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THE CIRCULATION OF MONTEREY BAY AND RELATED PROCESSES*

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Abstract The surface circulation of Monterey Bay is relatively weak. Early attempts to ascertain this circulation are initially summarized. Recent results indicate that the surface circulation is predominantly northward (i.e. cyclonic) with speeds usually in the range of 5 to 20 cm/sec; however, major reversals in flow direction do occur. The influx of fresh water, although relatively small, plus seasonal heating and residual tidal influence may all contribute to northward flow inside the Bay. During spring and summer, cooler waters which often occur across the entrance of Monterey Bay are most likely due to both local and advective processes.

Temperatures at intermediate depths in Monterey Bay (~25 to ~150 m) suggest that geostrophic flow within the thermocline may be opposite to that at the surface (i.e. anticyclonic). However, reversals in flow direction at depth from anticyclonic to cyclonic may occur when offshore flow in the California Undercurrent is weak. Seasonal changes in the deep circulation in Monterey Bay may be related to seasonal changes in the strength of the California Undercurrent. The deep flow in Monterey Submarine Canyon is vigorous (up to ~100 cm/sec) and frequently upcanyon, and oscillations in current speed and direction are often supertidal (i.e. of higher frequency). Nonlinear effects associated with very high amplitude internal waves may contribute to onshore flow within the Canyon. Supertidal frequency oscillations may also arise from nonlinear effects, and superinertial frequency oscillations may occur due to the narrowness of Monterey Submarine Canyon.

Residence times for bay waters estimated from sea surface temperatures (SSTs) inside and outside the Bay range from 5 to 12 days. Mean internal Rossby radii of deformation range from 10 km over Monterey Submarine Canyon to about 1 km around the periphery of the Bay, reflecting the strong influence of bottom depth. A scale analysis suggests that several processes, in addition to those usually indicated for the deep ocean may be important in the Monterey Bay coastal region.

Coastal upwelling through advection from outside the Bay, open ocean upwelling through positive wind stress curl and deep upwelling in Monterey Submarine Canyon may all contribute to the upwelled waters found in Monterey Bay. These waters, which are enriched through this unique combination of upwelling-related processes, most likely account for the very high biological productivity that characterizes this region.

A number of additional processes affect the circulation of Monterey Bay including winds, internal waves, mixing, tides, local heating and river discharge, eddies, oceanic fronts, spring transition events, 40–50 day oscillations and El Niño episodes. These processes are described.

The circulation in Monterey Bay is also strongly influenced by the circulation offshore. The circulation offshore is complex, consisting of eddies, interleaving alongshore flows involving

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the interaction of different water masses, and offshore jets. This complexity may be due, in part, to the presence of the Bay itself and the Canyon.

Finally, a conceptual model of bay circulation is presented that reflects a synthesis of the available observations and theory. Because of the importance of Monterey Submarine Canyon in influencing the circulation within Monterey Bay, 16 other bay/canyon systems are identified globally where canyons may influence the local circulation.

Introduction

Monterey Bay (MB) is located along the central California coast between 36.5 and 37°N (Fig. 1); it is 37km wide (between Point Pinos and Terrace Point) and is 19km from a line connecting Point Pinos and Terrace Point at the point of maximum depth (950m), to Moss Landing. It is semi-enclosed, has a free connection with the open sea and receives limited amounts of fresh water from several streams. MB is not an estuary because it is deep and broad and is not significantly diluted by the fresh water it receives except locally during brief periods of high river discharge.

The Bay is symmetrical in shape and covers an area of approximately 550km^2. The Monterey Submarine Canyon (MSC) is the major topograhic feature in MB and divides it more-or-less equally into northern and southern sectors. MSC is the largest submarine canyon along the west coast of North America (with the exception of the Bering Sea), having a volume of 420km^3 (Martin 1964). Two transects across the Canyon are shown in Figure 2. From Terrace Point to Point Pinos (transect B–B′), the distance across the Canyon (i.e. across the entrance of MB) at a depth of 150m (just below the canyon lip) is about 12km. Along this transect the maximum bottom depth approaches 900m. Further offshore, the Canyon width increases rapidly. Further inside the Bay, the Canyon width decreases significantly, to as short as approximately 3km at 150m, along transect A–A′. Although MSC represents a major depression that cuts across the continental shelf and slope, a significant fraction of the Bay is shallow. Approximately 80% of the Bay is shallower than 100m and 5% is deeper than 400m. There are two bights in MB, one to the south near Monterey and a second to the north between Santa Cruz and Aptos.

Offshore, the California Current transports relatively cool, fresh, subarctic water equatorward along the California coast (Reid et al. 1958). The mean flow is weak and instantaneous flows may often be poleward, especially near the coast. Along the central California coast, winds from the NW associated with the Subtropical High Pressure Cell produce coastal upwelling. Coastal upwelling influences coastal circulation and thermal structure strongly in this region and often starts abruptly with the so-called spring transition. At the latitude of MB, coastal upwelling usually occurs between March and October in accordance with the usually persistent upwelling-favourable winds. A frequently used indicator for the intensity of coastal upwelling is the upwelling index which provides a quantitative measure of wind-driven offshore Ekman transport (e.g. Bakun 1973). Figure 3 shows the daily (and weekly-averaged) time series of upwelling index on the coast at 36°N just south of MB for 1980. Coastal upwelling is relatively intense at this location compared to other locations along the US West Coast. The index becomes strongly positive in March at the time of the spring transition and gradually weakens during late summer and fall.

By early November, when coastal upwelling relaxes, a narrow (50 to 100km) near-shore countercurrent becomes established. Below depths of ~150m, this flow is termed the California Undercurrent and is present more-or-less year-round. During winter, the

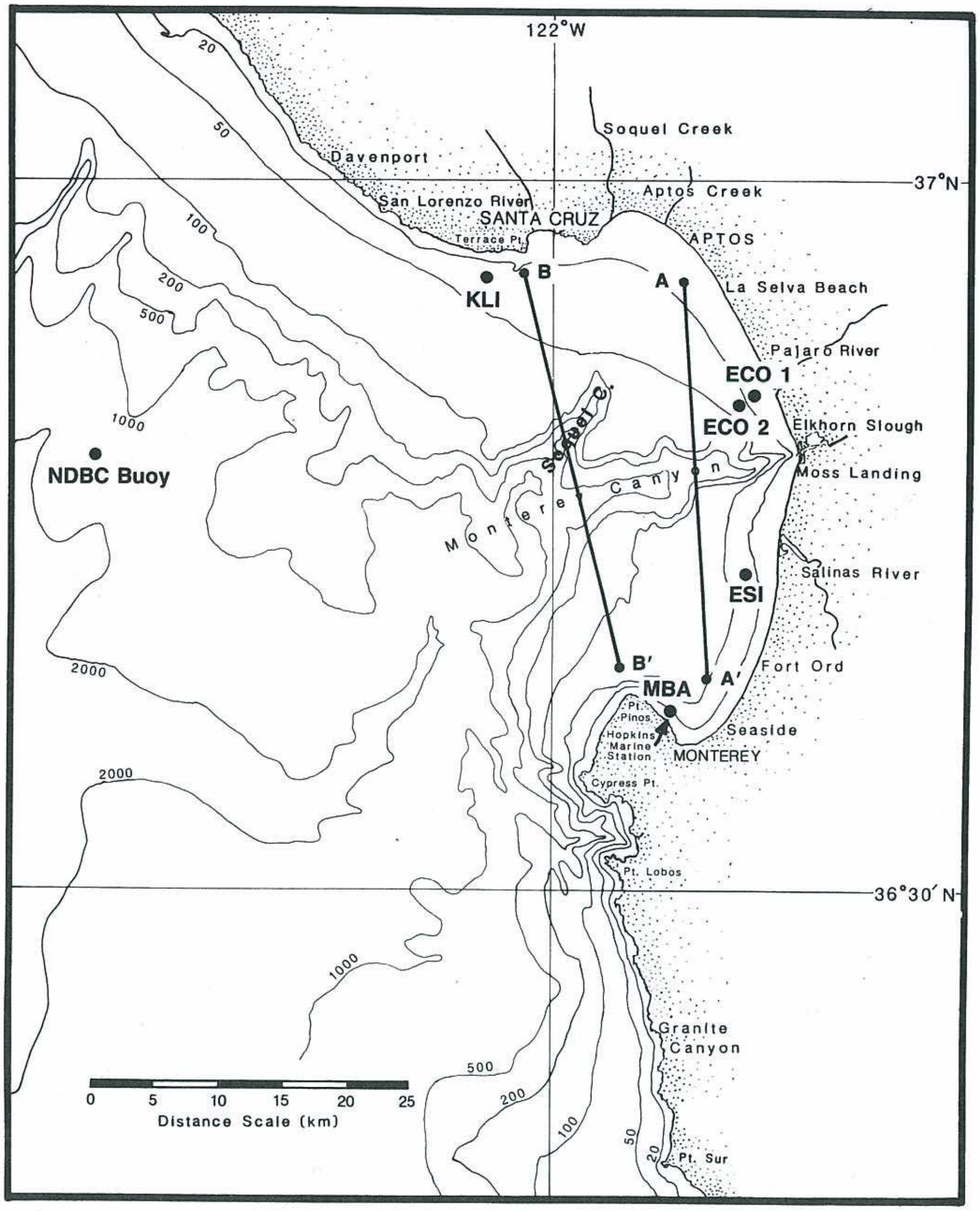

Figure 1 Map of Monterey Bay and surrounding area. Lines A–A′ and B–B′ represent the locations of vertical sections shown in Figure 2. Depths in metres.

Undercurrent and the northward-flowing surface current (often called the Davidson Current) may be indistinguishable. Pacific Subarctic Waters originating in the West Wind Drift and Pacific Equatorial Waters originating at lower latitudes are found in varying, but roughly equal, amounts in MB throughout the year.

A well-defined and rather unique relationship exists between biological productivity and nutrient enrichment in MB which results from upwelling-related processes. Phytoplankton

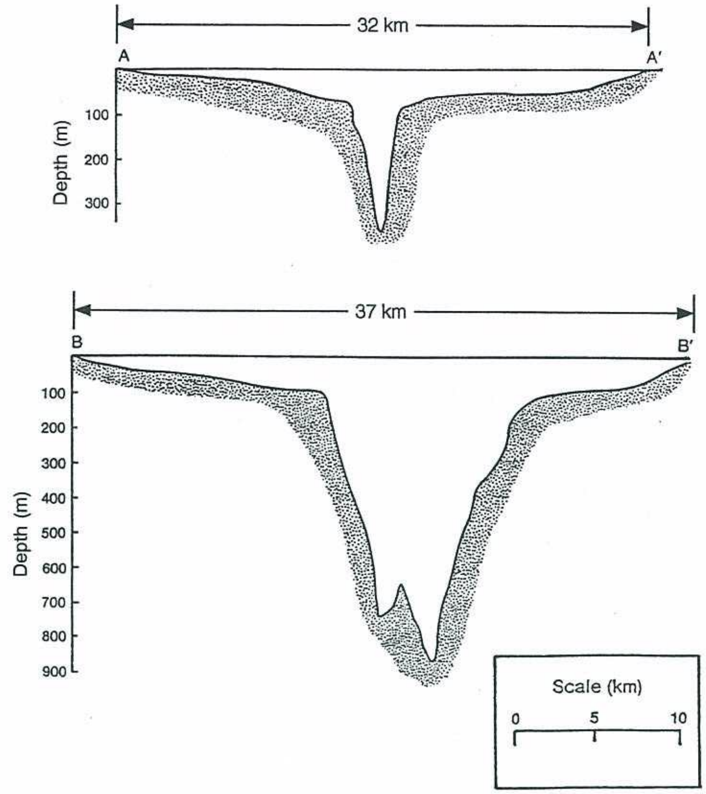

Figure 2 Vertical sections of bottom depth (A–A′ and B–B′) across Monterey Submarine Canyon (shown in plan view in Fig. 1). (Adapted from Scott 1973).

volume lags behind the seasonal cycle of upwelling by 1–2 months (Barham 1956). This lag apparently corresponds to the time required for the enriched waters which rise in MSC to spread out from the centre of the Bay to the shallower shelf areas where phytoplankton growth exceeds that which is usually observed over the centre of the Bay. An example of biological productivity which depicts the distribution of surface chlorophyll in MB and the surrounding area is shown in Plate 1. This chlorophyll map is based on a Coastal Zone Color Scanner (CZCS) satellite image from 15 June 1981 (areas depicted in white indicate the highest concentrations of surface chlorophyll encountered, $\geq \sim 10\,mg/m^3$). The highest chlorophyll concentrations along the central California coast occur within and just beyond MB. Also, slightly lower levels of chlorophyll occur over the centre of MB. These attributes are generally consistent with other CZCS satellite coverage of the MB area, particularly during the upwelling season (Hauschildt 1985).

Because conditions in MB reflect much of the variability that occurs offshore in the California Current (e.g. Skogsberg 1936), this bay differs from many of the other bays and estuaries bordering the US West Coast. Overall, the state of knowledge with respect to the general circulation of MB is less complete than for a number of other bays bordering the US

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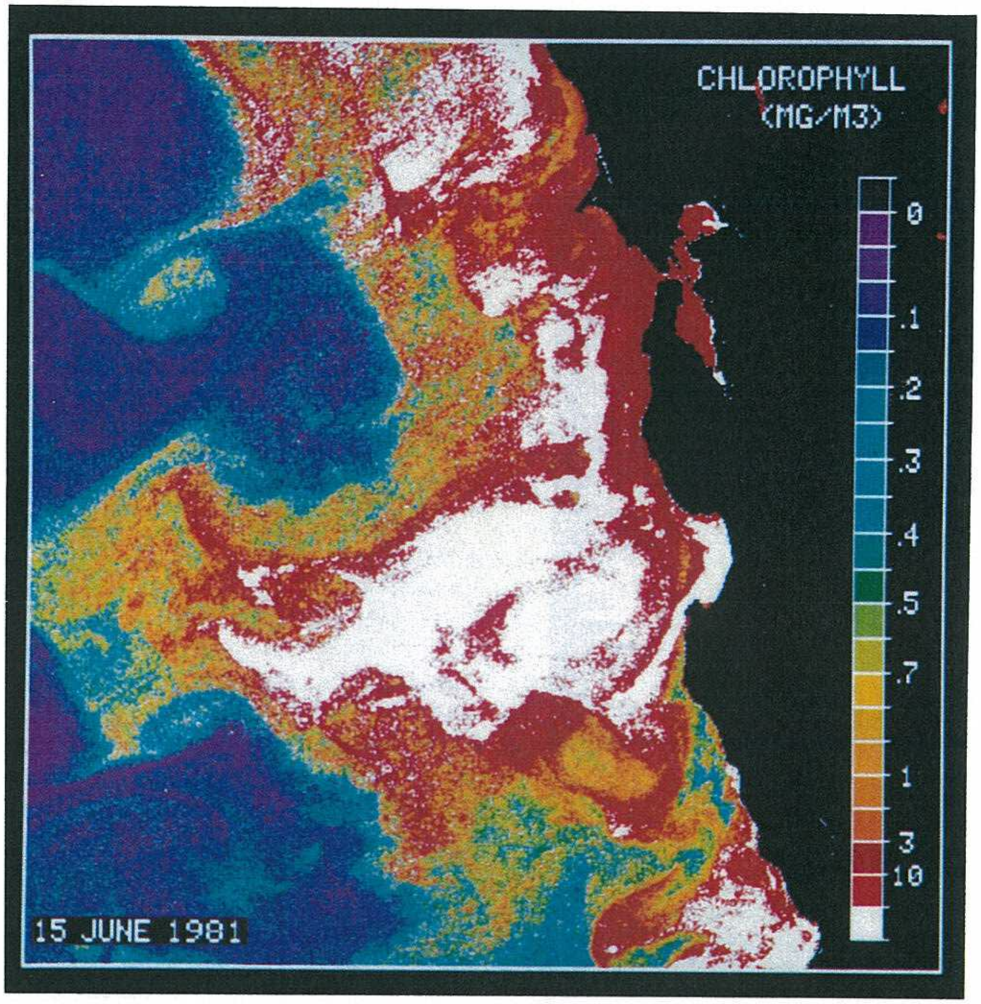

Plate 1 Coastal Zone Color Scanner (CZCS) satellite image from 15 June 1981. Image shows the concentration of surface chlorophyll (i.e. upper few tens of metres) along the central California coast including Monterey Bay. Areas in white indicate surface chlorophyll concentrations of ~10 mg/m^3, or greater; because of the uncertainties inherent in such measurements, no attempt has been made to resolve greater concentrations, if they exist.

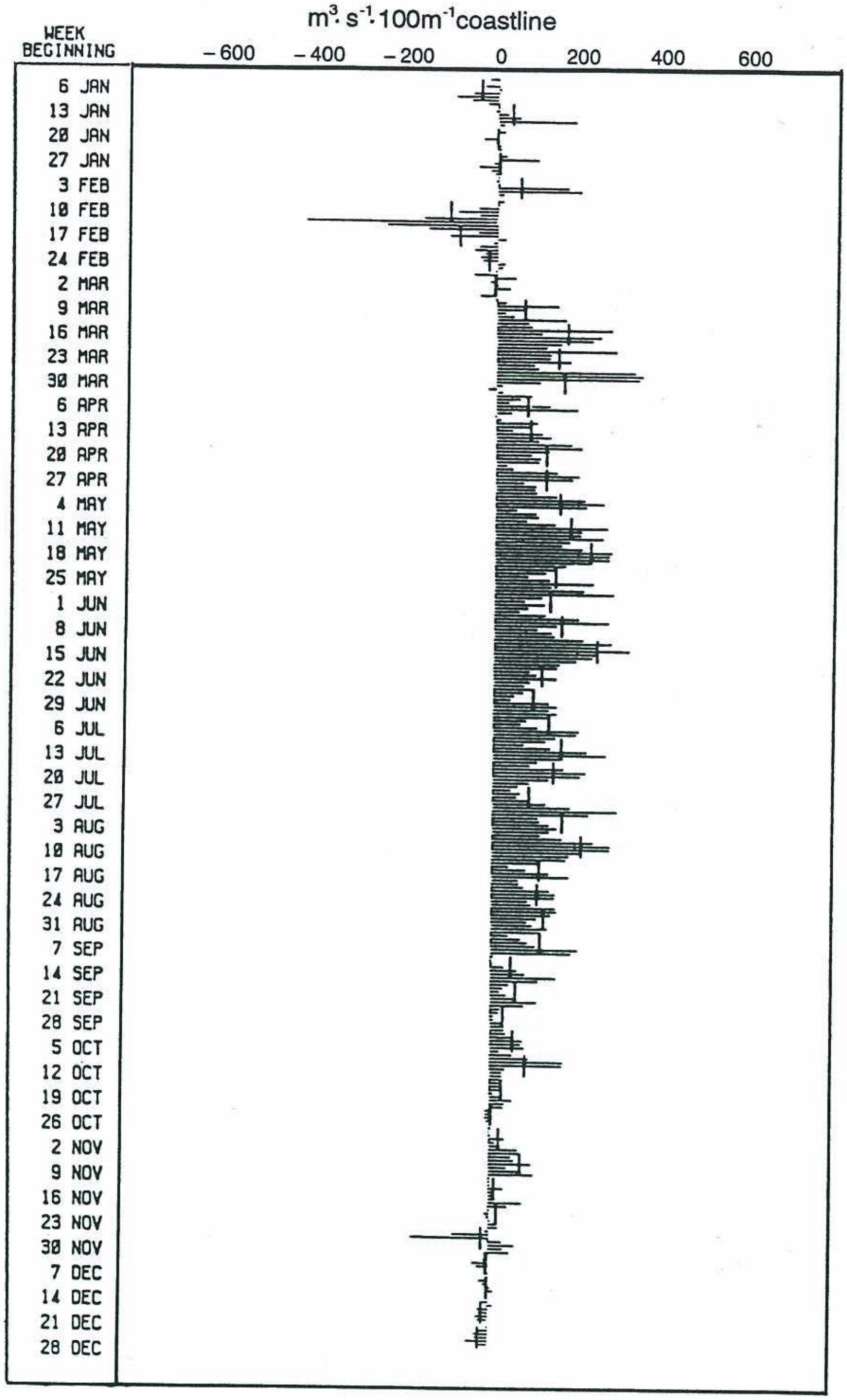

Figure 3 Daily and weekly mean coastal upwelling indices for 1980 at 36°N, 122°W. Positive values indicate upwelling and negative values, downwelling. (Adapted from Mason & Bakun 1986).

such as the Chesapeake Bay (Blumberg 1977, Wang & Elliott 1978, Scheffner et al. 1981) and San Francisco Bay (Conomos 1979, Cheng & Gartner 1985, Walters et al. 1985).

Over the past 50 years or so, studies related to the circulation of MB have been primarily motivated by two factors. During the 1930s and 1940s, the circulation of MB was studied because of interest in several local fisheries, particularly the sardine fishery. Since the 1960s, circulation studies have been motivated by concern regarding the disposal of municipal sewage. More recently (since the late 1980s), there has been considerable interest in designating MB as a National Marine Sanctuary to protect the area from the dangers associated with offshore drilling for gas and oil off the central California coast. In 1988, Congress designated MB as a National Marine Sanctuary. Final approval was granted in September 1992 and MB is now the 11th and the largest National Marine Sanctuary to be authorized. Also, interest in the marine biology of MB has grown considerably since the opening of the Monterey Bay Aquarium in 1984 and the creation of the Monterey Bay Aquarium Research Institute in 1986.

A variety of information on various aspects of circulation in MB exists, and most of it has been acquired since the pioneering studies of Bigelow & Leslie (1930) and Skogsberg (1936). The quality of the circulation data which have been acquired in and around MB over the past 40 years or so varies significantly. Much of the data on MB circulation prior to circa 1970 have come from drifters and drogues where the effects of windage were often neglected. Also, sampling problems arise with respect to drifter observations since only those which are recovered can be used to estimate the flow. The dynamic method has frequently been used to infer the circulation in the MB area. However, because of the high amplitude internal waves which occur in MSC, estimates of the local circulation based on calculations of dynamic height may be aliased.

Because of these problems, we have given careful attention to the selection of the data that are presented in this study. In keeping with the above, many of the drifter and drogue observations that have been reported and, in some cases, currents inferred from dynamic heights, have not been included because they were considered to be unrepresentative of the actual conditions which prevailed. The quality of the deep current measurements acquired in MSC was also considered. Of particular concern was the brevity of a number of the current meter records which were acquired in this region. In some cases, the records were too short to resolve tidal effects and as a result we have given preference (but not exclusively) to data which were of sufficient duration to clearly resolve tidal effects. Finally, in considering the quality of the circulation data acquired offshore beyond the Bay, in many cases only drifter (including drift bottles) and hydrographic data were available. In this case we have given preference to the hydrographic data (i.e. dynamic topographies) since they are based primarily on stations far removed from MSC.

The primary purpose of this study is to bring together and synthesize the existing information on MB circulation and to create a view of that circulation which is generally consistent with the available data.

