NOAA Technical Memorandum ERL PMEL-48

AN OVERVIEW OF CIRCULATION
IN THE PUGET SOUND ESTUARINE SYSTEM

Glenn A. Cannon

Pacific Marine Environmental Laboratory
Seattle, Washington
June 1983

UNITED STATES
DEPARTMENT OF COMMERCE

Malcolm Baldrige,
Secretary

NATIONAL OCEANIC AND
ATMOSPHERIC ADMINISTRATION

John V. Byrne,
Administrator

Environmental Research
Laboratories

George H. Ludwig
Director

NOTICE

Mention of a commercial company or product does not constitute an endorsement by NOAA Environmental Research Laboratories. Use for publicity or advertising purposes of information from this publication concerning proprietary products or the tests of such products is not authorized.

CONTENTS

An Overview of Circulation in the Puget Sound Estuarine System[1]

Glenn A. Cannon

ABSTRACT. The estuarine system that comprises Puget Sound and the Strait of Juan de Fuca, the southernmost, glacially carved waterways in western North America, is a complex composite of several connecting basins. Observations of circulation were made at a central location in the Main Basin of Puget Sound under various conditions during the past ten years. Intrusions of deep water below sill depth occurred at about fortnightly intervals, but not every fortnight. Flux calculations and water properties suggest a propagation of the intruded water along the Sound between Seattle and Tacoma in a little more than one week. A recent conceptual model, however, implies mixing of seaward-flowing surface water with the intruding deeper water. Present rough estimates are that below about 50 m only about half of the bottom water moving south past Seattle is new water from outside. Observations of mean daily current profiles in the Sound show significant deepening of the level of no-net-horizontal motion from about 50 m to as much as 100 m in approximately 200-m total depth. Winds along the Sound apparently can alter a fairly delicate balance of forces due to relatively small horizontal pressure gradients. In the Strait of Juan de Fuca, the waterway connecting the Sound with the Pacific Ocean, winter storms with predominantly southerly winds along the coast are capable of significantly reversing the normal estuarine flow and causing large intrusions of coastal water lasting several days. Outflow often occurs only in the deeper water with surface water being retained within the Strait.

1. Introduction and Geographic Setting

The Main Basin of Puget Sound is a fjord-like estuary connecting through Admiralty Inlet to the Strait of Juan de Fuca and thence westward to the Pacific Ocean (Fig. 1). The entrance sill extends about 31 km within Admiralty Inlet. It is shoalest at about 64 m just seaward of Port Townsend, but there is a secondary shoal of about 106 m near the entrance to Hood Canal (Ebbesmeyer and Barnes, 1980) (see e.g., Fig. 4); also see Geyer and Cannon, 1982, for a detailed chart of the sills). The Main Basin extends southward about 60 km from the major junction with Admiralty Inlet near Possession Point to the Narrows, a constriction with a 44-m sill separating it from the southern basin. Within the Main Basin, depths exceed 200 m in a section approximately 50 km long and three to five km wide. The Skagit River entering Whidbey Basin to the north supplies more than 60% of the freshwater.

---

[1]Contribution number 512 from the Pacific Marine Environmental Laboratory.

Prior to 1970 few direct observations of circulation in Puget Sound and along its connection to the ocean had been made for more than three to four days. Tidal current observations by the U.S. Coast and Geodetic Survey (USGS) to develop predictive tables were perhaps the only exception. Their observations, however, were not made in conjunction with water property observations and were of limited use in describing other flow characteristics. What was known was based on relatively short records. The estuarine characteristics of net seaward flow near the surface and net landward flow at depth were observed at several locations in Puget Sound from one to several days (Paquette and Barnes, 1951) and also in a section midway along the Strait of Juan de Fuca for about a day at each of three stations (Herlinveaux, 1954). Other observations, primarily of the distribution of water properties, contributed to the initial descriptions of flow characteristics (University of Washington, Department of Oceanography, 1954; Barnes and Collias, 1958; Herlinveaux and Tully, 1961). However, relatively little was known about variations in the circulation and, hence, about any consequences. The only known observations and descriptions of variations in mean flow were from electromagnetic flow measurements in Deception Pass (Morse et al., 1958). Those results suggest that prolonged high winds as well as the cessation of strong winds had a marked effect on net transport.

Because future decisions on questions regarding new and alternate uses of these waters require information on transport processes, we initiated in 1970 an experimental program utilizing unattended current meter moorings to characterize the temporal and spatial variability in the circulation and the large-scale dynamics of this estuarine system. Subsequently, many continuous observations in the Main Basin of Puget Sound have exceeded a month in duration, and in one case a mooring was maintained for an entire year (Fig. 1). A tabulation of known current observations through 1980 has been made by Cox et al. (1981).

These measurements, made beginning in 1970, initially involved use of Braincon current meters, supplemented by Nansen bottle casts to measure water properties. These instruments gradually were replaced by Aanderaa current meters, eventually equipped with temperature, conductivity, and pressure sensors, and later were supplemented with vector-averaging current meters, salinity, temperature, and depth (STD) and conductivity, temperature, and depth (CTD) profilers, and nearby land-based anemometers. The current meter observations described here were made in the 1970's and early 1980's. The records have been resolved into along-channel components in the direction of maximum variance, low-pass filtered (using fast Fourier transform techniques and more recently a Lanczos filter with a 35-hr, half-power frequency) to remove the tidal components (see Mofjeld and Larsen, 1983), and then daily averaged for conceptual convenience (Cannon, 1971).

This paper presents a summary of observations of tidally averaged flow and its variations that are important to understanding the overall circulation of Puget Sound and the Strait of Juan de Fuca (see also Ebbesmeyer et al., 1983b). Three main categories are considered: deep-water replacement, mean current profiles, and coastal interactions. Much of the material has appeared elsewhere in papers, reports, and abstracts.

2. Deep-water Replacement

Prior to the mid-1970's, our knowledge of bottom-water replacement in Puget Sound was based on approximately monthly observations of water properties. Deep-water replacement with denser, more saline water was thought to occur primarily in July-October in response to summer winds blowing from the north along the Pacific coast (Barnes and Collias, 1958). These northerly winds induced upwelling of deeper, denser oceanic water which then could be transported along the bottom of the Strait of Juan de Fuca and eventually into Puget Sound. Additionally, long-term hydrographic observations showed that bottom-water density in the Sound increased during April-May and decreased during May-June and October-February (Collias et al., 1974). During the lowering of density in winter, mixing and/or replacement results in the deeper water becoming relatively isothermal, less saline, and higher in dissolved oxygen.

Observations during the winters of 1972 and 1973 were made using moored instruments to record currents, temperature, and salinity at several levels at a location (mooring 6) in the Main Basin just north of Seattle (Fig. 1). The 1973 data (Cannon and Ebbesmeyer, 1978) showed that water near the bottom was cooled in a series of step temperature decreases of up to 0.6° C at about 2-week intervals, coincident with the onset of about 5-day intervals of net landward flow near the bottom (Fig. 2). Between the 5-day intervals, bottom flow was tidal with a small seaward component. A vertical temperature front was observed close to the mooring following the first sharp decrease, but at the beginning and end of the month there was no evidence of this front. Also, there apparently was less dense water (colder but less saline) intruding to near bottom just north of the mooring and the passing temperature front, as well as an overall decrease in density during the month.

Intrusion of new deep water occurred when the flood-tide range at Seattle exceeded 3.5 m. At this range, Strait of Juan de Fuca water had been shown by hydraulic model studies to completely transit Admiralty Inlet sill in one flood period and thus be least mixed (Farmer and Rattray, 1963). The 1973 study proposed that an 8- to 10-day time lag was required to transit the distance from the sill to the mooring site, and that the largest temperature decrease occurred following flood tides concurrent with below-freezing air temperatures over Admiralty Inlet. Density differences during the temperature steps were small, and it was not possible to determine whether initially the new water was more dense.

An overall rate of replacement of intermediate and deep water south of the mooring was calculated for the one-month 1973 winter observations using average daily currents which were assumed uniform across-channel, but without considering effects of entrainment and diffusion (Cannon and Ebbesmeyer, 1978). The mean landward (southward) transport (4.3×10^4 m^3/sec) was divided into the volume beneath the average depth of no-net-motion (about 35 km^3 beneath 52 m) which yielded a mean effective replacement time of about 9 days. The replacement time varied between 5 and 40 days (Fig. 2). More recent calculations by Ebbesmeyer et al. (1983) indicate that the flux measured from this central location should be multiplied by about 0.75 to account for cross-channel variations.

Cannon and Ebbesmeyer (1978) speculated that a replacement process similar to that observed in winter might occur in late summer when the most dense bottom water can enter the Sound. Current observations in July and August on the sill of Port Susan (mooring 13; Fig. 1), a secondary basin in Puget Sound, showed that water flowed into the basin three times for about 5 days each time (Cannon 1975). There was also speculation that deep water might be added quasi-continuously at about 2-week intervals. In an attempt to answer these questions, a year-long mooring (1975-76) was maintained at the same location (mooring 6). In addition, a survey of water properties was repeated five times during the year (Cannon and Laird, 1978). These observations also showed that winter intrusions (December-February) occurred 8 to 10 days following periods when tidal ranges exceeded 3.5 m (Fig. 3). Characteristics of water properties along the Sound during an intrusion usually showed a tongue-like contour or a single contour near bottom and extending southward from the south end of Admiralty Inlet (Fig. 4). However, note that an intrusion does not necessarily imply that all deep water is being replaced. Unlike the 1973 observations, all intrusions during 1975-76 were characterized by increased density of bottom water. Also, the bottom-flow reversals indicated that the intrusions in winter 1976 were very pronounced (Fig. 3). Horizontal excursions in December, January, and February (implied from the flow measured at the one location) were 35, 35, and 140 km, respectively. The 140-km excursion coincident with the largest salinity intrusion was remarkably large if one considers that the length of the Main Basin at depths of 180 m is only about 50 km. Daily average speeds at the mooring for the 5 largest consecutive days of this intrusion were about 20 cm/sec, which corresponds to an excursion of 86 km. Two intrusions in March were equally large, and one in April was only slightly smaller (Fig. 3).

The overall replacement rate of intermediate and deep water of all the major intrusions during the 1975-76 study agree with those calculated for the 1973 observations. However, how much new water is entering during any given intrusion could not be determined from these observations. The possible deep-water renewal rate of one to two weeks indicated by these studies is significantly shorter than earlier estimates of two to ten months for all of Puget Sound based on monthly averages of water properties (Friebertshauser and Duxbury, 1972). Their estimates of flushing were for the total water column for all the Puget Sound basins. Their overall averages were significantly influenced by the particularly slow renewal of Hood Canal, and they were unable to make specific estimates for the Main Basin. However, they indicated a possible quick renewal during late-summer bottom-water intrusions. Thus, the results here are not contradictory, but provide added information for deep-water replacement in the Main Basin.

Replacement Processes. Even though the yearlong study confirmed several of the 1972 and 1973 winter observations, some major differences were found. First, there were occurrences of large tidal ranges during winter (November-January) with either little or no inflow and with no apparent changes in water properties. Second, there were occurrences of inflow with different water properties during low tidal ranges during August-October and during March-April. The most dense water can enter the Sound during the summer-fall interval, and the least dense water is in the Sound during the winter-spring interval.

Therefore, other factors must be important in determining when there will be intrusions of deep water. One apparent factor is the degree of mixing that takes place when crossing the finite-length (31-km) entrance sill. There is always more dense water available at the sill than at the bottom in the Main Basin (Fig. 4 and Collias et al., 1974). The yearlong observations indicated that during summer to early fall, the water apparently was sufficiently saline (dense) that it did not matter that more than one tide was required to transport it across the sill. Also, at that time, only minimal amounts of freshwater entered the Sound to be available for mixing with the salt water over the sill. These intrusions during low-tide ranges also occurred at about fortnightly intervals. However, the lack of significant inflow later, during the fall and early winter when there were large tides, probably was influenced by there being an abundance of freshwater available for mixing over the sill. The largest temperature step decrease in December 1975 occurred following an extreme cold spell coincident with the large tidal range (Fig. 3). This cold spell also was characterized by clear weather and a lack of freshwater, as was the case in winter 1973 (Fig. 2). The two large intrusions in March 1976 occurred during small tides. They represented excursions as large as the one in February. This case is least understood. Apparently the combination of greater density outside the sill with minimum density in the basin is sufficient to cause the intrusion, even though more than one tidal cycle may be required to cross the sill, but still at fortnightly intervals. Earlier work showed renewal of oxygen in the secondary basin of Port Susan to be dominant during these winter intrusions (Cannon, 1975).

