and salinity in the offshore direction. Sea level is affected by upwelling because warm, less dense surface water is replaced with cooler, more dense water increasing the mean density of the water column. When the water mass tends to isostatically readjust, the mean sea surface is lowered in the upwelling zone. Stewart (1960) related coastal water temperature to sea level along the Pacific coast and attributed some of the variation to upwelling.

An investigation of early upwelling off the southern coast of Oregon by Smith, Pattullo and Lane (1966) was unique in that it provided data for computing offshore transport by three different techniques and the results for all three were in good agreement. The techniques were: by changes in the distribution of temperature and salinity (similar to the Ekman-Sverdrup technique), by heat budget considerations, and by the Ekman transport. Surface wind data were obtained indirectly by computations from twelve-hourly sea level barometric pressure charts and also from shipboard wind measurements. The significant result of the investigation was the good agreement of the Ekman transport with values obtained by the other two methods.

A statistical analysis of upwelling in the Gulf of Panama was made by Schaefer, Bishop and Howard (1958), and the technique was similar to the technique used in this study. They anticipated that wind measurements at Cristobal, where the winds come in from the open Caribbean, would be more closely related to effects in the Gulf than the winds at Balboa which were influenced by topography. It was determined that the Cristobal winds were not as well correlated with the temperature in the Gulf as were the Balboa winds. A multiple regression analysis determined that the upwelling temperature anomalies were affected by conditions which affect the total temperature structure of the Gulf and local effects of northerly winds, both operating independently.

Geostrophic and observed winds have been used in other studies to obtain values for wind stress. Observed winds from ships or land stations are sometimes questionable because of inaccuracies in measurement systems, unrepresentative locations for wind sensors and sparsity of reports. Montgomery (1936) concluded that, if observations are meager, winds computed from surface pressure analyses are often more consistent and accurate than a few wind observations. Pressure observations from a few ships can often reasonably define a pressure system with fair accuracy because its features do not fluctuate rapidly, but the same number of direct wind observations may prove misleading.

The geostrophic winds were calculated by Smith (1964) and Panshin (1967) from sources that describe the barometric pressure field. The pressure gradient was measured and used in the geostrophic wind equation:

$$\mathcal{V}_{geo} = \frac{1}{\rho_{air} f} \frac{dp}{dx}$$

where $\mathcal{V}_{geo}$ = the speed of the geostrophic wind

ρ_{air} = the density of air

f = the Coriolis parameter

$\frac{dp}{dx}$ = the pressure gradient

The effect of surface friction on the geostrophic wind is to reduce the velocity. This results in a reduction of the Coriolis force and causes the wind direction to shift to the left, if looking downwind in the northern hemisphere (Hess, 1959).

The relation between daily mean sea level and winds along the Oregon coast for the period of May 1933 through March 1934 was studied by Panshin (1967) using the method of regression analysis. Wind data were obtained from daily sea level pressure charts of the Historical Weather Map series by calculating geostrophic winds for 45^{o} N. and 125^{o} W. The vector was then rotated 10^{o} to the left of the downwind direction and the velocity multiplied by 0. 65 to estimate the speed and direction of the surface wind. The components of the wind stress were calculated and various combinations were used as independent variables with the departure of adjusted sea level for a day from the average sea level as the dependent variable. Sea level correlated most highly with the north-south component summed over a four day period. The next highest correlation was with the east-west component summed over a three day period. This indicated a lowering of sea level was associated with a northeast wind.

Bourke (1969) concluded that fluctuations in temperature measured in the Yaquina estuary during the upwelling season were directly related to upwelling off the coast. He considered a regression analysis of temperature in the estuary with the north-south component of winds observed every four hours by the U. S. Coast Guard at the entrance to Yaquina Bay. He determined that the correlation of temperature and wind components was best when the north-south component was averaged over the four days prior to the time of the temperature observation. The regression equation determined by Bourke was:

$$\hat{T} = 11.000^{\circ}C + 0.221\ A(V)$$

where: $\hat{T}$ = the estimated temperature ($^{\circ}$C)

$A(V)$ = the north-south component averaged over the four days prior to the time of the temperature observation

From this study a prediction equation was obtained which would allow one to predict the water temperature in the Yaquina estuary by observing the past wind data.

COLLECTION OF TEMPERATURE AND WIND DATA

Sea water temperature was recorded as part of a project designed to study physical processes during coastal upwelling by means of moored arrays containing recording instrumentation. The first phase of this program began in July 1965 when thermographs and current meters were installed at several depths for sites on the Oregon continental shelf. The observed data and basic statistics were presented in a data report (Oregon, 1966), and the project was unofficially referred to as the Coastal Currents Project. Instrument arrays were again moored during the 1966 summer season and a data report was prepared (Oregon, 1968).

The temperature data for this study were recorded during the 1967 and 1968 spring and summer seasons. The author was actively involved in preparing the instrumentation, mooring and recovering the arrays during the 1968 field work and processing the data for both 1967 and 1968. Sea water temperature was measured by a Braincon type 146 recording thermograph (Brainard, 1964). The project used four thermographs of which two were fitted with -2^{o}C to +25^{o}C thermometers, and two were fitted with +5^{o}C to +15^{o}C thermometers. In the thermograph a mercurial thermometer was placed between a phosphorescent source and photographic film, and the temperature was recorded as a dark line where the mercury prevented film exposure. A Dagmar II 35 mm viewer was used to project the image of the 70 mm film onto a grid graduated in tenths of a degree centigrade, and the values were recorded.

Each array included two thermographs which were placed at depths of approximately 20 and 40 meters. The temperature fluctuations in the deeper thermograph records were all much smaller in magnitude than those in the shallower thermograph records. The temperature profiles, drawn from hydrographic data collected during the installation program, indicated a very small temperature gradient below a depth of 20 meters. This would result in small temperature fluctuations even though large vertical displacements had occurred. Recording periods for the shallow thermographs are presented in Table I.

During 1967 one thermograph was set to advance every five minutes, and the pairs of five minute samples were later averaged to provide uniform ten minute sampling intervals for this study. The other thermograph in 1967 and the two in 1968 were set to advance every ten minutes. According to Braincon specifications the thermographs achieved a 95% response to changes in temperature in ten minutes, and thermometer readout accuracy was $\pm 0.1^{\circ}C$ (Brainard, 1964).

The installation program was very successful in that all arrays were moored and recovered without loss of equipment. However,

Table I

Temperature Records From Shallow Depth Thermographs

Installation Code	Beginning of Record Date	Beginning of Record Time*	End of Record Date	End of Record Time*	Length of Record (days)	Meter Depth (meters)
DB-5** 1967	05/07/67	1910	06/03/67	1700	26.9	20
DB-7 1967	05/07/67	2120	06/08/67	0230	31.2	20
DB-7A 1968	04/17/68	2020	05/27/68	1520	39.8	22
DB-7B 1968	05/24/68	2040	06/01/68	0050	7.1	28
DB-7A+DB-7B 1968***	04/17/68	2020	06/01/68	0050	44.1	22 to 28

* Time is in Greenwich Mean Time.

** Distance offshore from Depoe Bay in nautical miles.

*** Records DB-7A and DB-7B were joined at 1400 GMT on May 26, 1968.

there were some equipment failures, mainly due to a malfunctioning of the film advance mechanism, resulting in partial loss of the long time series of temperature. A laboratory test of thermograph reliability was made in February and March 1968. The results indicated that the advance mechanisms on three of the thermographs were very poor and the mechanisms were replaced. The test of the fourth thermograph indicated that it was reliable and, in fact, it provided the longest temperature record for 1968.

The processing of a thermograph film record required that the date and time, in Greenwich Mean Time, be established for the beginning and end of the record. The beginning was determined by noting the recorded change from air temperature to a stable sea water temperature, and the end of the record was determined by a change in the opposite direction. A logbook of the installation provided the date and time of the sampling interval in which this change took place. An overall test of accuracy was to count the number of intervals observed, convert to total time and, after adding this total to the beginning date and time, compare it with the date and time of retrieval. The results of this test were very good for the two records where it could be applied. A thermograph did not always record from mooring to retrieval due to the malfunctioning of the film advance mechanism. In the other cases, only the beginning of the record could be determined in the usual manner and the end of the record was determined by converting the number of successful sampling intervals to time. No assessment of the accuracy could be made for such cases.

A result of the installation program of 1968 was that the two thermographs at DB-7 recorded simultaneously for approximately three days. The spatial separation of the two thermographs is shown in Figure 1 and the meter depths are given in Table I. A comparison of the records during the period when the thermographs recorded simultaneously indicated that both meters were recording similar temperature fluctuations without any apparent lag in time and fluctuations were of approximately the same amplitude. To increase the available data for the analyses the records were joined at 1400 GMT on May 26, 1968, when they were nearly identical. This time series and the two recorded in 1967 are shown in Table I to be of adequate length for the data reduction and analyses intended in this study.

The principle geostrophic wind data were calculated from pressure gradients on the surface weather analyses prepared by the Northwest Regional Forecast Center, U. S. Weather Bureau. The analyses for 0000 GMT, 0600 GMT, 1200 GMT and 1800 GMT were used because they included all of the more recent ship reports. A template was prepared from the geostrophic wind equation for determining the geostrophic wind velocity and direction at 45^{o} N.

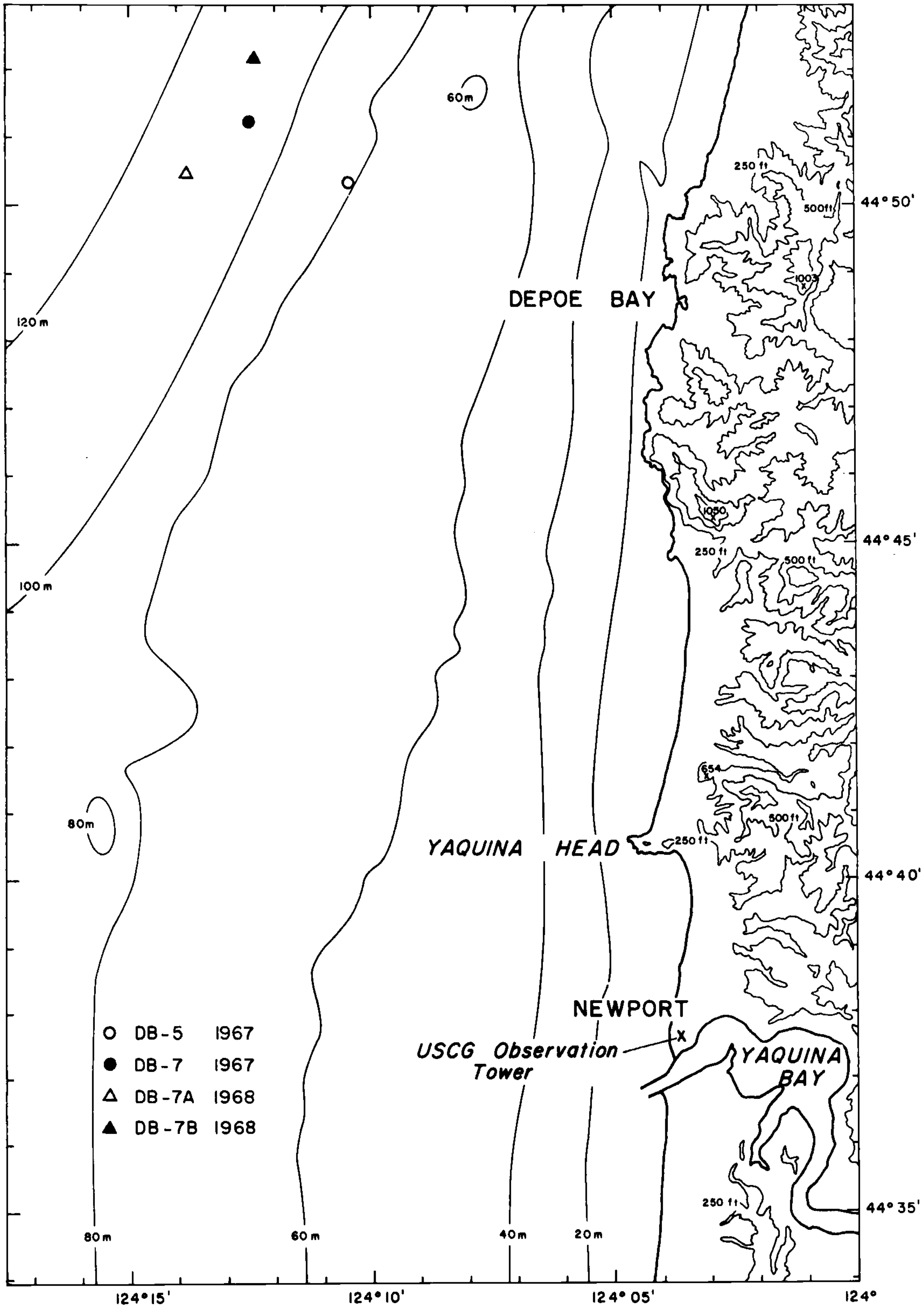

Figure 1. Locations of array stations off Depoe Bay, Oregon (1967 and 1968), and topography near Newport, Oregon.