The time scales of primary interest range from roughly the semidiurnal tidal period to interannual. The space scales range from a few km to the maximum dimension of the Bay (~40km), and to several hundred km, outside the Bay.

The subsequent discussion is presented in six parts. First, major results obtained prior to circa 1950 are presented under historical views on the circulation of MB. Next is a section on direct and indirect observations of bay circulation based on more recent studies. A section on time, space and dynamical scales follows. Then a section on the processes that affect the circulation follows next. The results of this study are synthesized in the following section, after which our conclusions are presented. Finally, a comprehensive bibliography is included.

Historical views on the circulation of Monterey Bay

Bigelow & Leslie (1930) reported the results of the first major oceanographic study of MB. Based on hydrographic data acquired during the summer of 1928, the vertical and horizontal circulations were inferred. Sea surface temperatures (SSTs) were elevated in the sheltered bight areas and generally cooler temperatures were observed over the Canyon. The internal temperature and salinity fields over the Canyon were influenced by upwelling to depths of about 250 m. In particular, "updrafts" of cold, saline water tended to follow the slopes of the Canyon in the direction of its head. Subsurface waters, elevated through upwelling, tended to rise within the Canyon and then spread out over the shelf areas to the north and south.

Bigelow & Leslie attempted to infer the surface circulation by calculating dynamic heights (0/500 db). While various uncertainties in making such calculations were acknowledged, their results depicted an anticyclonic gyral circulation pattern centred over MSC.

Because of the small range of temperatures they encountered over the Bay (~5.0°C), Bigelow & Leslie concluded that bay waters exhibited "great regional uniformity". They apparently associated this spatial uniformity with relatively slow temporal changes, since they created property maps for the Bay from measurements acquired over three to four weeks. They clearly did not recognize the rapidity with which bay waters could change (as was later shown by Skogsberg 1936).

An extensive oceanographic study of MB was conducted by Skogsberg (1936) between 1929 and 1933. Skogsberg presented a thorough analysis of numerous temperature observations acquired at several locations in southern MB. Perhaps the most often quoted result of Skogsberg's study was his description of the annual temperature cycle. He divided the annual temperature cycle into three (not necessarily contiguous) hydrographic seasons, the "cold water phase" extending from mid-February through November, a "warm water phase" extending from mid-August to mid-October and a "low thermal gradient phase" extending from December to mid-February.

The "cold water phase" or "upwelling period" was caused by upwelling, which Skogsberg attributed to a combination of wind-driven coastal upwelling and "an onshore pressure of unknown origin". Near MSC, the effects of upwelling could be detected as deep as 200 m in certain years and to depths of 500 m or greater in other years. Upwelling occurred along the edges of the Canyon, and the upwelled water ultimately made its way up and onto the shelves in the northern and southern portions of the Bay. Finally, he noted extraordinary interannual variability in the intensity of upwelling in MB and that this process did not start at the same time each year.

The "warm water phase" or "oceanic period" was attributed to the onshore movement of oceanic waters associated with the California Current and corresponded to the period of weaker and variable winds. According to Skogsberg, the warm water phase always had its maximum in September.

The "low thermal gradient phase" or "Davidson Current period" coincided with the local occurrence of the northward flowing Davidson Current. Winds from the south during this period produced surface convergence near the coast and thus lower thermal gradients in the upper 50 to 100 m.

Skogsberg was keenly aware of the tremendous variability that characterized his observations, and he continually emphasized the importance of the rôle that variability played in the interpretation of his results. For example, he observed that sometimes coastal waters moved through the Bay in a northward direction and sometimes in a southward direction. Also, he found distinctly different water masses entering the Bay at irregular intervals and concluded that much of the variability encountered within the Bay had its origins outside the Bay.

Skogsberg was also very aware of the rapidity with which ocean (thermal) conditions could change within the Bay. On one occasion waters in the southern bight were exchanged completely within less than a week (see section on "Time, space and dynamical scales"). In other cases, he found that the waters in this part of the Bay were renewed almost daily.

Local eddies frequently occurred in the northern and southern bights of MB. Strong currents flowed into and out of the Bay past Point Pinos. Skogsberg also concluded that the tidal influence in southern MB was not significant, based on drift patterns from a limited number of surface drifter trajectories.

Skogsberg's study had several shortcomings. First, he attempted to characterize the hydrography and circulation of MB using only one physical property (temperature). Secondly, his data were acquired mainly at a few selected locations in the southern half of the Bay; consequently, he could not construct maps depicting the distribution of temperature over the Bay. Thirdly, he made very few direct current measurements; hence, most of his observations on the circulation of MB were, by necessity, inferred from his temperature distributions.

In a sequel, Skogsberg & Phelps (1946) presented additional results on the hydrography of MB for the years 1934 to 1937. This work essentially confirmed the earlier results on the thermal environment of MB. The only major difference they found for the latter period was that there was no distinct rhythm in the amplitudes of the thermal variations near the surface. In fact, the variations in surface temperature were so pronounced and irregular that the notion of a "normal" annual temperature cycle for this region was untenable. Conversely, temperatures at 50 and 100 m were more stable from year to year, and the temporal variations were similar to those obtained earlier (i.e. 1929 to 1933). The periods associated with each of Skogsberg's (1936) original three "seasons" were somewhat different. In this later study, the upwelling season extended from mid-February to late August (versus mid-February through November), the oceanic period, from late August to mid-November (versus late August to late October), and the Davidson Current period, from mid-November to mid-February (versus December to early February). These differences underscored the uncertainties in defining the different oceanic regimes for MB.

Skogsberg's seasonal description for MB has been questioned on several occasions (e.g. Barham 1956). The existence of the upwelling and non-upwelling (Davidson Current) periods is generally accepted. More in question is the existence of a separate oceanic period in the fall, when warm offshore waters intrude into the Bay. Such intrusions might be expected following the relaxation of upwelling-favourable winds. This description should be amenable to water mass analysis, if, in fact, different water masses were involved. However, no such analysis has been undertaken. In recent studies, satellite imagery suggests that intrusions of warmer water from offshore during the oceanic period might be related to the breakdown of an eddy located beyond the entrance to MB (e.g. Fig. 22). Satellite data also suggest that such intrusions probably occur on an irregular basis and are subject to considerable interannual variability. This could explain why Barham (1956) was not able to reproduce Skogsberg's three-season description from hydrographic data acquired over MB in 1954 and 1955. Skogsberg's three-season model for MB has been extended to other portions of the California coast (e.g. Griggs 1974, Pirie et al. 1975, Pirie & Stellar 1977), where its applicability may be even more questionable.

Prior to 1950, the only deep current measurements made in MB were acquired in MSC by Shepard et al. (1939). Observations acquired over several hours near the bottom in a water depth of 91 m revealed maximum speeds of about 27 cm/sec, considerably higher than expected.

Recent observations on the circulation of Monterey Bay

In this section, observations of the circulation in MB acquired within the past 25 years or so are considered. First, we consider the surface circulation which includes direct current measurements, inferred currents and satellite observations. By "surface", we refer to the surface mixed layer which ranges in depth from about 20 m in summer to about 50 m in winter (Husby & Nelson 1982). In this study, we take 25–30 m as representative for the depth of the mixed layer. Then the circulation at intermediate depths (~25 to ~150 m) is considered and, finally, we consider observations of the deep circulation in MB that are mainly restricted to MSC and efforts to model the circulation of MB.

Direct and inferred observations of the surface circulation

Based on property distributions, Smethie (1973) and Broenkow & Smethie (1978) examined the surface circulation in MB. Distributions of temperature, nitrate and salinity showed the penetration of cool, high-nitrate waters into the Bay from the south around Point Pinos about one-third of the time. Based on property distributions acquired over a 27-month period, surface flow to the north occurred about 50% of the time, whereas southward flow was indicated less than 10% of the time. Northward flow inside the Bay during the winter was not unexpected due to the poleward-flowing Davidson Current offshore. However, prevailing northward flow inside the Bay during spring and summer was not anticipated because of the prevailing northwesterly winds.

Coincident drogue trajectories and contours of surface density also provide a consistent picture of cyclonic circulation at the surface in MB (Fig. 4; Moomy 1973). The drogues were deployed at 12 m and tracked over a two-day period from 30 to 31 August 1972. The hydrographic data from which the densities were calculated were acquired between 28 and 30 August 1972.

Direct current measurements have been acquired at several locations in MB using current meters moored in relatively shallow water. In each case the moorings were located between 1 and 3 km offshore.

Current meter data were acquired off the mouth of the Salinas River [Fig. 1 (ESI)] at depths of 9 and 25 m between January 1976 and January 1977 (Engineering Science, Inc. 1978). The daily mean flow at this location was northward (Fig. 5). Northward flow occurred 65% of the time and southward flow, 35% of the time. Variations in flow were episodic, with intermittent bursts of flow to the south, occasionally lasting several days. Daily mean speeds were generally greater to the north, reaching almost 20 cm/sec in a few cases. Currents at 25 m were similar in speed and direction to currents at 9 m, with the

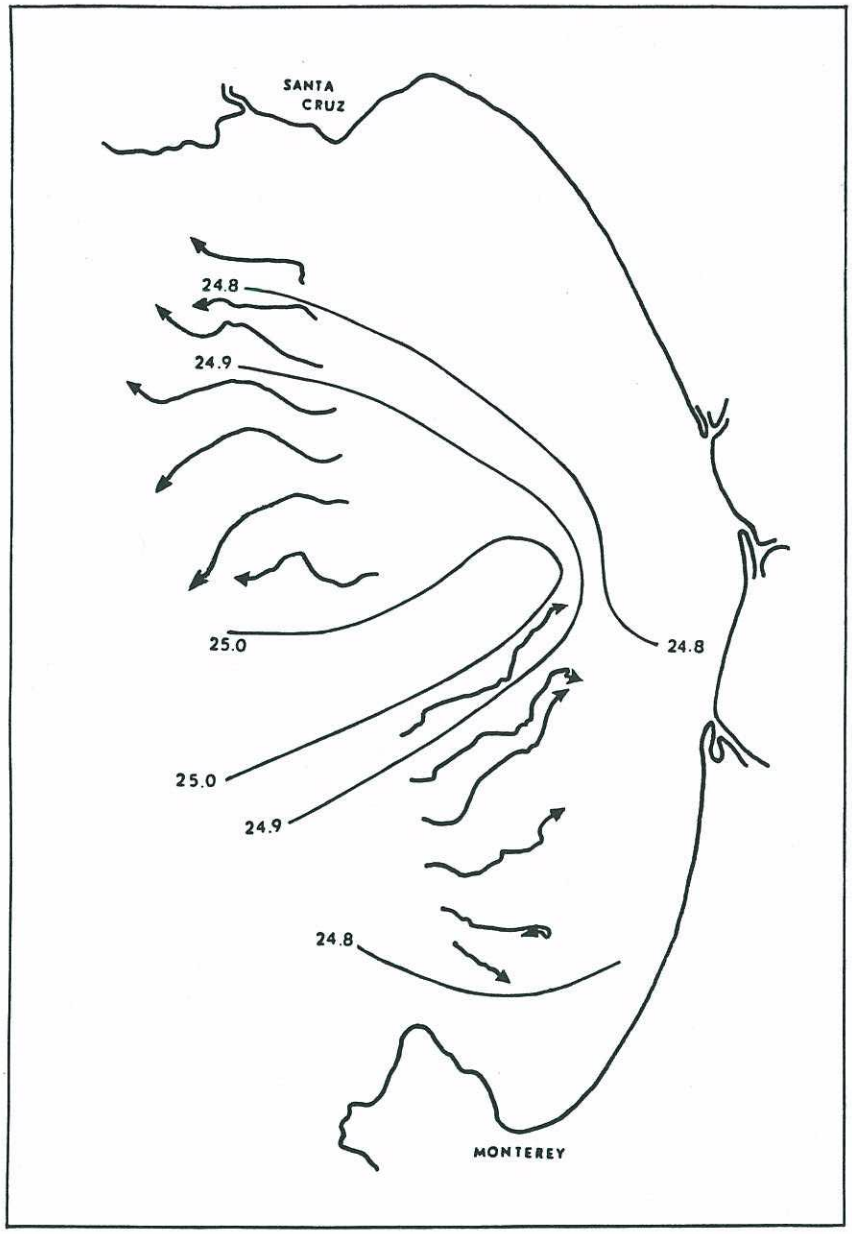

Figure 4 Drogue trajectories at 12 m for 30 to 31 August 1972 (uncorrected for wind drag) and contours of surface density based on hydrographic data acquired between 28 and 30 August 1972. (From Moomy 1973).

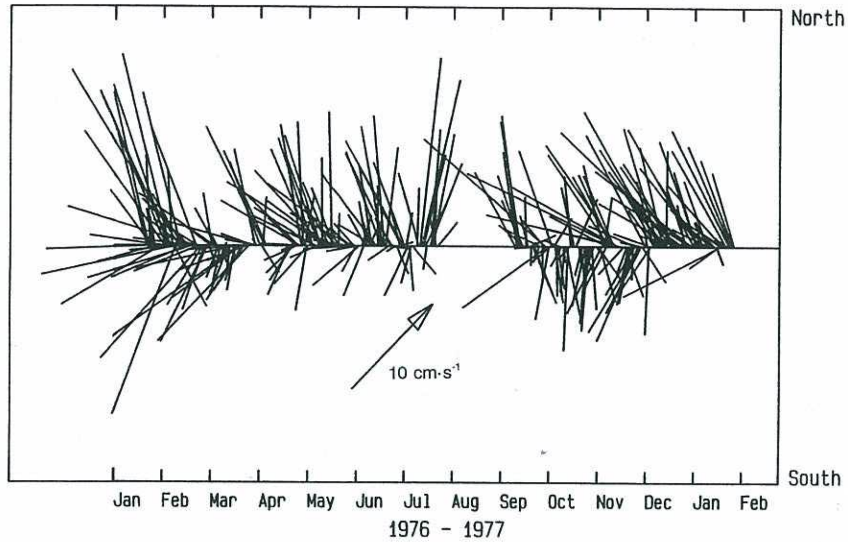

Figure 5 Stick diagram for daily-averaged currents at 9 m (ESI) near the mouth of the Salinas River. The top of the page is north.

deeper currents following the underlying bathymetry more closely than the shallower currents.

Between December 1979 and December 1980, current meter data were obtained near the Pajaro river mouth at two locations (Stations ECO 1 and ECO 2, Fig. 1; ECOMAR, Inc. 1981). One meter was deployed at 6 m in a water depth of 12 m (ECO 1), and two meters (at 6 and 20 m) were deployed slightly further offshore in a water depth of 24 m (ECO 2). At ECO 1, the monthly mean flow at 6 m was consistently toward the NNW, and mean speeds were between 6 and 13 cm/sec (Fig. 6). The near-surface flow at ECO 2 was primarily to the NW but varied more in direction with a major reversal occurring in March 1980, coincident with a major spring transition in that year. The deep flow at ECO 2 was often to the NW with speeds similar to those at 6 m (7 to 13 cm/sec). At both locations, the monthly mean, near-surface flow tended to be aligned with the local bathymetry.

Another long-term current meter record was acquired at the north end of MB off Terrace Point between May 1976 and January 1977 (Station KLI, Fig. 1; Brown & Caldwell 1979). Current meters were located at 9, 15 and 25 m in a water depth of 30 m. Between June and September 1976, the flow was predominantly westward, consistent with prevailing northward flow within the interior of the Bay. However, between November and January, there was a tendency for the flow off Terrace Point to be eastward, or into the Bay. Current speeds generally ranged between 10 and 25 cm/sec. Flows at 15 and 25 m were generally in the same direction as the flow at 9 m but reduced in speed. Again, the direction of flow at all three depths tended to be aligned with the local bathymetry.

Long-term current meter observations have also been acquired in southern Monterey Bay beginning in 1987 (MBA, Fig. 1). However, due to the sheltered nearshore location of this current meter mooring, the observed currents are not representative of the larger-scale circulation further into the Bay.

Recently, Coastal Ocean Dynamics Applications Radar (CODAR) data were acquired to

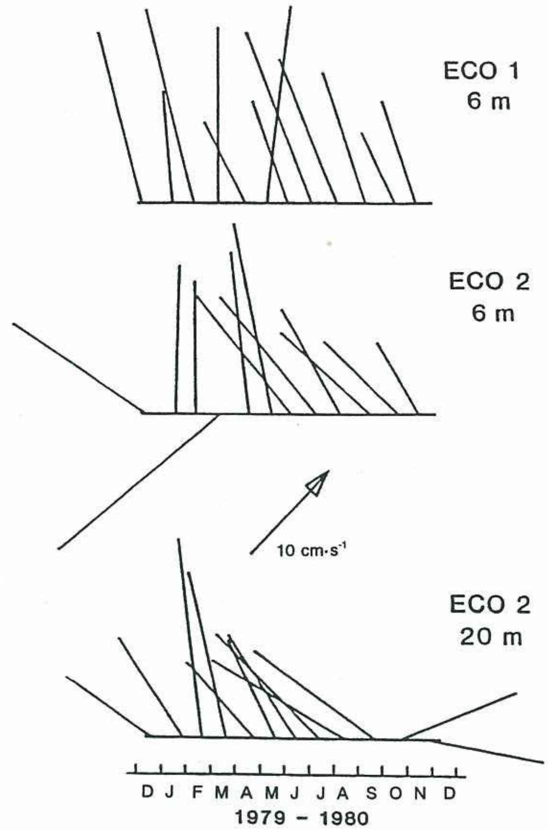

Figure 6 Stick diagrams for monthly-averaged currents at 6m (surface) for ECO 1 (top) and ECO 2 (middle), and for currents at 20m (deep) at ECO 2 for December 1979 to December 1980. (From ECOMAR, Inc. 1981).

estimate the surface circulation in MB (Paduan & Neal 1992). Variability of the surface flow was large and contained a significant diurnal contribution. The mean circulation between March and May 1992 revealed a single, cyclonic gyre with maximum current speeds on the order of 20cm/sec further offshore in the region where southward flow was observed. These observations tend to confirm the previously established cyclonic circulation in MB.

Remote observations of the surface circulation

Both visual and infrared (IR) satellite imagery are often useful in identifying patterns of circulation in coastal areas (e.g. Nihoul 1984). Advanced Very High Resolution Radiometer (AVHRR) infrared imagery from the NOAA polar-orbiting satellites has often been used to observe SST along the central California coast (e.g. Broenkow & Smethie 1978, Broenkow 1982, Breaker & Mooers 1986, Tracy et al. 1990).

An ERTS-1 visual satellite image from 22 January 1973 shows a pattern of suspended sediment originating from the mouth of the Salinas River (Fig. 7). This clearly implies

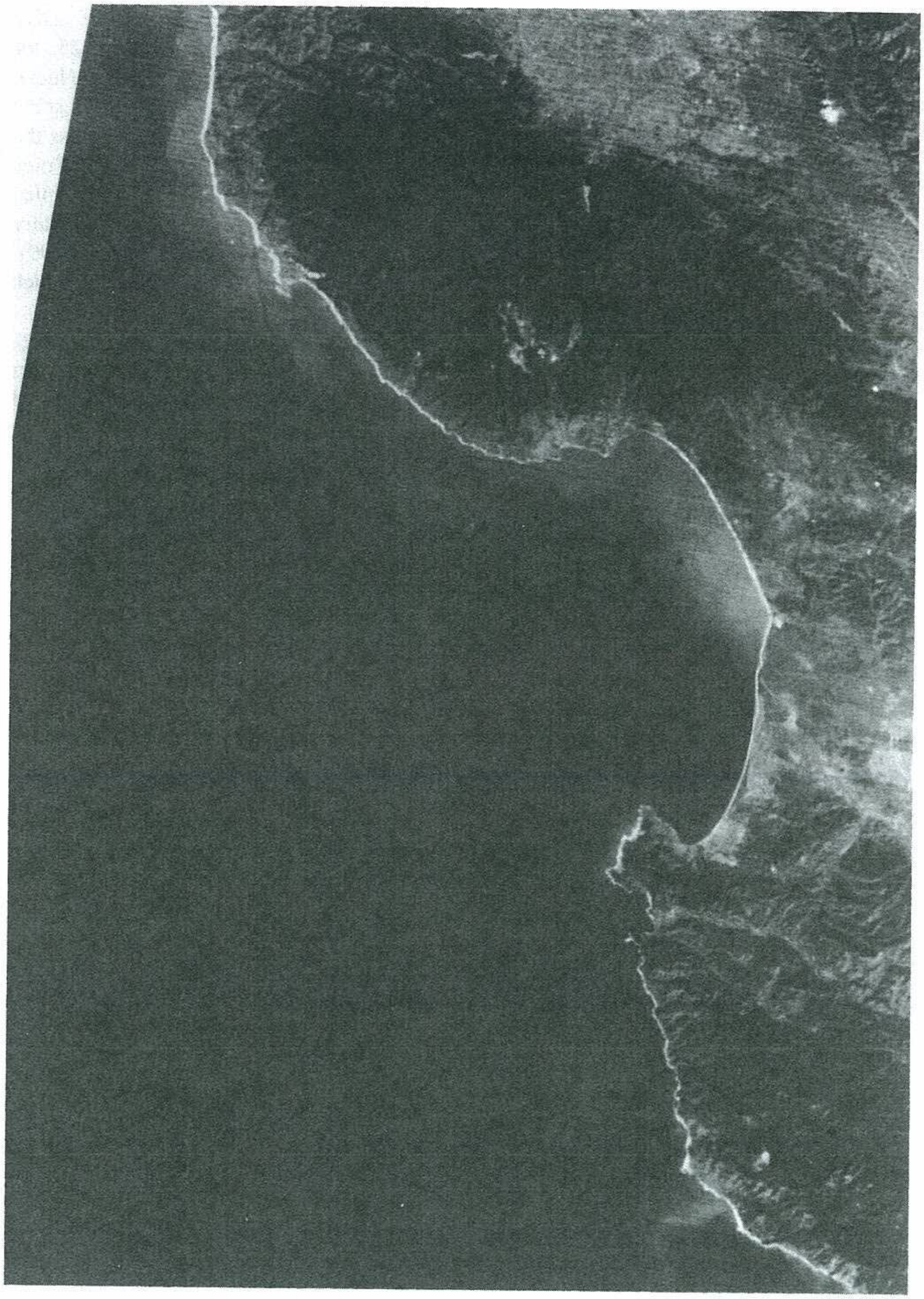

Figure 7 Visual image from Band 4 of the Multi-Spectral Scanner aboard the ERTS-1 satellite showing a northward trend for suspended sediment discharged from the Salinas River, 22 January 1973.

nearshore flow to the north inside the Bay.

AVHRR imagery shows certain common characteristics. First, during periods of active coastal upwelling, a band of cold water often crosses the entrance of MB (see Fig. 25, for example). These upwelled waters often originate in the upwelling centres off Ano Nuevo (a small cape 17 km NW of Davenport) north of MB, and Point Sur, south of MB. In some cases, it appears that cold water which originates north of MB is advected south across the entrance of the Bay, in agreement with the interpretation of recent satellite imagery by Tracy et al. (1990). Tracy also indicated the importance of Ano Nuevo as a source of upwelled water that is often advected south across the entrance of, or into, MB. This upwelled water occurs in a shallow surface layer somewhat less than 50 m in depth (Tracy et al. 1990). Overall, the satellite imagery is consistent with the view that cold upwelled water which frequently occurs inside MB often originates outside the Bay.