Observations during fall 1977 showed step decreases in bottom-water temperature at the long-term mooring location (mooring 6) and also, for the first time, at a second location in the Main Basin (mooring 5) at the south end of East Passage (Fig. 5). There was a time lag of about seven days for the 45 km between the moorings (Cannon and Laird, 1980). Observations in Admiralty Inlet, also in fall 1977, showed that new bottom water entered the Main Basin during both high and low flood tidal ranges. During low flood tide range, new bottom-water inflow was characterized by lower maximum tidal currents, but of longer duration, than during higher flood-tide range. On one occasion when inflow occurred during neap tides, the implied excursions for a successive flood-ebb-flood were 26, -4, and 12 km at the north end and 14, -3, and 12 km at the south end of Admiralty Inlet. The net excursions were 34 and 23 km, respectively. The relatively short ebb excursions provided little time for mixing, and the resultant salinity at the south end was considerably more (about 0.5°/oo) than at the beginning of the first flood (Fig. 5). During the following spring tides, the excursions for a successive flood-ebb were 23 and -14 km at the north end and 11 and -9 km at the south end, and the net excursions were more nearly equal. In these cases, more mixing occurred, and the resultant salinity at the south end was less during spring tides than during neap tides. The tidally averaged flow during these intrusions was characterized by increased bottom inflow and by increased outflow of upper water at about fortnightly intervals. The outflowing upper water became fresher and warmer during the bottom-water intrusions which were saltier and colder. Figure 5 shows an intrusion of new bottom water across Admiralty Inlet into the Main Basin under these conditions. Holbrook and Cannon (1983) have found that once the intrusion enters the basin, it propagates first as a temperature change followed by an up-estuary current pulse and then as a maximum in salinity and density.

It is still uncertain how to predict those inflows occurring during the flood-tide ranges of less than 3.5 m. Some possible tidal explanations are indicated by Mofjeld and Larsen (1983) and by Geyer and Cannon (1982). Apparently for ranges less than 3.5 m, more mixing occurs during the relatively higher tides and blocks formation of any gravitational circulation, while during the relatively smaller tides a two-layer regime develops. The optimum tidal conditions for inflow of bottom water in this case seems to occur during the equinox when the maximum diurnal inequality is in phase with neap tides. This process probably is applicable during both of the low-tide-range conditions indicated above for late summer and late winter.

Reflux of Water. Ebbesmeyer and Barnes (1980; also Barnes and Ebbesmeyer, 1978; and Cannon and Ebbesmeyer, 1978) have developed a conceptual model of the Sound's circulation based on many years of water property observations and on some of the recent current observations reported here. The hypothesis is that considerable seaward-flowing surface water is mixed with new bottom water just south of the Admiralty Inlet 106-m sill and is refluxed back into the Sound as part of the new deep water (Fig. 6). If this hypothesis is valid, it will have significant implications for predicting how quickly water (and thus dissolved pollutants) can be removed from the Sound. Month-long-averaged currents in the middle part of Admiralty Inlet are used here to make a first approximation of the volume flux into the Sound of about 2×10^4 m^3/sec. This is about half of what was estimated to flow south past mooring 6 in the Main Basin, thus supporting the hypothesis. However, significant variations are implied if the daily average currents (Fig. 6) are used for the flux calculation. The wave-like nature of some mid-depth isopleths is partial indication that the mixing of Puget Sound and Admiralty Inlet waters to increase the inward volume transport may take place at least partly through internal waves (Fig. 4).

Narrows Influence. Finally, a major difference between Puget Sound and other fjords is that the head of the Main Basin, the Narrows, is a relatively narrow, shallow passage; it is connected with the southern basin which has a relatively large tidal prism. Extremely high tidal currents through the Narrows appear capable of pumping deep water up from about twice the depth of the sill (44 m), thus greatly assisting the movement of deep water through Puget Sound. Hydraulic model studies blocking the Narrows have shown a reduction of about 75% in volume transport southward in East Passage (Ebbesmeyer and Barnes, 1980). Attempts in winter 1977 to determine whether this process occurs continuously on all flood tides or whether it occurs only on the largest, fortnightly tides were inconclusive because an unusually dry winter produced very uniform water properties in the Sound (Cannon et al., 1979).

3. Mean Current Profiles

Current meter observations of at least one month's duration for several years at the same location in the Main Basin (mooring 6, Fig. 1) are available to define vertical profiles of tidally averaged longitudinal currents. The depth of no-net-motion along-channel occurred in a range of 40-60 m (Fig. 7a, top). Although the number of data points in the vertical is small, the three winter profiles appear similar, with the bottom layer inflow maxima being a little shallower than during fall and spring. The 1979 summer profile showed the deepest inflow maximum. The average depth of zero longitudinal flow of about 50 m is about 25% of the total depth, which is intermediate between the shallower level for most fjords (10-15%) and the deeper level for partially mixed estuaries (40-50%). Application of Rattray's (1967) general fjord circulation theory to the Main Basin of Puget Sound (Winter, 1973; Winter et al., 1975) indicated approximate agreement in magnitude of the mean currents, but the depth of no longitudinal motion was about 10-20 m shallower than indicated by the observations. However, wind effects were not included in the model.

Wind Effects. Significant variations in the depth of zero longitudinal flow from these longer term means, however, have been observed in mean daily along-channel current profiles at the same location. During predominantly southerly winds in winter (blowing out-estuary, to the north), a quasi-steady profile is established with a depth of no-net-motion of 40-50 m. Relaxation of the wind or reversal to northerly (blowing in-estuary, to the south), even for brief intervals, results in apparent deepening of this level. During winter 1972, winds were southerly except for one two-day reversal midway through the record and a brief interval at the beginning. The zero crossing was lowered, with a short time lag, from 50 m to more than 100 m in a total depth of about 200 m, then returned to its original level (with a possible overshoot) over a period of a few days when the winds returned to southerly.

During winter 1973, there were more wind events that indicated the same effect. Winds which were primarily southerly were, except at the beginning of the record, fairly steady (Fig. 8). However, there was a letup in the wind on 17 January, a complete reversal on 25-28 January, and a major reversal commencing on 3 February. The average daily along-channel profiles initially showed a zero crossing of 40-50 m with the southerly winds. Following the wind letup on 17 January, the zero crossing deepened slightly by about 10 m, then shoaled as southerly winds continued. Following the 25-28 January wind reversal, the zero crossing deepened to 70 m and then returned to about 50 m before the major wind reversal on 3 February, after which the level deepened to almost 100 m. Note also the obvious surface layer reversal with up-estuary winds and the deep inflow discussed earlier (see Fig. 2). Similar changes in the depth of no-net-motion along-channel occurred during the 1975-76 observations (Cannon and Laird, 1978).

These relative shifts in the deeper flow caused by wind stress appear consistent with a general study of fjords (Rattray, 1967). Out-estuary winds raised the inflowing layer, including the depth of the maximum, to shallower depths. The main differences from the model were that inflow extended all the way to the bottom, that significant pulses of inflowing water occurred at the bottom, and that vertical excursions appeared larger than predicted by theory. STD observations along the Main Basin indicated relatively small horizontal gradients compared to those of Admiralty Inlet (see Fig. 4). Possibly a quasi-steady state existed for winter conditions of moderate-to-strong southerly (out-estuary) winds. Significant relaxation or reversal of the winds apparently changed the barotropic part of the horizontal pressure gradient, thus changing the total horizontal pressure gradient which was probably in relatively delicate balance. Observations by Pickard and Rodgers (1959) in Knight Inlet, British Columbia, have shown similar shoaling/deepening of the zero crossing from one-day observations during opposite wind conditions.

Observations across-channel from a two-month interval in the yearlong study (summer 1976) at mooring 6 show deepening of the zero level to the west (down to the left looking seaward; Fig. 9). The depth of the zero level about halfway to the western side from mooring 6 was about 70 m, and it was about 40 m halfway to the eastern side, indicating a depth of 56 m at mid-channel. This seems to imply that the observed deepening in Figure 8 is not due to cross-channel changes. However, cross-channel effects can be important in calculating volume fluxes through a section as was shown for Admiralty Inlet (Fig. 6).

Spatial Variations. These profiles in the Main Basin are probably accurate for mid-channel because they are based on long-term measurements. However, few long-term observations have been made at other Main Basin locations. Near the ends of the basin and in other parts of the system, the profiles can be expected to be considerably different. Concurrent month-long observations in East Passage and Colvos Passage, near the southern end of the Main Basin, indicate clockwise circulation around Vashon Island (Fig. 7a, bottom) as was previously deduced from hydraulic models. Volume fluxes assuming that the profiles in these two areas are uniform across-channel (as in the Main Basin studies) indicated 3.8 and 2.7 × 10^4 m^3/sec southward and northward, respectively, for the two passages (cross sections on right in Fig. 9). The 2.7 × 10^4 m^3/sec flux going north in Colvos Passage is the same as calculations based on the tidal prism south of the Narrows (Barnes and Ebbesmeyer, 1978). The 3.8 × 10^4 m^3/sec southward flux is about the same as that calculated for the middle part of the Main Basin (4.3 × 10^4, Cannon and Ebbesmeyer, 1978). The differences between the northward and southward fluxes have not been resolved, but they may indicate that some near-surface water is going northward in East Passage. Also, the effects of cross-channel variations at each of the sections have not been resolved.

Observations in Saratoga Passage in Whidbey Basin averaged over a shorter interval (15 days) show what has been considered a classical fjord current profile (Fig. 7). The location of the station can be thought of as being in the middle reach of a fjord, downstream from the Skagit River at the head of the fjord and upstream from the Admiralty Inlet sill (Fig. 1). Significant variations from the two-week average exist in the daily average current profiles (Fig. 10). These variations like those in the Main Basin appear to be related to wind effects.

Recent observations from the middle reach of Hood Canal during winter 1980 are for a longer interval (2 months) and are still being analyzed (Fig. 7). The profile differs from that of Saratoga Passage, possibly because Hood Canal lacks a significant or single source of freshwater. It also differs from those in the Main Basin. Both Hood Canal and Saratoga Passage are dead-end basins. Deception Pass flow has little affect in Saratoga Passage. The Main Basin, however, is not a dead-end but connects to the southern basin through the Narrows. The large currents in the Narrows help maintain relatively stronger flows in the bottom of the Main Basin (Ebbesmeyer and Barnes, 1980).

Current profiles in Admiralty Inlet resemble those of a coastal plain estuary (Figs. 6 and 7). When new bottom water enters the Main Basin, it shows as an increased inflow in the bottom layer of Admiralty Inlet, and because of continuity, it is accompanied by increased outflow in the surface layer (Fig. 7b and Geyer and Cannon, 1982). However, they show that when this increased inflow reaches the Main Basin, it results in a flow reversal in the central part of the water column, and there is negligible response near the surface.

4. Coastal Interactions

Because Puget Sound connects with the Pacific Ocean through the Strait of Juan de Fuca (Fig. 11), some major circulation variability characteristics of the Strait are included from Cannon and Holbrook (1981), who reviewed recent studies. Earlier studies in the outer basin, which has no sill, revealed characteristics similar to those of a partially mixed estuary, with seaward transport near the surface and landward transport near the bottom (Herlinveaux and Tully, 1961). The primary source of freshwater to the Strait is the Fraser River entering far upstream in the Strait of Georgia at Vancouver. However, strong tidal currents in the San Juan Island passages mix the freshwater lens, which is characteristic of fjords, before it reaches the Strait.

One of the first indications of significant variations in the estuarine flow in the Strait were observed during a study primarily of flow in the vicinity of the Juan de Fuca canyon in October-November 1971 (Cannon et al., 1972) based on an earlier study of the canyon (Cannon, 1972). Observations at mooring A (Fig. 11) at the mouth of the Strait indicate that major winter storms could result in reversal of the estuarine flow for a few days with inflow near the surface and outflow near the bottom. In fact, these flow reversals were also observed at the outer edge of the continental shelf in the glacial trough that extends entirely across the shelf. Subsequently, oceanographic studies in the outer basin of the Strait (Fig. 11) were carried out during two winters (February-April 1976, November 1976-February 1977) and one summer (June-August 1977). The first winter study (subsurface moorings N, M, S; Fig. 11) was designed, in part, to investigate the possible eastward propagation of the flow reverals and, if found, to determine its affect at the inshore end of the western basin (Cannon, ed., 1978). The 1976-77 winter study (surface moorings A, B, C) documents the relationship between coastal storms and the near-surface flow reversals along the western basin (Holbrook and Halpern, 1982). The summer studies (moorings D, E, F, G) show that reversals also occurred in the eastern basin of the Strait (Holbrook et al., 1980a; Frisch et al., 1981; Muench and Holbrook, 1980).

