Observation of Mesoscale Ocean Features in the Northeast Pacific Using Geosat Radar Altimetry Data

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In this paper, Geosat altimetry data from the northeast Pacific are examined. The data are (1) binned data on 1° latitude by 2° longitude blocks and (2) collinear track data. The binned data cover the region from 20°N to 60°N from the North American coast westward to 150°W and span from December 15, 1986, to June 1, 1989. To examine the data for evidence of westward propagation of sea level anomalies, three zonal transects (45°N, 49°N, and 57°N) are made. Complex empirical orthogonal function analysis of the data transects at these latitudes reveals phase speed and amplitude information. The first eigenmode at each latitude has a very strong annual signal. Two-dimensional spectral analysis of the first eigenmode at 45°N and 57°N shows westward propagation of sea level anomalies at 45°N only. The propagating features are annual in period and of order 1000 km in length. At 57°N, westward propagation is evident (as at 45°N), but there is also an eastward component. To examine the variability of the coastal sea level signal, collinear track data from three ascending tracks crossing the Gulf of Alaska are analyzed revealing mesoscale features of the order of 100 km that parallel the Canadian and Alaskan coast. The most notable feature is the Sitka eddy. Westward propagation of several of these features is observed. The results from the analysis of both types of Geosat data are compared with other observational data.

Introduction

1.1. Motivation and Objectives

The principle focus of this paper is the use of altimetry data to understand the mesoscale circulation in the Gulf of Alaska. Figure 1 shows the Gulf of Alaska region and the major oceanographic features including the California Current, the Alaska Current, the Alaskan Gyre, the Subarctic Current, the West Wind Drift, and a portion of the North Pacific Gyre. A drawback to past studies of the Gulf of Alaska is that observations in this region have been limited. Only recently have comparatively thorough studies been completed [e.g., Royer, 1981; Tabata, 1982; Mysak, 1986]. Theoretical studies [e.g., Cummins and Mysak, 1988] have been hampered by a poor observational data base as well. In recent years, new technologies such as deep moored current meters, satellite-tracked buoys [Tabata, 1982], and remote sensing have increased this observational data base.

One of the newest observational tools for determining the ocean's circulation is radar altimetry such as Geosat. The purpose of this study is to analyze Geosat sea surface height (SSH) data from the northeast Pacific region for specific mesoscale ocean features. Comparison of Geosat data will be made to other observational data and also to data produced by numerical models.

This paper is divided into five sections. The remainder of this section presents a brief overview of the regional circulation. Section 2 presents the two types of Geosat data and the analysis techniques used in this paper. The first Geosat data set is SSH binned in 1° latitude by 2° longitude boxes and then analyzed using complex empirical orthogonal functions (CEOFs) and two-dimensional spectral analysis. The second type of data are from several ascending Geosat tracks and analyzed as time series. The results of the Geosat data analysis are presented in section 3. Also presented are observations of mesoscale eddies and fluctuations in the mean surface circulation. Section 3 also includes comparison of Geosat data to other observational data. Section 4 contains comparisons of Geosat results with the sea surface signal produced by numerical models. In section 5 we present our conclusions. In this paper we demonstrate evidence of westward propagation of sea level anomalies, show that Geosat SSH data are also in good agreement with model solutions of the seasonal variability of the sea level off the Oregon and Washington coasts, and show that the Geosat data are in good agreement with drifter data.

1.2. Overview

The mean atmospheric circulation in the Gulf of Alaska is dominated in the winter by the intense low-pressure Aleutian low. To the south of this region is the Subtropical high which is centered over an area just west of southern California. In the spring, the Aleutian low begins to weaken and shift westward, and the Subtropical high strengthens and expands [Emery and Hamilton, 1985]. By midsummer, the Subtrop-
The high-pressure system dominates the entire northeast Pacific. The Aleutian low is outside the region. In the fall, high pressure weakens, and the atmosphere begins to return to wintertime conditions. Figures 2a and 2b show the winter and summer mean sea level atmospheric pressure for the Northeast Pacific.