Another feature common to most of the satellite images is generally warmer temperatures inside MB, suggesting (a) that coastal upwelling *per se* does not occur or that upwelled waters do not reach the surface directly inside the Bay particularly within the first 5 km or so of the coast, or (b) that local heating may be important.

Observations of flow in and around MB are presented using the method of feature-tracking applied to AVHRR satellite imagery (e.g. Vastano & Bernstein 1984, Breaker et al. 1986). The displacements of submesoscale thermal features identified in the patterns of SST are estimated from sequential, co-registered satellite images separated in time by a day or less. This technique assumes that the apparent feature displacements are due solely to horizontal advective processes.

AVHRR images from 9 and 10 December 1982 were chosen for analysis (Fig. 8), a period when SSTs do not usually vary significantly over MB. In this case, however, a small area of cold water was present off Point Pinos on 9 December. By the following day, this feature appeared to have moved northward. The motion estimated by feature-tracking was northward outside the Bay, and around Point Pinos it was into the Bay; the associated speeds were 10 to 12 cm/sec. These flow directions are consistent with the poleward-flowing Davidson Current which is usually established by mid-November, and the observations of Smethie (1973) and Broenkow & Smethie (1978).

Circulation at intermediate depths

Due to the paucity of observations in MB at intermediate depths (~25 to ~150 m), little is known about the circulation in this region that overlies the Canyon and the deeper shelf areas but underlies the surface layer.

Although direct observations of flow at intermediate depths in MB are generally lacking, the internal temperature field at depths between about 30 and 150 m has been mapped using historical data (Lammers 1971). Lammers computed monthly mean temperatures at 19 locations over the Bay on a uniform grid (7.4 km × 7.4 km) based on data acquired over the 40-year period from 1929 to 1969. Over 25 000 temperature observations were included in this study. Because of the extensive averaging employed to obtain these temperature fields, the confounding effects of high amplitude internal waves in MSC should have been minimized.

Lammers's monthly-mean maps of selected isothermal surfaces (10, 11 and 12°C) for January, March, August and November show that these isothermal surfaces range in depth from 30 to 90 m (Fig. 9). These analyses also indicate a region of warm water located near

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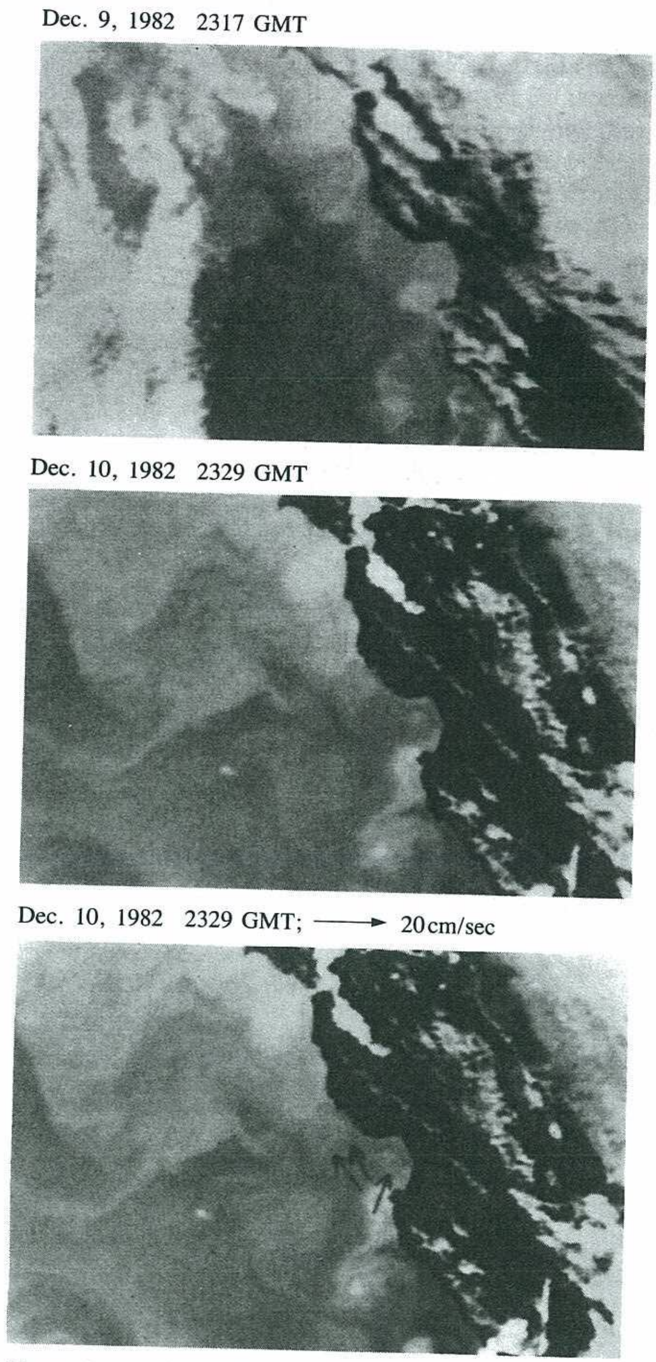

Figure 8 AVHRR satellite images (Channel 4) from 9 (upper) and 10 (middle) December 1982. Lower panel shows results of feature tracking which indicate northward flow offshore at Monterey Bay and flow into the Bay around Point Pinos. The origins of the flow vectors are plotted halfway between the initial and final locations.

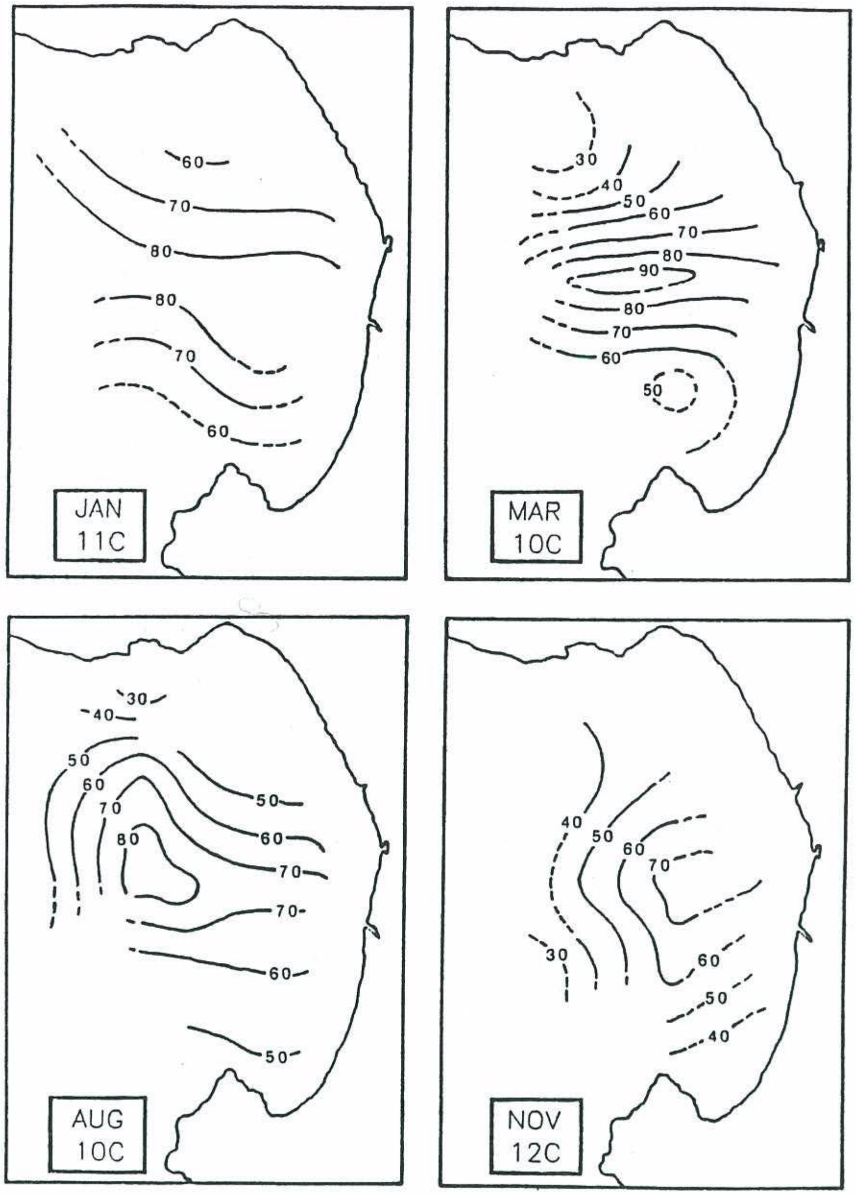

Figure 9 Monthly-mean topographies (m) of selected isotherms for months where the thermal high located over Monterey Submarine Canyon is particularly well developed. (Adapted from Lammers 1971).

the centre of the Bay over MSC. For the months shown, this warm feature is particularly well developed. According to Lammers's results, the "thermal high" over MSC appears to intensify during February and March and then weakens somewhat between April and July. The high strengthens again in August and then gradually weakens during September and October. From November through January, the high is again significantly stronger.

A number of mechanisms might explain the existence of, and variations in, the thermal high described by Lammers. For example, during periods of poleward flow offshore, a separate anticyclonic circulation cell within MB could be generated from waters originating in the California Undercurrent. In this case, we would expect the thermal high not only to be warmer than surrounding waters but also to be of higher salinity. Temperature-Salinity (T-S) analysis of hydrographic data acquired over MSC during 1971 and 1972 (Scott 1973) indicates that three water types coexist in this region – subarctic, transitional and equatorial. Subarctic waters occupy the upper portion of the water column down to 100–150 m, transitional waters occupy the region between about 100–200 m, and equatorial water occurs below 200 m. A seasonal increase in equatorial water was noted between August and December. Although this seasonal increase in equatorial water at deeper levels in MB does not correspond precisely with Lammers's observations on the seasonal behaviour of the thermal high over MSC, these differences could be due to interannual variability associated with Scott's data (since Lammers's data were averaged over many years). Also, variations in the intensity of the thermal high could be related to seasonal variations in the strength of the California Undercurrent (Chelton 1984, Wickham et al. 1987).

Lammers's results indicate that geostrophic flow below the surface layer (i.e. within the thermocline) may be opposite to the flow at the surface. To examine this possibility in greater detail, we have used the maps of Lammers plus other data available for MB as a basis for estimating geostrophic velocities below the surface layer from the thermal wind equation.

First, we convert Lammers's surfaces of constant temperature to horizontal gradients of temperature. According to the relationship between topographic and horizontal scalar fields (e.g. Saucier 1955), the horizontal temperature gradient can be calculated from

$$\frac{\partial T}{\partial x} = -\left(\frac{\partial z}{\partial x}\right)_T \left(\frac{\partial T}{\partial z}\right) \tag{1}$$

where we select representative values for $\partial z / \partial x$, based on the elevation maps of Lammers (Fig. 9), and representative values for $\partial T / \partial z$ (e.g. Skogsberg & Phelps 1946). Then we calculate the corresponding values for $\partial T / \partial x$. Based on Mamayev's simplified equation of state for sea water (Mamayev 1975), where

$$\sigma_T = 28.152 - 0.0735T - 0.00469T^2 + (0.802 - 0.002T)\ (S-35) \tag{2}$$

We calculate the corresponding density gradient [where $\sigma_T = (\rho - 1) \times 10^3$] by differentiating Mamayev's equation of state with respect to x, using the previous values of $\partial T / \partial x$ and selecting representative values for temperature (11°C), salinity (33.75 ppt) and the salinity gradient (1.7×10^{-2} ppt/km). The corresponding vertical shears were then calculated from the thermal wind equation, where

$$\frac{\partial v}{\partial z} = -\frac{g}{\bar{\rho} f}\frac{\partial \rho}{\partial x}\ , \tag{3}$$

following the notation of Pond & Pickard (1983). f was calculated for 36°N, $\bar{\rho}$ was taken equal to 1.0 g/cm^3 and g equals 980.0 cm/sec^2.

By considering only the case where the vertical shear acts to reduce the flow in the surface layer (i.e. where velocity decreases with increasing depth), we can estimate the depth at which the geostrophic flow reverses direction, if we know the depth and velocity at the bottom of the surface layer. Based on the above calculations, the depth of geostrophic current reversal is plotted as a function of the vertical shear for an assumed layer depth of 30m, for various velocities at the bottom of the surface layer (Fig. 10). For a vertical shear of 4.0×10^{-1} cm/sec/m and a velocity at the bottom of the surface layer of 15cm/sec, the depth of geostrophic current reversal is approximately 68m. The vertical shear, of course, can also be used to estimate the velocities at deeper levels below the depth of flow reversal by simple extrapolation.

Although the thermal high depicted by Lammers over MSC implies the possibility of anticyclonic circulation at depth, inflow data acquired across MB in May 1988 indicates cyclonic rather than anticyclonic circulation at intermediate depths in MB (Koehler 1990). The stations shown in Figure 11a were occupied on five occasions between 8 and 11 May and the resulting data averaged to obtain mean fields, a procedure that significantly reduced aliasing by the semidiurnal internal tide. Both geostrophic and Acoustic Doppler Current Profiler (ADCP) velocities indicated anticyclonic circulation in the upper 50m or so, and cyclonic circulation between ~75m and ~400m. A maximum inflow of about 16cm/sec was observed at the surface in northern MB with a maximum outflow of about 18cm/sec in the southern half of the Bay (Fig. 11b). Maximum inflow at depth (~8cm/sec) occurred at approximately 150m and maximum outflow (~8cm/sec) at approximately 250m. This deep circulation was primarily restricted to MSC. The circulation in MB inferred from the observations of Koehler followed a period of strong upwelling off the central California

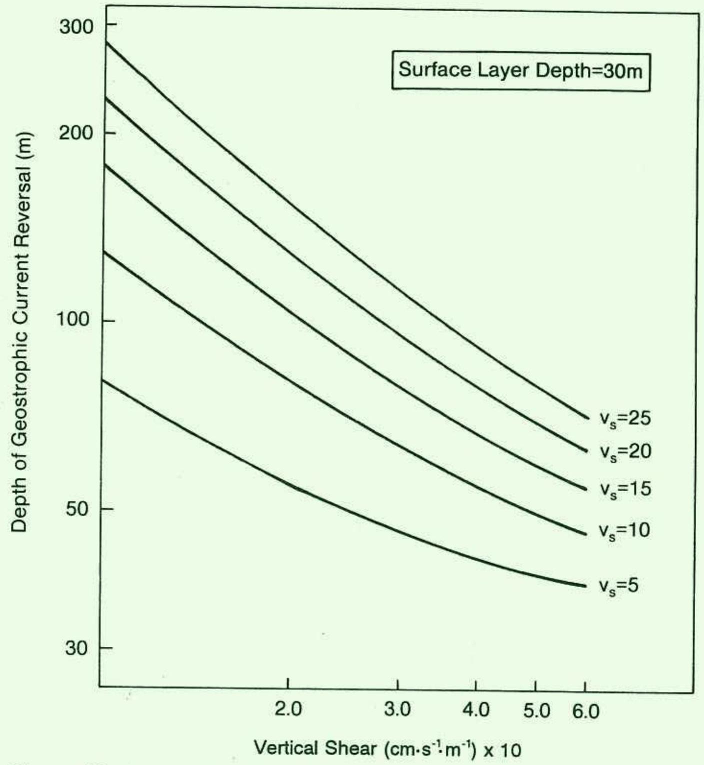

Figure 10 Depth of geostrophic current reversal versus vertical shear for various current velocities at the bottom of the surface layer over MSC, based on the internal temperature fields of Lammars. A surface layer depth of 30m has been assumed.

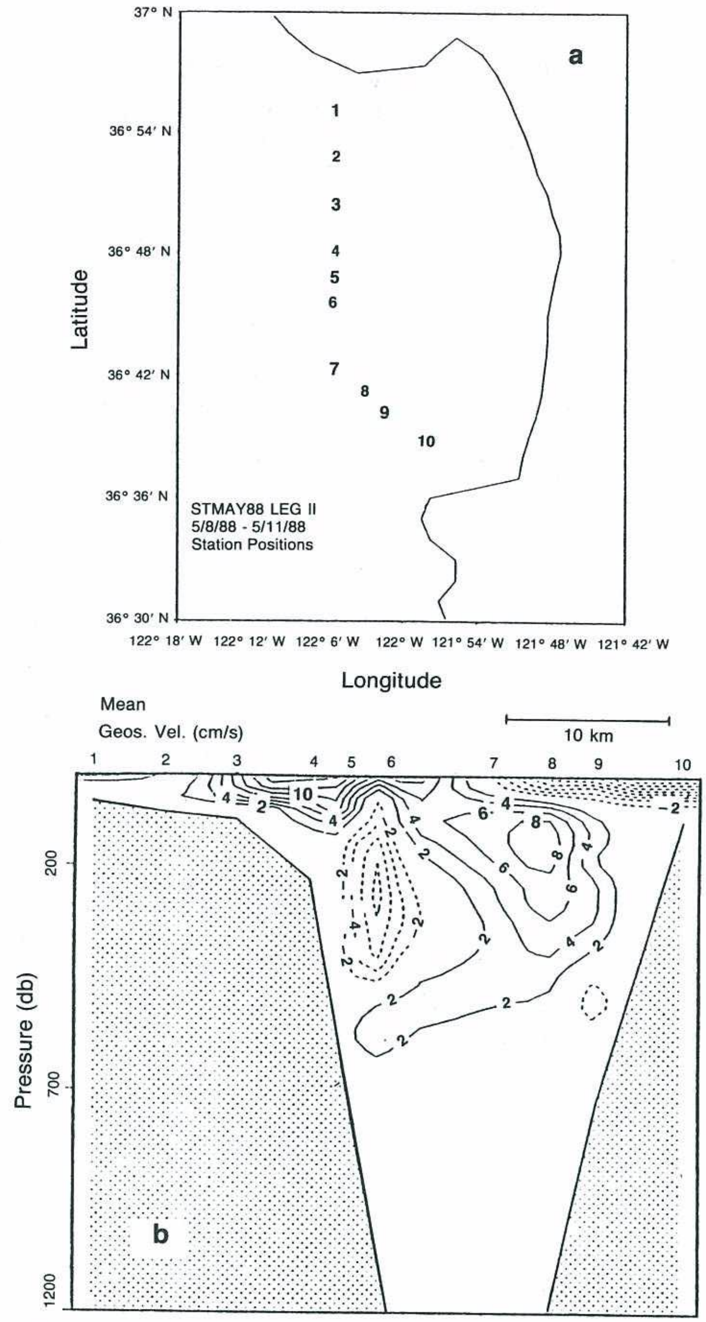

Figure 11 Sequence of ten hydrographic stations acquired across MB between 8 and 11 May 1988, at the locations indicated (a). To obtain the geostrophic velocities shown in the lower panel (b), the temperature and salinity data acquired at each location were averaged to reduce the effects of aliasing by the internal tide. The level of no motion (assumed) was chosen to correspond to the deepest common depth between stations. Solid contours depict flow into the Bay and dashed contours, flow out of the Bay. (Adapted from Koehler 1990).

coast. Periods of locally strong coastal upwelling may contribute to the previously described flow reversals that occasionally take place in the surface layers of MB. The advection of upwelled waters from the north may simply dominate surface circulation in the Bay for brief periods. The deep circulation coincided with weak or nonexistent flow in the California Undercurrent based on hydrographic data acquired off Point Sur just prior to Koehler's observations in MB (Koehler 1990). Weak flow in the Undercurrent is expected during the spring in this area (Chelton 1984) and coincides with a period when the thermal high in MB observed by Lammers is not well developed. Thus, the cyclonic circulation observed at depth in MB during May 1988 may represent a second mode of flow that occurs when the flow offshore in the California Undercurrent is very weak.

Skogsberg also recognized that the circulation at depth in MB could be different from that at the surface. According to Skogsberg (1936), "surface waters on the one hand and those of the lower levels on the other, presented nearly diametrically different thermal trends, strongly suggesting that the movements of these waters were more or less independent of each other."

The deep circulation in Monterey Bay

In this section, observations of the circulation in MSC are considered. Direct measurements of the currents in MSC were made in a series of studies conducted by the Naval Postgraduate School (Gatje & Pizinger 1965, Dooley 1968, Njus 1968, Caster 1969, Hollister 1975). These measurements were made near the head of MSC in water depths not greater than 485 m for periods of usually less than a week and sometimes as short as a few hours.

Current meter elevations ranged from about 5 to 60 m above the bottom with a tendency for current speeds to decrease with increasing elevation. Generally, currents near the canyon head tended to follow the canyon axis (Gatje & Pizinger 1965, Dooley 1968, Njus 1968). However, farther offshore in deeper water, Caster (1969) and Hollister (1975) found cases where the flow was cross-canyon as well. At 485 m, for example, Hollister found strong cross-canyon flow at 60 m above the bottom.

In each of these studies, the relation between the observed currents and the barotropic tide was considered. Spectrum analysis of the current speeds usually indicated the presence of several significant peaks in addition to the expected peak at the semidiurnal period. For example, both Dooley and Njus found spectral peaks at 6, 4 and 2 to 3 hours. These higher-frequency components (i.e. supertidal) were generally thought to be related to the semi-diurnal (internal) tide. Dooley also observed periodic bursts (i.e. sudden increases) in current speed upon occasion in the upcanyon direction. These bursts, together with coincident temperature data, implied the movement of colder water up the Canyon.

Direct current measurements have also been made in MSC by Shepard and his co-workers, the results of which are summarized in Shepard et al. (1979). Current measurements were acquired at six locations in MSC for bottom depths ranging from 155 to 1445 m. The measurements were obtained at elevations of 3 and 30 m above the bottom for periods of 3 to 4 days during March 1973, November 1974 and November 1977. Progressive vector diagrams at eight locations within the Canyon indicated the highly irregular nature of the flows encountered. At one location in a water depth of 1445 m, the flow was upcanyon at both 3 and 30 m above the bottom. However, at a depth of 1061 m, the net flow reversed direction from upcanyon at 3 m to primarily downcanyon at 30 m above the bottom. At a depth of 155 m, the net flow at 3 m was upcanyon, while at 30 m it was primarily downcanyon. In five out of eight cases, the net flow was upcanyon. Shepard et al. considered the results for MSC to be unique, because the net flows they observed in most other submarine canyons tended to be downcanyon.

According to Shepard et al. (1979), not only were the net flows upcanyon, but also the upcanyon velocities were stronger than those associated with downcanyon flow at the two deepest stations (1061 and 1445 m). In addition, they found that the currents generally showed little tendency to follow the axis of the Canyon out to the deepest observed depths (1445 m). They indicated that this lack of directional control is apparently a characteristic of MSC. In most canyons, Shepard et al. found that the period between reversals in current direction generally increased, approaching the tidal period as bottom depth increased. In MSC, reversals in current direction did not correspond with tidal periods, even at the greatest depths. The average cycle times for upcanyon and downcanyon reversals ranged from about six to nine hours. Shepard et al. have also interpreted their results in terms of propagating internal waves. For MSC, upcanyon propagation between depths of 384 and 155 m was found, while downcanyon propagation occurred between depths of 1445 and 1061 m.