Surface Changes. Observations during the second winter (moorings A, B, C; Fig. 11) show that decelerations in surface flow in the Strait generally occurred during strong southerly winds off the coast accompanied by increasing sea surface height at Neah Bay (Holbrook and Halpern, 1982). On five occasions intrusions of 1-3°C warmer northeast Pacific coastal water occurred for durations of 1-10 days. The 25 cm/sec up-strait speed of the intrusive lens agreed to within 20% of the gravity current speed computed from Benjamin's (1968) hydraulic model. The near-surface currents associated with the intrusions and southerly coastal winds were significantly correlated indicating that the intrusions were initiated when shoreward Ekman currents advected Pacific coastal water into the Strait (Fig. 12). In a subsequent study (Holbrook et al., 1983), two-year measurements of currents, water properties, winds, and sea level indicate that surface intrusions of less-dense coastal water into the Strait were a major, recurring feature of its general circulation throughout the year. These data and additional satellite images suggest that a wind-induced shoreward Ekman flux at the coast "pushes" less-dense coastal water into the Strait where it feeds a complex, 3-dimensional, density-driven flow, which has been observed to travel as far as 135 km up-strait. The isostatic head associated with the intruding coastal water establishes along-strait and across-strait sea surface slope fluctuations that are significantly coherent with reversals in current and coastal winds.

Deep-water Changes. During the first winter, the variations of flow through a section 90 km from the coast (moorings N, M, S; Fig. 11) showed two periods of inflow and two periods of outflow in the near-surface water (to about 15 m) on the south side of the Strait (Fig. 13; Cannon and Holbrook, 1981). Inflow occurred from mooring deployment (26 February) to 2 March and from 19-28 March. The second inflow started later and lasted longer at the middle (M) and north (N) moorings. Flow reversed at the north mooring earlier than at the middle one. The implied excursion during the second inflow at the south mooring was about 150 km, a distance from the mooring greater than that to the mouth of the Strait to the west, or to land toward the east. The excursions at the middle and north moorings were about 100 km. These reversals appeared directly related to storm winds along the coast measured on Tatoosh Island at the mouth of the Strait.

The different flow patterns through the section can be shown using average daily longitudinal currents (Fig. 14). The more normal estuarine flow (14 March) was similar to the month-long average, except that the level of no-net-motion was shallower indicating the deepening effect of the reversal on the no-motion level. The second reversal commenced at the south mooring on 19 March. The flow patterns continued to change until the 15-m average flow reached its maximum of more than 50 cm/sec on 25 March when the reversal was characterized by landward flow across the entire section. Seaward flow was observed at all deeper current meters except the bottom one on the north mooring. There was a mid-depth maximum greater than 30 cm/sec at the 60-m depth at the middle mooring. Flow on the south side commenced to weaken on 26 March, and a three-layer flow developed across the entire section. Maximum seaward flow remained at mid-depth, and this pattern remained for three days until the end of the reversal on 28 March after ten days.

During the same interval, water property observations showed that the flow reversals were accompanied by intrusions of relatively fresher and warmer coastal water, primarily along the south side. During the intrusion, the freshest water was less than $30^{o}/oo$, the $32^{o}/oo$ isohaline was relatively flat at about 100 m, and the $33^{o}/oo$ isohaline was near the bottom as was also indicated on a longitudinal STD section during another intrusion. During more normal estuarine flow, the freshest water was about $31^{o}/oo$, occurred on the north side, and all but the deepest isohaline sloped down toward the north. The $32^{o}/oo$ isohaline occurred at much shallower depths, and the $33^{o}/oo$ isohaline was about 50-70 m shallower than during the intrusion.

Summer Interactions. Although coastal low-pressure systems with southwesterly winds occur more frequently and persist longer during winter than during summer, occasionally a low-pressure system will lie off the coast long enough during summer to generate a response in the Strait. Such a system was observed in August 1978. Prior to and during an up-strait propagation of a storm-induced reversal, surface currents were monitored in the eastern basin of the Strait using surface drifters, a Coastal Ocean Dynamics Applications Radar (CODAR), and surface-moored, vector-averaging current meters (Frisch et al., 1981; Holbrook et al., 1980b). The CODAR system utilizes two shore-based, transportable HF Doppler radars to map hourly surface currents over a 900-km^2 area with a spatial resolution of 1.2 km (Barrick et al., 1977). Comparisons between the near-surface currents measured by the current meters at three sites and the CODAR data showed good agreement for both the mean and tidal flows (Holbrook and Frisch, 1981). The 3-day-averaged flow prior to the storm showed net seaward velocities of 20-40 cm/sec throughout the eastern Strait, and the subsequent 24-hr-averaged CODAR current map corresponded to the time of arrival of the reversal between Port Angeles and New Dungeness (Fig. 15). The core of the eastward flow was located near mid-strait, with shoreward components on either side. In the convergence zone near the reversal front, the seaward estuarine flow was diverted southward with an eastward recirculation north of New Dungeness Spit. These CODAR maps dramatically show the complexity of surface current reversals. The Tongue Point current and salinity sections (Fig. 14) show the vertical complexity.

5. Implications

An overall rate of replacement of intermediate and deep water south of the Seattle long-term mooring (6) (from a line between Meadow Point and Point Monroe) has been approximated using daily-average currents assumed uniform across-channel. The average replacement time was less than two weeks when new bottom water was entering, and observations of water properties along the Sound showed the same transit time. New deep water has been observed to enter the Sound at about fortnightly intervals usually associated with the largest flood tides through Admiralty Inlet, but also associated with the smallest flood tides. The first occurs when the flood-tide range (higher high water minus lower low water) exceeds about 3.5 m. Incoming bottom water can transit Admiralty Inlet in one flood cycle and is least mixed. The second occurs during the lowest flood tide ranges during the equinoxes in late winter and late summer. Incoming bottom water transits most of Admiralty Inlet on the larger flood cycle. However, the subsequent small ebb is almost nonexistent at the bottom, and the following small flood is capable of completing the transit of the sill with minimal mixing. At intermediate tide ranges, floods and ebbs are more nearly equal causing maximum mixing of water in Admiralty Inlet. However, denser water is always more readily available at sill depth in the Strait than it is in the bottom of the Sound, and the intrusions do not occur every fortnight. The effects of tides and of winds in Admiralty Inlet on deep-water replacement are only now receiving attention (Mofjeld and Larsen, 1983; Ebbesmeyer et al., 1983).

A recent conceptual model of the Sound's circulation implies that considerable seaward-flowing surface water is entrained or mixed downward and becomes part of new deep water entering the Sound at the south end of the Admiralty Inlet entrance sill (Ebbesmeyer and Barnes, 1980). At least part of this downward mixing may take place through internal waves generated at the sill. Ongoing studies are attempting to determine how much "new" water actually enters during any given intrusion. Flux calculations using month-long average currents midway along Admiralty Inlet show about half the inward volume transport as observed midway along the Main Basin. However, the variations due to seasonal changes are just beginning to be investigated. A recently developed box model of the Main Basin implies that winter transports may be about twice as large as summer transports (Hamilton et al., 1983). Ebbesmeyer et al. (1983b) find a similar difference in a recent synthesis of current observations.

In addition, the constriction of the Narrows, the landward sill zone, appears to act as a "pump" to remove deep water from the southern end of the Main Basin (Barnes and Ebbesmeyer, 1978). Thus, a combination of effects apparently results in the Main Basin of the Sound being somewhat unique in its capability to renew its deeper water at a much more rapid rate than a classical fjords (Gade and Edwards, 1980). Historical data have never indicated occurrences of low-oxygen water in the Main Basin (Collias et al., 1974). However, we need to understand the flushing processes better in order to determine what potential upper limits exist to use of the Sound for waste disposal.

Longer term average current profiles show characteristics of the Main Basin somewhat between those of classical fjords and partially mixed estuaries. However, significant variations in these profiles have been observed when averaging over intervals of about a day. The concept of a mean current is obviously a function of time. Apparently, there is a delicate balance of forces along the basin because of the relatively small horizontal gradients. Thus, the wind contribution to the total horizontal pressure gradient results in major changes in flow throughout the water column. Ebbesmeyer et al. (1983) indicate that wind effects are correlated to variations in southward flow in the Main Basin in the 90- to 120-m range. Similar flow variations have been observed in the more nearly classical fjord flow in Saratoga Passage. Little work has been carried out on this aspect of fjord circulation.

Observations have demonstrated the role of nonlocal meteorological forcing in modifying circulation in the Strait of Juan de Fuca, which connects Puget Sound with the open Pacific Ocean (Cannon et al., 1981; Holbrook and Halpern, 1982; Holbrook et al., 1983). The wind-stress field off the Washington coast can indirectly influence the along-channel currents through induced sea surface height fluctuations. Storm-related southerly winds may be expected to pile up water along the coast through onshore Ekman transport. This increases the coastal sea surface elevation, which in turn decreases the longitudinal sea surface slope in the Strait causing a deceleration of the seaward near-surface circulation. When the generated current response in the Strait becomes larger than the net gravitational circulation, current reversals propagate up-strait. Although the variations associated with coastal wind forcing decreased with distance up-strait and with depth, the effects of coastal storms have been observed to reverse surface flow as far as 135 km up-strait and to reverse flow throughout the water column 90 km from the coast. At 90 km, the reversals are accompanied by large vertical changes in water properties indicating a seaward retreating of the deeper salt water and a landward intruding of coastal surface water lasting from a few days to more than a week.

Observations in the Strait have many characteristics that are in agreement with a two-layer model relating coastal and fjord circulation (Klinck et al., 1981). During the downwelling of southwesterly winds, the model predicts an increase in sea surface slope across the shelf to the mouth of a fjord and then a decrease up the fjord. The interface has the opposite slope, down toward the coast and up along the fjord. Although the observations show continuously stratified water, the slopes of the isopycnals across the shelf and in the Strait generally are consistent with the model results (Cannon and Holbrook, 1981). However, Klinck et al. (1981) give results for model runs simulating the Strait of Juan de Fuca that indicate that only a limited region of the Strait (perhaps the outer 10-30 km) is affected by a disturbance at the coast, even after five days of steady forcing. Observations by Holbrook and Halpern (1982), Frisch et al. (1981), and Holbrook et al. (1983) indicate that coast intrusions are a major feature of the circulation up to 135 km into the Strait. This lack of agreement may, however, be expected since the model does not include such important effects as the less-dense coastal water mass and Coriolis acceleration in the fjord. Similar results have been observed in other fjords and partially mixed estuaries (Svendsen, 1981; Wang and Elliott, 1978). Two recent theoretical studies have attempted explanations of some of these observations. Wright (1980) showed that baroclinic instabilities may have a significant influence on this type of motion, but they probably are not the major source of energy. Rattray and Proehl (1982) have demonstrated that coastal storms can generate internal Kelvin waves which propagate up-strait favoring the southern side and are accompanied by cross-channel standing waves trapped near the mouth of the estuary. Finally, changes in flow in the Strait are related to flow in the Juan de Fuca canyon extending entirely across the shelf (Cannon and Holbrook, 1981; Cannon and Lagerloef, 1983).

6. Acknowledgements

This work was partially sponsored by the MESA Puget Sound Project of NOAA. Supporting services by many people have been necessary to carry out the numerous field observations and subsequent analyses. Pat Laird, who was lost at sea off Hawaii in December 1978, was a major contributor to all aspects. Many discussions over the years with Professors Clifford Barnes and Maurice Rattray, and more recently with Curtis Ebbesmeyer and James Holbrook, have provided much insight regarding estuarine circulation and the Puget Sound system. Many students have also contributed. All this support has been greatly appreciated. Because this paper is largely a review, related contributions from others have been freely used and appropriately referenced. I am particularly grateful to Jim Holbrook for critically reviewing the manuscript. Michael Grigsby and Jean Chatfield provided assistance with the final editing of figures and text, respectively.

7. References

Barnes, C.A., and E.E. Collias, 1958. Some consideration of oxygen utilization rates in Puget Sound. Journal of Marine Research, 17, 68-80.

Barnes, C.A., and C.C. Ebbesmeyer, 1978. Some aspects of Puget Sound's circulation and water properties. In: Estuarine Transport Processes, B. Kjerfve (ed.), University of South Carolina Press, Columbia, South Carolina, 209-228.

Barrick, D.E., M.W. Evans, and B.L. Weber, 1977. Ocean surface currents mapped by radar. Science, 198, 4313.

Benjamin, T.B., 1968. Gravity currents and related phenomena. Journal of Fluid Mechanics, 31, 209-248.

Cannon, G.A., 1971. Statistical characteristics of velocity fluctuations at intermediate scales in a coastal plain estuary. Journal of Geophysical Research, 76, 5852-5858.

Cannon, G.A., 1972. Wind effects on currents observed in Juan de Fuca submarine canyon. Journal of Physical Oceangraphy, 2, 281-285.

Cannon, G.A., 1975. Observations of bottom-water flushing in a fjord-like estuary. Estuarine and Coastal Marine Science, 3, 95-102.

