An Offshore Eddy in the California Current System

Part II: Surface Manifestation

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Abstract — Ship and satellite observations taken over the last thirty years show that mesoscale patterns of sea surface temperature (SST) in the California Current System are consistently found throughout the year and usually occur in approximately the same geographical locations. Typically, these patterns are more pronounced in fall/winter than in spring/summer. The temporal and spatial characteristics of these persistent features were examined with satellite infrared (IR) measurements during winter 1980-81. In January 1981, a ship surveyed the vertical structure of several physical, chemical, and biological parameters beneath one of these SST features centered near 32°N, 124°W. The surface IR pattern had a length scale of 200 km and a time scale of about 100 days. It disintegrated following the first two storms of the winter season. Motion studies of the pattern in late October indicated an anticyclonic rotation with maximum velocities of 50 cm s⁻¹ at 50 km from the axis of rotation. As a unit, the pattern advected southward with an average speed of 1 cm s⁻¹. Thermal fronts, determined from the satellite imagery, were strongest (0.4°C km⁻¹) along the rim of the pattern and were advected anticyclonically with the pattern; their length scales were 20-30 km in the along-front direction and less than 10 km wide. The hydrographic data revealed a three-layer structure beneath the surface pattern: a 75 m deep surface layer, a cold-core region from 75 to 200 m depth, and a warm-core eddy extending from 250 to 1450 m. The anticyclonic motion of the surface layer was caused by a geostrophic adjustment to the surface dynamic height anomaly produced by the subsurface warm-core eddy. The IR pattern observed from space reflects the horizontal structure of the surface layer and is consistent with a theoretical model of a mean horizontal SST gradient perturbed by a subsurface density anomaly. Ship of opportunity SST observations collected by the National Marine Fisheries are shown to resolve mesoscale patterns. For December 1980, the SST pattern near 32°N, 124°W represented a 2°C warm anomaly compared with the 20-year mean monthly SST pattern.

1. INTRODUCTION

The mean sea surface temperature (SST) off California increases (0.005°C km⁻¹) in a southwesterly direction. This gradient results from both the southward advection of cool surface water by the California Current System (CCS) along the eastern boundary of the Pacific and the upwelling of cold water along the coast. Every fall and winter for the last thirty years, shipboard observations have shown that mesoscale patterns of warm SST intrude the cooler coastal water and perturb the mean gradient (Anonymous, 1963; Burkov and Pavlova, 1980; Wyllie and Lynn, 1971; Simpson, 1982). In particular, a mesoscale feature centered near 32°N, 124°W has consistently been found in ship data since 1950 (Wyllie, 1966) and in satellite infrared (IR) data since 1975 (e.g., Bernstein, Breaker and Whritner, 1977). A detailed satellite and ship study of this feature was made during the winter of 1980-81. This paper compares remotely sensed mesoscale surface

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brightness temperature patterns in the CCS to observed subsurface structure and processes.

2. OBSERVATIONS

In this study satellite IR observations of the CCS were monitored daily from September 1980 through March 1981. These observations were used to determine the time and length scales of several mesoscale surface IR brightness temperature (IRBT) patterns. The relationship between IRBT and SST is discussed below. A ship survey examined the physical (Simpson, Dickey, and Koblinsky, 1984), chemical (Simpson, 1984), and biological (Haury, 1984) properties of the ocean beneath the mesoscale IRBT pattern centered near 32°N, 124°W (feature A in Fig. 1).

Fig. 1. An infrared image from the NOAA-6 AVHRR 11 μm band on December 11, 1980 at 0345 GMT. The 3000 m bathymetric contour, separating the continental shelf and slope regime from the deep ocean is shown in white. See the text for a discussion of the square region outlined in white and features labeled A-F.
The satellite IR observations were made with the Advanced Very High Resolution Radiometer (AVHRR) on the polar orbiting NOAA-6 and TIROS-N satellites. This instrument senses the upwelled irradiance in the 0.7, 0.9, 3.7 and 11 μm bands of the electromagnetic spectrum at a distance of 833 km above the surface (Schwalb, 1978). The spatial resolution of the AVHRR is 1.1 km at nadir. The field of view of the instrument directly available at an earth receiving station is approximately 2000 km east/west (perpendicular to the flight path) and 4000 km north/south (parallel to the flight path), because the orbit inclination angle is about 99 degrees. NOAA-6 passes over California at approximately 0400 and 1600 GMT, whereas TIROS-N crosses at 1000 and 2200 GMT. The AVHRR on TIROS-N ceased to operate properly after November 1980, but AVHRR data from NOAA-6 were available for the entire study period.