The results of the Naval Postgraduate School and those of Shepard et al. are consistent, although Shepard's data were generally acquired in deeper water. A strong tendency for flow to follow the canyon axis was observed only in relatively shallow water near the head of MSC. Generally, both sets of studies found frequent cross-canyon flow farther out into the Canyon, and both observed supertidal frequency oscillations in current speed and direction.

Recent beam transmissometer profiles (29 July 1986) acquired near the mouth of MSC show several discrete turbidity layers throughout the water column that suggest the importance of the near-bottom flows (Fig. 12). The turbidity layers above 500 m may be caused by baroclinic tidal currents (described later), but the turbid waters extending from the bottom to about 500 m above the bottom are probably the result of the vigorous currents which have been observed in this region. For near-bottom current speeds of 25 cm/sec or more, bottom scouring may provide the mechanism by which these silty sediments are resuspended (Dyer 1986). Scouring along the slopes above the bottom may also contribute sediment to this turbidity layer. Although benthic boundary layers along continental margins are typically less than 100 m thick, nepheloid layer thicknesses as great as 500 m have been observed in Willapa Submarine Canyon (Baker 1976).

Recent observations using a remote operating vehicle (ROV) near the bottom of MSC indicate highly turbid waters and very strong currents just west of MSC head (36°47′N, 121°50′W) in a water depth of 250 m (C. Harrold, personal communication). During the period of observation (15 September 1988), highly turbid waters extended to 15 m above the bottom. The bottom sediments in this area were highly scoured, and from floating debris, current speeds of at least 100 cm/sec were estimated. Finally, in retrieving the ROV along the north wall of the Canyon, strong upcanyon flows were encountered.

During a series of dives aboard the deep submersible research vehicle ALVIN during October and November 1988, Eittreim et al. (1989) found evidence for vigorous bottom currents in the steeper portions of MSC from depths of about 800 to 1200 m. In this depth range, they discovered frequent sediment clouds, rippled bottoms, winnowing and current scouring, and occasional debris from submarine landslides, all indications of vigorous bottom circulation. In several cases, bottom currents were estimated to be in the range of 25 to 60 cm/sec, although based on the observations from the ROV, these estimates appear to be somewhat conservative. The currents were, more often than not, upcanyon. Further offshore in the fan valley region, they found indications of much weaker bottom flows.

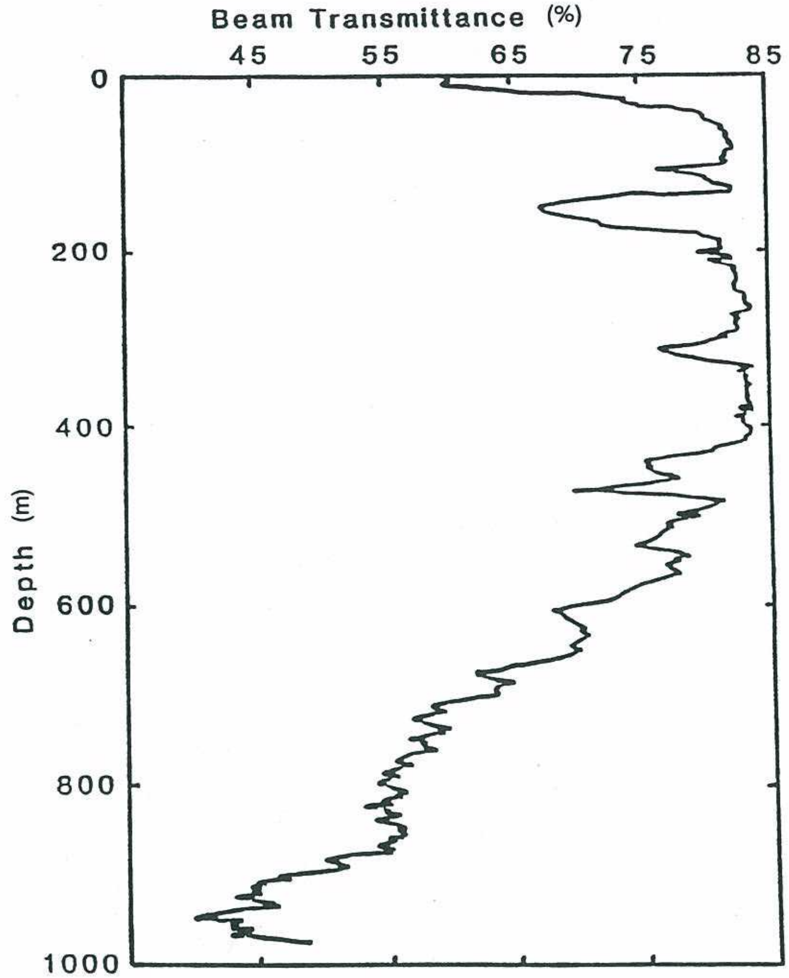

Figure 12 Vertical profile of percent beam transmittance (percent per m at 490 nm) near the mouth of Monterey Submarine Canyon (36°47′N, 122°02′W) for a bottom depth of 960 m, acquired on 29 July 1986.

Circulation modelling

The first attempt to simulate MB circulation was a time-dependent, finite-difference, hydrodynamic model for homogeneous flow (Garcia 1971). Circulation within the Bay was driven by flow outside the Bay through momentum transfer (i.e. by the lateral shearing stresses). Initially, the Bay geometry was treated as a simple cavity, and solutions were obtained for various Reynolds numbers and length-to-width ratios for the cavity. A more refined model included the effects of bottom topography and bottom friction. The volume transport stream function indicated an extensive anticyclonic gyre within the Bay driven by poleward flow outside the Bay (Fig. 13). Flow inside the Bay, particularly in the shallower regions, tended to follow the bottom topography. Other combinations of Reynolds number and bottom friction yielded similar results. Garcia noted that the effects of bottom friction diminished

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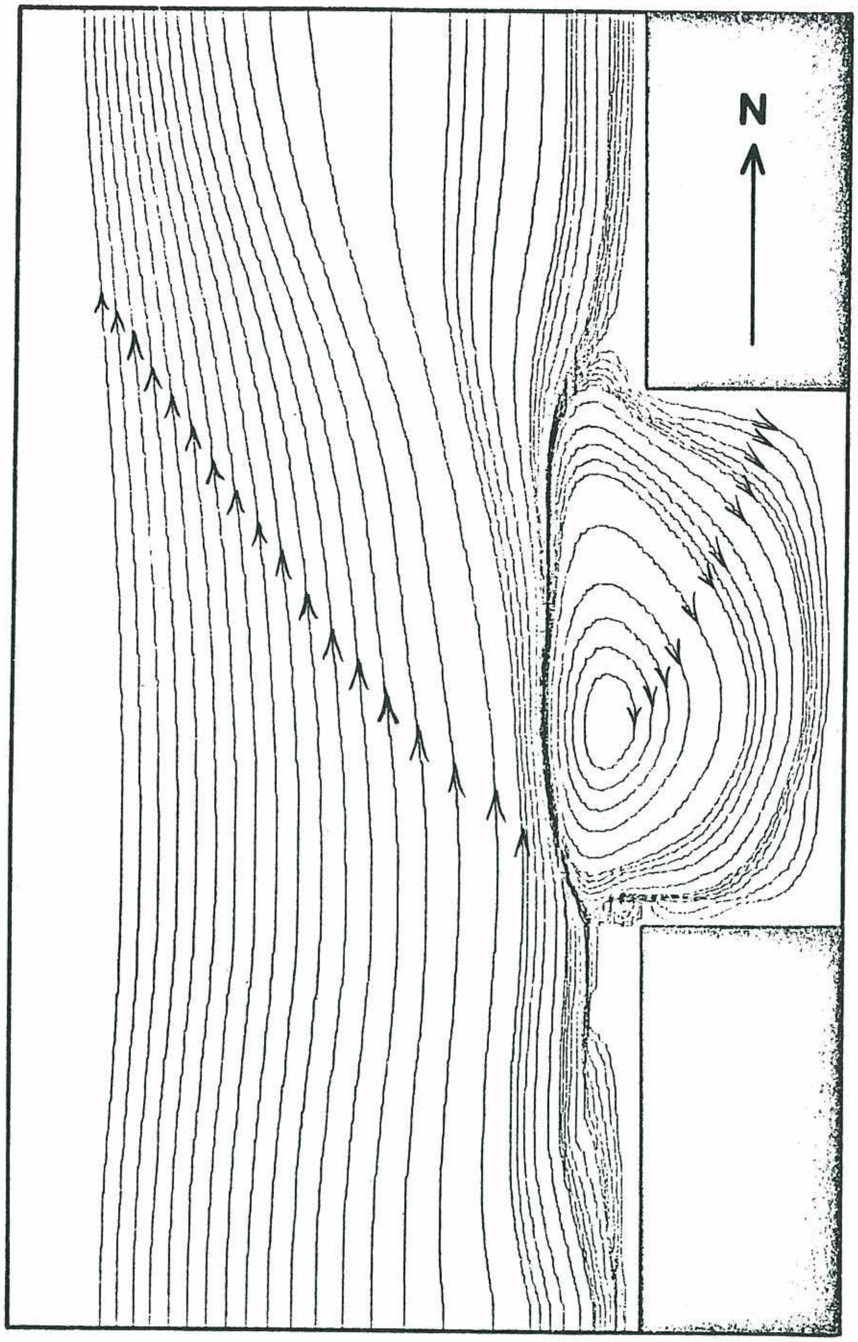

Figure 13 Results from a numerical circulation model for Monterey Bay, including realistic bottom topography but neglecting density stratification. The volume transport stream function shows northerly flow outside the Bay (i.e. toward the top of the page) and an anticyclonic gyre inside the Bay. (Adapted from Garcia 1971).

rapidly with increasing depth, but in the shallow portions of the Bay, bottom friction significantly affected the circulation.

Garcia's one-layer model is particularly significant in that it established a framework for comparison with future data. For example, the anticyclonic circulation pattern predicted by Garcia's model applied to intermediate depths in MB, provides a basis for explaining the origin of the thermal high observed by Lammers.

More recently, Bruner (1988) employed a two-layer, primitive equation, numerical model to study the circulation of MB. The upper layer thickness was 50m. The model was run for periods of 11 to 15 days to achieve steady state conditions. The effects of winds, tides and bottom friction were not included.

Due to constraints on model geometry, alongshore flow could not be specified as a boundary condition for model forcing. Alternatively, zonal flow across the open boundary (i.e. across the mouth of the Bay) was specified. Thus, model calculations were performed with various combinations of zonal forcing. In particular, the effects of topography, the distribution of zonal forcing, and variations in vertical (velocity) shear between the layers were examined. The results of two cases in particular, are considered, the first with inflow in both layers at the north end of the Bay (with a vertical shear of 5:1), and the second with inflows in both layers at the south end of the Bay (with a vertical shear of 5:2). In the upper layer, northward flow inside the Bay occurred for both inflow conditions, whereas in the lower layer, anticyclonic circulation occurred for inflow at the north end versus cyclonic circulation for inflow at the south end. Deep inflow at the north end of the Bay is consistent with poleward flow offshore and thus the offshore forcing employed by Garcia.

Bruner concluded that circulation in MB was greatly influenced by MSC, which led to the presence of a large vorticity gradient. Conservation of potential vorticity coupled circulation in the upper and lower layers. However, when flow in the lower layer was weak (1cm/sec), upper layer flow became decoupled from flow in the lower layer. For stronger lower layer flow (5cm/sec), the influence of bottom topography was exhibited in both the upper and lower layers.

A third modelling study of MB was recently undertaken by Koehler (1990) using the model employed by Bruner. In this case, however, wind forcing and bottom friction were included. The major difference in this study was that the boundary conditions along the western boundary were specified using the geostrophic velocities obtained between 8 and 11 May 1988 (see section on "Circulation at intermediate depths"). The upper layer thickness was maintained at 50m (as in Bruner's case). The results indicated that the flow patterns observed during Koehler's study could be reasonably well reproduced by the model. However, the results were sensitive to whether or not the mass transport across the western boundary was balanced. Better agreement was found when the mass transport was balanced. Also, for the case of strong wind forcing (winds of up to 8m/sec from the NW – i.e. upwelling-favourable), the observed surface flow patterns could be approximately reproduced. Weaker winds had little effect on the surface flow, however. Finally, bottom friction was found to suppress the flow out of MSC at deeper levels.

Garcia's and Bruner's models were both limited by the inability to specify the actual inflow into MB. Koehler, on the other hand, used observed inflow data along the western boundary and achieved reasonable success in reproducing the flow fields obtained during his study. In each case, these models were not able to isolate clearly the effects of density stratification; thus, more critical analyses are needed in this area.

Time, space and dynamical scales

Residence times for bay waters are considered initially by first reviewing past work in this area and then by examining the time required for water to circulate through the Bay from its point of entrance to a point along the inner coast.

In this section, we also consider the internal Rossby radius of deformation because it provides a width scale for geostrophic flow in the Bay, and because it may determine whether or not a deep inflow can occur in MSC.

Finally, we consider the important dynamical scales for the MB coastal region by conducting a scale analysis of the governing equations of motion and continuity.

Time scales

Skogsberg (1936) regularly occupied a line of hydrographic stations from the Hopkins Marine Station (Fig. 1) to a point just south of the Salinas River mouth. A sequence of temperature sections acquired by Skogsberg over a five-day period from 26 November to 1 December 1930 showed a rapid replacement of water. During the first two days, warmer water intruded over the southern portion of the section. By the end of the 5-day period, the original cooler water in this region was almost completely replaced by the warmer water. According to Skogsberg, this rapid change in hydrographic conditions was associated with northward, cyclonic circulation in the southern bight.

Broenkow & Smethie (1978) investigated residence times for surface waters (i.e. in the upper 30m) using salinity, temperature, nitrate and phosphate as tracers. The ratios between volume integrals for each tracer and stream discharge, net insolation, and biological productivity rates were used to estimate residence times. From the analysis of observations made at monthly or bimonthly intervals for 27 months from 1971 to 1973, residence times varied between 2 and 12 days, with a mean of around 6 days (which is approximately the time required for a water parcel to travel 50km at an average speed of 10cm/sec).

The change from winter to spring conditions along the US West Coast is frequently abrupt, resulting in the so-called "spring transition" to coastal upwelling (Huyer et al. 1979, Strub et al. 1987, Lentz 1987); however, an abrupt transition does not occur in all years. Along the central California coast, the spring transition typically occurs in March and lasts for about a week (Breaker & Mooers 1986). During this transition, SSTs decrease by 3 to 4°C and currents may temporarily reverse direction. This dramatic change in oceanic conditions is usually associated with (or driven by) the onset of upwelling-favourable winds which are associated with the seasonal northward shift of the Subtropical High Pressure Cell.

In years when the spring transition occurs, the sudden decrease in SST usually provides a distinct time reference against which the further development of coastal upwelling can be compared. The February/March 1977 and March 1980 spring transitions in SST at Pacific Grove and Granite Canyon (Figs 14 and 1) were particularly noteworthy. In each case, a clear lag can be observed between the onset of these transitions at Granite Canyon and at Pacific Grove. In 1977, the lag was about 10 days, and in 1980 it was about 12 days. There was usually no discernible lag between the occurrences of these transitions at Granite Canyon and at the Farallon Islands (both located along the open coast), which are separated by almost 200km. Thus, this time difference between Granite Canyon and Pacific Grove indicates the time required for the disturbance to travel between the coastal ocean and southern MB. Since this disturbance took 10 to 12 days to reach Pacific Grove, it may have

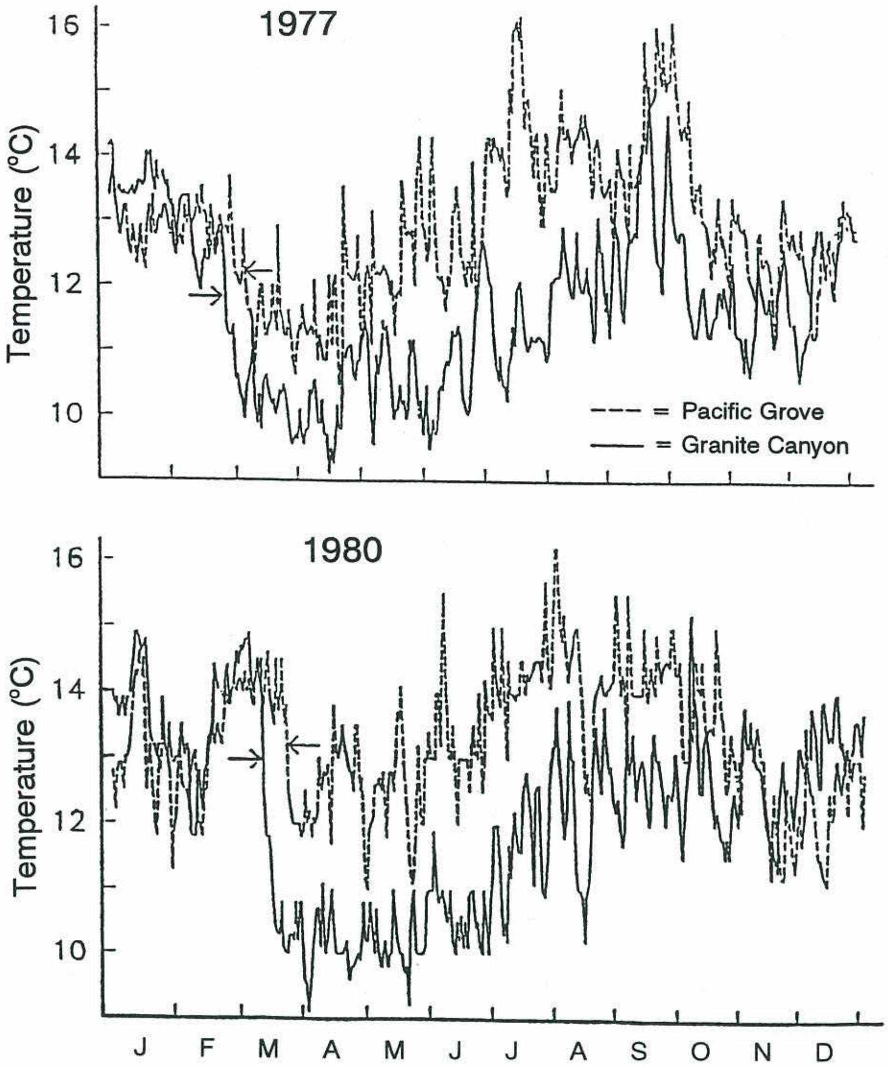

Figure 14 Daily sea-surface temperatures at Pacific Grove and Granite Canyon for the years 1977 (upper panel) and 1980 (lower panel). Sharp decreases in temperature during February and March 1977 and during March 1980 (indicated by arrows) correspond to the spring transitions to coastal upwelling in those years.

been advected into the Bay in the form of leakage from the main event outside the Bay. We note that even values of 10 to 12 days still fall within the range of residence times obtained by Broenkow & Smethie.

Fifteen years of daily SST (March 1971 to March 1986) at Granite Canyon and Pacific Grove were also compared using cross-correlation techniques. The results indicated that Granite Canyon led Pacific Grove by about five days. A similar cross-correlation analysis between Granite Canyon and the Farallon Islands indicated a lag of one day or less. Thus, another measure of the time required for "oceanic signals" to transit one portion of the Bay is obtained. Unlike the previous comparison between Granite Canyon and Pacific Grove, which considered only specific events, this comparison takes into account the entire series. Again, a value of five days falls within the residence time window indicated by Broenkow & Smethie.

The Rossby radius of deformation

In this section, the length scales relevant to bay circulation are considered. Klinck (1989) examined the process of geostrophic adjustment over submarine canyons and found that there are four important length scales that determine the distances over which perturbations in and around submarine canyons decay, the width of the coastal current, the width of the canyon itself, and the internal and external Rossby radii of deformation. The shortest of these scales ultimately governs these distances, dependent upon dynamical mode (i.e. internal or external). Because the internal Rossby radius is expected to be much smaller than either the width of the coastal current (of the order of 50 km) or the external Rossby radius (~2000 km; Emery et al. 1984) but of the same order as the width of the Canyon (Fig. 2), it is of interest to consider the internal Rossby radius of deformation, R_{bc}, because it provides a dynamical basis for determining the narrowness of MSC in relation to the types of flow that can occur there (Klinck 1988, Hughes et al. 1990). The R_{bc} also provides a width scale for geostrophic flow. Because this parameter depends on depth as well as stratification, R_{bc} is expected to vary significantly over the Bay.

The internal Rossby radius of deformation may be expressed as

$$R_{bc} = \frac{[N]\ H}{\pi f}\,, \tag{4}$$

where N is the Brunt-Vaisala frequency, f is the Coriolis parameter, and H is the total water depth. The brackets around N indicate a vertical average over H where N^2 is calculated from

$$N^2\,(z) = -\frac{g}{\bar{\rho}}\frac{\partial\rho\,(z)}{\partial z}\,, \tag{5}$$

g is the acceleration of gravity, ρ is the density at depth z, and $\bar{\rho}$ is the mean density averaged over depth. $[N]$ was calculated from density profiles acquired throughout the Bay during January, May and September 1972 (Broenkow & Benz 1973). To obtain the final value of R_{bc} at each location, the individual values for January, May and September were averaged.

The distribution of R_{bc} averaged over the three periods shows the largest values (10 km) over the mouth of MSC and the smallest values (1 km) around the periphery of the Bay (Fig. 15). The marked influence of depth on R_{bc} is obvious. Thus, the width scale for geostrophic flow within the Bay is governed primarily by the greater depths associated with MSC.

