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Date **August 27, 1996**
Abstract

Short-term variations of sea surface temperatures (SST) over Queen Charlotte Sound have been poorly understood, mainly due to the lack of data, and therefore hardly predictable. A potentially important consequence of SST variations is the choice of salmon homeward migratory route, which has a significant impact on commercial fisheries. Until recently, predictions of fish migration routes have been made by using SST data at Kains Island, one of the lighthouse stations at the northern end of Vancouver Island. Since the early nineties, AES buoy stations have provided a new set of hourly SST in offshore waters, which may be a better representation of the fish marine environment. This thesis is using visual inspection, statistical analysis and AVHRR satellite imagery to show that the SST at Kains Island do not represent those over the main portion of the Queen Charlotte Sound, but only the SST within 20 km to 30 km from the coast. The SST at the buoy 46207 gives a better representation of the area. Furthermore, the most significant SST variations are caused by upwelling associated with an offshore high pressure system and a lee trough along the west coast of Vancouver Island.
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Chapter 1

Introduction

The subject of this thesis is a study of the sea surface temperature (SST) off the northern end of Vancouver Island. This work is a contribution to a project on the Canadian west coast sockeye salmon behavior called: "Biophysical Controls of Salmon Migration and Production". This project seeks to advance the capability to predict the migration route and run-timing of adult sockeye salmon returning to the Fraser River by looking at the influence of oceanographic variables and how they are influenced by the weather.

1.1 Motivation for this work

The sockeye salmon spawn in the Fraser watershed, enter the marine environment in the spring of their second year, spend the third year into the open ocean and then return as mature fish in the summer of their 4th year (Hamilton, 1985). Despite the regularity of this cycle, uncertainties in homeward migration still remain. On their journey back to the spawning ground, the sockeye salmon have the choice between two different routes when they reach the northern end of Vancouver Island: the passage through Johnstone Strait and the Strait of Georgia or along the West coast of Vancouver Island and through Juan de Fuca Strait. The percentage of sockeye salmon migrating via the northern route is referred to as the northern diversion or Johnstone Strait diversion and has a significant impact on the commercial fishery. Time series of the northern diversion from 1953 to 1994 show a variation between 2% and 80% (Fig. 1). Before 1978, values were relatively
uniform with a mean diversion rate near 17%. However, after 1977, diversion rates were much larger with inter-annual fluctuations.

Over the past few decades, more research has been done to find the mechanisms responsible for the northern diversion's fluctuations. Water temperature, salinity and currents were found among the most important physical environmental parameters to determine the fish migration routes over the northeast Pacific due to their impact on swimming effort, availability of food and predation. In 1960, Tully et al. proposed a correlation between the northern diversion and warm water advection from the south, especially following an El Niño event. In this case, more fish tended to use the northern route due to either temperature preferences or food distributions. During the following year, Favorite (1961) postulated a correlation with an area of low sea surface salinity extending seaward from the Queen Charlotte Sound, which led Wickett (1977) to make a correlation with the springtime Fraser River discharge for the period 1953-77. He concluded that the proportion of Fraser River water discharged into the ocean northwest of Vancouver Island increased the rate of Fraser sockeye migrating through Johnstone Strait.

Thereafter, Mysak (1986) used long-term sea surface temperature observations to suggest a correlation between the northern diversion and long-term environmental trends. In particular, significant changes in the atmospheric circulation were found during ENSO events due to large scale re-distribution of the atmospheric pressure, temperature and humidity field. Furthermore, such changes were generally reflected in high sea level, high sea surface temperature and low salinity along the west coast of British Columbia, especially just after the mature phase of strong ENSO events, which then had an impact on fish behavior. Indeed, large catches of B.C. sockeye salmon occurred in several ENSO years (Mysak, 1986). Hamilton (1985)
found that the northern diversion is better correlated with long period SST trends rather than with SST at the time of the return migration. This result indicates a connection with the overlying atmospheric circulation. Then, Emery and Hamilton (1985) suggested a correlation between warming events over northeast Pacific and the northward transport in the oceanic surface layer induced by the surface wind stress.

Further analysis of the northern diversion indicated that the SST at Kains Island was found to be the best predictor for the period 1978-1983 (Groot and Quinn, 1987). Later, Xie & Hsieh (1989) developed a non-linear regression model for predicting the northern diversion rate using the SST at Kains Island and the Fraser river runoff. Their model works relatively well, however, further improvements to salmon migration route forecasts may be possible by including offshore SST data, the latter being more representative of the sockeye salmon's marine environment than coastal ones.

1.2 Objectives

The principal objective of this thesis consists of an examination of a new set of data provided by AES buoys to gain a better understanding of the SST distribution off northern Vancouver Island. The selected buoys, covering offshore waters over Queen Charlotte Sound (Fig. 1), have provided SST data since 1990. Also, a comparison is made with coastal data provided by selected lighthouse stations and available AVHRR satellite imagery.
A second objective of this thesis is to explain the daily SST variations. LeBlond & Thomson (1990) indicated in their study on the North Pacific Ocean Environment that surface waters off the coast of British Columbia are mostly influenced by local processes such as local winds, solar radiation and upwelling. While the wind forcing and induced upwelling are the most efficient mechanisms at changing the temperature characteristics of coastal water, the variations in the solar radiation may also have some impacts. Therefore, SST, wind and cloud cover time series are studied together. Following Emery and Hamilton (1985) results, the SST are also compared with synoptic weather maps of the northeastern Pacific.

1.3 Plan of the Thesis

The thesis is organized as follows:

Chapter 2 contains a description of the study area from previous studies. It includes bottom topography, temperature and salinity characteristics, water currents and meteorological forcing. Chapter 3 is a detailed description of data over the study area, including results from visual inspection of time series, data source and processing. Chapter 4 presents results from statistical analysis including correlation and regression analysis. Chapter 5 describes the background theory related to factors influencing SST including the conservation of heat energy and the oceanographic influences on SST variations such as upwelling, turbulent mixing and horizontal advection. Chapter 6 is a comparison between windstress and SST along with synoptic maps and associated weather patterns. Chapter 7 presents results from a satellite remote sensing analysis including oceanographic features and transects of
sea surface temperature over the Queen Charlotte Sound/northern Vancouver Island area. Finally, a conclusion is presented in chapter 8.
Figure 1. a) Proportion of adult sockeye salmon returning to the Fraser River via the northern route (through Johnstone Strait) (data from IPSFC). b) Schematic representation of Migratory routes of adult sockeye salmon around Vancouver Island (From Groot and Quinn, 1987).
Chapter 2

Description Of The Study Area.

The study area covers the Queen Charlotte Sound and northern Vancouver island including the following three lighthouse stations: McInnes, Egg and Kains islands and the following four buoys stations: 46207 (East Dellwood), 46208 (West Moresby), 46204 (West Sea Otter) and 46185 (South Hecate) (Fig. 2).

2.1 Background

2.1.1 Bathymetry (Fig. 3)

The Queen Charlotte Sound is a broad area between Vancouver Island and the Queen Charlotte Islands. It lies east of the Pacific ocean and the continental rise and west of the mainland of British Columbia. The bathymetry of the Queen Charlotte Sound is relatively complex due to extensive shallow areas and deep trenches reaching 350-400 m deep into the continental shelf. The main bathymetric features are the following:

- Three shallow banks: Cook, Goose Island and Middle banks.

- Goose Island trough, a deep trench lying between Cook and Goose Island banks and extending towards the Northeast Pacific between Vancouver Island and the mainland.
• Mitchells trough extending between Goose Island and Middle banks.