All AVHRR observations were directly received and processed at the Scripps Satellite Oceanography Facility in La Jolla, California. Routine processing of these observations included: digitization of the analog signal transmitted by the spacecraft; earth location and removal of instrument attitude and orbital variations (Legeckis and Pritchard, 1976); interpolation of the data to an equirectangular grid at 1.1 km resolution; and calibration of the 3.7 and 11 μm channel data to IRBT (Lauritson, Nelson, and Porto, 1979). The details of these routines are given in Young (1981) and Young and Fahle (1981). The resulting temperatures were then linearly mapped to a range of 256 grey shades and displayed as a black and white image on a cathode ray tube with 512 x 512 picture elements (pixels). The images were contrast enhanced for feature identification by varying the linear grey shade mapping, and photographed (e.g., Fig. 1). All images shown in this paper were mapped so that warmer temperatures are associated with lighter grey shades; hence, clouds are black. Because they have been enhanced separately for contrast, intercomparison of IRBTs is not possible with the images displayed in this paper. Additional analysis of the image data included the estimation of SST gradients by simple differences and the measurement of velocity from feature displacement. These latter methods are described in more detail below. The absolute positional accuracy of the processed images is approximately ± 0.10°K (Bernstein, 1982a).

The ship survey was conducted to examine the properties of the underlying ocean and to provide a dynamical interpretation of the satellite observations. From January 9th to 17th, 1981, R.V. New Horizon mapped the detailed vertical structure beneath the mesoscale feature A shown in Fig. 1. Vertical profiles of temperature and salinity were made to 1450 db with a Neil Brown CTD/O2 system. The vertical resolution was 0.5 db, which has been block averaged to 2.5 db. Profiles were made approximately every 20 km along north/south and west/east transects through the feature A. The station pattern is shown in Fig. 1 of Simpson, Dickey, and Koblinsky (1984) and further details of the sampling and data analysis are given in the same paper.

3. DESCRIPTION

3.1. Large-Scale IRBT Pattern

One of the clearest days for infrared observation during the study period was December 11, 1980. The 11 μm band image for this day is shown in Fig. 1. This figure reveals that there are three warm mesoscale patterns (A, B, C) which intrude into the cooler coastal water (D). Pattern C is partially obscured by clouds. South of A, a sharp front (E) extends from A parallel to the coast. It is important to notice that A, B, C, and E are deep ocean features that are seaward of the California continental shelf and slope region which lies between the coastline and the 3000 m bathymetric contour in Fig. 1. In the Southern California Bight a warm pool of water (F) is found. Some or all of these features are always present in fall-winter IR observations of the ocean off California. In this paper, we are concerned with features A, D, and E in the area outlined by the white box (440 x 440 km) in Fig. 1.
Fig. 2. A sequence of NOAA-6 AVHRR 11 μm band infrared images collected between September 30, 1980 and February 21, 1981. The data for each image come from the area outlined by the white box in Fig. 1.
The temporal evolution of the IRBT pattern A (Fig. 1) is shown in a sequence of IR images in Fig. 2. The time series of images are separated by approximately one-month intervals from September 30, 1980 to February 21, 1981. Pattern A, the dominant feature in Fig. 2, was first clearly discernible in mid-October (not shown) and remained coherent until February when its basic shape changed. The lifetime of the pattern was about 100 days, and the length scale was about 200 km. The feature remained remarkably stationary in space, especially from November 27 to January 15. The mean advection of A was 1 cm s\(^{-1}\) towards the south over the four-month period. Examination of consecutive images at 12-hour intervals revealed that the feature rotated anticyclonically at speeds less than 50 cm s\(^{-1}\) (details given below). By comparison, East Australian Current eddies (Nilsson and Cresswell, 1981) and Gulf Stream rings (Lai and Richardson, 1977; Olson, 1980) are usually found to advect zonally at 3-5 cm s\(^{-1}\) and rotate at speeds of 100 to 200 cm s\(^{-1}\) near the surface.

