Multiscale Dynamics, Mesoscale and Submesoscale Processes for the Monterey Bay Region Circulation

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Abstract

The multiscale dynamics of the August 2003 Monterey Bay circulation has been investigated in an attempt to understand the complex processes underlying the highly variable ocean environment of California coastal region. The processes are reconstructed on three mutually exclusive time subspaces: a large-scale window, a meso-scale window, and a sub-mesoscale window. The ocean is found to be most energetic in the upper layers, and the meso-scale structures are majority trapped above 200 m. Through exploring the window-window interactions, we find that the dynamics underlying the complex surface circulation is characterized by a well-organized self-sustained bimodal instability structure: a bay mode and a Point Sur mode which are located near the bay, and west of Point Sur, respectively. Both modes are of mixed type, but they are distinctly different in dynamics. The former is established when the wind relaxes, while the latter is directly driven by the wind. Either way the wind instills energy into the ocean, which is stored within the large-scale window and then released to fuel meso-scale processes. Upon wind relaxation, the generated meso-scale structures propagate northward along the coastline, in the form of a free thermocline-trapped mode of coastal trapped waves. Following these processes there exists a secondary instability, which bridges the two primary instability modes. The dynamical implication is that the Point Sur mode and the bay mode, though distinctly different in dynamics, are connected through sub-mesoscale processes. The homoclinic tangle observed in the surface trajectories shows its consistency with this claim. We have also studied the deep layer flow, which is unstable all the time through the experiment within the bay and north of the deep trench. We observe that the deep meso-scale flow within the bay may derive its energy from the sub-mesoscale window as well as the large-scale window. This study provides a real ocean example of how secondary upwelling can be driven by winds through nonlinear instability, and how winds may excite the ocean via an avenue different from the classical paradigms.
1 Introduction

Monterey Bay is a semi-enclosed marginal sea indented on the central California coast (Fig. 1). Distinguished with its high productivity and marine life diversity, it has attracted nationwide attention, both as a tourism resort and an arena of oceanographic research. Oceanographers began to show their interests in this area as early as 1930 (Bigelow and Leslie, 1930). Since then incessant efforts have been invested to understand its circulation and dynamical processes, either as an integrated part of the California current system, or in its own right. Research along this line includes, on an incomplete list, Bigelow and Leslie (1930), Skogsberg (1936), Shepard et al. (1939), Griggs (1974), Kelly (1985), Breaker and Mooers, 1986; Strub et al. (1987), Chelton (1988), Narimousa and Maxworthy (1989), Dewery et al. (1991), Breaker and Broenkow (1994), Rosenfeld (1994), Ramp et al. (1997), Collins et al. (2000), culminating in the project AOSN-II (Autonomous Ocean Sampling Network - II), a multi-institute multi-platform comprehensive experiment of observation and forecast conducted during August-September 2003 (Robinson et al., 2005; Lermusiaux, 2005; ...). This paper addresses the multiscale physics underlying the circulation reconstructed from the AOSN-II experiment. We expect to gain an understanding of the fundamental processes involving different scale ranges, and to bring these processes to relation with the upwelling-related events.

Since Skogsberg (1936), it has been observed that the Monterey Bay circulation is complex and rich in scale. This is partly due to the highly variable environment, partly due to the irregular domain geometry, and the interplay between the two. Winds, local heating, eddies, mixing, oceanic fronts, El Nino episodes, etc, all need to be taken in account for a full reconstruction of the complex processes as observed (Rosenfeld, 1994). Among these the most studied is the wind-driven variability, which is traditionally cast within the framework of coastal trapped waves (CTW) with idealized domain and background setting (see Brink, 1991, and references therein, and Wilkin and Chapman, 1990; Yankovsky and Chapman, 1995). While these theories have shown their success in interpreting the large-scale variability along the California coast, the underlying simplification in configuration no longer holds when processes on the Monterey Bay scale are under major concern. In other words, when the bay-scale circulation is considered, the submarine valley is of finite extent; the coastline cannot be treated as a perturbation; the bottom topography curves through the whole domain, the
background field may enter the balance in a way difficult to handle. The problem is therefore rather generic. One cannot rely on configuration simplification to reduce its complexity. Apparently, the irregular geometry and highly variable environment pose a great challenge to the investigation of the dynamics of the Monterey Bay circulation.

Liang and Robinson (2005a, 2005b, hereafter LR1 and LR2, respectively) have developed a hierarchy of analysis methodologies to answer this type of challenge. Their philosophy is: data-based geophysical fluid dynamics (GFD) theories may be developed independent of domain and other environment constraints, while numerical simulation can provide the data with an arbitrary configuration. They noticed that most existing GFD theories are based on global basis, tacitly invoking the assumption of homogeneity and/or stationarity. However, real ocean processes are in nature neither homogeneous nor stationary. A gap, therefore, exists between these theories and real ocean processes, and that is the reason why it is difficult to apply these theories to interpret real oceanic phenomena. Their methodologies are developed to bridge this gap. The fundamental idea is to localize these theories to cater for nonlinearity and intermittence. In doing so, dynamics is unfolded on a basis localized in space and time, and hence one is freed from referencing the complex domain boundaries and relieved from subjective choice of time intervals. So far this idea has been applied to build a system for generalized instability studies. The whole system is based on a mathematical apparatus called multiscale window transform by Liang and Anderson (cf. Liang 2002). Developed on it are the concepts of multiscale transport, multiscale conversion, and interscale transfer, which make a new analysis methodology called multiscale energy and vorticity analysis (MS-EVA). Currently Liang and Robinson (LR2) have rigorously cast the hydrodynamic stability problem in this framework and formulated a theory for localized barotropic instability and baroclinic instability (cf. section 2). Using an aspect of this system, Liang and Robinson (2004, LR3 henceforth) successfully unraveled the physics of a meandering oceanic front event. Here we want to put it to a similar application but with different focus, in anticipation of getting a clear picture of dynamics for the Monterey Bay circulation.