The mean ocean surface circulation of the northeast Pacific resembles (in a geostrophic sense) the atmospheric circulation. In Figure 1 there is cyclonic flow to the north of 45°N and a general anticyclonic flow to the south of that latitude. The mean surface circulation around 45°N is the combination of the Subarctic Current, the West Wind Drift, and the North Pacific Current. Near the North American continental shelf this flow bifurcates (due to the wind stress curl) forming the equatorward flowing California Current and the poleward flowing Alaska Current.

2. GEOSAT DATA AND ANALYSIS METHODS

2.1. GEOSAT DATA

The data for this study are from the Geosat Exact Repeat Mission (ERM). The orbit for this part of the Geosat mission was similar to that of Seasat, flown in 1978. The ERM was designed to repeat ground tracks within 1 km after 244 revolutions of Earth with a frequency of 17 days. The track spacing at the equator was approximately 150 km. This track spacing and repeat cycle gave an impressive spatial and temporal resolution over the entire globe [Cheney et al., 1987].

Kelly et al. [1991, Figure 1] diagram the essential measurements for determining SSH. Calman [1987] gives a simple yet complete explanation for determining SSH from satellite radar altimetry. In this paper we are interested in observing mesoscale ocean features in the northeast Pacific. Therefore only the variability of the SSH (as measured by Geosat) about a mean sea surface height is used. The mean sea surface was taken as the 1-year mean for the year August 1987 to August 1988. This definition of the mean was predetermined in the data set received from National Oceanic and Atmospheric Administration (NOAA) National Ocean Survey (NOS).

Corrections and adjustments were made to the raw data to produce the Geophysical Data Records (GDRs). They are explained in full detail by Cheney et al. [1987]. The weakest of all corrections made to raw Geosat data is the one for the wet troposphere effect.

The error induced by the wet troposphere is one that affects the path length of a radar pulse sent from the satellite. The moisture in the atmosphere effectively slows the speed of electromagnetic waves traveling from one point to another. The increased time is interpreted by the altimeter as a depression in the sea surface. In extreme cases it is estimated that this error can contribute as much as 35 cm in magnitude to the SSH. To date, there have been several methods utilized to correct the error caused by the wet troposphere.

The two data sets used in this research have different types of water vapor corrections. The first data set uses water vapor corrections derived from the U.S. Navy's Fleet Numerical Oceanographic Center (FNOC) water vapor model. This is a global model updated every 12 hours by a large observational data set [Cheney et al., 1987]. The second data set uses water vapor corrections derived from actual water vapor measurements from two satellite instruments: (1) TIROS Operational Vertical Sounder (TOVS) and (2) Special Sensor Microwave/Imager (SSM/I) (L. Miller, personal communication, 1991). The reader is encouraged to see the many articles concerning water vapor corrections which appear in the Geosat special issue of the Journal of Geophysical Research (1990).

The data used for this research are organized in two ways. The first data cover the region bounded by the North American coast west to 150°W and from 20°N to 60°N and have been binned in 1° latitude by 2° longitude boxes. The second data are collinear (along track) samples taken from
individual satellite tracks which cover a portion of the Gulf of Alaska. The data sets cover the entire ERM which lasted from November 1986 to January 1990.

2.1.1. Binned Geosat Data. The binning of Geosat data into the 1° by 2° boxes is accomplished in the following manner. Each altimeter cycle that passes through a given 1° by 2° region contributes a discrete height measurement as a function of time. This measurement is an estimate of an average SSH along the ground track segment that passes through the box. For boxes that have multiple cycles the measurement for each pass is collected with the others to form an average SSH for the entire box. This average SSH is considered to be located at the center of the box. For the ERM each box is sampled by the satellite 2-4 times in each 17-day period [Cheney et al., 1987].

The binned Geosat data were received with the corrections and adjustments listed in Table 1. The water vapor correction for these data is derived from the FNOC water vapor model. There is no inverse barometer correction.