The geographic coordinates 45° N. and 125° W. were used for wind computations because they were easily located on a weather map and were near the moored arrays. In the preparation of the surface weather analyses smoothed isobars were drawn by considering the atmospheric pressure observed at various weather stations. The geostrophic wind calculated from the pressure gradient near 45° N. and 125° W. appeared to reasonably represent the geostrophic wind over the area including the moored arrays and Newport, Oregon.

In May and June 1967 six-hourly geostrophic winds were calculated for the period when the thermographs were installed in order to correspond to the recorded temperature time series. The intent was to record temperature throughout the entire 1968 upwelling season so geostrophic winds were calculated for April through September 1968.

A second source of geostrophic wind data was the monthly mean sea level pressure charts prepared by the Extended Forecast Division of the National Meteorological Center. The geostrophic relationship was used with the monthly mean pressure gradient to determine the monthly mean geostrophic wind.

Wind observations recorded during the same periods by the U. S. Coast Guard in Newport, Oregon, were studied in relation to the geostrophic wind data. Observed wind data from instruments located near the entrance to Yaquina Bay, as shown in Figure 1, were recorded in the Coast Guard meteorological logbook on a four-hourly basis. The observed winds were instantaneous values and therefore did not necessarily give a valid estimate of the average winds over a four hour period (Bourke, 1969).

REDUCTION OF DATA

Temperature data were plotted as a time series for visual error detection and to aid in planning future analyses. This process proved to be quite effective for inspecting large amounts of data. An error detection program, described in the 1968 Data Report (Oregon, 1968), was also applied to this thermograph data and any questionable temperatures detected by either method were reread on the original film and corrected where necessary.

It was observed that tidal and higher frequency oscillations were present in the temperature time series. This study considered the lower frequency temperature domain which, according to Collins (1968), included the response to meteorological storms. A numerical tapering of the temperature records was used to obtain average temperatures representative of the longer period phenomena. The Cosine-Lanczos Filter-Taper, described in the 1968 Data Report, was used as a low pass filter and taper with a rapid "roll-off". The taper weights, normalization factor and response functions for the Cosine-Lanczos Filter-Taper are given in Appendix II.

The procedure was to first taper the temperature data sampled at ten minute intervals producing data at one hour intervals and then taper again. The resulting data was decimated to six hour intervals and called the "low-low pass". It had a half-amplitude point at $T \approx 36$

hours and it is shown in Appendix II that the "roll-off" rapidly approached zero for frequencies higher than 1/36 cycles per hour, which includes the frequencies of all significant tidal oscillations. The tapered temperature data are shown in Figure 2.

Wind data were recorded as direction in polar coordinates and speed in knots. In order to simplify the statistical analysis and relate to equations expressed in Cartesian coordinates, winds were resolved into their east-west (U) and north-south (V) components. These wind components were each plotted as time series and subjected to the error detection program (Oregon, 1968) to detect possible errors or regions of unusual variation.

Since geostrophic winds were calculated from barometric pressure charts, any suspected values were recalculated, resolved into components, and replotted to note any significant change. A small number of values were still rejected by the error detection program. A criterion established by the author for the Coastal Currents Project, shown in Appendix III, used the predicted values from the program to determine the corrections. This procedure was necessary for correcting the observed wind data because there was no graphic recording of the winds to check, only the values recorded by Coast Guard personnel.

The percentage of the number of components corrected for geostrophic wind data were 3.6% in 1967 and 1.9% in 1968. For the

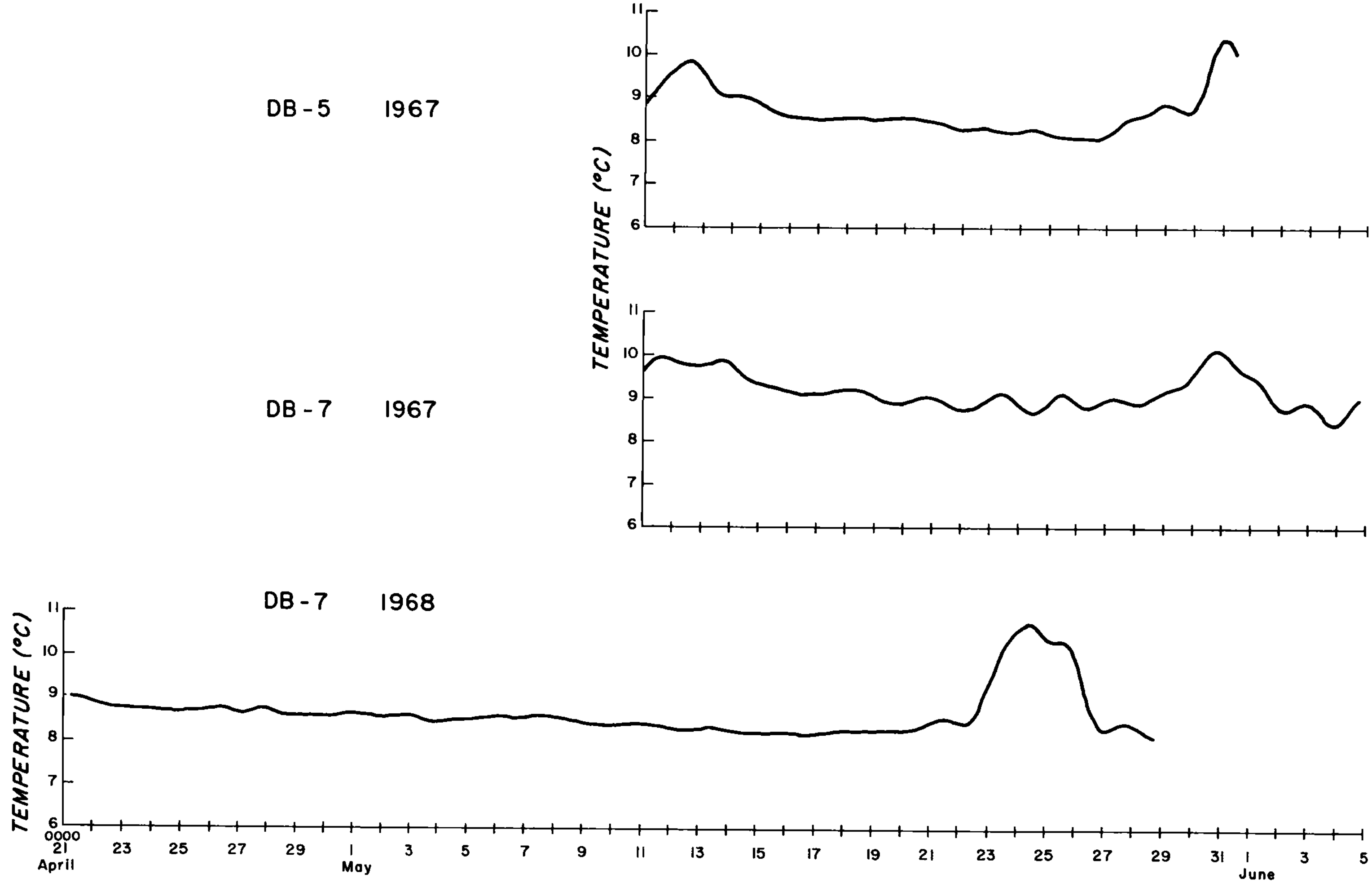

Figure 2. Tapered temperature data

corresponding wind data observed by the Coast Guard the corrections were 2.7% in 1967 and 5.8% in 1968. Monthly means and variances of the wind data before and after correction were compared as shown in Table II. The sign convention used in this thesis for wind components is +U and +V from the east and the north respectively. The result of a statistical test (Student's T Test, P = 0.95) using these means and variances indicated no significant change in the record. The corrected wind data were used for further analyses.

Since a regression analysis of temperature on wind data was considered, it was advisable to place the same frequency restriction on wind and temperature data. Both the geostrophic and observed wind time series were tapered with a half-amplitude point at $T \approx 36$ hours. The taper suppressed higher frequency oscillations such as diurnal wind variations.

The geostrophic wind data had a six-hourly sampling rate, and the winds at Newport, Oregon were observed every four hours. Therefore, a separate taper was derived for each type of data so that they would have the same frequency restriction. The taper weights, normalization factor and frequency response functions for each taper are given in Appendix IV.

Table II

(A) Monthly Means (Knots) For Velocities of Wind Components

Period	Original Observed Winds Newport, Oregon		Corrected Observed Winds Newport, Oregon		Original Six-Hourly Geostrophic Winds 45°N & 125°W		Corrected Six-Hourly Geostrophic Winds 45°N & 125°W		Geostrophic Winds From Monthly Mean Pressure Charts 45°N & 125°W	
	u	v	u	v	u	v	u	v	u	v
May 1967	-1.76	4.28	-1.96	4.73	0.04	9.21	-0.23	9.26	-1.47	6.15
June	-2.04	5.22	-2.00	5.07					0.42	9.22
June*	(-1.97)	(5.10)	(-1.91)	(5.10)	(0.21)	(15.47)	(0.21)	(15.27)		
April 1968	-5.28	7.41	-5.33	6.91	-6.02	8.89	-6.07	8.67	-4.49	7.14
May	-2.26	2.34	-2.37	2.65	-2.87	4.27	-2.87	4.27	-1.40	4.96
June	-5.87	2.56	-5.88	2.27	-1.51	10.23	-1.29	9.97	-1.89	6.15
July	-6.35	4.55	-6.30	4.43	0.92	12.94	0.92	12.50	0.63	7.14
August	-3.95	-1.69	-4.08	-1.67	-2.31	7.81	-2.11	7.92	-2.24	5.75
September	-3.38	1.13	-3.31	1.66	1.35	7.31	1.35	7.31	-0.98	4.36

*June 1967 six-hourly geostrophic wind data only obtained for June 1, 1967, to June 18, 1967. Other data was adjusted to this period.