The rotation of A caused a band of warm water from south of the front E to be wrapped around a patch of cooler water. This provided the strong contrast shown in the image. The large area of cold water shown in the upper right hand corner of the image for September 30 (marked with an 'X' in Fig. 2) extended from the cooler coastal water D into A. This evidence, along with chemical (Simpson, 1984) and biological (Haury, 1984) measurements, demonstrates that the surface water in the center of A is coastal in origin. IR images for October 18 (not shown) and October 28 (Fig. 2) indicate that cold water became entrained into the core of A and separated from D. In time, the surface water of A became mixed. By January 15 (Fig. 2), the contrast (temperature gradient) between A and surrounding waters was substantially reduced. The February 21 image (Fig. 2) shows that a dramatic distortion of the pattern occurred after January 15.

The front E seen in the southern half of each image in Figs. 1 and 2 moved to the southwest at about 2-3 cm s\(^{-1}\) over the six-month period. This resulted in a separation between the pattern A and the front E. The strength of the front E exceeded .3°C km\(^{-1}\) at many locations throughout November and December, but was weaker (<.15°C km\(^{-1}\)) in late February. The ship data indicated a salt change of 0.5 in less than 20 km across E. Saltier water was associated with warm temperatures. Historical data (Wyllie, 1966; Evans, 1971) show that a thermohaline front is a common feature in this region, although its precise geographical position is somewhat variable (+300 km).

Figure 3 shows the IRBT data within the square region outlined in Fig. 1 for January 5th and 15th. The CTD stations taken in January are marked with crosses and the position of the ship on January 15 is pointed out. The general vicinity of the rotation axis of the subsurface eddy determined by Simpson, Dickey, and Koblinsky (1984) is shown with an 'X' near the intersection of the transects. The warm IRBT pattern of A is over the rotational axis of the subsurface eddy. The two images show the marked change of small-scale (10-20 km) IRBT features over a 10-day period. The large-scale (100-200 km) pattern, however, did not change. The warm band of IRBT along the boundary of A was found on both days, but was more intact on January 5. The front E has moved south by January 15 and was obscured by clouds.

3.2. Surface Comparison

If the atmosphere were perfectly transparent to radiation in the 11 \(\mu\)m band, the radiometer brightness temperature (IRBT) would be nearly identical to the temperature of the upper 10 \(\mu\)m of the sea or the true SST (Mitnik, 1979). The atmosphere, however, is not completely transparent, and the sea surface radio brightness in both the 3.7 and 11 \(\mu\)m bands is reduced in a nearly linear fashion by a few degrees before reaching the satellite (Deschamps and Philp, 1980). Furthermore, \textit{in situ} SST is rarely measured in the upper 10 \(\mu\)m; usually it is obtained at 1 m or more below the surface. The 1 m SST is generally warmer than the 10 \(\mu\)m SST by a few tenths of a degree because of the heat transfer at the air-sea interface, especially under low wind-speed conditions (Simpson and Paulson, 1980; Paulson and Simpson, 1981). Therefore, the IRBT measured
by the satellite is less than the true SST. We hypothesize, however, that spatial changes of IRBT exceeding a few tenths of a degree over tens of kilometers should correspond with similar spatial changes in the $1\text{ m }\text{in situ}$ SST because atmospherically induced spatial variability is negligible at these scales under dry air conditions.

Fig. 3. The NOAA-6 AVHRR 11 $\mu\text{m}$ band infrared images for January 5th and 15th, 1981. The location of CTD stations is identified with small crosses on each image. The position of R.V. New Horizon on January 15th 0400 GMT is shown. The 'X' near the intersection of the CTD transects represents the general vicinity of the vertical axis of the subsurface eddy.
To check this hypothesis, we compared the IRBT measurements on January 5 and 15 to the 3 m in situ bottle temperature measured at each CTD station of the N/S transect taken between January 9 and 11 (Fig. 4). As hypothesized, the mesoscale variations of the measurements are highly correlated, even though the IRBTs were 2 to 3°C cooler than in situ SSTs. A major source of error in this comparison is caused by temporal differences in the two sampling methods. This is especially noticeable in the southern end of the transect, where the front E advects to the southwest changing its position in each measurement.

