COASTAL UPWELLING ALONG THE WEST COAST OF VANCouver Island

By

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We accept this thesis as conforming to the required standard

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ABSTRACT

Wind-driven upwelling events near Brooks Peninsula off the west coast of Vancouver Island have been identified during the summers of 1988 and 1989 from sea surface NOAA AVHRR thermal imagery obtained at the UBC Satellite Oceanography and Meteorology Laboratory. Software has been developed to characterize the strength and extent of the surface cooling associated with the upwelling. A two-dimensional, 2-layer finite difference model with 1 km resolution has been formulated to examine the small-scale dynamics of the upwelling events. The model uses local wind and includes realistic coastline and bathymetry. The wind-stress and coastline configuration appear to be the main factors contributing to favourable upwelling regions. The results of the model compare favourably to the observed starting location of the upwelling.
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1.0 Introduction

This thesis consists of an examination of upwelling 'events' off the west coast of Vancouver Island. Sea surface temperature measurements from the Advanced Very High Resolution Radiometer (AVHRR) on board the National Oceanic and Atmospheric Administration (NOAA) series satellites are used to identify regions of cold water. A numerical model using high resolution (1 km), realistic bottom topography, coastline, and wind stress is developed to try to identify which regions along the west coast are susceptible to upwelling, and to compare calculated sea-surface temperature distributions with those seen in satellite imagery.

Chapter 2 describes the background of the upwelling process and the use of the Bakun index as a coarse indicator of likely upwelling regions along the west coast of North America. Chapter 3 describes the processing and results of using satellite AVHRR data to identify cold water regions along the west coast of Vancouver Island.

Chapter 4 describes the development of a 2-layer numerical model based on realistic wind stress coastline and bathymetry. Recent studies (Peffley et. al. 1976, Haun et. at. 1983, Cai 1990) have shown how important local coastal bathymetry and coastline are to the development of upwelling along a coast. Experiments, in chapter 5, using the numerical model developed in chapter 4, are carried out to determine the features important to upwelling along the west coast of Vancouver Island. Chapter 5 also describes the results of using the wind stress observed, during a 1989 upwelling event, in the numerical model. The results of the model are compared to observations made using AVHRR satellite data in chapter
6. The success and possible future developments of using this model to determine upwelling regions along the west coast of Vancouver Island are discussed in the conclusions, presented in chapter 7.
2.0) **UPWELLING**

2.1) **Background**

Upwelling is the term used to describe the phenomenon by which water from depths of between 100 m and 300 m is brought up to the surface. The water in these deeper layers is usually cooler than the surface water and is also richer in nutrients because the nutrients in the warmer water have been depleted by the growth of phytoplankton near the surface. Upwelling areas bring subsurface nutrient-rich water to the light-rich surface water thus stimulating the development of phytoplankton blooms. These support a greater zooplankton concentration and this results in regions of high fish populations (Thomson, 1981). Coastal upwelling areas account for only 0.01 percent of the world's oceans but supply over 50 percent of the world's seafood (Packard, 1981).

Upwelling can occur when warm surface water along a coast is transported offshore by a favorable wind stress, allowing the colder water below the surface to replace it. Upwelling can also occur when cold water from below forces its way up to the surface in tidal currents that are deflected upwards by underwater ridges. Examples of cold water being deflected to the surface by underwater ridges can be observed at Active Pass in the Gulf Islands and at the Cobb seamount off the southwest coast of Vancouver Island (Thomson, 1981). Tidally induced vertical mixing is also important in such events.
In the northern hemisphere the near-surface, wind-driven mass transport (or Ekman transport) is directed 90 degrees to the right of the wind vector (see figure 1).

Figure 1: Wind Driven Upwelling (Thomson, 1981)

In the winter, a low pressure system dominates the atmospheric pressure distribution in the northeast Pacific and, as a result, winds are predominantly from the southeast, and no upwelling occurs (Thomson, 1981). In the late winter cold Arctic air usually occupies the interior of the province; occasional spill-over of this air mass occurs down B.C. coastal fjords, leading to intense winds and surface cooling (Tyner 1951). Although some upwelling may be caused by this process this is not the phenomenon which I consider here.
A major reversal of the prevailing winds along the west coast of North America occurs between late February and early May. A reversal in the current over the shelf break (200 m - 1000 m), lowering of the mean sea level, the re-stratification of the shelf-slope waters and favorable wind conditions for upwelling, are all associated with this "spring transition" (Thomson et. al. 1989).

In the summer, the prevailing winds are from the northwest along the west coast of Vancouver Island. Figure 2 shows the typical monthly mean wind vectors, calculated using 6-hourly pressure-derived geostrophic winds (Thomson et. al 1989). The winds shown are mainly from the north during the summer. Wind-induced upwelling during the summer months occurs each year and appears especially pronounced at the edges of the southern banks. (Freeland and Denman 1982).

Figure 2: Typical winds along west coast of Vancouver Island

The reversal of the prevailing "summer" winds occurs in mid-September to late October. A rise in mean coastal sea level, enhanced wind and convective mixing of the surface waters, and the lack of
Figure 3: Cross-shore structure of temperature, salinity, dissolved oxygen and density (sigma-t) seaward of Tatchu Point central coast of Vancouver Island for July 1980. Note the logarithmic depth scale (Thomson, R.E. et al. 1989).
favorable upwelling conditions are all associated with this "fall" transition (Thomson et. al. 1989)

During favourable upwelling conditions, the coldest surface waters appear right along the shores of Vancouver Island and the Queen Charlotte Islands. The upwelling conditions are characterized by upward sloping isotherms (temperature), isohalines (salinity), isopycnals (density), and dissolved oxygen levels over much of the water column (see figure 3) (Thomson et. al. 1989). Maximum upwelling is indicated by a near shore minimum in temperature and a maximum in salinity, density and dissolved oxygen (Landry, J.R. Postel et. al. 1989).

The general circulation of the ocean currents for winter and summer is shown in figure 4a. The near shore currents for winter and summer are shown in figure 4b. At the latitude of the southern B.C. coast the North Pacific Drift splits in two, with the Alaska current flowing north and the California current flowing south (see figure 4a). The Alaska current, is a weak, broad and ill-defined drift and flows at about 10 cm/s (Thomson, 1981). The location and intensity of the Alaskan gyre have been shown to have seasonal and interannual variability (Royer and Emery, 1987).

The current within the first 20 km of the shore is northward and fluctuates seasonally, but is present all year-round. Near the coast the currents are subject to significant seasonal variability. The Davidson current is observed only in winter months (Hickey, 1979). In January and February the near shore shelf current is predominantly northward, and is
Figure 4 (a): General ocean circulation off Canada's west coast (adapted from Thomson 1981).
Figure 4 (b): General coastal circulation off west coast of Vancouver Island (adapted from Thomson 1981).
most intense at the coast. The speeds vary from 8 cm/s on the southern shelf to 12 cm/s off Estevan Point. The flow over the shelf, in the Brooks Peninsula and Estevan Point areas, are northward in March, weak and indeterminate in April, and Southward in May. The flow remains predominantly southward until the fall transition in September to October, when the flows return to the winter regime (Freeland et. al., 1984)

Tidal currents are weak in the deep ocean and, since tidal currents are of a periodic nature, they are, in general, only minor contributors to the total water transport in the offshore region. At large distance offshore the deep-sea circulation in northward, and fluctuates substantially. Observations by Freeland et. al. (1984) suggest Vancouver Island is in the Alaska gyre system.

