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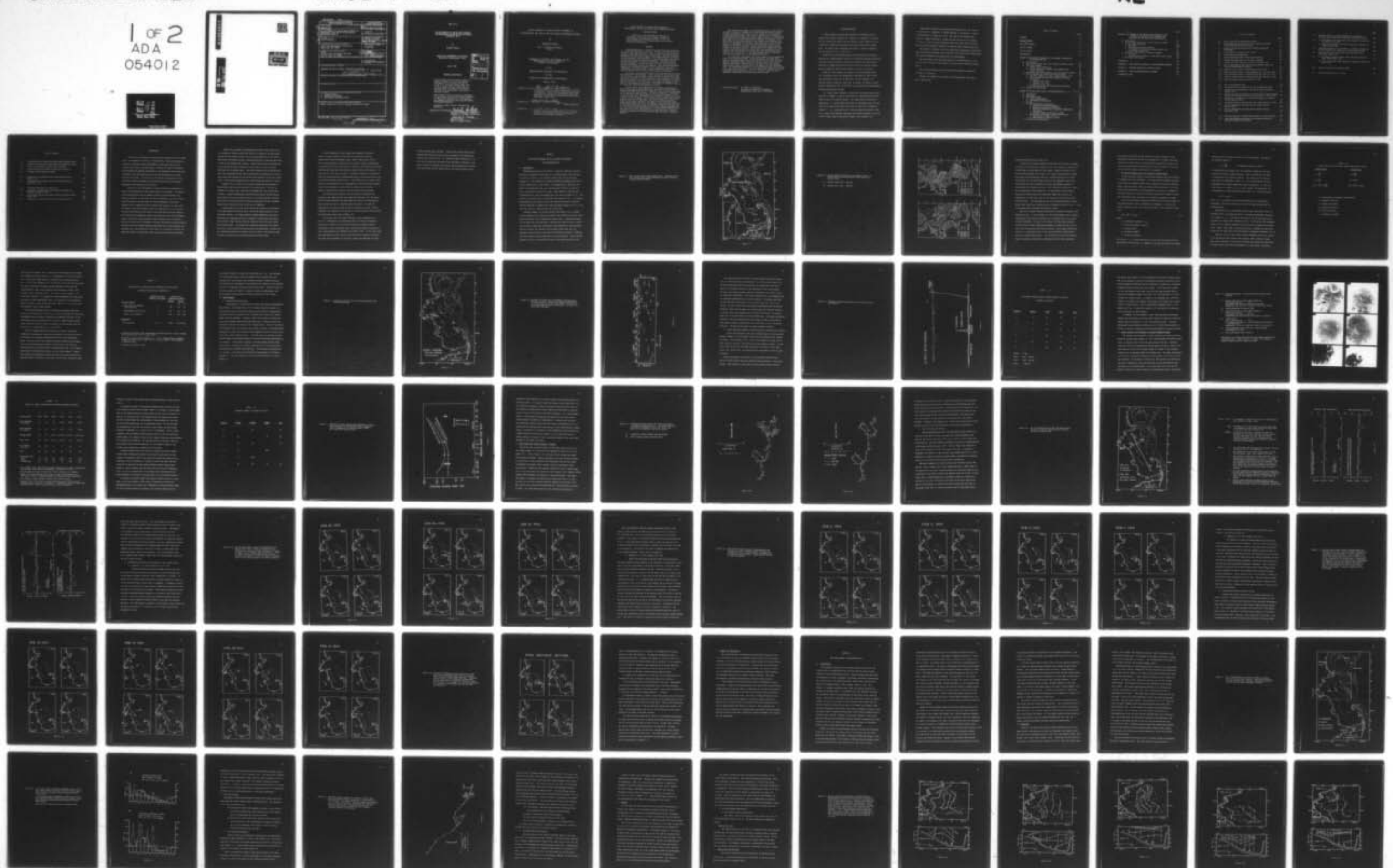

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by

Bradford Butman

WOODS HOLE OCEANOGRAPHIC INSTITUTION
Woods Hole, Massachusetts 02543

April 1978

DOCTORAL DISSERTATION

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ON THE DYNAMICS OF SHALLOW WATER CURRENTS IN MASSACHUSETTS BAY AND ON THE NEW ENGLAND CONTINENTAL SHELF

by

BRADFORD BUTMAN

A.B., Cornell University
(1969)

SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY

at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

and the

WOODS HOLE OCEANOGRAPHIC INSTITUTION

April, 1975

Signature of Author... Bradford Butman ...
Joint Program in Oceanography, Massachusetts Institute of Technology - Woods Hole Oceanographic Institution, and Department of Earth and Planetary Sciences, and Department of Meteorology, Massachusetts Institute of Technology, April, 1971

Certified by... Robert C Beardsley ...
Thesis Supervisor

Accepted by...
Chairman, Joint Oceanography Committee in the Earth Sciences, Massachusetts Institute of Technology - Woods Hole Oceanographic Institution

ON THE DYNAMICS OF SHALLOW WATER CURRENTS IN MASSACHUSETTS BAY AND ON THE NEW ENGLAND CONTINENTAL SHELF

BRADFORD BUTMAN

Submitted to the Massachusetts Institute of Technology — Woods Hole Oceanographic Institution Joint Program in Oceanography on April 7, 1975, in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

ABSTRACT

Massachusetts Bay is a coastal Bay 100 km long and 40 km wide located in the western Gulf of Maine. The Bay is closed by land to the north, west and south, but is open to the Gulf to the east; the opening is partially blocked by a shallow bank. The bottom sediment distribution in the Bay is complex; fine grained material is found in the deep basin, sand and gravel on the shallow bank, and mixtures of sand, gravel and fine material nearshore. Richardson current meters were moored 1 m from the bottom over a one year period at several locations in the Bay to study the bottom currents and the equilibrium between current and sediments. The current measurements suggest that the bottom sediments can be expected to move only occasionally in certain areas. The maximum bottom speeds are principally determined by the strong tidal currents in the basin.

In winter, the near bottom currents are dominated by wind stress associated with strong storms. Bottom currents in the shallow areas are generally in the direction of the wind while currents in the deep portion of the basin are often opposite to the direction of the wind. Sea surface setup in the direction of the wind is observed, as well as absolute changes in sea level as the Bay adjusts to changes in the level of the adjacent Gulf of Maine. Adjustment of the bottom currents to wind events requires approximately 12 hours.

Moored current meter measurements and synoptic hydrographic observations made in Massachusetts Bay show that freshening from the spring runoff dominates the low frequency currents and the hydrography of the Bay in the spring months. The major freshening is attributed to the Merrimack River which empties into the Gulf of Maine 30 km to the north of the Bay; discharge of the Merrimack increases by at least a factor of two in spring. Flow directly into the basin from several smaller rivers is not important. Two major features are found: a fresh surface plume confined to the upper 10 m of the water column which becomes more distinct as the seasonal thermocline develops, and a large deep fresh lens. Flow is clockwise around the deep lens and is consistent with the thermal wind relation. Sustained currents of 10 - 20 cm sec^{-1} with time scales of 5 - 10 days were observed as the deep lens (or lenses) slowly advected through the basin. Current observations made in the previous spring show similar low frequency behavior.

Two simple linear models of the semidiurnal tide on the continental shelf are used to estimate the vertical turbulent eddy viscosity, a linear bottom drag coefficient, and the change in the bottom drag coefficient during storms. The analytic solution for the response of a homogeneous water column with constant eddy viscosity to a sinusoidal body force with a slip bottom boundary condition is presented. With measurements of the tidal current at two depths, four parameters are shown to be independent of the body force: the ratio of the clockwise current at two depths, the ratio of counterclockwise current at two depths, the change in the tidal ellipse orientation, and the change in phase of the tidal ellipse. Observations of the semidiurnal tidal current on the New England continental shelf are consistent with a vertical eddy viscosity of 20 - 50 $cm^2 sec^{-1}$ and a bottom drag coefficient of .02 - .05 cm sec^{-1}. The Ekman depth is thus 10 m and the integrated adjustment time is approximately 28 hours.

An integrated linear model with linear damping of the semidiurnal tide on the continental shelf, forced uniformly at the shelf edge, shows an increasing phase lag of the tide at the coast with increased damping; amplitude remains relatively constant over a wide range of damping coefficient. Observations of the tide at the coast during storms shows a phase lag of as much as 10 degrees for the semidiurnal tide. For approximate dimensions of the New England shelf, this implies an increase by a factor of 3 - 5 of the bottom drag coefficient and an integrated motion adjustment time of 6 - 9 hours. Waves may be an important contribution to the increased bottom stress.

Thesis Supervisor: Dr. Robert C. Beardsley
Associate Professor of Oceanography
Massachusetts Institute of Technology

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J.S. Schlee kindly allowed me to work with the bottom current data presented in Chapter I, and to incorporate the sediment-bottom current study as part of my thesis research. His enthusiasm and patience is appreciated. V. Manohar-Maharaj provided the hydrographic data for the sections of March 29-30, April 15-16, May 5-6, and June 2-3 shown in Chapter II. R.C. Beardsley and C.F. Flagg provided data from the 1974 M.I.T. Winter Shelf Dynamics Experiment (NSF Grant GA 41075) to test the vertical Ekman model of the bottom boundary layer (Chapter III).

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The help of J. Maltais, use of the Woods Hole Oceanographic Institution Buoy Group computer programs, and assistance of the Woods Hole Oceanographic Institution Information Processing Center is acknowledged.

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Mrs. May Reese cheerfully and professionally typed the manuscript for which I am grateful.

Finally, the support of my parents and friends during this thesis work is appreciated.

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INTRODUCTION

This thesis investigates several physical processes which are important in the dynamics of currents in shallow water. Among oceanographers, coastal or continental shelf oceanography has generally been of less interest than study of the deep oceans. However, the need to understand the biological and physical environment of the nearshore coastal areas has increased in the last several years as use and development of coastal resources have grown. To assess the environmental consequences of these activities requires knowledge of the processes which mix and transport substances in the water and on the bottom.

Perhaps the slow development of coastal physical oceanography is a result of the complexity of the coastal physical environment. In shallow water meteorological forcing (heating, cooling, and wind stress), the physical boundaries of the coast and bottom topography, the flux of fresh water from rivers, and the fluxes of mass and momentum at the shelf boundary will be important factors affecting the currents. On the northeast coast of the United States, much of our knowledge of the circulation has come from studies of drift bottle and seabed drifter returns (Bumpus, 1973) which do not adequately resolve short time and space scale phenomena. However, short time scale (periods of days) and space scale (horizontal scales of several tens of kilometers) currents associated with meteorological forcing or with the seasonal density distribution may in fact dominate the nearshore flow. The objective of this thesis is to identify, describe and understand several of these short time and space scale physical processes.

Chapter One, prepared in collaboration with Dr. John Schlee of the U.S. Geological Survey, reports the results of a study of the equilibrium between the near bottom currents and the bottom sediments and the factors which drive the bottom currents in Massachusetts Bay, a semi-enclosed basin located in the western Gulf of Maine. Bottom currents were monitored over a one year period using Richardson current meters in areas of widely different bottom sediment types. The study suggests that the bottom sediments can be expected to move only occasionally in certain areas, and that the strong tidal currents in the Bay basically determine the bottom current speed. Strong storms dominate the net near bottom flow pattern in winter.

Chapter Two is a study of the current and density distribution in Massachusetts Bay during the period of the spring runoff. The Bay is located 30 km from a major river which discharges into the western Gulf of Maine. Synoptic hydrographic observations and moored current measurements show that the low frequency current in the spring months is primarily driven by the strong density gradients associated with the increased spring river discharge and that winds are not important.

Chapter Three is a study of the effects of friction on the currents on the New England continental shelf, and on the parametrization of the frictional effects. Two simple dynamical models appropriate for winter conditions are developed and used to estimate a bottom drag coefficient and a vertical eddy viscosity, and the change in the bottom drag coefficient during storms. The bottom drag coefficient determines the bottom stress for a given current and thus specifies the adjustment or decay time for integrated motions on the shelf, while the value of the vertical eddy viscosity determines the vertical distribution of the stress.

A brief chronology of this thesis work explains the author's interest in these aspects of the physical oceanography of Massachusetts Bay and the continental shelf. When I began thesis work in 1972 I was particularly interested in the field of coastal oceanography and especially in field observations of important time dependent processes; unfortunately there was little ongoing research within the community. At that time, Dr. John Schlee of the U.S. Geological Survey was conducting a survey of bottom currents in Massachusetts Bay as an extension of his work on the bottom sediment distribution. I had worked with Dr. Schlee on the currents in Massachusetts Bay in preparation for this monitoring program, and he kindly allowed me to work with the bottom current records. Chapter I presents the results of the bottom monitoring experiment.

The spring current records from the monitoring experiment and historical data suggested that the freshening associated with the spring runoff strongly influenced the hydrography and currents in Massachusetts Bay in spring, and was an important physical process. To study the importance of the spring runoff in the low frequency currents, a field experiment was conducted in the spring of 1973 using moored current meters and hydrographic observations (Chapter II).

At the same time, the author conducted a pilot experiment with Prof. R.C. Beardsley on the continental shelf to study the response to strong winter storms, and to test the feasibility of making current measurements on the continental shelf using similar mooring techniques as used in Massachusetts Bay (Beardsley and Butman, 1974). It was clear from that experiment and from my work on the wind driven currents in Massachusetts Bay that an estimate of frictional effects was important in understanding shallow water currents. Similar work had been begun using Massachusetts Bay data, but the shelf presented a less complicated geometry with which to work. Dr. Beardsley kindly allowed me to use data from a second more extensive shelf experiment conducted in late winter of 1974 to estimate the vertical eddy viscosity and the bottom drag coefficient using an Ekman model of the bottom boundary layer.

CHAPTER I

NEAR BOTTOM CURRENTS AND THE SEDIMENT DISTRIBUTION IN MASSACHUSETTS BAY

A. Introduction

Massachusetts Bay and its southeast extension, referred to as Cape Cod Bay, are bounded on three sides by the Massachusetts coast and open to the Gulf of Maine on the northeast between Cape Cod and Cape Ann (Figure 1.1). The opening is partially blocked by Stellwagen Bank which rises to within 20 m of the surface. Stellwagen Basin, located in the center of Massachusetts Bay, has a maximum depth slightly in excess of 100 m, though most is about 80 m deep. Depth changes in Cape Cod Bay, Stellwagen Basin, and on Stellwagen Bank occur gently (grades of .1 - .5%), except on the western side of Stellwagen Bank (grades of 6%). In contrast, the bottom along the western side of the Bay from Cape Ann to Plymouth (42°20'N 70°40'W) is hummocky and rough with depth changes of 5 m in .1 km (see C&GS chart No. 0808N-50 Cape Cod to Cape Ann).

Although complex, the bottom sediment distribution can be grouped in four categories by location and sediment type (Schlee and others, 1973; Oldale and others, 1973; Tucholke and others, 1972). Nearshore adjacent to the rocky coast from Cape Ann to Plymouth the rough bottom is a patchwork of gravel, sand, mud, and bedrock, while adjacent to constructional features (outwash and moraines) from Plymouth around Cape Cod, the generally smooth bottom is well sorted sand mixed with gravel. Offshore, the shallow bank is well sorted sand, or sand and gravel. Finest grained sediment of clay, silt and sand is found in the deep basin and in the

Figure 1.1 Map of Cape Cod and Massachusetts Bays. Smoothed 40 and 80 m contours show major bathymetric features (Plymouth located 42°20'N 70°40'W).

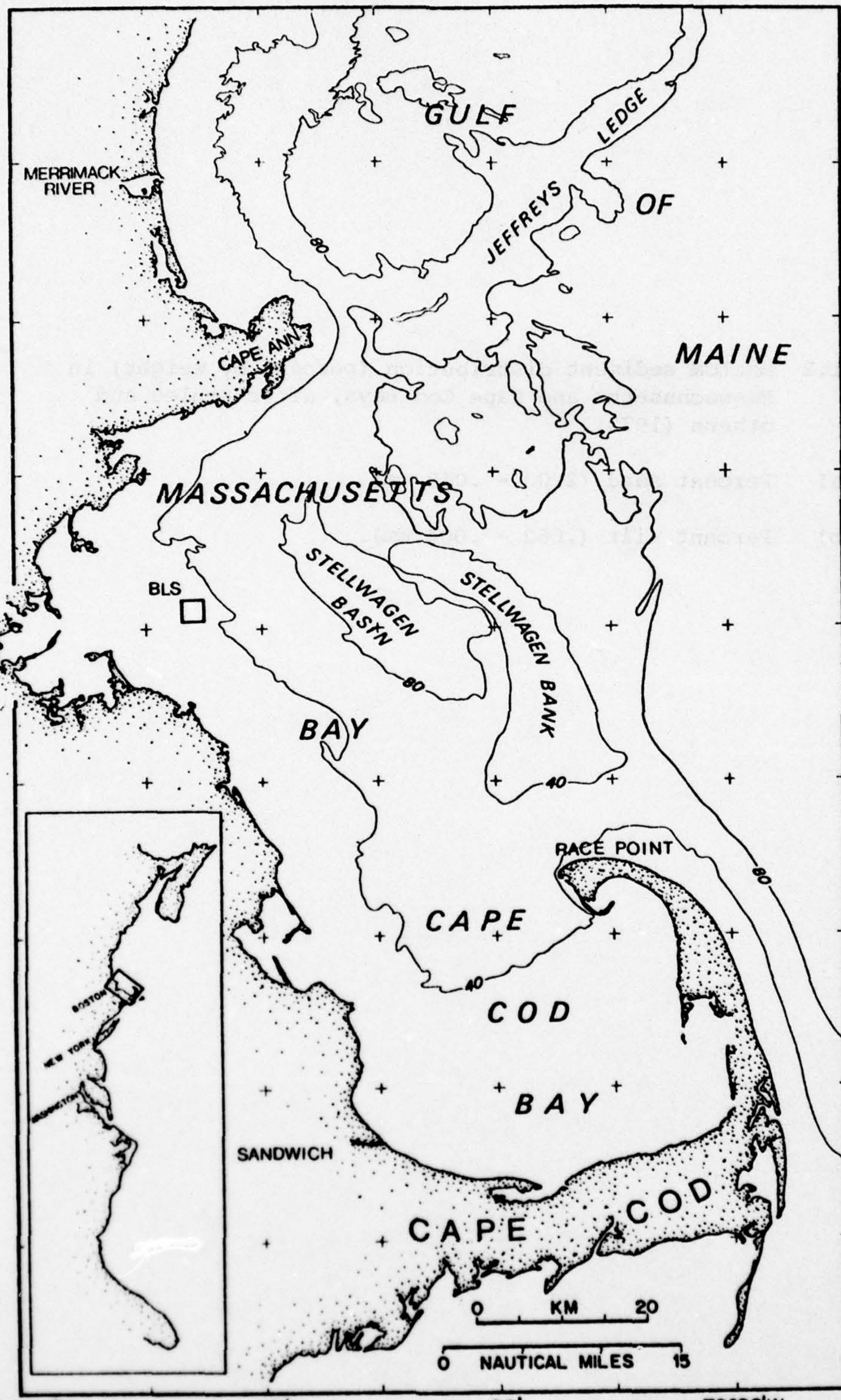

GULF
OF
MAINE
JEFFREYS LEDGE
MERRIMACK RIVER
CAPE ANN
MASSACHUSETTS
BAY
BOSTON
BLS
STELLWAGEN BASIN
STELLWAGEN BANK
RACE POINT
CAPE COD BAY
SANDWICH
CAPE COD
80
40
0 KM 20
0 NAUTICAL MILES 15
50'
40'
30'
20'
10'
42°00'N
50'
40'
71°00'W
40'
20'
70°00'W

Figure 1.1

Figure 1.2 Bottom sediment distribution (percent by weight) in Massachusetts and Cape Cod Bays, after Schlee and others (1973).

a) Percent sand (2.00 - .062 mm).

b) Percent silt (.062 - .004 mm).

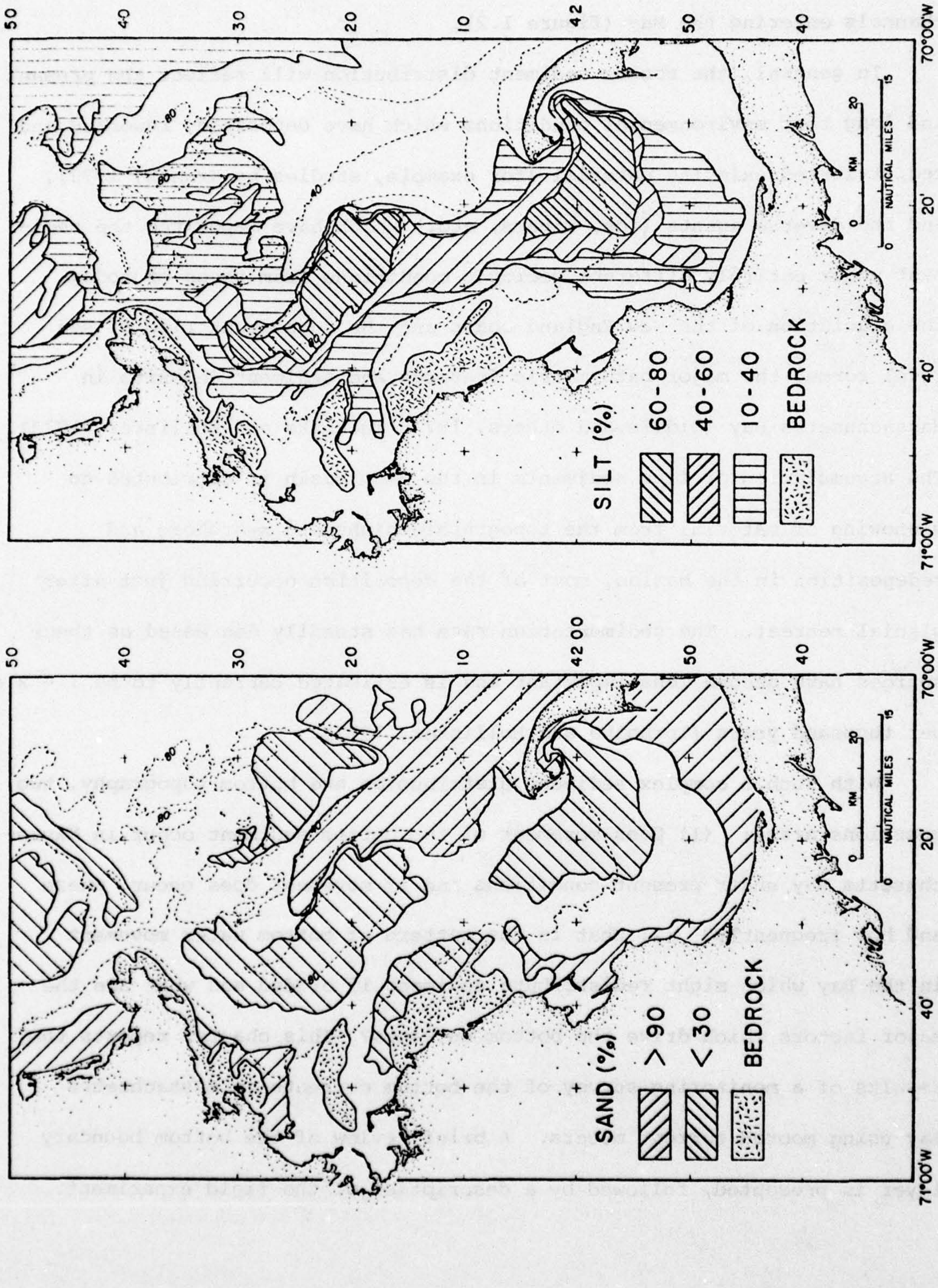

SAND (%)
>90
<30
BEDROCK
SILT (%)
60-80
40-60
10-40
BEDROCK
71°00'W
70°00'W
42 00
50
40
30
20
10
KM
NAUTICAL MILES
0
20'
40'
15

Figure 1.2

channels entering the Bay (figure 1.2).