The paper is organized as follows: We first give a brief introduction of how multiscale processes on a generic basis can be analyzed using the multiscale energy and vorticity analysis (MS-EVA) and a summary of the theory of MS-EVA-based localized instability. Followed is a
brief account of the dataset generated through the AOSN-II simulation. The analysis starts in section 4, where we present how a large-scale window, a meso-scale window, and a sub-mesoscale window are demarcated. The MS-EVA is set up in section 5. Processes are then unfolded on the three scale windows, and the synthesized features are described in section 6. In sections 7 and 8, we focus on how kinetic energy and potential energy are transferred from the large-scale background to meso-scale eddy structures. Particularly, section 7 gives a detailed analysis for the baroclinic and barotropic instabilities associated with the variation of the prevailing winds. In section 9, we carry the above analysis down to the sub-mesoscale window, and substantiate the multiscale dynamics inference with observations from a nonlinear dynamics perspective. This study is summarized in section 10.
2 Multiscale energy and vorticity analysis (MS-EVA) and MS-EVA-based localized instability analysis

We use the multiscale energy and vorticity analysis (MS-EVA) developed by Liang and Robinson (LR1) and the MS-EVA-based theory of hydrodynamic instability (LR2) to fulfill the objectives for this research. MS-EVA has been used in the study of the dynamics of a previously identified process, the meandering of the Iceland Faeroe Front (IFF) (LR3); in this paper, we show how it can be adopted to investigate an area where characteristic process is previously unclear.

MS-EVA is built on the basis of a functional analysis machinery called multiscale window transform (Liang, 2002, chapter 2). In this framework, a function space is decomposed into a direct sum of several mutually orthogonal subspaces, each with an exclusive range of scales. Such a subspace is termed a *scale window*. Our research task is, in an abstract way, to represent ocean processes on appropriate scale windows, and then to study how these processes evolve and interact through exploring the window-window interactions. In this context, we need a large-scale window, a meso-scale window, and a sub-mesoscale window (denoted as $\varpi = 0, 1, 2$ respectively). Theoretically, demarcation of these windows does not require a basis, but for simplicity we appeal to wavelet analysis to fulfill it, as in LR3. This yields three “window bounds”: $j_0$, $j_1$, and $j_2$, which are the wavelet scale levels marking the upper bounds of large-scale, meso-scale, and sub-mesoscale windows, respectively. In other words, given a time series scaled with its duration, $2^{-j_0}$, $2^{-j_1}$, and $2^{-j_2}$ are the lower time scale bounds for the three windows.

MS-EVA deals with multiscale dynamics through exploring the energetics on the specified scale windows. For a geofluid flow, the kinetic energy ($K_n^{\varpi}$) and available potential energy ($A_n^{\varpi}$) grow as, in a symbolic form (horizontal dissipation/diffusion ignored),

\[
\dot{K}_n^{\varpi} = \Delta Q_{K_n^{\varpi}} + \Delta Q_{P_n^{\varpi}} + T_{K_n^{\varpi}} - b_n^{\varpi} + F_{K_n^{\varpi},z}, \quad (1)
\]

\[
\dot{A}_n^{\varpi} = \Delta Q_{A_n^{\varpi}} + T_{A_n^{\varpi}} + b_n^{\varpi} + F_{A_n^{\varpi},z}, \quad (2)
\]

on window $\varpi$ ($\varpi = 0, 1, 2$) and at time step $n$, where the $\Delta Q$-terms represent the multiscale transport process on the specified scale window $\varpi$, and the “T-terms” are *perfect transfers*.
among different windows in the sense that they vanish when averaged over \( \omega \) and \( n \). In the equations, the symbol \( \left( \widetilde{\gamma}_n \right)^{\sim \omega} \) indicates a multiscale window transform (LR1, Sec. 2) on time
window \( \omega \) and at times \( n \). Other notations are summarized in Table 1. Note all these terms are horizontally treated with a two-dimensional large-scale window synthesis (LR1, section 7).

The major difference between Eqs. (1) and (2) and the classical formalism lies in that all the
terms here are field-like, and hence intermittent processes are naturally embedded. Amongst
these terms, of particular importance are the perfect transfers, which connect processes be-
tween different scale windows. It has been shown that these transfers are closely related to the
classical GFD stability (LR2), and the classical hydrodynamic stability theory in the sense
of Lyapunov (Liang et al.). A localized hydrodynamic instability analysis was henceforth
rigorously established, which we briefly introduce hereafter.

Consider first a system with only two windows involved (window 0 and window 1). Its
stability can be explored either with a large-scale window, or a meso-scale window, depending
on which window you are in when viewing the problem (except for the opposite sign). This
makes sense, as stability/instability is related to the transfers between two windows. In LR2
and LR3, Liang and Robinson introduced the theory based on the meso-scale window, in this
study we adopt the other stance to express it. Let

\[
BC = T^{0 \rightarrow 1}_{A_0^{-1}} = -T^{1 \rightarrow 0}_{A_0},
\]

where the superscript \( \omega_0 \rightarrow \omega_1 \) \( (\omega_0, \omega_1 = 0, 1) \) denotes an interaction analysis operator that
selects out the transfer component from window \( \omega_0 \) to window \( \omega_1 \) (see LR1, Sec. 9), and

\[
BT = T^{0 \rightarrow 1}_{K_0^{-1}} = -T^{1 \rightarrow 0}_{K_0}.
\]

This definition of \( BC \) and \( BT \) differs from that in LR2 and LR3 in that we are considering
\( T^{0 \rightarrow 1}_{A_0^{-1}} \) instead of \( T^{0 \rightarrow 1}_{A_0} \), and \( T^{0 \rightarrow 1}_{K_0^{-1}} \) instead of \( T^{0 \rightarrow 1}_{K_0} \) (note the difference in subscripts). The
two formulations need not be the same, but they are equivalent on the large scale window
according to the property of perfect transfer (cf. LR1). For this particular problem where no
\textit{a priori} process is specified, the advantage of Eqs. (3) and (4) over those used in LR3 for the
IFF meandering is that (3) and (4) allow one to stand on a larger scale to view the problem
and therefore to get the basic physics upfront which would be otherwise very complicated to
Table 1: Symbols for multiscale energetics (time step \( n \), scale window \( \omega \)).