3.3. The Surface Layer

The CTD profiles (Fig. 2 in Simpson, Dickey, and Koblinsky, 1984) demonstrated the presence of an anticyclonic eddy beneath the IRBT pattern A (Figs. 1, 2, and 3). The vertical axis of rotation of the subsurface eddy coincided at the surface with the approximate geometric center of this pattern (Fig. 3). The diameter of the eddy was about 160 km at 700 m, whereas the diameter for the associated IRBT pattern on January 5th and 15th was 220 km. The eddy did not extend to the surface. Instead, a three-layer structure was found. A warm subsurface eddy extending from at least 1450 m to 250 m dominated the vertical structure (Simpson, Dickey, and Koblinsky, 1984). Directly above the eddy was a cold-core region extending from 200 m to 75 m. A surface layer from 75 m to the surface was separated from the cold-core region by a strong vertical gradient. Figures 5a and 5b compare the horizontal surface view seen by the satellite on January 5 with vertical CTD sections of temperature (T), salinity (S), and density (σt) to 100 m. These sections have been spatially aligned with the satellite data along the axis of the ship transects to show the vertical extent of the surface pattern. It is evident in Figs. 5a and 5b that the satellite observations reflect the horizontal distribution of temperature throughout the surface layer.
Fig. 5a. A comparison of the horizontal view of surface brightness temperatures observed by the NOAA-6 AVHRR 11 μm band on January 5th, 1981 (see Fig. 3) and the vertical section of T, S, and $\sigma_\theta$ to 100 m contoured from the CTD profiles along the north to south section.
Fig. 5b. A comparison of the horizontal axis of surface brightness temperatures observed by the NOAA-6 AVHRR 11μm band on January 5th, 1981 (see Fig. 3) and the vertical sections of T, S, and $\sigma_\theta$ to 100 m contoured from the CTD profiles along the west to east section.
Within the surface layer an annular region of light ($\sigma_\theta < 24.5$ kg m$^{-3}$), warm, and salty water is found in all but the eastern leg of the hydrographic survey, approximately 75 km from the axis of the eddy. The density of this region increases from the western to the northern transect. The band of warm IRBT around A is associated with this annulus of lighter water. In the center of the eddy and north of the eddy, denser, cooler and less saline water was found. The biological (Haury, 1984) and chemical (Simpson, 1984) data confirm that warm, salty water originating to the south of A and E is advected in an anticyclonic sense around a pool of fresh, cool and dense coastal water. In particular, the biological data (Haury, 1984) show that some southern, warm water surface living species of zooplankton are more abundant in the annulus of warm water in A, whereas species common to the California Current were more abundant in the cooler water in the center of A and outside of A to the north and east.

Across the front E, the temperature changes almost 1 degree centigrade and the salinity changes more than 0.3. There is, however, very little change in density. The temperature and salinity compensation across this front is common to this region (Evans, 1971).

3.4. Surface Kinematics

Surface velocities can be estimated with satellite IR data. The displacement in time of small-scale (10-20 km) features is measured from a sequence of images similar to the way winds are estimated from cloud movements. These features are apparently moved about by the large scale advection. They remain identifiable for only a few days, perhaps as a result of local turbulence and/or vertical heat exchange. Therefore, image sequences must be closely spaced in time. The principle error, approximately 5 cm s$^{-1}$, in estimates of velocity results from the difficulty in determining the position of the feature in the second image. For two images separated by 12 hours in time, an error in relative position of $\pm 2.2$ km is equal to a maximum velocity error of $\pm 5$ cm s$^{-1}$. Errors associated with the difference in the latitude/longitude mappings of each image caused by small variations in the spacecraft attitude result in a bias error which is removed by aligning land features. The frequency of large and cloudy atmospheric systems moving through an area limits the days when sequences of clear ocean images are available. During the six months of our study, we were able to collect only two sequences of two consecutive images that were sufficiently cloud free for detailed velocity analysis.

The surface velocity field derived from IR images separated by 12 hours on November 27, 1980 is shown in Fig. 6. The dashed lines outline the position of the major fronts between A and D or E. Velocities as high as 50 cm s$^{-1}$ were found. The maximum velocities were associated with the southward flow along the eastern boundary of A. Unlike the 0/1450 db geostrophic flow in January, northward flow along the western boundary was not present in November. Eddy-driven southward flow could be enhanced and northward flow diminished at the surface by the southward advection of the California Current. The California Current typically ranges from 5 to 25 cm s$^{-1}$. The eddy-driven surface flow could also be masked by wind-driven surface currents; this problem is discussed further below.