(B) Monthly Variances ($Knots^2$) for Velocities of Wind Components

Period	Original Observed Winds Newport, Oregon		Corrected Observed Winds Newport, Oregon		Original Six-Hourly Geostrophic Winds 45°N & 125°W		Corrected Six-Hourly Geostrophic Winds 45° & 125°W	
	u	v	u	v	u	v	u	v
May 1967	55.26	139.24	50.93	147.82	89.32	98.79	87.09	99.44
June*	24.11	65.26	24.89	65.26	30.45	48.55	30.45	45.23
April 1968	81.49	78.61	88.64	76.99	74.79	60.77	76.71	61.04
May	71.87	121.95	69.27	112.68	84.60	180.32	84.60	180.32
June	71.99	126.99	70.31	120.88	68.48	126.83	60.97	115.65
July	66.08	103.67	59.18	104.47	58.12	129.40	58.12	119.29
August	57.62	86.50	53.59	86.08	97.52	168.13	102.04	166.60
September	42.18	76.79	43.20	70.20	168.16	141.06	168.16	141.06

*June 1967 six-hourly geostrophic wind data only obtained for June 1, 1967, to June 18, 1967. Other data was adjusted to this period.

STATISTICAL COMPARISON OF GEOSTROPHIC AND OBSERVED WINDS

Since geostrophic winds (Panshin, 1967) and winds observed by the Coast Guard (Bourke, 1969) had been used for studies of upwelling near Newport, Oregon, it was of interest to determine whether or not these studies described the same wind regime, excluding high frequency oscillations. The tapered time series were plotted, as in Figures 3-A, 3-B and 3-C to inspect for correlation between components of geostrophic and observed winds. It was noted that there appeared to be some correlation, especially in the north-south components.

Both types of tapered wind data were equivalently tapered so their correlation coefficients could be computed by using standard statistical techniques. Each tapered time series was decimated to 0000 GMT and 1200 GMT, daily, to provide values representative of the same sample periods for each record. The U and V components of the observed winds were each separately considered to be the dependent variable, while the geostrophic wind components were taken to be the independent variables. The square of the correlation coefficient, expressed as a percentage, is a measure of the amount of variation in the dependent variable which is accounted for by a linear relationship to one or more independent variables. These percentages for each combination over the total observed period and for each

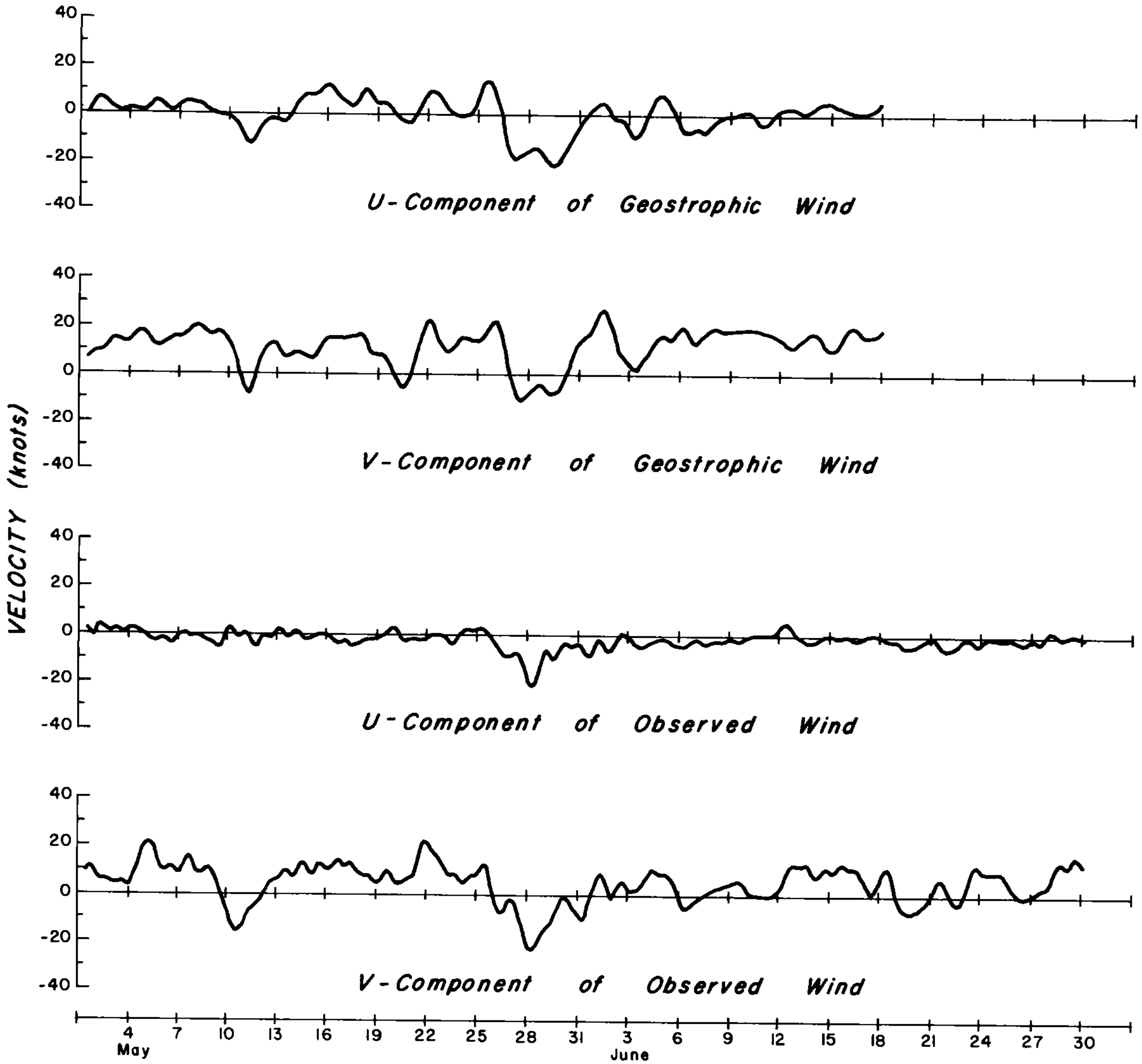

Figure 3-A. Tapered wind components (May-June 1967)

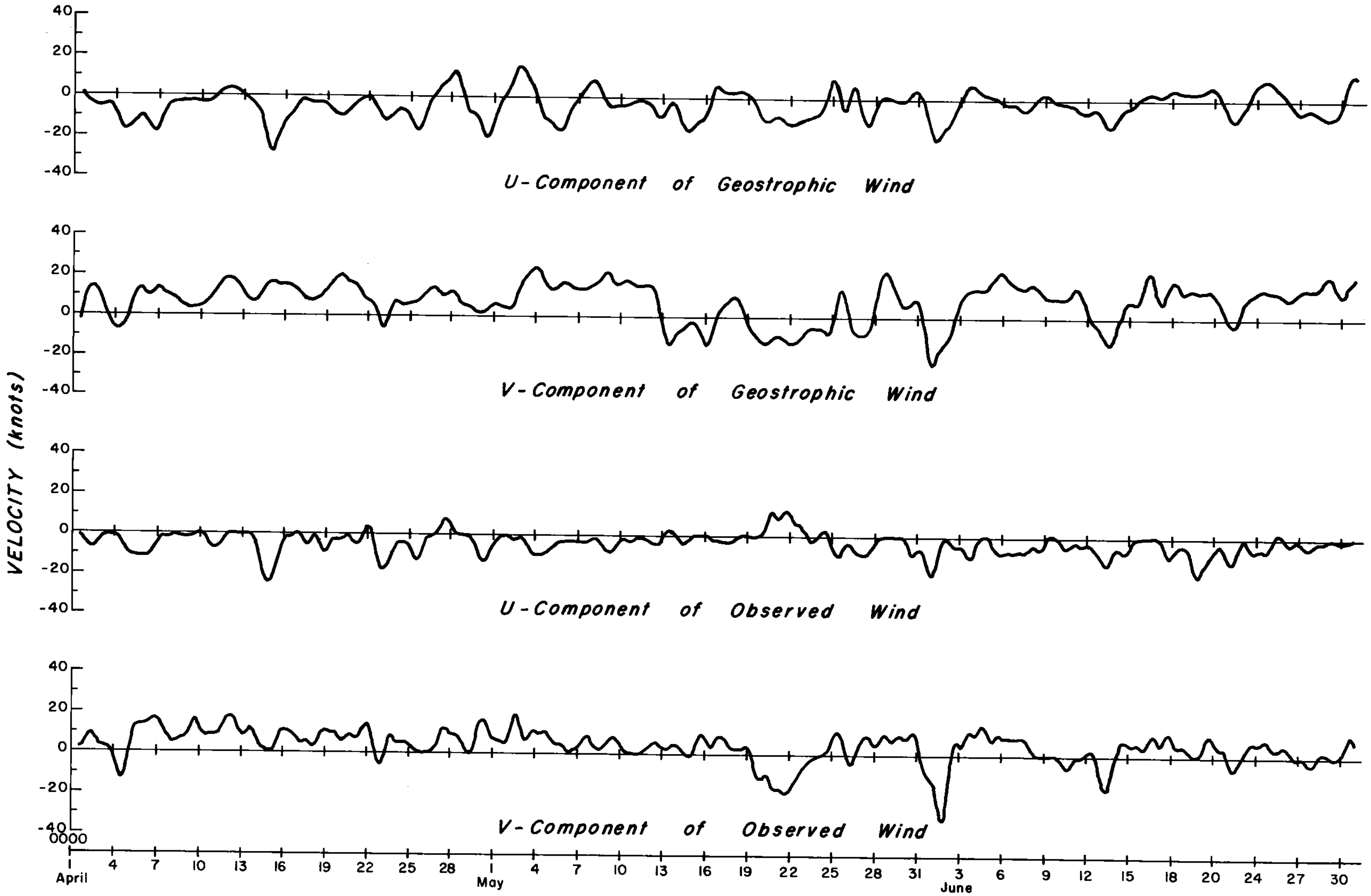

Figure 3-B. Tapered wind components (April-June 1968)