• Moresby Trough running from Cape St. James to Banks Island.

Numerous coastal indentations and seaways close to the main channels, especially the Goose Island Trough, allow oceanic water to flow into the coastal regions.

Near northern Vancouver Island, the isobaths run generally parallel to the coast. The continental shelf, extending seaward to the 200m depth contour, has a width of approximately 20km near Cape Scott and narrows to approximately 15km at Kains Island. Beyond the shelf, the depth drops quickly to more than 1500m.

2.1.2 Tides

This thesis concentrates on the effect of wind-driven currents and, to some extent, on the heat budget, but does not consider the tides explicitly. Nevertheless, the tides are the most energetic phenomena over B.C. waters and should be discussed briefly to appreciate their effect on SST.

The waters in the vicinity of northern Vancouver Island and Queen Charlotte Sound are dominated by the semi-diurnal (twice daily) tide ($M_2$). The semi-diurnal tidal wave propagates around the northeastern Pacific in a counter-clockwise direction at a speed of approximately 200 m/s. The tidal currents are rotary in the open water of Queen Charlotte Sound and become rectilinear as they enter Hecate Strait. Typical peak $M_2$ speeds over that area are 20 to 40 cm/s (Crawford, 1994). During spring tides, maximum speeds are near 50 cm/s but decrease to
approximately half this value during neap tides. Near the shores, the tidal currents become aligned with the channels, and in many cases, the flow is accelerated as it enters the narrow mouth of these channels. The tidal range varies from approximately 2.4 m across the mouth of the Sound to around 3.0 m just north of Aristazabal Island, but may reach up to 5.0 m at the head of some channels. Figure 4 shows the co-range and co-phase values for the semi-diurnal tide (Thomson, 1981).

The tidal current can affect the SST by its friction on the bottom. The latter generates mixing in the water column and can destroy the stratification. As a result, the SST tends to remain more uniform and cooler in areas of strong tidal mixing compared with surrounding waters.

2.1.3 Surface Salinity (Fig. 5)

The water properties of the Queen Charlotte sound result from a combination of freshwater discharge from the coast and more saline offshore oceanic water. At the end of the winter, before the freshwater river discharge has started, the isohalines run generally parallel to the coastline with typical values near 31.5 along the mainland coast to 32.5 at the line joining the southern tip of the Queen Charlotte Islands to the northern end of Vancouver Island. In May, as the freshwater discharges begin, the salinity distribution shows the formation of a trough of lower salinities in the southeastern area near Calvert Island. This trough extends seaward just north of Vancouver Island. As the summer season progresses, the fresh water input increases with the most important discharge over the southeastern section. This is reflected in lower salinities of coastal waters and an intensification of the southeastern trough. Minimum values can be 2-3 lower in August than at the
beginning of the summer (Dodimead, 1980). Further south, surface waters along the west coast of Vancouver Island are more saline and show an annual cycle. A typical monthly summer value for Kains Island is near 32.0 in August, and can be 0.1-0.3 higher during an upwelling event. During the winter monthly sea surface salinity can be as low as 28.5 (Webster and Farmer, 1976).

2.1.4 Surface Water Temperature (Fig. 6)

The monthly sea surface temperatures vary mainly with the amount of incoming solar radiation and the input of fresh water runoff. The monthly average values range from approximately 8°C in May to 14°C in August. Records of sea surface temperatures (Dodimead, 1980) show a weak temperature gradient over the Queen Charlotte Sound in early summer. Maximum sea surface temperatures are generally found near Goose Island and the Middle banks. Jardine et al. (1993) looked at several years of AVHRR infrared imagery from the NOAA satellites and noticed an area of warm water developing late in summer over Goose Island bank. This can be explained by solar heating and weak tidal mixing preventing cold bottom water to rise to the surface. Just north of Vancouver Island, SST are lower due to fresh water discharge from rivers, especially the Bella Coola and Wannock rivers (Fig. 3) (Leblond et al, 1983). Then, a band of cool water has been observed off the northern tip of Vancouver Island which may be the result of horizontal advection of colder northern waters by the southward flow (Ikeda and Emery; 1984, Fang and Hsieh, 1993). However, from EOF analysis of AVHRR satellite imagery off Vancouver Island, Fang and Hsieh (1993) found their third mode indicating upwelling along the shelf break as another possible source.
Further north, a cold plume forms near Aristazabal Island, extends to the south, then west between Goose Island and Middle banks. Crawford et al. (1995) suggest that wind-driven upwelling may be responsible for this cold plume, while Jardine et al (1993) attribute some of the cooling to mixing. Just south of Cape St. James an area of cold and dense surface water is caused by intense mixing (Crawford et al, 1995).

2.1.5 Subsurface Temperature and Salinity

During the summer season, when the northwest component of the wind prevails, the relative offshore water mass transport increases near the surface over the Queen Charlotte Sound which triggers a compensating inshore movement of deeper waters and allows oceanic water to approach the coastal regions. Below approximately 50m depth, the salinity is everywhere greater than 32.8 and increases with depth to near 33.6 which is characteristic of the oceanic halocline.

Figures 7 and 8 show the vertical profiles of temperature and salinity at two different locations (stations A and C) within the Queen Charlotte Sound for the months of July, August and September. The station A is located on the western side of the study area near 129.5W/51.2N and represents water properties just off the Queen Charlotte Sound. The station C, located near 129.2W/51.9N, represents water properties in the centre of the Sound. From these graphs, the main features are a thin mixed or near-mixed surface layer and strong thermocline and halocline. Under weak surface mixing, the thermocline extends from near surface to a depth of 75-100 m at the station A and to 100-125 m at station C. The halocline extends from
near surface to 125-150 m at both stations. Below the thermocline and halocline, temperature decreases and salinity increases slowly with depth (Dodimead, 1980).

2.1.6 Currents

The general circulation over northeastern British Columbia consists of an eastward flowing current known as the Subarctic Current or West Wind Drift. This current crosses the Pacific Ocean between 45 and 50°N and divides near Vancouver Island. One part turns northward to become the Alaska Current and the second part shifts southward to form the California Current (Thomson, 1981) (Fig. 9).

Over the study area, the main dynamical forcing for currents are tides and wind systems, the latter being much more efficient at changing the temperature and salinity characteristics of the water (Freeland et al, 1984) through turbulent mixing and upwelling, but also less predictable. While turbulent mixing is not a dominant factor during the summer months due to weaker atmospheric circulation and wind forcing, upwelling favorable winds are often observed along the west coast of B.C. from April through the end of September.

The summer near-surface current along a line from the north end of Vancouver Island to Cape St. James flows southward as a result of dominant northwesterly wind forcing (Fig. 10). Freeland et al (1984) used a simple one layer model to show that the core of the southward current is the result of an interaction between variable bottom topography and the wind forcing, so that upwelling and cold waters should appear first near the shelf break, especially under northwesterly winds.
Over the Queen Charlotte Sound the surface flow is more variable as a result of frictional effects and there is a tendency for the flow to be parallel to the local isobaths. A complex bottom topography lies below a shallow layer of water resulting in a clockwise gyre around Goose Island bank which is thought to be caused by tidal rectification (Freeland et al, 1984). There is also strong residual tidal currents from the Sound near Cape St. James (Foreman et al, 1992). Freeland et al raised the importance of local winds at 2 to 30 day periods for surface currents.