Dynamical scales

In this section we conduct a scale analysis of the governing equations of motion and continuity which are appropriate to the MB coastal region. The results of such an analysis allow us to make order-of-magnitude estimates concerning the forces and thus the processes

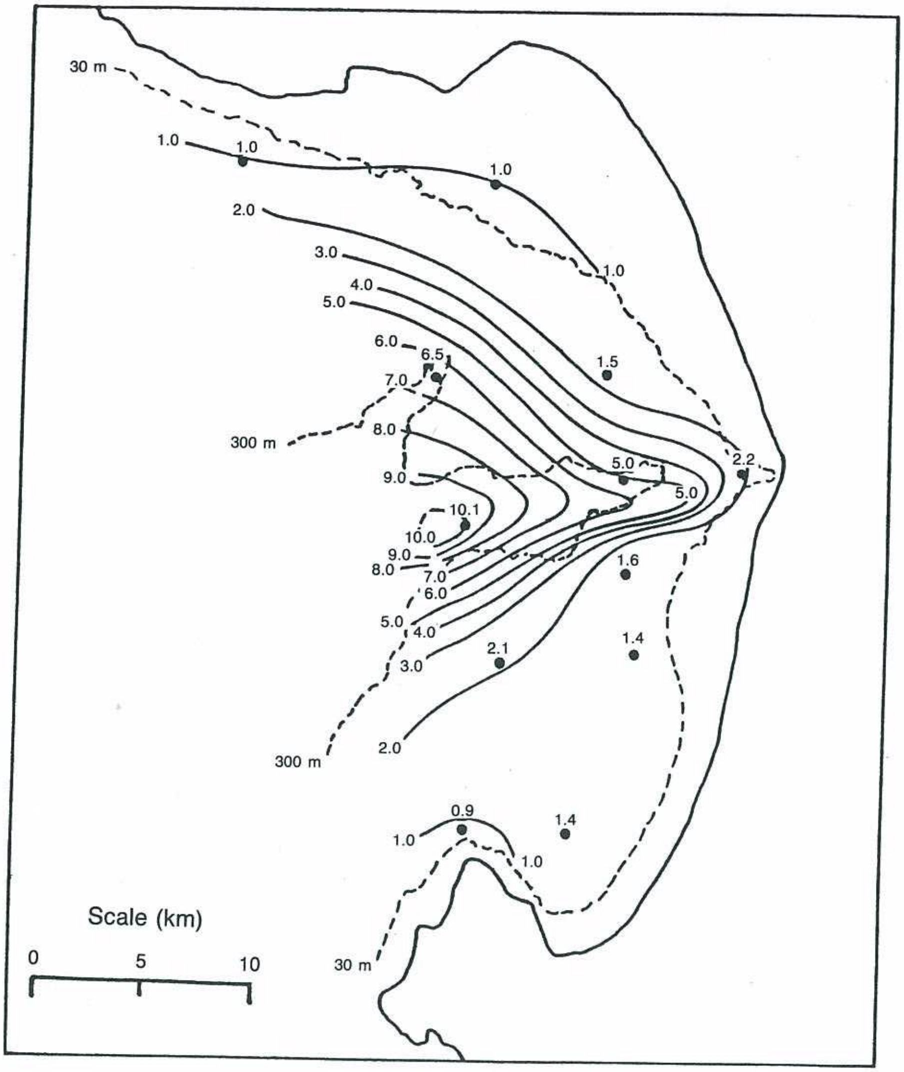

Figure 15 Mean internal Rossby radius of deformation (km) calculated from hydrographic data acquired in January, May and September 1972.

that are likely to be important. Scale analysis in the coastal ocean in some cases may be more precise than it would be in the deep ocean because the important time and space scales are known with slightly greater accuracy. MB is well suited to a dynamical scale analysis because of its relatively simple geometry and its well-defined boundaries. However, in part because these scales are generally smaller (and/or shorter) in coastal regions, more terms in the governing equations are likely to be important, thus complicating the force balances which result.

For incompressible flow, we consider only the following simplified horizontal equations of motion (the hydrostatic balance has been assumed for the vertical equation of motion) and continuity and the appropriate boundary conditions, given as:

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$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}+w\frac{\partial u}{\partial z}=-g\frac{\partial}{\partial x}(\zeta_t-\zeta_c)+fv+K_H\frac{\partial^2 u}{\partial x^2}+K_H\frac{\partial^2 u}{\partial y^2}+K_V\frac{\partial^2 u}{\partial z^2} \tag{6}$$

$$\frac{\partial v}{\partial t}+u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}+w\frac{\partial v}{\partial z}=-g\frac{\partial}{\partial y}(\zeta_t-\zeta_c)-fu+K_H\frac{\partial^2 v}{\partial x^2}+K_H\frac{\partial^2 v}{\partial y^2}+K_V\frac{\partial^2 v}{\partial z^2} \tag{7}$$

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0 \tag{8}$$

$$w=-\left(u\frac{dH}{dx}+v\frac{dH}{dy}\right),\ \text{at } z=-H \tag{9}$$

The notation used here follows that of Csanady (1982) where a conventional (x, y, z) Cartesian co-ordinate system with the z-axis pointing up, has been employed. The pressure is related to the surface elevation ζ, which is divided into barotropic (ζ_t) and baroclinic (ζ_c) components to separate tidal effects from the horizontal density gradient. f is the Coriolis parameter, H, the water depth, g, the acceleration of gravity, and K_H and K_V, the horizontal and vertical kinematic eddy viscosities, respectively.

The last term on the right-hand side of equations (6) and (7) represents frictional effects arising from shearing stresses due to wind at the surface and flow over the bottom plus internal friction. The vertical friction terms can be related to the kinematic wind and bottom stresses as

$$\mathbf{F}=(F_x, F_y)=K_V\left(\frac{\partial u}{\partial z},\frac{\partial v}{\partial z}\right)_{z=0} \tag{10}$$

and

$$\mathbf{B}=(B_x, B_y)=K_V\left(\frac{\partial u}{\partial z},\frac{\partial v}{\partial z}\right)_{z=-H} \tag{11}$$

where **F** and **B** are parameterized as

$$\mathbf{F}=(\rho_a/\rho_w)\ C_s\,|\,\mathbf{W}\,|\,\mathbf{W} \tag{12}$$

and

$$\mathbf{B}=C_b\,|\,\mathbf{U}\,|\,\mathbf{U} \tag{13}$$

where **W** is the surface wind velocity at a height of 10m, ρ_a and ρ_w are the densities of air and water, respectively, C_s is a surface friction co-efficient, **U** is the near-bottom current velocity and C_b is a frictional co-efficient for the bottom.

Because the horizontal scales for x and y are expected to be of the same order, we only consider the x equation for horizontal motion. However, as will be shown using the equation of continuity, under certain conditions the vertical velocity w, is not insignificant and must be taken into account.

Guidance in selecting the appropriate characteristic scales for the field variables has been obtained in part from information presented in several of the following sections. Conversely, the results of this section in turn helped us to decide which of the oceanic processes in and around MB deserved primary consideration. The characteristic scales used in the following scale analysis are given in Table 1.

Of the parameters specified in Table 1, selecting representative values for the kinematic eddy viscosities, K_H and K_V, is probably the most difficult task since their values can only be estimated to within several orders of magnitude and in reality they are not constant in either space or time.

Next, we substitute these scales into equation (6), yielding

$$\begin{array}{cccccccccccccccccccc}
\dfrac{U}{T} & + & \dfrac{U^2}{L} & + & \dfrac{U^2}{L} & + & \dfrac{U^2}{L} & = & g\Bigg(& \dfrac{\zeta_1}{L_1} & + & \dfrac{\zeta_2}{L_2} & \Bigg) & + & fU & + & \dfrac{\rho_a C_s V^2}{\rho_w H} & - & \dfrac{C_b U^2}{H} & + \; K_V \dfrac{U}{H^2} \\
\downarrow & & \downarrow & & \downarrow & & \downarrow & & & \downarrow & & \downarrow & & & \downarrow & & \downarrow & & \downarrow & \downarrow \\
1\times10^{-4} & & 1\times10^{-4} & & 1\times10^{-4} & & 1\times10^{-4} & & & 0.5\times10^{-3} & & 1\times10^{-3} & & & 1\times10^{-3} & & 5\times10^{-5} & & 2.5\times10^{-5} & 5\times10^{-6}
\end{array}$$

where $U^2/L = WU/H$ for the nonlinear term in z. These results indicate that to within two orders of magnitude (except for internal friction), all of the terms in the horizontal equations of motion are approximately the same. However, based on these estimates, the barotropic tide and the Coriolis force, followed by the internal density field are the most important, within the limitations of this approach.

Substituting the appropriate scales into Equation (9) yields a value for w of the order of 1 cm/sec which is extremely high but is due to the steep slopes that occur in MSC. For bottom currents which exceed 10 cm/sec, the corresponding vertical velocities will be even higher.

Wind stress curl may be important in producing offshore upwelling (Ekman pumping). Based on an average value for wind stress curl over the first 70 km off the coast of central

Table 1 Characteristic scales used in the scale analysis.

Type (Abbreviation)	Scale Magnitude	Source/Explanation
Length (L_1)	1×10^6 cm	Figure 20 – Rossby radius
Length (L_2)	1.5×10^8 cm	Wavelength for semidiurnal tide for H = 100 m
Depth (H)	1×10^4 cm	Typical water depth
Hor. velocity (U)	1×10^1 cm/s	Typical horizontal velocity
Vert. velocity (W)	1×10^0 cm/s	Value expected in MSC
Time scale (L/U)	1×10^5 s	Advective; an inertial period
Coriolis (f)	1×10^{-4} sec^{-1}	
ζ_1	0.5×10^2 cm	Typical change in surface elevation over MB
ζ_2	1.5×10^2 cm	Mean tidal range for MB
K_H	1×10^6 cm$^2 \cdot$s^{-1}	Thompson (1978)
K_V	5×10^1 cm$^2 \cdot$s^{-1}	Thompson (1978)
N	5×10^{-3} rad$\cdot$s^{-1}	Broenkow & Benz (1973); calculated
g	1×10^3 cm$\cdot$s^{-2}	
dH/dx(dH/dy)	1.5×10^{-1}	Maximum mean slopes in MSC
W	5×10^2 cm$\cdot$s^{-1}	Wind speed; Nelson (1977)
C_s	2×10^{-3}	Csanady (1982)
C_b	2.5×10^{-3}	Soulsby (1983)
ρ_w	1×10^0 gm/cm^3	Density of water
ρ_a	1×10^{-3} gm/cm^3	Density of air

California, Breaker & Mooers (1986) estimated a vertical velocity of about 2 m/day during June 1980. Using the time scale given in Table 1, a dimensionally compatible value for the magnitude of the wind stress curl is roughly 1×10^{-7}, a value considerably lower than most of the other terms considered. However, this value may be unrealistically low compared to the wind stress curl adjacent to the coast where frictional effects due to coastal mountains may be important. Also, localized alongshore wind jets which were observed off northern California may contribute to increased wind stress curl within 200 km of the coast (Beardsley et al. 1987, Zemba & Friehe 1987). Thus, the importance of wind stress curl in the region around MB could be considerably greater than that indicated above.

The most important outcome of this scale analysis is perhaps that most of the terms that were evaluated turned out to be roughly similar, within two orders of magnitude. Since the range of uncertainty associated with such an analysis is probably ± one to two orders of magnitude in any case, we conclude that all of the terms in the horizontal equations of motion that were included (and thus the processes they imply) may be important in and around MB. For example, the barotropic tide and the geostrophic balance are expected to be important in the MB area. Very high vertical velocities (of order 1 cm/sec) may also be expected to occur in MSC due to the relatively steep slopes and high horizontal bottom velocities encountered there. These results differ significantly from the deep ocean where only the pressure gradient and the Coriolis terms are usually important.

Processes affecting the circulation in Monterey Bay

In this section we consider a number of the processes that affect the circulation in MB. Our selection has been influenced, in part, by the results of the previous section.

Winds

The daily and monthly-averaged winds acquired from the National Data Buoy Center (NDBC) environmental data buoy located just outside MB (36.8°N, 122.4°W; Fig. 1) for the year 1988 are shown in Figure 16. The monthly-mean winds for the one degree square closest to MB, based on over 100 years of ship reports, have also been included (Nelson 1977). These winds are highly persistent in direction being primarily from the NW throughout the year, except from November to February when winds from the south and SW are encountered which often result from winter storms.

The monthly-mean winds do not reflect the smaller-scale temporal and spatial variability associated with the winds in and around MB. The winds back slightly upon entering the Bay, becoming more westerly (Hayes et al. 1984). This backing is also evident by comparing the monthly-averaged winds from the buoy with Nelson's climatological winds, particularly between April and October. Winds at Moss Landing, for example, are predominantly from the west year-round (Broenkow & Smethie 1978), whereas winds at Davenport are typically from the NW. Within the Bay, near Santa Cruz and elsewhere closer to shore, the sheltering effects of coastal mountains and the orientation of the coastline produce winds which are somewhat weaker than those observed near the centre of the Bay and further offshore. Winds at the NDBC buoy and at Davenport typically approach 8 m/sec during the summer whereas further inside the Bay at Monterey and Fort Ord, for

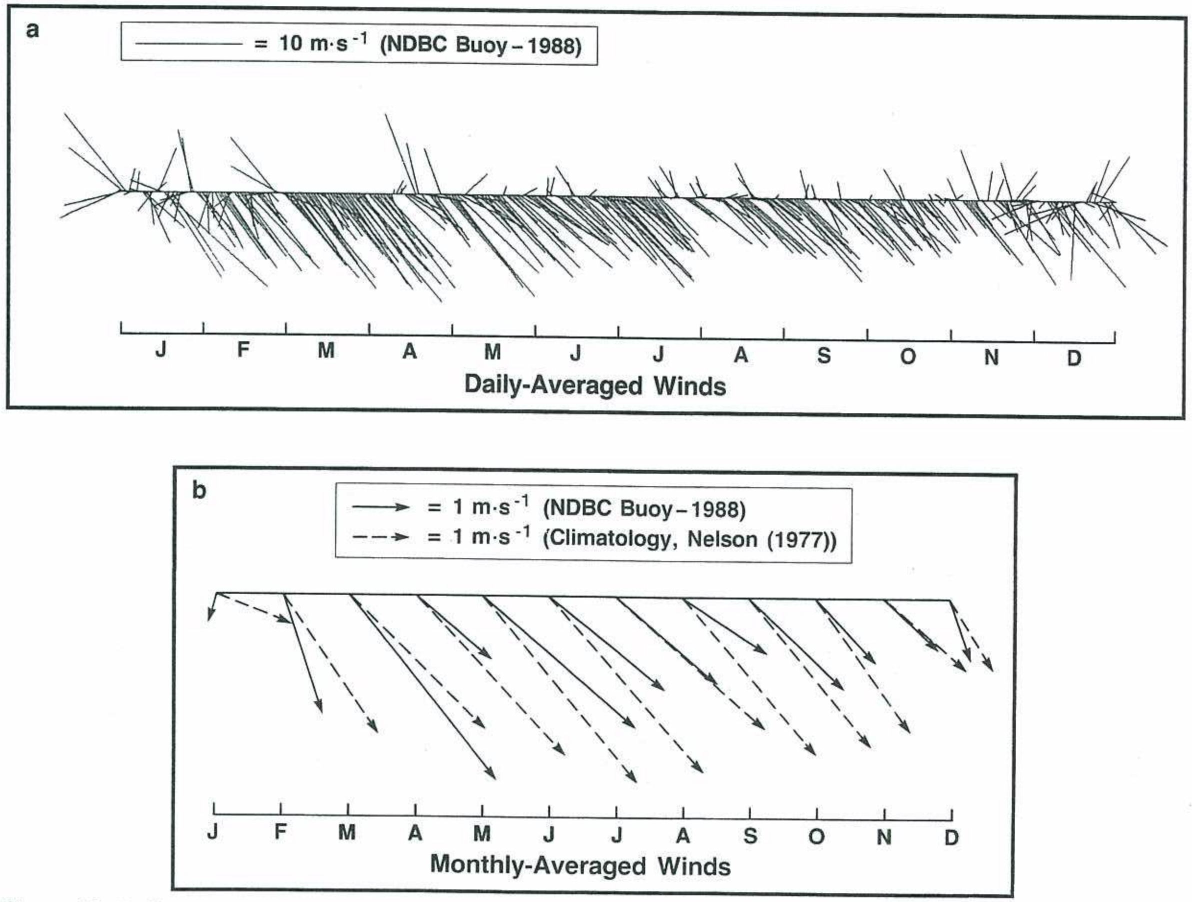


Figure 16 Daily (a) and monthly-averaged winds (b) for 1988 from the NDBC environmental data buoy located at 36.8°N, 122.4°W. Monthly-averaged climatological winds (b) for the one degree square centred at 37°N, 122°W are also included. (Adapted from Nelson 1977).

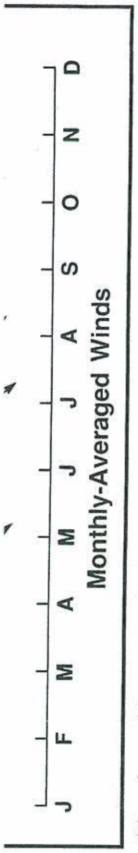


Figure 16 Daily (a) and monthly-averaged winds (b) for 1988 from the NDBC environmental data buoy located at 36.8°N, 122.4°W. Monthly-averaged climatological winds (b) for the one degree square centred at 37°N, 122°W are also included. (Adapted from Nelson 1977).

example, the winds are usually closer to 5 m/sec (Hayes et al. 1984).

A vigorous sea breeze circulation occurs in MB due to the influence of the Salinas Valley. This sea breeze is especially pronounced between May and August, when heating in the Salinas Valley reaches a maximum (Hayes et al. 1984). The sea breeze also affects the surface circulation in MB. Blaskovich (1973) found that the direction along which drift cards moved was affected by the diurnal sea breeze. Reise (1973), in a drift-bottle study in southern MB, found a significant difference in the number of returns between morning and afternoon. More returns from morning releases were attributed to longer exposure to onshore winds associated with the sea breeze.

Upwelling

In the MB region, three types of upwelling may occur: (a) coastal upwelling, which requires an adjacent coastal boundary and is primarily wind-driven, (b) open ocean upwelling or Ekman pumping, caused by positive wind stress curl, and (c) bathymetrically-influenced upwelling, which may occur when the flow is onshore and guided by the presence of a submarine canyon.

Satellite imagery and hydrographic data indicate that coastal upwelling occurs along central California, both north and south of MB, between about March and October (e.g. Breaker & Mooers 1986). That coastal upwelling does not occur to any appreciable extent inside MB was indicated by Broenkow & Smethie (1978) and is consistent with Lammers's (1971) monthly-mean maps of SST (Fig. 17). If significant coastal upwelling *per se* occurred inside the Bay, relatively cool SSTs at least near the coast would be expected during spring and summer. However, between March and October, distinctly warmer temperatures are usually found inside the Bay. AVHRR satellite imagery also indicates generally warmer temperatures inside MB during spring, summer and early fall.

As discussed previously, satellite imagery and hydrographic data indicate that a band of cold water frequently crosses the entrance of MB. This cold-water band often originates north of the Bay in the upwelling center off Ano Nuevo and may simply be advected south; in some cases it may originate south of the Bay and be advected north, into MB. In either case, these waters are most likely the result of coastal upwelling and, as such, clearly influence the circulation in MB.

Offshore upwelling or Ekman pumping may also be important in MB. Breaker & Mooers (1986) estimated the magnitude of offshore upwelling off Point Sur due to positive wind stress curl to be 2 m/day during June 1980, a value comparable to that for coastal upwelling ($\sim$1–10 m/day; e.g. Smith, 1968). Even using a climatological mean value of wind stress curl for this area and period (Nelson 1977), an upwelling velocity of 0.5 m/day is obtained.

Through the process of coastal upwelling, water is transported to the surface from as deep as 300 m (e.g. Smith 1968). Skogsberg (1936) observed the upward displacement of water at the mouth of MSC to depths of 700 m or greater in 1933. The 40-year mean annual temperature cycle at the mouth of MSC from Lammers indicates upward trending isotherms between March and July to depths of at least 500 m (Fig. 18a). The 1951 to 1955 mean annual salinity cycle at the mouth of MSC (Bolin et al. 1964) shows seasonal upward displacements to a depth of 700 m (Fig. 18b). These studies demonstrate that the internal density field at the mouth of MB oscillates vertically to depths of 500 m or more. But, as indicated by Skogsberg, upwelling at depths considerably greater than 300 m cannot be due to local wind-forcing alone. Other factors must be taken into account.

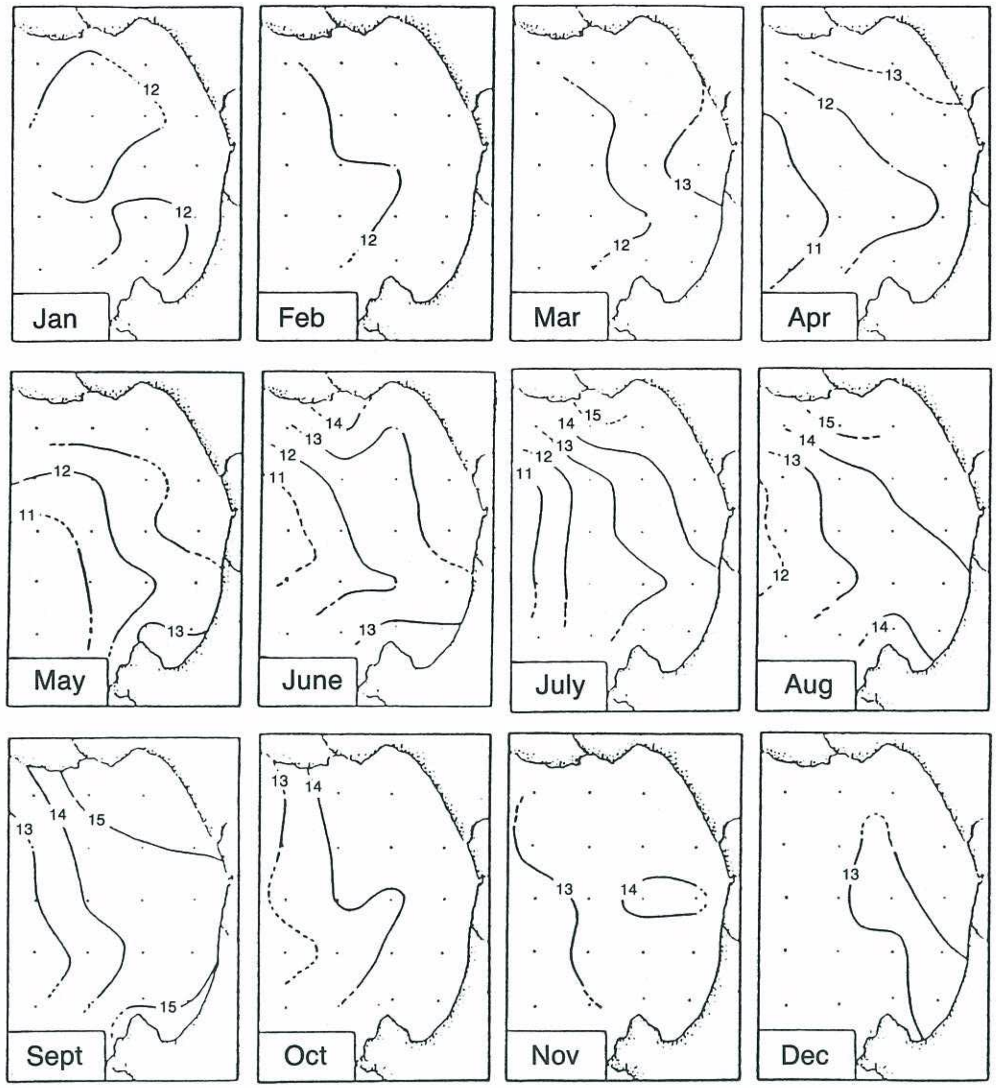
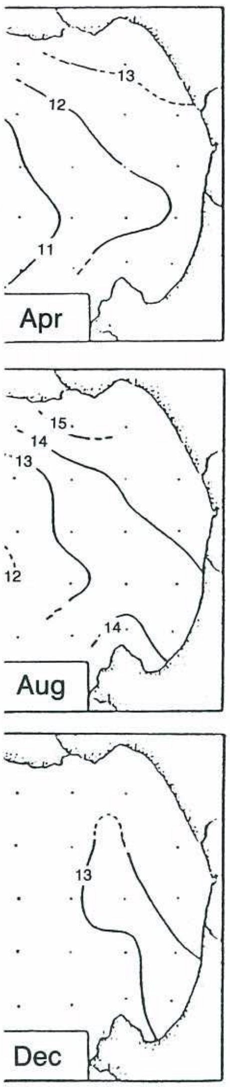


Figure 17 Monthly-mean maps (1929–1968) of sea-surface temperature (C) for Monterey Bay. (Adapted from Lammers 1971).