The data have had a 1-year mean (August 1987 to August 1988) removed, and the resulting time series of sea level anomalies is spaced on average 17 days apart. After examining the distribution of SSH in the data, all values greater than 5 standard deviations from the mean were removed. Linear interpolation between the 17-day points in time filled the missing data. The final data set begins December 15, 1986, and ends June 1, 1989. A great deal of data were removed at the end of the mission because of the large data gaps during the last 8 months of the ERM. The data gaps were due in part to tape recorder failure on the satellite and spacecraft attitude variability [Doyle et al., 1990]. A 1-2-1 Hanning window repeated 92 times in time filtered periods

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<th>Correction</th>
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<tr>
<td>1. Solid Earth tide</td>
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<td>2. Fluid tide (ocean tides)</td>
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<td>3. Ionospheric correction</td>
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<td>4. Wet troposphere</td>
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<td>5. Dry troposphere</td>
<td>FNOC barometric pressure model</td>
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<td>6. Geoid</td>
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<td>7. Electromagnetic bias</td>
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<td>8. Orbit error</td>
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From Cheney et al. [1987].
less than 55 day. The frequency response function for this filter was chosen to retain Rossby wave propagation and mesoscale features that persist for 2 months or longer. The data were also smoothed in space with a Hanning window once in each of the latitude and longitude directions removing two-dimensional noise remnant from the binning process.

2.1.2. Collinear track data. The collinear track data set consists of descending and ascending tracks that crossed a portion of the Gulf of Alaska. These data had all the corrections and adjustments as the binned data except that the water vapor correction is derived from the TOVS SSM/I. Also, a better orbit correction has been included as well as an inverse barometer correction (L. Miller, personal communication, 1991). Each groundtrack is repeated every 17 day. Due to the large data gaps in the descending tracks (many cycles were missing altogether), only the ascending tracks are considered for this study.

The data samples along each ground track are spaced, on average, 6.7 km apart. In order to get evenly spaced data along track, a cubic spline with 5-km spacing was applied to each cycle of each track. Missing data were filled by averaging nearest neighbors forward and backward in time.

2.2. Data Analysis Techniques

2.2.1. Complex empirical orthogonal functions. CEOFs have been used successfully on oceanographic and meteorologic data [Shriver et al., 1991; White et al., 1990, 1987; Barnett, 1983; etc.]. The CEOFs extract information about propagating physical features from a space-time data set by Hilbert transforming the original data matrix and creating a new complex data set with the Hilbert transform as the complex part. Eigenmode decomposition is performed on the new data matrix. As this new matrix is Hermitian, its eigenmodes contain phase and amplitude information about the principle components of the original data. White et al. [1987] and Barnett [1983] present a detailed discussion of this technique.

2.2.2. Two-dimensional spectral analysis. White et al. [1990] used CEOF analysis on Geosat data in the California Current region and found evidence of Rossby wave propagation in the Geosat data. There was very little quantitative information about the phase speed and direction of the propagating features. Therefore they used two-dimensional spectral analysis, choosing east-west wavenumber and frequency as the independent variables. In this study we do the same on the first mode results of the CEOF analysis.

3 Results

3.1. Analysis of Binned Geosat Data

One of the most prominent features in the SSH is the annual signal. Each winter in the 3-year data set shows the same general pattern as any other, and the same is true for the summer pattern. Westward propagation of sea level anomalies is evident throughout the data set, suggesting Rossby wave dynamics. This is especially true in the southern portion of the data domain. Propagation of anomalies is also evident along the coast of North America, although the evidence is very segmented and sporadic. Poleward Kelvin wave propagation in this region has been shown to exist [Pares-Sierra and O'Brien, 1989] but cannot be verified by Geosat because it does not measure sea level adjacent to the coast.

Seasonal upwelling and downwelling are both evident along the coasts of California, Oregon, and Washington. In this area, close to the coast, there is also evidence of the winter formation and summer disappearance of the Davidson current. Propagation of sea level anomalies, parallel to the coast, is especially evident in the Gulf of Alaska. This is strong evidence of the seasonal and interannual intensification and relaxation of the Alaska Current and the Alaska Stream.

3.2. CEOF Analysis of Binned Geosat Data

CEOFs are used to isolate westward propagating features in the region from 40°N northward into the Gulf of Alaska. One of the objectives is to compare results from this analysis of Geosat data with data produced by reduced gravity, baroclinic models of the same region.

At 45°N, a space-time data matrix (from 125°W to 137°W) of the 900-day Geosat data set was constructed. A spatial mean along the transect was removed, and the data were averaged into a composite one year data set (Figure 3). There is westward propagation of sea level anomalies at this latitude and a strong annual signal.