Cannon, G.A. (ed.), 1978. Circulation in the Strait of Juan de Fuca: some recent oceanographic observations. NOAA Technical Report ERL 399-PMEL 29, 49 pp.

Cannon, G.A., and C.C. Ebbesmeyer, 1978. Winter replacement of bottom water in Puget Sound. In: Estuarine Transport Processes, B. Kjerfve (ed.), University of South Carolina Press, Columbia, South Carolina, 229-238.

Cannon, G.A., and J.R. Holbrook, 1981. Wind-induced seasonal interactions between coastal and fjord circulation. In: The Norwegian Coastal Current, R. Saetre and M. Mork (eds.), University of Bergen, 131-151.

Cannon, G.A., and G.S.E. Lagerloef, 1983. Topographic influences on coastal circulation (a review). In: Coastal Oceanography, Physical Aspects, H. Gade, H. Svendsen and A. Edwards (eds.), Plenum Press, New York (in press).

Cannon G.A., and N.P. Laird, 1976. Wind effects on tidally averaged current profiles in a fjord estuary. Transactions American Geophysical Union, 57, 933 (abstract).

Cannon, G.A., and N.P. Laird, 1978. Variability of currents and water properties from year-long observations in a fjord estuary. In: Hydrodynamics of Estuaries and Fjords, J.C.J. Nichoul (ed.), Elsevier, Amsterdam, 515-535.

Cannon, G.A., and N.P. Laird, 1980. Characteristics of flow over a sill during deep water renewal. In: Fjord Oceanography, H.J. Freeland, D.M. Farmer, and C.D. Levings (eds.), Plenum Press, New York, 549-556.

Cannon, G.A., N.P. Laird, and T.L. Keefer, 1979. Puget Sound circulation: final report for FY77-78. NOAA Technical Memorandum ERL MESA-40, 55 pp.

Cannon, G.A., N.P. Laird, and T.V. Ryan, 1972. Currents observed in Juan de Fuca submarine canyon and vicinity, 1971. NOAA Technical Report ERL 252-POL 14, 57 pp.

Collias, E.E., N. McGary, and C.A. Barnes, 1974. Atlas of physical and chemical properties of Puget Sound and its approaches. Washington Sea Grant Publications, University of Washington Press, Seattle, Washington, 235 pp.

Cox, J.M., C.C. Ebbesmeyer, C.A. Coomes, L.R. Hinchey, J.M. Helseth, G.A. Cannon, and C.A. Barnes, 1981. Index to observations of currents in Puget Sound, Washington from 1908-1980. NOAA Technical Memorandum OMPA-5, 51 pp.

Ebbesmeyer, C.C., and C.A. Barnes, 1980. Control of a fjord basin's dynamics by tidal mixing in embracing sill zones. Estuarine and Coastal Marine Science, 11, 311-330.

Ebbesmeyer, C.C., C.A. Coomes, J.M. Cox, J.M. Helseth, L.R. Hinchey, G.A. Cannon, and C.A. Barnes, 1983a. Synthesis of current measurements in Puget Sound, Washington. Vol. 1. Circulation in Puget Sound: an interpretation based on historical records of currents. NOAA Technical Memorandum OMPA, (in press).

Ebbesmeyer, C., C. Coomes, J. Cox, J. Helseth, L. Hinchey, G. Cannon and C. Barnes, 1983b. Chatacteristics of circulation and water properties in Puget Sound. In: Puget Sound. Oceanography of the Inshore Waters of Washington, D. Henry (ed.), Univ. of Wash. Press, Seattle (resubmitted).

Farmer, H.G., and M. Rattray, 1963. A model of the steady-state salinity distribution in Puget Sound. University of Washington, Department of Oceanography, Technical Report 85, 33 pp.

Friebertshauser, M.A., and A.C. Duxbury, 1972. A water budget study of Puget Sound and its subregions. Limnology and Oceanography, 17, 237-247.

Frisch, A.S., J.R. Holbrook, and A.B. Ages, 1981. Obervations of a summertime reversal in the circulation in the Strait of Juan de Fuca. Journal Geophysical Research, 86, 2044-2048.

Gade, H.G., and A. Edwards, 1980. Deep water renewal in fjords. In: Fjord Oceanography, H.J. Freeland, D.M. Farmer, and C.D. Levings (eds.), Plenum Press, N.Y. 453-489.

Geyer, W.R., and G.A. Cannon, 1982. Sill processes related to deep-water renewal in a fjord. Journal of Geophysical Research, 87, 7985-7996.

Hamilton, P., J.T. Gunn, and G.A. Cannon, 1983. A box model of Puget Sound. Journal of Geophysical Research, 88, (submitted).

Herlinveaux, R.H., 1954. Tidal currents in Juan de Fuca Strait. Journal Fisheries Research Board Canada, 11, 799-815.

Herlinveaux, R.H., and J.P. Tully, 1961. Some oceanographic features of Juan de Fuca Strait. Journal Fisheries Research Board Canada, 18, 1027-1071.

Holbrook, J.R., and G.A. Cannon, 1983. Propagation and entrainment of renewed bottom water in a fjord. Transactions American Geophysical Union, 64, 248.

Holbrook, J.R., and A.S. Frisch, 1981. A comparison of new-surface CODAR and VACM measurements in the Strait of Juan de Fuca, August 1978. Journal of Geophysical Research, 86, 10908-10912.

Holbrook, J.R., and D. Halpern, 1982. Wintertime near-surface currents in the Strait of Juan de Fuca, Atmosphere-Ocean, 20, 237-339.

Holbrook, J.R., G.A. Cannon, and D. Kachel, 1983. Two-year observations of coastal-estuarine interaction in the Strait of Juan de Fuca. In: Coastal Oceanography Physical Aspects, H. Gade, H. Svendsen, and A.Edwards (eds.), Plenum Press, New York, (in press).

Holbrook, J.R., R.D. Muench, and G.A. Cannon, 1980a. Seasonal observations of low-frequency atmospheric forcing in the Strait of Juan de Fuca. In: Fjord Oceanography, H.J. Freeland, D.M. Farmer, and C.D. Levings (eds.), Plenum Press, New York, 305-317.

Holbrook, J.R., R.D. Muench, D.G. Kachel, and C. Wright, 1980b. Circulation in the Strait of Juan de Fuca: recent oceanographic observations in the eastern basin. NOAA Technical Report ERL 412-PMEL 33, 42 pp.

Klinck, J.M., J.J. O'Brien, and H. Svendsen, 1981. Simple model of fjord and coastal circulation interaction. In: The Norwegian Coastal Current, R. Saetre and M. Mork (eds.), University of Bergen, 178-214.

Mofjeld, H.O., and L.H. Larsen, 1983. Tides and tidal currents in Puget Sound and the Strait of Juan de Fuca-Strait of Georgia system. In: Puget Sound. Oceanography of the Inshore Waters of Washington, D. Henry (ed.), Univ. of Wash. Press, Seattle (resubmitted).

Morse, R.M., M. Rattray, R.G. Paquette, and C.A. Barnes, 1958. The measurement of transports and currents in small tidal streams by an electromagnetic method. University of Washington, Deptartment of Oceanography, Technical Report 57, 70 pp.

Muench, R.D., and J.R. Holbrook, 1980. Vertical structure of fluctuating currents in the Strait of Juan de Fuca. In: Fjord Oceanography, H.J. Freeland, E.M. Farmer, and C.D. Levings (eds.), Plenum Press, New York, 319-327.

Paquette, R.G.,and C.A. Barnes, 1951. Measurement of tidal currents in Puget Sound. University of Washington, Department of Oceanography, Technical Report 6, 28 pp.

Pickard, G.L., and K. Rodgers, 1959. Current measurements in Knight Inlet, British Columbia. Journal Fisheries Research Board Canada, 16, 635-678.

Rattray, M., 1967. Some aspects of the dynamics of circulation in fjords. In: Estuaries, G.H. Lauff (ed.), American Association Advancement Science, Washington, D.C., 52-62.

Rattray, M., and J.A. Proehl, 1982. Baroclinic response of wide, deep estuaries to offshore atmospheric forcing. Transactions American Geophysical Union, 63, 1012 (abstract).

Svendsen, H., 1981. Wind induced variations of water level and circulation in fjords. In: The Norwegian Coastal Current, R. Saetre and M. Mork (eds.), University of Bergen, 229-262.

University of Washington, Department of Oceanography, 1954. Puget Sound and approaches, a literature survey. Vol. 3, 175 pp. (unpublished manuscript).

Wang, D.-P., and A.J. Elliott, 1978. Non-tidal variability in the Chesapeake Bay and Potomac River: evidence for non-local forcing. Journal of Physical Oceanography, 8, 225-232.

Winter, D.F., 1973. A similarity solution for steady-state gravitational circulation in fjords. Estuarine and Coastal Marine Science, 1, 387-400.

Winter, D.F., K. Banse, and G.C. Anderson, 1975. The dynamics of phytoplankton blooms in Puget Sound, a fjord in the northwestern United States. Marine Biology, 29, 139-176.

Wright, D.G., 1980. On the stability of a fluid with specialized density stratification. Part II: mixed baroclinic-barotropic instability with application to the Northeast Pacific. Journal of Physical Oceanography, 10, 1307-1322.

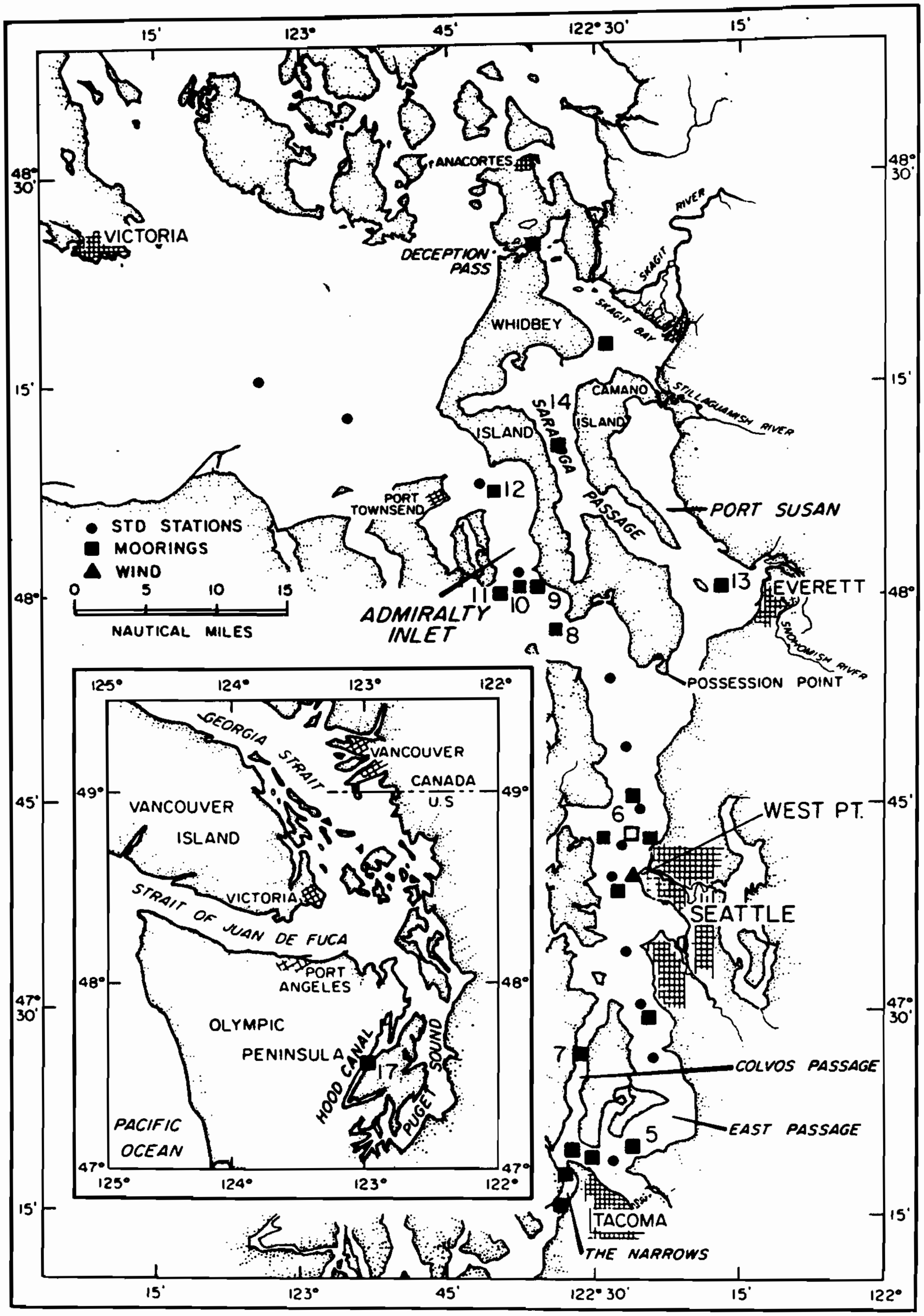


Figure 1. Station locations in Puget Sound. Mooring 6 (open square) just north of Seattle is location of studies in 1972, 1973, 1975-1976, 1977, 1979, 1981, 1982, and 1983. Numbered stations are referenced in text.