2.1.7 Meteorological Forcing

a) Large Scale Pressure Patterns

The large-scale atmospheric circulation in the central North Pacific is determined by two major persistent centres of action: the Aleutian Low south and east of the Aleutian Islands and the sub-tropical high pressure of the North Pacific with the westerly Jet Stream running between them. Superimposed on this climatological pressure pattern are synoptic weather systems (highs, lows and fronts) responsible for relatively rapid wind fluctuations. The low pressure systems (lows and fronts) develop and travel along the westerly Jet Stream at a speed corresponding to approximately 50% of its speed. A typical weather cycle is near 2 to 3 days over northeastern Pacific. On the other hand, the high pressure systems (anticyclones) tend to be slower and may influence the weather for several days.
1. Fall/Winter Pressure Pattern

During the cold season, up to April, the Aleutian low lies near 45° lat. N/175° long. W and the North Pacific High, relatively weak, has its center near 30-40° lat. off the California coast (Emery and Hamilton, 1985) (Fig. 11). The mean position of the Jet Stream is near 50°N. This pattern produces prevailing southwesterly geostrophic surface winds from the subtropical Pacific Ocean into the Gulf of Alaska. The superimposed synoptic lows, most intense over this period, tend to remain well offshore and track into the Gulf of Alaska along the Jet stream, while the associated frontal systems move across the B.C. coast (Fig. 12). The intensity of the lows and winds is determined by the contrast between cold and warm sectors (temperature gradient), the amount of tropical or sub-tropical moisture being injected into the system and the relative stability of the airmass. These systems usually give southeasterly gale (34 to 47 knots) to storm-force winds (48 to 63 knots), and occasionally hurricane-force winds (64 knots and more) to the B.C. coast. However, when the dynamics allow a rapid deepening of the lows near the coast (more than 24 mb deepening at the centre of the low within 24 hours), east to southeast winds may occasionally reach 70 knots with gusts up to 100 knots (Marine Weather Hazards Manual, 1990).

2. Spring/Summer Pressure Pattern

As the spring arrives, the baroclinic zone retreats northward due to increased solar radiation over the northern hemisphere. The Pacific Northwest begins to feel the effects of the Californian high pressure system, which strengthens, moves further north and deviates the storm track into the northern Gulf of Alaska. The ridge,
extending over the Gulf of Alaska and off the B.C. coast, reaches its full development by the month of July. Meanwhile the Aleutian low has weakened significantly and retreated to the north so that it is no longer evident, especially in July (Fig. 13). From May to September, the main position of the Jet Stream is around 53-55°N. Since the contrast between warm and cold air is much smaller, the strength of the Jet Stream has decreased significantly from winter, leading to weaker storms moving over the B.C. coast. The offshore ridge tends to dominate the northeastern Pacific, which gives prevailing northwesterly winds along much of the B.C. coast up to gale-force strength. However, wind direction and speed fluctuate when an occasional front tries to penetrate through the ridge.

3. Deviation From The Main Pressure Patterns

In other situations, especially during the winter, an Arctic airmass may spread over the southern interior of the province and push very cold air down the mountain passes. This situation usually brings cold, dry weather over much of B.C. and produces strong winds (gale or storm-force) through the coastal inlets that may persist for several days.
b) Summer Regional Scale Pressure Patterns Over Queen Charlotte Sound

1. Lee Trough

During the summer, the usual pressure pattern is dominated by the offshore ridge and a trough of low pressure in the southern interior of British Columbia (Fig. 14a). The latter is a thermal trough which develops after many days of sunshine. In more detail, when the land becomes relatively warm over southeastern B.C., the overlying airmass gets lighter which triggers an upward vertical motion, divergence of air aloft and lower surface pressures. At high levels into the atmosphere, the divergent flow of air causes north-northeasterly winds which subside on the west side of the coastal ranges. Then, these subsiding winds converge with the outflow winds produced by the offshore high pressure system to form a trough of low pressure (called lee trough) just west of Vancouver Island (Fig. 14b). When the trough is relatively strong and the offshore ridge builds, a band of strong northwesterly winds develops over the water just behind the trough. Strong winds, rising at times to gale force, are often observed just off the northern end of Vancouver Island. However, further south just along the coastline of west Vancouver Island, the winds remain generally light from the east at 15 knots or less. This pressure pattern may persist for many days until a low pressure system moves eastward from the Pacific, forcing the offshore ridge to get closer to the coast and cutting the heating source by spreading clouds. Under such circumstances, the lee trough breaks down and the northwest winds (if still present) become much weaker. The combined effect of a thermal trough and the offshore ridge is probably the most
suitable situation for upwelling-favorable wind with speed occasionally up to 65 km/hr.

2. Summer Fronts

When a frontal system approaches Queen Charlotte Sound, winds are generally from the south or southwest except southeast in the waters close to the coast because of the topography (Fig. 14c). Behind the front, as the ridge of high pressure rebuilds along the coast, the pressure may rise quite rapidly (0.5 to 1 millibars per hour) which produces strong northwesterly winds over the water. These Northwesterlies can be quite strong in the spring or early summer (May, June) as the fronts still have some of the strength of winter storms.
Figure 2: The northwest coast of British Columbia including the Queen Charlotte Sound. The study area lies within the dashed lines and the coastline. The buoys' positions are indicated by stars sign (*). (From Crawford et al, 1995).
Figure 3. Bathymetry of the northwest coast of B.C. including the study area. Depth are in metres. (From C. Hannah, 1992.)
Figure 4. Co-range and co-phase values for the semi-diurnal tide. Tidal range (broken line) in meters, tidal phase (solid line) in degrees. Difference of 29° corresponds to time difference of 1 hr. (From Thomson, 1981).
Figure 5. Salinity distribution (ppt) at 3 m depth a) for May 3rd-28th, 1954, b) for June 29th-July 22nd, 1954 and c) for August 17th-September 9th, 1954. The arrows indicate the direction of flows. (From Dodimead, 1980).
Figure 6. Temperature distribution in degree Celsius (°C.) at 3 m depth a) for May 3rd-June 20th, 1954, b) for June 29th-July 22nd, 1954 and c) for August 17th-September 9th, 1954. The arrows indicate the direction of flows. (From Dodimead, 1980).
Figure 7. Temperature and salinity structures for waters in Queen Charlotte Sound at station A. (Circled numbers indicate years of survey). (From Dodimead, 1980).
Figure 8. Temperature and salinity structures for waters in Queen Charlotte Sound at station C. (Circled numbers indicate years of survey). (From Dodimead, 1980).
Figure 9. Prevailing surface currents in the North Pacific (From Thomson, 1981).

Figure 10. July-August average near surface currents over the Queen Charlotte Sound and southern Hecate Strait. (From Crawford et al, 1995).
Figure 11. Mean sea level pressure pattern for January (pressure values are in millibars, mb). The arrows indicate winter storm tracks. (From Env. Canada, 1991).
Figure 12. Mean sea level pressure (mb) pattern associated with a deep winter low over the Gulf of Alaska with the related frontal system approaching the B.C. coast. (From Env. Canada, 1991).
Figure 13. Mean sea level pressure (mb) pattern for July. The arrows indicate summer storm tracks. (From Env. Canada, 1991).
Figure 14. Typical summer sea-level pressure (mb) pattern. a) Summer pressure trough over southern B.C. interior, b) lee trough (arrows and wind barbs indicate direction of surface winds. Wind barbs also indicate strength of winds). (From Env. Canada, 1991).
Figure 14. continued. c) Summer front.
Chapter 3

Description of Data

3.1 Data sources: Meteorological and Oceanographic Data.

3.1.1 Lighthouse stations.

Daily sea surface temperature observations have been made at different locations along the B.C. coast; data were provided by the Institute of Ocean Sciences. This thesis examines 5 years of data (from 1990 to 1994) collected from three lighthouse stations which are: McInnes, Egg and Kains Islands (Table 1, Fig 2). Daily observations are made within one hour before the time of high tide, light schedules and other duties permitting, and taken at a depth of 0.9 metre. The temperature is measured with a mercury thermometer with an accuracy of 0.3 °C. or better for individual measurements.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude N</th>
<th>Longitude W</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kains Is.</td>
<td>50°27'</td>
<td>128°02'</td>
</tr>
<tr>
<td>Egg Is.</td>
<td>51°15'</td>
<td>127°50'</td>
</tr>
<tr>
<td>McInnes Is.</td>
<td>52°16'</td>
<td>128°43'</td>
</tr>
</tbody>
</table>

Table 1. Lighthouse geographic locations.