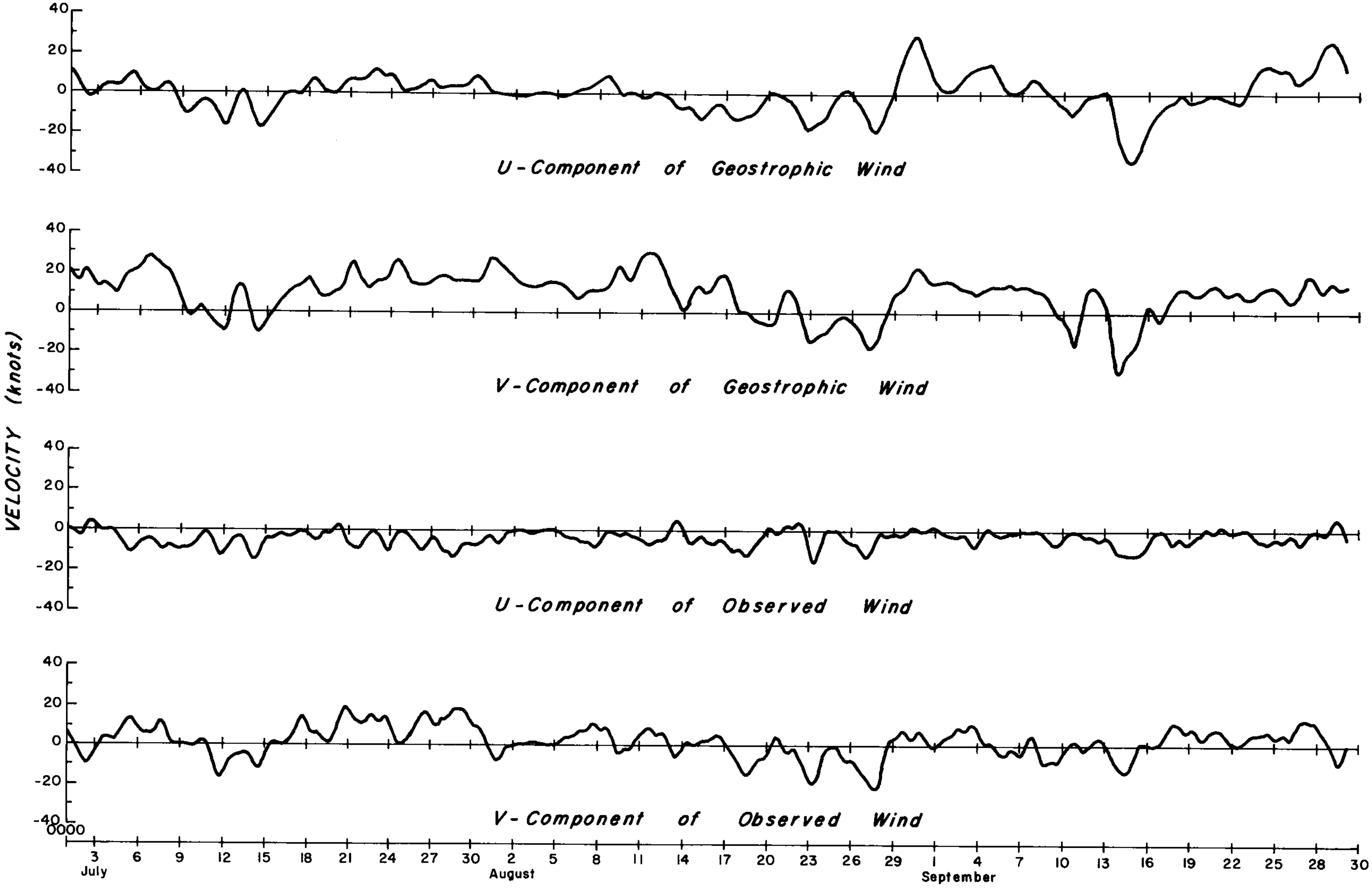

Figure 3-C. Tapered wind components (July - September 1968)

month are shown in Table III. The monthly correlation coefficients for certain combinations of components are plotted in Figure 4 and indicate that both components of the geostrophic wind are well correlated with the V-component of the observed wind.

It is interesting to note the effect of considering the second independent variable after the correlation between similar components such as geostrophic-U and observed-U had been determined. After correlating the preceding components the increase of the percentage of correlation due to considering the geostrophic-V was usually small. However, the increase of the percentage of correlation due to considering the geostrophic-U, after correlating the observed-V and geostrophic-V, was significant (P=0.95). This may indicate that the far offshore wind, represented approximately by the geostrophic wind, shifted to become more north-south at the coast. This study did not indicate the amount of change in direction nor the decrease in speed due to friction, but it did show such a change existed and that it was not the same for both wind components. Certainly some of the difference was due to the effect of topography in the vicinity of the Coast Guard's equipment for observing wind speed and direction (Figure 1).

The orographic features might increase the frictional effect on the surface wind thereby reducing speed and changing direction, or act as a barrier to the winds causing a change in direction and

Table III

Percentage of Correlation ($R^2 \times 100 = \%$)

	U_{obs} correlated with U_g	U_{obs} correlated with V_g	U_{obs} correlated with U_g & V_g	V_{obs} correlated with U_g	V_{obs} correlated with V_g	V_{obs} correlated with U_g & V_g
May 1967	38.29	24.44	38.31	58.66	42.43	59.27
Total*	32.45	15.40	32.47	57.03	27.98	57.09
April 1968	58.46	0.63	58.47	8.31	25.48	30.33
May	5.77	23.88	24.38	20.15	31.72	38.77
June	4.71	17.07	17.07	56.75	56.75	67.38
July	5.52	4.21	5.62	57.47	32.98	57.64
August	17.31	8.84	17.93	42.22	56.10	63.91
September	34.41	32.16	34.41	15.93	28.34	28.34
Total	9.57	0.43	11.05	19.33	34.90	36.92

*May 1, 1967 to June 18, 1967.

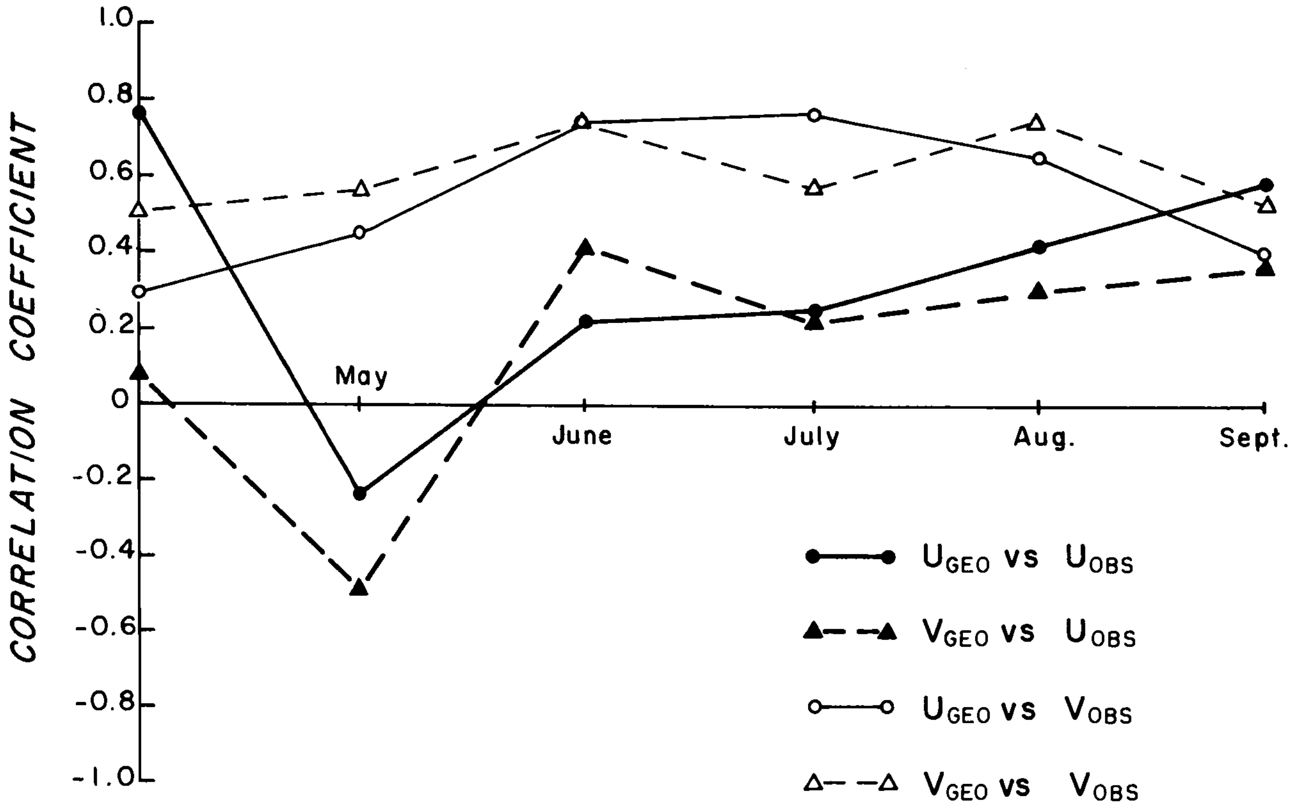

Figure 4. Correlations of wind components (1968).

increased turbulence. The most dominant features affecting the north-south wind component are Yaquina Head and the coastal mountains which act as a barrier to a northerly wind. A secondary factor is that the southerly winds move over water or wide beaches near the mouth of Yaquina Bay so that the reduction in northerly velocity causes the southerly velocity to appear to be increased in relative magnitude. The plots of wind data in Figures 3-A, 3-B, and 3-C show that if the observed and geostrophic wind components are compared, the northerly component of the observed wind appears to be reduced relative to the southerly component.

The monthly mean geostrophic wind components, calculated from the monthly mean barometric pressure charts, were compared to the monthly means of the six-hourly geostrophic wind components and the Coast Guard observed wind components. These monthly means, shown in Table II, indicate that the best agreement was between the two types of geostrophic winds, because they both represented smoothed large scale meteorological phenomena. Based on overall considerations, the six-hourly geostrophic winds were considered the best wind data for this study.

DESCRIPTIVE ANALYSIS OF MAJOR CHANGES IN TEMPERATURE AND WIND DATA

The results of a descriptive analysis of the major changes in sea water temperature associated with changes in geostrophic wind components are shown in Tables IV-A, IV-B and IV-C. It is evident that wind components from the north and east are associated with a decrease in temperature, and those from the south and west are associated with an increase. The change in wind components usually preceded the temperature change, but it is difficult to determine, by inspection, which wind components had the most effect on temperature.

A major difficulty in using temperature as an indicator of upwelling is that the thermal structure of the whole water column has a great effect on how the near-surface temperature will respond to the wind. The temperature profiles in Figure 5 show that the water column fluctuated from a nearly isothermal state to one strongly structured in temperature. Before the early stage of upwelling there was usually a well established thermocline at a moderate depth, but by the equilibrium stage the thermocline had moved toward the surface and the water column was nearly isothermal near shore. In the early stage the thermocline rose during an intensification of upwelling and fell during a relaxation. The shallow thermograph was within the range of rise and fall of the thermocline resulting in

Table IV-A

Changes in Temperature Associated with Changes in Wind Components

DB-5 1967

Period of Temperature Change			Period of Major Wind Change						Temperature Profile		
Beginning Date and Duration (Days)		Description	U_g Direction and Maximum Velocity	Beginning Date and Duration (Days)		V_g Direction and Maximum Velocity	Beginning Date and Duration (Days)		Date		Description
1910	May 7* (3)	Slightly Decreasing	East 7 kts.	0000	May 1 (8 1/2)	North 20 kts.	0000	May 1 (9 3/4)	0245	May 10	Small vertical temperature gradient. No thermocline.
1900	May 10 (2)	Increasing	West 13 kts.	1200	May 9 (4 1/3)	South 8 kts.	1800	May 10 (11/12)	1528	May 12	Thermocline from surface to 35 m.
1800	May 12 (13 3/4)	Decreasing	East 12 kts.	2000	May 13 (6 1/6)	North 17 kts.	1600	May 11 (8 1/3)			
1200	May 26 (2 1/2)	Slightly Increasing	West 3 kts.	0000	May 20 (4 5/8)	South 6 kts.	0000	May 20 (2 5/8)			
0000	May 29 (3/4)	Slightly Decreasing	East 14 kts.	1500	May 24 (1 3/4)	North 22 kts.	1500	May 22 (4 1/4)			
1800	May 29 (1 1/2)	Increasing	West 21 kts.	0900	May 26 (5 1/8)	South 21 kts.	2000	May 26 (3 5/12)			
0600	May 31 (3 1/2)	Decreasing	East 5 kts.	1200	May 31 (1 1/2)	North 27 kts.	0600	May 30 (3 1/6)	2140	June 2	Thermocline from surface to 10 m. Small vertical temperature gradient below 10 m.
1700	June 3*	End of Temperature Record		0000	June 2		1000	June 2			

* Tapered temperature record does not include this time. Description is made of untapered temperature record.