During the summer a 10 to 20 meter layer of relatively warm water overlies the southern portion of the Vancouver Island shelf (Thomson et. al, 1989). In the winter, the inner shelf region ( depth < 200 m) is characterized by well mixed cold water with low salinity. Warmer and more saline water is found over the outer shelf (Thomson et. al. 1989).

The extent of wind-driven upwelling off the west coast of Vancouver Island has been observed to cover a surface area ranging from 1 to 100 km² (Thomson, 1981). The temperature difference between the colder upwelled water and the relatively warm surface water may be as much as 7° C. In this thesis, I will use satellite imagery to identify wind-driven upwelling areas and compare them with the results of a numerical model.
2.2) Upwelling Index

Along a coast the offshore-directed Ekman transport may be characterized by an 'index of upwelling', which is tabulated for the west coast of North America by the National Oceanic and Atmospheric Administration (NOAA). These upwelling index reports give the upwelling index for various locations in daily, weekly and monthly plots (Bakun 1973).

An upwelling index, defined as mass transport per 100 meters of coastline, can be generated by calculating the component of the Ekman transport that is perpendicular to the coast (Bakun 1973). The mass transport in the Ekman layer is given by:

\[ M = \frac{\tau}{f} \]  

--- (1)

where:  
\( M \) is the magnitude of the mass transport,  
\( \tau \) is the magnitude of the wind stress, and  
\( f \) is the Coriolis parameter.

To calculate the wind stress, the geostrophic wind is computed from surface pressure maps:

\[ u_g = -(f \rho_a R)^{-1} \left( \frac{dP}{d\phi} \right) \]  

--- (2)

\[ v_g = (f \rho_a R \cos \phi)^{-1} \left( \frac{dP}{d\lambda} \right) \]  

--- (3)

Where:  
\( \rho_a \) is the density of air,  
\( P \) is the atmospheric pressure,  
\( R \) is the mean radius of the earth.  
\( v_g \) is the northward component of the wind velocity.
\( u_g \) is the eastward component of the wind velocity
\( \phi \) is the latitude
\( \lambda \) is the longitude
\( f \) is the Coriolis parameter.

The wind near the sea surface is formed by rotating the geostrophic wind vector by 15 degrees to the left and reducing it by 30% to account for frictional effects (Bakun 1973). The sea surface stress can be computed from:

\[
\tau = \rho_a \, C_d \, |v| \, v
\]

--- (4)

Where: \( \tau \) is the stress vector,
\( \rho_a \) is the density of air,
\( C_d \) is an empirical drag coefficient, and
\( v \) is the estimated wind vector near the sea surface.

The National Oceanic and Atmospheric Administrations (NOAA), calculates a coastal upwelling index using wind stresses obtained for points near the west coast of North America. A negative value of the index indicates accumulation of wind-transported surface water at the coast resulting in downwelling. A positive upwelling index is a measure of how much water is upwelled, from below the Ekman layer, to replace water driven offshore. Both upwelling and downwelling occur as distinct events at various times of the year. Upwelling predominates, however, during the summer months (May through September), and downwelling dominates during the winter months (October through March).
The upwelling index is calculated at 15 stations along the west coast. The stations closest to the B.C. coast, are located at 51N, 131W (station 5) and 48N, 125W (station 6) (see figure 5). The upwelling index is at a maximum earliest in the south and progresses northward. Off the coast of southern California the peak is usually in May. Along the west coast of Vancouver Island the peak is usually between the end of June and the beginning of August (see figure 5).

Neither the bathymetry nor the coastline geometry are taken into account in the calculation of the upwelling index which, therefore, provides only a very coarse indicator of possible upwelling along a coast. Satellite images, on the other hand, provide a very accurate and cheap way of determining when and where upwelling actually occurs. They are available, however, only under cloud free conditions.

3.0) UPWELLING EVENTS

3.1) Background

I define an upwelling "event" as a time series of images showing the development of a region in which the surface water became cooler than the surrounding surface water and then gradually returned to 'normal'. Two "events", one in July 1988 and one in August 1989 (referred to as event 1 and event 2 respectively), were found from searching through the available images in the U.B.C. Satellite Oceanography and Meteorology Laboratory. These images, chosen for further analysis, were geometrically corrected (navigated) and
Figure 5: Upwelling index values for all the stations along the west coast of North America
processed, using a program developed to highlight and calculate the area of any possible upwelling event.

The upwelling index is only calculated at two stations near the west coast of Vancouver Island. The stations, located at Cape Flattery (48N, 125W) and just south of the Queen Charlotte Islands (51N, 131W), were examined to find out when wind-driven upwelling would be most likely to occur off the west coast of Vancouver island. The upwelling index was found to be highest from 10th July to 7th August in 1988 as well as in 1989. The upwelling index for these two stations, during event 1 is positive from approximately April through September with peaks at station 5 during the weeks of June 4, and 25, July 30, August 20, and September 3, and 10 1989. Unfortunately, there are no cloud-free satellite images available at these times. Peaks at station 6 occur in the weeks of May 7 and 14, June 4, July 2 and 9 and between the period of July 25 and August 27 1988 (see figure 6 a,b)

3.2) Satellite Imagery

Satellite images for the period June to September of 1988 and 1989 were processed to determine whether any upwelling events could be detected within periods of positive upwelling index. It was impossible to obtain a reasonable time series (3-5 days of clear weather) for any winter images because of the excessive amount of cloud cover on the west coast of B.C. during the winter months. Images of upwelling events were obtained in the summer of 1988 and 1989 from the advanced very high resolution radiometer (AVHRR) on board the NOAA 9 and NOAA 11 satellites. The
Figure 6 (a): Coastal upwelling indices, daily and weekly means during Jan-Sept. 1988 at 48 N 125 W (NOAA/NMFS Pacific Environmental Group - Monterey, California)
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Figure 6 (b): Coastal upwelling indices, daily and weekly means during Jan-Sept. 1989 at 48 N 125 W (NOAA/NMFS Pacific Environmental Group - Monterey, California)
satellites are in a near-polar (98.89° inclination), sun-synchronous orbit 870 km above the earth. The digital telemetry was received by the U.B.C. Satellite Oceanography and Meteorology Laboratory, and could be processed in near real time or stored on tapes for later retrieval and processing.