The first CEOF mode accounts for 88% of the variance of SSH. We believe the remaining variance is noise, therefore, only the first mode will be discussed. In the temporal pattern (Figures 4a and 4b) of the first eigenmode, the annual signal is clear. Note the negative phase shift between the real and imaginary components of the temporal pattern. The real and imaginary components of the spatial pattern (Figures 4c and 4d) also show a phase shift but otherwise show little information. A more interesting and informative representation of the spatial information is the amplitude of the spatial
1.5

The maximum variability of SSH occurs near 127°W, approximately 235 km from the coast of North America. A mean phase speed of \(-4.0 \text{ cm s}^{-1}\) was calculated. The negative value implies westward propagation along 45°N. As this speed is higher than the theoretical Rossby wave speed for this latitude, we cannot conclude that the westward propagation of sea level anomalies is due to Rossby wave dynamics.

A space-time picture of the first eigenmode (Figure 6) (reconstructed from the temporal and spatial patterns) shows several noticeable features. First, as observed in the amplitude of the spatial pattern, the maximum variability occurs near 127°W. Due to the nature of the binning process, any data recorded within 2° of the coastline (located at approximately 124°W) at 45°N show up as a value at 123°W. For that reason, and because the satellite generally does not acquire the sea surface signal until some distance from the coastline, there is no accurate representation of coastal sea level in Geosat SSH. The second feature observed in the first eigenmode is between 125°W and 131°W. Here, the sea level anomalies are negative in the winter and positive in the summer. This agrees with steric height calculations made by Huyer [1977], who showed that the steric height at 300 km (about 128°W) offshore is least in late winter and greatest in summer. This is a result of the seasonal heating and cooling cycle. The associated sloping pattern (less negative anomalies in winter and less positive in summer) toward the coast is an indication of the formation of the Davidson current in the winter and its disappearance in summer. The westward propagation of sea level anomalies is strongest from 125°W out to 132°W. This is in agreement with the numerical study of Pares-Sierra and O'Brien [1989], which will be discussed in more detail in section 4. West of 132°W the variability of sea level is very low in amplitude, again in agreement with Pares-Sierra and O'Brien [1989]. For reference, the calculated phase speed from the CEOF analysis is shown superimposed on the contours of SSH. Some features appear to travel faster than the calculated phase speed, while others travel slower.

Progressing northward into the Gulf of Alaska, the same analysis was performed at 49°N. The data prior to processing are shown in Figure 7. A difference between the data at this latitude and that at 45°N is that the amplitude of the SSH variability is smaller. The first eigenmode of this spatial series accounted for 66% of the variance. The amplitude of
the spatial variability is greatest near the coast at this latitude than at 45°N. This is significant because this latitude is in the bifurcation region of the West Wind Drift and Subarctic Current. The sea surface signal propagating from the coast at this latitude is most likely suppressed by the onshore transport from these two currents. The average phase speed calculated is 5.3 cm s\(^{-1}\) westward. Combining the temporal and spatial patterns to form the reconstructed first mode reveals (Figure 8), an annual signal with sea level close to the coast being negative in the winter and positive in summer as at 45°N. The features propagating from the coast travel to the west at a somewhat faster speed than the average.

Approximately 5° from the coast the phase speed slows considerably, but the amplitude of the SSH signal is so small that these measurements are suspect. After about 2° more to the west the slope of the traveling features is unclear but eventually suggests eastward traveling waves. At the western edge of the domain the contours of SSH slope up and to the right. As shown by Kirwan et al. [1978], the mean flow of satellite tracked buoys at this latitude is directly to the east from about 145°W to about 136°W, precisely where we see evidence of eastward (positive slope) propagation.

The processed SSH data at 57°N are displayed in Figure 9. The CEOF yields similar results to those at 45°N and 45°N. However, the amplitude of the spatial pattern reveals a very different structure from those at more southern latitudes (Figure 10). As before, the magnitude of the sea level variability is maximum nearshore and decreases rapidly going offshore. About 6°-7° offshore, the variability increases to a second maximum. This indicates a second energetic region away from the coastal domain. The phase speed of this mode (which accounts for 67% of the total variance) is \(-2.47 \text{ cm s}^{-1}\). The fact that the phase speed at this latitude is less than at 45°N is consistent with the reduction in phase speed of Rossby waves with latitude, even though 49°N is beyond the critical latitude for the annual wave.