Table IV-B

Changes in Temperature Associated with Changes in Wind Components

DB-7 1967

Period of Temperature Change		Period of Major Wind Change				Temperature Profile	
Beginning Date and Duration (Days)	Description	U_g Direction and Maximum Velocity	Beginning Date and Duration (Days)	V_g Direction and Maximum Velocity	Beginning Date and Duration (Days)	Date	Description
2120 May 7* (2 7/8)	Slightly Decreasing	East 7 kts.	0000 May 1 (8 1/2)	North 20 kts.	0000 May 1 (9 3/4)	0153 May 10	Small vertical temperature gradient. No thermocline.
1800 May 10 (3)	Increasing	West 13 kts.	1200 May 9 (4 1/3)	South 8 kts.	1800 May 10 (11/12)	1645 May 12	Weak thermocline from surface to 20 m.
1800 May 13 (8 1/2)	Slightly Decreasing	East 12 kts.	2000 May 13 (6 1/6)	North 17 kts.	1600 May 11 (8 1/3)		
0600 May 22 (1)	Slightly Increasing	West 3 kts.	0000 May 20 (1 5/12)	South 6 kts.	0000 May 20 (1)		
0600 May 23 (1 1/4)	Slightly Decreasing	East 10 kts.	1000 May 21 (1 7/12)	North 23 kts.	0000 May 21 (1 5/8)		
1200 May 24 (3/4)	Slightly Increasing	West 1 kt.	0000 May 23 (1 5/8)	North Minimum 9 kts.	1500 May 22 (2 7/8)		
0600 May 25 (1)	Slightly Decreasing	East 14 kts.	1500 May 24 (1 3/4)	North 22 kts.	1200 May 25 (1 1/3)		
0600 May 26 (4 1/2)	Increasing	West 21 kts.	0900 May 26 (5 1/8)	South 11 kts.	2000 May 26 (3 5/12)		
1800 May 30 (4)	Decreasing	East 5 kts.	1200 May 31 (1 1/2)	North 27 kts.	0600 May 30 (3 1/6)	2222 June 2	Shallow thermocline, surface to 15 m. Small vertical temperature gradient below 15 m.
1800 June 3 (4 1/3)	Increasing	West 10 kts.	0000 June 2 (1 1/6)	North Minimum 2 kts.	1000 June 2 (1 5/6)		
0230 June 8*	End of Temperature Record		0400 June 3		0600 June 4		

* Tapered temperature record does not include this time. Description is made of untapered temperature record.

Table IV-C

Changes in Temperature Associated with Changes in Wind Components

DB-7 1968

Period of Temperature Change		Period of Major Wind Change				Temperature Profile	
Beginning Date and Duration (Days)	Description	U_g Direction and Maximum Velocity	Beginning Date and Duration (Days)	V_g Direction and Maximum Velocity	Beginning Date and Duration (Days)	Date	Description
2020 April 17* (32 1/6)	Slightly Decreasing	Variable West 20 kts. East 16 kts.	0000 April 17 (25)	North 24 kts.	2100 April 4 (38)	0720 April 19	Small vertical temperature gradient.
0000 May 20 (1 1/2)	Slightly Increasing	West 16 kts.	0000 May 12 (4 5/12)	South 14 kts.	2000 May 12 (4)		
1200 May 21 (3/4)	Slightly Decreasing	East 6 kts.	1000 May 16 (2 7/8)	North 10 kts.	2000 May 16 (2)		
0600 May 22 (2 1/4)	Increasing	West 13 kts.	0700 May 19 (2 3/4)	South 13 kts.	2000 May 18 (4 1/6)		
1200 May 24 (3/4)	Slightly Decreasing	East 10 kts.	1200 May 24 (11/12)	South Minimum 5 kts.	0000 May 23 (1 1/4)	0315 May 25	Thermocline from 10 m. to 50 m.
0600 May 25 (1/2)	Slightly Increasing	West 6 kts.	1000 May 25 (3/4)	South 8 kts.	0600 May 24 (7/8)		
1800 May 25 (1 1/4)	Decreasing	East 6 kts.	0400 May 26 (3/4)	North 13 kts.	0300 May 25 (1)		
0000 May 27 (3/4)	Slightly Increasing	West 13 kts.	2200 May 26 (1 1/3)	South 9 kts.	0400 May 26 (1 5/8)	1356 May 27	Strong thermocline from 10 m. to 30 m. Small vertical temperature gradient below 30 m.
1800 May 27 (1)	Decreasing	East 5 kts.	0600 May 28 (2 1/12)	North 23 kts.	1900 May 27 (2 1/2)		
1800 May 28	End of Temperature Record		0800 May 30		0800 May 30		

* Tapered temperature record does not include this time. Description is made of untapered temperature record.

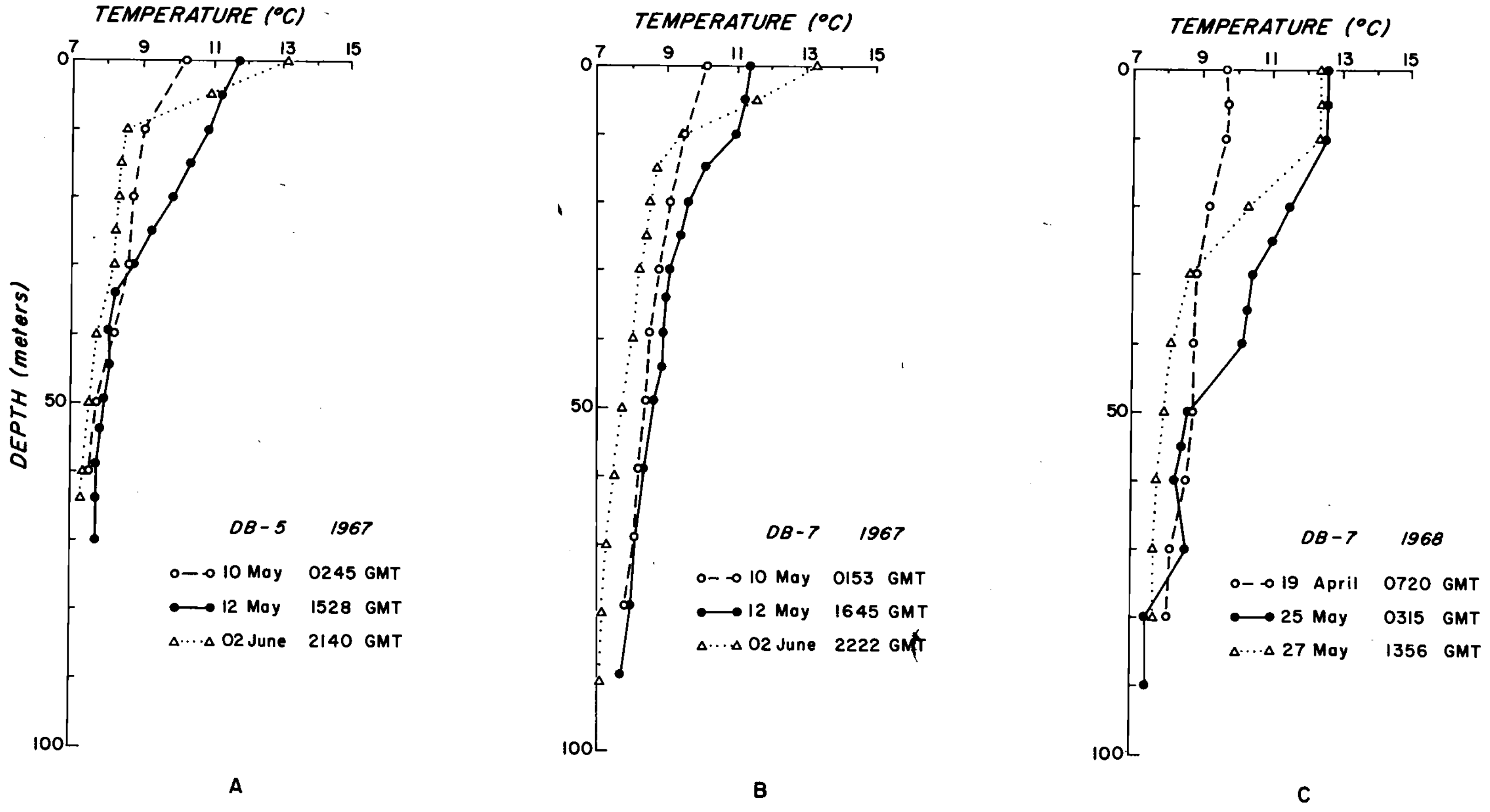

Figure 5. Vertical profiles of sea water temperature.

relatively large fluctuations in the thermograph record. It is interesting to note how rapidly the thermocline fluctuated in depth during May 1967. Temperature records indicated that the duration of temperature fluctuations differed from 1967 to 1968. In 1967 there were numerous fluctuations of a relatively short duration throughout the temperature record. The longest periods of monotonic temperature change in 1967 were noted when the temperature decreased over a period of approximately 14 days at DB-5 and 8-1/2 days at DB-7. The temperature record for 1968 at DB-7 indicated that upwelling had already commenced, and that the temperature decreased over a period of approximately 32 days. During the long decrease in temperature in 1968 the northerly component of the geostrophic wind persisted for 25 days with a relatively high velocity. No similar condition was observed in the 1967 winds. After the long period of gradual temperature decrease in 1968 there were numerous fluctuations of a much shorter duration which included a fluctuation with a greater increase in temperature than observed in 1967.

CORRELATION OF TEMPERATURE AND GEOSTROPHIC WINDS

The purpose of this study was to determine the effect of wind on temperatures observed off the Oregon coast during the early stage of upwelling. The length of time required for temperatures to reflect the effect of wind was not definitely known so the relationship between the preceding geostrophic winds and temperature was studied. Bourke (1969) suggested the major wind effect on temperature in Yaquina estuary was from the north-south or V-component of the observed wind averaged over a period of four days prior to the time of the temperature observation. The east-west or U-component of the observed wind was eliminated from his study because it was observed that the V-component was much greater in magnitude and that there was almost no temperature dependence on the U-component. Weighted averages of V-components were used to determine whether the correlation between temperature and these components could be improved. An increase in correlation was noticed when the third day was most heavily weighted. However, the actual effect of the V-component three days prior to the time of the temperature observation was not determined because a four day average of the V-components was used in the regression analysis.

Wind data used in this study were the six-hourly tapered geostrophic winds. Observed temperature was correlated with each component of the wind for each of the preceding twelve-hourly wind values. In other words, temperature was correlated with all of the wind components, which were successively lagged by 12 hour intervals, to a total of 7-1/2 days. A regression analysis was later made on temperature and the lagged wind components.

The correlations between the observed sea water temperature and geostrophic wind components for each successive time lag are plotted in Figures 6-A, 6-B and 6-C. The meteorological convention for a wind system was used. A positive U-component, $+U_g$, was a component from the east and a $-U_g$ was from the west. Similarly, a positive V-component, $+V_g$, was a component from the north and a $-V_g$ was from the south. The positive components were expected to cause upwelling, and the negative components were expected to reduce upwelling. In this case a negative correlation of U_g and temperature indicated that a $+U_g$ was related to a decrease in temperature and a $-U_g$ was related to an increase in temperature. The same principle applied to the correlation of V_g and temperature.