There were only very few missing data (Table 2) that have been filled by linear interpolation. No spike was found since the data set has already been checked at IOS.
### Lighthouse station

<table>
<thead>
<tr>
<th>Lighthouse station</th>
<th>Year</th>
<th>Missing days (day number/date)</th>
<th>Number of missing data</th>
</tr>
</thead>
<tbody>
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<td>0</td>
</tr>
<tr>
<td></td>
<td>1991</td>
<td>none</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>1992</td>
<td>116 (August 24th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
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</tr>
<tr>
<td></td>
<td>1994</td>
<td>none</td>
<td>0</td>
</tr>
<tr>
<td>McInnes Isld.</td>
<td>1990</td>
<td>94 (August 02nd)</td>
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</tr>
<tr>
<td></td>
<td>1992</td>
<td>55-57 (June 24th-26th)</td>
<td>3</td>
</tr>
<tr>
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<td>1993</td>
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<td>0</td>
</tr>
<tr>
<td></td>
<td>1994</td>
<td>none</td>
<td>0</td>
</tr>
<tr>
<td>Egg Island</td>
<td>1990</td>
<td>6,13 (May 06th,13th)</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>151 (September 28th)</td>
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</tr>
<tr>
<td></td>
<td>1991</td>
<td>05 (May 05th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>118,121 (August 26th, 29th)</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>147 (September 24th)</td>
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</tr>
<tr>
<td></td>
<td>1992</td>
<td>10 (May 10th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>43,52 (June 12th, 21st)</td>
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<tr>
<td></td>
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<td>75 (July 14th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>102 (August 10th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>1993</td>
<td>67 (July 06th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>1994</td>
<td>17,27 (May 17th, 27th)</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>69 (July 08th)</td>
<td>1</td>
</tr>
</tbody>
</table>

Table 2. SST missing data.

#### 3.1.2 Buoy data

Four buoys from the Marine Weather Programs, Environment Canada - Pacific & Yukon Region, have been selected for the present analysis. They are located in the Queen Charlotte Sound, west of the Queen Charlotte Islands, off Cape Scott and over southern Hecate Strait (Table 3, Fig. 2).

<table>
<thead>
<tr>
<th>Station name</th>
<th>Station ident.</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water depth(m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>East Dellwood</td>
<td>C46207</td>
<td>50.8°N</td>
<td>129.9°</td>
<td>2125</td>
</tr>
<tr>
<td>West Moresby</td>
<td>C56208</td>
<td>52.5°</td>
<td>132.7°</td>
<td>3000</td>
</tr>
<tr>
<td>West Sea Otter</td>
<td>C46204</td>
<td>51.4°</td>
<td>128.8°</td>
<td>244</td>
</tr>
<tr>
<td>South Hecate St.</td>
<td>C46185</td>
<td>52.4°</td>
<td>129.8°</td>
<td>220</td>
</tr>
</tbody>
</table>

Table 3. Buoy geographic location.

32
Hourly buoy data used in this analysis are water temperatures, wind speed and wind direction. Water temperature is measured by a YSI 703 Linearized thermistor with an accuracy of 0.1 °C. Wind direction and wind speed are measured by a RM Young 5103 sensor with an accuracy of +/- 5 degrees and 0.6 m/s.

Weather and ocean information was not reported at the buoy stations on a regular basis during the study period, especially for the first two years. Also, hourly water temperature below 5 °C and above 18°C have been considered as errors and removed from the data set. Gaps shorter than 10 percent of the time series have been filled by linear interpolation. Time series where gaps were between 5 and 10 percent must be interpreted with care. In some cases, the full time series are not available (see Table 4 for missing data). Missing data are relatively numerous, especially in 1990 and 1991, which does restrict the analysis.
<table>
<thead>
<tr>
<th>Buoy station</th>
<th>Year</th>
<th>Missing days (day number, date)</th>
<th>Nbr of missing days</th>
</tr>
</thead>
<tbody>
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<td>46207</td>
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<td>29 (May 29th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>103-107 (August 11th-15th)</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>62-72 (July 04th-11th)</td>
<td>11</td>
</tr>
<tr>
<td>1991</td>
<td>Series not available</td>
<td>too many</td>
<td></td>
</tr>
<tr>
<td>1992</td>
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<td>0</td>
<td></td>
</tr>
<tr>
<td>1993</td>
<td>none</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>1994</td>
<td>29 (May 29th)</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>71 (July 10th)</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>128-148 (September 05th, 25th)</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>46208</td>
<td>1990</td>
<td>Series not available</td>
<td>too many</td>
</tr>
<tr>
<td>1991</td>
<td>1-61 (May 01st-July 30th)</td>
<td>61</td>
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</tr>
<tr>
<td>1992</td>
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<td></td>
</tr>
<tr>
<td>1993</td>
<td>1-24 (May 01st-24th)</td>
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</tr>
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<td>1994</td>
<td>29 (May 29th)</td>
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</tr>
<tr>
<td>46204</td>
<td>1990</td>
<td>29 (May 29th)</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>58-72 (June 27th-July 11th)</td>
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</tr>
<tr>
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<td>103-107 (August 11th-15th)</td>
<td>5</td>
</tr>
<tr>
<td>1991</td>
<td>129-139 (September 06th-16th)</td>
<td>11</td>
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</tr>
<tr>
<td>1992</td>
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<td>0</td>
<td></td>
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<tr>
<td>1993</td>
<td>none</td>
<td>0</td>
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<tr>
<td>1994</td>
<td>29 (May 29th)</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>71 (July 10th)</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>82 (July 21st)</td>
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<tr>
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<td>128-148 (September 05th, 25th)</td>
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</tr>
<tr>
<td>46185</td>
<td>1990</td>
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<td>too many</td>
</tr>
<tr>
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<td>61 (June 30th)</td>
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<tr>
<td>1992</td>
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<td>1994</td>
<td>29 (May 29th)</td>
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<td>71 (July 10th)</td>
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<td>128-139 (September 05th, 16th)</td>
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</tr>
<tr>
<td></td>
<td>148 (September 25th)</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

Table 4. SST missing data.
Prepared hourly wind data have been obtained from Josef Cherniawsky, IOS. Hourly data obtained from the Pacific Weather Centre in Vancouver have been checked for suspicious records and gaps shorter than 2 days have been filled by linear interpolation.

3.1.3 Other data

Synoptic weather maps (12Z standard time) for northeast Pacific/western Canada and cloud cover data for three lighthouse stations (Cape Scott, McInnes Island and Cape St. James, See Fig. 2) have been provided by the Pacific Weather Centre, Environment Canada. Cloud cover data available every 3 hours, except every hour for Cape St. James, have been converted into daily data for three specific events analyzed in chapter 6.