<table>
<thead>
<tr>
<th>Kinetic energy (KE)</th>
<th>Available potential energy (APE)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( K_n^{\omega} ) Time rate of change of KE</td>
<td>( A_n^{\omega} ) Time rate of change of APE</td>
</tr>
<tr>
<td>( \Delta Q_{K_n^{\omega}} ) KE advective working rate</td>
<td>( \Delta Q_{A_n^{\omega}} ) APE advective working rate</td>
</tr>
<tr>
<td>( T_{K_n^{\omega}} ) Total KE transfer</td>
<td>( T_{A_n^{\omega}} ) Total APE transfer</td>
</tr>
<tr>
<td>( \Delta Q_{P_n^{\omega}} ) Pressure working rate</td>
<td>( b_n^{\omega} ) Rate of buoyancy conversion</td>
</tr>
<tr>
<td>( F_{K_n^{\omega},z} ) Rate of vertical dissipation</td>
<td>( F_{A_n^{\omega},z} ) Rate of vertical diffusion</td>
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perceive. Equations (3) and (4) have been applied successfully in the structure study and control of a turbulent wake (Liang and Wang, 2004).

The criterion of instability analysis based on (3) and (4) is the same as that of LR2 and LR3. Specifically,

1. A flow is locally unstable if \( BT + BC > 0 \) and vice versa;
2. For an unstable system, if \( BT > 0 \) and \( BC \leq 0 \), the instability is then barotropic;
3. For an unstable system, if \( BC > 0 \) but \( BT \) is not, the instability is baroclinic;
4. If \( BT > 0 \) and \( BC > 0 \) are positive, the system is undergoing a mixed instability.

In these statements, both BT and BC are field-like (localized in time and space). The criterion is hence applicable to problems on a generic basis. For convenience, BT and BC may also be loosely referred to as barotropic transfer and baroclinic transfer, respectively.

The above instability analysis may be carried down to the transfers between the meso-scale and sub-mesoscale windows. All the criteria are the same expect a replacement of window index 2 for 1, and 1 for 0. In the three-window case, if the energy source is ultimately traced to the basic background, the perfect transfer between the large-scale and meso-scale windows can be viewed a primary instability, while that between the meso-scale and sub-mesoscale windows is a secondary instability. To avoid confusion, we only assign the shorthands BC and to the former. When a secondary instability is concerned, the full expressions \( T_{A_1^{1-2}} \) and \( T_{K_1^{1-2}} \) will be written out out.
3 Dataset

In August-September 2003, a multi-institute collaborated multi-platform survey was conducted in the Monterey Bay region in a domain as shown in Fig. 1. Five cruises, which last respectively from 2 August through 6 August (Model initialization and start of science experiments), 7 August through 20 August (Science experiments and model skill assessment), 21 August through 25 August (Model re-initialization, and continued science experiments), 26 August through 2 September (Science experiments and model skill assessment), and 3 September through 7 September (Model skill assessment and final science experiments). Data are collected and taken into several numerical models as input to reconstruct the whole circulation. Details about the project are referred to Robinson et al. (2005). In this study, the output from the Harvard Ocean Prediction System (HOPS) is chosen for our analysis.

We use a well validated nowcast to form the dataset. The research domain is horizontally discretized into a $83 \times 96$ mesh grid, with an equal grid spacing $\Delta x = \Delta y = 1.5$ km. In the vertical direction a double sigma coordinate is adopted (see Robinson et al., 2004, for details). The whole procedure starts on August 2, initialized with climatology, with data taken in every day at noon upon availability until the initialization survey is completed. The nowcast is fully initialized as of August 6. For reference convenience, we label this day as day 1. Sequentially August 7 through September 7 are called day 2, day 3, and so forth. A nesting strategy is adopted to provide the horizontal open boundary conditions. In the vertical, data of external forcing are derived from the COAMPS atmospheric fields and objectively mapped onto the grid. For details, refer to Robinson et al. (2004).

Winds play a dominant role in the upwelling process in this region (e.g., Rosenfeld et al., 1994). Shown in Fig. 2 are the vectors of winds at Mooring station M1 ($36.755^\circ$N, 122.025$^\circ$W). As usual, northwesterlies prevail, though in summer winds are observed weak (Nelson, 1977; Bakun and Nelson, 1991). During this period, the maximum is attained on August 11. (If the small-scale features are filtered, the maximum occurs around August 15.) The winds relax during August 18 - 23. After that, another cycles starts and a second peak appears around August 27.

We focus on the processes between these two wind stress peaks. Time sequences of tem-
perature and flow for depths 10 m and 150 m are shown in Fig. 3. These two depths are typical of the processes in the surface layer and the deep layer, respectively. It is not our intention to give them a complete description, but from these plots it is easy to identify some features with dynamical significance. As is clear, the surface distribution is very complicated and much affected by the winds. There are two cold centers along the coast, one residing offshore Point Sur, another one between the bay and offshore Point Ano Nuevo. When the wind relaxes, warm water quickly takes over, in agreement with previous observations (e.g., Rosenfeld et al., 1994). But the most intensive upwelling appears on August 19, the day when the wind has relaxed, while no strong indication of propagation has been identified. This observation, among many others, is counterintuitive in light of what we know about coastal upwelling-related dynamics. In contrast, the deep layer is less affected by the winds and the processes seem to be simpler. Warm patches appear all the time outside Point Sur and within the bay, and propagate northwestward along the coast, regardless of the winds. It would be of interest to understand the dynamics of the emergence and propagation of the meso-scale features. Toward the end of this work, we expect to give these observations a satisfactory explanation.
Figure 1: Left: Bottom topography of the Monterey Bay region (depth in meters). Boxed is the AOSN-II research domain. Right: The research domain rotated clockwise by 30°. Indicated are the six locations where time series are extracted for spectral analysis.

4 Spectral analysis and window bound determination

As in Liang and Robinson (2004), the window bounds needed for the MS-EVA analysis is determined through wavelet spectral analysis of point time series. We have examined the following six points for this purpose (cf. Fig. 1): point 1: (65, 40); point 2: (30, 15); point 3: (60, 45); point 4: (60, 80), point 5: (55, 30), point 6: (45, 50). These points are typical of the geography and certain dynamical processes observed in the simulation. Specifically, point 1 and 3 are at the mouth of the bay, and the latter is also over the canyon at the outer bay; point 2 is in a region very energetic; points 4 and 5 are located offshore Point Ano Nuevo and Point Sur, the two crucial places in dynamics, while point 6 is within a large surface anticyclonic eddy observed most of the time through the experiment. As surface upwelling events are of special interest for this region, we focus on processes in the upper layers. To facilitate visual inspection, the means have subtracted from all the time series before plotting the spectra.