Figures 6-A, 6-B, and 6-C show the correlation that each wind component, if considered separately, had with temperature. In 1967 at DB-7, U_g had a negative correlation with temperature for lags of 0 to 4 days and V_g had a negative correlation with temperature for lags of 0 to 3-1/2 days. The correlations of temperature with U_g and V_g were positive after these lags which was indicative of a

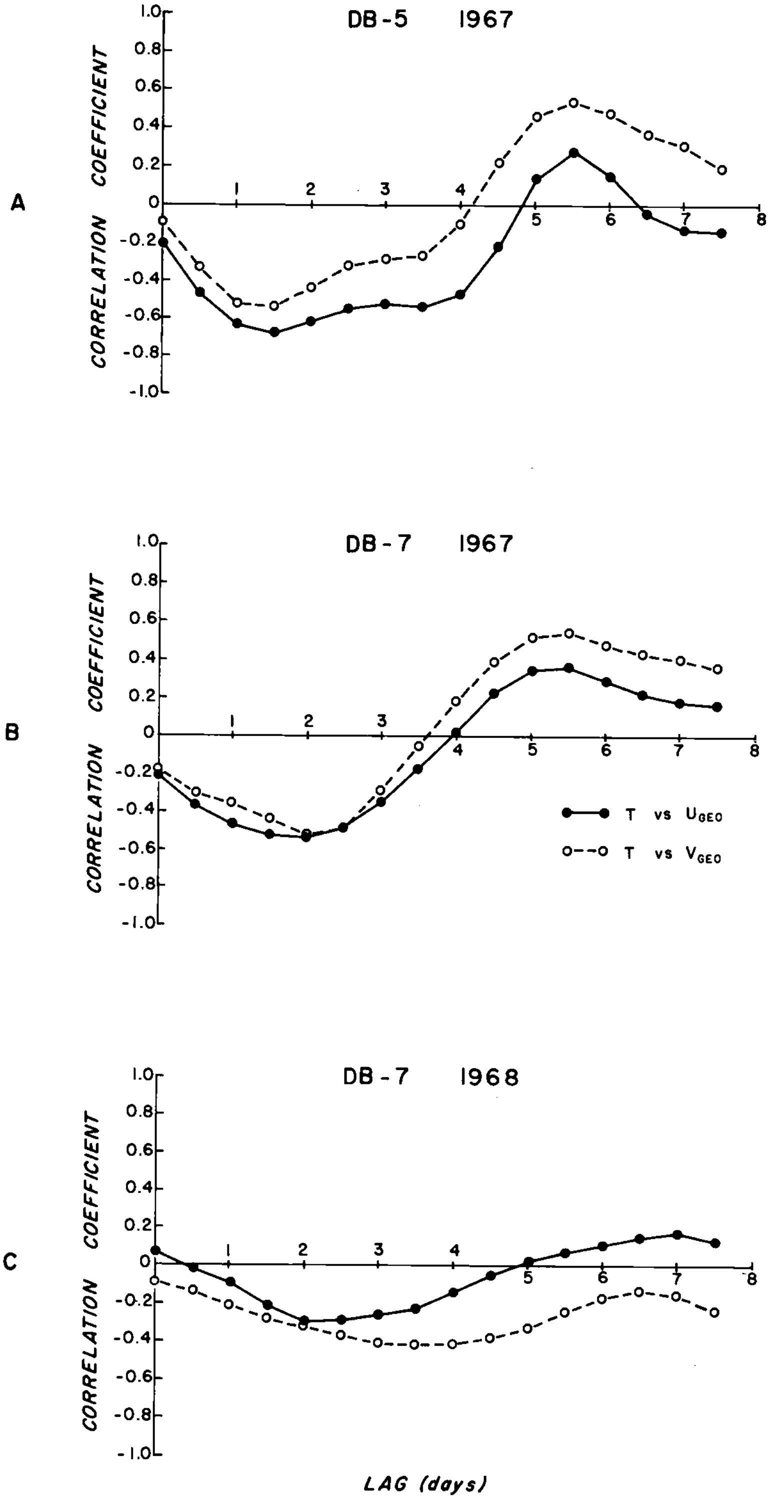

Figure 6. Correlations of temperature with lagged wind components.

periodicity in the wind of about eight days. Only half a cycle was considered in the regression analysis in order to study the effect of only one atmospheric pressure system on the temperature.

The correlations of temperature and wind components at DB-5 in 1967 also exhibited a change from negative to positive which reflected an oscillation with a period of approximately eight days. The correlations for DB-7 in 1968 did not show as distinct an oscillation but still suggested an oscillation of wind components with a period greater than seven days.

The relative magnitudes of U_g and V_g were considered when studying the relationship of correlations of temperature with V_g, at each lag. This was necessary because the magnitude of the wind component is important when computing the transport of surface waters. An estimate of the relative magnitudes of U_g and V_g are the monthly means which are shown in Table II.

The temperatures at DB-5 and DB-7 were recorded throughout most of May 1967 during which the mean U_g was 0.23 knots from the west and the mean V_g was 9.26 knots from the north. The plots of wind components in Figure 3-A also show V_g to be much greater in magnitude than U_g except for two periods when the component from the west was large relative to the component from the south.

The temperature at DB-7 was recorded during the latter portion of April and most of May 1968. The means for U_g were 6.07 and

2.87 knots from the west for April and May respectively, and the means for V_g were 8.67 and 4.27 knots from the north for April and May respectively. This indicated that while V_g had the larger magnitude, the magnitude of U_g was substantial. The plots of the wind components in Figures 3-B and 3-C show that the V-component was generally greater in magnitude than the U-component, but the U-component approached the size of the V-component in 1968.

The effect of each wind component, lagged in time, with respect to the observed temperature was studied by considering relative magnitudes of the wind components and the correlation of temperature with each component. In 1967 at DB-5 there was a striking difference in the correlations of temperature with U_g and temperature with V_g, at each lag. The U-component was consistently more negatively correlated with temperature than the V-component, with a maximum correlation of -0.68 at a lag of 1-1/2 days. However, the correlation of temperature with V_g was also significant, with a maximum correlation of -0.53 at a lag of 1-1/2 days. Since the magnitude of V_g was greater, the V-component probably had more effect on the observed temperature than did the U-component.

In 1967 at DB-7, U_g and V_g showed approximately the same negative correlation with temperature at each lag up to a lag of 3-1/2 days. The maximum correlation of -0.54 was at a lag of 2 days for both components. However, the magnitude of V_g was greater, so the V-component with a lag of 2 days probably affected the temperature more than did the U-component.

In 1968 the relationship of the correlations of temperature and each lagged component was not the same as for 1967 at DB-7. At each lag, V_g consistently correlated more negatively with temperature than U_g. The V-component's maximum correlation was -0.42 at a lag of 3-1/2 days compared with a lag of 2 days in 1967. The U-component had a maximum correlation of -0.30 at a lag of 2 days. This was the period when the magnitude of U_g was nearer that of V_g. However, the result was probably similar to 1967 in that V_g had a greater effect on temperature.

The V-component explained only a part of the effect of geostrophic wind on temperature. Even though U_g was smaller in magnitude than V_g, temperature and U_g at certain lags were well correlated. This indicated the U-component also had an effect on the temperature and suggested a coastal modification of the wind system. The correlation of U_g and V_g with the components of the winds observed at Newport showed both U_g and V_g were well correlated with the V-component of the observed winds, but the correlation with the U-component of the observed winds was poor. This indicated that U_g may contribute to the magnitude of the longshore surface wind.

There are four combinations of U_g and V_g, representing the four quadrants, and each combination is studied to note how U_g and

V_g affect the longshore surface wind. A combination of $+V_g$ (north) and $+U_g$ (east) will produce an offshore wind from the northeast. It is important to note that if the theory given by Hidaka (1954) is applied to this area, a wind from about 025^o would produce the most intense upwelling. The correlations of temperature with U_g and V_g, considering the lag restrictions imposed due to passing storms, may indicate that a wind from the northeast quadrant does decrease the water temperature. The correlations, however, do not indicate the magnitude of the effect.

There are few combinations of a $-V_g$ and a $+U_g$ shown in Figures 3-A, 3-B and 3-C. In general, a southeasterly wind with a duration of a day or longer blowing far from shore is not observed. This is because lows usually travel eastward north of the area studied causing mostly southwesterly and northwesterly winds.

The orographic features of the coast from Yaquina Head to Depoe Bay have a major effect on the combination of $-V_g$ (south) and $-U_g$ (west) (Burdwell, 1969). The contours in Figure 7 show that the coastal mountain range is very close to the ocean boundary and that the 250 foot contour approximates a broad circular arc. The 500 foot contour parallels most of this arc, and there are elevations inside the arc that exceed 750 feet. The effect of a circular mountain barrier was studied by Dickey (1961) where he approximated the Brooks Range in Alaska with a cylindrical model. He drew a

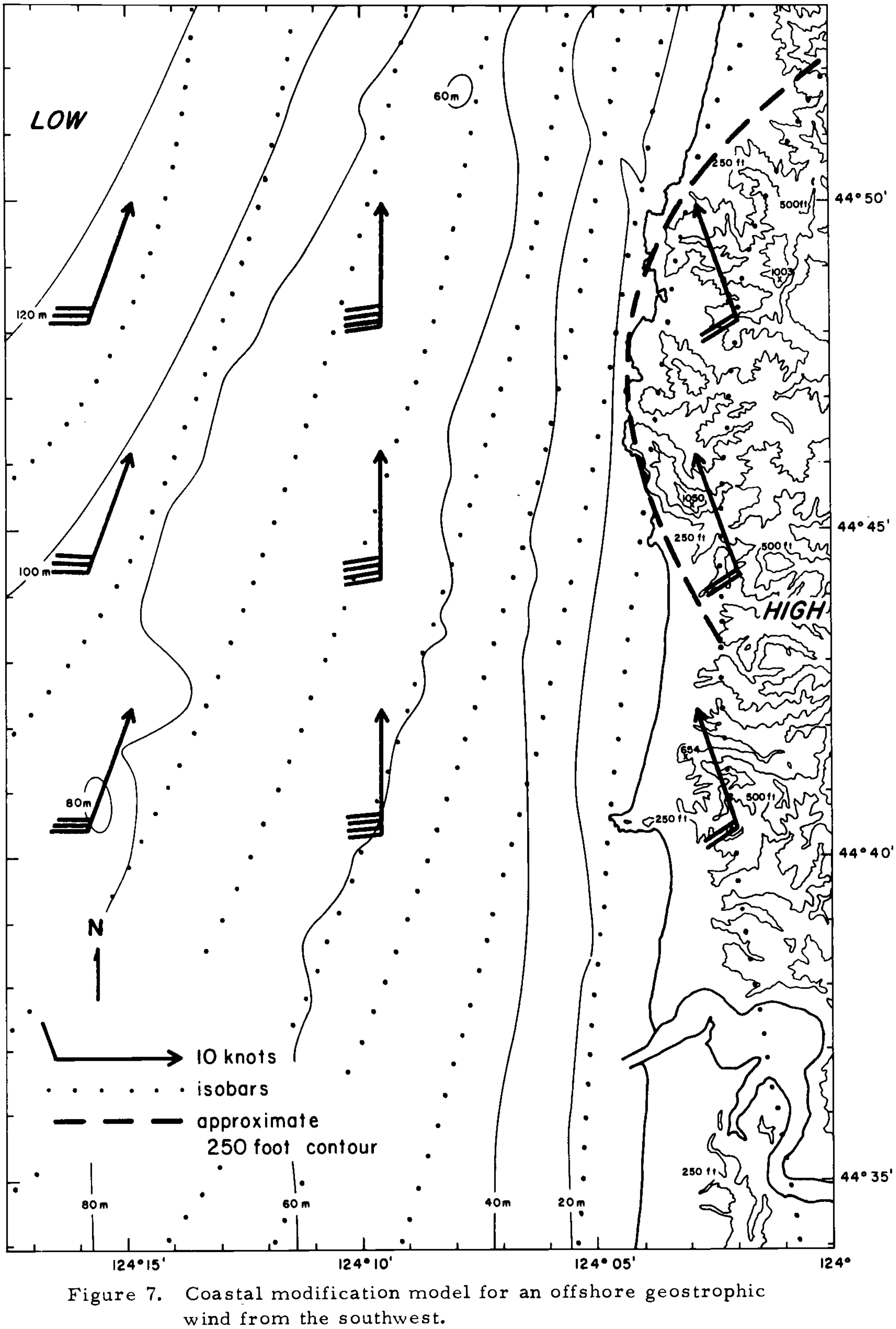

Figure 7. Coastal modification model for an offshore geostrophic wind from the southwest.

circular arc along the average 2,000 foot contour with a radius of 170 miles. This was a greater barrier to the winds than the coastal mountains near Depoe Bay, Oregon, which must be considered when applying his results to the present study. The barrier near Depoe Bay could affect the onshore winds but the magnitude of the effect would probably be smaller than that for the Brooks Range. Dickey determined that the speed and direction of the surface wind near such a barrier was dependent on the direction of the basic wind flow and the distance from the barrier. The geostrophic winds were converted to basic wind flow by considering the frictional forces and cross isobaric flow. He calculated the maximum velocity of the surface wind to be 1.9 times that of the basic wind flow.