3.2 Data descriptions

3.2.1 Temperature

3.2.1.1 Variation In Time.

Time series of daily sea surface temperature have been plotted for 5 consecutive summers (day 1(May 1st) to day 153(September 30th) from 1990 to 1994) for the following lighthouse stations: Kains, McInnes and Egg Islands, and buoy stations: 46207, 46208, 46204 and 46185 (Fig. 15).
A summer thermal cycle in the upper ocean has been observed by Webster and Farmer (1976) at these lighthouse stations and is evident from SST time series with the seasonal maximum SST reached in late summer. Time series show a temperature increase generally until approximately day 108 (mid-August) followed by a slow temperature decrease to the end of the season. The fact that the SSTs begin to fall when the sea is still gaining heat has been explained by advective and/or eddy processes (Tabata, 1958). The highest seasonal mean SST (14.0°C) was recorded in 1990 at the buoy 46204 while the lowest one (11.3°C) happened in 1991 at Egg Island (Table 5). From time series, daily fluctuations tend to have bigger amplitudes after day 75 (mid-summer). Pronounced variations occurred in 1992 after day 75 at the buoys 46185 and 46208, in 1993 after day 63 at the buoy 46185 and in 1994 after day 78 at the buoy 46185. The data have been checked, but no error has been detected. From seasonal mean SST, the warmest year appears to be 1994, while the coolest one corresponds to 1991.

A previous study has shown that during an El Niño event daily sea surface temperatures off the southern west coast of Vancouver Island may increase by as much as 3°C above normal (Freeland, 1990). However, the seasonal mean SSTs computed in this thesis for the years 1990 to 1994 show that the maximum values do not coincide with the last two El Niño events (1991 and 1992-93) at any of the stations.
<table>
<thead>
<tr>
<th>May 1st-Sept 30th.</th>
<th>T.Min. (°C)</th>
<th>T.Max (°C)</th>
<th>T.Mean (°C)</th>
<th>St.Dev. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Kains Island.</strong></td>
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<td></td>
</tr>
<tr>
<td>1990</td>
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<td>12.7</td>
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<td>1994</td>
<td>10.1</td>
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<td>13.4</td>
<td>1.9</td>
</tr>
<tr>
<td><strong>McInnes Island.</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1990</td>
<td>8.3</td>
<td>16.3</td>
<td>12.5</td>
<td>1.8</td>
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<td>12.8</td>
<td>1.6</td>
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<td><strong>Egg Island.</strong></td>
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<td></td>
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Table 5. Statistics for SST (minimum, maximum, mean and standard deviation for the summer).
3.2.1.2 Variations In Space.

Isotherms have been drawn from the mean SST distribution at every 0.5°C (Fig. 16) for each summer. The number of data points is certainly not sufficient to get detailed SST distributions. However the following features seem to emerge: A gradient of temperature perpendicular to the coast with relatively low values along the coast. The lowest SST are shown near Egg Island while the warmest are revealed by the buoys over Queen Charlotte Sound, especially at buoy 46207. Similar information can be seen from Table 5: Seasonal means and summer maximum SST are generally higher at the buoy stations than at the lighthouses. The summer maximum values were not observed simultaneously at all stations, although the seasonal cycle is assumed to be the same over the scale of the study area for any particular year. The difference is relatively small between Kains and McInnes Islands (1 to 5 days), but can be as much as 15 days between Kains and Egg Islands. The buoy stations show as many variations, except for 1993 when summer maxima happen within 3 days for all buoys. The standard deviation of SST varies from 1.0°C to 2.1°C. Egg Island shows the smallest variations around the mean temperature, while SST fluctuations are more pronounced at the buoy 46185.

3.2.2 Synoptic And Wind Analysis

Wind data at the buoy stations are available for the years 1990 to 1994. However due to numerous missing points, a relatively complete data set is only available for the last three years (1992, 1993 and 1994). The following analysis combines observations from time series and synoptic maps.
3.2.2.1 Variations In Time

The months of May and June (day 1 to day 61) are usually transition months between the winter wind regime and the summer one. During the first part of the summer winds are often light and variable. In the second part of the summer, when the high pressure system is well established over the northeastern Pacific, stronger northwesterly winds become dominant along the west coast of British Columbia.

a) Summer 1990:

The summer of 1990 is typical with variable winds from day 1 to approximately day 61 and mostly northwesterly winds after day 73 (Fig. 17a). The seasonal mean wind speed, averaged over data from two buoys (46204 and 46207), is near 5.2 m/s from the west-southwest and the seasonal maximum, recorded at buoy 46207 on day 32, is 13.1 m/s from the east-southeast (Table 6).

From synoptic maps (Fig 18a), a north-south ridge of high pressure located approximately 450 km offshore on day 69 (July 8th) gave rise to northwesterly winds over part of the area. By day 73 (July 12th) the Northwesterlies were established over much of the B.C. coast. On day 75 (July 14th) the pressure gradient strengthened over Queen Charlotte Sound as a result of a thermal trough over southern B.C. interior and eastern Washington State and a low pressure system approaching from the west into the Gulf of Alaska (Fig. 18b). Northwesterly winds reached approximately 10-13 m/s (20-25 knots) over Queen Charlotte Sound on that day. Thereafter, the pressure pattern was dominated by a thermal trough over southern B.C. interior and a ridge of high pressure off the B.C. coast except for a
front that moved across the area on day 77 (July 16th). As a result, northwesterly winds prevailed over Queen Charlotte Sound until day 93 (August 1st). Between day 94 (August 2nd) and day 114 (August 22nd), a series of frontal systems moving across the area gave variable winds. For example, Figure 18c shows a frontal wave just off the northern end of Vancouver Island moving eastward. The pressure gradient on that day (day 94/August 2nd) was relatively weak with southeasterly winds generally less than 10 knots (5 m/s) in the vicinity or the northern end of Vancouver Island and along the eastern boundary of Queen Charlotte Sound. Behind the wave, winds were west to southwest near 10 knots (5m/s).

The Northwesterlies were re-established along much of the B.C coast on day 115 (August 23rd) due to an offshore high pressure system (Fig. 18d). These winds were maintained until the end of the summer (day 153) with variable speeds generally below 13.0 m/s. However, winds were variable at times, especially at the end of August and the beginning of September, as two weak fronts moved across the B.C. coast.

b) Summer 1991:

For 1991, time series (available only from the buoys 46204 and 46185) show relatively strong winds oscillating between northwesterly and southeasterly during almost the entire season (Fig. 17b). The period of oscillation varied approximately between 4 to 12 days. The spatially averaged seasonal mean wind speed is around 4.9 m/s from the west-southwest and the maximum, recorded at the buoy 46185 on day 122, is 12.8 m/s from the south-southeast (Table 6).
The synoptic pattern for the summer of 1991 was slightly different from the seasonal mean: there were a few relatively strong frontal systems that moved across the B.C. coast during the summer which kept the Californian ridge further south and gave relatively strong southeasterly winds along much of the B.C. coast. The most significant frontal systems moved through the area during the following times: between days 9-16 (May 9th-May 16th), 39-42 (June 8th-June 11th), 60-64 (June 29th-July 3rd) and 118-128 (August 26th-September 5th). Between these systems, the ridge off California was able to build over the northeastern Pacific and a thermal trough to develop over southern B.C. interior and Washington State. Both systems were responsible for northwesterly winds along the B.C. coast, especially from day 17 to day 27 (May 17th to May 27th) and from day 104 to day 116 (August 12th to August 24th). In particular, Figure 18e shows a thermal trough getting established on day 17 (May 17th) while the offshore ridge is relatively strong near 140°W. Thereafter, the thermal trough deepened into the interior of B.C. and the ridge moved towards the coast so that the pressure gradient and the Northwesterlies increased over Queen Charlotte Sound. However, by day 27 (May 27th) (Fig. 18f) the ridge was forced onto the coast due to an approaching frontal wave from the west, which ended the northwesterly event.