The AVHRR is a scanning radiometer with five spectral bands:

<table>
<thead>
<tr>
<th>Band</th>
<th>Wavelength Range</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Band 1</td>
<td>0.58 - 0.68 mm</td>
<td>(visible)</td>
</tr>
<tr>
<td>Band 2</td>
<td>0.725-1.10 mm</td>
<td>(near infrared),</td>
</tr>
<tr>
<td>Band 3</td>
<td>3.55 -3.93 mm</td>
<td>(thermal infrared)</td>
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<tr>
<td>Band 4</td>
<td>10.5-11.30 mm</td>
<td>(thermal infrared)</td>
</tr>
<tr>
<td>Band 5</td>
<td>11.5-12.50 mm</td>
<td>(thermal infrared)</td>
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</table>

The spatial resolution of the AVHRR is 1.1 km by 1.1 km at the nadir. The radiometer output is calibrated to degrees Celsius by using the internal temperature references of the satellite which are based on the assumption that the ocean surface radiates like a perfect blackbody. The satellite generates calibration data with every scan by aiming at a target of pre-determined temperature and at outer space (40K). A calibration program reads the first 50 lines of data to produce a calibration table for each orbit. An algorithm combining data from 2 or 3 channels could take advantage of different wavelength effects to partially compensate for atmospheric effects but this is not strictly necessary since, in looking for regions of upwelling, only the relative difference between the upwelled water and the surrounding water is needed. The resulting resolution in local temperature differences is approximately 0.1°C. Large absolute errors may still be introduced by absorption and re-emission by the atmosphere, refracted and scattered solar radiation (clouds) and sub-pixel
(i.e. $\leq 1.1$ km) clouds which can contaminate the observed radiance. Variations in atmospheric moisture may also introduce error of the order of 1 degree which could be removed by comparing the observed sea surface temperature with in-situ data. We are counting, however, on the ability of the imagery to resolve relative sea surface temperature differences.

In order to use satellite imagery for detecting upwelling events it is necessary to obtain a cloud-free time series of images showing upwelling. Off the west coast of B.C., however, this is a major problem. Only 5 images out of a total of 300 available, from 1986 to 1990, could be used because there is so much cloudiness in that area.

3.3) Image Processing

To obtain images containing an upwelling event, a series of steps were followed. First, a NOAA AVHRR image was loaded from tape and displayed on a color monitor. It was then determined, by visual inspection, whether the image was worth processing. Only images with a low cloud content ($< 20\%$) were selected for further processing.

An upwelling identification program was developed for this thesis to 'process' the image so as to detect and enhance cold water areas. The program has the capability of performing any combination of the following operations:

i) high and low frequency filtering,
ii) Laplacian edge enhancement,
iii) Sobel nonlinear edge enhancement,
iv) Roberts edge detector,
v) percentage linear contrast
vi) thresholding,
vii) land masking, and
viii) image reconstruction.

Various combinations of the above procedures were used to find out techniques which would produce the best 'visual' images, sharply identifying cold water areas.

The images were first corrected for the earth's rotation and sphericity and registered to a common Mercator grid (centered at lat. 49.5 N, long. 128.0 W) by linear interpolation. The position accuracy is approximately ±2 km after navigation. The surface temperature is then obtained by using an empirical formula, developed by Stewart (1985). An unprocessed navigated image, July 17 1988 (image number n9.18516), is shown in figure 7.

The next step in processing the image is to cloud mask and land mask the image to reduce the amount of data. I experimented with a variety of image analysis techniques to identify and quantify upwelling areas. I found that cold water areas are best highlighted by first applying a low pass filter. This image 'smoothing' removes periodic 'salt and pepper' noise but blurs the image, especially at the edge of areas such as thermal fronts where there is a sharper contrast between cold water and surrounding warmer water. To reduce this blurring an unequal-weighted smooth mask was used, defined by the following weights:
Figure 7: NOAA satellite image (raw data)
The new value of the central (c5) pixel becomes (see figure 8):

\[
BV_{5,\text{out}} = \text{int} \left( \frac{1}{9} \sum_{i=1}^{9} c_i BV_i \right)
\]

--- (5)

where \(BV_i\) is the brightness value (0-255) of the original image, and \(BV_{5,\text{out}}\) is the new or processed brightness value (0-255) at the central point (with weight c5).

The logic of the 3x3 low pass filter is shown in figure 8.

\[\text{Figure 8: Logic of a 3 x 3 low frequency filter, the numbers represent brightness values (0-255) (adapted from Jensen (1986))}\]
Using a 3 x 3 kernel results in the output image being two lines and two columns smaller than the input image. This problem was overcome by replacing the average brightness values at the borders with the average values within 3 pixels of the border.

After the low pass filter was applied, "thresholding" was used to isolate the upwelling region. The lower and upper thresholds are defined repectively as:

\[
TL = \frac{512}{512} \left( \frac{\sum_{ij} (BV_{ij}/n)}{\sum_{ij} (BV_{ij} - \bar{BV})^2} \right)^{0.5}
\]

\[
TH = \frac{512}{512} \left( \frac{\sum_{ij} (BV_{ij}/n)}{\sum_{ij} (BV_{ij} - \bar{BV})^2} \right)^{0.5}
\]

Where:  
- TH is upper threshold  
- TL is lower threshold  
- n is 2500 (number of pixels in a 50 by 50 sub-area)

The threshold was applied so that only values between TL and TH were kept. The 50 by 50 sub-area was used to prevent errors arising from the presence of large scale regional gradients which may occur in the image. This results in an image containing various non-connected areas ("blocks"), as seen in figure 9. Figure 9 shows one main area of cold water off the west coast of Brooks Penensula, and several scattered areas along the south west coast of Vancouver Island. The scattering of the cold water areas is due to clouds covering part of the area. As a result of the nature of the filtering and thresholding procedure a cold water area may be broken up into several neighbouring blocks. The cold water areas may be made to include any number of neighbouring blocks by varying a
Figure 9: NOAA satellite image showing results of filtering and thresholding
program parameter. This allows blocks which are separated by a small distance (e.g., 2 km) to be included in the same area of a potential upwelling event.

The last step in processing the image is to identify cold water areas by surrounding them with a colored (blue) border, and to reconstruct the image. The area of the potential upwelling event is calculated and displayed in the upper right part of the image.

The final result of applying this procedure to the images of 1988 and 1989 was to characterize two upwelling events during the period of August 6 to 8, 1989 and July 31 to August 4, 1988. The final processed images for event 1 and event 2 are shown in figures 10 and 11; they include estimates of the area of each cold water pattern.

There are many more upwelling events, as indicated by the upwelling index, but the climatic conditions made it impossible to get a cloud-free time sequence of upwelling at other times in these years.

3.3) Wind Observations.