For this region, we have observed that temperature resembles density anomaly in spectrum. Temperature hence characterizes well the available potential energy distribution. We have also found that points 3 and 5 are typical of the six chosen temperature series. Contoured in Fig. 4 are their respective spectra. In both of them there is an obvious transfer of energy
Figure 2: Stickplot of winds (in knots) at M1 (36.755°N, 122.025°W) during the period of AOSN-II experiment.

Figure 3: Sequence of simulated temperature for depths 10 m (top) and 150 m (bottom).
Figure 4: Spectral energy for the surface temperature at points 3 (top) and 5 (bottom).

from low scale levels (large scales) to higher scale levels as time moves on. The transfer starts on day 8 or August 14, and is enhanced from day 12 through day 16 (August 18-22), which is the wind relaxation period. At point 3, there is a strong transfer on day 16 to a meso-scale event from both large scales and smaller scales, forming a very clear maximum at scale level 2 (8 days). All these events are related to the wind relaxation.

A similar trend is also seen on the velocity spectra. Points 4 and 5 are found to be two typical points. Shown in Fig. 5 are the spectra for $u$ at these two points. Again, the energy transfer is clearly seen, but the timing differs. At point 4, it begins on day 8 (August 14), gets enhanced on day 16, and lasts through day 24 (August 30). The peak takes place at scale level 2, corresponding to a period of 8 days. At point 5, the transfer period is from day 12 (August 18) through day 20 (August 26), and the maximum is at scale level 3 (4 days).

According to the above analysis, if one focuses on the relaxation events, both the temperature (density anomaly) spectra and the velocity spectra support a window partitioning with $j_0 = 2$ and $j_1 = 5$. That is to say, processes with scale levels smaller than 2 should be included in the large-scale window, while those higher than or equal to 5 are put in the sub-mesoscale window. This completes the time window determination.
Figure 5: Spectral analysis for the surface $x$-velocity at points 4 (top) and 5 (bottom).

As regard to the spatial scale window used for the horizontal filtering, the window bound choice has been found insensitive once the time window bounds are appropriately set. This is the same as the observation in LR3. Through this study, we choose $\text{Sp}_0 = 5$, which corresponds approximately 11 km in length.

5 MS-EVA setup

In the MS-EVA package, flat $z$-coordinates are used. The double sigma coordinate adopted in the HOPS forecast must be interpolated onto flat levels before the MS-EVA application. The $z$-levels chosen are summarized in Table 2. We select depths 0 m, 10 m, 30 m, 150 m, and 500 m for our analysis. Among them, depth 10 m is within the mixed layer (either the top, middle or bottom depending on the winds), depth 30 m lies in the middle of the thermocline, depth 150 m is roughly at the bottom of thermocline and in the core of the California undercurrent, while depth 500 m cuts across the deep California current system (see Robinson et al., 2005). The necessary fields are linearly interpolated onto these levels.

A mean profile of $\bar{p} = \bar{p}(z)$ is needed for the available potential energy analysis. The mean
Figure 6: Profiles of mean density anomaly $\bar{\rho} = \bar{\rho}(z)$.

is computed through taking average of $\rho$ over all the available data points and time instants from the simulation. The resulting profile is plotted in Fig. 6. In a brief summary, these parameters are listed in Table 2.

6 Multiscale analysis

With the parameters in Table 2, it is straightforward to make variability reconstructions on different scale windows. This section gives a description of these multiscale variabilities. The governing dynamics will be presented in the next section.

6.1 Temperature

Understanding the temperature distribution and its variation in response to the upwelling-favorable wind is a major goal for the Monterey Bay circulation study. The response is reflected on different time scale windows. Fig. 7 presents a sequence of the large-scale reconstructions of temperature for levels 2 and 12. In the upper layer, a belt of cold water dominates along
the coastline. The upwelled water is organized into two cold centers: one within the bay, another outside Point Sur. This horizontal structure is most conspicuous around August 15, and becomes weakened through the end of the month. In contrast, the interior region is characterized by a large pool of warm water. As time goes on, the pool gets enlarged and eventually takes over the whole western part of the domain at depth 10 m.

There is a distinct vertical structure on the large-scale temperature field. In deep layers, the water is generally warmer along the coastline and colder seaward, as shown in Fig. 7c. The trend reversion of offshore temperature gradient occurs roughly at 100 m (figure not shown).

The meso-scale temperature is rich in structure and process. This is particularly so in the surface layers. Shown in Fig. 8 is the meso-scale temperature evolution at level 2. Three major events are clearly seen during the period day 6 (August 11) through day 22 (August 27). On day 6, a dipole appears outside the bay, with warm water in the south. There is no evidence that this dipole propagates. Rather, it grows and oscillates in time, with shape changing accordingly. On day 14 (August 19) it reaches its peak in low temperature, and on day 18 (August 23) the phase is reversed, implying a period of 8 days for this process. As a result, a large pool of cold water appears in the middle of the domain on August 19. Comparing to
Fig. 3 and Fig. 7, the sudden cooling of water on this day should be a secondary upwelling. We shall come back to this issue later on in the MS-EVA analysis.

A new structure begins to emerge north of the bay outside Point Ano Nuevo on day 12, which becomes evident by day 14 (August 19), and strong enough on day 16 (August 21) that it since then propagates northward along the coast, with a period of about 8 days. The third event is found on August 13 as a dipole near the southwestern corner, with a cold center to the left. This dipole keeps growing until the wind relaxes, when it begins to propagate northward on August 21. The patterns of August 25 and August 21 are approximately out of phase, implying this event also has a period of roughly 8 days and a wavelength 60 km, which implies a speed of 0.08 m/s. We have computed a coastal trapped wave (CTW) properties with the averaged bounancy frequency profile and topography for this region (see Appendix A). This observation agrees with a surface-trapped mode of CTW with an eigenvalue (inverse of phase speed) 11 s/m.

There is also a vertical structure on the meso-scale field. A distinct feature is that the meso-scale variation amplitude descreses with depth. A descreasing trend is also seen on the horizontal scale, as shown in the distributions for depth 10 m (Fig. 8) vs. depth 150 m (Fig. 9). Below 250 m, meso-scale variations give away to sub-mesoscale processes.

The processes in deep layers are relatively simple. At level 12 (150 m), we see cold and warm pools being alternatively generated within the bay, and propagating northward along the coast. Comparing the patterns of August 19 and August 23, the period is also roughly 8 days, and its wavelength approximately 60 km, with a speed of 0.08 m/s. That is to say, the deep layer meso-scale structure also propagates in the free CTW mode as computed in the Appendix.