The influence of submarine canyons on promoting a net upward flow has been investigated in a number of studies (e.g. Peffley & O'Brien 1976, Shaffer 1976, Freeland & Denman 1983, Hickey 1989). Based on model results, submarine canyons are locations where the enhanced upwelling of cold, nutrient-rich waters occur (Peffley & O'Brien 1976). Short, deep canyons act to guide the onshore flow associated with coastal upwelling off NW Africa (Shaffer 1976). Upwelling from great depths off southern Vancouver Island is driven by the interaction of a large-scale coastal current and a narrow canyon on the continental shelf (Freeland & Denman 1983). Deep waters are upwelled in Astoria and Quinalt Canyons but they do not reach the surface directly; rather, they must first cross the continental shelf and be entrained into the upwelling region along the inner shelf (Hickey 1989). The previous



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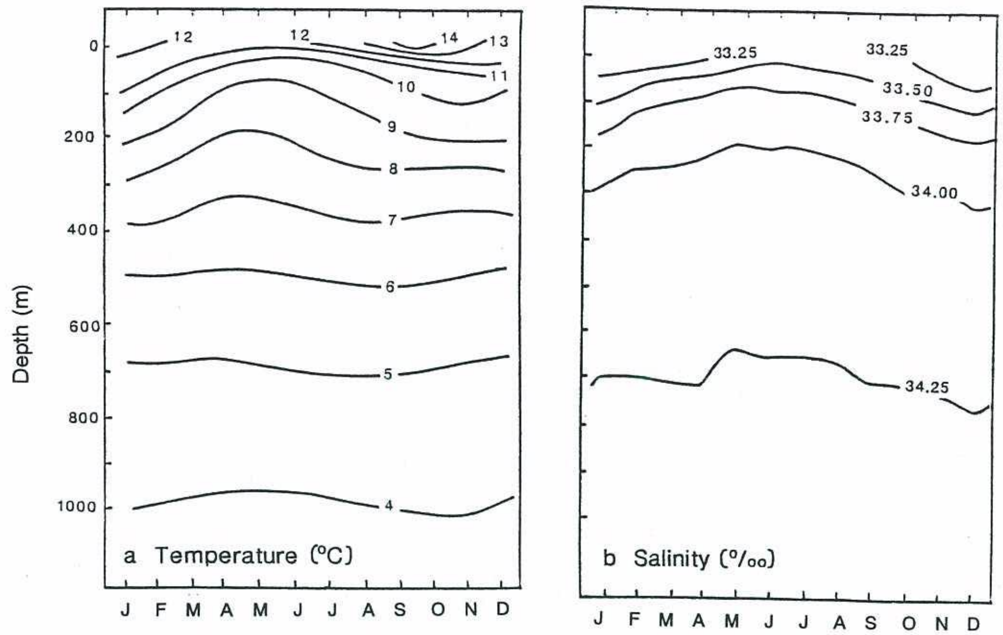

Figure 18 Mean annual cycles of temperature (1929–1968; adapted from Lammers, 1971), left panel (a), and salinity (1951–1955; adapted from Bolin & collaborators 1964), right panel (b).

scale analysis suggested that relatively high vertical velocities should occur in MSC (i.e. bathymetrically-induced upwelling).

A net upward flow near the head of MSC may be produced through tidal pumping (Shea & Broenkow 1982). Through this mechanism, the baroclinic tides raise deep water within the Canyon above the surrounding shelf. As this water spreads out laterally, a residual volume of deep water remains on the shelf during the ebb tide due to mixing, surface heating and inertia effects. Such tidal rectification should be important in the presence of steep continental slopes (Huthnance 1981).

Klinck (1988) studied the influence of canyons on initially geostrophic flow using an analytic model which assumed a homogeneous ocean. Flow within narrow submarine canyons can be forced by the pressure gradient associated with a sloping free surface. By narrow, the canyon width must be small compared to the Rossby radius of deformation (external) according to Klinck, a condition easily satisfied for MSC. This mechanism was originally proposed by Cannon (1972) for driving the deep circulation along the axis of Juan de Fuca Canyon and later expanded upon by Freeland & Denman (1983). Recent support for the importance of this mechanism arises from observations in Lydonia Canyon, located on the southern flank of Georges Bank, which indicate that currents were driven along that canyon by a cross-shelf pressure gradient in geostrophic equilibrium with the alongshelf currents (Noble & Butman 1989). The observations of Shepard et al. (1979) of frequent upcanyon flow in MSC are also consistent with Klinck's barotropic model results. In the case of MB, it is possible that higher surface elevations offshore due to the California Current and prevailing positive wind stress curl, for example, could provide the pressure gradient required to drive the deep inflow in MSC.

In a recent study, Hughes et al. (1990) developed an analytic model to examine the response of a boundary undercurrent such as the California Undercurrent to major variations in bottom topography (i.e. submarine canyons). Their results indicate that for canyons wider than the internal Rossby radius, undercurrent waters enter the canyon and topographic steering occurs. For deep canyons (~1000m, or greater), inertial forces become significant and the assumption of geostrophy fails. In this case, the undercurrent separates from the bottom, creating a wake deep within the canyon. The dynamics of the wake in turn indicate that upwelling occurs along the axis of the canyon with anticyclonic circulation in the upper part of the wake (due to vortex compression) as the flow follows the topography, and cyclonic circulation at deeper levels where vortex stretching occurs. Cyclonic circulation at the deeper levels would lead to cross-canyon flows, consistent with the cross-canyon flows observed in MSC by Caster, Hollister, Shepard et al. and Shea & Broenkow. Observations in Quinalt Canyon by Hickey (1989) are also consistent with the results of Hughes et al.

According to the definitions employed by Hughes et al., MSC is both wide and deep. It is wide since the entrance of the Canyon (transect B–B′ in Fig. 2) is of the same order as the Rossby radius (~10km). However, the width of MSC decreases rapidly inside the Bay, invalidating this requirement closer to the coast. MSC is deep since the scale of the topography (i.e. the maximum canyon depth is almost 900m along transect B–B′, in Fig. 2) is less than the Rossby number where the length scale is taken to be the Rossby radius.

From the above, we expect that waters from the California Undercurrent will enter MB at intermediate depths and follow the topography within the interior of the Canyon until decreasing canyon width precludes its further incursion. At deeper levels (>300–400m), flow separation may occur, producing cyclonic circulation and upwelling in this region. The observations presented earlier are generally consistent with these expectations.

It is important to recognize that the upwelling in MSC originates at deeper levels and thus can not be primarily induced by the local winds and as a result may occur more-or-less continuously over periods of several months or more. The deep, steady upwelling that occurs in MB may have a significant impact on conditions in MB; conversely, upwelling along the open coast which is primarily wind-driven, is subject to wind reversals and event-scale fluctuations and thus will contribute intermittently to upwelling conditions in MB.

The importance of upwelling in MSC and its contribution to the surface waters of MB, although frequently referred to, has not been firmly established. Earlier studies (in the 1950s and 1960s) generally concluded that MSC was the primary source of upwelling in MB and that deep, upwelled waters from MSC ultimately reached the surface and gradually spread out to the shallower shelf areas. Bolin & Abbott (1963) referred to upwelling at the surface near the centre of the Bay which was markedly colder and more saline than waters in the bight areas; Abbott & Albee (1967) indicated that upwelling over MSC lowered temperatures in the centre of the Bay more so than it did elsewhere. During the upwelling season, the lowest temperatures were observed directly over MSC approximately 70% of the time (Marine Research Committee 1958). On one occasion (25–26 July 1973), a cluster of five drogues at 15m depth revealed a well-defined pattern of divergence near the head of the Canyon consistent with localized upwelling over MSC (Broenkow & Smethie 1978). Higher surface salinities in MB between May and October (Fig. 21, p. 42) are likewise consistent with the seasonal upwelling of higher salinity waters in MSC. Perhaps the strongest case for the importance of upwelling in MSC and the appearance of deep, upwelled water at the surface over the centre of the Bay is based on the unique relationship between phytoplankton growth and the physical and chemical properties that exists in MB. According to Barham (1956) and CalCOFI studies (Marine Research Committee 1952, 1953), phytoplankton growth lags the thermal cycle in MB by 1 to 2 months. The relatively long lag between changes in temperature and salinity (and nutrients), and productivity, is apparently related to the time required for the enriched waters that have surfaced to spread out to the surrounding areas of the Bay. This process takes place so rapidly that significant increases in productivity do not occur until these waters have had time to reach the surrounding areas. This sequence of events is well documented and results in a well-defined relationship between the physical, chemical and biological cycles in MB. The lower concentrations of chlorophyll over MSC indicated in CZCS satellite imagery are consistent with this picture (Hauschildt 1985). The exact rôle that upwelling in MSC plays in the above sequence, however, is still not clear.

Thermal patterns observed in satellite imagery, acquired over the past five years or so, do not clearly suggest the surfacing of upwelled waters over MSC. As indicated earlier, satellite data suggest rather that upwelled waters are often advected into MB from source regions outside the Bay. The recent satellite observations are not necessarily inconsistent with the results of the earlier studies, however. Similar to the upwelling observed by Hickey in Astoria and Quinalt Canyons, where deep, upwelled waters reach the surface indirectly, upwelled waters originating in MSC may likewise reach the surface indirectly and thus be difficult to detect.

In summary, all of the processes indicated above most likely contribute to the enrichment of the surface waters in MB. However, carefully designed experiments will be required to distinguish between, and prioritize the importance of, these contributing factors.

Tides

The tides in MB are mixed, being mainly semidiurnal. The tidal range is about 1.6 m with the K1, O1, M2 and S2 constituents contributing approximately 80% of the tidal variation.

Measurements of tidal phase between Monterey and Santa Cruz indicate that the incoming flood tide arrives approximately seven minutes earlier at Monterey than it does at Santa Cruz (US Department of Commerce, 1984). This time delay suggests that the incoming tide arrives at an angle to the direction normal to a line connecting Santa Cruz and Monterey. For high water, this angle is roughly 20°, assuming a water depth of 100 m. Tidal amplitude can be estimated from the tide tables if we assume an average water depth. In this simplified treatment, frictional and nonlinear effects are not included. To calculate the tidal current under these conditions, the expression for progressive, shallow water waves is used, where

$$u = \frac{gA}{C}\, Cos(kx - \omega t) \tag{14}$$

and

g = acceleration of gravity (9.8 m/sec^2)
A = amplitude of incoming semidiurnal tide (15 cm)
$C = \sqrt{gh}$, h equals water depth (100 m)
$kx - \omega t = \phi$, the phase function for propagating waves with phase speed equal to C

In the above expression, we have not included a reflected or outgoing wave, since the amplitude of the ebb tide is relatively small. When Equation (14) is evaluated, we obtain a value for u_{max} (i.e. for $kx\text{-}\omega t = 0$) of approximately 18 cm/sec. Resolving this value into eastward and northward components we obtain a value of approximately 7 cm/sec for the northward component for deep water (i.e. near the centre of the Bay). When averaged over a complete tidal cycle, such an unbalanced tidal component could lead to a residual circulation within the Bay that is directed to the north. Although this effect is small, it may not be insignificant in view of the relatively weak currents in MB, overall.

Baroclinic tidal currents are largely unknown in MB except for estimates made by Broenkow & McKain (1972) and Shea & Broenkow (1982) from observations near the head of MSC (Fig. 19). Tidal amplitudes of up to 120 m have been observed just below the main thermocline near the canyon head in a water depth of approximately 250 m. These internal tides are among the largest ever reported. Internal tidal heights decreased in the offshore direction, and the phase relationship between the barotropic and baroclinic tides varied during the two-week period of observation. From consideration of volume continuity between isopycnal surfaces, Shea & Broenkow (1982) computed alongcanyon current speeds of up to 8 cm/sec, and cross-canyon tidal current speeds of up to 13 cm/sec. These speeds are in reasonably close agreement with the nearly coincident cross-canyon speeds observed by Caster (1969), which ranged from 4 to 15 cm/sec.

Intense tidal motions have been observed in a number of submarine canyons in addition to MSC. For example, strong tidal currents were observed in Hudson (Hotchkiss & Wunsch 1982) and Baltimore (Hunkins & Wuntsch 1988) Canyons along the east coast of the US, and in Quinalt Canyon off the coast of Washington (Hickey 1989). According to Hotchkiss & Wunsch (1982), tidal amplification in Hudson Canyon was most likely due to the focusing of tidal energy near the canyon floor. Baines (1983) indicated that high amplitude tidal waves in steep, narrow submarine canyons could arise from the interaction of an incident wave at the mouth of the canyon with reflected waves from the external continental slope. His results are applied to MSC in the following section.

Dooley (1968), Njus (1968) and Shepard (1975) found periodic near-bottom currents in MSC having semidiurnal and shorter-period oscillations which they attributed to tidal forcing. Caster (1969), Hollister (1975), and Shepard et al. (1979) also found indications of supertidal frequencies in their current meter data from MSC.

Oscillations related to the internal tide with periods shorter than semidiurnal are frequently observed along continental slopes and shelves (Huthnance & Baines 1982, Jones & Padman 1983). One explanation for the occurrence of these motions is that they originate from nonlinear processes and may be enhanced in regions where water depths are small and/or bottom slopes are relatively steep (Huthnance 1981). Higher harmonics may also account for, or at least contribute to, these shorter period oscillations. As will be discussed in the following section, internal waves of semidiurnal tidal period may be of very high amplitude near the bottom of MSC; thus, it is likely that the above nonlinearities are associated with the expected high-amplitude internal waves in this region.

Another explanation for the presence of supertidal frequency oscillations in narrow canyons was proposed by Klinck (1988). According to Klinck's analytic model, superinertial frequencies are predicted for narrow canyons, with periods on the order of 0.2 of the local inertial period. MSC easily satisfies the conditions of "narrowness" in the sense defined by Klinck for the homogeneous case. For the latitude of MSC (36°N), this period is about 4 hours. We note the close agreement between Dooley's (1968) and Njus's (1968) observations of bottom currents and the results of Klinck's model. These oscillations may be forced by a pressure gradient associated with a sloping free surface which produces rotationally modified standing waves in the canyon which are constantly forced by the

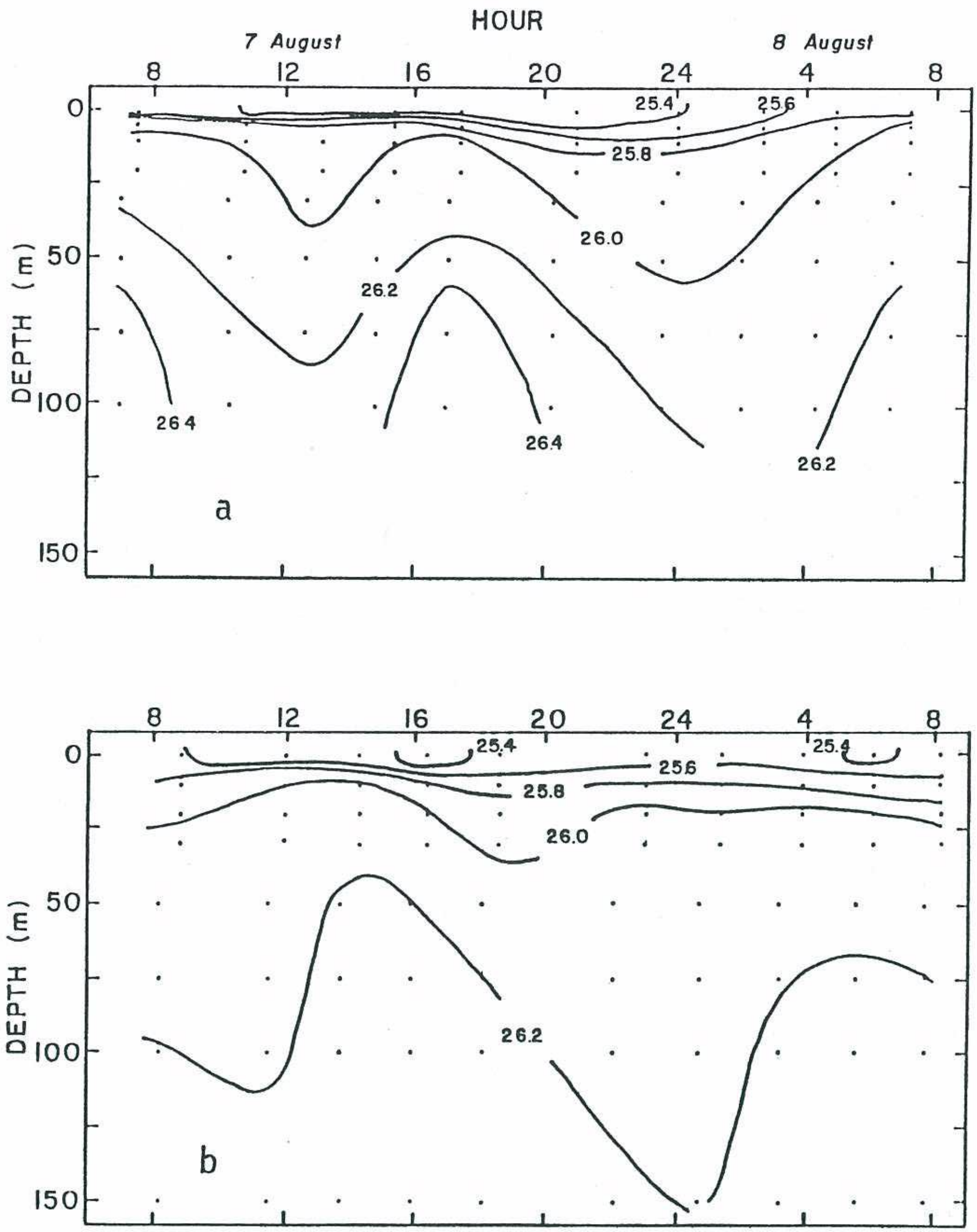

Figure 19 Vertical density (σ_t in gm/l) time series for anchor stations occupied on 7 and 8 August 1971, (a) 1 km, and (b) 4 km, west of Monterey Submarine Canyon head. (From Broenkow & McKain 1972).

oscillating free surface. For a stratified water column, Klinck indicated that pressure gradients near the bottom may produce these effects. Thus, both nonlinear and Coriolis effects associated with the semidiurnal internal tide may be important in generating supertidal frequency oscillations in MSC.

Internal waves

Internal waves are expected to occur in MB since they are often generated at the shelf break and in the vicinity of other major topographic features (Baines 1986). Internal waves on the continental shelf tend to propagate toward shore (e.g. Cox 1962). An example of internal waves at a depth of 9 m in southern MB (MBA, Fig. 1) is indicated by a sequence of week-long temperature time series (Fig. 20). The effects of seasonal stratification are apparent by examining the seasonal behaviour of the semidiurnal internal tide. During January, when the mixed layer depth exceeded the sensor depth, temperature variance was small, but as near-surface thermal stratification increased from March to May, the amplitude of the internal wave signature increased to almost 5°C.

Internal wave activity in MSC has been reported by Broenkow & McKain (1972), Shepard (1975), Shepard et al. (1979) and Shea & Broenkow (1982). Shepard (1975) observed upcanyon internal wave propagation in MSC, with propagation speeds of 25 cm/sec. Shepard et al. (1979), however, reported both upcanyon and downcanyon propagating internal waves. Broenkow & McKain (1972) occupied two anchor stations near the head of MSC and observed an internal wave of semidiurnal tidal period having an amplitude of 120 m (Fig. 19). The onshore phase speed was about 25 cm/sec, in agreement with Shepard.

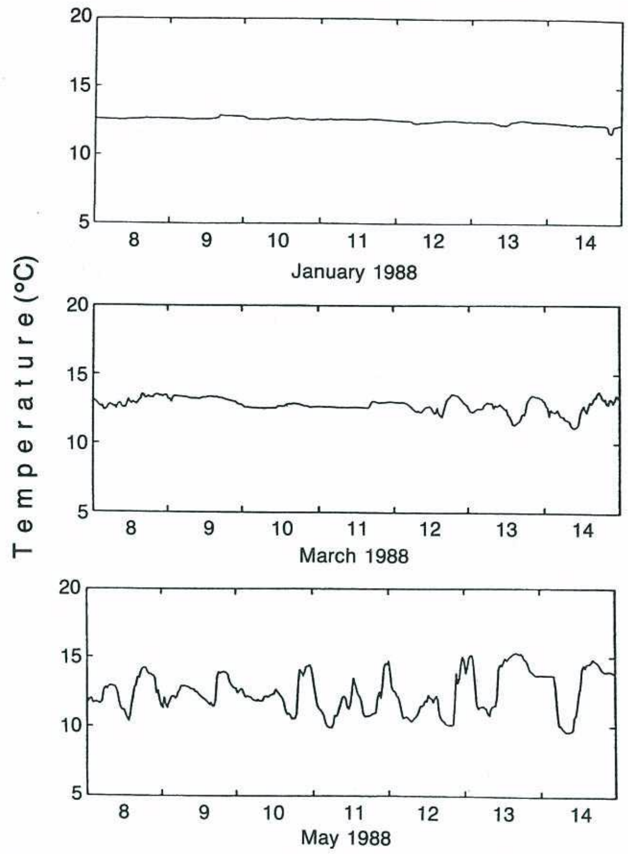

Figure 20 Temperature at 12 m in southern Monterey Bay near Monterey Bay Aquarium (MBA-Fig. 1) during one week periods in January, March and May 1988. Effects of the internal tide at this depth increase with development of seasonal stratification.

Shea & Broenkow (1982) suggested that the narrowing and shoaling of MSC may cause the energy associated with these internal waves to be focused and concentrated at the head of the Canyon where the energy may be dissipated by breaking internal waves and mixing. Hotchkiss & Wunsch (1982) similarly reported that internal wave energy is dissipated through mixing at the head of Hudson Canyon.