In general, the bottom sediment distribution will reflect the present and long term environmental conditions which have deposited, reworked and redistributed existing material (for example, studies by Kranck, 1972), and the diverse agents (ice, rivers, etc.) which have deposited the sediment under entirely different geologic conditions from those of today. The glaciation of the New England coast and the subsequent rise of sea level formed the major bathymetric features and sediment deposits in Massachusetts Bay (Oldale and others, 1973; Tucholke and Hollister, 1973). The accumulation of fine sediments in the deep basin is attributed to winnowing of material from the topographic highs and nearshore and redeposition in the basins, most of the deposition occurring just after glacial retreat. The sedimentation rate has steadily decreased as these sources have become less important and is estimated currently to be 1 - 2 cm per thousand years (Tucholke and Hollister, 1973).

With such a complex sediment distribution and bottom topography, two questions arise: (1) Does movement of the bottom sediment occur in Massachusetts Bay under present conditions and if movement does occur, where and how frequently? (2) What is the pattern of bottom water movement in the Bay which might redistribute sediment if eroded and what are the major factors which drive the bottom currents? This chapter reports the results of a monitoring survey of the bottom currents in Massachsetts Bay using moored current meters. A brief review of the bottom boundary layer is presented, followed by a description of the field experiment.

The equilibrium between bottom sediments and bottom currents at the monitoring locations in the Bay is discussed next. Finally, winds were found to be a major factor in generating net bottom flow in winter, and the bottom flow pattern and associated sea level adjustment in Massachusetts Bay during strong winter storms is presented.

B. The Bottom Boundary Layer and Incipient Sediment Motion

A vast literature exists on the benthic boundary layer and processes of sediment motion and transport. (A review article by Wimbush and Munk (1971), work of Weatherly (1972), and a collection of articles edited by Swift, Duane and Pilkey (1972) were found useful references.) For this study of the adjustment of bottom sediments to currents, an estimate of the stress required to erode the existing material and a method to determine the bottom stress occurring in the marine environment is needed.

Field and laboratory studies have shown that the velocity profile in the bottom boundary layer is logarithmic, extending 1 - 4 m from the bottom (Sternberg, 1971, 1972; Miller and others, 1972; Weatherly, 1972). Thus

$$u(z) = \frac{u^*}{\kappa} \ln (z/z_o) , \qquad (1)$$

where

κ = VonKarman's constant = .4 ,

u^* = friction velocity = $(\tau/\rho)^{1/2}$,

τ = bottom stress,

z_o = roughness length,

z = height from bottom.

The value of z_o is found empirically to vary with flow characteristics. Specifically, the value of z_o depends on the relative size of the laminar sublayer and the roughness elements (d) of the boundary. The depth of the laminar sublayer is

$$\delta_{\ell} \simeq \frac{12\nu}{u*} , \qquad \nu = \text{kinematic viscosity } cm^2 sec^{-1} .$$

For hydrodynamically smooth flow, the roughness elements are enclosed completely within the laminar sublayer, while for rough flow the layer is disrupted by the roughness elements. The conditions for smooth or rough flow are summarized in Table 1.1. For a smooth bottom, velocity measurements at a single height will give an estimate of $u*$ but for a rough bottom an estimate of z_o is also required. Alternatively, for measurements at a fixed height above the bottom (1) can be written as

$$u*^2 = \rho C_D u^2 ,$$

where C_D is a property of the bottom roughness and is empirically determined. The drag coefficient is larger for rough bottoms than for smooth bottoms.

Measurements of the logarithmic bottom layer, particularly those of Sternberg (1971) and Weatherly (1972), have shown considerable variability in estimates of z_o . Variability decreased if the velocity profile was averaged over longer times (Weatherly, 1972). It may be that the roughness height is not a physical property of bed, but more a property of the fluid flow. Despite these rather serious reservations, estimates of the bottom stress will be made using bottom photographs to determine roughness, and (1).

Empirical curves of the bottom velocity (Sundborg, 1956; Allen, 1965), or of the bottom stress (Inman, 1963; Bagnold, 1962) required to erode well sorted sediment of a given grain diameter and density have been found to agree roughly with field measurements in a tidal channel (Sternberg,

TABLE 1.1

CONDITIONS FOR FLOW OVER SMOOTH AND ROUGH BOTTOM

Smooth Bottom	Rough Bottom
$d < \frac{4\nu}{u^*}$	$d > \frac{40\nu}{u^*}$
$z_o \sim \frac{.1\nu}{u^*}$	$z_o \sim d/30$
$u(z) = \frac{u^*}{\kappa} \ln \left(\frac{zu^*}{.1\nu}\right)$	$u(z) = \frac{u^*}{\kappa} \ln (z/z_o)$

d = characteristic roughness element height

ν = kinematic viscosity

z = height from bottom

z_o = effective roughness

u^* = friction velocity

κ = VonKarman's constant

1971; Miller and others, 1972). The bottom stress required for movement is a minimum in the sand range (.1 - 1 mm diameter) and is about 4 dynes ($u* = 2$ cm sec^{-1}); larger stress is required to move coarser material ($u* = 8$ cm sec^{-1} for diameter 1 cm). For material in the silt and clay range, the stress required for incipient movement depends on the degree of consolidation or 'age' of the sediment after deposition (Postma, 1967; Southard, Young, Hollister, 1971). The curves of Inman suggest a $u*$ of at least 4 cm sec^{-1} is required to erode consolidated silts and clays, but there is little experimental data; one exception is the laboratory study of Southard, Young, Hollister (1971), where a $u*$ of 1.37 cm sec^{-1} was required to move a deep sea mud.

If the bottom boundary layer is sufficiently turbulent deposition of fine grained material will not occur, or will be resuspended immediately. The Sundborg curve indicates no deposition of material less than .05 mm in diameter (silt) for velocities of 1 cm sec^{-1}. McCave (1972), summarizing laboratory studies, reports deposition of fine material will not occur for $u*$ in the range of .6 - .9 cm sec^{-1}.

It must be emphasized that the effects of cohesion, biological reworking and resuspension, aging after deposition, bed form and sorting on conditions for incipient sediment motion are not well understood. Rhoads and Young (1971) show intense biological reworking and resuspension of fine grained bottom sediment in Cape Cod Bay, and biological activity may be a significant factor in Massachusetts Bay. Also, bedform contributes to measured shear stress, not just grain diameter. Despite these problems the approximate bottom stress and velocities required for incipient motion of material in the sand, silt and clay, and gravel range

TABLE 1.2

ESTIMATES OF BOTTOM STRESS REQUIRED FOR INCIPIENT SEDIMENT MOTION AND DEPOSITION.

	Required Friction Velocity (U*) cm/sec	Speed 100 cm from bottom[2] cm/sec		
		Smooth	Rough	
Incipient Motion[1]			(a)	(b)
Sand, recently deposited silt and clay	2	61	40	32
Consolidated silt and clay	4	129	80	64
Gravel (1 cm diameter)	8	272	160	128
Deposition[3]				
Silt and clay	.6 - .9	17-26	12-18	10-14

[1]Estimated from Inman (1949), reproduced in Miller and others (1972); Southard, Young and Hollister (1971), Postma (1967).

[2]Velocities estimated from equation 1.1 . For a rough bottom two roughness heights are used: a) d = 1 cm, b) d = 5 cm (z_o = d/30). All estimates to nearest cm/sec.

[3]Estimates from McCave (1972).

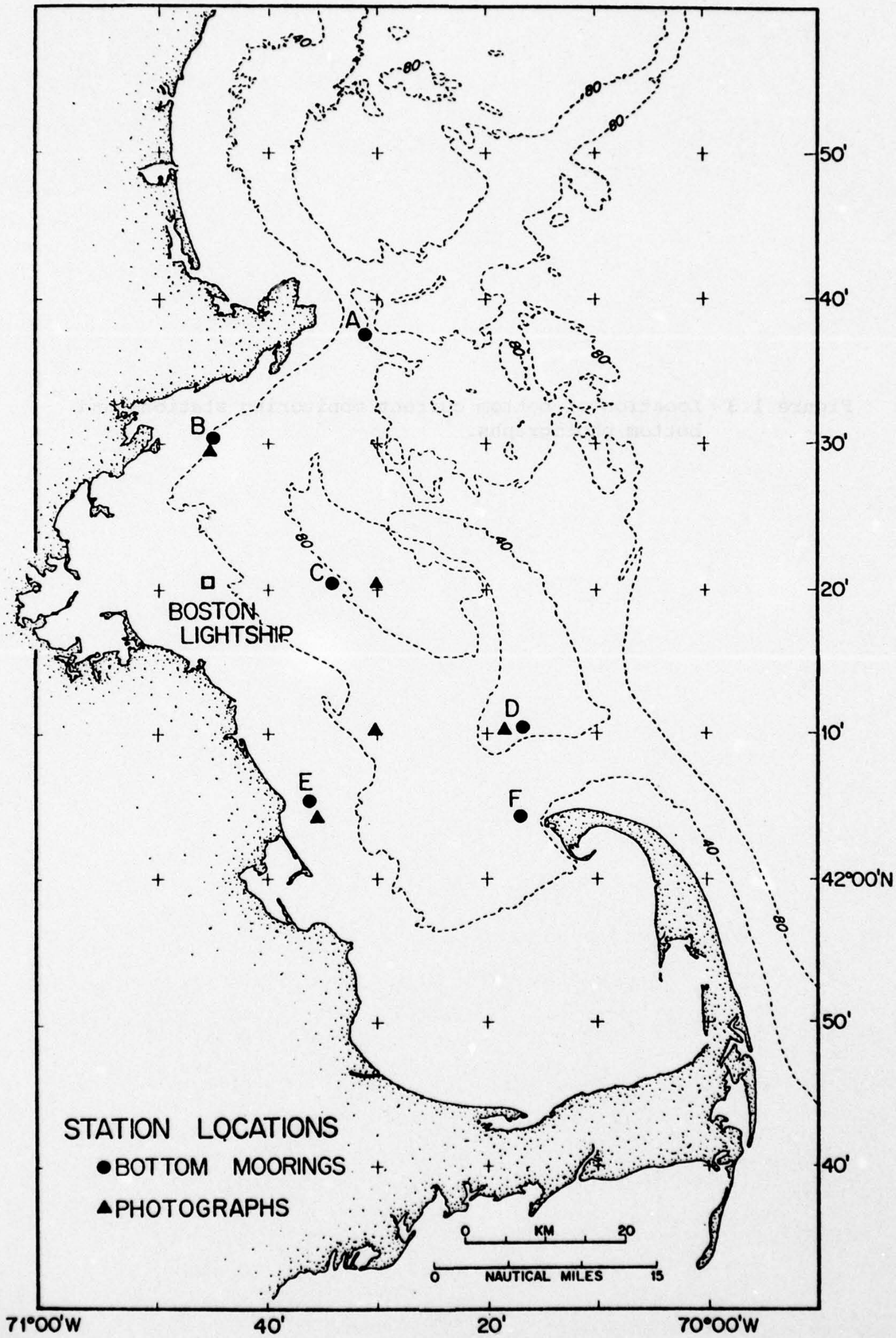
are estimated (Table 1.2) using the logarithmic law (1). The estimates in the sand and gravel range are probably fairly accurate while the estimates for finer material are somewhat uncertain, primarily because of uncertainty in the degree of consolidation and cohesion of the sediments and lack of laboratory and direct field erosion data. Although stress and velocity estimates in Table 1.2 should be viewed as approximate, they should provide useful ranges of erosion velocities for this study.

C. Field Program

1. Bottom Current Monitoring

Measurements of the near bottom currents were made in Massachusetts Bay from May, 1971 through July, 1972 at several locations (figure 1.3) which were selected to define the bottom current pattern in Massachusetts and Cape Cod Bays, if possible, and to measure the currents over various types of bottom sediments. Stations A and F were placed to monitor flow in the two channels leading into Stellwagen Basin; stations B and E were selected to monitor circulation on the coastal shelf. Station C was placed in Stellwagen Basin, and station D was placed to monitor on Stellwagen Bank, Stations A, C and F are in areas of fine sediments, while B, D and E are in areas with coarser bottom material. A current meter was deployed throughout the study at the Boston Lightship to provide continuous measurements at one location. It was hoped to obtain a one month bottom current record each season at stations A-F, however temporary loss of one current meter and malfunctions prevented this objective. The latitude and longitude of the stations and the dates of the successful measurements are listed in Appendix A . The time sequence of current meter deployments is shown in figure 1.4.

Figure 1.3 Location of bottom current monitoring stations and bottom photographs.



80
A
B
C
D
E
F
40
BOSTON
LIGHTSHIP
50'
40'
30'
20'
10'
42°00'N
STATION LOCATIONS
BOTTOM MOORINGS
PHOTOGRAPHS
KM
0
20
NAUTICAL MILES
0
15
71°00'W
40'
20'
70°00'W

Figure 1.3

Figure 1.4 Schedule of current meter deployment in Massachusetts and Cape Cod Bays. Records are numbered sequentially at each station and dashed lines indicate records with inaccurate sampling rate and time base. See figure 1.3 for station locations.

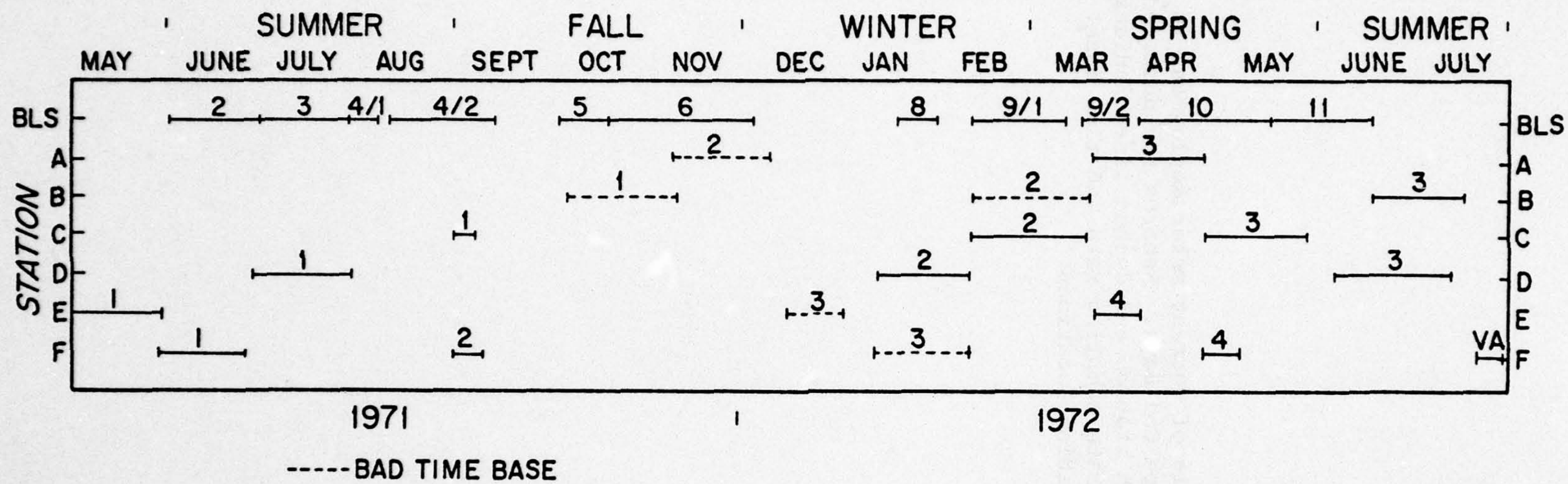

SUMMER
FALL
WINTER
SPRING
SUMMER
MAY
JUNE
JULY
AUG
SEPT
OCT
NOV
DEC
JAN
FEB
MAR
APR
MAY
JUNE
JULY
STATION
BLS
A
B
C
D
E
F
2
3
4/1
4/2
5
6
8
9/1
9/2
10
11
VA
1971
1972
-----BAD TIME BASE

Figure 1.4

The current meters at stations A-F were moored 1 m from the bottom, with all floatation within approximately 5 m of the bottom (figure 1.5). The mooring was designed with groundlines on either side of the main anchor to facilitate recovery of the instrument if the surface marker was lost, and a small acoustic beacon was attached to the meters to aid in locating the instrument. The Boston Lightship meter was suspended from the lightship approximately 8 m from the bottom. One Geodyne model 850 and the two model 102 Richardson type current maters were used in the survey; the 850 current meter was deployed continuously from the Boston Lightship, while the 102 meters were rotated among the offshore stations. No current meters were permanently lost during the program, and approximately 60% of the instruments returned useable data. Instrument malfunction occurred in the 102 current meters which record on film, and was due to film advance problems, malfunction of the circuitry, or to an inaccurate time base. The 850 current meter was found extremely reliable.

2. Bottom Sediment Composition and Texture Near Current Meter Stations

Grain size analysis of the bottom sediment near the current meter stations (Table 1.3) was obtained from an earlier sediment survey (Schlee and others, 1973; Hathaway, 1971). Most of the samples are sandy silt or sand, except at station E where there is a large amount of gravel. At the deeper stations (A,C,F) the bottom composition is similar: 44-47% sand, 32-39% silt, 16-19% clay. At station B and D, the bottom is sand or sand and gravel.

Bottom photographs (see figure 1.2 for locations) obtained from previous studies clearly show the different bottom sediments, sorting and texture. Near Station B a thin layer of fine material overlies gravel

Figure 1.5 Schematic diagram of mooring for near bottom current measurements.

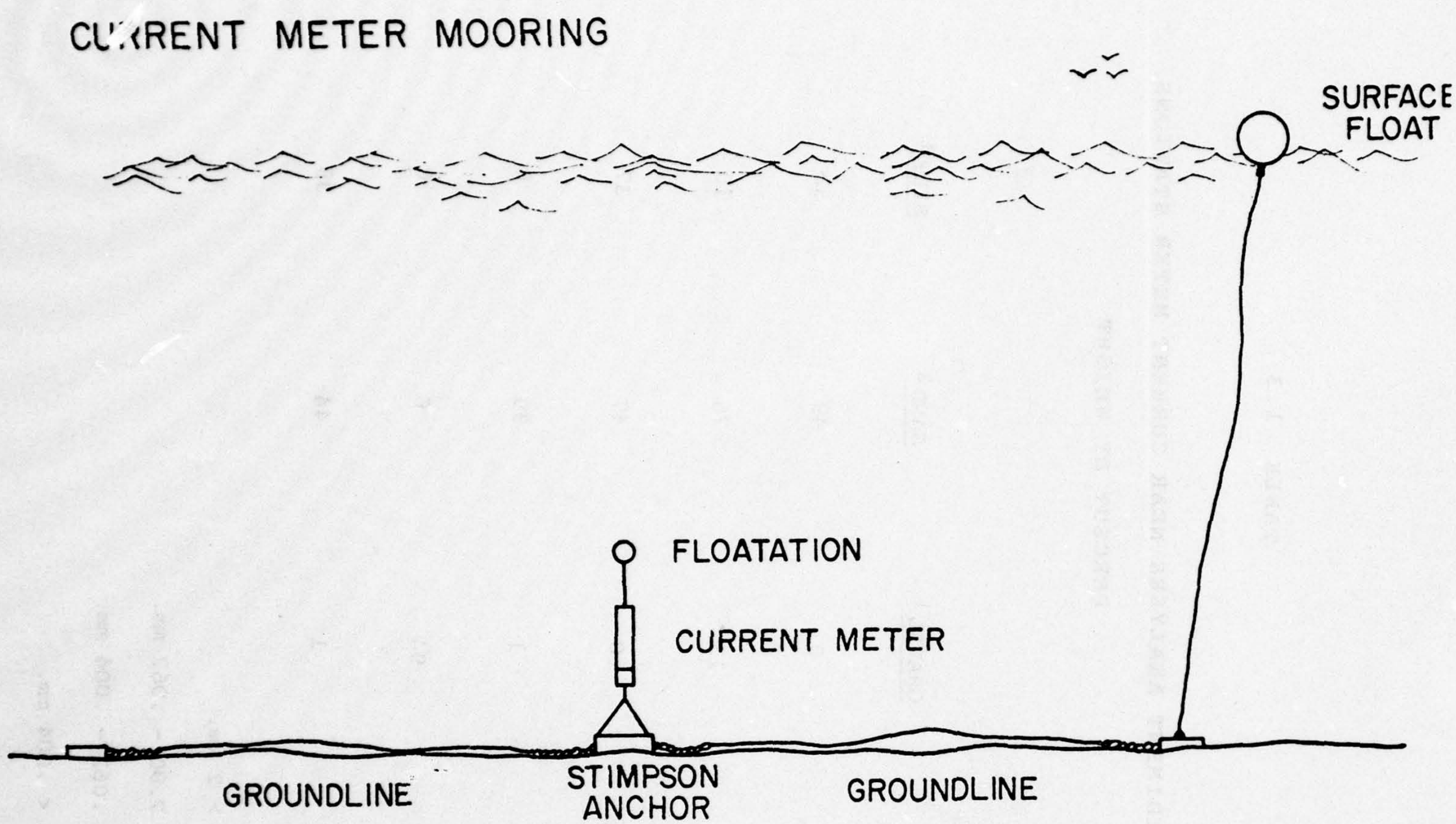

CURRENT METER MOORING
SURFACE FLOAT
FLOATATION
CURRENT METER
GROUNDLINE
STIMPSON ANCHOR
GROUNDLINE

Figure 1.5

TABLE 1.3

SEDIMENT ANALYSES NEAR CURRENT METER STATIONS

PERCENT BY WEIGHT

STATION	GRAVEL[1]	SAND[2]	SILT[3]	CLAY[4]
A	0	49	32	19
B	13	76	10	0
C	0	47	37	16
D	1	99	0	0
E	94	6	0	0
F	1	44	39	16

[1]Gravel > 2 mm.

[2]Sand 2.00 - .062 mm.

[3]Silt .062 - .004 mm.

[4]Clay < .004 mm.

and coarse sand (figures 1.6 a,b) suggesting little active current erosion. The two photographs at station B illustrate the typical patchiness of the bottom sediment distribution and the difficulty in estimating a meaningful roughness height in areas with poor sorting. In the deep basin (figure 1.6 c,d) the bottom is soft silt, sand and clay, much smoother than at station B (roughness elements less than 1 cm), and again there is little evidence for active erosion. In contrast, on Stellwagen Bank (station D, figure 1.6 e) there is no apparent fine material and ripple marks suggest active or recent erosion. Here an approximate roughness height is estimated to be 5 cm. At station E (figure 1.6 f) the bottom is coarse sand with no evidence of active erosion.