Surface winds for the periods of observed upwelling events were obtained from surface pressure charts and lighthouses located along the west coast of B.C. (see figure 4). The surface pressure charts were issued every 6 hours and the lighthouse data were issued hourly from automatic stations, and every 6 hours from manned stations. Both sets of data were obtained from the Pacific Weather Center (Environment Canada). The
Figure 10 (a): NOAA satellite images of August 2nd 1988 (event 1)
Figure 10 (b): NOAA satellite images of August 3rd 1988 (event 1)
Figure 11 (a): NOAA satellite images of August 6th 1989 (event 2)
Figure 11 (b): NOAA satellite images of August 7th 1989 (event 2)
Figure 11 (c): NOAA satellite images of August 8th 1989 (event 2)
Figure 12: Event 1 wind (stick Diagram)
Cape Scott:
(50° 49' N, 127° 54' W)

(knots)

July 29  July 30  Aug. 1  Aug. 2  Aug. 3  Aug. 4  Aug. 5  note July 31
data missing

Cape St. James
(50° 47' N, 128° 26' W)

(knots)

July 29  July 30  Aug. 1  Aug. 2  Aug. 3  Aug. 4  Aug. 5  note July 31
data missing

Figure 12: (cont'): Event 1 wind (stick Diagram)
Figure 13: Event 2 wind (stick Diagram)
Figure 13 (cont'): Event 2 wind (stick Diagram)
wind velocities have been plotted for both events using six lighthouse stations (Cape St. James, Cape Scott, Kains Island, Sartine, Solander and Estevan Point) and are shown in figures 12 and 13.

3.4) **Temperature Observations**

No regular sea surface temperature measurements are available near Brooks Peninsula. The closest location for which a time series of temperature data is available is Kains Island.

The temperature observations are taken at the surface on the coast at noon each day. The data for the dates of the two upwelling events Aug 1989 and July 1988 are shown in Figure 14

There is no substantial drop in temperature during the upwelling events studied. Surface temperature observations at Kains Island are, therefore, inadequate to determine whether or not upwelling is in progress.

3.5) **Summary of Observations**

Two upwelling events, lasting between 2 and 5 days, have been observed by satellites designed to measure the sea surface temperature. Remote sensing is ideally suited for this task and is cost effective in covering large areas over a long period of time. From the NOAA images it appears that the upwelling off the northern coast of Vancouver Island is limited to about 40 km off Brooks Peninsula.
Figure 14 (a): Sea surface temperature plot for event 1

Figure 14 (b): Sea surface temperature plot for event 2
The first available image for event 1 is on August 2\textsuperscript{nd} 1988 (figure 10a). The cold water region is already very well developed, covering over 10,000 km\textsuperscript{2}. The cold water region extend from the north west tip of Vancouver Island to about 100 km south of Brooks Peninsula. The average temperature for the cold water region is 9.8°C, and the surrounding surface water is at approximately 14°C. The next day, August 3\textsuperscript{rd} 1988 (figure 10b), the area of the cold water region has increased by about 7 percent, but scattered cloud over the region makes it difficult to get an accurate measure of the cold water area. The average temperature of the area is 9.7°C and the surrounding water is still at 14°C. The extensive cloud cover before August 2\textsuperscript{nd} 1988 prevents the acquisition of a useful time series, showing the development of the cold water area for event 1.

The satellite images for August 6-8 1989 are much more successful in showing the development of a cold water region of the coast of Brooks Peninsula. Although there is extensive cloud cover in the western most part of the August 6\textsuperscript{th} satellite image (figure 11a), the area around Brooks Peninsula is cloud free. There does not appear to be any cold water regions around Brooks Peninsula at this time. A cold water region of 16 km\textsuperscript{2} appears southwest of Brooks Peninsula on August 7\textsuperscript{th} 1989 (figure 11b). The temperature of the water is approximately 10°C and the surrounding water is at 13°C. Another cold water region appears 25 km south of Estevan Point and covers 98 km\textsuperscript{2}. By August 8\textsuperscript{th} (figure 11c) the area off Brooks Peninsula has increased to 1338 km\textsuperscript{2}. The temperature has decreased to 9°C and the surrounding water is approximately 12°C.
The cold water region south of Estevan Point has increased slightly to 119 km².

After August 8th 1989 the images become too cloudy to be analysed. The satellite images for August 6-8 1989, however, are sufficient to show the initial stages of an upwelling event off Brooks Peninsula.

From the data it appears that the upwelling off Brook's Peninsula is induced by strong local winds blowing from the north west. These winds are observed by lighthouse stations along the west coast.

The temperature observations obtained from lighthouses along the coast do not appear to be useful in determining whether upwelling is in progress.
4.0) MODEL FORMULATION

4.1) Background

A numerical model was developed to explore the influence of nearshore bathymetry and coastline shape and to see if a simple model could reproduce the observations from satellite imagery.

Almost all numerical modelling of coastal systems makes use of layered models. Layered models allow the computational effort to be minimized and yet still retain some baroclinic resolution.

The first significant contribution to multi-layered coastal modelling was from O'Brien and Hurlburt (1972). Their model was a multi-layered, two-dimensional, nonlinear, time-dependent coastal upwelling model and neglected all alongshore variations since it was based on a cross-shelf transect.

A model without alongshore variation may exclude some basic dynamics of coastal upwelling since the omission of longshore derivatives excludes barotropic and baroclinic Kelvin and continental shelf waves. Kishi and Sugino-hara (1975) included both coastline geometry and bottom topography and found that longshore variation of the topography could cause strengthening or weakening in the process of generation and propagation of internal Kelvin and shelf waves.

From the early 70's up until the 80's regional studies of the effects of bottom topography were carried out. Peffley and O'Brien (1976) carried out a study of the Oregon shelf and Preller and O'Brien (1980)
studied upwelling on the Peruvian shelf. Topography was recognized as a major factor in determining the preferred locality for upwelling in these studies.

Hua and Thomasset (1983) studied the effects of coastline geometry on wind-induced upwelling in the Gulf of Lions and showed that preferred upwelling locations could arise from the coastline geometry alone.

Much more complicated models exist, such as O'Brien and Heburn's (1983) two-layer model which includes both heating and interfacial mixing, and the Heaps (1980) model which was hydrodynamically fully three-dimensional, although without thermodynamics. Since it is not possible to match the sophistication of these models with the facilities available, a simpler linear, two-layer model, including both bottom topography and coastline geometry at a 1 km resolution has been developed for this thesis.

4.2) Formulation of a 2-Layer Model.

This study requires a simple approach similar to that used by Cai and Lennon (1988). The viscous friction term, thermodynamic effects, nonlinear terms and mixing of heat and salt are expected to be small on the time scales involved and are, therefore, neglected. The depth-integrated equations over each layer are set up in a right-handed cartesian coordinate system. The x coordinate increases towards the south, the y coordinate increases towards the coast and the z coordinate increases vertically upwards. The origin of the coordinate system is set at (50° 19' N, 129° 14'
and at a reference level 2000 meters below the sea surface. A schematic diagram of the model geometry is shown in figure 15.