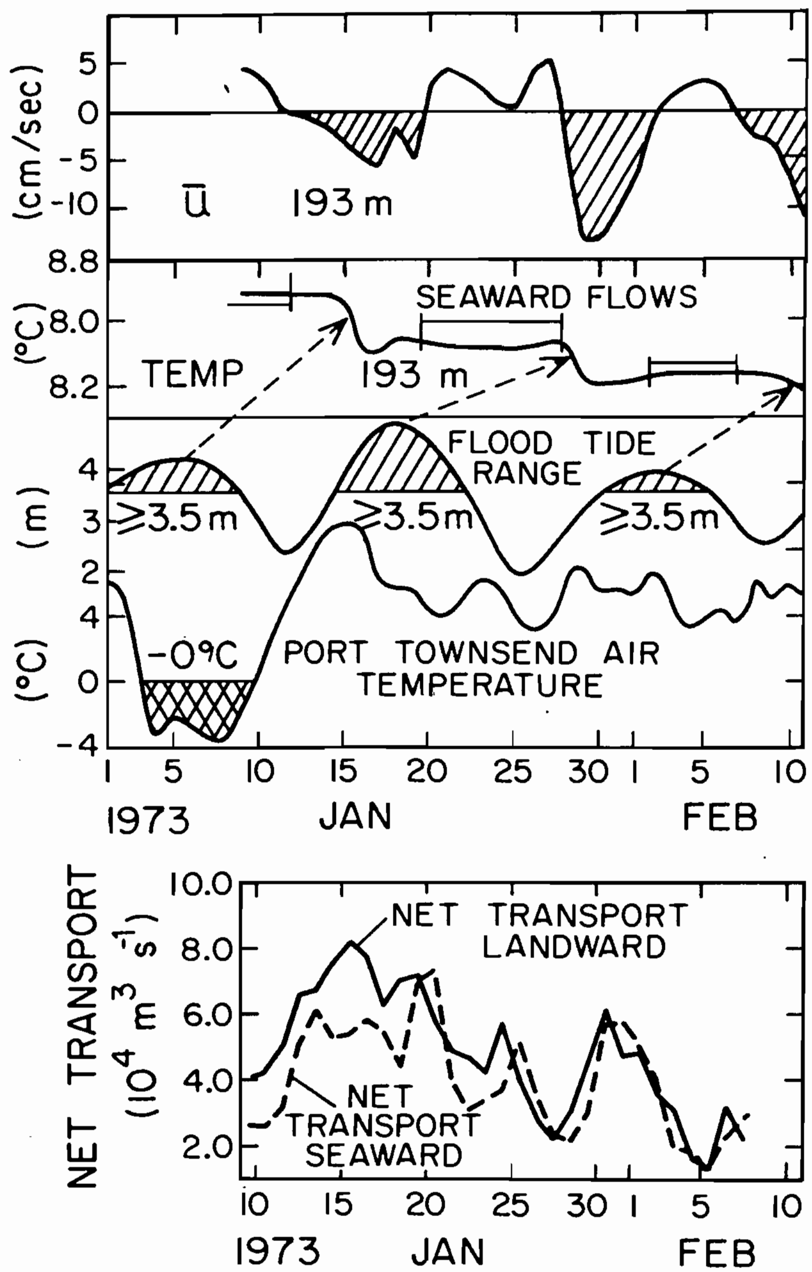


Figure 2. Daily average longitudinal bottom currents and temperature at mooring 6 (Fig. 1); range of greatest daily flood tide at Seattle; daily average air temperature at Port Townsend; and net landward and seaward transport past mooring 6. Adapted from Cannon and Ebbesmeyer (1978).

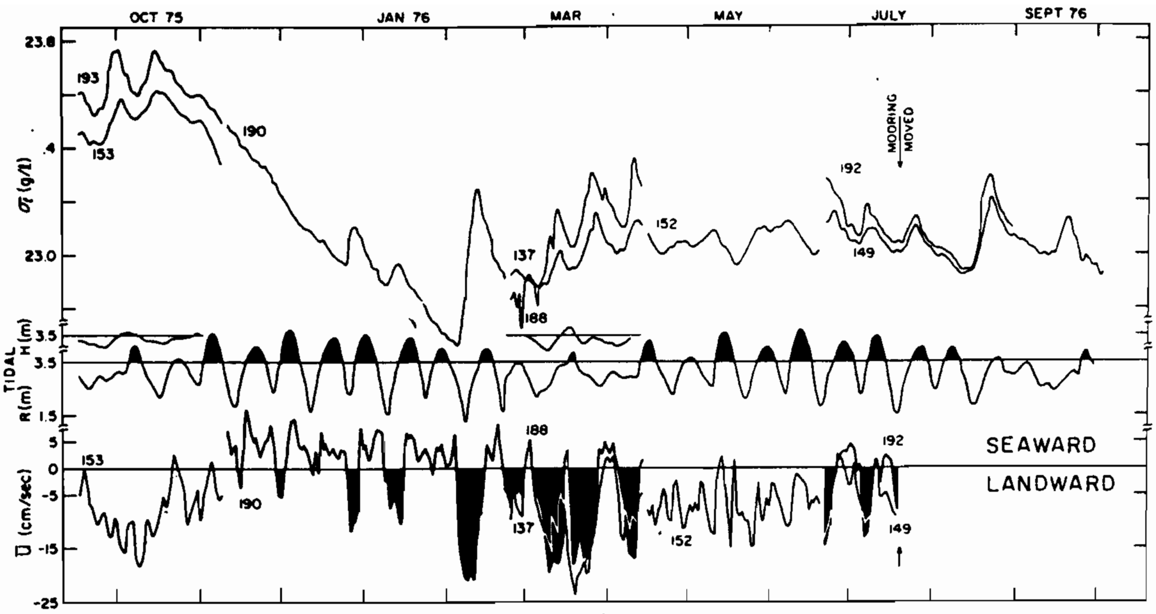


Figure 3. Daily average sigma-t at 5 and 50 m above the bottom at mooring 6 (Fig. 1); height (H) and range (R) of greatest daily flood tide at Seattle and daily average along-channel currents at 5 and 50 m above the bottom. From Cannon and Laird (1978).

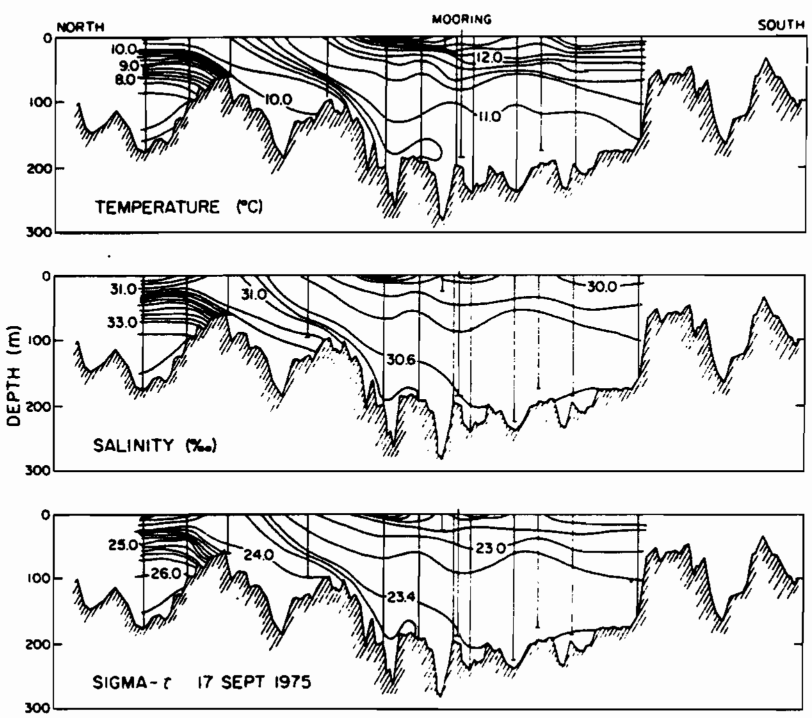
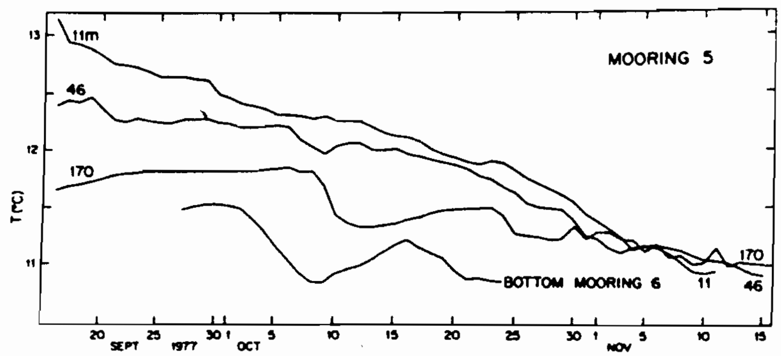


Figure 4. Temperature, salinity, and sigma-t from Strait of Juan de Fuca through Admiralty Inlet and along Puget Sound Main Basin during the first intrusion of deep water indicated by the year-long mooring (Fig. 3). Vertical lines show stations shown by dots on Fig. 1. From Cannon and Laird (1978).



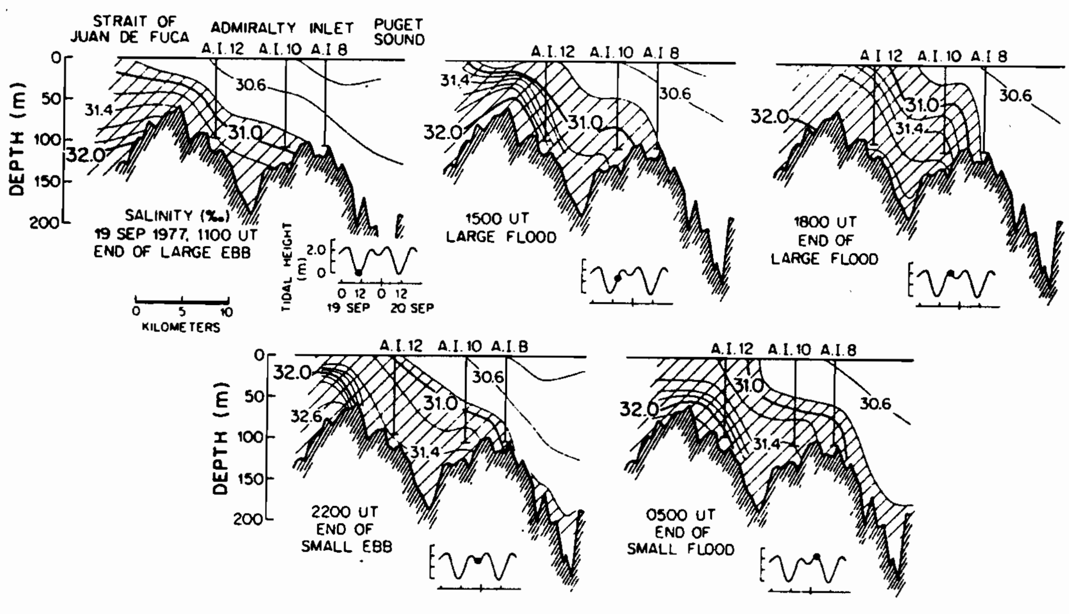


Figure 5. *Top:* Daily average temperature at the indicated depths at mooring 5 in East Passage near Tacoma (Fig. 1) and from near bottom at mooring 6 near Seattle about 46 km north. From Cannon and Laird (1980). *Bottom:* Salinity sections at various phases of a neap tide. Shaded part is more saline than that in the Main Basin. From Geyer and Cannon (1982).