3.5. Surface Fronts

A simple centered difference scheme was used on the satellite 11 $\mu$m band IRBT images to measure thermal gradient. A difference width of 4.4 km was determined to be the shortest distance that could be used without contaminating the calculation with the high wavenumber instrument noise that is common in the data. Prior to differencing, a two-dimensional running average was applied to the data to smooth out variations that occurred over distances of less than 4.4 km. The amplitude of the IRBT gradient is dependent on the atmospheric transmittance (Deschamps, Frouin, and Wald, 1981), as well as the SST gradient. The former effect, however, was negligible.
Fig. 6. The surface velocity field for November 27, 1980 as determined from two NOAA-6 AVHRR 11 μm band images separated by 12 hours. The data are from within the box outlined in Fig. 1. The dashed line traces the frontal gradient of E and the exterior of A (compare with the image for November 27, 1980 in Fig. 2). The letters A, D, and E reflect the location of features identified in Fig. 1.

because of the small dynamic range of the temperatures considered here and because of the dry air mass prevalent in most of the clear images that were analyzed. Therefore, the IRBT gradient closely approximates the SST gradient (see Fig. 4).

Figure 7 shows the front on the outer boundary of A between the warm water of A to the west and cooler coastal water D to the east for several different days over a three-month period. The IRBT patterns for these days are shown in Figs. 1, 2, and 3. Only the expression of this front in the northeast quadrant (220 × 220 km) of the box outlined in Fig. 1 is shown in Fig. 7. Almost all of A is seen on November 27, whereas only a part of A can be seen on January 15 because of its southward movement. There is an indication in Fig. 7 that the southward movement of A may have increased in early January. Similar to calculations by Van Woert (1982) for the North Pacific subtropical front, thermal fronts were found to be less than 10 km in width. The largest gradients, nearly 0.4°C km⁻¹, are almost 100 times greater than the large scale mean gradient shown by Robinson (1976). The strongest gradients do not extend more than 30 km in the along-front direction. In time, the maximum gradient of the frontal pattern weakens and becomes less continuous.
Fig. 7. The position and strength of the frontal gradient around the rim of feature A for various days between November, 1980 and January, 1981 (compare with images for same days shown in Figs. 1, 2 and 3). The area shown comes from the northeast quadrant (220 x 220 km) of the box outlined in Fig. 1.

A series of frontal maps similar to those discussed above, but with only a 12-hour interval between each of them, is shown in Fig. 8. These images were obtained on November 27 and 28, 1980. The images demonstrate that patches of intense gradient within the front are advected with the anticyclonic circulation of A. The most intense gradients have short length scales (20 km) and short time scales (2-3 days) relative to the mesoscale pattern.

4. DISCUSSION

4.1. Dynamics

The IRBT pattern in our data aligns with the surface layer density field (Figs. 5a and 5b) and the rotational axis of the subsurface eddy (Fig. 3). The surface flow in November is anticyclonic. Therefore, we conclude that surface feature A results from a geostrophic adjustment of the surface layer to the perturbation in the dynamic height field associated with the subsurface warm-core eddy.
Our observations confirm the model of Nelepo, Kufarkov, and Kosnyrev (1978). They modeled the effect of a subsurface eddy on a vertically-homogeneous surface layer with a mean horizontal temperature gradient. They predict that within the surface layer bands of fluid will be horizontally advected around an isothermal interior (e.g., Figs. 5a and 5b). They also predict that the strongest frontal gradients will: (1) occur along the exterior rim of the eddy (e.g., Fig. 7), (2) be more than 25 times the mean gradient, and (3) be less than 10 km wide. In their model, the dominant source of energy is the subsurface eddy. They find that moderate winds (<1000 cm s\(^{-1}\)), vertical heat flux, and translation of the eddy do not change the basic surface pattern described above. These predictions are confirmed by our observations and those of Saunders (1971) for Gulf Stream rings, Voorhis, Schroeder and Leetmaa (1976) for MODE eddies, and Nilsson and Cresswell (1981) for East Australian Current eddies.