The depth integrated equations for the 2-layer, rotating, incompressible fluid, on the \( f = \) plane, are as follows:

**Layer 1:**

\[
\frac{dV_1}{dt} + k \times f V_1 = -g \ h_1 \nabla h_1 (h_1 + h_2 + D) + \frac{T_S}{\rho_1} - \frac{T_B}{\rho} - \frac{T_B}{\rho_2} + g' h_2 \nabla h_1
\]  \( \tag{8} \)

\[
\frac{dh_1}{dt} + \nabla \cdot V_1 = 0
\]  \( \tag{9} \)

**Layer 2:**

\[
\frac{dV_2}{dt} + k \times f V_2 = -g \ h_2 \nabla h_1 (h_1 + h_2 + D) + \frac{T_B}{\rho_2} + g' h_2 \nabla h_1
\]  \( \tag{10} \)

\[
\frac{dh_2}{dt} + \nabla \cdot V_2 = 0
\]  \( \tag{11} \)

Where: \( x, y \), are horizontal coordinates defined in figure 15. 
\( k \) is a unit vector in the \( z \)-direction (cf. figure 15)
\( \nabla \) is the horizontal gradient operator
\( V_1 = \int_{D + h_2 + h_1}^{D + h_2} v_1 \, dz \), and \( V_2 = \int_{D + h_2}^{D} v_2 \, dz \)
\( h_i(x, y) \) = instantaneous local thickness of the layers \( i = 1, 2 \)
\( \rho_i \) = the density of the sea water in layer \( i \).
\( \rho = \frac{(\rho_1 + \rho_2)}{2} \)
\( V_i \) = the transport vector of layer \( i = 1, 2 \), with components \( (U, V) \).
\( v_i \) = the velocity vector of layer \( i = 1, 2 \), with
components \((u, v)\).

\(f = \) the Coriolis parameter.

\(g = \) the acceleration due to gravity.

\(g' = g (\rho_2 - \rho_1)/\rho \) (reduced gravity)

\(D(x,y) = \) height of the bottom topography above the reference level.

\(TS = \) the surface stress vector with horizontal components \((TSx, TSy)\).

\(TI = \) the interface stress vector \(\rho CI q (v_1 - v_2)\), with components \((TIx, TIy)\)

\(q = (q_1 + q_2)/2\)

\(q_1 = (u_1^2 + v_1^2)^{0.5}\)

\(q_2 = (u_2^2 + v_2^2)^{0.5}\)

\(TB = \) the bottom stress, \(\rho_2 CB q_2 V_2\), with components \((TBx, TBy)\)

\(CB = \) drag coefficient for bottom stress.

---

**Figure 16**: Model Geometry Across the Continental Shelf
Solutions to equations (8)-(11), subject to appropriate boundary conditions, were obtained by using a centered differencing scheme and were evaluated on a staggered grid with a spacing of 1000 m (see figure 16). The leapfrog method was used to advance the solution in time using an interval of 10 seconds. Typically a one and a half day integration period was used, stopping when the interface surfaced. Positive elevation anomalies of the interface were taken to represent upwelling and negative anomalies, downwelling.

There is the problem of dealing with the interface height as it nears the surface \( h_1 \Rightarrow 0 \). Cai (1990) showed that model performance until the time of surface contact was equally acceptable both by simple hydrodynamics and by the more complex models which incorporate thermodynamical principles. By contouring the interface height before it surfaces, it is possible, therefore, to show favorable upwelling and downwelling regions. Our model is, therefore, capable of dealing with the initial stages of upwelling, within the limitations of a two-layer approximation.

4.3) **Boundary Conditions.**

Initially the heights \( h_j \) are taken to be undisturbed (i.e. horizontal) and the initial velocities are set to zero (i.e. no motion).

The surface area of the model is rectangular, with one side aligned with the coast. There are, therefore, one coastal boundary and three open
Figure 16: Staggered grid used in model
ocean boundaries. For the coastal boundary the normal velocity is set to zero. For the off-shelf boundary the heights \( h_i \) are taken to be undisturbed, and the transport vector obtained from equations (3) and (4). A Sommerfeld radiation condition is used for the north cross-shelf boundary to allow interior disturbances to propagate north, and a sponge layer is used for the south cross-shelf boundary to absorb any interior disturbances propagating southward.

4.4) Model Area.

The NOAA images show that in the presence of north-westerly winds, cooler upwelled water is present about 20 km off the coast of Brook’s Peninsula and extends over the continental shelf. The computing facilities available allowed only a small area to be used in the model. The area chosen included 225 km of the northwestern part of Vancouver Island from Estevan point (49°23' N 126°33' W) to Cape Scott (50°49' N 127°54' W) and extends about 100 km out to sea (see figure 17).

The coastal boundary axis is aligned with 315°T to be parallel to the coast. The model grid has grid square elements with 1 km sides and there are 250 elements in the x-direction and 200 elements in the y direction. The total area covered by the model is \( 50,000 \) km\(^2\).

4.5) Bottom Topography and Coastline

The bathymetry off the west coast of Vancouver island is characterized by an extensive shelf, steep slope and broad continental rise
Figure 17: Model area and bathymetry used in model
Thomson et al. 1989). The continental shelf (shoreward of the 200m depth contour) has a maximum width of approximately 65 km at the southern end of Vancouver island and narrows to approximately 5 km off Brook's peninsula near the north end of the island (see figure 17). The depth contours parallel the coastline except at the shelf edge where there are several canyons. A steep slope centered along the 500 m depth contour approximately 10 km offshore of Brook's peninsula is expected to have a significant effect on the coastal upwelling regime. No extensive fishing banks exist north of Amphitrite bank, located north of Barkley sound (Thomson et al. 1989), but by identifying upwelling regions along the northern part of Vancouver Island it may be possible to find new fishing grounds.

The minimum depth considered was 50 m and this value was used everywhere within 4 km of the shore. The maximum depth considered was 2000 m. Any depths greater than this were set equal to a flat bottom with $H = 1000$ m since the gradient of the bottom topography oceanward of the 1000 m contour is relatively small. Based on Lane (1963), the upper layer thickness was taken, initially, at 20 m and a density of $1.024 \times 10^3$ kg/m$^3$ was chosen for the first layer of the model.

The bottom topography and coastline used in the model were digitized from the Natural Resource Map series (1:250,000). The map was digitized using 25 km$^2$ sub grids divided into 1 km squares. The grid was drawn such that the axis was aligned (315°T) to the nearest compass rose to help offset any errors due to the transverse Mercator projection. The digitized map was overlaid on the original maps (19318, 19316, 19308,
19306, 15798, 15796) and spot checks made to ensure the quality of the digitized map. The nature of the depth contours on the original maps permitted a spatial resolution of approximately 2 km.

The coastline is very irregular with many promontories and interconnecting inlets. Two of the most prominent features are Brook's peninsula and Hesqinent peninsula (Estevan point). Brook's peninsula is expected to have a significant effect on the coastal upwelling regime.

4.6) Wind Stress

The wind stress is considered to be the initial driving force for upwelling and takes the form already given in equation (4):

\[ \tau = \rho_a C_d |v| v \]

Where: \( \tau \) is the stress vector,
\( \rho_a \) is the density of air,
\( C_d \) is an empirical drag coefficient, and
\( v \) is the estimated wind vector near the sea surface.

The wind stress was computed from data obtained from lighthouse stations along the coast according to equation (4) and the resulting data rotated 45° to correspond to the x-y axis chosen for the model. The wind stress data were smoothed using a low frequency filter (12 hour moving average) to remove short period signals (see figure 18).

No offshore wind data were available, thus wind stress at locations offshore are obtained from interpretation of the surface pressure maps.
Figure 18 (a): Wind stress near the coast for event 1

Figure 18 (b): Wind stress near the coast for event 2
The main area of interest is within 40 km of the coast; therefore, the open ocean area serves only to produce realistic boundary conditions for the coastal zone.