Baines (1983) studied tidal motions within submarine canyons using a nonrotating, stratified tank model and found that flow within steep, narrow canyons depends on the ratio of the bottom slope, d, to the internal wave characteristic slope, c. Bottom slopes in the 1000 to 2000m depth range for MSC are of the order of 0.15 (Table 1). The internal wave characteristic slope, neglecting earth rotation, is given by

$$c = \omega / (N^2 - \omega^2)^{1/2} \tag{15}$$

where ω is the semidiurnal tidal frequency ($\approx 1.4 \times 10^{-4}$ rad/sec), and N is the vertically-averaged Brunt–Vaisala frequency (taken to be 5.0×10^{-3} rad/sec). (According to Baines, the effects of earth rotation are small for motions inside a "narrow" canyon. For the "narrow canyon" approximation to be valid, it is assumed that the motion in the canyon is driven by an imposed pressure of external origin.) Thus, c is approximately 0.03 and d/c is 5. For values of d/c in the range of 1 to 10, Baines's results indicate for the bottom steepness conditions appropriate to MSC, that internal waves associated with the semidiurnal tide propagate to the bottom of the canyon where they are reflected upward into a number of "energy-propagating modes". The superposition of these modes may produce large amplitudes near the bottom of the canyon. According to Baines, including the effects of earth rotation for not-so-narrow canyons introduces additional complications in the form of internal Kelvin waves and associated cross-canyon variations. These experimental results are generally consistent with, and provide a basis for interpreting, the current measurements which have been made in MSC.

As indicated, internal waves have been observed on several occasions to propagate upcanyon in MSC. Also, deep current measurements have frequently indicated upcanyon flow. These observations may be related through nonlinear processes associated with high-amplitude internal waves (Bretherton 1969, Grimshaw 1972). Nonlinear effects become important when internal wave amplitudes become large compared to water depth. According to Grimshaw (1972), a wave-induced mean flow is associated with nonlinear internal gravity waves that is proportional to the square of the wave amplitude. Because high amplitude internal waves have been observed on more than one occasion and may be expected to occur frequently, based on theoretical considerations, these waves may contribute to the frequently observed upcanyon flow in MSC.

Within MB, mixing processes may be especially important near the head of MSC. The density field shown in Figure 19b indicates an asymmetric profile with a rapidly rising wave front followed by a gradually decreasing trailing edge. Cairns (1967), Thorpe (1971) and Thorpe & Hall (1972) indicate that shoreward-propagating internal waves become steeper and more asymmetric as they enter shallow water. When the wave amplitude becomes large compared with water depth, the nonlinear effects that produce these asymmetries become important. Wave profiles which are characterized by a sudden rise followed by a gradual decrease take on the form of a tidal surge or bore (Thorpe & Hall 1972). As the incoming tidal wave becomes steeper, instabilities and wave breaking often occur. Breaking internal waves in turn produce turbulence and mixing, processes which may be important in dissipating (the expected) internal wave energy near the head of MSC.

Local heating and river discharge

The residence time for surface waters is sufficiently long (~one week) that local warming causes SSTs inside MB to be higher than the SSTs of adjacent coastal waters. Increased SSTs occur between March and October and are clearly illustrated in the monthly maps of Lammers (1971; Fig. 17) and in AVHRR satellite imagery. Warmest surface waters are usually found in the northern bight near Santa Cruz. At times, warmer water can be traced in a narrow band extending NW around Terrace Point (Fig. 25, p. 48). Near-surface increases in temperature result in increased stratification within the upper 10 m or so of the Bay. This heating may create a horizontal pressure gradient which balances enhanced northward geostrophic flow within the Bay.

Freshwater input from the rivers that border MB is small compared to the total volume of bay waters. Thus, MB does not exhibit estuarine circulation. However, during winter and spring, the Salinas, Pajaro and San Lorenzo Rivers discharge modest amounts of fresh water into the Bay, contributing about 93% of the total river discharge received. Over 90% of the stream discharge occurs between December and May (US Geological Survey, 1985), reducing salinities significantly (0.5 to 1.0 ppt) in the nearshore regions during this period. This reduction in salinity during the winter and spring is clearly evident in the annual cycle for salinity at Pacific Grove (Fig. 21; Cayan et al. 1991). Monthly long-term means (1919–75) are shown together with the annual long-term mean (33.52 ppt) plus one indication of dispersion. The increase in salinity during spring and summer is most likely caused by at least two factors: the surfacing of upwelled water of higher salinity which originates, in part, in the California Undercurrent, and an excess of evaporation over precipitation. As discussed earlier, Broenkow & Smethie (1978) used the seasonal reduction in salinity to estimate residence times for water replacement in MB. Increased stratification due to lower salinities in winter and spring may also contribute to enhanced northward geostrophic flow within the Bay on a seasonal basis.

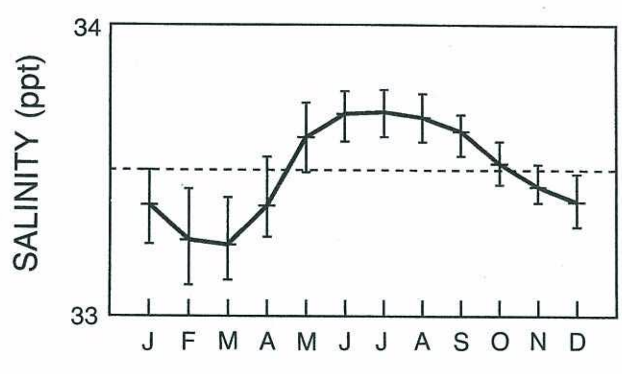

Figure 21 The annual cycle for salinity at Pacific Grove defined by the monthly means (1919–75). The monthly medians are indicated by the longer horizontal bars. The dashed line represents the long-term mean. A measure of dispersion is given by the 25th and 75th percentiles (shorter horizontal bars). (Adapted from Cayan et al. 1991).

Eddies

Satellite observations often show eddies along the central California coast. Off Point Sur, for example, cyclonic eddies have been found to entrain recently upwelled, cool, high salinity waters (Traganza et al. 1981). Satellite observations suggest that a meander or warm-core eddy, 50 to 100 km in diameter, may be present upon occasion just west of MB. Figure 22 shows an AVHRR satellite image from 8 October 1982 suggesting the presence of an anticyclonic circulation pattern just offshore of MB. An anticyclonic eddy in this area is also indicated in the surface dynamic topography shown in Figure 23. A cyclonic eddy has also been identified in the area west of MB (Broenkow & Smethie 1978).

Smaller eddies may also occur inside MB. According to Skogsberg (1936), eddies occur in the northern and southern bights of MB. Bolin & Abbott (1963) also indicate the occurrence of relatively stable eddies at the northern and southern extremities of the Bay. The existence of eddies in the bights is poorly documented, however, and in some cases there has been apparent confusion between the elliptical motion of drogues caused by tidal currents and the presence of closed circulation cells or eddies. We note that eddies within the Bay, when and where they occur, contribute to increased residence times for bay waters.

Offshore circulation

Circulation in MB is strongly influenced by the circulation outside the Bay (e.g. Skogsberg 1936). However, the available information on currents along the central California coast indicates a complex and occasionally contradictory picture of the circulation in this region. In this section, we summarize previous results that pertain to the circulation along the central California coast, giving particular attention to the region just outside MB.

In a summary of drift-bottle studies conducted along the west coast of the US between 1954 and 1960, Schwartzlose (1963) indicated that flow along central California was generally northward between October and February and southward during May and June. These results may be somewhat biased, since they were acquired during a period which included the 1957–58 El Niño, a period when increased poleward flow might be expected. Drift-bottle studies by Crowe & Schwartzlose (1972) indicate northward flow along central California from October through February and southward flow from April to July (1955 to 1969).

Dynamic topographies from the CalCOFI (California Cooperative Oceanic Fisheries Investigations) programme (Wyllie 1966) indicate a cyclonic circulation pattern along the central California coast for all months except April for the period 1949 to 1965. This cyclonic circulation implies a persistent northward flow near the coast along central California. Because of the sampling grid employed by CalCOFI, the station spacing may have been too sparse to resolve the narrow, equatorward jet that occurs near the seaward edge of the coastal upwelling zone.

Wickham (1975) found complex patterns of poleward flow at 50 and 200 m in the area out to 50 km off the Monterey Peninsula in a drogue and hydrographic study during August 1972 and August 1973. Flow at both depths near the continental shelf was poleward and apparently into the Bay around Point Pinos. Further offshore, the poleward flow turned to the northwest, perhaps deflected by an eddy that is often observed just west of MB. Based on geostrophic calculations, poleward flow predominated over the first 20 km offshore and

Figure 22 AVHRR satellite image (Channel 4) from 8 October 1982 showing an eddy-like circulation pattern offshore to Monterey Bay.

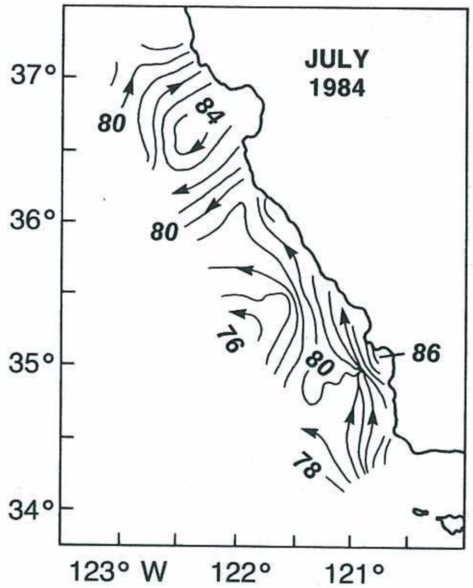

Figure 23 Surface dynamic topography relative to 500 m (1 dyn-cm contour interval) based on hydrographic data. (Adapted from Chelton et al. 1988).

down to a depth of at least 300 m. Offshore transects of temperature and salinity revealed a complex interleaving of filaments, often 10 km or less in width, apparently flowing in opposite directions. This complex structure was attributed to the mixing of equatorial and subarctic waters within the study area. These results also suggest that geostrophic calculations from earlier studies may not have adequately resolved the nearshore circulation in this region. (Also, the possibility that some of the observed variability was due to internal tides cannot be overlooked.)

Hickey (1979) summarized previous drogue and geomagnetic electrokinetograph (GEK) data and charts of dynamic height, and concluded that (a) poleward flows occur inshore of the southward-flowing California Current between Cape Mendocino and Point Conception except in April, (b) a southward, wind-driven coastal current occurs inshore of these poleward flows during spring and summer, and (c) poleward flow including subsurface geostrophic flow associated with the California Undercurrent is enhanced during winter. (It is doubtful that nearshore poleward flow between Cape Mendocino and Point Conception is always interrupted only in April. This reversal is probably related to the spring transition to coastal upwelling, a phenomenon that usually occurs anywhere between February and May.) Hickey also suggested that the earlier drift-bottle results would be consistent with the dynamic topographies if the nearshore, southward flow was narrow enough not to be resolved by the CalCOFI sampling grid.

In a more recent drift-bottle study (1977 to 1983), Dewees & Strange (1984) found generally northward flow north of Point Conception from November through February and southward flow from May through August. A striking exception occurred in July 1980, when all drift bottles released off Davenport moved to the north.

During the first half of 1984, currents on the continental shelf and upper continental slope were generally poleward from Point Conception to Point Sur (Chelton et al. 1988). The currents north of Point Sur to San Francisco were generally equatorward. During a period of unusually weak winds in July 1984, poleward flow south of Point Sur extended 100 km offshore. Surface dynamic heights (0/500 db) from high-resolution hydrographic data in July 1984 indicated a jet-like circulation directed offshore just north of Point Sur (Fig. 23). Salinities at 10 m further indicated that this jet separated two distinctly different water masses, lower salinity water from the north located above higher salinity water of southern origin. According to Chelton et al., somewhat similar conditions occurred off the central California coast in July 1981, as well. These results, together with those of Wickham (1975), indicate that in the region off MB, waters of equatorial and subarctic origin often meet and interact in complex ways.

Subsurface currents south of MB at 34.7°N over the shelf break (35 and 65 m) and at mid-shelf (35 and 65 m) were predominantly poleward throughout the year between May 1981

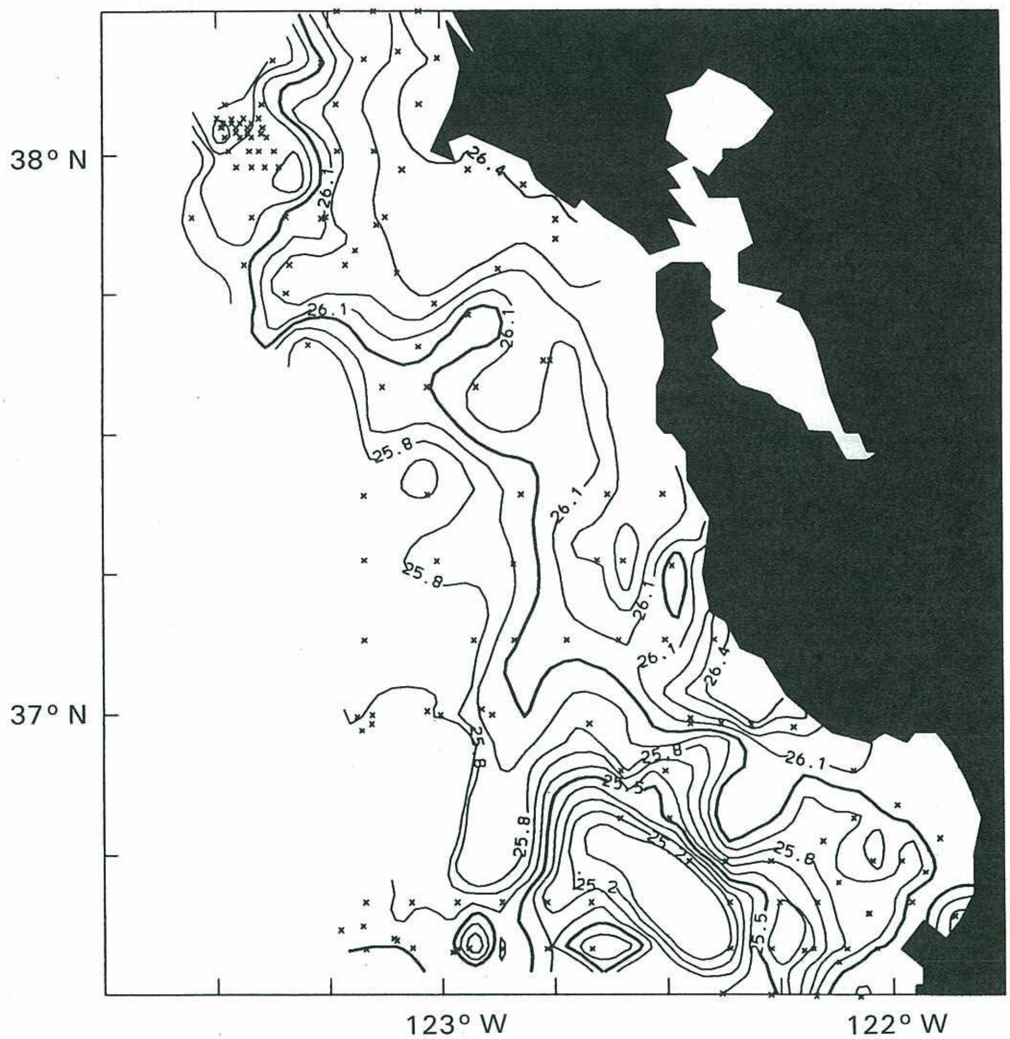

Figure 24 Distribution of density at 30 m off the central California coast based on hydrographic data acquired between 22 May and 3 June 1989. (From Schwing et al. 1990).

and May 1982 (Strub et al. 1987). There was a semiannual cycle in the flow with poleward maxima occurring in January and July, similar to the geostrophic results of Chelton (1984) for the California Undercurrent at the same latitude. Equatorward flow occurred only between March and May at 35 and 65 m. At 125 m, the flow was poleward throughout the year at speeds of 5 to 10 cm/sec.

A recent map of density at 30 m based on hydrographic data acquired between 22 May and 3 June 1989 (Fig. 24) emphasizes the convoluted nature of the flow off central California and, in particular, off MB (Schwing et al. 1990). The alongshore flow meanders significantly over the entire region shown (36.5° to 38.25°N). The 26.0 sigma-t contour which is located 25 to 50 km offshore north of MB, actually enters (and exits) the Bay at the southern end of the area, clearly indicating a pathway by which offshore waters may enter the Bay. Finally, a major anticyclonic circulation cell or eddy is located at 36.7°N, 122.5°W, approximately 65 km west of Moss Landing, again stressing the complexity of the circulation just outside MB, itself.

From the observations cited above, it is clear that circulation off the central California coast is complex. The region off MB contains eddies, interleaving alongshore flows, offshore jets, and differing water masses. The bathymetry associated with MSC and the perturbation in coastline geometry associated with the Bay itself, undoubtedly add, and may in fact be a primary contributing factor, to the complexity of the alongshore circulation in this region. A somewhat more coherent picture of the circulation emerges when only the hydrographic data acquired over a somewhat larger domain along the central California coast are considered. These data (i.e. dynamic heights) acquired since 1949 indicate that poleward flow within 50 km or so of the coast occurs most of the year. Closer to the coast, equatorward flow may occur particularly during spring and summer when coastal upwelling is important.

Fronts

Oceanic fronts often produce localized but intense flows along their boundaries. Such features are common in the upwelling zone along the central California coast (Breaker & Mooers 1986). Sharp frontal boundaries often separate recently upwelled water, which is cold and saline, from California Current water, which is warmer and fresher. Fronts arising in this manner are not density compensated, and thus strong baroclinic alongfront flows are expected.

The occurrence of a front in MB is suggested in the AVHRR image from 7 July 1981 (Fig. 25). This image and other AVHRR imagery show a major thermal boundary or front at the north end of the Bay extending roughly from Santa Cruz and paralleling the coast up to Ano Nuevo. This apparently recurrent feature may represent interaction between locally warmed bay waters exiting past Terrace Point and cold upwelled water that originates in the upwelling centre off Ano Nuevo. Alongfront flow to the north is suggested. This front has also been identified by Graham et al. (1992) and Graham & Largier (1992). Temperature and salinity data acquired in northern MB indicate what they refer to as an "upwelling shadow" inshore of the front that separates the cold upwelled water that originates north of MB from the warm coastal waters off Santa Cruz. This front persists throughout the upwelling season (Graham & Largier 1992). It is possible that the northward flow observed by Dewees & Strange (1984) off Davenport in July 1980 was due to entrainment just inshore of this frontal region. Another frontal region within the Bay occurs where the cool upwelled waters that often extend across the Bay from Ano Nuevo, meet the warmer waters inside

Figure 25 AVHRR satellite image (Channel 4) from 7 July 1981 showing a sharp thermal boundary or front at the northern end of MB and along the coast up to Ano Nuevo.

ow

rp thermal boundary

the Bay (Schwing et al. 1991). This front is similar to the previous front except that it extends further south and further into the Bay.

A front occurs at each ebb tide at the head of MSC, where waters from Elkhorn Slough (Fig. 1) and Moss Landing Harbor meet the cooler, saltier bay waters. Fronts may also occur upon occasion near the mouths of the Salinas, Pajaro and San Lorenzo rivers where freshwater plumes frequently occur.

According to Skogsberg, on the occurrence of fronts in MB,

> . . . on one occasion, the transition between the coastal and the oceanic waters was so sharp that a difference of a couple of degrees was observed between the temperature of the water under the bow and that of the water under the stern of our boat. In other words, there was a distinct rip at this place, and such rips were seen quite frequently, although we did not stop to examine them. These rips were often marked by the accumulation of drifting kelp and other debris.

Additional factors affecting the circulation of MB

Other factors, including inertial oscillations, coastal-trapped waves, 40–50 day oscillations, spring transition events and El Niño episodes, affect the circulation in MB. Near-inertial motions can be generated by impulsive wind episodes, and inertial motions near the continental shelf may be highly energetic (Baines, 1986). Wind-induced, near-inertial motions in the vicinity of MB probably occur more frequently in winter because of greater storm activity and associated wind events. However, such motions generated well inside the Bay may be restricted in their development due to the limited size of the Bay; thus, their trajectories may be distorted and hence difficult to identify.

Wind-forced, sea-level fluctuations propagate poleward as coastal-trapped waves along the California coast (e.g. Halliwell & Allen 1984). Along the coasts of southern and central California, these fluctuations tend to originate along northern Baja California. Sea level data from the Monterey tide gauge showed evidence of coastal-trapped waves with periods of 2 to 20 days.

Due to local wind forcing, 40–50 day oscillations in SST occur along the central California coast (Breaker & Lewis 1988). They have been observed inside the Bay at Pacific Grove as well. The impact of these periodic variations on the local circulation has not been established.

The change from winter, non-upwelling conditions to upwelling in spring along the central California coast occurs abruptly in some years (Breaker & Mooers 1986) and is often referred to as the "spring transition". Occasionally, this transition is dramatic in its abruptness and magnitude. The spring transition in 1980 can be easily identified in the previous upwelling index time series (Fig. 3, p. 5) by the abrupt change to positive values that occurs during the week of 9 March. SSTs at Granite Canyon dropped by 4°C over a period of 6 days, starting on 11 March 1980 (Breaker & Mooers 1986). Currents throughout the water column on the continental slope, just south of Point Sur, reversed from poleward to equatorward temporarily during this event (Wickham et al. 1987).

Spring transitions in SST in 1977 and 1980 inside the Bay at Pacific Grove are shown in Fig. 14 (p. 26). Although there was a significant delay in the onset of the spring transition at Pacific Grove during both years and its magnitude was reduced, its occurrence inside the Bay is readily apparent. A sudden decrease followed by a significant increase in near-surface salinity at two locations in MB in February and/or March 1973, observed by Broenkow & Smethie (1978), may reflect the spring transition which occurred in that year (9 March 1973; Breaker & Mooers 1986).

Variability on interannual time scales is strongly influenced by El Niño episodes at mid-latitudes along the California coast (Enfield & Allen 1980, Chelton & Davis 1982, Breaker et al. 1984, Breaker 1989). Daily SSTs at Pacific Grove from 1940 to 1985 (not shown) reveal significantly warmer temperatures (2–4°C) in 1941, 1957–58, and 1982–83, periods during which strong El Niño episodes occurred. The El Niño warming influence along central California is usually seasonal, being most apparent during the fall and winter (Breaker 1989). Freshwater input from local streams may also increase during major El Niños, reducing salinities significantly in the nearshore regions (Hauschildt 1985). Sea-level anomalies at Monterey also indicate significant interannual variability associated with El Niño episodes (Bretschneider & McLain 1983).

Synthesis of results

In this section, the diverse observations, model results and hypotheses presented earlier are synthesized in an effort to create a generally consistent picture of the circulation of MB. As a part of this synthesis, we include a conceptual model for the circulation of MB.

MB does not exhibit estuarine circulation because of its broad contact with the coastal ocean and because the influx of fresh water from local river discharge is small compared to the volume of the Bay. The mean surface circulation of MB is cyclonic, a result recently reaffirmed by high-frequency radar (CODAR) measurements. The surface circulation of the Bay is characteristically weak in magnitude with speeds typically in the range of 5–20 cm/sec. Flow into the Bay frequently takes place around Point Pinos. Satellite imagery suggests that cold water upwelled along the coast north and south of MB may, at times, be advected across the entrance of the Bay. Reversals in current direction from several days to several weeks (from prevailing northward to occasional southward) also occur. Such reversals may follow periods of intense coastal upwelling where the advection of upwelled waters, primarily from the north, enter the Bay and dominate the local circulation at least temporarily.