Along the outer boundary and within feature A are small, growing disturbances of 10-30 km which extend radially outward. Voorhis and Bruce (1982) found small-scale disturbances around an eddy-like surface feature in the Saragasso Sea. They suggest that such disturbances could result from shear instabilities similar to those found in laboratory rotating vortices (Griffiths and Linden, 1981). In particular, they showed that the wavenumber for maximum growth rate for their eddy, derived from Griffiths and Linden (1981), was consistent with the measurements. If we follow this analysis and consider the surface layer of the CCS eddy A to be a rotating vortex the estimated wavenumber for maximum growth rate is 5, which is much less than the observed 10-15. Maximum growth rates with wavenumbers of 10 to 15 are more frequently found in systems which are ageostrophically unstable (Griffiths, Killworth, and Stern, 1982).

Small-scale (10-20 km) disturbances are also found along the front E. These features grow in size, have a time scale of about 5 to 10 days, and cause a distortion and intensification of the front. Such disturbances are common in frontal observations and have been modeled with some success by James (1981). As the models require density driven flows, our observations may be somewhat
surprising because they show similar instability in a front which is temperature-salinity compensated and has a very weak density signature (Fig. 5a).

The small-scale disturbances in both the eddy A and the front E are important in frontogenesis. Figs. 7 and 8 show that the most intense fronts in both A and E are associated with these small-scale disturbances. This mechanism of frontogenesis is confirmed in the models of James (1981). Fig. 8 also indicates that the phase speed of the small-scale disturbances in A is much less than the geostrophic flow induced by the subsurface eddy.

We have described the dramatic distortion of feature A in Fig. 1 between January and February (Fig. 2). Bernstein, Breaker, and Whitmer (1977) found a similar distortion in 1975-76 imagery for the same region and time of year. They attributed the existence of the eddy to baroclinic instability of the California Current. They found that the growth rate in a simple instability model was comparable to the distortion and onshore movement of the observed mesoscale feature. In 1981, eddy A in Fig. 1 was not the result of a local instability of the California Current but an isolated vortex of coastal origin (Simpson, Dickey, and Koblinsky, 1984). Such a vortex could become unstable and change its shape (e.g., Griffiths and Linden, 1981), but, because of the strong winter wind systems common to this area, it may also be possible for atmospheric forcing to decouple and distort the surface layer. It is important to note that if the latter mechanism is operative then the satellite imagery cannot be relied upon to follow the temporal development of the subsurface eddy in the storm season.

In order to compare the strength of the wind driven surface currents with the surface gradient currents in January and February, we obtained the geostrophic wind field for this area from FNOC (Fleet Numerical Ocean Center) and computed the surface wind stress ($\tau$) following Garratt (1977). At mid-latitudes the FNOC wind field approximates the actual wind field quite well in regions with moderate to heavy merchant shipping (Freihe and Pazan, 1978). A 10-day mean surface wind drift current was computed with an Ekman model (Gonella, 1971) for the period September, 1980 to February, 1981. We assumed a constant eddy viscosity of 100 cm$^2$ s$^{-1}$ and a surface layer depth of 75 m. The cube of the surface friction velocity $u_f^3 = (\tau/\rho)^{1.5}$, a common measure of wind forcing, and the estimated surface wind drift velocity at 33°N, 125°W (the closest location in the FNOC grid) are shown in Fig. 9. From September through mid-January the winds are light and the wind drift is southward. On November 27 the spatial mean velocity derived from the satellite observations shown in Fig. 6 is consistent with the estimated wind drift. We conclude that the wind drift on November 27 suppresses the northward flow and enhances the southward flow at the surface in feature A. Between mid-January and the end of February two storm systems moved through the study region. Both systems had winds that were strong enough to generate drift currents of the same magnitude or larger than the estimated surface gradient currents in January. These storm driven currents could distort the surface signature of the subsurface eddy. The strong eastward drift current caused by the first storm is consistent with the eastward distortion of feature A on 21 February (Fig. 2). This result suggests that caution must be used in interpreting the winter growth and distortion of mesoscale IRBT patterns in the CCS.

4.2. **Sea Surface Temperature Anomalies**

Barnett (1981) has shown that SST anomalies in the CCS are significantly correlated with seasonally averaged air temperatures over North America and may play an important role in climate problems. We have shown that subsurface eddies within the CCS transfer the large scale variability of the mean SST field into synoptic scales. We consider here the role of mesoscale variability in the SST anomalies of the CCS.