6.2 Velocity

The flow field is also rich in scale. Fig. 10 shows the large-scale reconstructions for levels 2 (10 m) and 12 (150 m). They are typical of the flows for the surface layers and deep layers. At level 2, the flow is a little complicated due to the direct influence of winds. Roughly it can be classified into two types: one costal current flowing southward, one northward current offshore pertaining to the California Counter Current (CCC). These two currents connect off
Point Sur and make an anti-cyclonic gyre. For the deep layers, the flow pattern is much simpler. The whole system is characterized by a strong northward along-slope current, with a weaker California Under Current to the left. The general trend of the large-scale current is getting weaker toward the end of August.

On the meso-scale window, the flow exhibits itself in a complicated pattern in upper layers, particularly in surface layers. In the 10-m sequence, as that in Fig. 8, the strength of variability does not change much until the wind relaxation (cf. the distribution of August 17 in the figure), when a burst of variability appears which lasts toward the end of the month. Eddy structures are generated, advected, and diminished here and there, making the flow a very complex system. We will return to the analysis later with more powerful methodologies.

The deep layer variability is relatively simpler. Drawn in Fig. 9 is a sequence of the 150-m meso-scale flow. Generally, the variability is in the form of a dipole outside the northern part of the bay. The variability keeps being generated within the bay area during August 11 through August 23. It then propagates northward, becomes weakened and finally disappears just near the northern boundary. It may be summarized as a source outside the bay, plus a sink near the northern boundary. On August 19 and August 23, the patterns are roughly out of phase, implying a period of about $4 \times 2 = 8$ days. One may also measure the wavelength from the pattern of August 23 to be 60 km. This gives a speed of about $0.08 \text{ m/s}$, in agreement with the free CTW mode described above. In addition to the major feature, there is also a process occurring near the southwestern corner. A cyclonic eddy and an anticyclonic eddy appear alternatively as time goes. The amplitude increases but no propagation is identified. Underlying there could be a local instability. We will see more dynamics details later.

6.3 Multiscale energy

Multiscale energy distributions could expose more information about the ocean response to the external forcing. As we will see later, the effect of winds enters the energy balance mainly on the large-scale window. For this reason, we only look at the large-scale energy and its variation.

At depth 10 m, the large-scale potential energy (Fig. 13, upper panel) exhibits itself mainly in two regions: One lies in the bay, another off Point Sur. For convenience, we will refer to
them as the bay mode and the Point Sur mode, respectively. Both the bay mode and the Point Sur mode get strengthened as time goes starting from August 11, and reaches their respective maxima on August 15, and then decrease afterwards toward the end of the month. Note that sometimes the bay mode extends northward to Point Ano Nuevo.

The large-scale bimodal structure also exists on the surface layer large-scale KE sequential maps (Fig. 13, lower panel), and their variations also follow a similar fashion. The only difference is that the Point Sur mode lies far offshore, and it becomes strengthened again after the relaxation period (August 18 - 23). It seems that the Point Sur mode on the KE map has a better relation to the wind stress. Apart from these two modes, a third hotspot resides in the northwest. It is strongest on August 11 during the period in study. We will see in the next section that underlying this hotspot the dynamics is completely different from that of other two modes.

The deep-layer large-scale energy follows a different pattern on the distribution. Drawn in Fig. 14 is the large-scale APE and KE of August 15 for depth 150 m. Clearly the flow stores its APE (left panel) on the coastal side, while kinetically it is most energetic along the shelf-break. The largest kinetic energy occurs on August 11. It decreases afterwards through the survey period (figure not shown).
Figure 8: Sequence of meso-scale temperature at depth 10 m.
Figure 9: Sequence of meso-scale temperature at depth 150 m.
7 MS-EVA analysis: Barotropic and baroclinic transfers

The multiscale processes are governed by multiscale dynamics. MS-EVA is a problem-independent methodology to unravel the dynamics. In this section, we focus on the interaction between large-scale and meso-scale windows, which is measured by two field-like metrics, the barotropic transfer and baroclinic transfer, or BT and BC for short. These metrics have been connected to barotropic instability and baroclinic instability in a generalized sense (cf section 2).

7.1 Surface layers

As is shown in the preceding section, the dynamics of the upper layer is relatively complex. We calculate its BT and BC distributions to distinguish intrinsic mechanisms from extrinsic mechanisms. Again, level 2 (10 m) is found typical of the upper layer in this regard. Figs. 15 and 16 depict how BC and BT evolve with time on this level.

Look at BC first. A remarkable feature is its bi-modal structure: Clearly shown in the sequence of Fig. 15 are two positive BC centers. We look at the southern one first. Starting from August 11, it emerges in the domain outside Point Sur. It becomes stronger as time goes on and reaches its maximum on August 15. After that, the transfer strength decreases until August 21 when it splits into two parts. The upper part eventually merges into another
Figure 11: Sequence of meso-scale velocity at depth 10 m.
Figure 12: Sequence of meso-scale velocity at depth 150 m.
Figure 13: A time sequence of large-scale available potential energy (upper panel) and kinetic energy (lower panel) for level 2 (10 m).

Figure 14: Large-scale available potential energy (left) and kinetic energy (right) at level 12 (150 m) for August 15, 2003.
center further north which we will describe soon. The bottom part rejuvenates after August 23. This BC hotspot correlates well in location and variability to the Point Sur mode on the large-scale APE maps (Fig. 13). Notice positive BC means loss of energy to the meso-scale window. So the high correlation between $BC$ and the large-scale APE shows that the wind stores energy in the large-scale window, and then releases it to meso-scale processes.

Another positive center on the sequential maps is located between the Bay and Point Ano Nuevo. In comparison to the bay mode on the maps of Fig. 13, the location is a little northward. But for convenience we will still refer to it as the bay mode. This mode does not emerge until August 15, when the wind slows down. It reaches its maximum on August 21, and gets weakened afterwards.

Both the two BC hotspots are highly correlated with the wind. The Point Sur mode generally follows the wind stress variation, with a phase lag of 2-3 days. In contrast, the bay mode has a negative correlation with the wind. The signs of correlation indicate that the two baroclinic instabilities are triggered through different mechanisms. This will be clear soon.