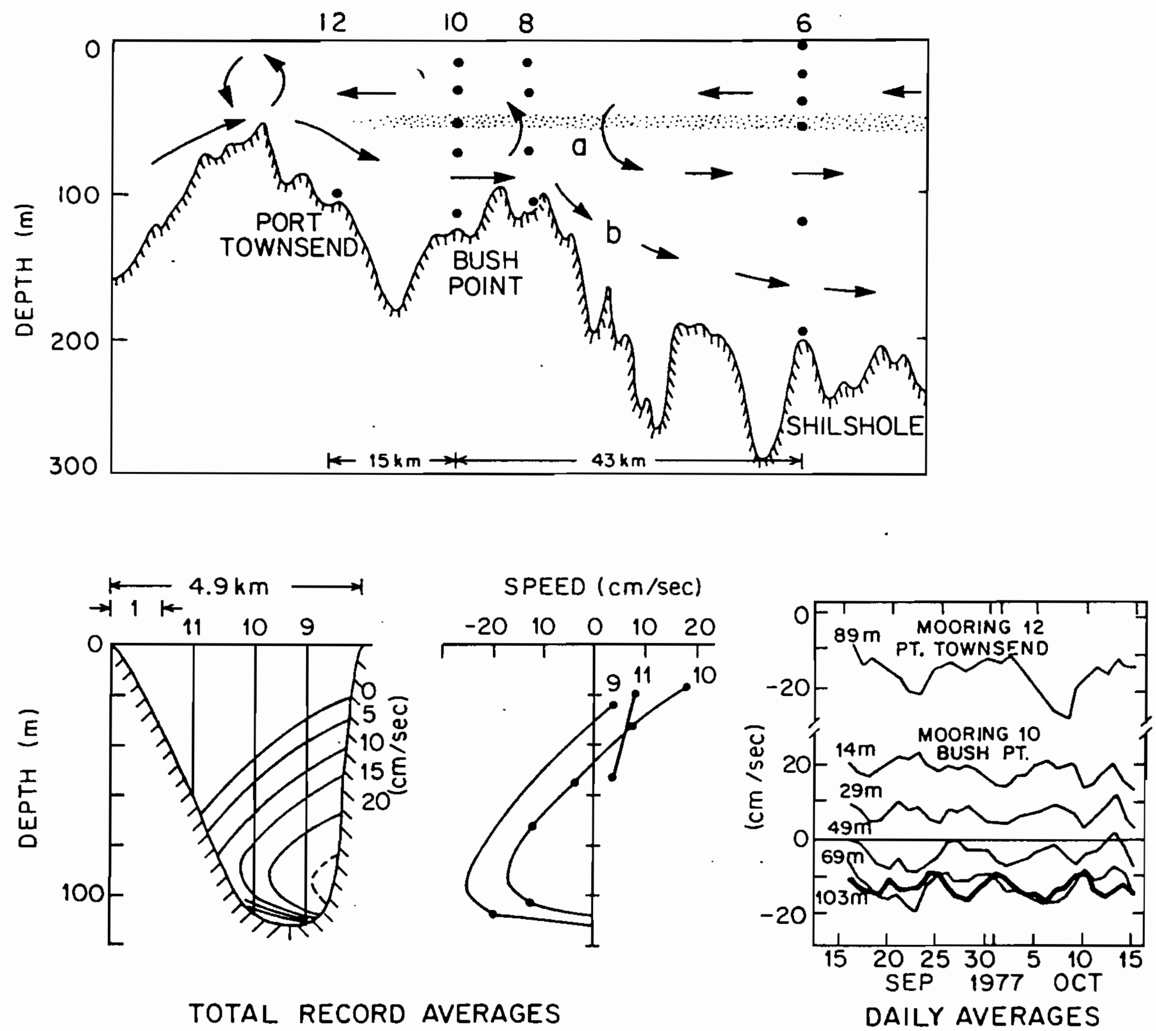


Figure 6. Top: Schematic representation of the recirculation or entrainment of seaward flowing surface water (a) with incoming new bottom water (b). Stippling represents approximate depth of no motion. Dots show locations of current meters on numbered moorings (Fig. 1). Adapted from Cannon and Ebbesmeyer (1978). Bottom: Currents averaged over 2 month-interval (September-November 1977) through the 3-mooring section off Bush Point midway along Admiralty Inlet and daily average currents (12) and at Port Townsend at the middle mooring off Bush Point (10) at the indicated depths. Positive flow is seaward.

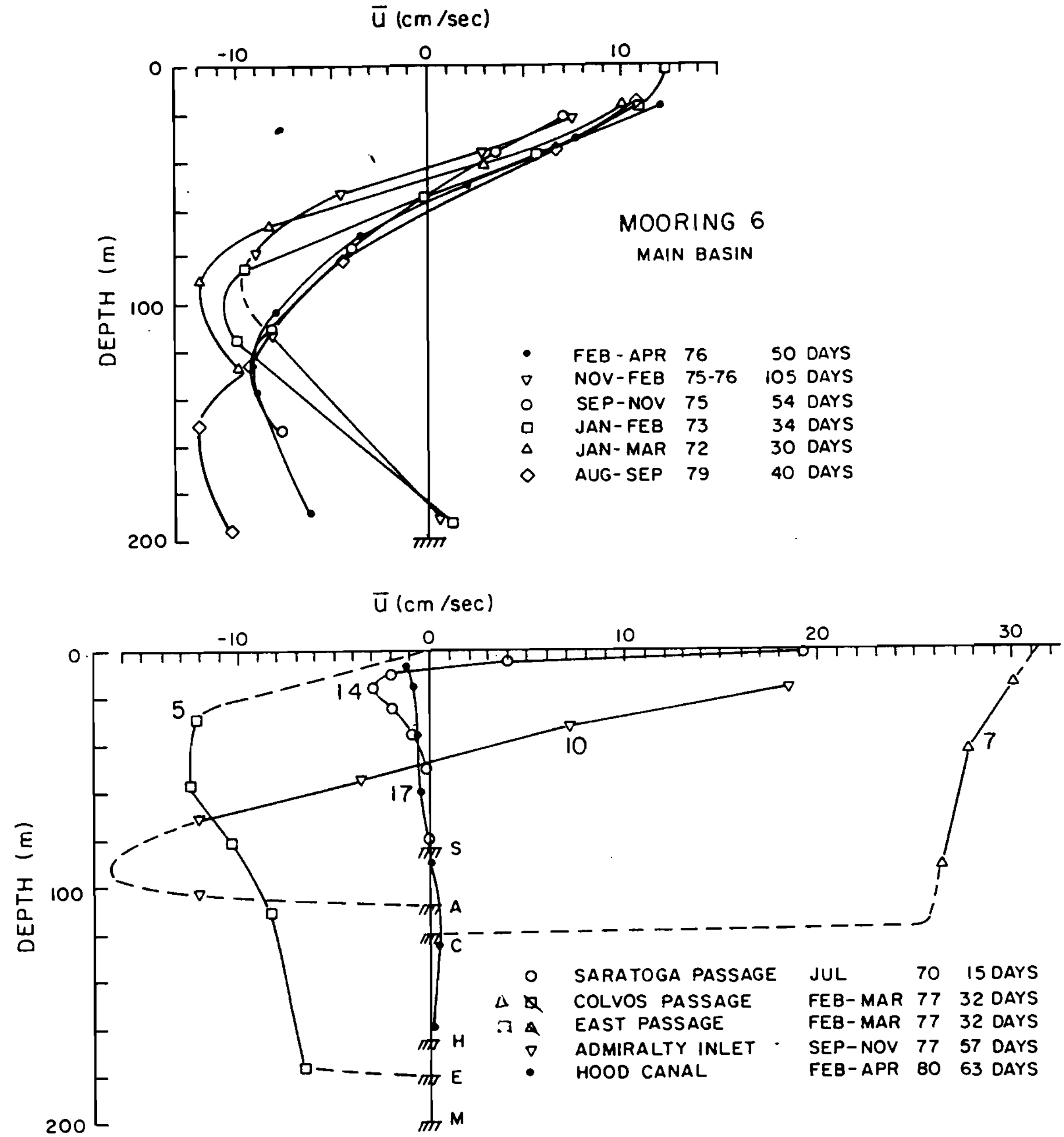


Figure 7a. *Top:* Along-channel current profiles averaged over indicated observation intervals at mooring 6 in the Main Basin. *Bottom:* Profiles at five other locations (mooring numbers on profiles; Fig. 1) in Puget Sound. Positive flow is seaward.

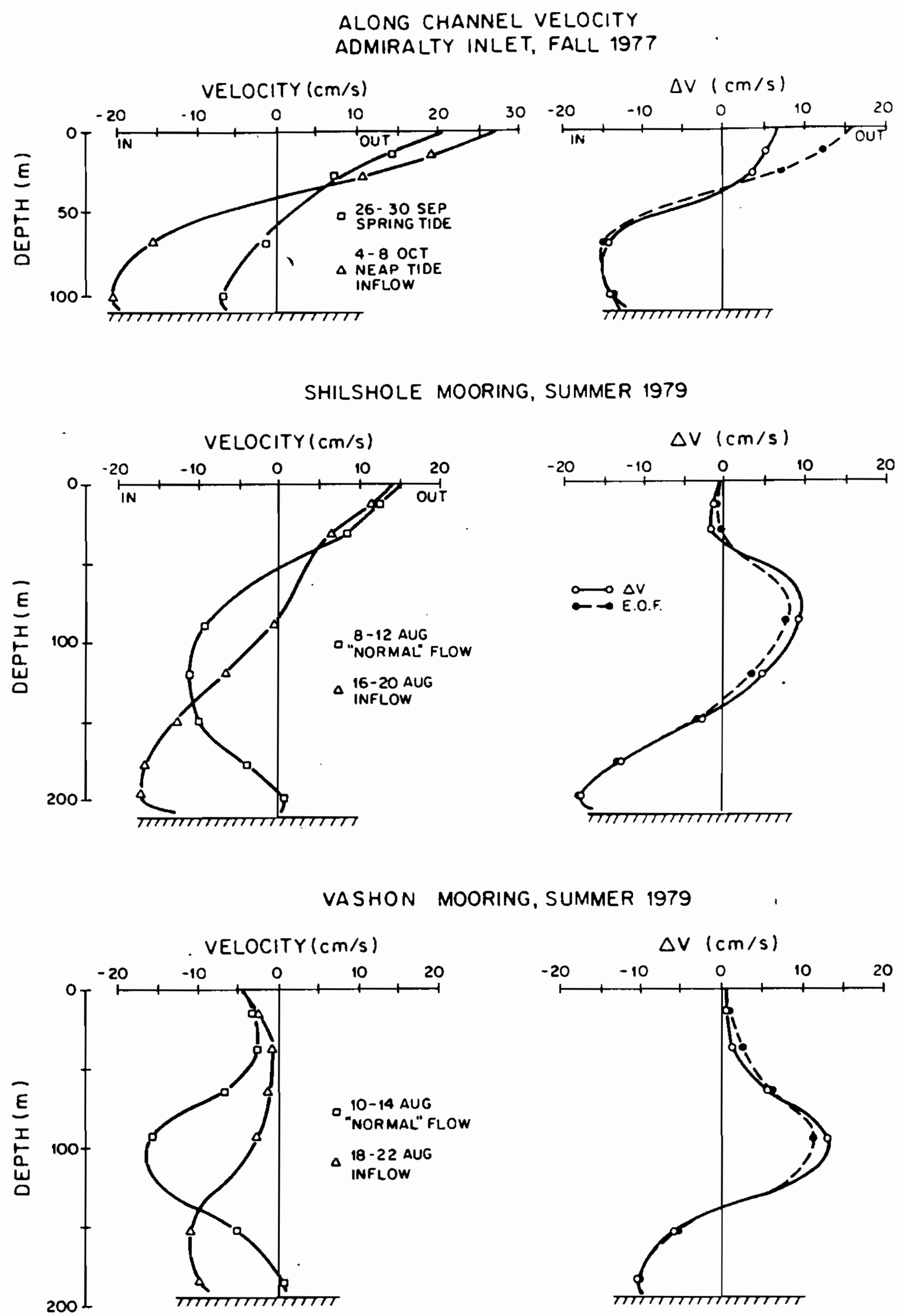


Figure 7b. Vertical profiles of tidally averaged velocities during periods with (open squares) and without (open triangles) inflow in the bottom of the Main Basin, and profiles of velocity difference (ΔV) during the two regimes (open circles) and the first eigenvector of 35-hr-filtered velocity variance (solid circles). From Geyer and Cannon (1983).