In summary, for the purposes of this study the bottom photographs suggest that the bottom in the deep basin (station C) should be considered smooth, while at stations B, D, and E the bottom is rough. No bottom photographs were available at stations A or F, but the bottom sediment is similar to station C suggesting that the bottom might be considered smooth.

D. Speed Statistics and Estimates of Bottom Sediment Movement

The stations may be grouped in three classes by the observed maximum speed and average speed (Table 1.4); the largest maximum and average speeds occur at stations D and F at the southern mouth of the Bay. Nearshore (stations B and E) the maximum and average speeds are weaker, while at the deep stations (A and C) the average speeds are similar to the nearshore stations, but the maximum speeds are somewhat less. The speed distribution in the basin is primarily controlled by the strength of the tidal current and the depth. At the mouth of the Bay the tidal currents are strong and the water is shallow so that wind generated currents and possibly waves contribute to the maximum speed. In the deep basin and on the shallow border of the Bay the tidal currents are substantially weaker, but maximum

Figure 1.6 Bottom photographs in Massachusetts Bay showing bottom texture.

a) Station 1200 (near current meter station B)
42°29.8'N 70°44.7'W 43 m
Grab sample description: grey silty fine sand
Composition (%): gvl 13.4; sand 76.3; silt 10.4; clay 0.

b) Second photograph at station 1200.

c) Station 1202 (near current meter station C)
42°20.6'N 70°30.0'W 91 m
Grab sample description: grey clay
Composition (%): gvl 0; sand 4.7; silt 49.3; clay 43.0.

d) Station 1203
42°10.1'N 70°30.2'W 55 m
Grab sample description: 5 cm soft brown clay overlying stiff grey clay
Composition (%): gvl 0; sand 39.0; silt 49.5; clay 11.7.

e) Bottom photograph near station D. Rod in picture approximately 1 m long.

f) Bottom photograph near station E.

Photographs at stations 1200-1203 from continental margin study (Hathaway, 1971). Others from Dr. D. Cooper, U.S. National Marine Fisheries Service, Woods Hole, Mass.

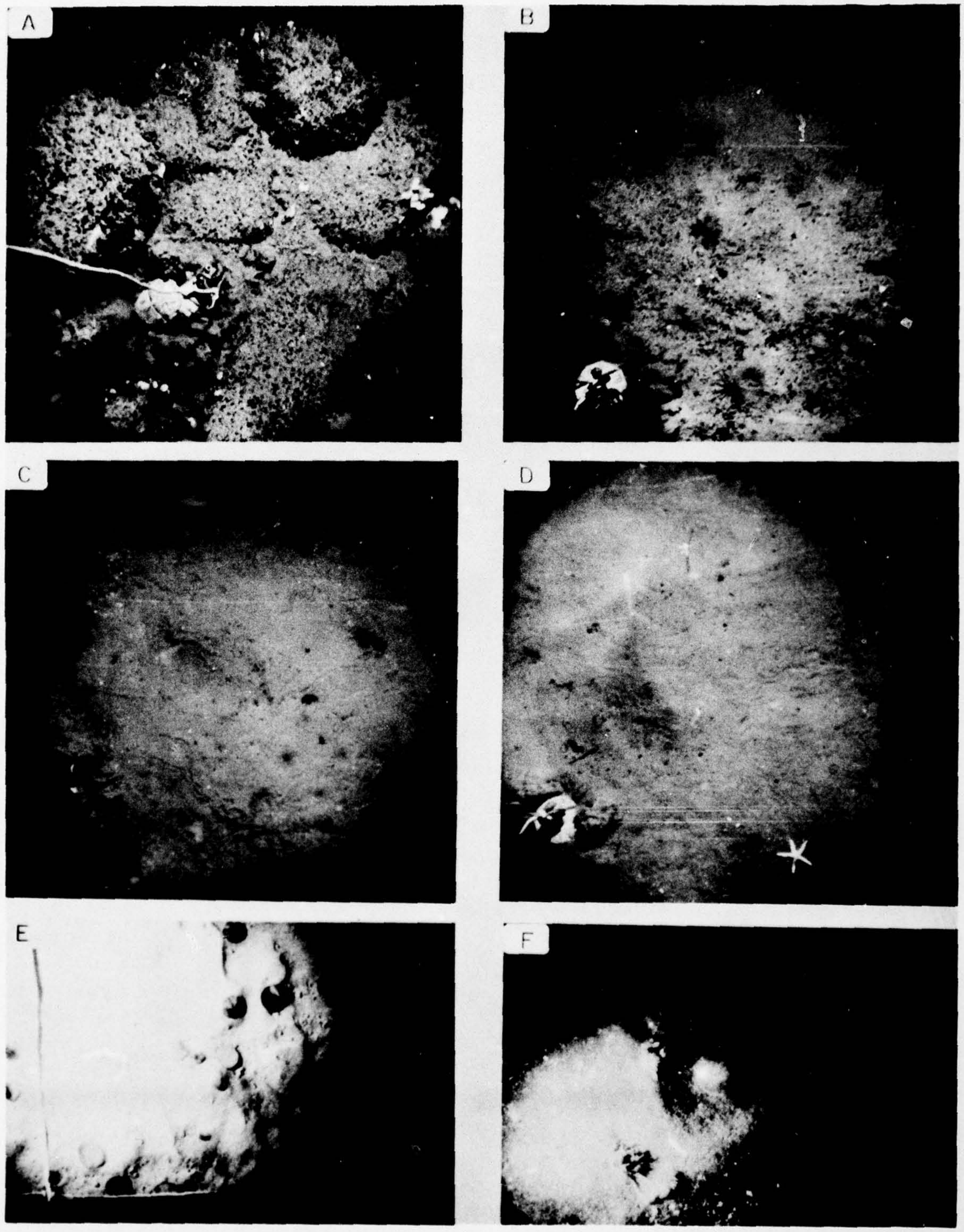

A
B
C
D
E
F

TABLE 1.4

RANGE OF SPEED STATISTICS FOR BOTTOM CURRENT RECORDS[1]

	A	B	C	D	E	F
Maximum Speed	24-34	25-43	26-29	32-47	29-37	35-47
Speed Exceeded 1% of Time	18	14	20	26-36	20-26	27-33
Speed Exceeded 5% of Time	12	10	12-13	23-29	17-21	24-28
Average Speed[2]	5.7	4.7	4.7-6.1	13.1-17.0	4.2-7.9	14.5-15.9
Stand. Dev.	3.2	3.8	3.0-3.5	5.2-8.1	4.2	5.5-7.7
No. Records, Time Base[3]	2,1	3,2	3	3	3,1	2
Days[4]	36	28	66	98	42	44
Semidiurnal Tide[5]						
Major	2.6	3.5	6.2	20.2	5.2	17.5
Minor	.1	2.6	1.2	-3.6	.4	2.2
Orientation	17	5	79	72	170	34

[1]Only records longer than 15 days included except for max. speed. Records not simultaneous, so not directly comparable. All numbers in cm sec^{-1}.

[2]Average speed and stand. deviation for vector averaged 1 hr. samples.

[3]Number of records included (first digit); number with time base error (2nd digit); only max. speeds tabulated for records with time base error.

[4]Total number of days excluding records with time base error.

[5]Computed from 15 day pieces; ellipse orientation with respect to north. Estimated error ±.5 cm sec^{-1} in speed, ±5° in direction. Positive minor axis indicates vector rotates counterclockwise, negative clockwise.

speeds are larger in the shallow regions probably because of wind and wave action.

At station B, D, and F, the observed maximum speeds are above the estimated minimum critical erosion speeds (Table 1.2), assuming a rough bottom. None of the observed speeds are large enough to move sand if the bottom is smooth. At stations A and C, the maximum speeds are substantially below critical erosion speeds for consolidated or unconsolidated silt and clay. At all stations during most of the measurement period, the current speed was substantially less than the critical erosion speed, and thus bottom movement of sediments, if it occurs at all, is infrequent. There is some suggestion that the maximum bottom speeds occurred in the fall and winter months (Table 1.5); however the data is too sparse to make any generalizations with statistical reliability. The average current at stations D and F is sufficiently high to prevent deposition of fine material, while at the stations A,B,C deposition is possible 95% of the time.

A graph of median grain size versus the maximum one minute average current speed observed during each of the bottom current meter records (figure 1.6) shows that most of the points fall within the limits of the competency curves of Sundborg. It should be noted that these empirical curves do not take into account bottom roughness except through grain diameter. Only at station E do the values fall far below the competency curve, and this is because most of the sample used to characterize the bottom sediments at station E is gravel which is relict glacial material.

In summary in the well sorted sand regions (station B and D), we can expect occasional movement if the bottom is assumed to be rough with roughness elements of at least 1 cm. Estimates of critical erosion stress for the silt-sand bottoms are uncertain, but the data suggest that the

TABLE 1.5

MAXIMUM SPEEDS BY SEASON (cm/sec)

STATION	WINTER	SPRING	SUMMER	FALL
A		24		34
B	43		25	26
C	29	26		27
D	47		32,40	
E	37	29,37		
F	42	41	35	47

Figure 1.7 Maximum one minute average speed observed at current meter stations A-F and Sundborg (1956) competency curves (reproduced in Miller and others, 1972). Note logarithmic velocity scale.

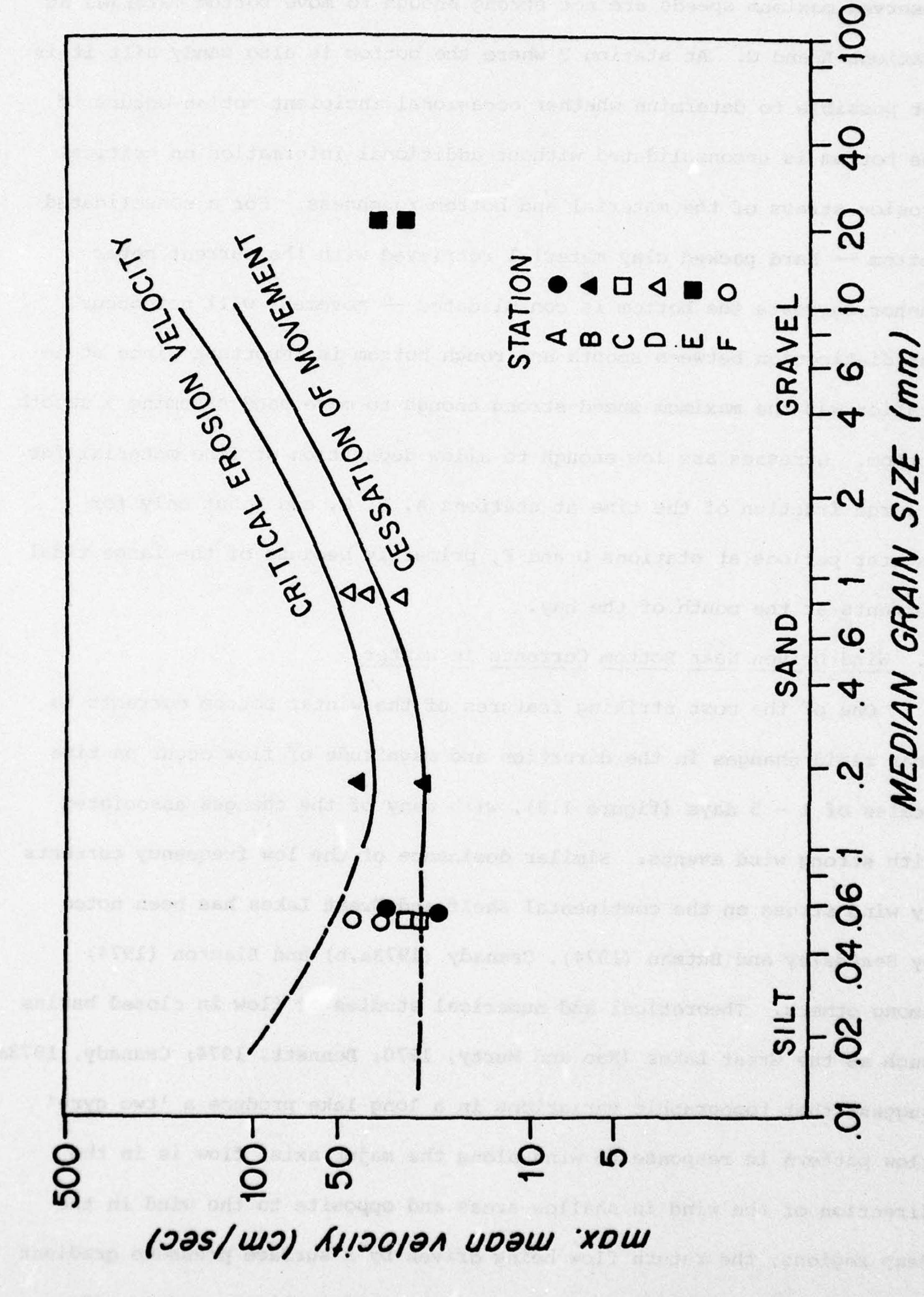

CRITICAL EROSION VELOCITY
CESSATION OF MOVEMENT
STATION
A ●
B ▲
C □
D △
E ■
F ○
SILT
SAND
GRAVEL
.01 .02 .04 .06 .1 .2 .4 .6 1 2 4 6 10 20 40 100
MEDIAN GRAIN SIZE (mm)
500
100
50
10
5
max. mean velocity (cm/sec)

Figure 1.7

observed maximum speeds are not strong enough to move bottom material at stations A and C. At station F where the bottom is also sandy silt it is not possible to determine whether occasional incipient motion occurs if the bottom is unconsolidated without additional information on critical erosion stress of the material and bottom roughness. For a consolidated bottom — hard packed clay material retrieved with the current meter anchor suggests the bottom is consolidated — movement will not occur. The distinction between smooth and rough bottom is important since at no station was the maximum speed strong enough to move sand assuming a smooth bottom. Stresses are low enough to allow deposition of fine material for a large fraction of the time at stations A, B, C, and E but only for shorter periods at stations D and F, primarily because of the large tidal currents at the mouth of the bay.

E. Wind Driven Near Bottom Currents In Winter

One of the most striking features of the winter bottom currents is that rapid changes in the direction and magnitude of flow occur on time scales of 1 - 5 days (figure 1.8), with many of the changes associated with strong wind events. Similar dominance of the low frequency currents by wind stress on the continental shelf and Great Lakes has been noted by Beardsley and Butman (1974), Csanady (1973a,b) and Blanton (1974) among others. Theoretical and numerical studies of flow in closed basins such as the Great Lakes (Rao and Murty, 1970; Bennett, 1974; Csanady, 1973a) suggest that topographic variations in a long lake produce a 'two gyre' flow pattern in response to wind along the major axis; flow is in the direction of the wind in shallow areas and opposite to the wind in the deep regions, the return flow being driven by a surface pressure gradient or setup. The large depth changes and the semi-enclosed geometry of

Figure 1.8 Progressive vector diagram of current at station D in January and February, 1972. Two major departures from the net southwest flow on Jan. 25-26, and Feb. 3-6 are associated with winter storms.

a) Progressive vector diagram from hourly data.

b) Daily average current and wind stress.

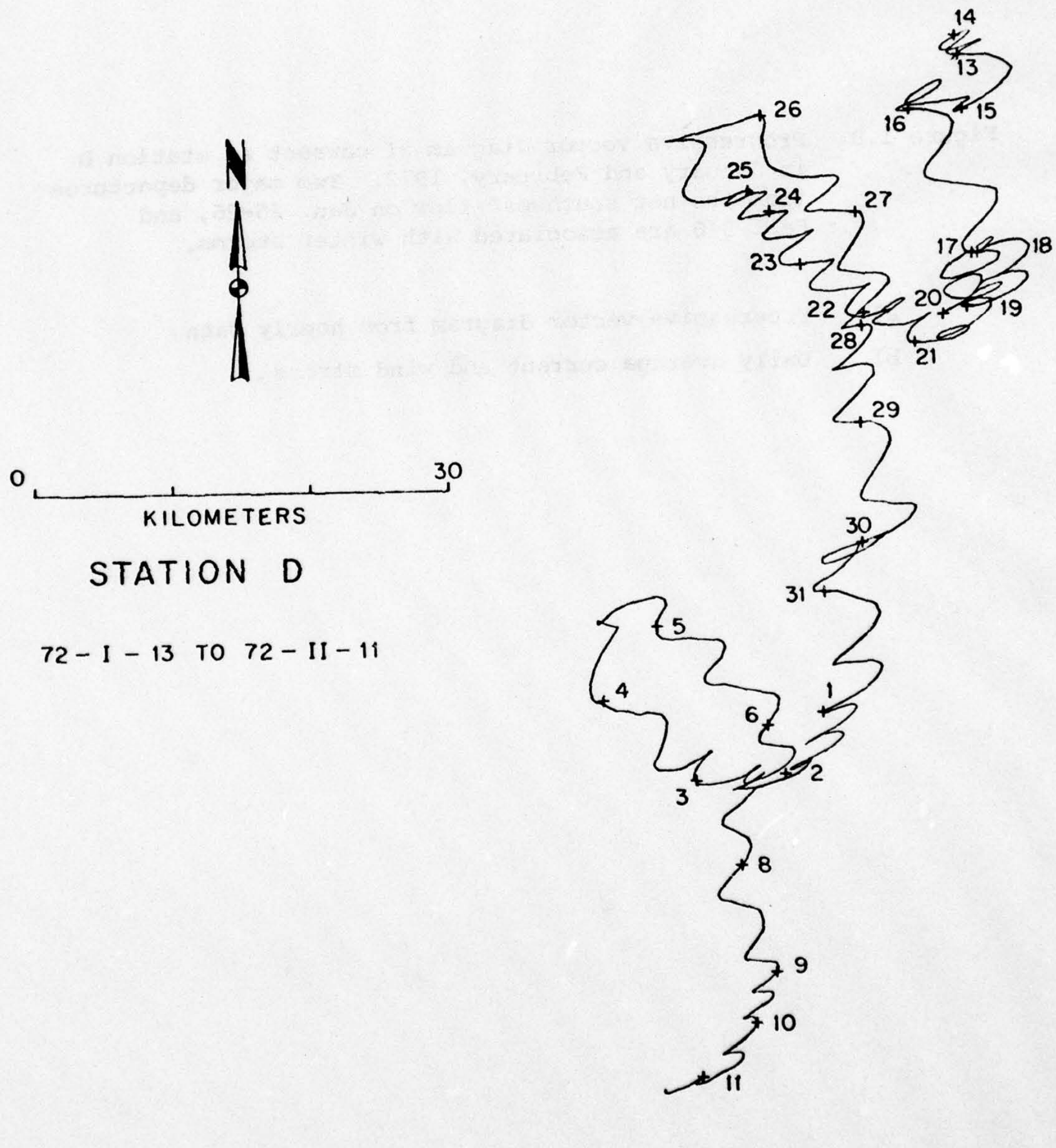

N
0
30
KILOMETERS
STATION D
72 - I - 13 TO 72 - II - 11
14
13
15
16
26
25
24
27
17
18
23
22
20
19
28
21
29
30
31
5
4
1
6
2
3
8
9
10
11