A second northwesterly event began near day 104 (August 12th) as a thermal trough developed over southeastern B.C. (Fig. 18g). Meanwhile, the Californian ridge extended across the Queen Charlotte Islands. As a result, Northwesterlies were relatively strong over Queen Charlotte Sound and persisted for several days. However, by day 116 (August 24th), the Californian ridge was located further away offshore, the thermal trough was filling (Fig. 18h) and the Northwesterlies began to weaken. Shortly after (day 118/August 26th), a strong frontal wave moved across the Queen Charlotte Islands, followed by a series of waves.
c) Summer 1992:

During the summer of 1992, northwesterly winds were relatively frequent and also stronger (Fig. 17c). However, the seasonal maximum wind speed of 17.8 m/s is from the south-southeast and recorded at the buoy 46185 on day 151. The spatially averaged seasonal mean wind speed is near 5.5 m/s from the west (Table 6).

Time series show three main northwesterly (upwelling-favorable) wind events. The first one was observed early in the season between day 10 and day 21 (May 10th and May 21st) as a result of a low pressure system over Central Alberta with a trough extending along the Alaska panhandle and a broad high pressure system far offshore. There was also a thermal trough over southeastern B.C. after day 15 (May 15th). The associated northwesterly flow along the northwest coast generated wind speeds up to 11 m/s near the northern end of Vancouver Island. Figure 18i shows the pressure pattern for day 10 (May 10th) with a trough extending over northern B.C. and a northwest-southeast ridge off the B.C. coast, both responsible for the northwesterly winds over much of the B.C. coast. On day 14 (May 14th), there was a thermal trough relatively well developed over southeastern B.C. while the offshore ridge had expanded over northeastern Pacific (Fig. 18j). Then, winds were relatively aligned with the coast (upwelling-favorable), especially in the northern part of Queen Charlotte Sound and around the Queen Charlotte Islands. Wind speeds were up to 15 knots (7.8 m/s) especially north of 51°N.

Thereafter, a series of frontal systems moved over northeastern Pacific and the B.C. coast during the last part of May and much of June (Day 23 (May 23rd) to day 55 (June 24th)). Another ridge of high pressure started to form offshore on day 62 (July...
1st) (Fig. 18k), but winds remained light and variable over much of the study area for the next few days. On day 78 (July 17th), the offshore ridge built stronger and a thermal trough developed over southeastern B.C. (Fig. 18l). Further west, there was a low pressure moving into the Gulf of Alaska, increasing the pressure gradient along the B.C. coast. On day 78 (July 17th) northwesterly winds up to 11 m/s were recorded over much of the area. This northwesterly event ended on day 85 (July 24th) as a deep low pressure system moved into the Gulf of Alaska forcing the ridge onto the B.C. coast (Fig. 18m). Thereafter, a series of frontal systems moved over the B.C. coast giving variable winds over much of the coast. The last northwesterly event was initiated by a ridge that started to develop across the Queen Charlotte Islands during the second week of August (day 104/August 12th), combined with a thermal trough over southern B.C. interior and Washington State (Fig. 18n). However, it is not until day 110 (August 18th) that northwesterly winds were relatively strong over much of the area, when the ridge moved just north of the Queen Charlotte Islands in the wake of a cold front (Fig. 18o). From day 115 to day 139 (August 23rd to September 16th) northwesterly winds were interrupted every 4 to 5 days by frontal waves moving across Queen Charlotte Sound.

d) Summer 1993:

Wind time series for the summer of 1993 are similar to those of 1990 (Fig 17d). Variable winds prevailed from day 1 to day 65 (May 1st to July 4th). From day 66 (July 5th) to almost the end of the summer relatively strong, sustained northwesterly winds were maintained over the area, especially from day 66 to day 81 (July 5th to July 20th) and from day 120 to day 145 (August 28th to September 22nd). The spatially averaged seasonal mean wind speed is 5.1 m/s from the west and the
maximum, recorded at the buoy 46185 on day 124, is 13.2 m/s from the north-northwest (Table 6).

Ridges and lows alternated every 4 to 5 days during the first part of the summer, explaining the variable winds mentioned above. For example, there was a weak low pressure area covering much of B.C. on day 35 (June 4th) giving light and variable winds along much of the coast (Fig. 18p). A few days later (day 38/June 7th), the low had moved along the B.C. and Alberta border, leaving a relatively strong ridge of high pressure near 135°W (Fig. 18q). As a result, the Northwesterlies up to 15 knots (7.8 m/s) were reported over Queen Charlotte Sound. However, on day 40 (June 9th), an intense frontal wave had replaced the ridge and the winds shifted to southeast (Fig. 18r).

During the second part of the summer, the synoptic pattern was dominated by a quasi-stationary and strong high pressure system over northeastern Pacific (Fig. 18s) giving northwesterly winds up to 15 knots (7.7 m/s) along much of the B.C. coast. This ridge was also combined at times with a thermal trough over the southern interior of B.C. and a lee trough along the west coast of Vancouver Island. Figure 18t shows these features, with additionally a trough of low pressure extending along the north coast of B.C. and the Alaska panhandle.

e) Summer 1994:

Wind time series for the summer of 1994 look similar to the one of 1991 (Fig. 17e). A northerly wind component alternated with a southerly one with a period varying generally between 4 to 15 days. Optimal conditions for upwelling were met between day 50 and day 86 (June 19th and July 25th). The spatially averaged wind speed is
near 4.8 m/s from the west-southwest and the maximum wind, recorded at the buoy 46204 on day 44 (June 13th), is 13.9 m/s from the east-southeast (Table 6).

The synoptic pattern cannot be described as maps for 1994 are unavailable.

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Table 6. Statistics for wind data. Wind direction (degree) indicates where the wind originates. It increases from 0° (westerly) to 90° (southerly) and to 180° (easterly). Negative wind direction decreased from 0° (westerly) to -90° (northerly) to -180° (easterly). See the following sketch.
3.2.2.2 variations in space

Winds at the buoys tend to be highly correlated since they are located relatively far from topographic influences. Also, distances between buoys are small compared with the synoptic scale (scale of weather systems).