As noted in the spatial pattern, the increased sea level variability well offshore is evident in Figure 11. At the western edge of the domain there is evidence of eastward propagation of the SSH signal. The Alaska Stream is quite developed here and flows to the southwest along the Alaskan coastline. Variability in the Alaskan Stream (such as mean- der offshore or broadening of the stream) would be observed as eastward propagation of SSH.

3.3. Two-Dimensional Spectral Analysis of Geosat Data

To determine if the westward propagating features were related to Rossby wave dynamics, wavenumber/frequency spectra were calculated for the first eigenmodes of Geosat...
data at 45°N and 57°N. The method utilized is similar to that of White et al. [1990], except that they performed the analysis on Geosat data that had not been examined using EOFs, and they used data from the California Current region south of 40°N.

The maximum power of the wavenumber/frequency spectra for the first eigenmode of the Geosat data at 45°N is centered at a period of 357 days and a wavelength of 936 km (Figure 12). Using a value of \( c = 2.5 \text{ m s}^{-1} \) (where \( c \) is the gravity wave speed), the dispersion relation falls well below the maximum spectral density. The critical latitude for annual Rossby waves using \( c = 2.5 \text{ m s}^{-1} \) is between 35°N and 40°N. By increasing the value of \( c \) (by almost a factor of 2) the linear dispersion curve coincides more closely to the maximum spectral density. This indicates that these annual waves travel considerably faster than the speed of the expected Rossby wave speed for this region. These results are similar to those found by White and Tabata [1987] along line P in the northeast Pacific when examining constant density surfaces. They conjectured, because of the proximity to the critical latitude for annual Rossby waves, that these waves are merely forced by the annual cycle of the wind. Another explanation for the discrepancy between the theoretical Rossby dispersion curve and the observations is that nonlinear processes are manifested in the SSH signal as observed by Geosat. In essence, there is evidence of westward propagation of sea level anomalies at this latitude but conclusive evidence of Rossby wave propagation is not supported.

The first eigenmode of the Geosat data at 57°N is similar to that at 45°N (Figure 13). However, the wavelength of the annual wave is shorter (≈725 km), with considerably more energy in the positive wavenumber quadrant than at 45°N. For the linear Rossby dispersion relation to intersect the maximum spectral density, a value of \( c = 5.2 \text{ m s}^{-1} \) is required. It is interesting to note that Mysak [1986] and Tabata [1982] found evidence of what they described as Rossby wave propagation of mesoscale features using other types of oceanographic data in this region. We will show, as did Gower [1989a, b], that by examining only ascending Geosat tracks, several mesoscale features propagate to the west at very near the theoretical Rossby wave speed for this latitude.

3.4. Analysis of Collinear Geosat Data in the Gulf of Alaska

In an effort to examine the variability of sea level on shorter time and smaller spatial scales than those studied using the binned Geosat data, SSH data from independent ascending tracks were analyzed. Ascending satellite tracks traverse the region of study from the southeast to northwest. Three ascending tracks have been examined. Using the numbering system described by Gower [1989a, b], we examined ascending tracks 171–173 (see Figure 14). One of the primary motivations for this analysis is to compare the near-coast, sea level signal as measured by Geosat to the near-coast signal of a reduced gravity baroclinic model discussed in section 4.

The data for the three tracks are shown in Figures 15–17. Track 173, which is the closest track to the coast, has, as observed by Gower [1989a, b], sea level anomalies with spatial scales of the order of 100 to 200 km and temporal scales of a few weeks to several months. The northwestern end of the satellite track appears to be the most energetic in the sense that the variability of sea level is more intense here. Two very noticeable positive anomalies (first observed by Gower [1989a]) are located near 57°N and 138°W. This is the location of the Sitka eddy first described by Tabata [1982].