Figure 1.8a

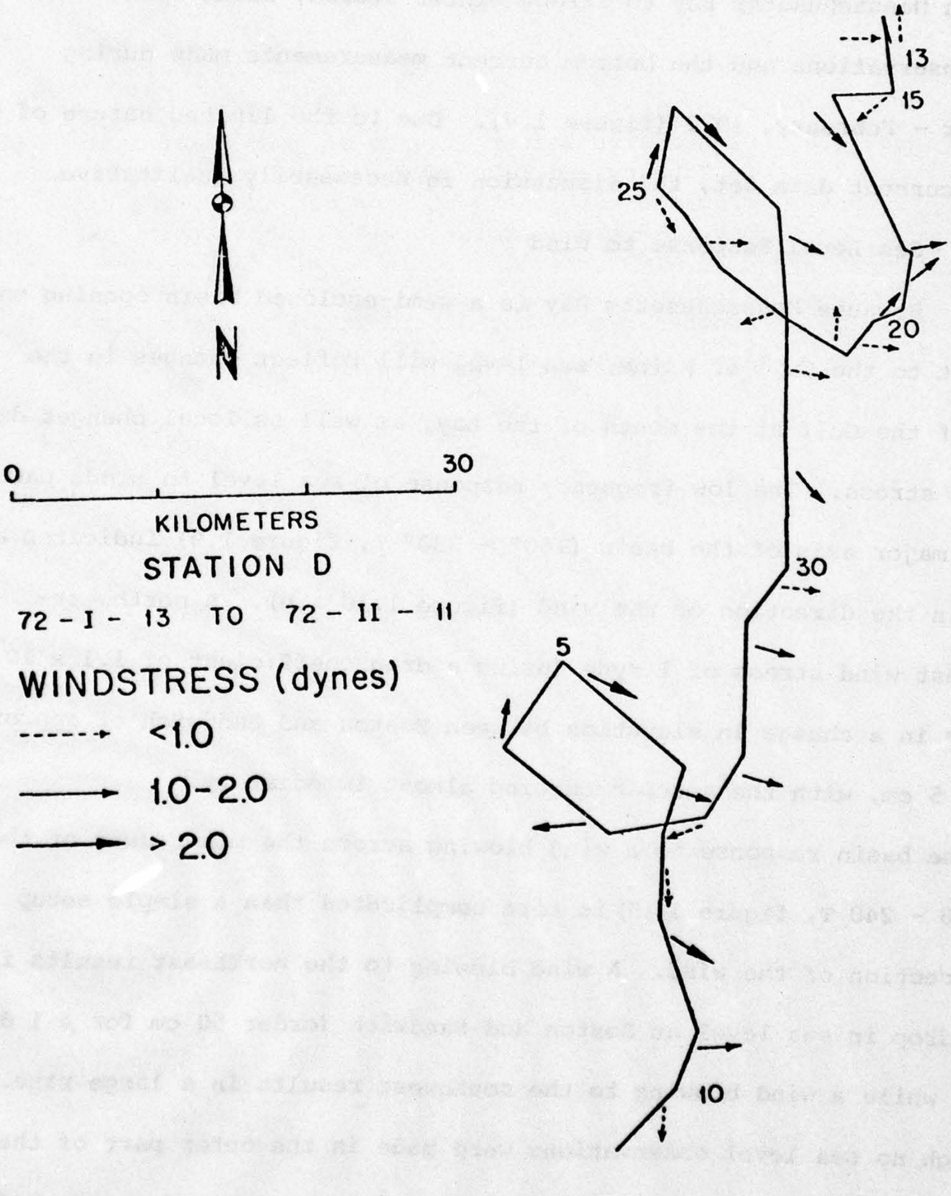

N
0
30
KILOMETERS
STATION D
72 - I - 13 TO 72 - II - 11
WINDSTRESS (dynes)
<1.0
1.0-2.0
> 2.0
13
15
25
20
30
5
10

Figure 1.8b

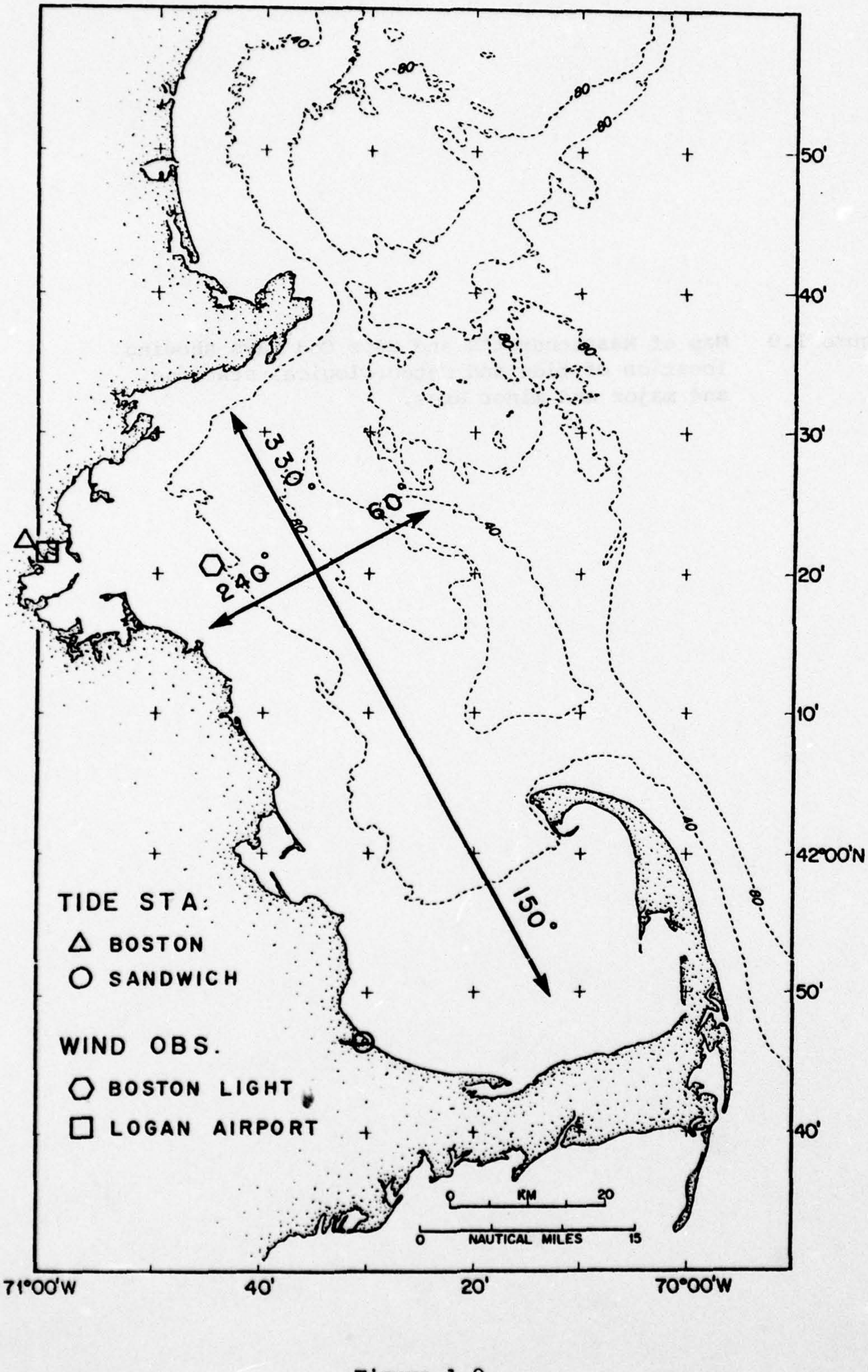
Massachusetts Bay suggests that a similar flow pattern may occur during strong wind events which could be important in redistributing the fine bottom sediments from the nearshore areas and bank to the deep basin. In this section we investigate the response of the bottom currents and sea level in Massachusetts Bay to strong winter storms, using coastal sea level observations and the bottom current measurements made during December - February, 1972 (figure 1.4). Due to the limited nature of the bottom current data set, the discussion is necessarily qualitative.

1. Sea Level Response to Wind

Because Massachusetts Bay is a semi-enclosed basin opening on the east to the Gulf of Maine, sea level will reflect changes in the level of the Gulf at the mouth of the Bay, as well as local changes due to wind stress. The low frequency response of sea level to winds parallel to the major axis of the basin (150° - 330° T, figure 1.9) indicates a setup in the direction of the wind (figure 1.10 a,b). A northwest-southeast wind stress of 1 dyne (using a drag coefficient of 1.1×10^{-3}) results in a change in elevation between Boston and Sandwich of approximately 5 cm, with the setup occurring almost immediately.

The basin response to a wind blowing across the minor axis of the Bay (60 - 240 T, figure 1.10) is more complicated than a simple setup in the direction of the wind. A wind blowing to the northeast results in a large drop in sea level at Boston and Sandwich (order 50 cm for a 1 dyne wind), while a wind blowing to the southwest results in a large rise. Although no sea level observations were made in the outer part of the Bay it is unreasonable to expect that such a large setup (an order of magnitude larger than the setup for winds along the long axis) occurs

Figure 1.9 Map of Massachusetts and Cape Cod Bays showing location of tide and meteorological stations, and major and minor axes.



80
80
80
80
80
80
40
40
40
80
330°
60°
240°
150°
50'
40'
30'
20'
10'
42°00'N
50'
40'
TIDE STA:
BOSTON
SANDWICH
WIND OBS.
BOSTON LIGHT
LOGAN AIRPORT
0 KM 20
0 NAUTICAL MILES 15
71°00'W
40'
20'
70°00'W

Figure 1.9

Figure 1.10 a,b — Low frequency response of sea level in Massachusetts Bay to wind, Jan.-Feb., 1972.

Top: Difference in sea level between Sandwich and Boston and component of wind stress parallel to major axis of the Bay (330°-150°T, figure 1.9).

Bottom: Deviation of sea level at Boston from mean and component of wind stress parallel to minor axis of the Bay, and across the open side (60°-240°T). Deviation from mean level is corrected for the inverse barometer effect using atmospheric pressure measured at Logan Airport.

Notes:

1. All series have been filtered with a Gaussian filter (½ power at 56 hr., see Appendix B).
2. Note difference in sea level scale between top and bottom figures; upper scale expanded five times.
3. Wind stress is computed from Logan wind data using a constant drag coefficient of 1.1×10^{-3}. The wind stress would be approximately a factor of two higher for wind observations from the Boston Lightship. The drag coefficient may vary by as much as a factor of two - three over the period (Csanady, 1972b; Parker, 1974) as the stability of the air-water interface changes.
4. Sea level difference between Sandwich and Boston has not been corrected for atmospheric pressure differences. Parker (1974) indicates that the difference is less than 3 mb, and usually less than .5 mb, at least in summer.
5. Wind at Logan and Boston Lightship shown for the period Feb. 18-22 where the two observations differed significantly (Logan, solid line; Lightship, dotted line).

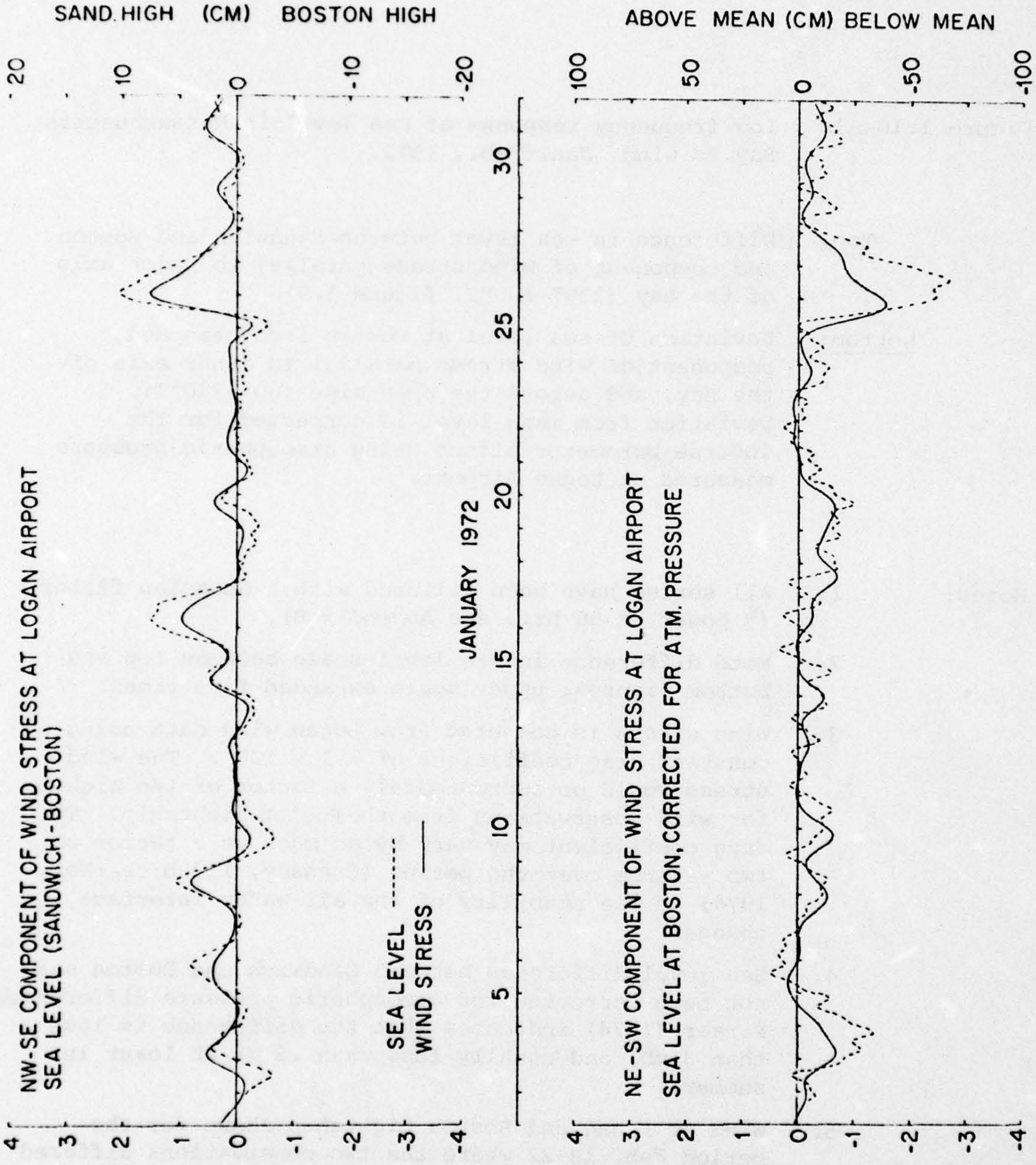

SAND HIGH (CM) BOSTON HIGH
20
10
0
-10
-20
ABOVE MEAN (CM) BELOW MEAN
100
50
0
-50
-100
NW-SE COMPONENT OF WIND STRESS AT LOGAN AIRPORT
SEA LEVEL (SANDWICH-BOSTON)
SEA LEVEL
WIND STRESS
JANUARY 1972
5
10
15
20
25
30
NE-SW COMPONENT OF WIND STRESS AT LOGAN AIRPORT
SEA LEVEL AT BOSTON, CORRECTED FOR ATM. PRESSURE
FROM NW (DYNES) FROM SE
4
3
2
1
0
-1
-2
-3
-4
FROM NE (DYNES) FROM SW

Figure 1.10a

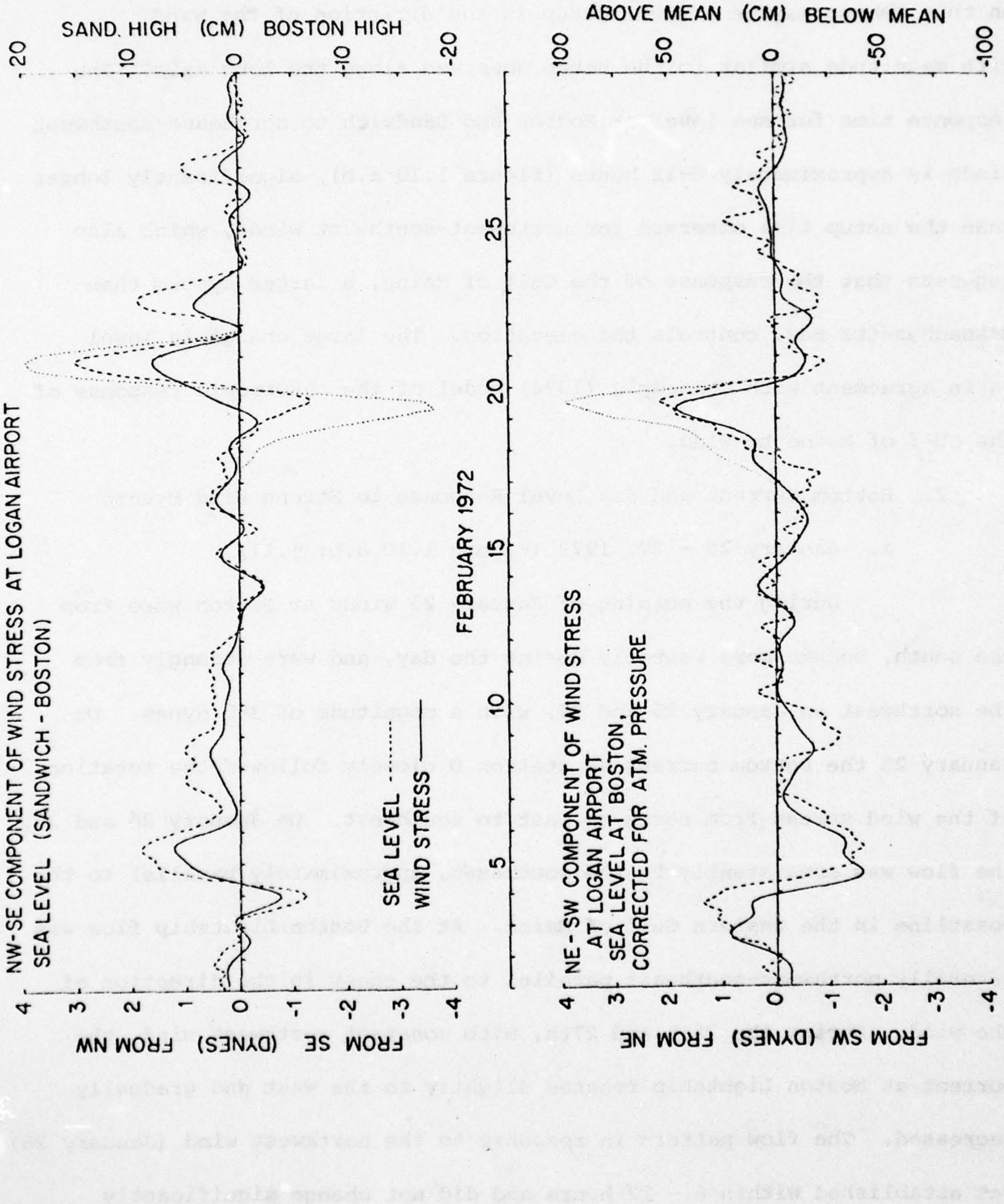

NW-SE COMPONENT OF WIND STRESS AT LOGAN AIRPORT
SEA LEVEL (SANDWICH - BOSTON)
SAND. HIGH (CM) BOSTON HIGH
20
10
0
-10
-20
FROM SE (DYNES) FROM NW
4
3
2
1
0
-1
-2
-3
-4
SEA LEVEL
WIND STRESS
FEBRUARY 1972
5
10
15
20
25
NE - SW COMPONENT OF WIND STRESS
AT LOGAN AIRPORT
SEA LEVEL AT BOSTON
CORRECTED FOR ATM. PRESSURE
ABOVE MEAN (CM) BELOW MEAN
100
50
0
-50
-100
FROM SW (DYNES) FROM NE

Figure 1.10b

across the short axis of the Bay. The large change in sea level in response to northeast-southwest winds probably reflects a change in the level of the Gulf of Maine to which the Bay must adjust. Superimposed on this change may be a small setup in the direction of the wind with magnitude similar to the setup observed along the long axis. The response time for sea level at Boston and Sandwich to northeast-southwest winds is approximately 6-12 hours (figure 1.10 a,b), significantly longer than the setup time observed for northwest-southeast winds, which also suggests that the response of the Gulf of Maine, a larger system than Massachusetts Bay, controls the elevation. The large change in level is in agreement with Csanady's (1974) model of the barotropic response of the Gulf of Maine to wind.

2. Bottom Current and Sea Level Response to Strong Wind Events

a. January 25 - 27, 1972 (Figure 1.10 a,b; 1.11)

During the morning of January 25 winds at Boston were from the south, became more westerly during the day, and were strongly from the northwest on January 26 and 27, with a magnitude of 3-5 dynes. On January 25 the bottom current at station D closely followed the rotation of the wind stress from north to east to southwest. On January 26 and 27, the flow was consistently to the southeast, approximately parallel to the coastline in the western Gulf of Maine. At the Boston Lightship flow was generally northwest-southeast parallel to the coast in the direction of the wind. During the 26th and 27th, with constant northwest wind, the current at Boston Lightship rotated slightly to the west and gradually decreased. The flow pattern in response to the northwest wind (January 26) was established within 6 - 12 hours and did not change significantly throughout the 26th.

Figure 1.11 Wind and near bottom current in Massachusetts Bay January 25-27, 1972. Current records have been filtered with a Gaussian filter (½ power at 33 hours, see Appendix B) to remove tidal oscillations. Values of current and wind stress shown every six hours. Current meter located 1 m from the bottom at station D and F, 10 m from the bottom at the Boston Lightship.