The model is "run" using real data from the lighthouse stations located nearest to the upwelling area, at Solander (50°07' N, 127°56' W) on Brook's Peninsula, Sartine (50°49' N, 127°54' W), and at Estevan Point (49°23' N, 126°33' W), for both event 1 and event 2. The wind field is calculated by linear interpolation between lighthouse stations along the coast. The wind at the western most part of the model is set to zero since this area's only purpose is to supply realistic boundary conditions for the near shore area. The results of the model will be compared to the upwelling features observed in the NOAA satellite images (see figure 10 and 11).

4.6) Time Step

The time step (Δt) used in the model is constrained by the Courant-Friedrichs-Lewy linear stability criterion (CFL) and was chosen such that:

\[
Δt \leq \frac{\min(Δx, Δy)}{\sqrt{2(g \max(h_1 + h_2))}}^{0.5}
\]  

The CFL criterion for this model is Δt \leq 13.1 s. A Δt of 10 s was chosen. The model takes about 30 minutes, to simulate 1.5 days, running on an IBM 3090.
4.7) **Summary of Parameters**

Below is a summary of the values of the parameters used in the model:

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air Density</td>
<td>1.2 kg/m³</td>
</tr>
<tr>
<td>Acceleration due to Gravity ((g)):</td>
<td>9.81 m/s²</td>
</tr>
<tr>
<td>Bottom Drag Coefficient ((CB)):</td>
<td>(1 \times 10^{-3})</td>
</tr>
<tr>
<td>Coriolis Parameter ((f)):</td>
<td>(9 \times 10^{-5}) s⁻¹</td>
</tr>
<tr>
<td>Density of water ((\rho)):</td>
<td>(1.024 \times 10^3) kg/m³</td>
</tr>
<tr>
<td>- Layer 1 ((\rho_1)):</td>
<td>(1.022 \times 10^3) kg/m³</td>
</tr>
<tr>
<td>- Layer 2 ((\rho_2)):</td>
<td>(1.026 \times 10^3) kg/m³</td>
</tr>
<tr>
<td>Grid Increment ((x\text{-Direction})):</td>
<td>1000 m</td>
</tr>
<tr>
<td>Grid Increment ((y\text{-Direction})):</td>
<td>1000 m</td>
</tr>
<tr>
<td>Interfacial Drag Coefficient ((C_I)):</td>
<td>(5 \times 10^{-5})</td>
</tr>
<tr>
<td>Initial Thickness of Layer 1 ((h_1)):</td>
<td>20 m</td>
</tr>
<tr>
<td>Initial Thickness of Layer 2 ((h_2)):</td>
<td>30 - 1000 m</td>
</tr>
<tr>
<td>Time Step ((\Delta t)):</td>
<td>10 s</td>
</tr>
<tr>
<td>Wind Stress Drag Coefficient ((C_d)):</td>
<td>(1 \times 10^{-3})</td>
</tr>
</tbody>
</table>
5.0 MODEL APPLICATION

5.1) Background

The model developed in the previous section was run using the bathymetry and coastline near the Brooks Peninsula area. The upward movement of the interface layer from its initial point will be used to identify areas where wind-induced upwelling occurs.

Several experiments were carried out using the model developed in the previous section. The experiments were designed to determine the effect of bottom topography, wind strength, and wind direction on the location, extent and time for upwelling to occur in the vicinity of Brooks Peninsula.

The model was run using the observed wind stress for the period July 29-Aug. 6 1988 (event 1) and Aug. 6-8, 1989 (event 2). Both the 'real' and 'flat' bottom topographies were used. To determine the effect of the coastline on the location and extent of upwelling the model was also run using 'half' of Brooks Peninsula.

The success of the model is assessed in terms of its ability to reproduce the upwelling features observed in the satellite images, including areas of cold waters and timing of their appearance.
5.2) **Bottom Topography**

It was possible to evaluate the significance of the bottom features by running the model with a modified geometry simulating a flat bottom. The initial thickness of the two layers were defined by $h_1 = 20 \text{ m}$, and $h_2 = 280 \text{ m}$, in which a total depth of 300 m is regarded to be sufficient to include the continental shelf as well as most of the near-shore topographic features.

The bottom topography has little effect on the location or extent of the upwelling, as illustrated in figure 19. The upwelling time is measured by the upwelling time taken for the interface ($h_1 \rightarrow 0$) to surface at the coast (point P in figure 17). The upwelling time is increased slightly if a flat bottom is used (figure 21). The "upwelling time" for the flat bottom experiment was 35.15 hours compared to 37.39 hours for the 'real' bottom experiment.

5.3) **Wind Stress**

The model was initially run using the real bottom topography, and a constant, uniform wind. The wind was from the on-shore at 9.65 m/s, 25.6° relative to the coast (see figure 20 a). The resulting interface height after 1.4 days is plotted in figure 20(c).

The model was run again using a wind speed of 6.82 m/s and from the same direction (see figure 20 b) to determine the effect of the magnitude of the wind stress on the development of the interface. The resulting interface height, after 1.4 days is shown in figure 20 (d).
Figure 19: Effect of bottom topography on upwelling time.
Figure 20: Effect of wind stress on upwelling time.
The upwelling and downwelling in figure 20 occurs in the same place but for the reduced wind stress the interface only reached a height of 6 m, compared to 14 m for the 'full wind stress case. It took the full wind stress case 37 hours to surface whereas the half wind stress case took 57 hours (see figure 21). This indicates that the wind strength is of paramount importance in producing the upwelling phenomenon.

5.4) Wind Direction

The importance of wind direction was tested by running the model with the same wind strength but varying the angle relative to the coast. Real bottom topography was used together with an artificial wind stress time series in which the magnitude increases smoothly from 0 to 10 m/s in 12 hours and remains constant. The upwelling time at point P (see figure 17) for a number of angles ranging from +26° (onshore) to -26° (offshore) are shown in figure 22. The maximum upwelling time occurs when the wind is 6.6° offshore. This is consistent with previous models (O'Brien et. al., 1972 and Cai, 1990) which obtain an angle of approximately 6°.

5.5) Model Results

5.5.1) Event 1

The model is started on July 29th at 0:00 hours (12 midnight on July 28th). The winds, in the vicinity of Brooks Peninsula, start 45° offshore at approximately 10 m/s, and remain constant for the first 8 hours (see
Figure 22: Plot showing the effect of wind direction on upwelling time.
Figure 23 (a): Wind stress, near Brooks Peninsula, used in model run on July 29 and 30 1988

Figure 23 (b): Development of interface height at point P over 40 hours
The interface does not show any noticeable response during the first 8 hours (see figure 23 b). The winds become light and variable for the next couple of hours. The winds gradually increase, becoming slightly onshore, to 50 km/hr for the end of July 29th and the first couple of hours of July 30th. The interface height increases to approximately 2 meters during this time. The winds change direction becoming offshore, and decreasing to 30 km/hr by mid-day on July 30th. Towards the end of July 30th the winds are alongshore and reach 50 km/hr. The interface increases rapidly to 15 meters during the first 12 hours of July 30th and reaches a height of 20 meters at 1:10 pm on July 30th 1988.