The opposite correlations with the wind and hence the relative variation between the two modes suggest some kind of relation between the two transfer centers. It is not clear whether this means a conservation law while the two modes interact. But we do observe interactions between them, and particularly interactions are observed during the relaxation period. They are marked by the the splitting of the Point Sur mode, which should exchange energy and information as well.

On the BT maps (Fig. 16), there is also a bi-modal structure, and a similar evolution pattern is observed. The slight difference between BT and BC is that the two modes of BT are located a little southward and eastward before the wind relaxes, and that the Point Sur center becomes negative during the relaxation period. In this case, the “bay mode” is really within the bay before August 17, but the Point Sur mode is far from the coast. The correlations of the two centers with the wind are also in opposite, just as that on the BC maps.

Although BC and BT follows a similar variability pattern with the wind, the wind influence is more conspicuous on the BT evolution. In the south, negative transfer appears outside Point Sur during August 18-23; in the middle, the hotspot jumps from within the bay on August 17 to where the BC bay mode resides. Both these prominent variabilities occur in response to
the wind relaxation, indicating a closer relationship of BT to external forcing.

The BC and BT distributions show that the system is unstable in the upper layer within the bay after August 11, because BC + BT is positive, and the instability is of a mixed type. The most unstable case occurs during the relaxation period, August 18 through 23. Off Point Sur, the flow is also unstable except during the relaxation period, and the instability is also of a mixed type. But it is dominated by baroclinic instability before August 18, and switched to barotropic instability after August 23. During the relaxation period, the flow tends to be stabilized outside Point Sur. The high correlations of BT and BC with the wind indicate that these instabilities are to a large extent triggered externally. The whole story will be clear after a deeper study of the energetics.

### 7.2 Deep layers

The deep layer transfer patterns are much more simpler than their surface counterparts. Shown in Fig. 17 are the BC (left panel) and BT (right panel) on August 15 for the depth 150 m. On the BC map, the whole domain is characterized by a positive center within the bay, and a weak negative center to the west of Point Sur. All the rest regions are virtually zero in transfer. This simple distribution keeps its structure through the whole experimental period, though the hotspot gets weakened, and the negative center disappears during relaxation (figures not shown).

The 150-m BT and its evolution is also simpler than the surface BT. In the right panel of Fig. 17, it is basically composed of three distinct centers: a positive hotspot in the middle, a negative center to the north, and another weaker positive center to its south. The southern center varies in a way similar to the Point Sur mode at depth 10 m as shown in Fig. 16, and also becomes negative when the wind relaxes. The northern center is always negative, forming a sink to certain perturbations. This is consistent to what we have observed in the meso-scale flow plots (Fig. 12).

What warrants our while is the middle hotspot. It exists from August 12 to August 23, and does not vary much during the period, both in strength and location. Clearly it does not show correlation to the wind. If we compare it to the topography in Fig. 1 and the flow in Fig. 10, it sits just downstream of the deep canyon. Obviously it is caused by the California Under
Figure 15: Potential energy transfer at level 2 (10 m) from large-scale window to meso-scale window (BC). (Units: $m^2/s^3$)
Figure 16: Kinetic energy transfer at level 2 (10 m) from large-scale window to meso-scale window (BT) (in m$^2$/s$^3$). For comparison purpose, contours with values exceeding ±3.5 × 10$^{-7}$ are not drawn.
Figure 17: Potential energy transfer (left) and kinetic energy transfer (right) at level 12 (150 m) from large-scale window to meso-scale window. (Units: $m^2/s^3$)

Current flowing over the canyon, a mechanism distinctly different from that of other positive centers.

8 MS-EVA analysis: Energy balances on multiscale windows

We have examined the baroclinic and barotropic transfers during the large-meso-scale interactions, and explored the correlation between these transfers and the wind stress. In this section, we study the other MS-EVA terms on the energy balance. As it is observed that the external forcing enters the equation mainly through the large-scale window, we consider the large-scale MS-EVA only.

8.1 Point series

We first choose two locations, S: (30,15), and B: (57,55), to study the problem. These locations are associated with the two distinct transfer centers over the experimental period. For convenience, we refer to them as S (“Sur”) and B (“Bay”), respectively.
Plotted in Fig. 18 is the point series of the large-scale MS-EVA terms at point S. One observation is that external forcing contribution dominates the balance. Another observation is that buoyancy conversion is rather weak. This is different from that of the previous MS-EVA application case, the Iceland-Faeroe frontal variability problem (LR3). Balancing the work due to the external forcing is mainly from the horizontal advection in the APE equation, and the horizontal pressure work in the KE equation. The importance of other terms varies with time and location.

It merits mentioning that both the baroclinic transfer and the barotropic transfer make significant contributions to their respective equations. As we note before, our BC and BT are equal to $-T_{A_n}$ and $-T_{K_n}$ followed by an interaction analysis which selects out the part from large-scale window to the meso-scale window (indicated as superscript $0 \rightarrow 1$ in the text). The variations of $T_{A_n}$ and $T_{K_n}$ then from an aspect reflect the BC and BT, except for a negative sign. Since we have seen that energy does not convert much between the two types, these transfers are the main indices for the temperature fluctuation and flow variability, respectively. In this sense, although the instability has a ingredient of baroclinicity, it is not Eady-like, as the buoyancy conversion is rather weak.

The correlation between the external forcing and the transfers is clearly seen from the plots. In Fig. 18a, the transfer strength is positively correlated to $F_{A_n}$ with high correlation. Particularly, $-T_{A_n}$ reaches its maximum just a couple of days after the $F_{A_n}$. (Recall related to instability is negative $T_{A_n}$.) The same observation is made in Fig. 18b before August 23, when the wind relaxes. The largest negative value of $T_{K_n}$ occur on August 13, while on the same day the wind instills the largest part of energy into the ocean. When the wind relaxes, the transfer becomes positive, i.e., the flow is stabilized, in agreement with our previous observation with the BT sequence. After August 23, this region experiences another instability, but the driving mechanism seems to be changed. Nonetheless, one can safely say that the Point Sur instability mode is directly driven by the external forcing.