JAN 25, 1972

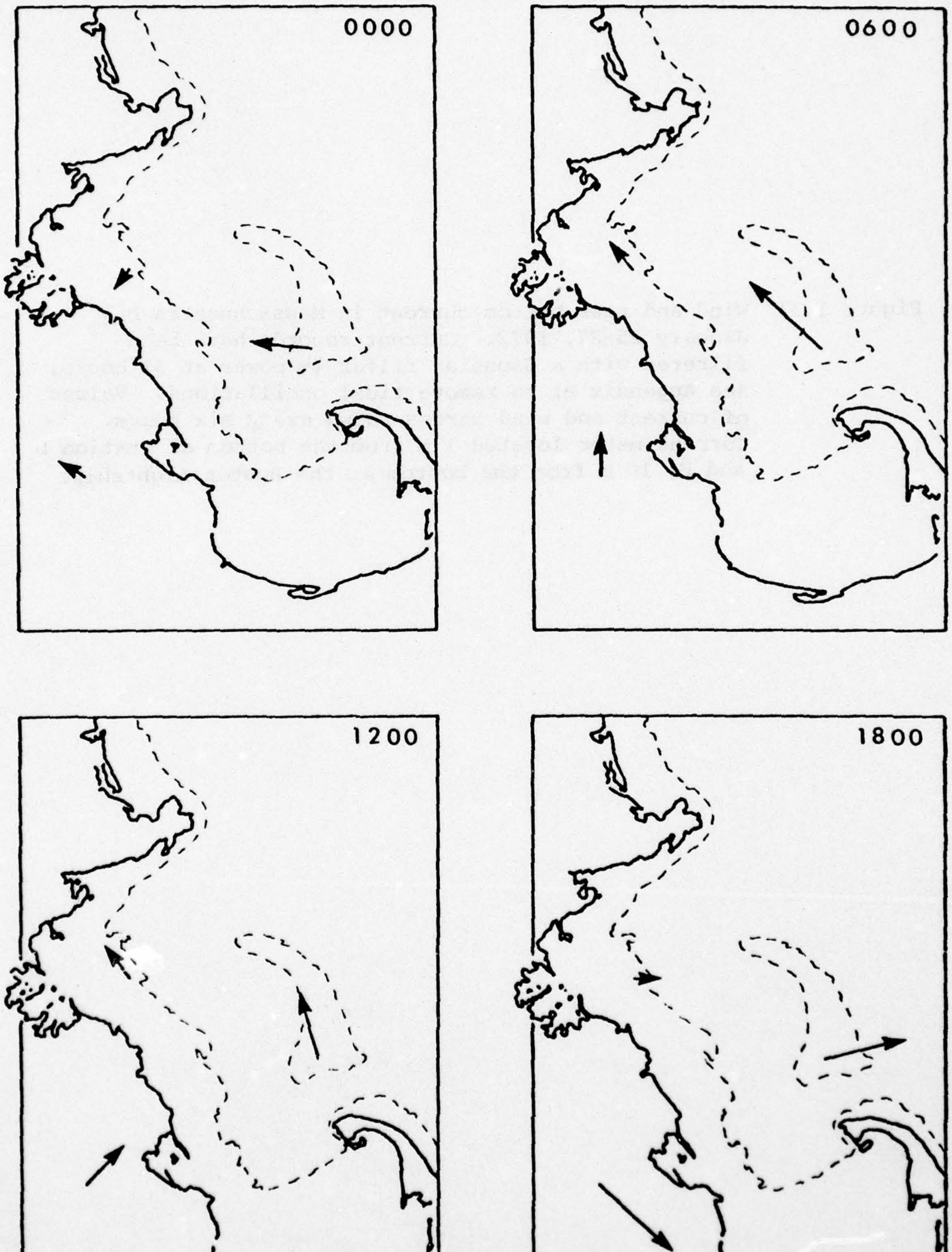


Figure 1.11

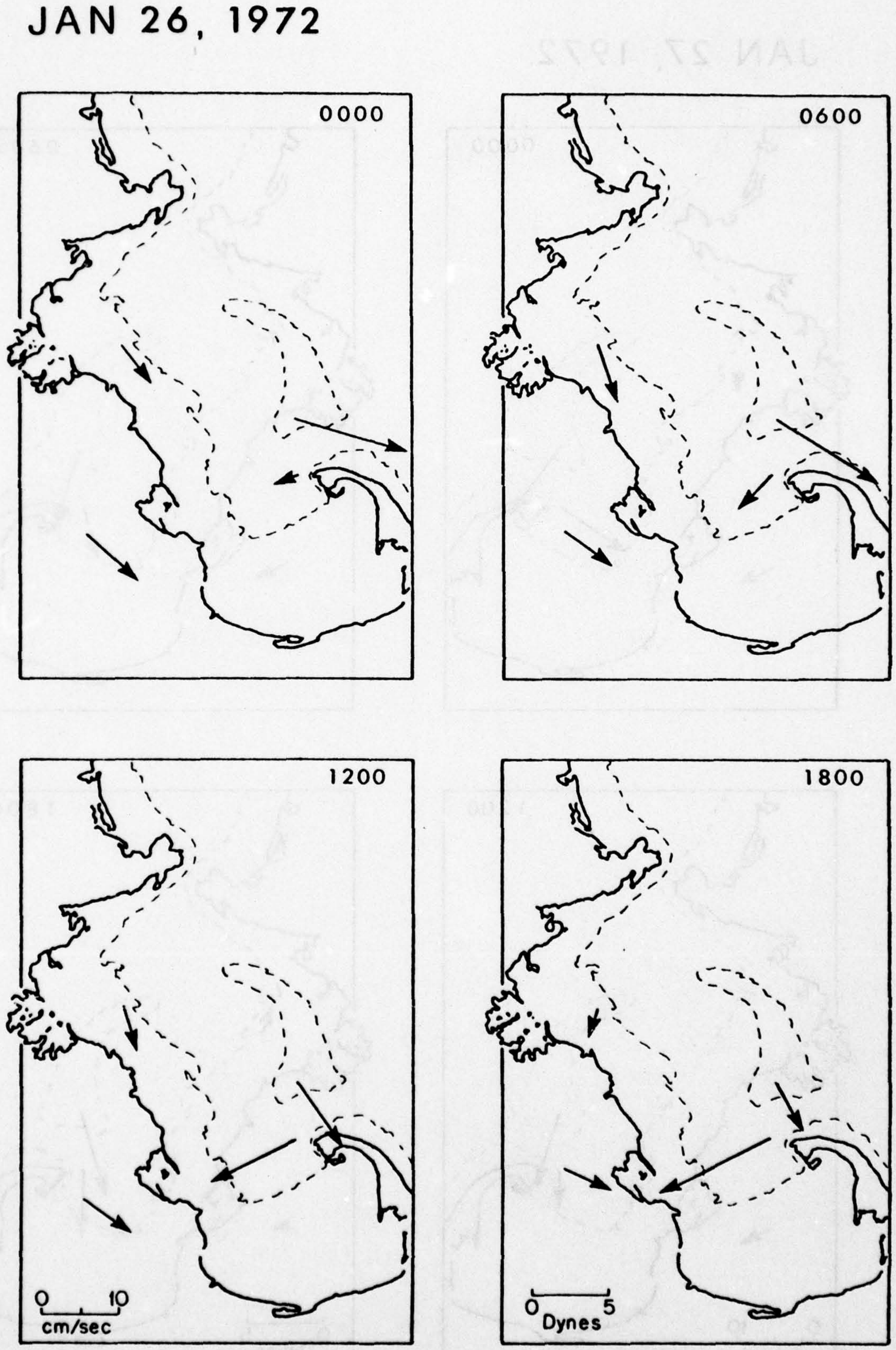


Figure 1.11

JAN 27, 1972

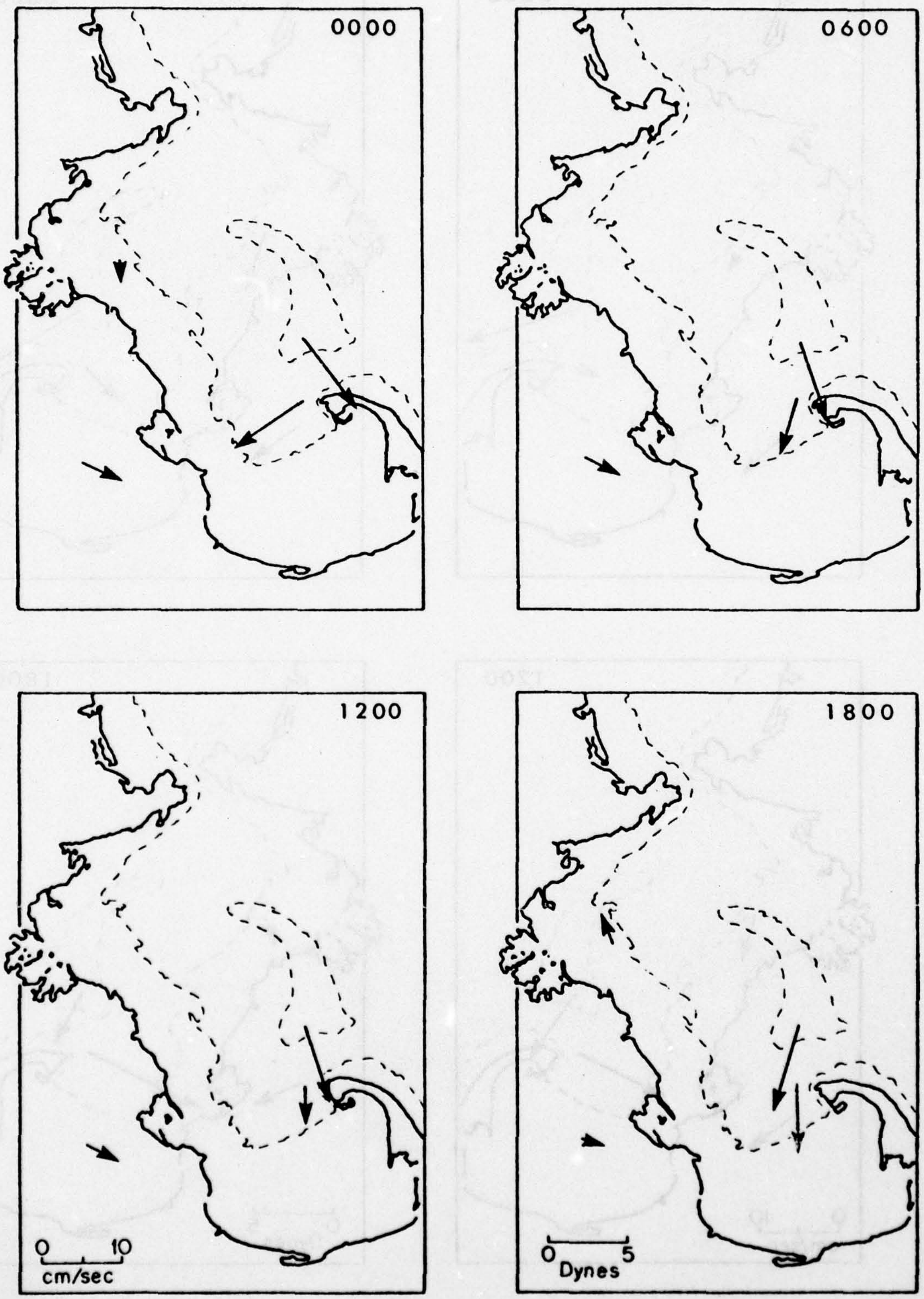


Figure 1.11

Sea level difference between Sandwich and Boston (figure 1.10a) shows a slight setup on the 24th and early on the 25th in response to the southerly wind. The strong northwest wind on the 26th and 27th produced a large setup, with Sandwich higher than Boston by approximately 10 cm. As the wind rotated through north to south, the absolute level at Boston dropped 65 cm; the outflow is clearly seen at station D at 1800 hr on January 25. The inflow on the 27th is somewhat more gradual and is not obvious from the current record at station D.

b. February 2 - 5, 1972 (Figure 1.10, 1.12)

Winds over Massachusetts Bay were light on the morning of the 2nd, gradually became easterly in the afternoon and continued from the east on the 3rd, then changed to westerly on the 4th. On the 5th, winds were from the northwest throughout the day. The bottom current at station D on the 2nd was to the west into the Bay as sea level at Boston and Sandwich rose. The flow, at least early on the 2nd does not appear to be locally wind driven. It is suggested that the flow at stations D and F was driven by a rise in sea level at the western end of the Gulf of Maine; wind over most of the Gulf on the morning of the 2nd was to the southwest, but was only to the southwest at Boston in the afternoon. The flow at station D became more parallel to the western coast of the Gulf on the 3rd as the rise in sea level at Boston diminished. Flow at station F late on the 3rd and early on the 4th was in the direction of the pressure gradient caused by the slight setup of the bay to the north. On February 4th the wind changed from southeast to south to southwest to northwest. The bottom flow at station D closely followed the rotating wind; sea level in the Bay fell below mean level as the bottom flow at station D passed through east. The current at station F rotated from south to north as the wind

Figure 1.12 Wind and near bottom current in Massachusetts Bay February 2-5, 1972. Current records filtered with a Gaussian filter (½ power 33 hours, see Appendix B) and plotted every six hours. Current meters located 1 m from the bottom.

FEB 2, 1972

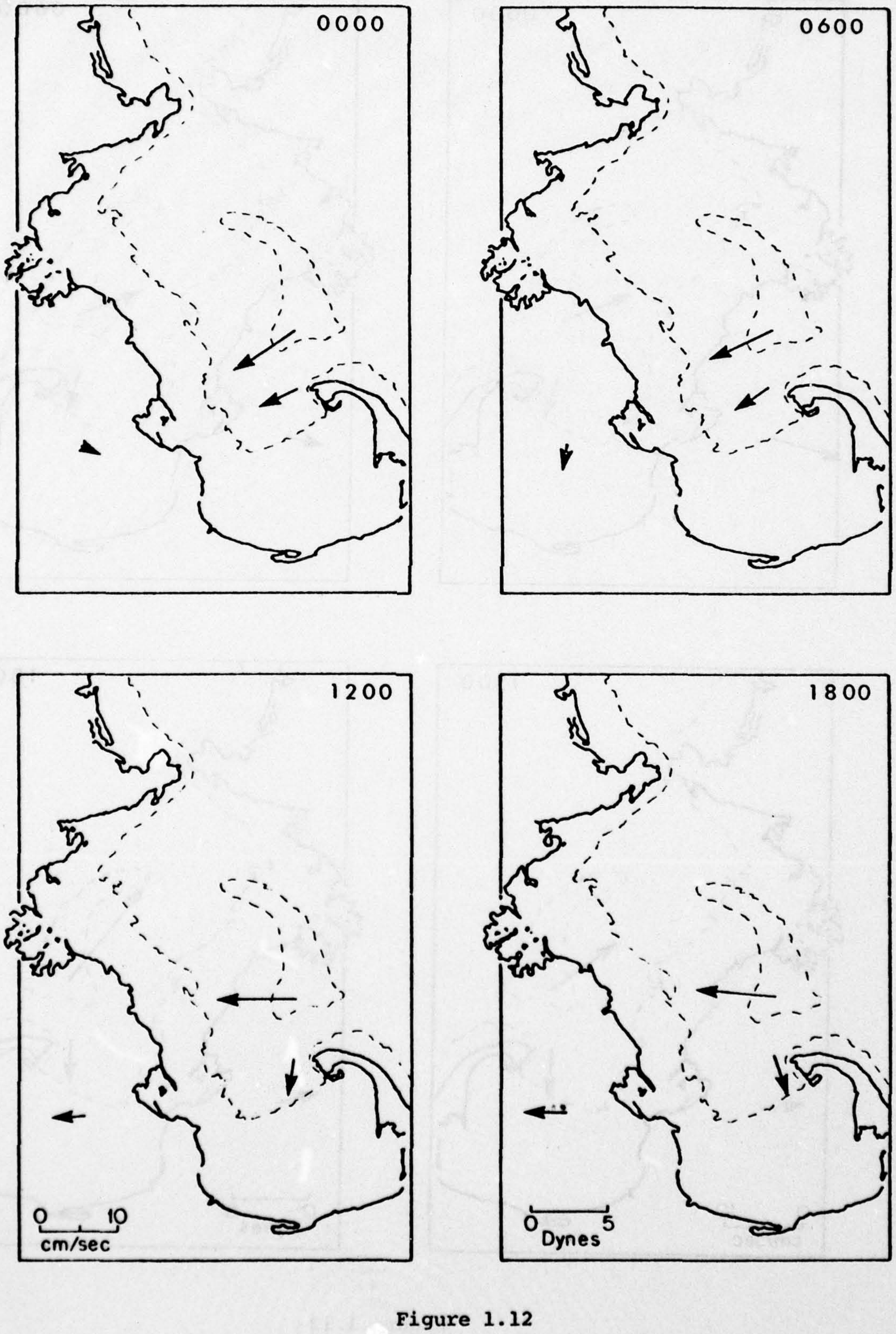


Figure 1.12

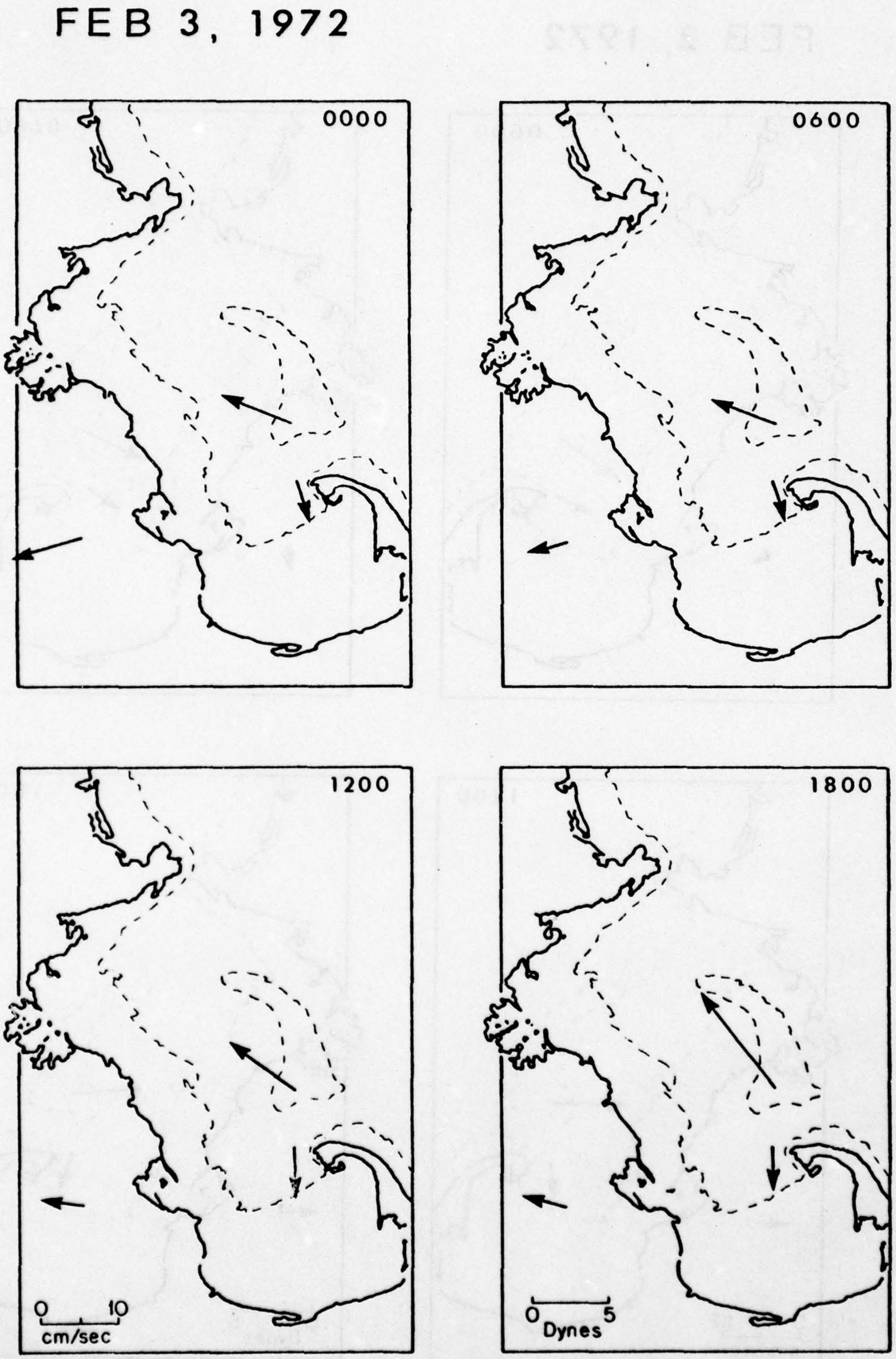


Figure 1.12

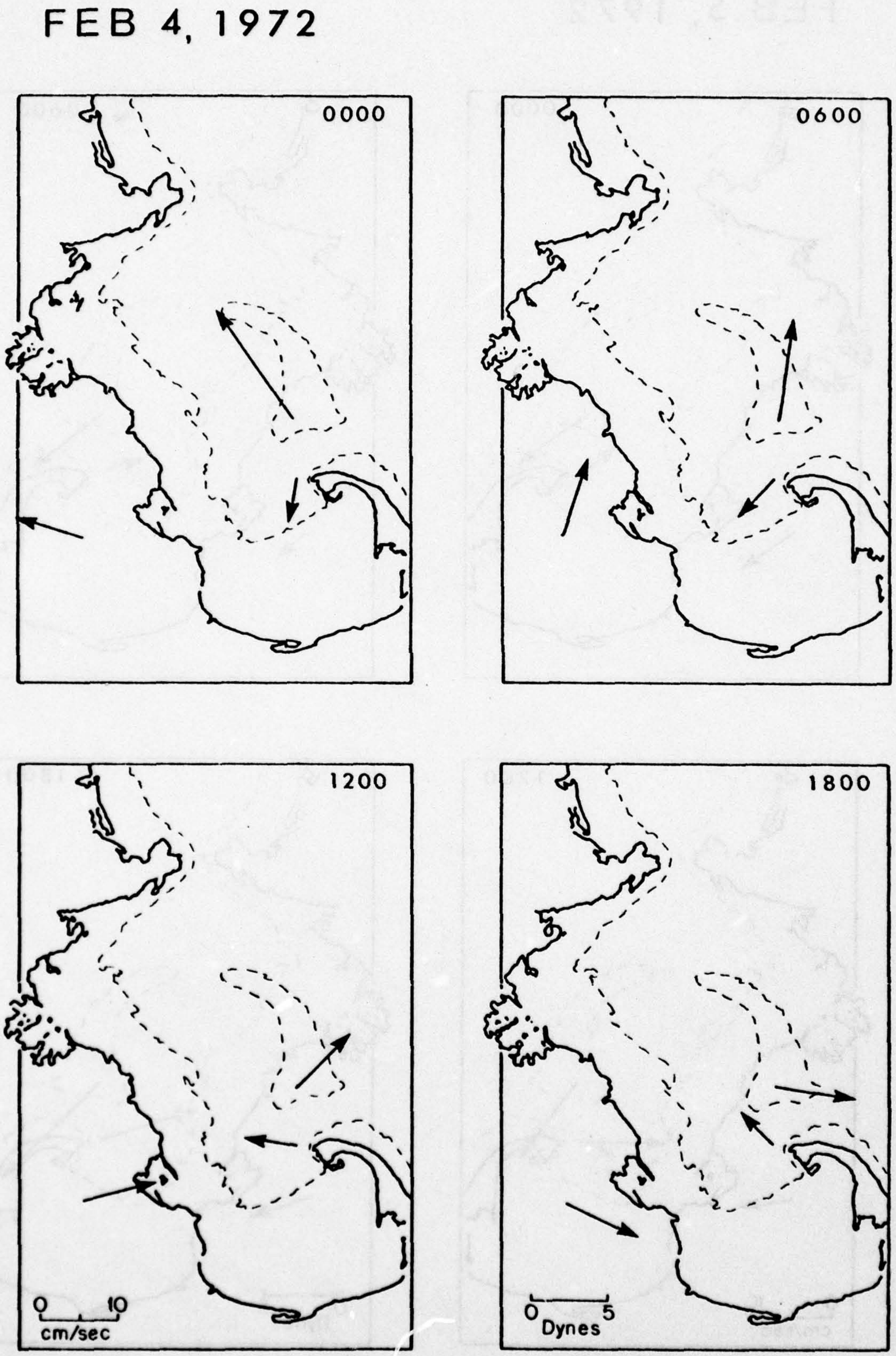


Figure 1.12

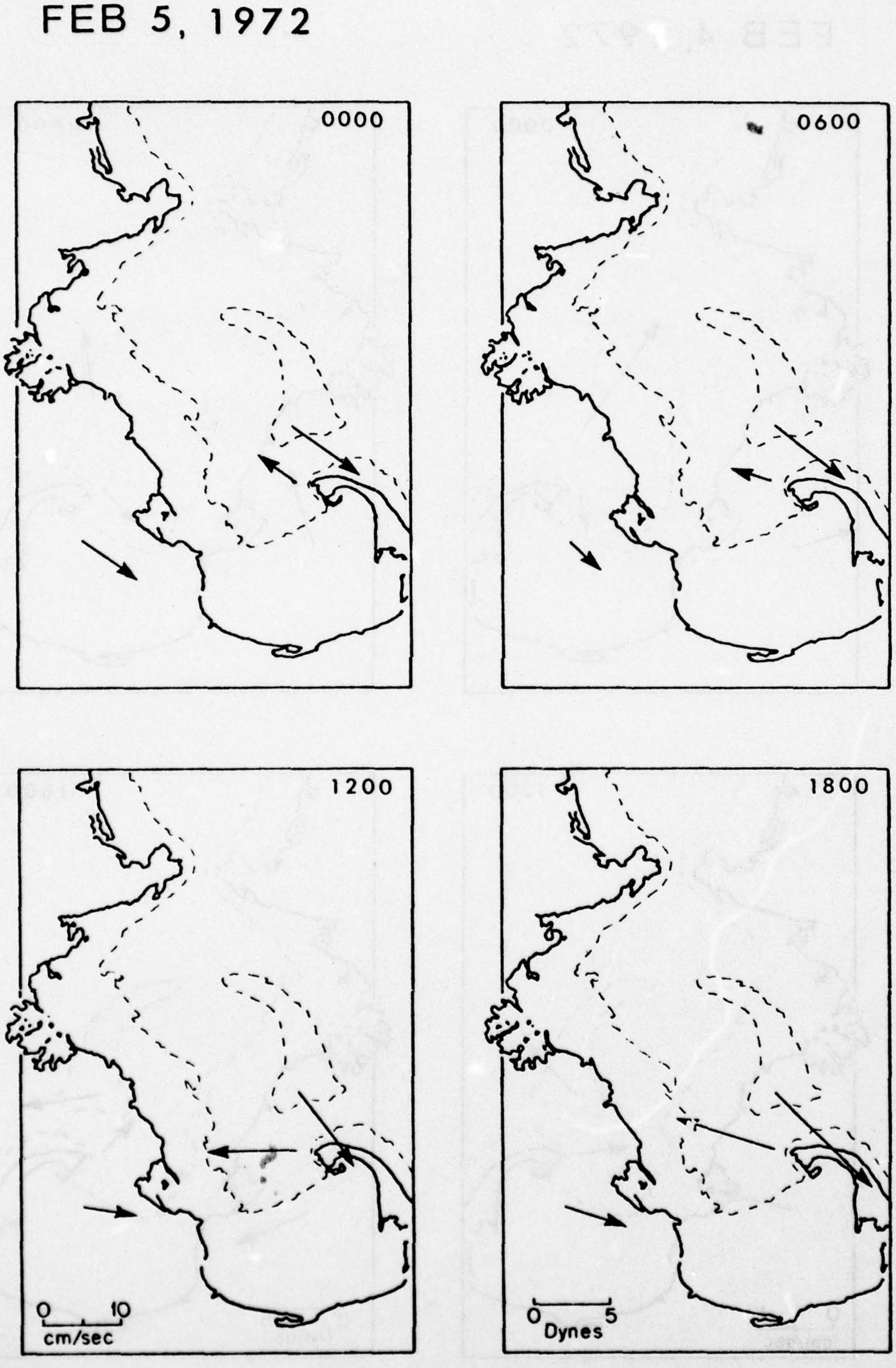


Figure 1.12

shifted. The strong northwest wind continued on the 5th with little change in the bottom flow pattern.

c. February 18 - 21, 1972 (Figure 1.10, 1.13)

On February 18, winds over Massachusetts Bay and the Gulf of Maine were from the east and southeast. Flow at station B and at the Boston Lightship was southerly parallel to shore while flow at station C in the deep central basin was to the east, opposite to the direction of the wind. The wind stress increased from the southeast during the morning of the 19th; in response the velocities at all stations increased, but the flow pattern remained basically unchanged. Late on the 19th and early on the 20th the wind rotated from southeast to northwest. Flow at station C remained to the south for the first twelve hours of the northwest wind, but then gradually rotated to the northwest after sea level at Boston reached equilibrium. Flow at station B was to the northwest at midday on the 20th and became northeasterly on the 21st. The flow pattern at 0600 on the 21st strongly suggests a double gyre flow pattern, with the northwest flow at station C feeding both shore parallel flows at station B and at the Boston Lightship.