The first time that the eddy is seen along this track is in the late winter, early spring of 1987 (feature A in Figures 15–17). It persists well into the summer of that year. The along-track
Fig. 12. The contour of the wavenumber/frequency spectra of the first eigenmode of SSH at 45°N. Only the negative wavenumber, positive frequency quadrant is plotted. The theoretical, linear Rossby dispersion relation is superimposed on the contour. (a) A value of $c = 2.5 \, \text{m s}^{-1}$ is plotted, and (b) a value of $c = 4.7 \, \text{m s}^{-1}$ is plotted. Because there is very little energy in the positive wavenumber quadrant at this latitude, it is not shown.

The size of the eddy at its maximum diameter as seen by track 173 is of the order of 200–300 km. This occurs in May 1987. The eddy forms again in February 1988 and persists until at least June 1988 (feature B). Gower [1989a] uses the argument that the negative sea level anomaly is due to reversal of the eddy. If true, then the eddy actually persists much longer than suggested by these data. The most important new observation, as compared to that of Gower [1989a, b], is that the Sitka eddy shows no evidence of forming in the spring of 1989. This agrees with Tabata [1982], who found that the eddy does not form every year.

There is a great deal of variability to the north of the Sitka eddy at the annual period. Some of the positive and negative anomalies are very intense. This segment of the satellite track crosses the transition area between the Alaska Current and the Alaska Stream. Thomson et al. [1990] show that this transition is not smooth in nature and the region is highly variable. The anomalies observed in this segment of track 173 are also indicative of the seasonal intensification and relaxation of the coastal currents in the Gulf of Alaska.

To the south of the Sitka eddy, off the coast of the Queen Charlotte Islands, there is evidence of another intense eddy-like feature. This feature forms in January 1987 and by summer of the same year the signal is negative. This feature is labeled C in Figures 15–17. There is no evidence that this feature recurs in subsequent years.

It is interesting to examine tracks 171–173 together to observe westward propagating sea level anomalies that are evident from track to track. The positions along each track of features A, B, and C are marked in Figure 14. The SSH anomalies propagate to the southwest. The positions are plotted along track by subjectively marking the position when the satellite observed the maximum positive amplitude of each feature. By doing this for each of the features on
each of the three tracks, an estimate of the speed of propagation can be made. From track 173 to track 171, feature A travels at an average speed of 1.74 cm s$^{-1}$, B at 3.53 cm s$^{-1}$, and C at 3.9 cm s$^{-1}$. This is consistent with the calculations made by Gower [1989a, b] and the results of Cummins and Mysak [1988]. The northern segments of the other ascending tracks show as much variability as track 173.

4. COMPARISON OF GEOSAT DATA TO MODEL RESULTS

One of the objectives of this paper was to compare Geosat SSH to the sea level signal produced by numerical models. We use 1-1/2 layer, reduced gravity, baroclinic models with bottom topography for this comparison. For complete descriptions of the models see Pares-Sierra and O’Brien [1989], Johnson and O’Brien [1990], and Heim et al. [1992]. There are two models that overlap the region of the Geosat study. The first is a model of the California Current region that extends to 50°N. It is forced by climatologic monthly mean wind stress at the local scale and also has a remote forcing component from coastally trapped Kelvin waves which are produced by a model of the equatorial region of the Pacific Ocean. The second model overlaps the first by 5° beginning at 45°N and extends along the coast of Alaska. It is forced only by climatologic monthly mean wind stress.

We begin by comparing the sea level signal as seen by Geosat at 45°N to the upper layer thickness (ULT) produced by the California Current region model. Figure 7 of Pares-Sierra and O’Brien [1989] shows the long-term monthly ULT anomaly derived from the local forcing mechanisms of their model. Referring to the region from 40°N to 50°N, patterns of ULT and Geosat SSH are similar. Because Geosat does not resolve the sea level signal at the coast, we focus our attention to 126°W and westward.

In winter the sea level is characterized by a negative sea level anomaly (see Figure 6). The maximum negative value
occurs at approximately 127°W and propagates to the west. This is similar to the pattern of model ULT. The negative values of ULT (meaning the ULT is thinner than the mean) are manifested in the sea level signal as negative anomalies. By March of each year, a positive anomaly of sea level starts to propagate into the Geosat domain, and a positive anomaly of ULT propagates from the coast to 126°W. By July the sea level is positive from 126°N all the way to 130°N. Again this is the same pattern that is displayed in the model ULT. By late fall the process begins to repeat. It is observed in both the Geosat data and the model data that the maximum variability occurs around 127°–128°W. Proceeding westward of 132°W in both data sets, the anomalies are very weak and most generally approach the mean sea level and ULT. Paredes-Sierra and O’Brien [1989] attribute this phenomenon to the existence of the theoretical critical latitude for an annual wave (~35°N–40°N). Their conclusion (supported here by Geosat observations) is that south of the critical latitude, waves can propagate to the west as free Rossby waves but north of the critical latitude they present a decaying component.