The upwelling region, off point P, is limited to within 15 km of the coast (see figure 24(a)). The upwelling occurs in a region in which the bathymetry drops off sharply from 0 to 1200 meters (see figure 24 (b)). The upwelling regions remain essentially the same over 40 hours; only the magnitude of the interface height changes.

Downwelling occurs over a 30 km region and is limited to a region within 40 and 70 km of the coast. The downwelling region occurs seaward of a 600 m deep ridge. The bathymetry drops off sharply in this area, changing 1200 meters over a distance of 10 km (see figure 24 (b)). The downwelling region shifts slightly shoreward, but the general location does not change significantly over the course of 1.5 days (see figure 24(a)).

The development of the upwelling region over a 1.5 day period is shown, for event 1, in figure 25 in 3 periods 4.8 hours apart. The figure covers a 64 by 64 km area, showing the area around Brooks Peninsula.
Figure 24 (a): Interface height at point P at 28.8, 33.6, and 36.0 hours along a transect oceanward of point P (event 1)

Figure 24 (b): The bathymetry along a transect oceanward of point P
The interface height is indicated by a series of color values shown in the legend, the dark blue color indicates regions where the interface height has decreased from its initial zero value, indicating regions of downwelling activity. The series of lighter colors, yellow to red, show regions where the interface height has increased from the initial zero value. The color red indicates regions in which the interface is within 2.0 meter of the surface, indicating very strong upwelling regions.

The interface height at 04:48 am on July 30th, 1988 is shown in figure 25 (a). There is no significant development of the interface height up to this time. I have defined the upwelling anomaly as the region in which the interface height is greater than zero (i.e. initial condition). It can be seen from figure 25 (a) that the upwelling anomaly covers an 8 km by 4 km region and a small downwelling region is located at point D. Uniform weak downwelling occurs about 20 km off the west coast of Brooks Peninsula and about 40 km from the rest of the coast.

At 09:36 am on July 30th, 1988 secondary interface anomalies appear at 'jagged' edges along the coast (see figure 25 b). The most significant development occurs just south of Brooks Peninsula. The interface anomaly has increased to a 20 km by 10 km region extending from just west of Brooks Peninsula to about 10 km south of the peninsula. The downwelling at point D has been replaced by an upwelling region extending just south of point D. Uniform downwelling occurs about 20 km off the west coast of Brooks Peninsula and has increased in intensity from the previous image (figure 25 b)
Figure 25 (a): Contour map of interface at 04:48 on July 30th 1988.
Figure 25 (b): Contour map of interface at 09:36 on July 30th 1988.
Figure 25 (c): Contour map of interface at 14:24 on July 30th 1988.
At 2:24 pm on July 30th, 1988 extensive upwelling is occurring along the coast with the most intense area south of Brooks Peninsula (see figure 25 c). A strong uniform downwelling region occurs about 18 km off the west coast of Brooks Peninsula. The intense upwelling occurs in a 20 by 20 km region along the west and south coast of Brooks Peninsula. A secondary upwelling region is located at point D. A small downwelling region occurs at point E about 10 km south of Brooks Peninsula and covers only a 1 by 2 km region.

5.5.2) Event 2

The model is started on August the 6th at 0 hours (12 midnight on August 5th). The wind stress, in the vicinity of Brooks Peninsula is fairly constant for the first 14 hours (see figure 26). The winds are blowing offshore with a wind speed of approximately 25 km/hr. This wind stress does not cause a significant change in the interface height, which increases only by about 0.8 meters during the first 14 hours. During the next 8 hours the wind increases to about 54 km/h and starts to parallel the coast, as indicated by the near zero value of the y-component of the wind stress (TSy). The wind remains constant at 54 km/h parallel to the coast, for the last four hours of August 6th and for the first 6 hours of August 7th. The interface height begins to increase and reaches 2 meters by the end of August 6th. Although the wind changes direction to slightly offshore, the wind speed remains essentially constant at 45 km/h. The interface height shows a dramatic increase of 18 meters and reaches a height of 20 meters at 10:06 am on August 7th, about 3 hours earlier than it did in event 1.
Figure 26 (a): Wind stress used in model run on August 6th and 7th 1989 (event 2).

Figure 26 (b): Development of interface height at point P over 36 hours.
Figure 27 (a): Interface height at point P at 26.4, 31.2, and 33.6 hours along a transect oceanward of point P (event 2)

Figure 27 (b): The bathymetry along a transect oceanward of point P
The interface surfaces more quickly for event 2 because the wind stress is larger.

There is not a significant difference in the distribution of upwelling and downwelling areas off point P, between event 1 and event 2. The upwelling is limited to within 15 km of the coast (see figure 27 a), and remains essentially the same over 1.5 days. Only the magnitude of the interface height changes. The downwelling occurs in the same 30 km region as event 1.

The development of the upwelling region over a 1.5 day period is shown for event 2, in figure 29 in 3 periods 4.8 hours apart. The figure covers a 64 by 64 km area, showing the area around Brooks Peninsula.

The interface height at 00:00 August 7th is shown in figure 28 (a). There is no significant development of the interface height up to this time. Five hours later the interface anomaly begins to appear and covers a 4 km by 13 km region 2 km south of Brooks Peninsula, A small downwelling feature is located at Point D (see figure 28 a). Downwelling occurs uniformly about 20 km and more off the west coast of Brooks Peninsula.

At 09:36 secondary interface anomalies appear at 'jagged' edges along the coast. The most developed area appears just south of Brooks Peninsula, covering a 16 by 7 km region. The downwelling occurs uniformly about 18 km off the coast of Brooks Peninsula, while the downwelling at point D has disappeared.
Figure 28 (a): Contour map of interface at 00:00 on August 7th 1989 (event 2).
Figure 28 (b): Contour map of interface at 04:48 on August 7th 1989 (event 2).
Figure 28 (c): Contour map of interface at 09:36 on August 7th 1989 (event2).
The influence of bathymetry and coastal configuration on upwelling characteristics for event 2 were studied by running simulations with a flat bottom and with a truncated Brooks Peninsula respectively.

In the flat bottom case a slight intensification occurs and the upwelling area is shifted slightly shoreward, while the downwelling is shifted offshore (see figure 28 c and 29). The bottom topography does not appear to have a significant effect on the development of the upwelling.

The model was also run with half of Brooks Peninsula missing to determine the effect of coastline on the location of upwelling (see figure 28 c and 30). The reduced coastline results in a much less intense upwelling region, covering a 4 by 4 km region south of the reduced peninsula. The downwelling occurs further offshore and is also less intense.

The coastline configuration appears to be a much more important factor than the bathymetry in determining the location and extent of upwelling in the vicinity of Brooks Peninsula.