Finding mesoscale anomalies in the observed SST field is difficult because of sparse observations. Operational maps of SST are routinely produced by averaging, smoothing, and contouring the in situ surface observations from the merchant fleet. One such map is produced by the U.S.
Fig. 9. Time series of the cube of the surface friction velocity $u_2^3 = (\tau/\rho)^{1.5}$ in the upper panel and estimates of the surface wind drift velocity in the lower panel. Both data sets were derived from the FNOC (Fleet Numerical Ocean Center) quasi-geostrophic wind field at 33°N, 125°W, and averaged over 10 days except for the arrow in the lower panel below the dashed line. The latter data point is the spatial mean surface velocity of the ocean derived from the image analysis shown in Fig. 6.

The highest resolution maps of NMFS are temporally averaged biweekly and spatially averaged over one-degree squares. While the temporal average is adequate, the spatial average (because of the sparse observations) tends to smooth out the mesoscale features. The extent to which NMFS SST data is useful in studying mesoscale anomalies was examined with data from December 1-15, 1980. These data were found to be consistent with the satellite observation shown in Fig. 1. The 19-year (1960-1978) mean field for December 1-15 was then subtracted from the NMFS SST field for December 1-15, 1980. Figure 10 shows the mean bi-weekly SST field, the average number of observations in each one-
degree square for December 1-15, the December 1-15, 1980 SST field, and the December 1-15, 1980 SST anomaly. The anomaly map and the 1980 SST field are visually consistent with the IRBT data in Fig. 1. A 2°C warm anomaly is associated with feature A, and a 1°C warm anomaly is aligned with features B and C. A 1°C cold anomaly is associated with water to the south of E. This latter cold anomaly may represent the transport of heat from the normally warmer water south of E to the normally cooler water at A, B, and C. A surface feature 200 km in diameter and 50 m deep which is 2°C warmer than previous years represents a heat content anomaly of about $6.5 \times 10^{16}$ joules.

The monthly variation of mesoscale anomalies was studied by examining the monthly NMFS anomaly maps for September, 1980 through March, 1981 (Fig. 11). The 1949-1968 mean monthly SST fields were used to construct these charts (F. Miller, personal communication). From September through November, the anomaly field is predominantly negative. Starting in December, the anomalies become more positive and there is a strong anomaly from December through February in the region of A. However, the length scales of the monthly anomalies are predominantly 500 to 1000 km. Further understanding of the wavenumber-frequency spectrum of the SST field...
Fig. 11. SST anomaly maps derived from ship observations collected at NMFS for one-month periods. The anomaly is determined by subtracting a 20-year monthly mean (1948-1967) from the current month. Data have been averaged over one-degree square and one month. All maps align with and cover the area shown in Figs. 1 and 10. Units are degrees Celsius.

must await the development of higher resolution SST measurement systems (e.g., Bernstein, 1982b).

5. CONCLUSIONS

Mesoscale variations in the surface brightness temperature field were observed in the California Current (Fig. 1). These variations are most evident in IR imagery from the fall through the winter months. Measurements taken from a ship showed that beneath the SST pattern near 32°N, 124°W, there was a three-layer structure. The surface pattern reflects the horizontal variability of the 75 m deep surface layer. The center of the surface brightness temperature pattern geometrically aligns with the rotational axis of a subsurface warm core eddy. In the fall and early winter the dominant source of kinetic energy in the surface layer is the geostrophic adjustment induced by the subsurface eddy. The mesoscale surface pattern advects to the south at 1 cm s⁻¹ and rotates anticyclonically up to 50 cm s⁻¹. The strongest thermal fronts associated with the pattern are 0.4°C km⁻¹, have scales of 10-30 km, and may result from ageostrophic processes. As winter progresses and storm systems increase in strength, wind driven surface currents achieve parity with the geostrophic surface flow. This may result in distortion of the SST pattern as observed from space and a “decoupling” of the surface layer from the subsurface eddy structure. The eddy field
breaks up the large scale SST variability into shorter length scales. Significant SST anomalies are associated with the mesoscale SST field because of the space-time variability of the eddies. Therefore, the variations in the reservoir of heat available to the ocean offshore of California can be controlled by the CCS and its eddy structures.

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