Figure 7 illustrates the effect of the barrier on a wind from the southwest considering only the effect of friction on the onshore wind and not the intensification derived with the cylindrical model by Dickey. In this study it is assumed that the frictional effect of the barrier is much greater than that of the sea surface. A geostrophic wind of 30 knots from a direction of 200^{o} is used as an example. This represents a $-U_g$ of 10 knots and a $-V_g$ of 28 knots. The speed and direction of the geostrophic wind is dependent on the pressure gradient and the Coriolis force as previously shown in the geostrophic relationship. These two forces are in balance producing a wind that is approximately parallel to the isobars. As the geostrophic wind approaches the barrier the effect of friction reduces the velocity which results in a reduction of the Coriolis force. Since the Coriolis force is no longer in balance with the pressure gradient force the resultant wind is shifted to the left if looking downwind (In the northern hemisphere). The example in Figure 7 shows that the wind over adjacent land has changed direction and is now from 160°, and the speed is reduced to 20 knots. These values are not exact but are representative of wind observations in this area (Burdwell, 1969).

The offshore wind and wind over adjacent land are convergent in a zone relatively near shore. This conforms to a classical case of convergence in which winds in the convergence zone are intensified. In the illustrative example the winds of this zone are intensified to 40 knots and are from 180° which represents a longshore wind. This example was based on a technical report on the use of weather maps by mariners prepared by the World Meteorological Organization (1966). Intensification due to the somewhat cylindrical barrier, as discussed by Dickey (1961), might also produce very strong longshore winds from the south.

The effect of the barrier on the combination of a $+V_g$ (north) and a $-U_g$ (west), representing a northwesterly wind, is not the same as in the previous case. The speed is reduced by friction as the northwesterly geostrophic wind approaches the barrier, and the direction is again shifted to the left when looking downwind. This effect is shown in Figure 8 and illustrates a reduction of the longshore flow instead of an intensification. The barrier may force the winds to move somewhat longshore, but there is no definite convergence zone and friction reduces the velocity. The wind near shore, where the temperature is observed, may exhibit a weaker V-component but friction will mainly affect the U-component. This implies that part of the negative correlation of temperature and these components is because the $-U_g$ which is reduced by friction, does not have a significant effect on temperature whereas the $+V_g$ causes an offshore transport of surface waters resulting in a decrease in temperature.

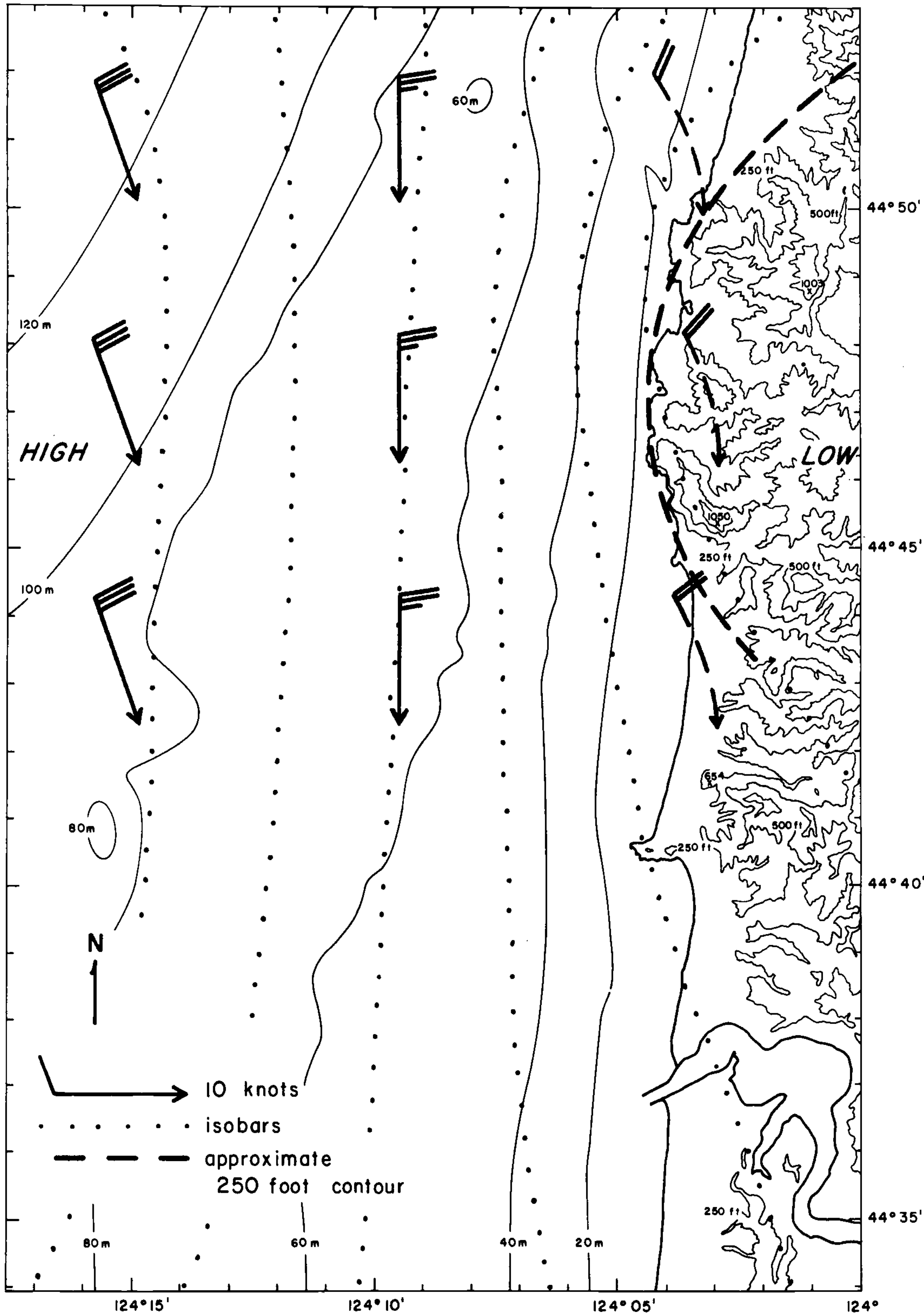

Figure 8. Coastal modification model for an offshore geostrophic wind from the northwest.

REGRESSION ANALYSIS OF TEMPERATURE ON GEOSTROPHIC WINDS

A regression analysis was made on each time series with temperature as the dependent variable and the geostrophic wind components at each lag as independent variables. The technique used was a stepwise multiple linear regression analysis, a library program of the Oregon State University Computer Center. A general discussion of this technique was prepared by Efroymsen (1960) and explained in a text by Draper and Smith (1966).

The linear model used in this analysis was assumed to be:

$$T = \beta_o + \sum_{t=0}^{180} \alpha_{-t} U_{-t} + \sum_{t=0}^{180} \gamma_{-t} V_{-t} + \epsilon$$

$t = 0, 12, 24, \ldots, 180$ hours

where:
- T = Temperature (^oC)
- β_o = True mean temperature with no wind effect
- U, V = Components of geostrophic wind (knots) at time $-t$
- α_{-t} = Rate of change of temperature at $t=0$ with change in U-component at time $-t$
- γ_{-t} = Rate of change of temperature at $t=0$ with change in V-component at time $-t$
- ϵ = Random error, NID $(0, \sigma^2)$

The estimated regression of temperature on all of the lagged components of the geostrophic wind was:

$$\hat{T} = b_o + \sum_{t=0}^{180} a_{-t} U_{-t} + \sum_{t=0}^{180} g_{-t} V_{-t}$$

$t = 0, 12, 24, \ldots, 180$ hours

where: $\hat{T}$ = Estimate of the temperature (^{o}C)
b_o, a_{-t}, g_{-t} = Estimates of β_o, α_{-t} and γ_{-t} respectively.

The sample residual was the difference between the observed temperature and the estimate of temperature. Assuming that the sample residuals had a zero mean, constant variance and were not correlated allowed the use of the method of least squares to estimate the regression coefficients. The theoretical coefficients β_o, α_{-t} and γ_{-t} were estimated by the coefficients which minimized the sum of the squares of the sample residuals.

It was not expected that all of the independent variables would have a significant effect on the dependent variable. Therefore, a stepwise procedure was used to determine which variables would be retained in the regression equation. In the first step the independent variable chosen was the lagged wind component which was most highly correlated with observed temperature. After the regression equation was determined a measure of the proportion of variation in temperature, which was accounted for by the linear relationship between the observed temperature and the most highly correlated wind component at a certain lag, indicated the effect of considering the first independent variable. This proportion, expressed as a percentage, was called R-square.

If a second step was required the sample residuals were correlated with the remaining wind components at each lag, and the most highly correlated component was used as the second independent variable in the new regression equation. The proportion of variation in temperature which was accounted for by this linear relationship indicated the additional effect of considering the second independent variable. This procedure continued until all the independent variables were considered. However, relatively few variables were usually required to define the linear relationship between the observed temperature and geostrophic wind components. There were statistics calculated for each step that indicated when the last significant (P=0. 95) coefficient was determined.

The stepwise regression procedure was applied to the three time series in which the relationships of temperature to the lagged U_g and V_g were considered. The maximum lag was limited to a half period of four days in all three time series because the periodicity of the winds in 1967 was approximately eight days. The period was approximately the same in 1968 and reduction to four days allowed a simple comparison of the three regression equations. The variables in each equation were chosen on the basis of statistical parameters computed in the stepwise procedure that indicated which independent variables had a significant linear relationship with the dependent variable. The regression equations and the R-square values were determined for each time series and are presented in Table V.

Table V

Regression of Temperature on Lagged Geostrophic Wind Components

Location Code	Regression Equation	R-square
DB-5 1967	$\hat{T} = 8.6438 - 0.0341U_{-1\ 1/2} - 0.0360U_{-3\ 1/2} + 0.0104V_{-3}$	61.56%
DB-7 1967	$\hat{T} = 9.4003 - 0.0143U_{-2} - 0.0146V_{-1/2} - 0.0140V_{-2\ 1/2}$	42.17%
DB-7 1968	$\hat{T} = 8.5395 - 0.0130U_{0} - 0.0316U_{-2} - 0.0174U_{-4} - 0.0172V_{-3\ 1/2}$	29.76%

A comparison of these equations indicated some basic similarities. In all three cases there was a term representing the effect of the V-component 2-1/2 to 3-1/2 days prior to the temperature observation. This agreed with the results of the study by Bourke (1969) in which he tested various weighting schemes on the data. He determined that the minimum variance from the regression line occurred whenever the V-component of the observed wind on the third day preceding the temperature measurement was weighted the heaviest. It was interesting to note that V_g with a lag of three days had a positive regression coefficient in the regression equation for DB-5 in 1967. However, the correlation between temperature and this V_g was negative as shown in Figure 6-A. In the stepwise procedure this variable was the least significant of the three independent variables which indicated that the two U-components had a greater effect on temperature than would be expected.

The statistical tests indicated that the effect of U_g with a lag of 1-1/2 days at DB-5 and a lag of 2 days at DB-7 was significant (P=0.95). In 1967 at DB-5 there was another significant U-component which had a lag of 3-1/2 days. Since both of the U-components in the regression equation for DB-5 were more significant than the V-component, the inference was that U_g had a stronger effect on temperature than V_g when five nautical miles from shore. The R-square term indicated that 61.56% of the variation in temperature had been accounted for by the linear relationship between temperature and these three independent variables. This was a higher percentage than was explained by the other two regression equations.