Visual inspection of wind time series show relatively strong correlations among the four buoys for both, speed and direction. The correlation is particularly strong between buoys 46207 and 46204, although wind speeds tend to be somewhat stronger at the buoy 46207. Winds tend to be a little stronger at buoys 46208 and 46185, especially when the direction is from the northwest or southeast due to the orientation of the land and the fetch.
Figure 15. Sea surface temperatures in degree Celsius (°C) at lighthouse and buoy stations for the summers (day 1 (May 1st) to day 153 (September 30th)) of 1990 to 1994. See Table 4 for missing data.
Figure 15 (continued)
Figure 15 (continued).
Figure 16. Distribution of seasonal mean SST (°C) for summers 1990 to 1994. Isotherms are drawn every 0.5°C. Missing data are indicated by "m". Relatively warm waters are indicated by "w".
Figure 16 (continued).
Figure 17. Wind time series in metre per second (m/s) at the buoy stations a) for the summer 1990 (Day 1 (May 1st) to day 153 (September 30th)). Data for the buoys 46185 and 46208 are missing.
Figure 17 (continued) b) for the summer 1991. Data for the buoys 46207 and 46208 are missing.
Figure 17 (continued) c) for the summer 1992,
Figure 17 (continued) d) for the summer 1993,
Figure 17 (continued) e) for the summer 1994.
Figure 18. Synoptic weather maps for northeastern Pacific/western Canada with sea-level pressures in millibar. a) July 8th, 1990, b) July 14th, 1990,
Figure 18 (continued)  c) August 2nd, 1990, d) August 23rd, 1990,
Figure 18 (continued) e) May 17th, 1991, f) May 27th, 1991,
Figure 18 (continued) g) August 12th, 1991, h) August 24th, 1991,
Figure 18 (continued)  i) May 10th, 1992, j) May 14th, 1992.
Figure 18 (continued) k) July 1st, 1992, l) July 17th, 1992,
Figure 18 (continued) m) July 24th, 1992, n) August 12th, 1992,
Figure 18 (continued) o) August 18th, 1992,
Figure 18 (continued) p) June 4th, 1993, q) June 7th, 1993,
Figure 18 (continued) r) June 9th, 1993,
Figure 18 (continued)  s) July 5th, 1993, t) July 9th, 1993.
Chapter 4

Statistical Analysis

4.1 Regression Analysis

A regression is applied to the time series of temperature to isolate and eliminate the low frequency seasonal signal. The residuals about a least square fit will be called SST anomalies.

Let the temperature at the sea surface be represented as follows:

\[ T = A + B \sin(\omega t) + C \cos(\omega t) + \varepsilon \]  

where \( \omega = 2\pi / 365 \) except \( \omega = 2\pi / 366 \) for the bissextile year (1992),

\[ t = \text{time}, \]

A, B and C are the model’s parameters;

T represents sea surface temperatures (SST) and

\( \varepsilon \) represents SST anomalies about the fit.

In order to find the values for A, B and C, I have used the least square fit method to minimize residuals. Therefore, we need to find the values of A, B and C so that the following expression:

\[ \sum \varepsilon^2 = \sum_{i=1}^{153} (T_i - A - B \sin(\omega t_i) - C \cos(\omega t_i))^2 \]  

is minimal.
4.1.1 Results From Regression Analysis

The coefficients A, B and C, the amplitude, phase and root mean square of residuals (RMS) resulting from the regression analysis are presented in the following tables (Tables 7 to 13) for the three lighthouses and the four buoys over the summer season (from May 1st to September 30th). Figure 19a and 19b show examples of the annual cycle superimposed on the SST for Kains Island and the buoy 46207 between 1990 and 1994. The main difference between the two sites is the amplitudes of the SST variations, the latter being generally larger at Kains Island. The residuals (SST anomalies) are analyzed in more detail in the following section.

From the mean annual SST (coefficient A), 1994 was generally the warmest year with the highest value found at Kains Island. Isolines of constant amplitude (Fig 20) indicate larger variations of the annual cycle at the buoys 46207 and 46204, which also correlates with spatial distributions of mean SST presented in the previous chapter.
Regression Analysis:

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Table 7. Kains Island.

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Table 8. McInnes Island.

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<td>0.67</td>
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Table 9. Egg Island.
Regression Analysis (continued):

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<td>0.44</td>
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Table 10. Buoy 46207.

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<td>0.80</td>
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Table 11. Buoy 46208.

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<td>8.28</td>
<td>9.19</td>
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<td>5.18</td>
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Table 12. Buoy 46204.

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<td>3.98</td>
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<td>C</td>
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<td>-0.19</td>
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<td>0.48</td>
<td>0.62</td>
<td>0.73</td>
<td>0.63</td>
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</table>

Table 13. Buoy 46185.
4.1.2 SST Anomalies Time Series Analysis (Fig 21)

After removing the low frequency signal, visual inspection of time series show the most evident cycle common to all stations at 2 to 3 days (Fig. 21). This is suggestive of a relation with synoptic weather systems as discussed in chapter 2. Then, variations in SST anomalies are not well defined. They seem rather uncorrelated from one station to the other and from one year to the next. However, a weak cycle seems to emerge at 6 to 7 days and up to 10 days which also suggests a connection with the atmospheric circulation. During the summer, slow moving weather systems giving quasi-stationary weather conditions for several days are common along the west coast of B.C which may be responsible for this cycle. A long cycle of 40 to 60 days and up to half the summer period can also be seen in some time series, especially for the summer of 1993 and 1994 (Fig. 21d to 21e).

In more detail, the time series show that high frequency fluctuations tend to have smaller amplitudes at the buoys than at the lighthouses, especially for the years 1990 and 1992. However, SST time series anomalies show episodic large amplitudes at the buoys 46185 and 46208 (see Fig. 21c,d and e). SST anomalies vary approximately between -2.0°C and 2.3°C at the lighthouses and between -1.7°C and 1.5°C at the buoys. Time series suggest similarities between the buoys in high and low frequency variations, especially for the buoys 46204 and 46207. By a careful inspection, one can see similarities between coastal and offshore stations, especially between Kains Island and the buoy 46204. For example, the 1993 data (Fig. 21d) show a correlation at 75 days relatively in phase for all stations with two crests near day 35 and day 110 and two troughs near day 70 and day 145. Low frequency oscillations also emerge from the 1994 time series, but do not appear correlated in space. For the years 1991 to 1992, such slow oscillations do not have much energy. However, correlations can be
found at higher frequencies. For example, the data for 1992 (Fig. 21c) indicate an oscillation at approximately 10 days which show a relatively strong correlation in space over much of the summer, especially between Kains Island and the buoy 46204.

4.2 Spectral Analysis

The spectral analysis technique using the fast Fourier transform (FFT) has been performed on detrended time series (SST anomalies time series) by using a Hanning window over blocks of length of 64 days (M). From Priestley (1989), the bandwidth for a Hanning window is given by

\[ B = 2.45 \frac{\pi}{M \Delta t} \]

With a sampling interval \( \Delta t \) equal to one day, the bandwidth is 0.12 cycle per day (cpd). The number of degrees of freedom is given by \( v = 2.45 N / M \), where \( N \) (153) is the number of observations. Therefore, there are approximately 6 degrees of freedom.

4.2.1 Results From Spectral Analysis

The figures 22a, 22b and 22c present examples of power spectral density (PSD) for Kains Island and the buoys 46207 and 46204 over the study period (1990 to 1994). The 95% confidence intervals indicated by the dashed lines are estimated by calculating the variance of the unaveraged spectral estimates under the assumption of a normal
distribution of nonoverlapping sections. Provided this assumption is correct, there is a 95% probability that the confidence interval covers the true PSD (For more details, see Signal Processing Toolbox, User's Guide for Matlab).

The main feature emerging from PSD is that there is generally more energy towards the low frequencies (red signal) compared with high frequencies. Most PSDs show a relative maximum near 14 days (0.07 cpd). This cycle may be due to the fortnightly tidal variation which affects the method of sampling. However, the weakness of these maxima indicates that other phenomena are also involved within a range of frequencies close to 0.07 cpd. At the other end of the spectrum, a cycle just above 2 days (near 0.44) is common over most stations and in most summers, which agrees with visual inspection (section 4.1.3). From a comparison between Kains Island and the buoys 46204 and 46207, there is also a 4 day cycle (0.25 cpd) showing up for almost every year (Fig. 22a, b and c). However, this energy peak is above the 5% level of significance (0.88) only for the year 1990 and 1994.