3. Bottom Wind Driven Circulation Pattern

Although the spatial coverage of the currents during any one storm is sparse, a composite picture of the bottom flow pattern in the basin in winter for several wind directions can be developed from measurements made at different times but under similar wind conditions (figure 1.14). The composite response suggests near bottom flow in the shallow parts of the Bay parallel to the wind and flow opposite to the wind in the deep central basin. Nearshore, flow is parallel to the coast. Flow on Stellwagen Bank is almost always in the direction of the wind, while

Figure 1.13 Wind and near bottom current in Massachusetts Bay February 18-21, 1972. Current records filtered with a Gaussian filter (½ power at 33 hours, Appendix B) and plotted every six hours. Flow at station B is shown only occasionally because of a failure in the timing circuitry of the current meter. Measurements at stations B and C 1 m from the bottom, measurements at Lightship 10 m from the bottom. Wind stress at both Boston Lightship and Logan Airport shown on Feb. 19, when direction of the two observations differ significantly. Scale for Logan stress is 0-2.5 dynes.

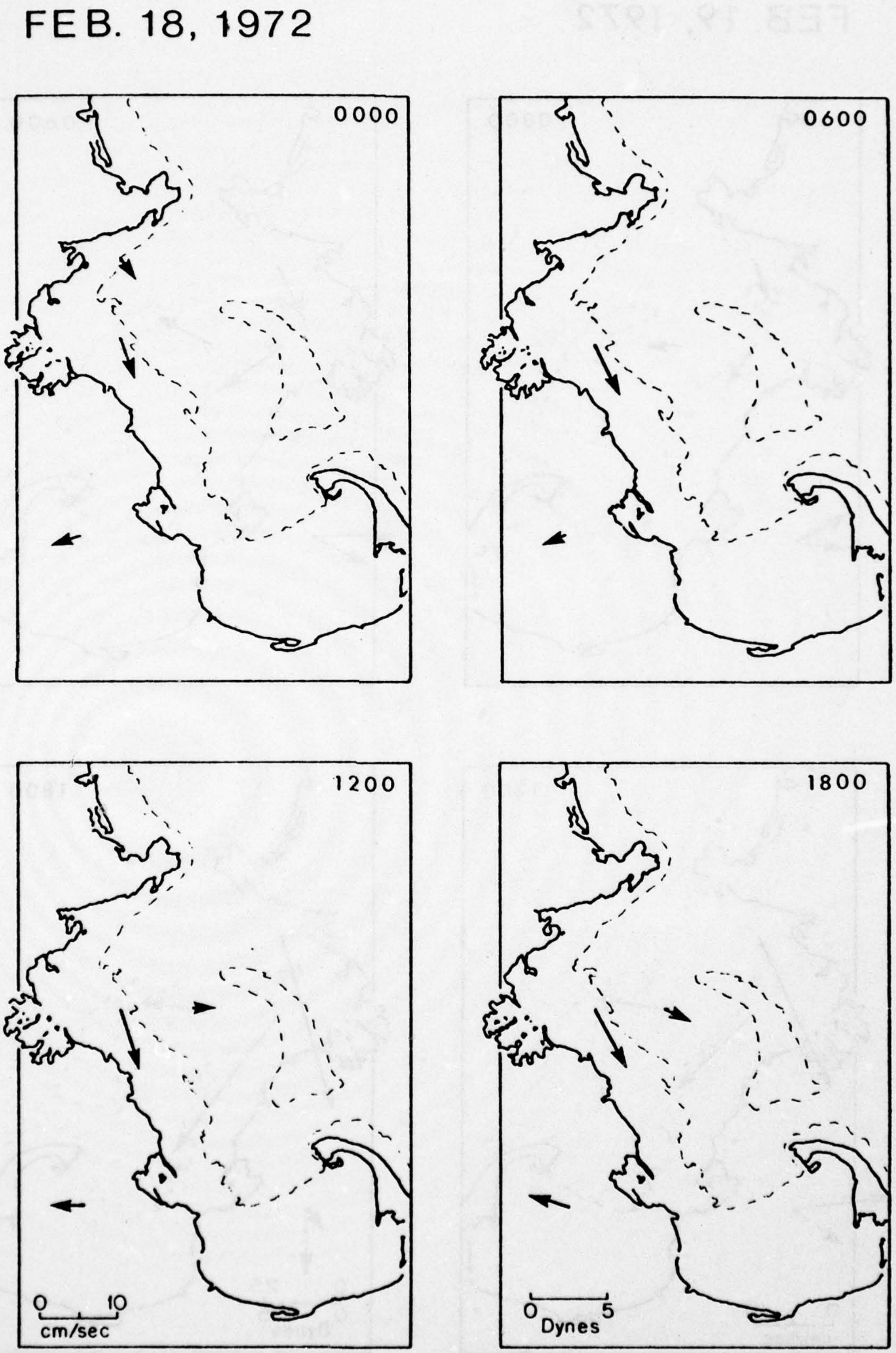
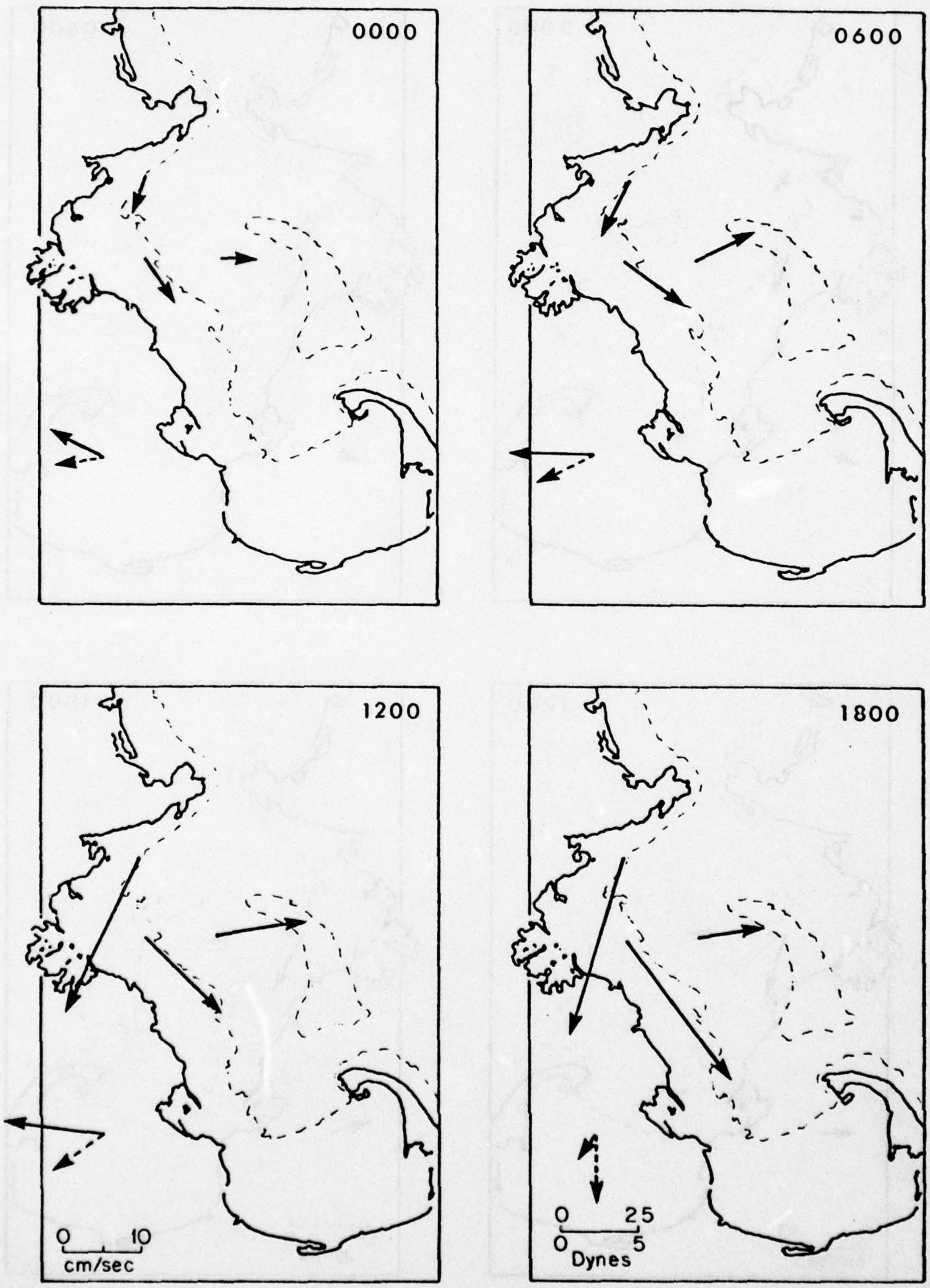
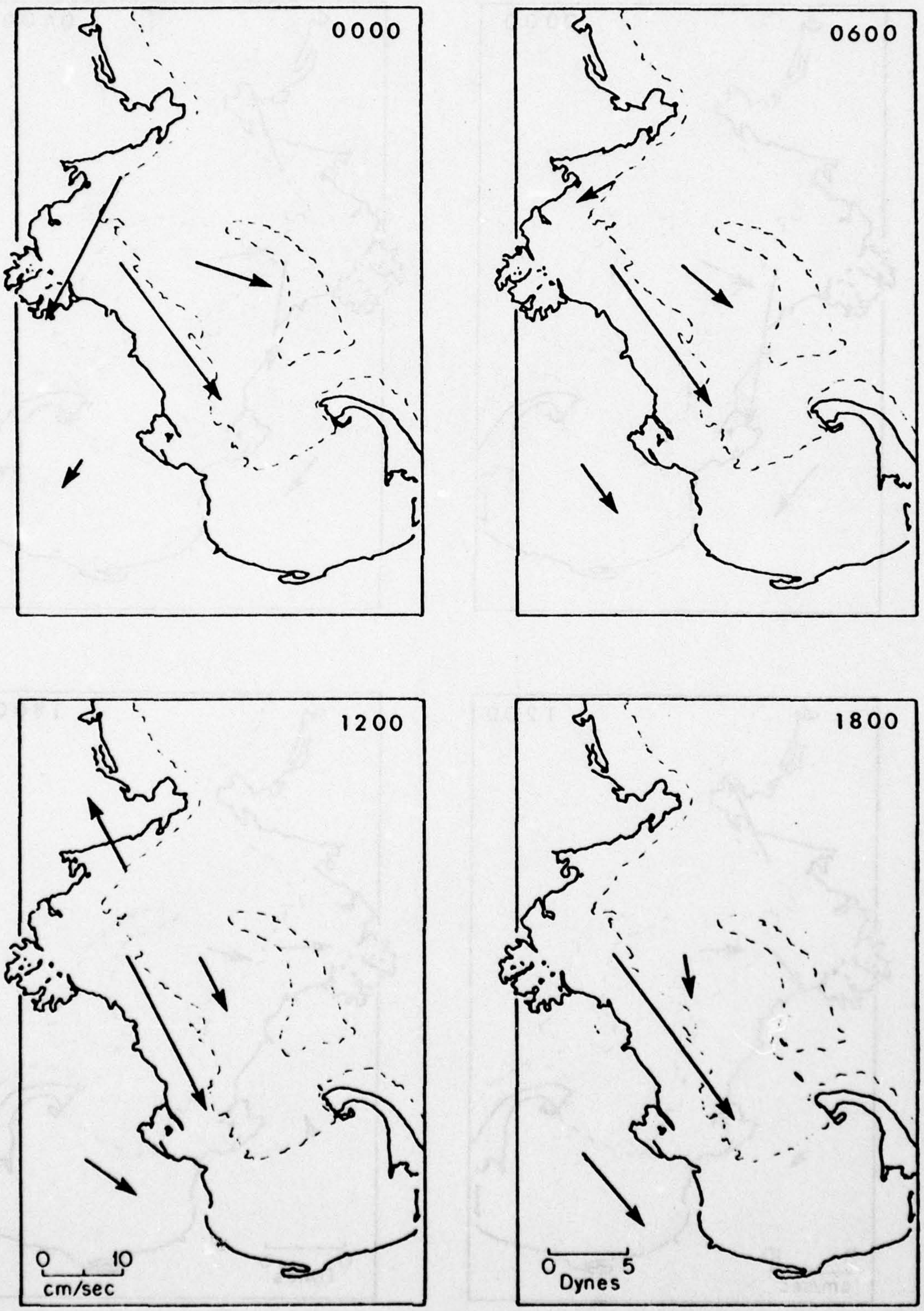


Figure 1.13

FEB. 19, 1972



FEB. 20, 1972



FEB. 21, 1972

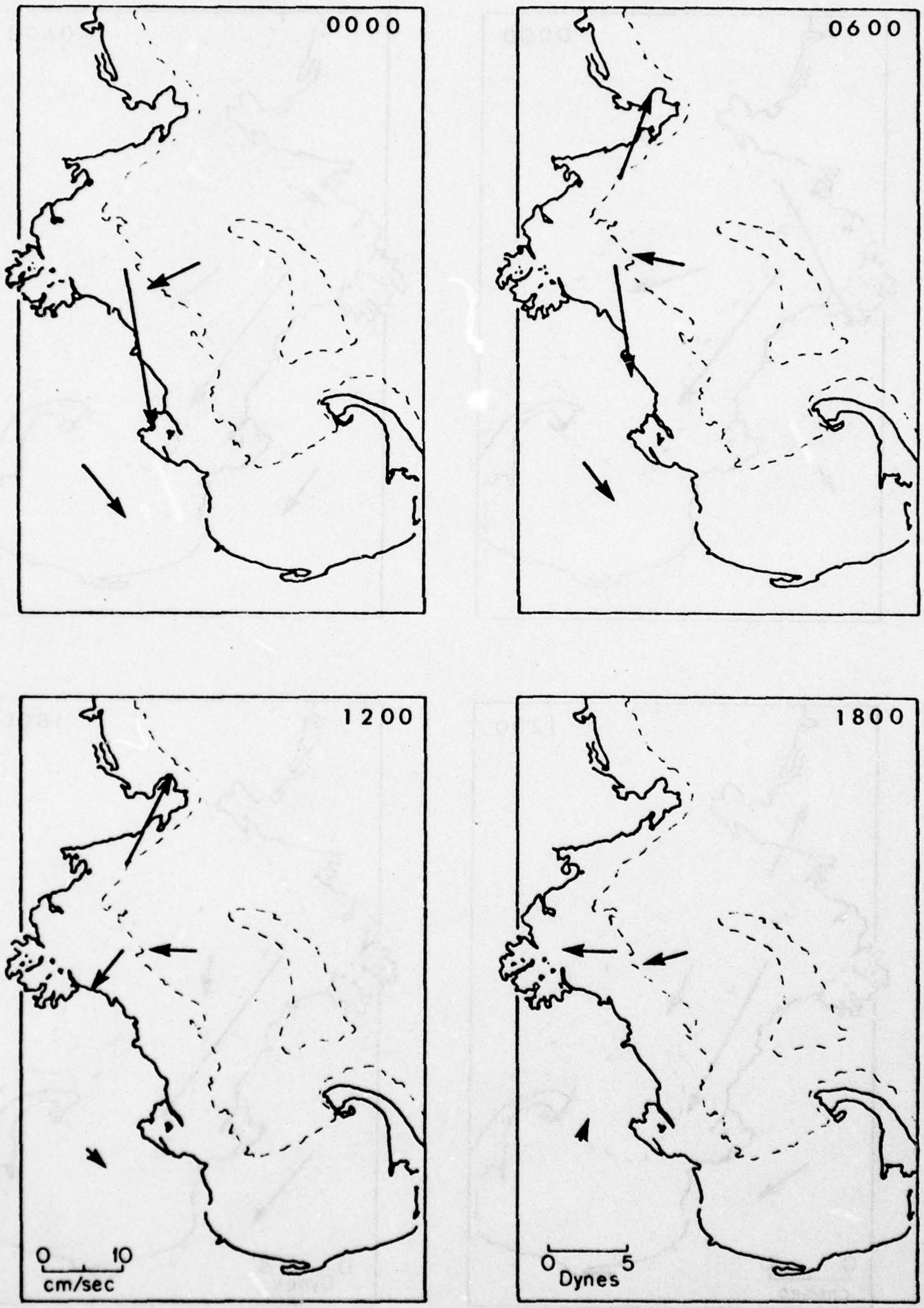
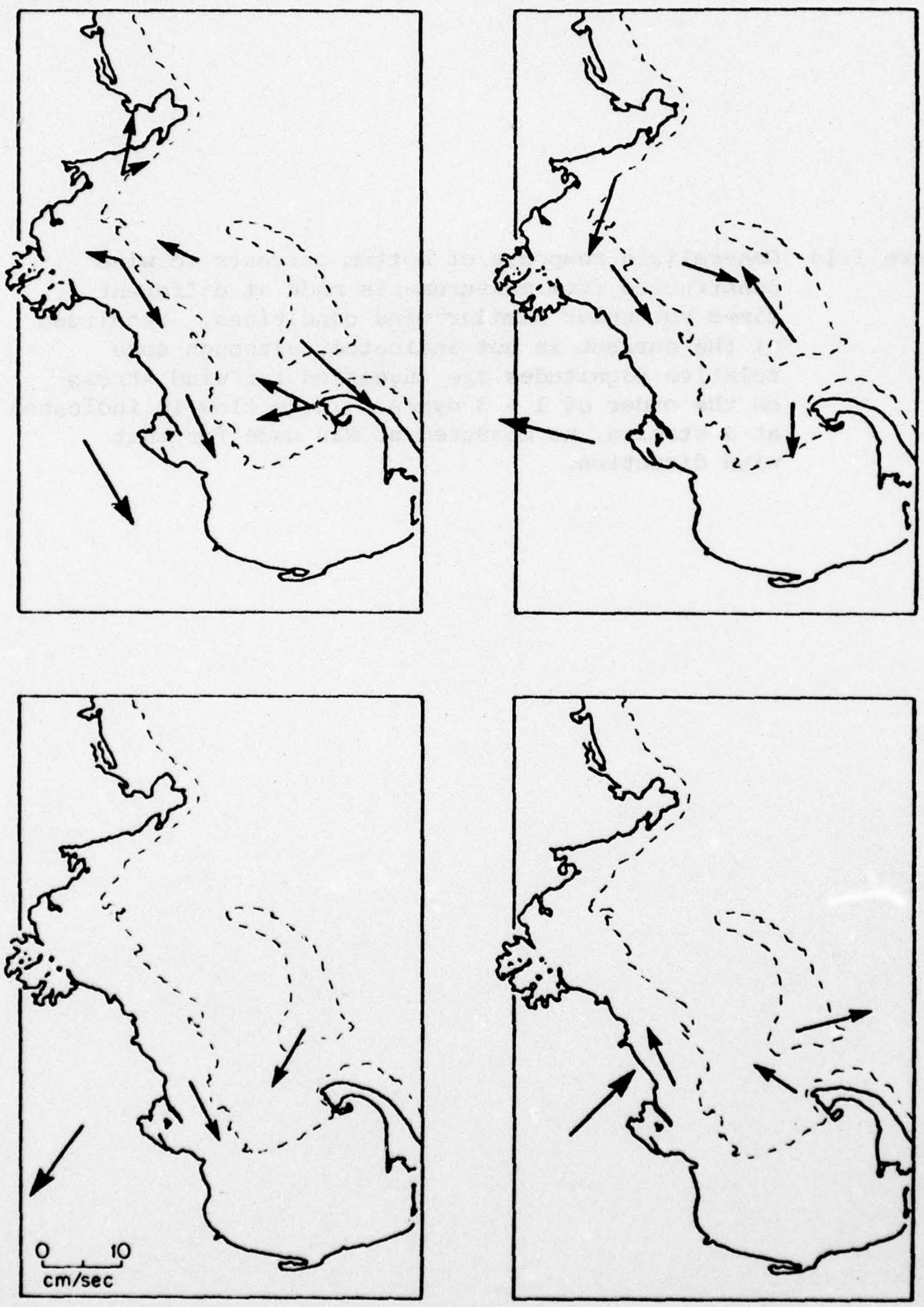


Figure 1.14 Generalized response of bottom currents to wind constructed from measurements made at different times but under similar wind conditions. Magnitude of the current is not indicated, although some relative magnitudes are suggested for wind stress on the order of 1 - 3 dynes. If no flow is indicated at a station, no measurement was made for that wind direction.

BASIN RESPONSE (BOTTOM)



flow in Stellwagen Basin for a northwest or southeast wind is nearly opposite to the wind direction. The observed northwesterly flow at Boston Lightship with a northwest wind appears to accept the flow from the deep basin and feed the northeast flow at station B, or the southerly flow at station E. Similarly, the southerly flow at Boston Lightship associated with a southeast wind may feed the easterly flow in the central portion of the basin, and not continue down the coast.

In summary, the response of sea level and of the bottom currents in Massachusetts Bay in winter to strong wind stress is as follows:

(1) A sea surface setup in the direction of the wind. Superimposed on the setup are changes in the absoluted level of the Bay controlled primarily by the response of the Gulf of Maine. Local setup is established in less than 1 hr; absolute changes require 6 - 12 hours.

(2) The bottom current is coherent over basin scales during strong wind events, with flow in the direction of the wind in the shallow parts of the basin and opposite to the wind in the deep basin. Flow is more complicated near the ends and corners of the Bay where the current must adjust to the coast. On Stellwagen Bank, flow is in or out of the basin as sea level adjusts to the level of the Gulf of Maine.

(3) The local bottom wind driven current is established approximately 12 hours after the wind stress is applied and remains basically unchanged even if the wind stress lasts as long as 24 hr. In some cases, however, the time to establish the flow pattern is as long as 18 - 24 hours, particularly in the corners of the basin, and there is a slight modification of the flow pattern with time. (The rapid adjustment of winter wind driven currents is also observed on the New England continental shelf, and is investigated in Chapter 3.)

F. Summary and Conclusions

The bottom currents in Massachusetts Bay have been monitored over a one year period in areas of different sediment types, bottom roughness and depth. With the reservations about stress estimates, critical erosion stress and the effects of bioturbation, it appears that the bottom sediments are in equilibrium with the bottom currents: the observed currents are not sufficient to move existing material regularly, except possibly on Stellwagen Bank and in the shallow nearshore regions. The speed distribution is primarily determined by the strong tidal currents.