During the fall, water may occasionally enter the Bay directly from the west as the upwelling-favourable winds begin to relax and the alongshore flow is in seasonal transition. Although fewer observations are available to document conditions within the Bay during winter, poleward flow associated with the Davidson Current apparently enters the Bay directly from the south, producing northward flow inside the Bay which more-or-less parallels the flow offshore. Upon departure from the Bay, these waters then merge with the poleward flow along the coast. The results of a feature-tracking analysis using AVHRR satellite imagery support this view of bay circulation during winter.

It is not clear why the prevailing surface circulation in MB is northward. Whether this circulation is due primarily to forcing from outside the Bay, or from within, needs to be established. However, because the incoming tide arrives earlier at Monterey than it does at Santa Cruz, it is possible that this phase difference produces a net, unbalanced tidal residual that contributes to northward flow inside the Bay. As indicated earlier, during spring and summer, local heating and river discharge increase stratification within the Bay consistent with locally enhanced northward geostrophic flow. However, the relative importance of these effects is not known.

Circulation at intermediate depths (i.e. between ~25 and ~150m) in MB is not well-known. Monthly mean distributions of temperature on selected isothermal surfaces indicate the existence of a thermal high over MSC at certain times of year (Lammers 1971). Application of the thermal wind equation to these data suggests that geostrophic circulation within the thermocline in MB may be anticyclonic, and hence opposite to the flow at the surface. A depth of geostrophic current reversal of the order of 70m is plausible, based on Lammers's data. The occurrence of the thermal high over MSC may be due to the combination of momentum transfer from the California Undercurrent, producing anticyclonic circulation at depth within the Bay (Garcia 1971, Bruner 1988), and topographic steering (Hughes et al. 1990).

Recent inflow data acquired across MB in May 1988 indicate cyclonic rather than anticyclonic circulation at depth (~75 to ~400m) at least on one occasion (Koehler 1990). This pattern of circulation coincided with a period when flow in the California Undercurrent was weak or nonexistent. On a seasonal basis, weak flow in the Undercurrent is expected during the spring along the central California coast (e.g. Chelton 1984). This is also a period when the subsurface thermal high described by Lammers is poorly developed. Thus, during periods of weak flow in the Undercurrent, the offshore circulation may simply follow the topography into MB, lacking the necessary momentum to produce typical cavity flow. These results suggest that flow at intermediate depths (and deeper) in the Bay may be either anticyclonic or cyclonic depending on conditions outside the Bay. Also, in either situation, flow at the surface may be opposite to the flow at depth, indicating the existence of at least a two-layer system of circulation in MB.

The model results of Hughes et al. (1990) suggest the possibility of a three-layer system in the case of MB where a deep bottom layer might rotate cyclonically in opposition to anticyclonic circulation at intermediate depths. Cross-canyon flows at deeper levels in MSC (> ~350m) observed on a number of occasions are generally consistent with this possibility. For the case where cyclonic flow was observed at depth in the Bay, the depth range for cyclonic flow (~75 to ~400m) far exceeded the range associated with "intermediate" depths. For this situation it is clearly possible that cyclonic flow at intermediate depths combines with the theoretically predicted cyclonic flow at deeper levels to produce continuous cyclonic flow over a much greater depth interval. Finally, the increase in stratification which results from river discharge and local heating may additionally promote the development of a baroclinic circulation in MB.

Waters that enter MB through MSC gradually rise due to onshore flow which is directed upward through the guiding influence of the Canyon (i.e. bathymetrically-induced upwelling). Onshore flow in MSC may be due in part to nonlinear effects arising from the very high amplitude internal waves associated with the internal tide which originates offshore. Deep flow in MSC may also be driven by a sloping free surface, in accordance with the theoretical results of Klinck (1988).

Waters entering MB via MSC most likely originate within the depth range of the California Undercurrent, consistent with the model results of Hughes et al. (1990), which indicate the intrusion of undercurrent waters for relatively wide canyons and the importance of topographic steering at intermediate depths. Thus, this current is expected to be the primary source for waters lifted from deeper to shallower levels within MSC. The deep circulation in MSC is related to circulation at lesser depths within the Bay in several ways. First, the deep waters are expected to rise within the Canyon on shorter time scales due to tidal pumping, and second, they rise on seasonal time scales due to bathymetric uplifting. In this regard, the frequently observed onshore, or upcanyon, flow is consistent with the upward vertical motion that occurs at deeper levels in MSC.

Seasonal changes in the deep circulation within MB may be related to seasonal changes in the strength of the California Undercurrent; thus, deep waters may enter the Bay producing anticyclonic flow at depth when flow in the California Undercurrent is strong. Deep, upward vertical motion in MSC often starts as early as December, a process that may be caused by the intensification of poleward flow in the California Undercurrent. The strongest subsurface poleward flow north of Point Conception occurs during winter. Thus, the intensification of the subsurface thermal high over MSC which occurs in February and March may be related to the winter increase in poleward flow associated with the California Undercurrent, allowing a month or so for the deep response to reach intermediate depths.

Early observations (late 1960s to mid-1970s) on bottom currents in MSC indicated unexpectedly strong flows, occasionally downcanyon but more often, upcanyon, in the depth range of 100 to 1500 m. Measured current speeds ranged from a few cm/sec to a few tens of cm/sec. More recently (1988–89), observations at the bottom of MSC using a ROV and the DSRV ALVIN suggest that flows in this region, particularly in water depths ranging from about 250 to 1200 m, may be more vigorous than previous measurements had indicated. Estimates of bottom current speeds at these depths ranged from a few tens of cm/sec to at least 100 cm/sec. Again, the flows tended to be more upcanyon than downcanyon.

A conceptual diagram for one possible mode of circulation in MB is shown in Figure 26. Mean flow in the surface layer based on the observations presented earlier is cyclonic or generally to the north inside the Bay with speeds of ~10 cm/sec. At intermediate depths, the flow may be anticyclonic inferred from the results of Lammers (1971) with speeds of 5 cm/sec or so. At deeper levels below 200–300 m, the horizontal circulation may again be cyclonic according to the theoretical results of Hughes et al. (1990) with mean speeds of the order of a few cm/sec. The situation for anticyclonic flow at the surface and cyclonic flow at depth is not included because it may be less representative of the long-term mean circulations in these layers. A second figure (Fig. 27) shows this circulation in a vertical plane centred along the axis of MSC. Here, we have included a vertical component to emphasize the importance of the upcanyon and downcanyon flows, with the upcanyon flows generally exceeding the downcanyon flows. As indicated, these flows are expected to be vigorous with speeds approaching several tens of cm/sec. Near the bottom along the axis of MSC, we consider the instantaneous flow rather than the mean flow because of its magnitude and periodic behaviour. These flows, however, are primarily restricted to the region along the axis of the Canyon with maximum elevations of only 100 m or so.

Oceanic conditions in MB can change within just a few days. Residence times for bay surface waters range from 2 to 12 days based on several methods of estimation. The Rossby radius of deformation (internal) ranges from about 10 km over MSC to about 1 km around the periphery of the Bay. The order-of-magnitude variation in the Rossby radius is due to the presence of MSC. A dynamical scale analysis for MB suggests that in addition to geostrophy, upwelling in MSC contributes significantly to the local circulation.

Barotropic tidal currents in MB are of the same order of magnitude as the mean flow (15–20 cm/sec). Significant tidal-related current variability was observed at deeper levels in MSC which may be partially explained on theoretical grounds (Baines 1983). For steepness conditions appropriate to MSC, internal waves of tidal frequency propagate to the bottom of the Canyon, where they are reflected upward and then combine constructively to produce

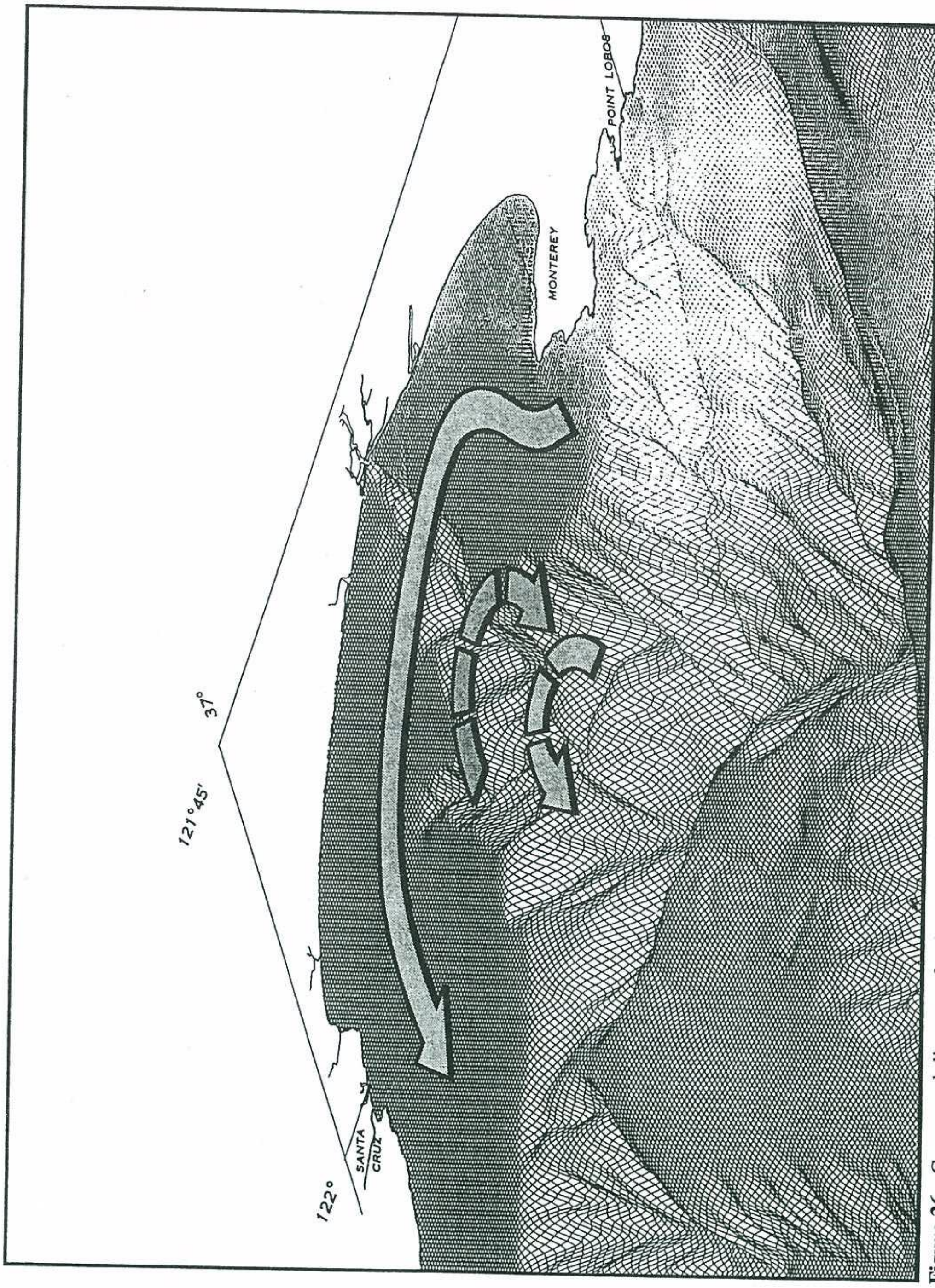

Figure 26 Conceptual diagram of a three-layered system of circulation in MB. Flow at the surface is cyclonic or northward, at intermediate depths the flow is anticyclonic and at deeper levels the flow is again cyclonic. Dashed curves indicate our uncertainty in characterizing the flow below the surface due to the lack of observations (except near the bottom along the axis of MSC). (Adapted from NOAA 3-D Map PI-1 of the central California coast).

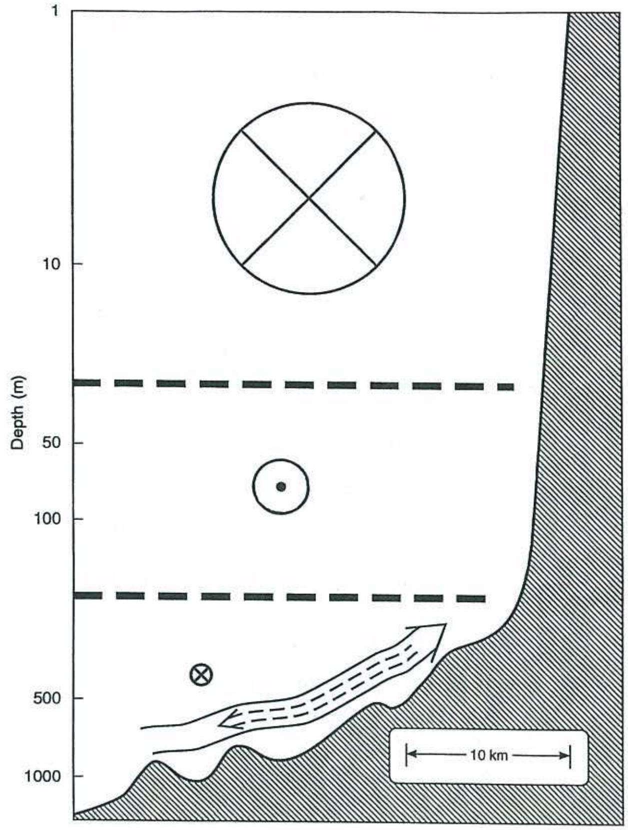

Figure 27 A vertical section of the flow in MB (i.e. conceptual) taken along the axis of MSC. The depth axis is logarithmic to emphasize the shallower levels. The relative speeds in each layer are indicated by the diameters of the circles that indicate the direction of flow (circles containing x's indicate flow into the page and the circle containing the dot indicates flow out of the page, following the standard convention). Dashed lines indicate approximate depths at which the flow may reverse direction. A vertical component has been included along the bottom to indicate the vigorous flows that have been observed there. In this case, the stronger upcanyon flow is indicated by the broader arrow and the weaker downcanyon flow by the narrower dashed arrow.

large amplitude variations. Supertidal frequency oscillations observed in current meter data acquired near the bottom of MSC are likely due to nonlinear processes associated with the expected strong semidiurnal internal tide and/or superinertial frequency oscillations arising in the manner described by Klinck (1988). The tidal and near-tidal variability in MSC is generally consistent with similar observations made in other major submarine canyons along the east and west coasts of the US. Very large amplitude internal tides (>100m peak-to-peak) occur near the head of MSC. These internal waves may take on the characteristics of a tidal bore as they approach the head of MSC and ultimately break in order to dissipate their internal wave energy. Breaking internal waves in turn produce turbulence and mixing and it is likely that these processes are important in this region.

Coastal upwelling *per se* does not appear to be important inside MB; however, the occurrence of cooler, nutrient-enriched waters in MB may be due to (a) advection of upwelled waters from the north (or south), (b) bathymetrically-induced upwelling within MSC, and/or (c) offshore upwelling (i.e. Ekman pumping) due to prevailing positive wind stress curl. Eddies may occur in the northern and southern bights of MB, although their presence at these locations is not well documented.

The circulation in MB is strongly influenced by the circulation offshore because the Bay is part of the coastal ocean, and because it is also connected to the waters offshore at deeper levels through MSC. The circulation offshore is complicated by many factors, including the presence of meanders and eddies, interleaving alongshore flows which may contain waters from different water masses, offshore jets and MSC. The complex nature of the offshore circulation, however, may be due in part to the presence of the Bay itself, and to MSC. For example, it is possible that the eddies, which are frequently observed offshore, are topographically coupled to MSC. Also, the theoretical results of Killworth (1978) and Wang (1980) indicate that irregularities in shelf topography may produce complex circulation patterns near canyons due to (a) the reflection and scattering of long shelf waves, and (b) the excitation of higher mode internal waves. The theoretical results of Klinck (1989) indicate that the width of a submarine canyon determines the strength of the cross-canyon flow and thus the importance of the canyon in affecting the alongshore flow above it. There have been numerous observations of cross-canyon flow in MSC suggesting the importance of the Canyon in influencing the coastal circulation around MB. Finally, El Niño episodes which are clearly evident in higher SSTs along the (central) California coast often enhance poleward flow nearshore, and thus are expected to influence the circulation in MB as well.

Because MSC appears to play such an important rôle in influencing the circulation in MB, we have endeavoured to generalize our results by identifying other bay/canyon systems around the world where the presence of canyons may influence the local circulation. A number of studies have been conducted that address the circulation in submarine canyons. In almost all cases, however, these canyons are located along the outer continental shelf and thus are usually far-removed from coastal embayments such as MB. One study, however, attempted to relate the circulation of a particular bay to the possible influence of an adjacent submarine canyon. Marcer et al. (1991) used a hydrodynamic model to ascertain the three-dimensional circulation of Cannes Bay off the southern coast of France where a deep (~ 800m) trench leads directly into this embayment. They found that deep upwelling occurred on the western side of the bay, clearly a result of the canyon.

Based on the above, 16 bay/canyon systems are presented in Table 2 where canyons may influence the bay circulation. Four references were gleaned to obtain the information given in this Table: Shepard & Dill (1966), Reimnitz & Gutierrez-Estrada (1970), Bouma et al. (1985), and Marcer et al. (1991). In a number of cases, these bays receive large amounts of fresh water from adjacent rivers which may contribute to significant estuarine circulation, unlike MB.

Table 2 Bay/canyon systems around the world where canyons may influence bay circulation

Bay	Canyon	Location	Contributory rivers	Comments
Cannes	Cannes Trench	Mediterranean Sea 43°31′N; 7°E		Very small bay
Carmel	Carmel	Central California coast 36°32′N; 121°56′W	Carmel (negligible inflow)	Just south of MSC, but bay and canyon are both much smaller
San Lucas	San Lucas	Southern tip of Baja Peninsula 22°53′N; 109°54′W		Relatively small bay – dominated by canyon
Los Frailes	Los Frailes	SE coast of Baja 23°22′N; 109°26′W		Very small bay
Palmas	Pescadero	East coast of Baja 23°42′N; 109°40′W		Two other small canyons also feed into Palmas bay
Petacalco	Petacalco	Pacific coast of Mexico – 280 km NW of Acapulco 17°59′N; 102°06′W		Very small bay – dominated by canyon
Tokyo	Tokyo	East coast of Japan 35°23′N; 139°44′E	Ara, Edo and Furu-tone plus several others	Canyon only becomes well-established beyond the bay and ultimately feeds into the Sagami Trough
Sagami	Sagami Trough	East coast of Japan – just west of the Miura Peninsula 35°13′N; 139°20′E	Sagami plus several smaller rivers	Large bay containing at least six smaller canyons
Suruga	Nankai Trough	East coast of Japan – adjacent to Izu Peninsula	Fuji plus several smaller rivers	

Suruga	Nankai Trough	East coast of Japan – adjacent to Izu Peninsula 35°05′N; 138°40′E	Fuji plus several smaller rivers	
Biscay	Cap Breton	SW coast of France 43°40′N; 1°40′W	Adour	Very large bay – canyon head comes very close to shore
Tagus	Lisbon	Coast of Portugal 38°35′N; 9°20′W	Tagus	Canyon extends seaward of the Tagus river estuary
Gulf of Lions	Petit-Rhone	South coast of France	Rhone	Several small canyons, including the Petit-Rhone enter the Gulf of Lions
Congo	Congo	Off the west coasts of Angola and Zaire 6°S; 12°E	Congo	The Congo river discharges a large amount of fresh water into the Congo river estuary
Bengal	Swatch-of-no-Ground	Off east coast of India 15°N; 90°E	Ganges and Bahmaputra	Very large bay with major input from both rivers
Manila	Manila	Phillipine Island of Luzon 14°31′N; 120°45′E	Pampanga and Angat plus others	
Koddiyar	Trincomalee	Off NE coast of Ceylon 8°N; 81°15′E	At least two rivers feed into Koddiyar Bay	

Conclusions

The currents in MB are generally weak and as a result, historically, they have been difficult to describe.

Indirect evidence suggests that flow at the surface and at intermediate depths may be different in MB because it is stratified. Thus, one possible mode of circulation is that waters at the surface move cyclonically, whereas the waters at depth (within the thermocline) move anticyclonically. If the above premise is generally correct, it has important implications concerning the eventual fate of sewage and other pollutants that are discharged into the Bay. For example, particulates discharged at or near the surface which are initially transported through the Bay past Santa Cruz may sink to deeper levels where they could be subsequently transported back into the Bay.

Because of the wide, direct connection with the coastal ocean, the circulation of MB is strongly influenced by the circulation offshore. However, the circulation offshore is complex, consisting of eddies, interleaving alongshore flows, offshore jets and water masses of different origin, and as a result is not completely understood. Some of this complexity may be due to the presence of the Bay itself, and MSC. Also, to obtain a better understanding of the circulation within MB, it may be necessary to obtain a better understanding of the circulation in this complex region beyond the Bay.

Although the importance of offshore flow on the circulation in MB is often emphasized, local factors such as heating, river discharge, and residual tidal effects may also contribute to northward flow inside the Bay. However, overall, we still do not understand why the surface circulation in MB is predominantly northward. Mechanistic models of bay circulation will be required to distinguish between the effects of local and remote forcing. Also, there is an urgent need to acquire observations on the circulation in the interior of MB to determine the predominant direction of flow, which in turn will help to provide more realistic initial and boundary conditions for hydrodynamic models of bay circulation. Finally, it will be important for hydrodynamic models to be fully stratified to obtain realistic results on the circulation of MB, and model domains should extend well beyond the boundaries of the Bay itself.

Variability associated with the semidiurnal internal tide in MSC and near its head is far greater than elsewhere in MB, where the barotropic tide predominates. Much of the hydrographic data that have been acquired in this area most likely contain strong tidal influence. Consequently, it will be important to delineate the region around MSC where the influence of the internal tide is significant.

Mixing at the head of MSC must be intense in order to dissipate the energy associated with the very high amplitude internal waves which have been observed and are expected to occur in this region. The observational techniques of Apel et al. (e.g. Apel et al. 1985) might be useful in detecting and monitoring this internal wave activity.

The availability of high-resolution satellite imagery has helped to improve our understanding of the surface circulation in MB significantly. For example, satellite data suggest that upwelled waters in MB often originate at locations outside the Bay; Ano Nuevo, just north of MB, is one such location.

Several types of upwelling contribute to the circulation in and around MB; they include coastal, offshore and bottom-enhanced upwelling. Additional surface and subsurface measurements are needed to help distinguish between the types (and sources) of upwelling and upwelled waters in MB and to prioritize their importance. For example, tracers injected at deeper levels in MSC could be helpful in determining how (and if) the deep waters in this region ultimately reach the surface.

Most of the circulation data for MB have been acquired during the spring, summer and fall, periods when upwelling-related processes are usually important. During the winter it has been assumed that flow at the surface is dominated by the poleward-flowing Davidson Current. More observations are needed to confirm the simple northward flow pattern that has been indicated within and beyond the Bay during this period.

To interpret correctly oceanographic data acquired in and around MB, it is important to consider whether or not El Niño conditions are present.

Finally, we conclude that most of Skogsberg's original conclusions have stood the test of time. He had great insight into the nature of the circulation that occurs in MB, and his results were remarkable in view of his limited data sets; temporally dense but spatially sparse.

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