CEO analysis was performed on a subset of model results in the same manner as that done for the Geosat data at 45°N. To prevent any contamination of the results from the choice of boundary conditions of the model, 46°N was chosen as the line of latitude for this analysis. The first eigenmode accounts for 97% of the variance of the sea level produced by this model. The reconstructed space/time matrix of that mode is displayed in Figure 18. The general structure of the contour plot is the same as that of the Geosat data with westward propagation of sea level anomalies being obvious. The calculated mean phase speed is higher in the model than for the Geosat data. Also there is a second region

Fig. 14. Selected ascending tracks used in examining the sea level variability along the Alaska and British Columbia coastlines. The three features A, B, and C are marked to show the westward propagation of sea level anomalies and are discussed in the text. The track numbers are as described by Gower [1989a, b].

Fig. 15. A time/latitude contour of ascending track 173. This track is the closest complete ascending track to the coastline. The data points are separated by 5 km in space and 17 days in time. The contour interval is 5 cm. Notice that there are several eddylike features in addition to those that are marked. A considerable amount of variability at the northern part of the track is evident at the annual period.

Fig. 16. A time/latitude contour of ascending track 172.

Fig. 17. A time/latitude contour of ascending track 171.
of high variability away from the coast, which is not evident in the Geosat data. A possible explanation is that this latitude is close to the open boundary of the model. Since the model is forced only with local wind stress, external mechanisms do not influence circulation in the model as they do in the SSH data. Performing CEOF analysis at other latitudes did not yield information that was comparable to the Geosat data.

Some of the physical features that are produced by the model and are described by Heim et al. [1992] are evident in the Geosat data of the region. The most prominent of these features is the propagation parallel to the coastline of sea level anomalies. As discussed in section 5, Geosat shows a seasonal signal of the intensity of the coastal jet. This is also a result of the model; however, there does seem to be a phase shift between the model data and the Geosat data. Heim et al. found that this was true when comparing model data to tide gauge data of the same region.

5. Conclusions

We have shown that when the Geosat satellite data over the northeast Pacific Ocean and Gulf of Alaska are examined, evidence of westward propagating features is present in sea level anomalies at 45°N. For this study it was necessary to remove the unresolvable interannual variability in the original 2.5-year record and to reduce the data set to a 1-year composite. As Shriver et al. [1991] have shown in expendable bathythermograph (XBT) and model data at 40°N, the evidence of such propagation diminishes offshore which they attribute to Rossby wave dispersion and the proximity to the critical latitude of the annual Rossby signal. Northward into the Gulf of Alaska, there is no direct evidence of Rossby wave propagation at the annual period. Instead, our results suggest that the westward propagating waves result from local wind forcing. A critical latitude exists for the annual Rossby wave signal which is generated at an eastern boundary. In the Gulf of Alaska, there is evidence that some mesoscale features do propagate with speeds that are close to the theoretical Rossby wave phase speed.

Individual ascending track information along the coast of Alaska and British Columbia was analyzed for the presence of mesoscale features. The analysis revealed the presence and formation of the Sitka eddy in the winter of both 1987 and 1988. The eddy did not form in the winter of 1989. There are a number of other eddies evident in the Geosat data as well as variability of the Alaska Current and the Alaskan Stream. The ascending track information confirmed the conclusion made by Thomson et al. [1990] that the transition region between these two currents is not smooth as originally presumed.

The annual sea level signal as seen by Geosat at 45°N is very similar to the yearly anomalies produced by a numerical model of the same region [Paredes-Sierra and O'Brien, 1990]. The Geosat results indicate that north of the critical latitude for free Rossby wave propagation the variability of sea level is most likely a result of local forcing. The sea level signal at the higher latitudes is highly correlated at the annual frequency with the wind patterns in the region.

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