In the winter, the net near bottom current is dominated by strong wind events. Despite the limited nature of the data and the only partially closed geometry of the Bay, there is indication of a two gyre flow pattern, with flow in the direction of the wind in the shallow areas and opposite to the wind in the central basin. Sea level is controlled both by local wind in the Bay and by the response of the Gulf of Maine; inflow and outflow over the shallow bank at the mouth of the Bay is observed as sea level in Massachusetts Bay adjusts to the Gulf. The net bottom flow pattern could be important in redistributing fine material from the shallow bank and nearshore areas, where occasional incipient sediment motion occurs, into the deep basin.

CHAPTER II

THE SPRING RUNOFF IN MASSACHUSETTS BAY

A. Introduction

This chapter investigates the currents during the period of the spring runoff in Massachusetts Bay, and the relation of the current pattern to the observed density field. Recent investigations have shown the importance of wind, topography, and thermal structure in determining the circulation on the continental shelf, shallow seas, and the Great Lakes (for example, Bennett, 1974; Csanady, 1973a,b, 1974 , and others). A somewhat neglected aspect (with the notable exception of Stommel and Leetmaa (1972)), or assumed to be less important, has been the circulation driven by the discharge of fresh water at the coast by rivers. This study is concerned with river influenced currents in regions somewhat distant, on the order of 10-20 km from a river mouth, but not distant enough to consider the in-flow as in Stommel and Leetmaa's model. The major discharge of fresh water from rivers in temporate regions is from the spring runoff; typically fifty percent of the yearly streamflow occurs in March, April and May. Because of this major increase in river flow and the generally weak wind stress, density gradients associated with river discharge may have an important influence on the currents in nearshore areas in spring, even those distant from major rivers.

Studies of rivers as they discharge into the sea fall into two broad categories: studies of the surface plume of fresh water near the river mouth (the near field), [for example, studies by Wright and Coleman (1971) of the Mississippi River, and by Garvine (1974a) and Garvine and Monk (1974) of the Connecticut River], and studies of the large scale salinity distribution associated with the river discharge far from the initial mixing zone (the far field), [for example, studies by Barnes et al (1972) of the Columbia River and by Gibbs (1970) and Ryther et al (1967) of the Amazon River]. The studies show a major influence of river discharge on the surrounding ocean. For example, freshening associated with the Amazon and Columbia Rivers extends several hundred kilometers seaward in a large plume. The shape and position of the plume may depend on the local currents, winds, and the river discharge; the position of a river plume with respect to the river mouth may vary on short time scales, for example due to tidal currents (Garvine, Connecticut River), or on a seasonal time scale due to seasonal large scale ocean currents (Barnes, Columbia River). The studies generally interpret the surface plume as passively advected on the prevailing currents. Garvine (1974a)did measure currents in the frontal zone of the plume, but none of these studies included direct current measurements of the large scale currents directly associated with the source of buoyancy.

Models of river discharge into the ocean have primarily been of the near field, or of the small scale structure of the river plume (Garvine, 1974b;Takano, 1954; Wright and Coleman, 1971) and have generally ignored the effects of rotation and the far field circulation, with two exceptions. One model of the medium to far field circulation driven by river discharge in a rotating system is of steady river flow distributed evenly along a coastal boundary. The discharge will result in fresher water nearshore, with the strength of the associated horizontal density gradients dependent on the magnitude of the fresh water discharge and on the extent of the vertical and horizontal mixing. Because of the thermal wind relation, offshore horizontal gradients will tend to produce flow parallel to shore, to the right (looking out from shore) in the northern hemisphere. Such a model is presented by Stommel and Leetmaa (1972) for the U.S. east coast continental shelf.

A second relevant model (Csanady, 1971) is of the inertial adjustment of a column of light warm water nearshore and a column of heavier water offshore. The two layer adjusted state is either a 'wedge-shaped' or a 'lens-shaped' thermocline with length scales of the internal Rossby radius. For the wedge-shaped thermocline currents are to the right in the surface layers and to the left in the bottom layer (looking offshore). In the lens-shaped thermocline flow in upper and lower layers is in the same sense, to the left on the inshore edge of the lens, looking offshore, and to the right on the outer edge. The model was developed to explain the thermally driven spring lake circulation, but the dynamics are similar if applied to coastal freshening by river flow.

The effects of river discharge on the currents in Massachusetts Bay will differ from these models in several ways: (1) the current pattern is three dimensional and varies over length scales of 10-20 km in the along-shore direction, (2) the major source of fresh water is not at the coast but to the north of the basin, (3) the major fresh water flow occurs as spring runoff and is thus not steady throughout the year, and (4) a summer thermocline gradually develops during the runoff period.

B. Background

Studies in the Gulf of Maine show a general freshening of the near-shore waters in the spring as a result of increased river runoff, and an increased local freshening adjacent to the large river systems (Meade, 1971; Graham, 1970; Colton, 1968; Bigelow, 1927). Associated with the nearshore freshening is a southerly flow of about 10 cm sec^{-1} along the western coast

(Bumpus, 1973; Graham, 1970; Bumpus and Lauzier, 1965), the western side of the Gulf of Maine gyre. The strength of the gyre is strongest in spring suggesting the importance of runoff, but interpretation of the gyre as a wind driven flow can also be made (Csanady, 1974).

Massachusetts Bay is a semi-enclosed basin, located on the southwestern end of the Gulf of Maine, 'downstream' from all the major rivers which discharge into the Gulf (figure 2.1). There are two river systems which affect the salinity distribution. A small system of several rivers, of which the largest is the Charles River, discharges directly into the Bay through Boston Harbor. This group of rivers will be referred to as the Charles River system. The spring runoff typically begins in late February and continues through March (figure 2.2a), with a maximum spring flow of approximately 30 $m^3\ sec^{-1}$. The second nearby source of fresh water is the Merrimack, a major river which discharges 30 km to the north of Massachusetts Bay. The peak spring runoff of approximately 500 $m^3\ sec^{-1}$ generally occurs in April, somewhat after the peak runoff from the Charles River system (figure 2.2b). The two fresh water sources are thus separated in time and space. On the average, the first inflow of fresh water to Massachusetts Bay is at the coast from the Charles River and is relatively small; the second source is much larger, dominated by the Merrimack, and is located to the north and east of the open side of Massachusetts Bay. In the spring of 1973, discharge was much larger than average and considerably more time dependent than the monthly average figures might suggest; the discharge over a five-day period may change by a factor of two during the peak runoff season.

During the period of the spring runoff a seasonal thermocline gradually develops in Massachusetts Bay. The water column in March and April is

Figure 2.1 Map of Massachusetts Bay showing study area (boxed), location of major river systems, current meter moorings, tide and meteorological stations. Smoothed 40 and 80 m contours indicate major topographic features.

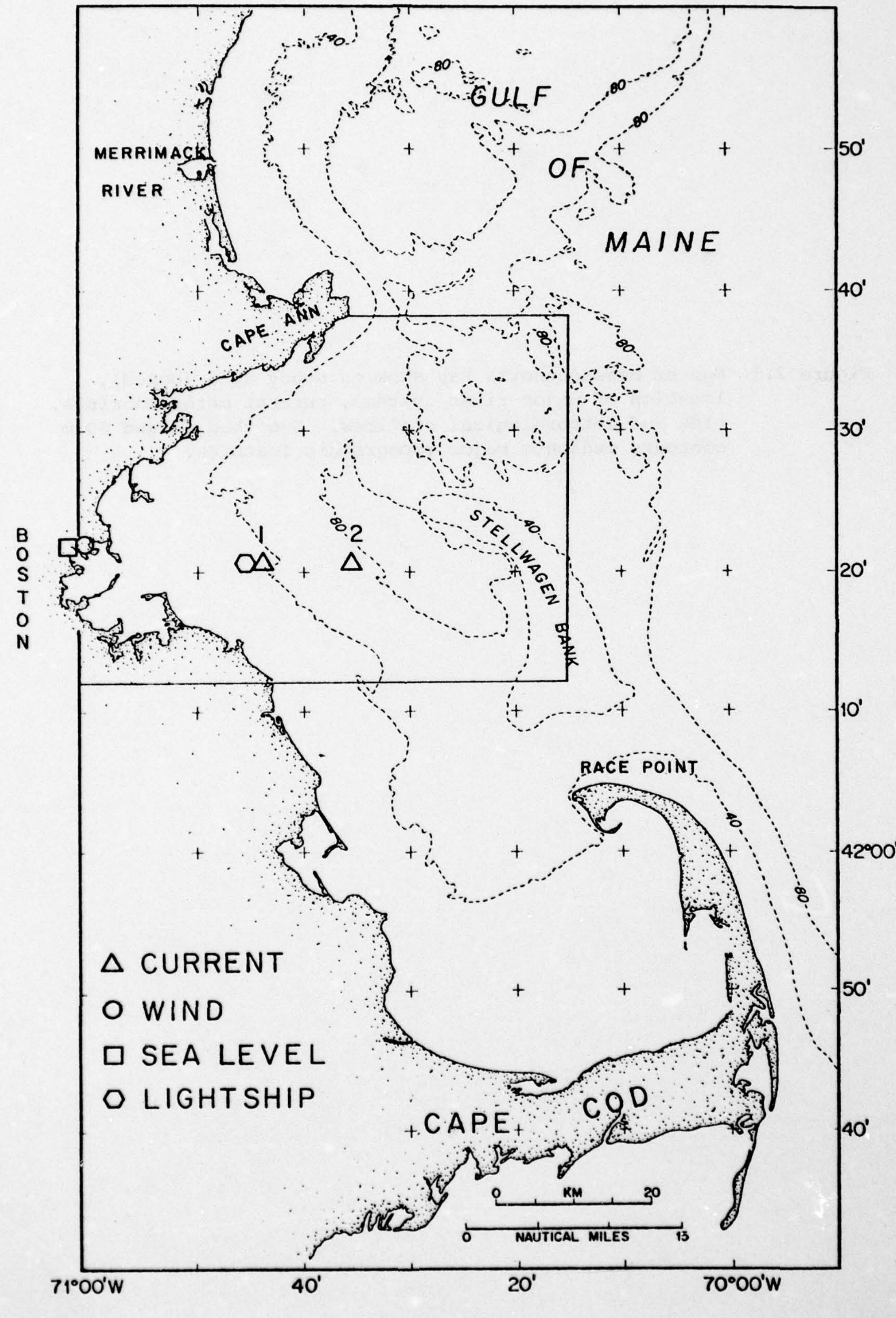
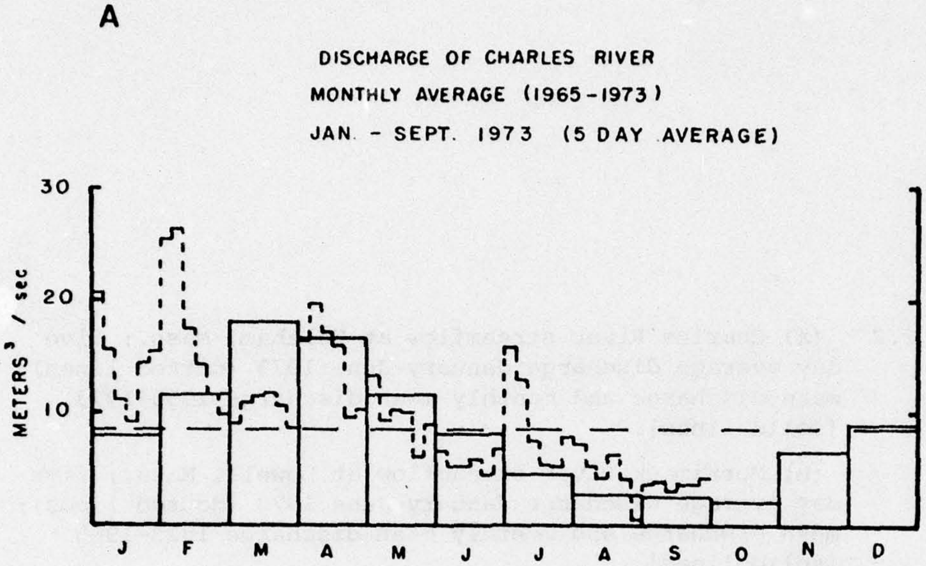

GULF
OF
MAINE
MERRIMACK
RIVER
CAPE ANN
BOSTON
STELLWAGEN BANK
RACE POINT
CAPE COD
1
2
80
40
△ CURRENT
○ WIND
□ SEA LEVEL
⬡ LIGHTSHIP
0 KM 20
0 NAUTICAL MILES 13
50'
40'
30'
20'
10'
42°00'N
50'
40'
71°00'W
40'
20'
70°00'W

Figure 2.1

Figure 2.2 (a) Charles River streamflow at Waltham, Mass.; five day average discharge January-June 1973 (dotted lines); mean discharge and monthly mean discharge 1956-1973 (solid lines).

(b) Merrimack River streamflow at Lowell, Mass.; five day average discharge January-June 1973 (dotted lines); mean discharge and monthly mean discharge 1923-1965 (solid lines).

Data from U.S. Geological Survey and UNESCO (1969).


A
DISCHARGE OF CHARLES RIVER
MONTHLY AVERAGE (1965-1973)
JAN. - SEPT. 1973 (5 DAY AVERAGE)
30
20
10
METERS³ / sec
J
F
M
A
M
J
J
A
S
O
N
D

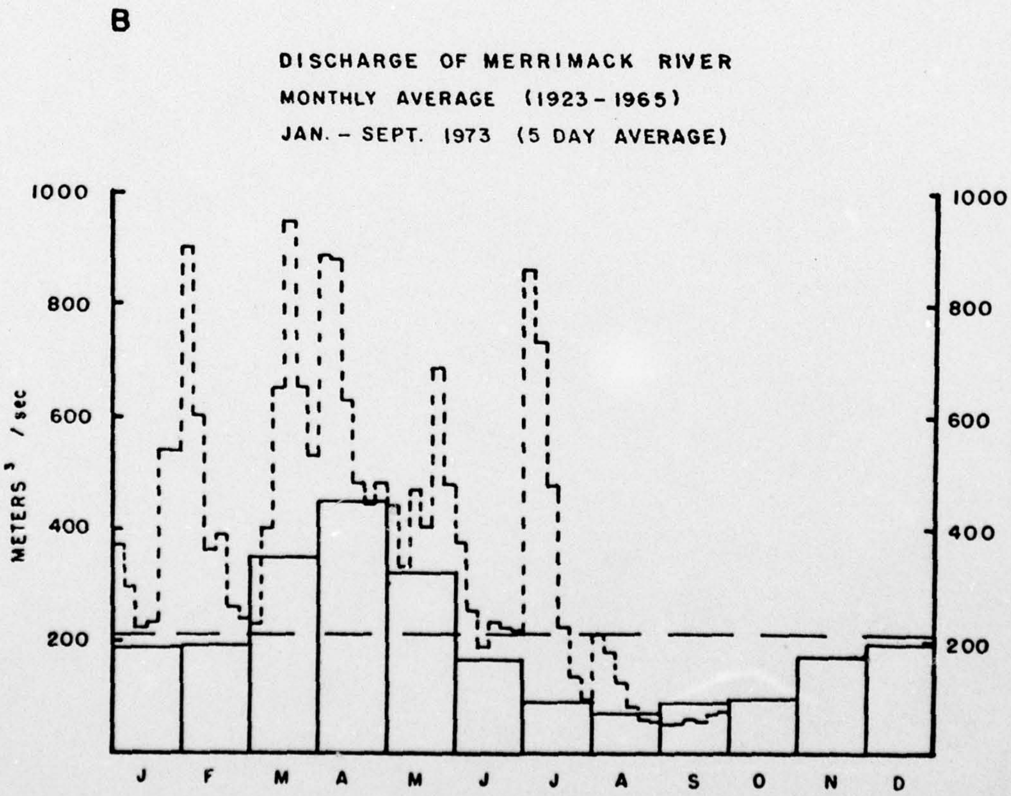

B
DISCHARGE OF MERRIMACK RIVER
MONTHLY AVERAGE (1923-1965)
JAN. - SEPT. 1973 (5 DAY AVERAGE)
1000
800
600
400
200
METERS³ /sec
J
F
M
A
M
J
J
A
S
O
N
D

Figure 2.2

isothermal, but by late May and early June the difference between surface and bottom temperature is 5-10°C (Bumpus, 1974). The early river discharge is into an unstratified water column, while the later discharge is into a more stable water column, both due to the surface heating and surface freshening. In addition, wind stress during the spring months is not large. Because of the increased stratification, downward mixing of the fresh water discharge may be significantly reduced as the spring progresses.

C. The Field Program

Hydrographic observations and moored current meter measurements were made during the spring runoff period of March-May, 1973. The objectives of the field program were to:

(1) describe the spatial and temporal variation of the salinity and density fields during the runoff period in the northern part of Massachusetts Bay (north of 42°10'),

(2) describe the major (subsurface) current patterns associated with the density field, particularly the competing influence of the local discharge of the Charles, and the expected freshening from the north and east.

1. The Current Measurements

Three Savonius rotor Richardson current meters were deployed in Massachusetts Bay from March 4 - June 3, 1973 (figure 2.1). Two instruments were located near the Boston Lightship at 20 and 30 m in water 35 m deep (station 1). A third current meter was located 15 km due west at a depth of 35 m in water 65 m deep (station 2).

Several considerations heavily influenced placement of the three available current meters. Current measurements at the Boston Lightship (figure 2.1) made in the spring of 1972 indicated strong currents

Figure 2.3 Progressive vector diagram of current at Boston Lightship 10 m from the bottom, spring 1972 (see figure 2.1 for location). The surface salinity dropped approximately 1‰ on April 25 and May 5, coincident with the major change in current flow. The bottom salinity did not change significantly, although the data is sparse (salinity from U.S. Coast Guard).

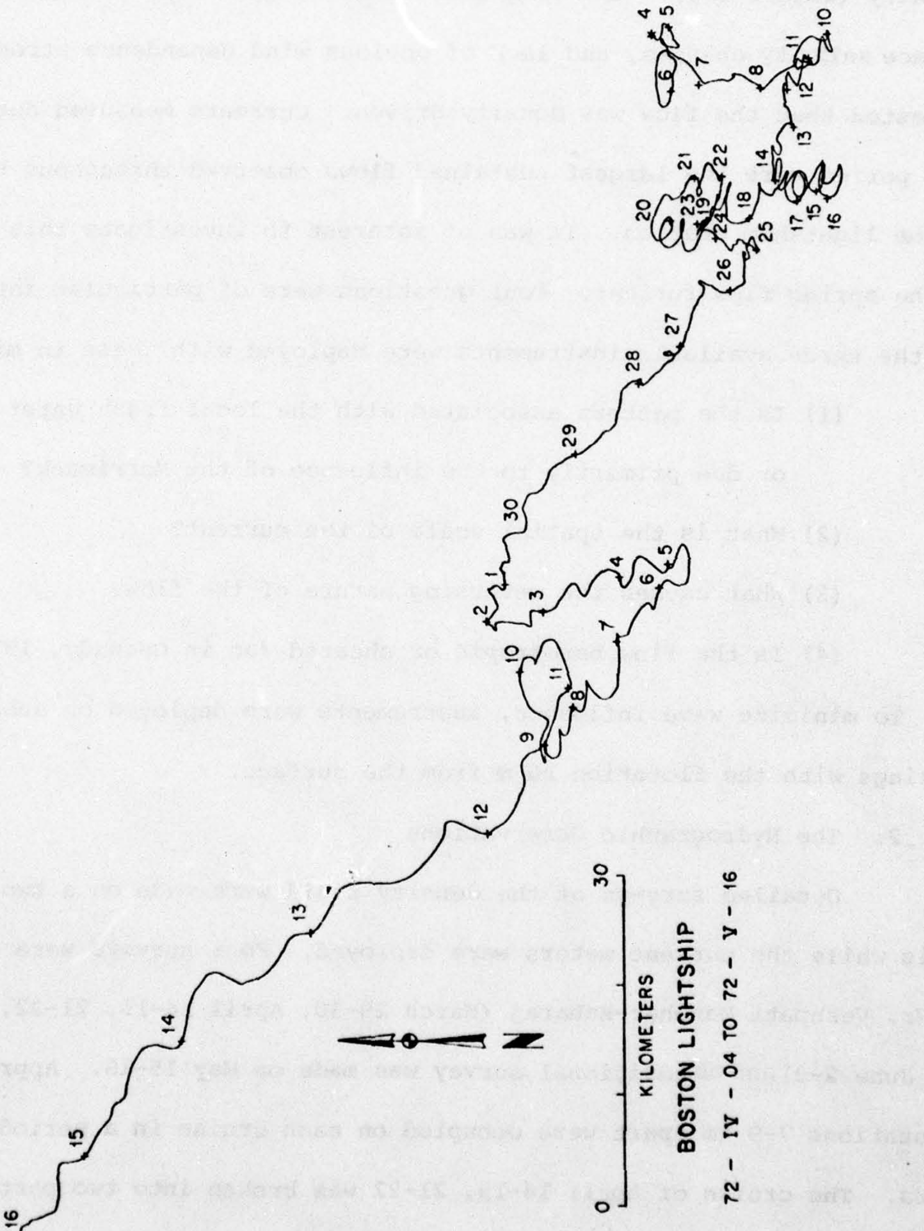


Figure 2.3

($10-20 \text{ cm sec}^{-1}$) aligned northwest-southeast parallel to the coast with periods of 5-10 days; abrupt changes of flow direction, occurring in less than one day, were often associated with a major change in the surface salinity (figure 2.3). The long time scale of the flow, the associated surface salinity changes, and lack of obvious wind dependence strongly suggested that the flow was density driven. Currents measured during this period were the largest sustained flows observed throughout the year at the lightship station. It was of interest to investigate this aspect of the spring flow further. Four questions were of particular interest, and the three available instruments were deployed with these in mind:

(1) Is the pattern associated with the local fresh water flow, or due primarily to the influence of the Merrimack?

(2) What is the spatial scale of the current?

(3) What causes the reversing nature of the flow?

(4) Is the flow barotropic or sheared (as in Csanady, 1971)?

To minimize wave influence, instruments were deployed on subsurface moorings with the flotation 20 m from the surface.

2. The Hydrographic Observations

Detailed surveys of the density field were made on a two-week basis while the current meters were deployed. Four surveys were conducted by Mr. Veshpati Manohar-Maharaj (March 29-30, April 14-15, 21-22, May 5-6, and June 2-3) and an additional survey was made on May 15-16. Approximately 30 stations 7-9 km apart were occupied on each cruise in a period of 24 hours. The cruise of April 14-15, 21-22 was broken into two parts due to bad weather. Unfortunately, due to equipment problems, the first hydrographic cruise was not made until late March.

An M.I.T. built C.T.D. was used to obtain vertical profiles of conductivity and temperature. Salinity was computed from conductivity and temperature. The C.T.D. was not well calibrated; a temperature dependent correction was made on the basis of surface C.T.D. readings and bottle samples (See Maharaj and Beardsley (1974) for details). Estimated relative errors in temperature, salinity, and sigma-t are T ± .1°C, S ± .1‰, sigma-t ± .1 . The relatively poor accuracy of the instrument will not affect the conclusions of this study.

D. Results

In this work, the main concern will be the spatial distribution of density, the changes of the distribution during the spring, and the relationship of the currents to the observed density field. Throughout the spring period the density is primarily controlled by the salt distribution, although temperature becomes an important factor later in the spring.