For Point B which is with the bay mode, the energetic scenario is a little different. In the APE equation, the most significant balance is still between the external forcing and the horizontal advection. But in the KE equation, the balance is between horizontal pressure work and the work done by winds together with vertical pressure force. Especially different is
Figure 18: Large-scale APE (a) and KE (b) balance for point (30,15) at depth 10 m. In (a), $Q_{A_{h}}$: solid; $Q_{A_{v}}$: dashdot; $b_{h}^{C}$: blue dashed; $F_{A_{h}}$: thick grey solid; $T_{A_{h}}$: thick black solid.
In (b), $Q_{K_{h}}$: solid; $Q_{K_{v}}$: dashdot; $Q_{p_{h}}$: thick grey dashed; $Q_{p_{v}}$: thick dashdot; $u_{h}^{C}$: dashed; $F_{K_{h}}$: thick grey solid; $T_{K_{h}}$: thick black solid. These mechanisms as a whole account for the growth of large-scale APE and large-scale KE, respectively.

Figure 19: Same as Fig 18, but for point B(57,55).
the correlation of the transfers to the external forcing work. In Fig. 19a, $T_{K^n}$ almost vanishes when the absolute value of $F_{n}$ reaches its peak on August 14-15, while $-T_{K^n}$ is maximized as wind relaxes. Same thing is shown in Fig. 19b. $T_{K^n}$ takes its significant negative value from August 17 to August 25, which correspond to the valley of the curve of $F_{K^n}$. Clearly, the instability mode at point B has a driving mechanism completely different from that of point S. It is the relaxation of winds that makes this region unstable.

8.2 Horizontal MS-EVA maps

We now look at the horizontal maps of the major MS-EVA terms. In Fig. 20, the most significant two terms in the APE balance are contoured for August 15. Observe the two negative centers on the $F_{n}$ map. They correspond well in location to the two instability modes we have discussed above.

The KE balance is mainly between horizontal pressure working rate $\Delta_h Q_{P0}$, vertical pressure working rate $\Delta_z Q_{P0}$, and the wind stress working rate $F_{K^n}$. We have seen from Fig. 19 that the three correlate well to each other. Indeed, this is true in most of the regions at depth 10 m, as seen in Fig. 21. Of particular interest are the two distinct positive regimes of $F_{K^n}$. They are related to the two barotropic transfer centers.

It merits notation that there is a strong negative regime in the top-left corner of Fig. 21c. Recall that in discussing the large-scale energy distribution (cf. Fig. 13), we have identified a large kinetic energy center in this place. Different from the two instability modal centers, we cannot identify a correspondence on the transfer maps. Now it is clear that this kinetic energy patch is highly related to the wind, and is therefore not due to intrinsic reasons.

To summarize, the large-scale MS-EVA balance is mainly between the the external forcing and the advective work together with the pressure work. Both the baroclinic transfer and the barotropic transfer are highly correlated to the wind, but the correlations in the two distinct transfer centers are quite different. The Point Sur mode is excited directly by the wind, while the bay mode is due to a loss of balance established as the wind applies. That latter provides an example of excitation of real ocean motion through storing energy first within large-scale
Figure 20: Large-scale potential energy balance at the second level (10 m) on August 15 (units: $m^2/s^3$).

Figure 21: Large-scale kinetic energy balance at the second level (10 m) on August 15 (units: $m^2/s^3$).
window and then releasing it to form meso-scale processes.

9 Sub-mesoscale processes and inter-modal communication

According to the previous spectral analysis, processes with scales less than half a day are belonging to the sub-mesoscale window. Sub-mesoscale processes are observed to be active for this region. In this study, however, it is difficult to investigate them on the sub-mesoscale window. The original simulation was generated such that observational data are taken in on a daily basis. Daily processes are interfered with by assimilation scheme, and a direct analysis of the sub-mesoscale energetics therefore could not be reliable.

Nevertheless, we may still gain some insight into the sub-mesoscale physics through investigating the transfers between the meso-scale window and the sub-mesoscale window. Transfer processes serve as a protocol between different scale windows, and as we did before, they can always be computed based on the larger window. In this case, we need only examine the meso-scale window.

We first look at the simple case. In deep layers such as level 12 (150 m), the sub-mesoscale processes are rather weak (figures not shown). We just give a brief description here. On the maps of $T_K^{1\rightarrow2}$, the significant transfer occurs in a region just near Point Ano Nuevo, particularly after the relaxation. On the $T_A^{1\rightarrow2}$ maps, the transfer is limited within the bay. Generally speaking, it is positive before August 20, and after that it turns into negative. Recall that mesoscale thermal structures are generated at this level within the bay all the time through the experiment. The negative $T_A^{1\rightarrow2}$ means that, although deriving their energy mainly from the large-scale background, these meso-scale structures after August 20 may also have a partial energy source from the sub-mesoscale processes.

The sub-mesoscale processes on the surface levels are more interesting. Contoured in Fig. 22 is a snapshot of the second level (10 m) energy transfers from the meso-scale window to the sub-mesoscale window. Both the APE (left) and KE (right) demonstrate a cascade of energy toward smaller scales on this day (August 15). This trend persists on the KE transfer map as of August 21, when it becomes more complicated. On the APE transfer map, the distribution is similar, except a negative center sneaks in from southwest when the wind relaxes (figure
Figure 22: Potential energy transfer (left panel) and kinetic energy transfer (right panel) between meso- sub-meso-scale windows on August 15 for depth 10 m (units: m²s⁻³). Positive value indicate a transfer from the meso-scale window to the sub-mesoscale window.

not shown).

Compared to Figs. 15 and 16, Fig. 22 indicates clearly that a secondary instability follows the primary instability between the large-scale and meso-scale windows, releasing energy to sub-mesoscale processes. This secondary instability exhibits itself in a type mixed with baroclinicity and barotropicity. If observing closely the day-10 (August 15) pattern, one finds it actually sitting in a location bridging the bay mode and the Point Sur mode on the maps of Fig. 16 and 15. That is to say, the energetic scenario here may be described as two primary instabilities within the bay and west of Point Sur, followed by a secondary instability lying in between.

The critical location of the meso-submeso-scale interaction suggests a possible mechanism connecting the two dynamically isolated regions. As shown in Figs. 15 and 16, the two centers hosting the bay mode and the Point Sur mode are dynamically developed almost independently. Whether these two regions interact and how they communicate with each other is therefore of interest. Fig. 22 implies that there could be a bridge on the sub-mesoscale window. That is to say, these two regions could talk with each other through a kind of sub-mesoscale processes caused by a secondary instability.