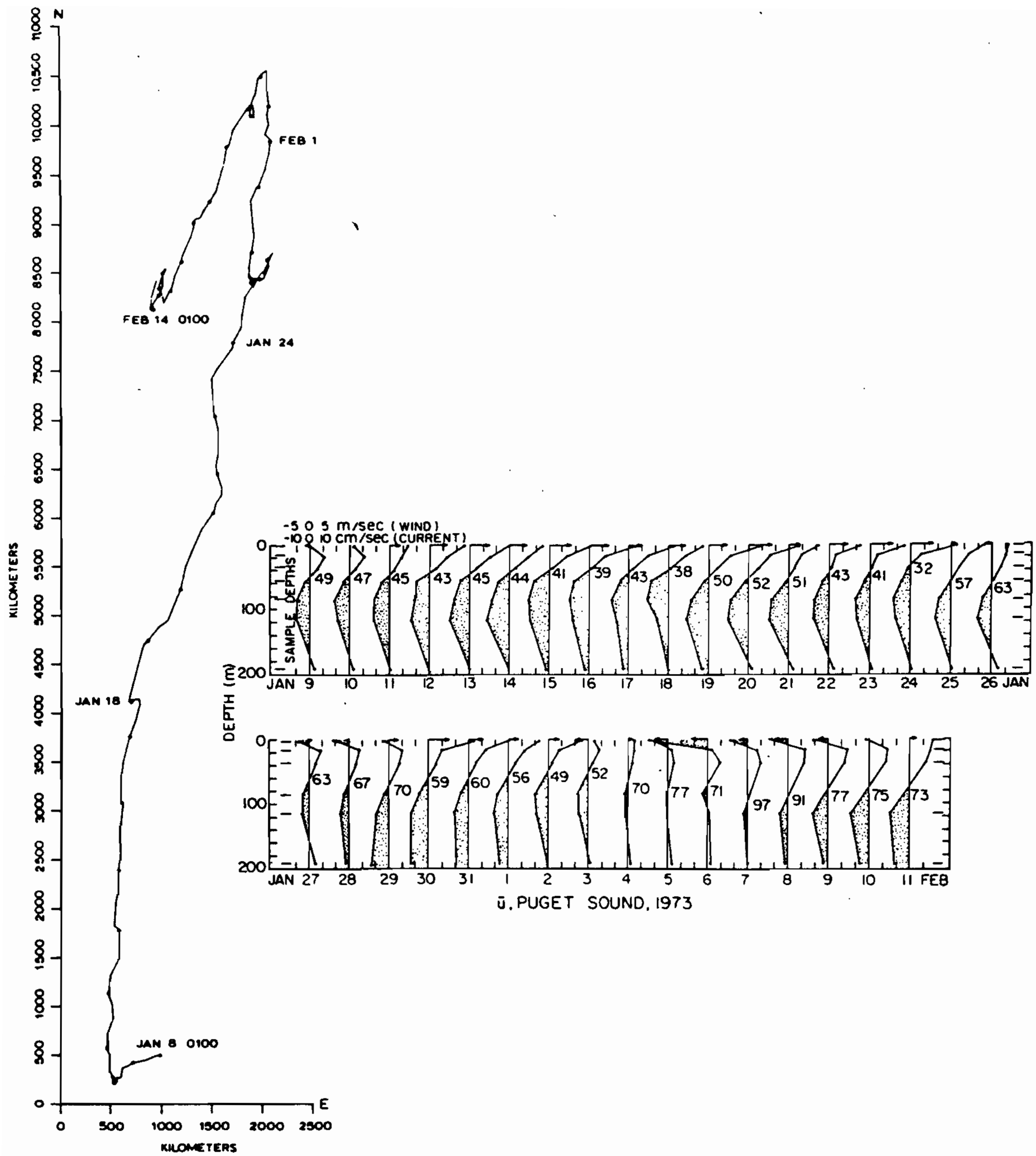


Figure 8. *Left:* Progressive vector diagram of winds at West Point (just south of mooring 6) showing direction to which winds blow. *Right:* Daily average along-channel current profiles at mooring 6. Daily average north-south winds are shown by the vectors at 0 m on each profile. Depth of no motion is given for each profile. Positive flow is northward (seaward).

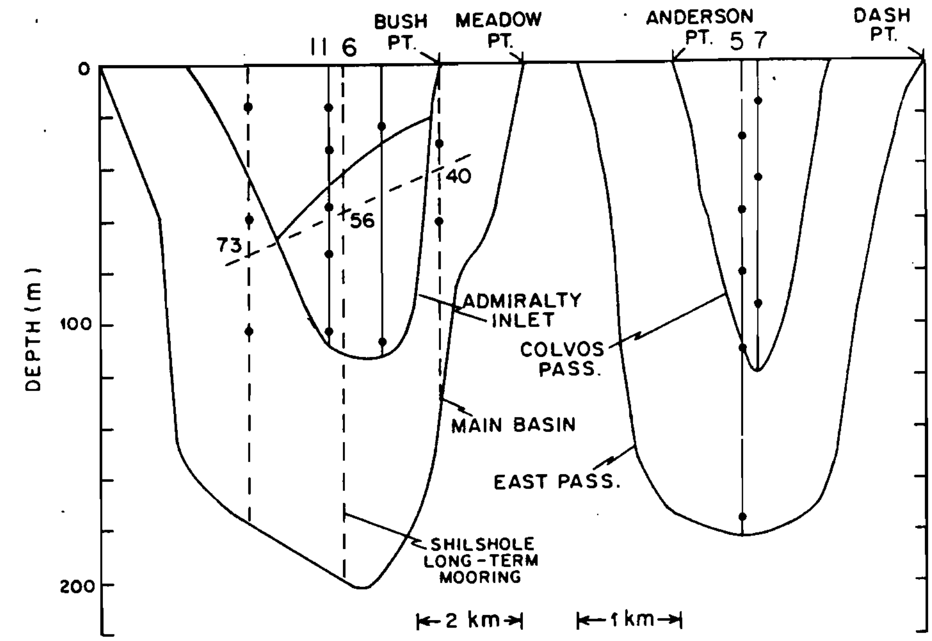


Figure 9. Cross sections (looking seaward) in Admiralty Inlet, Main Basin, East Passage and Colvos Passage where volume fluxes have been calculated. Locations are indicated by middle mooring number (Fig. 1) and the nearest point of land. Current meter locations except for the long-term mooring. Dashed and solid sloping lines indicate zero levels for the Main Basin and Admiralty Inlet, respectively. Horizontal scales are different for the two pairs of sections.

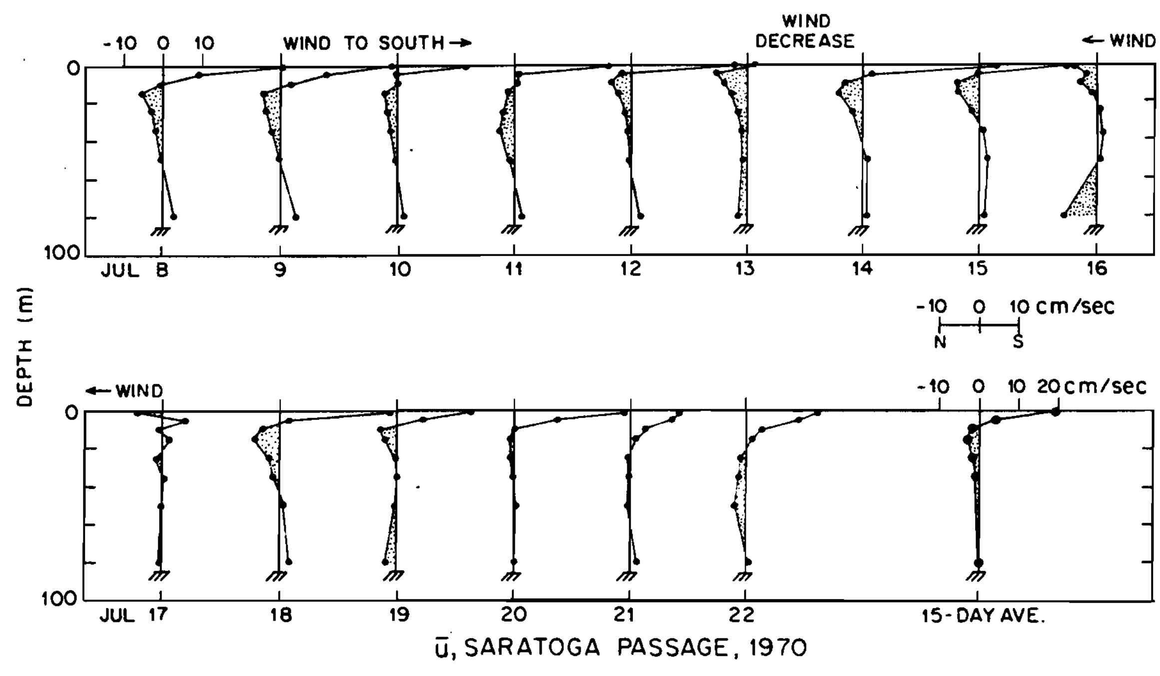


Figure 10. Daily average along-channel current profiles for Saratoga Passage showing along-channel wind direction measured at Sandy Point about 18 km south. Also shown is the total record average from Fig. 7 on the same scale. Positive flow is southward (seaward).

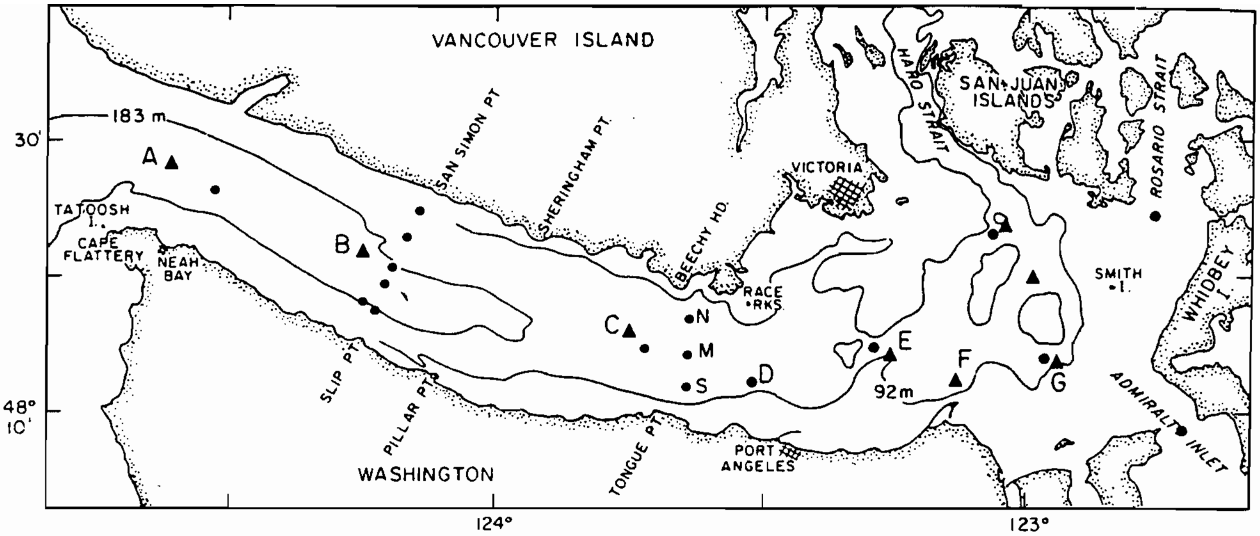


Figure 11. Location of surface moorings (triangles) and subsurface moorings (circles) in the Strait of Juan de Fuca. Stations with letters are referenced in text.

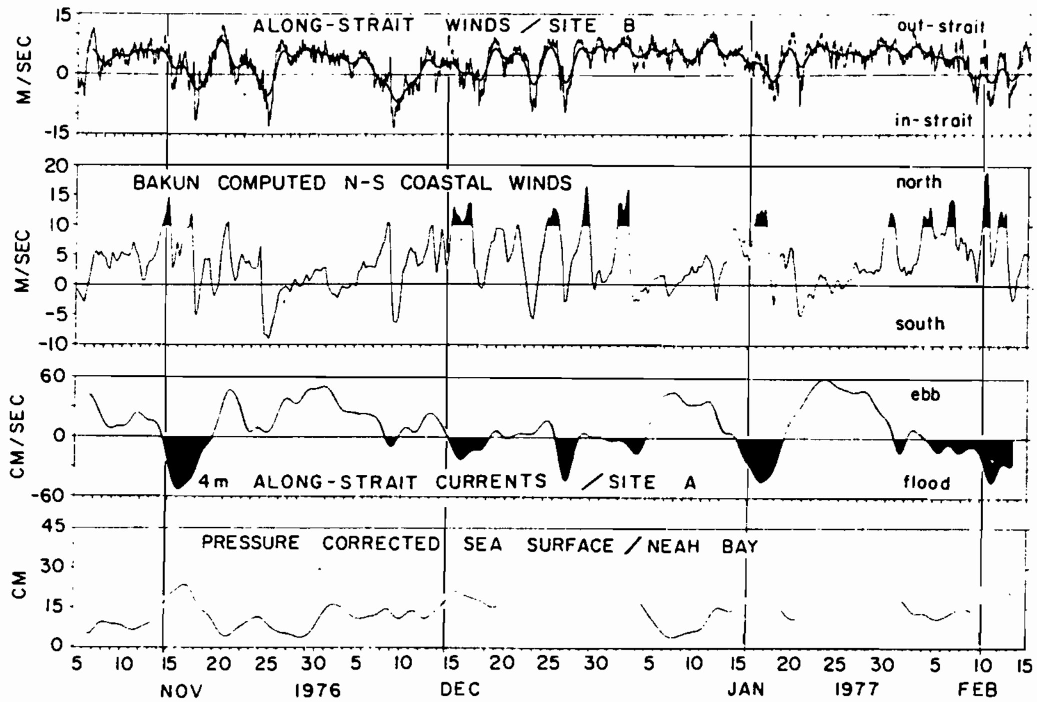


Figure 12. Along-channel (mooring B; Fig. 11) and coastal winds (computed), along-channel currents at the mouth of the Strait (mooring A), and sea-surface elevations (Neah Bay). Vertical lines indicate various correlations. From Holbrook and Halpern (1982).