A profile of the x and y components of the current normal to the coastline and across the width of the continental shelf is shown in figure 31. In the upper level, an offshore current of approximately 32 cm/s is generated, whereas in the lower layer the direction is onshore and the balance is achieved in the thicker layer by the lower speed of approximately 0.5 cm/s. This situation is entirely compatible with the expectation of forcing by an alongshore wind stress in a rotating system.
Figure 29: Contour map of interface at 09:36 on August 7th 1989 using the flat bottom topography.
Figure 30: Contour map of interface at 09:36 on August 7th 1989 using the reduced coastline.
Figure 31: Velocity profile at point P for the x and y components of the two layers
Figure 32(a): Velocity vector field plot for layer 1 at 09:36 August 7th 1989.

Figure 32(b): Velocity vector field plot for layer 2 at 09:36 August 7th 1989.
The current is essentially in the direction of the applied wind; near the coast, a baroclinic jet is found. This jet is a common feature in the upwelling zone (O'Brien et. al. 1972). The velocity vector field for layer 1 and layer 2 are shown in figure 32.

6.0 COMPARISON BETWEEN MODEL AND OBSERVATIONS

The model predicts that the interface should surface at 1:10 pm on July 30th 1988, using the wind stress observed for event 1. It is very hard to confirm that this time is correct because of the relatively poor satellite coverage at this time. The first available image is on August 2nd, and shows a very well developed upwelling region around Brooks Peninsula. The satellite images can neither confirm nor disprove the model for event 1.

Event 2 (Aug 6,7,8 1989) is more useful in that there is good satellite coverage showing the development of the upwelling region surrounding Brooks Peninsula. There is no upwelling development visible in the August 6th satellite image, consistent with the model results. The upwelling appears on the August 7th satellite image at 1:44 pm and covers a 16 km² area. When the model is run with the wind stress associated with event 2 the interface height surfaces at 10:06 am on August 7th, which is consistent with the satellite observations.
The upwelling in the model appears west and south along the coast of Brooks Peninsula. The satellite images indicate that the upwelling is occurring just off the west coast of Brooks Peninsula.

It is difficult to compare the interface height plots (figures 25 and 28) with the pictures observed in the satellite images (see figure 10 and 11). The model is limited to the onset of upwelling and is useful for showing the period up until the time that the interface reaches the surface. The satellite images are more sensitive to the later stages of 'mature' upwelling and are useful in defining the spatial scales occurring some days after the surfacing of the interface. The model cannot reproduce the upwelling features observed in the satellite images. The model does, however, pinpoint the 'source' of upwelling regions quite effectively and in predicting the correct time scales (1.5 days) with respect to the onset of the wind stress.

The interface height at time $t_0$ was converted to a temperature field in order to provide a more complete comparison with the the satellite images. An initial temperature ($t_0$) field is obtained from the satellite images (figure 10a and 11a). The interface height is converted to an temperature field ($t_1$) by assigning a temperature to each interface value based on a typical temperature profile of the region (Lane 1963). The temperature field is moved according to the velocity field, and the moved temperature field ($t_2$) is linearly mixed with the initial field ($t_0$), producing a final temperature field ($t_3$). The interface height at $t_0 + \Delta t$ is then converted to a temperature and the procedure is repeated, with the mixed temperature field ($t_3$) replacing the initial field ($t_0$). A flow chart
Initial temperature field (use satellite data)

Model results

Read in interface height

Read in velocity field

Convert interface height to temperature => t1

Move temperature field t2 = f(t1,u,v)

Mix moved temperature with initial temperature t3 = (to + t2)/2

Set: to = t3

Save temperature field to

Figure 34: Flow chart showing logic of temperature conversion program.
of this process is shown in figure 33. The temperature fields for event 1 and event 2 are shown in figures 34 and 35.

Event 1 shows a cold water region of approximately 30 by 30 km around the coast of Brooks Peninsula (see figure 35). The undisturbed water is at 13°C, and the cold water extends along the coast. The water gradually gets cooler closer to shore until it reaches a cold water core, of 9.7°C, extending 7 km offshore, and 15 km along the southern tip of Brooks Peninsula. It is difficult to compare the temperature field generated by the model to the satellite observations because the first available image is on August 2\textsuperscript{nd}, and the temperature plot is calculated for July 30\textsuperscript{th} (when the interface surfaces).

The temperature field for event 2 is more localized, the cold water area covers a 25 by 20 km region. The water is coldest just south of Brooks Peninsula reaching a temperature of only 9.4°C, and covering a 15 by 7 km area. The water along the coast is at approximately 11.7°C (see figure 35). The major difference between the model results and the temperature field is that there does not appear to be any cold water along the south-western part of Brooks Peninsula in the satellite images. This may be due to the coastal current transporting the cold water oceanward to the north-western tip of Brooks Peninsula (see figure 4b). By adding the coastal current into the model the results might more closely resemble the satellite observations.
Figure 34: Temperature field at 14:24 July 30th 1988 (event 1)
Figure 35: Temperature field at 09:36 August 7th 1988 (event 2).
7.0 CONCLUSION

Time-dependent, wind-driven upwelling along the west coast of Vancouver Island has been studied using NOAA AVHRR satellite images and a layered 2-dimensional numerical model. The satellite images were processed to highlight cold water regions along the west coast of Vancouver Island. The area and average temperature of the regions were calculated, and later compared to the results of the computer model. The model was able to pinpoint the 'source' of upwelling regions along the coast quite effectively, but could not reproduce the results of the satellite images since these images are more sensitive to the later stages of 'mature' upwelling. The model, while having some limitations, has proved to be successful in reproducing the correct time scales during two upwelling events. The model has also been used to identify sites favourable to upwelling in comparison to features evident in the AVHRR images.

The model indicates that the upwelling occurs after approximately 30 hours, and is limited to the area around the south coast of Brooks Peninsula. The model results are consistent with the observations made using the NOAA satellites. The velocity profiles indicate that the upwelled water comes from within a few kilometers of Brooks Peninsula. The upwelling Circulation is essentially restricted to the continental shelf area.

The model was run using various configurations to determine the effect of bottom topography, coastline, and wind direction on upwelling in the Brooks Peninsula region. Variations in coastline, rather than bottom topography, were found to be the dominant factor in determining sites
susceptible to upwelling. The maximum upwelling time at the coast was shown to occur with a wind direction of 6.6° offshore.

To eliminate the problem of cloud cover obscuring possible upwelling events satellite altimeter data could be used. The altimeter data might be used to measure the sea level height, and upwelling regions would be identified by the depressed areas along the coast.

The model focuses upon the processes which occur during the onset of upwelling. The model run, however, terminates when the interface reaches the surface, after approximately 1.5 days, and hence precludes the study of subsequent upwelling features. A more detailed model would be required to examine the later development of upwelling and hence be able to compare the model results to satellite images. The more detailed model would include the local current field, interfacial mixing and thermodynamics similar to that of Cai's (1990) model.

There does not exist, at the present time, a local measure of the upwelling time along the west coast of Vancouver Island. Satellite imagery coupled with a more detailed numerical model might be able to provide a more quantitative insight into our understanding of the local marine dynamics of the coastal waters and also in its role of provision of nutrients to the local coastal fishery.
7.0 Bibliography