Evidence has been identified from the surface trajectories to substantiate the claim, we
look at the tracks of a cluster of drifters (15 in total) launched within the bay in the original simulation. Most of the trajectories are either of type A like drifter 3 (Fig. 23a) or in a closed form (type B) as shown in Fig. 23b. Because of the limited forecast cycle, we do not know where the unstable trajectory or manifold of Type A extends. But it seems that it always directs toward the center of the Point Sur mode, reminding one of the separatrix surrounding two equilibrium points in a dynamical system. Though incomplete, the surface trajectories seem to support our bi-modal structure of dynamics.

Type B (Fig. 23b) trajectory fairly resembles a homoclinic manifold and a homoclinic tangle (e.g., Holmes, 1983). A homoclinic manifold is a trajectory issuing from a saddle point that returns back to the point, namely, both a stable and unstable manifold with respect to the saddle. It has been established that once the stable manifold intersects with the unstable manifold, then there will be infinitely many intersections within a limited region. These intersections are termed homoclinic tangle. Although we are still not quite sure that the manifold in Fig. 23b is homoclinic because of the resolution of the numerical scheme, the intersections it exhibits bears resemblance to some homoclinic tangle.

Homoclinic tangle implies vehement small scale intersections and hence enhanced mixing with the surrounding particles. As implied by Fig. 23b, the enhanced mixing could occur at the baymouth. This is consistent with what we have observed with the sub-mesoscale transfers. In this sense, it is possible that the two modes are interacted on the sub-mesoscale window.

10 Summary and conclusions

The multiscale dynamics of the August 2003 circulation in the Monterey Bay region has been investigated using the multiscale energy and vorticity analysis (MS-EVA) and the MS-EVA-based localized instability theory. The whole system is found to be governed by a bimodal instability structure, with two primary instabilities at two different locations and a secondary instability lying in between. We have studied how the wind instills energy into the ocean to drive this structure, and how the resulting two dynamically distinct centers communicate with each other.
Figure 23: Trajectories of two of the 15 drifters released within the Bay

The Monterey Bay system is found well organized in space and windowed in (time) scales. It can be roughly classified into two dynamically different types: a surface type and a deep type. During the experiment period, two regimes have been identified as intrinsic sources of the complicated surface flow system. Represented on the MS-EVA maps these regnies are two positive centers of baroclinic and barotropic perfect transfers (resp. BC and BT) from the large-scale window to the meso-scale window as constructed. They are located near the bay and offshore Point Sur, respectively, and has been since termed as the bay mode and the Point Sur mode for convenience.

The two instability modes in the surface layers have different driving mechanisms lying behind. We have correlated the BC and BT in these two regimes to the wind. Both the correlations are very high, but the responses are in opposite. A conclusion drawn from the observation is that the bimodal structure results from the wind, perhaps in cooperation with the coastal geometry. The difference between the two modes is that outside Point Sur the wind destabilizes the system directly, while within the bay, the wind tends to stabilize the southward coastal current, and instability occurs when the external constraint is relaxed.

The two centers on the surface bimodal structure are relatively isolated in dynamics. But they do communicate in some way. The communication is evidenced to be associated with sub-mesoscale processes, which is due to a secondary instability occurring in between the two regions. The claim is further substantiated by the homoclinic tangle frequently observed on the trajectories of the surface drifters launched within the bay.
Corresponding to the energy transfers toward the meso-scale window are the meso-scale fields. The meso-scale temperature and flow are found to be trapped above 200 m during the experiment, with a maximum near the thermocline. The potential and kinetic energy transfers generally correspond in timing to the meso-scale temperature and velocity respectively, though sometimes discrepancy in location does exist. We observed that a remarkable cooling during the wind recession on August 19. It is clearly not driven directly by the wind, nor from a remote region via wave propagation; rather it is caused by the local baroclinic instabilities. We have found the bay is a source region of perturbation. Disturbances are generated all the time during the experiment, and the generated disturbances propagate northward along the coast, in the form of a thermocline trapped mode of coastal trapped waves.

The multiscale dynamics for the Monterey Bay circulation has a distinct vertical structure. To complete the story, we have also conducted a study for the deep flows (below the thermocline), which dynamically are much simpler than their surface counterparts. During the experiment period, the bay is found baroclinically unstable all the time, and that is the only hotspot on the maps of large-to-mesoscale APE transfer. The corresponding large-to-mesoscale KE transfer is a little different. The flow is barotropically unstable in the middle of the domain, starting from August 11 to the end of wind relaxation (August 23). Attached to the north is is a negative transfer (stable center) all the time, while to the south is also so for most of the time just outside Point Sur. We have analyzed the transfers from the meso-scale window to the sub-mesoscale window. A remarkable observation is that within the bay they are negative after the wind relaxes toward the end of the month. That is to say, the deep layer meso-scale flow is not only fueled by the large-scale flow via instabilities, but also has its source derived from the sub-mesoscale processes.

We close the paper by remarking that this study shows an avenue how winds can excite the ocean through building up energy in the large-scale background, and then releasing it upon relaxation to fuel meso-scale eddies. We have also learned that a sudden cooling in the coastal ocean does not need to be directly driven by the wind, nor a result of remote cooling via wave propagation. It could be driven by a local instability process occurring in situ.

Acknowledgments
Figure A.1: (a) Topography and (b) buoyancy frequency profiles for the coastal trapped waves computation.

A Coastal trapped waves in the Monterey Bay region

We have computed the free coastal trapped waves for three cross-shelf sections for the Monterey Bay region. (They are similar in result.) One section is outside Point Ano Nuevo, where wave propagations are frequently observed during the AOSN-II experiment. The topography is slightly modified, as marked in Fig. A.1a. The basic buoyancy frequency, which is shown in Fig. A.1b, is computed from the density profile averaged over all the available simulated density data on the section through the experiment. The eigenvalue problem is formed and solved with the method by Brink (1980), Clarke and Gorder (1986), and Wilkin and Chapman (1987). The eigenfunction of the thermocline-trapped mode is shown in Fig. A.2, which corresponds to a phase celerity of 0.9 ms$^{-1}$. 
Figure A.2: The thermocline trapped mode of the coastal trapped waves in the Monterey Bay region. It corresponds to an eigenvalue \((1/c)\) of 11 m−1 s \((c \approx 0.09 \text{ m/s})\). The signs \(”+”\) and \(”−”\) indicate the positive and negative regions of contour lines.

References