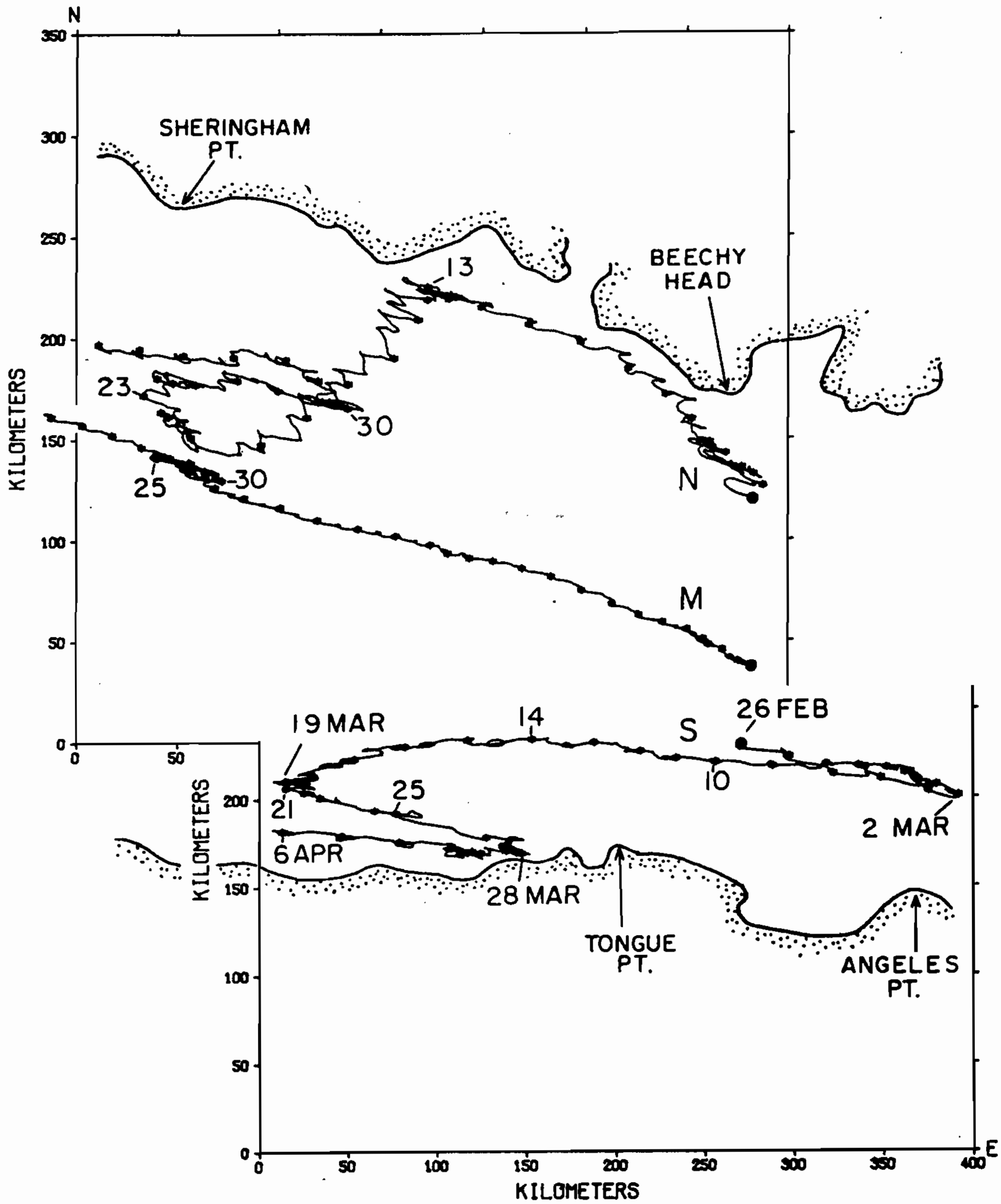


Figure 13. Progressive vector diagram of 15-m flow across Tongue Point section (N, M, S; Fig. 11). All records start at the mooring locations on 26 February 1976. Scale for mooring M is double that shown. Scale of the coastline is about 10% that shown. North is along the vertical axis. From Cannon, ed. (1978).

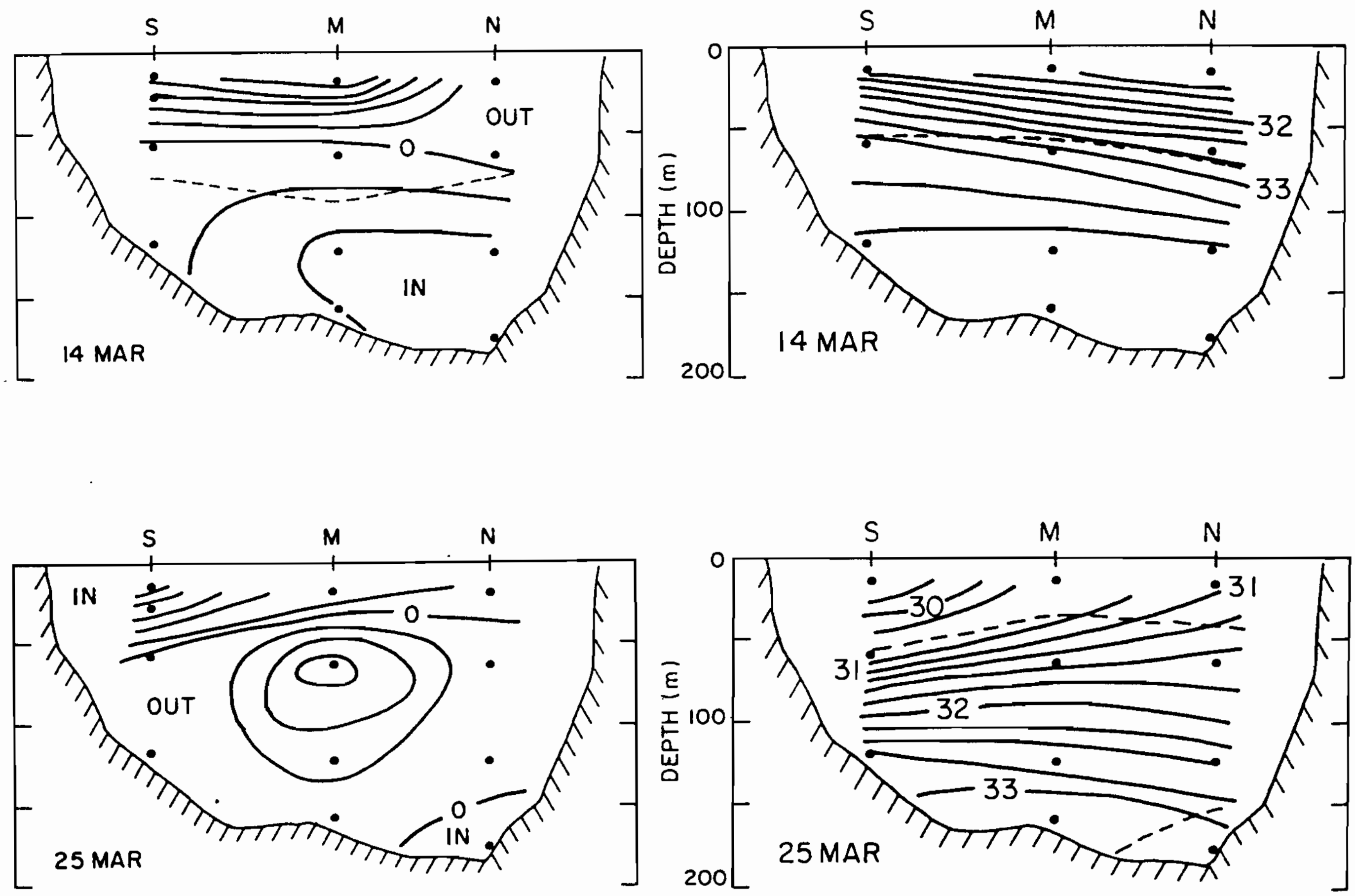


Figure 14. Daily average currents and salinity across Tongue Point section (Fig. 13) during normal estuarine flow (14 March) and during one reversal of flow including an intrusion of coastal water (25 March). Current contour interval is 5 cm/sec. Dashed line on 14 March current chart is total-record average zero level; dashed lines on salinity charts indicate zero currents for that day. From Cannon and Holbrook (1981).

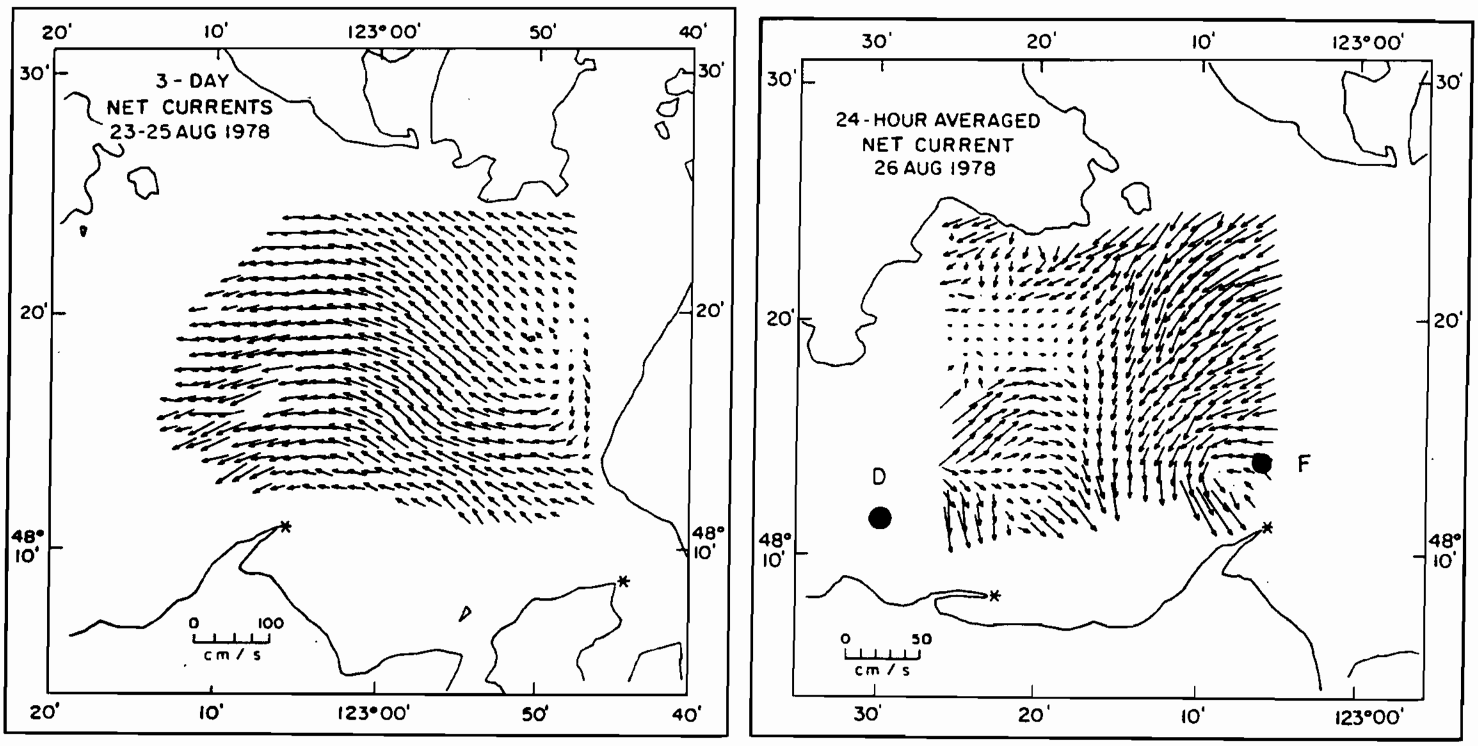


Figure 15. Three-day averaged and subsequent 24-hr averaged surface currents as measured by the CODAR current mapping system (Barrick et al., 1977). Asterisks mark locations of the shore-based HF Dopplar radar units. D and F indicate current meter mooring locations (Fig. 11). From Frisch et al. (1981) and Holbrook et al. (1980b).