One U-component and two V-components had a significant effect on temperature at DB-7 in 1967. It is suggested that the V_g with a lag of 1/2 day caused an additional change in the temperature. The regression equation for DB-7 in 1968 contained four significant independent variables and it has previously been shown that U_g with a lag of 2 days and V_g with a lag of 3-1/2 days in this equation were similar to independent variables in the other two regression equations. In 1968 U_g with a lag of 4 days was similar to a U_g with a lag of 3-1/2 days in 1967 at DB-5, and there was an additional U_g showing an effect at the same time as the temperature observation. Perhaps this indicated that the conditions which affected DB-5 in 1967 were effective as far offshore as DB-7 in 1968. This was not definite because the regression equation for DB-7 in 1968 only explained 29.76% of the variation in temperature.

CONCLUSIONS

The purpose of this study was to determine the effect of wind on temperature observed during the early stage of upwelling. The available wind data were the six-hourly geostrophic winds and winds observed every four hours by the U.S. Coast Guard at Newport, Oregon. The six-hourly geostrophic winds provided the best estimate of winds that affected the near surface temperature five and seven nautical miles offshore from Depoe Bay. The partial correlation of all the wind components showed the V-component of the observed wind was stronger and the U-component was weaker than the corresponding geostrophic wind components. It was indicated that both components of the geostrophic wind were strongly related to the actual longshore surface wind.

A descriptive analysis indicated that temperature was affected by wind components existing prior to the time of the temperature observation, and that temperature response varied from year to year and with distance offshore. The correlations of temperature with geostrophic wind components lagged in time indicated that temperature was decreased by north and east components and was increased by south and west components. A model was presented that would explain the effect on the temperature by each combination of a U and V component. The orographic effect of the coastal mountain range, considered in this model, indicated a process that was producing longshore winds from both components of the geostrophic winds. The combination of a south and west component would cause an intensification of the longshore surface wind from a southerly direction. North and west components would combine to produce a longshore surface wind from a northerly direction but the speed will be reduced due to friction.

The regression analysis indicated that the V-component 2-1/2 to 3-1/2 days prior to the time of the temperature observation had the most significant effect on the temperature. There were also indications that the U-component of the geostrophic wind was definitely affecting the temperature. The percentage of variation in temperature, which was accounted for by the linear relationship between the temperature and the significant wind components, indicated that there may have been factors other than wind affecting the near surface temperature during the early stage of upwelling.

In the study, sea water temperature was more frequently sampled and with more precision than the wind. Better wind data may explain more of the variations in temperature, and a statistical study might then determine a model to predict sea water temperature in the early stage of coastal upwelling based on a knowledge of the past winds.

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APPENDICES

APPENDIX I

REVIEW OF COASTAL UPWELLING THEORY

Ekman's classic paper of 1905 provided the first physical explanation of upwelling although it had been observed and described much earlier. He assumed steady state conditions in a uniform and homogeneous ocean of infinite extent, with a steady and uniform wind field at the sea-air interface. Due to the effect of the earth's rotation and wind stress, the net transport of surface water was 90^{o} to the right of the wind in the northern hemisphere. In the Cartesian coordinate system the equations for wind induced mass transport are:

$$M_x = \frac{\tau_y}{f} \qquad \text{and } M_y = \frac{-\tau_x}{f} \quad \text{given: } f = 2\,\Omega \sin\phi$$

and the wind stress on the sea surface can be estimated from:

$$\tau_x = \rho_{air}\, C_D V|\mathcal{V}| \quad \text{and } \tau_y = \rho_{air}\, C_D U|\mathcal{V}|$$

where:

M_x = mass transport in the X direction
M_y = mass transport in the Y direction
f = the Coriolis parameter
τ_x = wind stress in the X direction
τ_y = wind stress in the Y direction
ρ_{air} = density of air
C_D = drag coefficient of the wind
$\mathcal{V}$ = speed of the surface wind
U = velocity of the east-west component of the surface wind
V = velocity of the north-south component of the surface wind
Ω = angular velocity of the earth
ϕ = latitude

Therefore, a northerly wind parallel to the Oregon coast would transport the near surface water offshore. Ekman determined a depth of frictional influence which is of the order of 100 meters (Smith, 1964). This offshore transport requires a replenishment by water from deeper layers.

Sverdrup (1938, Sverdrup and Fleming, 1941) applied the Ekman theory to upwelling off southern California which resulted in a dynamic interpretation of coastal upwelling. Smith (1968) refers to this as the Ekman-Sverdrup Model. Sverdrup estimated vertical velocities from the displacement of isolines of physical properties over an interval of time. Assuming that the vertical velocity is zero at the sea surface and the coastline is at x = 0, Sverdrup showed that at a distance from the coast, x = -L, where the effects of the boundaries are negligible, the transport away from the coast is the Ekman transport.

$$M_x = \int_0^{-L} \rho_{sw} W_{-H} dx = \frac{\tau_y}{f}$$

where: ρ_{sw} = density of sea water
W = water velocity in the Z direction
H = depth of the wind induced current

Sverdrup suggested that a convection current flowed downwind parallel to the coast as a result of the strong horizontal density gradient at the offshore boundary between upwelled water and the less dense wind transported water. He argued that the stress in the top layer of the convection current would reduce the Ekman transport in this region.

Hidaka (1954) developed a steady state theory of upwelling in a homogeneous ocean. In the model, the wind is confined to an area of specified width parallel to a straight coast. He assumed no variation in the longshore wind direction, constant eddy viscosities, and horizontal velocities to be negligible at the coastline, bottom and at great distances from the coast. At the surface the shearing stress equaled the wind stress. He determined that the ratio of the vertical to offshore velocity equaled the square root of the ratio of the vertical to horizontal coefficients of eddy viscosity. This result is questionable because the values determined for the coefficients of eddy viscosity are uncertain. He also considered the case where the wind does not blow parallel to the coast and observed that when there was an offshore wind at an angle of 21.5° with the coast, from a northerly direction, the upwelling was most intense.

Yoshida (1955) considered upwelling in a two layer model, and assumed that the ocean responded quasi-isostatically in the coastal region with no variations in conditions in the direction parallel to the coastline. He examined the vertical velocity at the base of the upper layer in the early stages of upwelling and during the equilibrium state, assuming that lateral mixing had a greater effect than local accelerations. The vertical velocity during early upwelling becomes:

$$W_{-h} = \frac{-k}{\rho f} \tau_y e^{kx} \quad \text{given: } k = f\left(\frac{gh\Delta\rho}{\rho}\right)^{-1/2}$$

where: h = depth of the upper layer
ρ = density of the upper layer
$\Delta\rho$ = density difference between the upper and lower layers
g = gravitational acceleration

It has been shown by Smith (1967) that if the assumption of a uniform wind over the water is applied to the vertical velocity distribution obtained by Yoshida, the result is an offshore transport identical to the Ekman transport normal to the coast. This is important because the Ekman transport was derived for the equilibrium state of the ocean assuming homogeneity, infinite dimensions and constant wind stress. Yoshida, however, specified that his two-layer model theory of vertical velocity distribution for the coastal region was applicable during the period before equilibrium was attained. This led Smith (1967) to state "... the Ekman transport would appear to give a valid estimate of the offshore transport in the early (non-equilibrium) stage of coastal upwelling".

APPENDIX II

CHARACTERISTICS OF THE COSINE-LANCZOS FILTER-TAPER

Sampling Periods: ΔT = 10 min. (Ten minute temperature samples)
ΔT = 1 hour (Representative hourly temperature samples)

Taper Weighting Functions

m	f(m)	m	f(m)	m	f(m)
0	1.000				
1	0.989	21	-0.122	41	0.0284
2	0.970	22	-0.135	42	0.0263
3	0.943	23	-0.141	43	0.0235
4	0.900	24	-0.141	44	0.0200
5	0.847	25	-0.136	45	0.0163
6	0.787	26	-0.126	46	0.0128
7	0.719	27	-0.113	47	0.00935
8	0.648	28	-0.0981	48	0.00634
9	0.570	29	-0.0810	49	0.00385
10	0.490	30	-0.0643	50	0.00189
11	0.409	31	-0.0471	51	0.000452
12	0.331	32	-0.0309	52	-0.000468
13	0.256	33	-0.0163	53	-0.000961
14	0.184	34	-0.00329	54	-0.00112
15	0.118	35	0.00747	55	-0.00102
16	0.0589	36	0.0162	56	-0.000793
17	0.00676	37	0.0225	57	-0.000487
18	-0.0374	38	0.0266	58	-0.000219
19	-0.0738	39	0.0288	59	-0.0000452
20	-0.102	40	0.0295	60	0.0

Normalization Factor = 17.06

Taper Response Functions

(At one-hour sampling rate evaluated for the following periods)

T(hours)	$F(\sigma)$	T(hours)	$F(\sigma)$	T(hours)	$F(\sigma)$
100	1.004279	40	0.737569	28	0.154763
80	1.006590	36	0.583084	24	0.009995
60	0.989450	32	0.376798	12	-0.000028

APPENDIX III

PROCEDURE FOR CORRECTING RAW WIND DATA*

1. Convert raw wind data into U and V components.
2. Plot U and V as a function of time.
3. Visually mark any peaks that appear to be erroneous.
4. Subject the components to the error detection program.
   a. Error detect level is set at 2% of the data.
   b. Set original C-test at 10 knots for the first 25 values. Then the program will recompute the C-test for the remaining values. (It is easier to do this than convert the existing program.)
5. In the output mark all clusters of predicted values caused by adjacent errors, and decide which value was the major cause. This predicted value may be used but no adjacent ones may be used because they have been calculated using the major erroneous value as a data point.
6. Check all suspected values against the original data to detect card punching errors.
7. For geostrophic winds, check the suspected values against the original facsimile charts.
   a. Changes can only be recorded in polar coordinates which must be converted to U and V components.
   b. If the charts strongly verify readings as being correct,

I believe they should be kept in spite of the values predicted by the error detection.

8. If the charts are vague or do not exist, then look at the predicted values. Here there is still the option to choose whether or not to accept the predicted value.

9. If the predicted value is used, note the data on the corresponding vector plot to see if it should be corrected.

10. A value may appear to be erroneous, but it was not picked up by the error detection. Caution should be exercised in choosing a new value. It may be done by graphic means but the results of arbitrary smoothing are questionable.

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*For use with the error detection program shown in Data Report no. 30, Appendices I and II (Oregon, 1968).

APPENDIX IV

CHARACTERISTICS OF TAPERS USED TO SMOOTH WIND DATA

Sampling Period: ΔT = 4 hours (Wind observed every four hours by the Coast Guard)

Taper Weighting Functions

m	f(m)
0	1.0
1	0.88636
2	0.59833
3	0.27388

Normalization Factor = 4.51714

Taper Response Functions

T(hours)	F(σ)	T(hours)	F(σ)	T(hours)	F(σ)
100	0.92204	40	0.58326	28	0.29664
80	0.88021	36	0.50738	24	0.16388
60	0.79463	32	0.41313	12	0.013962

Sampling Period: ΔT = 6 hours (Six-hourly geostrophic wind)

Taper Weighting Functions

m	f(m)
0	1.0
1	0.79173
2	0.34568

Normalization Factor = 3.27482

Taper Response Functions

T(hours)	F(σ)	T(hours)	F(σ)	T(hours)	F(σ)
100	0.90821	40	0.52433	28	0.22438
80	0.86028	36	0.44157	24	0.094246
60	0.76178	32	0.34111	12	0.032950