The salinity observations, with the exception of the cruise of May 15-16, are presented in Maharaj and Beardsley (1974) but they do not discuss the density or temperature distribution. A volumetric analysis of the amount of fresh water in the basin as a function of time compared favorably with the amount discharged by the Charles River system and Merrimack River, with suitable corrections for time lag and mixing. Maharaj and Beardsley conclude that the major freshening of the bay is due to fresh water which enters from the north (inferred from a distinct surface plume), and that within experimental error, all of the fresh water runoff of the Merrimack and Charles River systems can be accounted for in the freshening of Massachusetts Bay north of 42°25'N and west of 70°20'W. The volumetric aspects of the salt distribution will not be discussed further.

The current observations must be discussed in the context of the entire spring runoff period. Thus, although Maharaj and Beardsley (1974) have previously described the salt distribution, a brief review which emphasizes the changes in the spatial distribution of salinity and density is appropriate. The previously unreported observations of May 15-16, 1973 as well as a discussion of the changes in temperature and sigma-t is included. With the exception of the surface salinity distribution, the observations have been recontoured. With this hydrographic background, the current measurements will be presented and the two hydrographic cruises made while the current meters were deployed will be discussed in detail.

1. The Hydrographic Observations

a. The surface salinity distribution

The surface salinity distribution clearly shows the effect of the spring runoff (figure 2.4,a-e). The major features are summarized below:

March 29-30, 1973

The central portion of the basin is occupied by water with salinity greater than 31‰. The surface water nearshore, offshore (east of approximately 70°30'), and north of 42°30'N is slightly fresher, between 30-31‰. A salty core occupies the central basin with fresher water to the west, north, and east. The nearshore freshening is attributed to local runoff and the offshore freshening to the general freshening of the Gulf of Maine.

April 15-16, 21-22, 1973

The surface salinity over the surveyed area is between 30.2‰ and 30.8‰. The major accumulation of fresh water in the Bay occurred between March 29-30, and April 21-22.

Figure 2.4 Surface salinity distribution in northern part of Massachusetts Bay and east-west salinity distribution along 42°20'N, spring, 1973. Surface distribution in a,b,c,e drawn from continuous sampling along indicated track (redrafted from Maharaj and Beardsley, 1974); distribution in d drawn from surface samples at indicated stations. East-west section contoured from CTD stations at indicated locations, usually starting between 1 - 3 m. Slight discrepancies between east-west and surface contours result from the different data sources (CTD casts starting at approximately 2 m vs. continuous surface sampling) used to draw the contours.

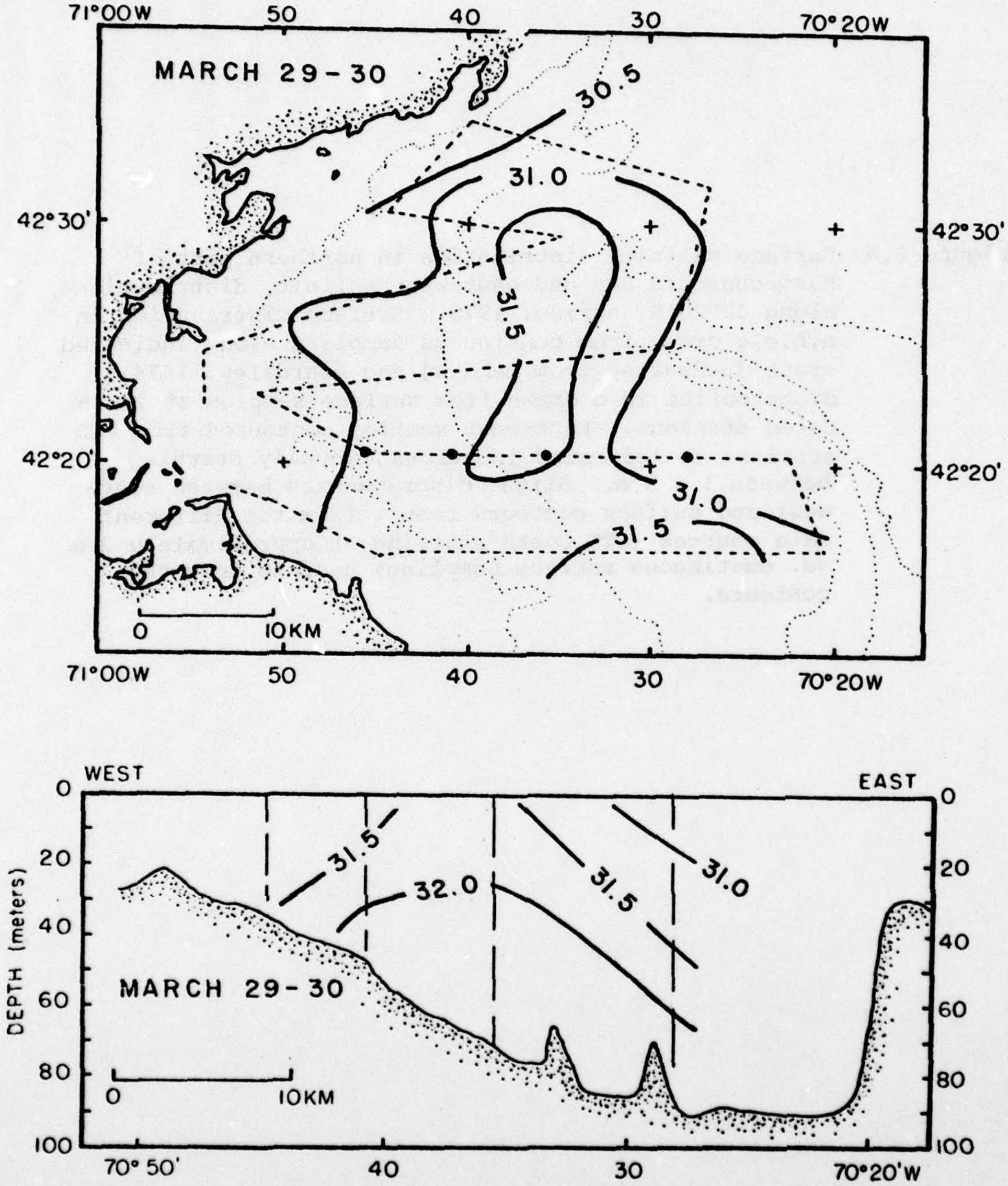
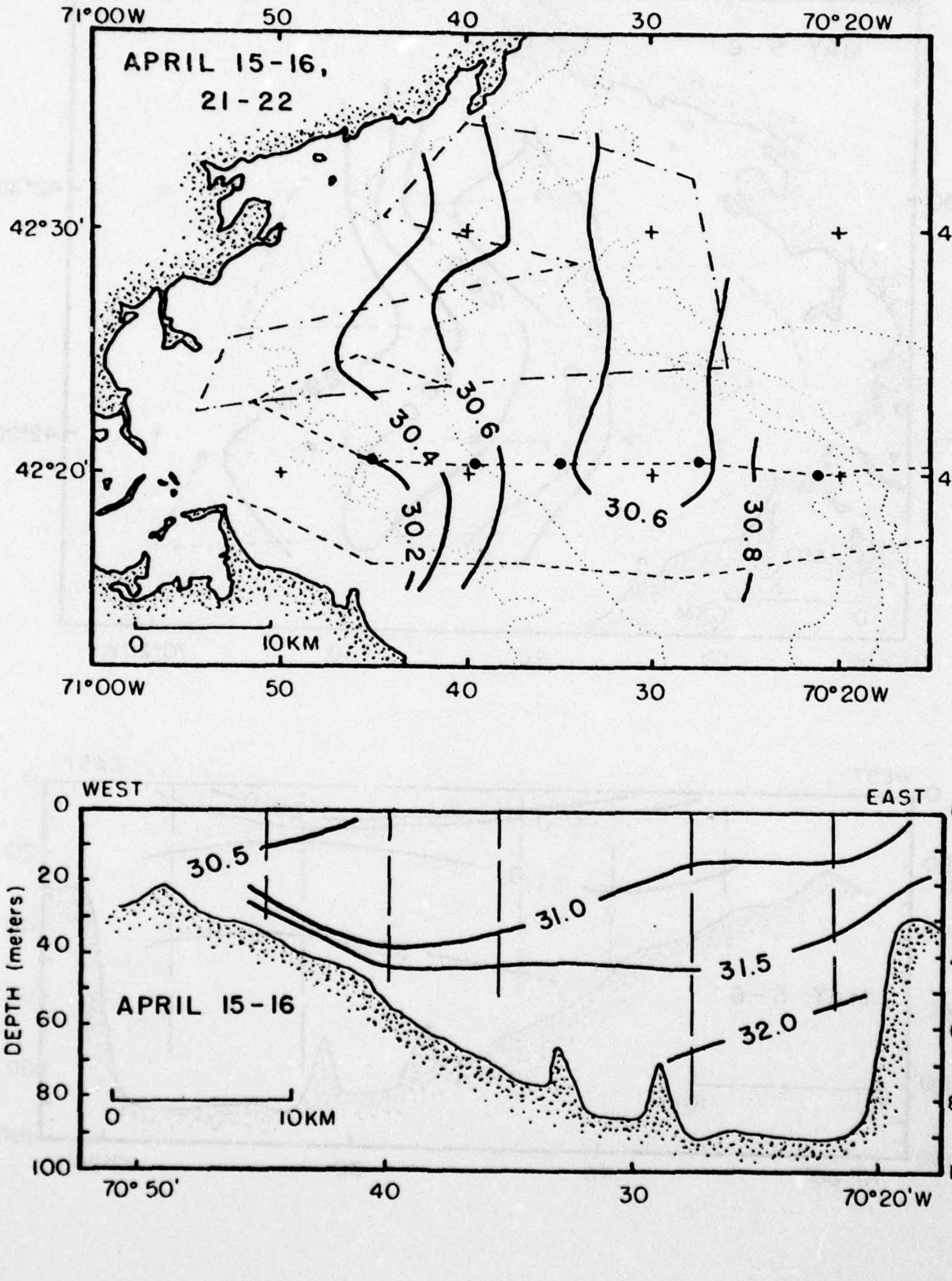
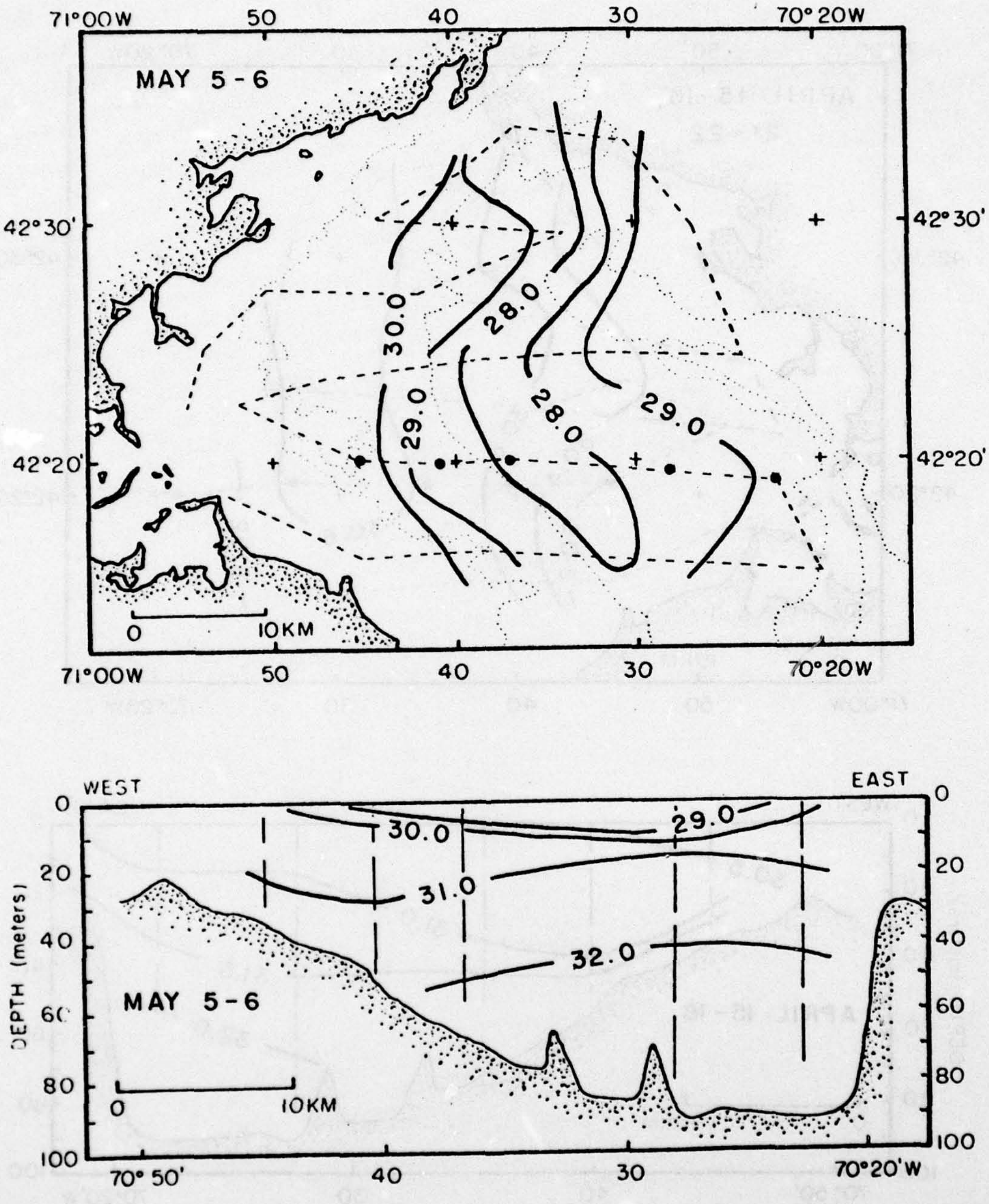

71°00W
50
40
30
70°20W
MARCH 29-30
30.5
31.0
31.5
31.0
31.5
42°30'
42°20'
0
10KM
71°00W
50
40
30
70°20W
WEST
EAST
MARCH 29-30
31.5
32.0
31.5
31.0
DEPTH (meters)
0
20
40
60
80
100
0
10KM
70° 50'
40
30
70°20'W

Figure 2.4a



71°00W
50
40
30
70°20W
APRIL 15-16, 21-22
42°30'
42°20'
30.4
30.6
30.2
30.6
30.8
0
10KM
WEST
EAST
DEPTH (meters)
0
20
40
60
80
100
30.5
31.0
31.5
32.0
APRIL 15-16
70° 50'
70°20'W

Figure 2.4b



71°00W
50
40
30
70°20W
MAY 5-6
42°30'
42°20'
30.0
28.0
29.0
28.0
29.0
0
10KM
WEST
EAST
DEPTH (meters)
0
20
40
60
80
100
30.0
29.0
31.0
32.0
70° 50'
70°20'W

Figure 2.4c

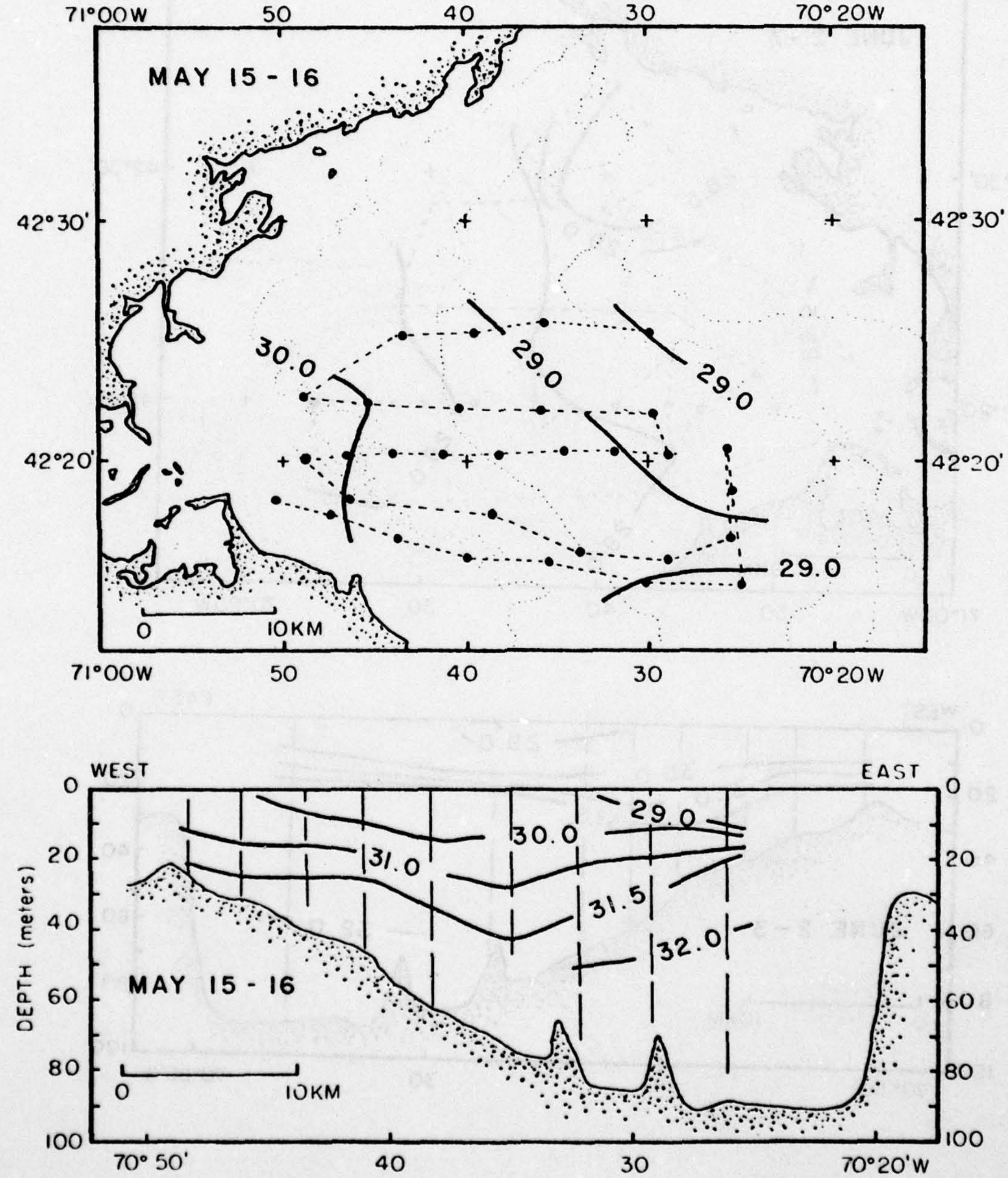

71°00W
50
40
30
70° 20W
MAY 15 - 16
42°30'
42°20'
30.0
29.0
29.0
29.0
0
10KM
WEST
EAST
DEPTH (meters)
0
20
40
60
80
100
29.0
30.0
31.0
31.5
32.0
MAY 15 - 16
0
10KM
70° 50'
40
30
70°20'W

Figure 2.4d

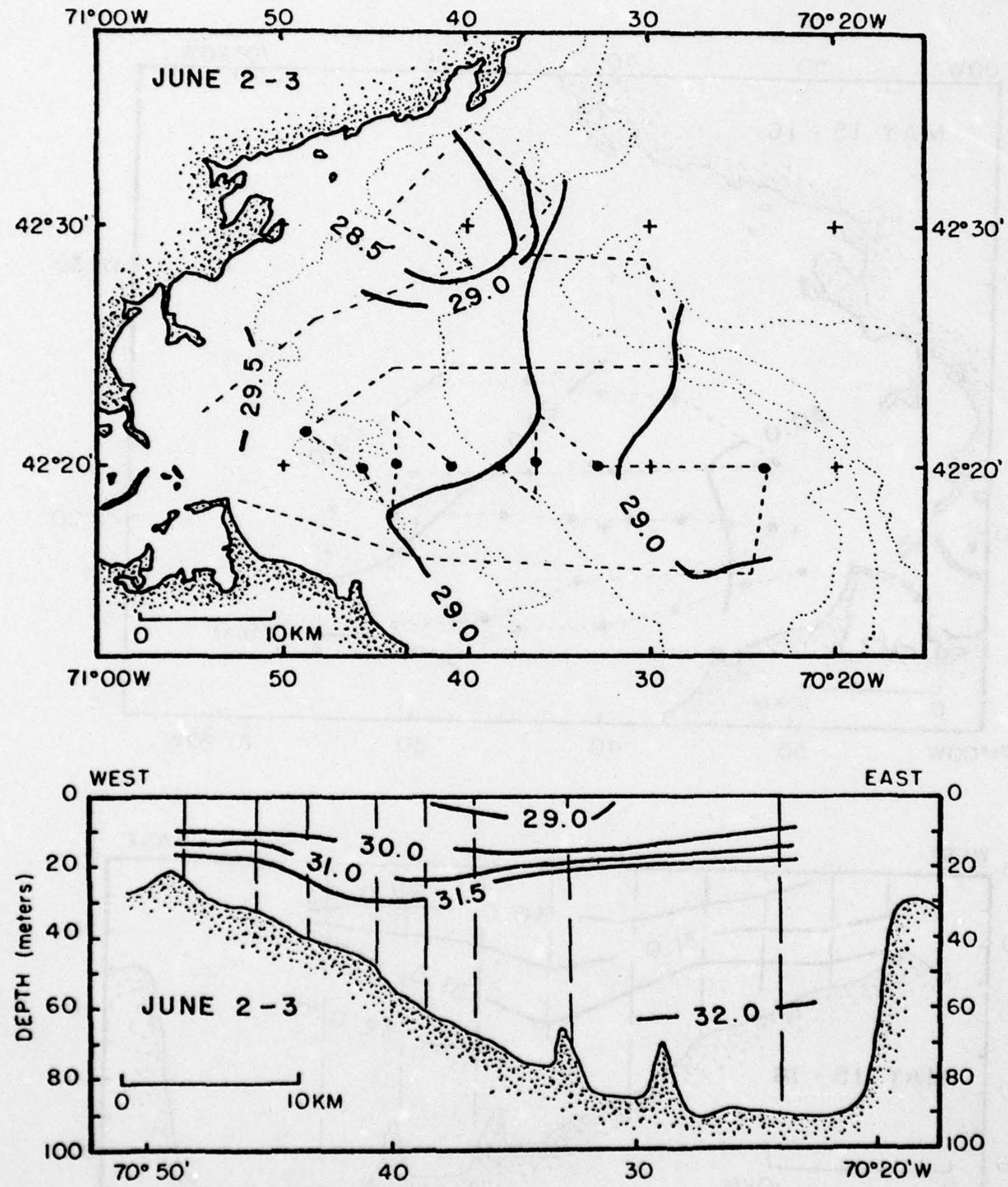

71°00W
50
40
30
70°20W
JUNE 2-3
42°30'
42°20'
28.5
29.0
29.5
29.0
29.0
0
10KM
WEST
EAST
DEPTH (meters)
0
20
40
60
80
100
29.0
30.0
31.0
31.5
32.0
JUNE 2-3
0
10KM
70°50'
40
30
70°20'W

Figure 2.4e