4.3 Correlation Analysis

4.3.1 Correlation Coefficient

The correlation coefficient, a measure of intensity of association between two variables, is calculated as follows:

\[
r = \frac{\sum (T_{ui} - \bar{T}_1)(T_{2i} - \bar{T}_2)}{\sqrt{\sum (T_{ui} - \bar{T}_1)^2 \sum (T_{2i} - \bar{T}_2)^2}} \tag{4.3}
\]
where \( T_{1i} \) and \( T_{2i} \) are the temperatures for series 1 and 2 observed at time "i", and
\( \bar{T}_1 \) and \( \bar{T}_2 \) are the seasonal mean temperature for series 1 and 2.

This coefficient does not have units and varies in the range \(-1 \leq r \leq 1\).

4.3.2 Results From Correlation Analysis - SST anomalies

The auto-correlation analysis for Kains Island reveals relatively short de-correlation scales from 1990 to 1992 (4 to 8 days/19 to 38 degrees of freedom(DF)) and longer ones for the years of 1993 and 1994 (near 18 days/8 DF). McInnes Island seems to have a longer memory with a de-correlation scale between 12 and 18 days (13 and 9 DF), except for the year 1992 where it is only 5 days (30 DF). Finally, the mean de-correlation scale is 7 days (21 DF) at Egg Island and varies generally between 6 and 18 days (8 and 25 DF) at the buoy stations.

The correlation coefficients among buoys and lighthouses vary from nearly 0 to 0.76 with highest values generally reached in 1993 (tables 14 to 18). The highest correlation between time series seems to be revealed by the buoys 46204 and 46207 (0.76) of that year, while nearly no correlation is shown by McInnes Island and the buoy 46204 (0.01) in 1990. However, from the 95% confidence intervals, only eight coefficients are significantly different from zero (marked with star sign in Table 14 to 18). Most of them are revealed by the buoy data in 1992 and 1993. Among the lighthouses, the highest significant correlation coefficient (0.47) is shown by Kains and McInnes Islands in 1992. A comparison between Kains Island and the buoy
coefficients shows the highest significant correlation with the buoy 46204 in 1993 (0.61).

Correlation coefficients (r) for SST time series:

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<tbody>
<tr>
<td>Kains/McInnes</td>
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<td>0.48</td>
<td>0.47 *</td>
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Table 14. Lighthouses

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<td>0.23</td>
<td>---</td>
<td>0.66</td>
</tr>
<tr>
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<td>---</td>
<td>---</td>
<td>0.52 *</td>
<td>---</td>
<td>0.46</td>
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<td>---</td>
<td>0.22</td>
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<td>0.35</td>
</tr>
<tr>
<td>46207/46185</td>
<td>---</td>
<td>---</td>
<td>0.59 *</td>
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<td>0.28</td>
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<td>---</td>
<td>0.55 *</td>
<td>0.76 *</td>
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</tr>
<tr>
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<td>0.59</td>
<td>0.47</td>
<td>0.67 *</td>
<td>0.56</td>
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Table 15. Buoys.

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<td>0.26</td>
<td>0.19</td>
<td>0.47 *</td>
<td>0.36</td>
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Table 16. Kains Island vs buoys.

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Table 17. McInnes Island vs buoys.

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</table>

Table 18. Egg Island vs buoys.
4.3.3 Results From Correlation Analysis - Alongshore Wind Component

The alongshore wind is obtained by rotating the u and v-components by 45°. Then, the correlation coefficients have been computed for the alongshore wind component. Numbers are relatively high (Table 19), as expected from visual inspection of wind time series, and they are all significantly different from zero. The lowest values (near 0.60) are found between the buoys 46208 and 46204, while the highest ones (near 0.90) involve the buoy 46207, 46204 and 46185. The decorrelation scales for wind time series (between 3 to 8 days) is generally shorter than the ones for SST time series. This is not surprising due to the great inertia of the ocean compared to that of the atmosphere.

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<td>0.60</td>
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<td>0.80</td>
<td>0.75</td>
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<tr>
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<td>0.82</td>
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<td>0.87</td>
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</table>

Table 19. Correlation coefficients for alongshore wind time series at the buoys.

4.4 General Observations

From most results in this chapter, it seems evident that statistical analysis does not give enough information about the temperature variations of the surface water. The main reason is the large number of variations caused by complex oceanic and/or atmospheric processes at different scales, both in time and space. However,
by visual inspection of the time series, there is evidence of some correlations for certain periods of time. In order to understand the variations of the surface water temperature, three events are analyzed in chapter 6. Details are given in terms of SST compared with wind and cloud data, along with synoptic pressure patterns. However, before we get into the analysis, a review of the relevant theory is presented in the next chapter.
Figure 19 (continued). b) for the buoy 46207.
Figure 20. Distribution of the amplitude of the annual cycle of SST (°C) for summers 1990 to 1994 (from regression analysis). Isotherms are drawn every 0.5°C. Missing data are indicated by "m". Relatively warm waters are indicated by "w".
Figure 20. (continued).
Figure 21. Sea surface temperature anomalies(°C) at lighthouse and buoy stations for a) summer 1990 (Day 1 (May 1st) to day 153 (September 30th)). The data for the buoys 46185 and 46208 are missing.
Figure 21. (continued) b) summer 1991. The data for the buoy 46207 are missing.
Figure 21. (continued) b) summer 1991.
Figure 21. (continued) c) summer 1992.
Figure 21. (continued) c) summer 1992.
Figure 21. (continued) d) summer 1993. The data for the buoy 46208 are missing.
Figure 21. (continued) d) summer 1993.
Figure 21. (continued) e) summer 1994.
Figure 21. (continued) e) summer 1994.
Figure 22. Power spectrum density for SST anomalies at a) Kains Island. The area within the dashed lines represents the 95% confidence intervals. The bandwidth is 0.12 cpd and the degrees of freedom is 6.
Figure 22. (continued). b) Buoy 46207,
Figure 22. (continued). c) Buoy 46204.
Chapter 5

Background Theory

The first part of this chapter (section 5.1) presents an overview of the main factors affecting the SST along the west coast area of British Columbia based on past studies. In particular, it outlines the mechanisms affecting the local stratification which relates to the variability of SST. The next section (5.2) is a review of the heat budget of the surface waters, covering the most important terms responsible for SST variations. Then, section 5.3 presents a summary of Ekman theory, relating the wind forcing to the ocean's response in the upper layer. In particular, it includes a detailed description of the upwelling phenomena which appear as a dominant factor in modulating SST over much of the study area as shown in chapter 6.

5.1 Vertical Stratification

The mechanisms responsible for the stratification of the shelf waters (within the 200m depth contour) fall under the following two categories: The first category involves mechanisms that create stability, i.e., the buoyancy forces resulting from freshwater input and surface heating. The second category includes processes that destroy it, i.e., the negative buoyancy forces resulting from evaporation and cooling and the mixing forces such as wind and bottom shear generated turbulence. Most of these mechanisms can lead to SST fluctuations with a time scale from a few to several days (2 to 10 days).

River runoff is likely to be the primary source of buoyancy that creates stratification especially in the inner and middle shelf near rivers (Atkinson and Jackson, 1986). In
comparison with solar heating that may be equally important over the entire shelf area, Garrett et al. (1978) have shown that the buoyancy flux at rate $R$ is equivalent to that due to a heat flux $Q$ if

$$R = \alpha Q c_p (\Delta \rho)$$

where $R$ is the fresh water input from rainfall, $\alpha$ is the thermal expansion coefficient of water, $c_p$ the specific heat and $\Delta \rho$ the density difference between fresh and salt water.

This relationship, in agreement with Atkinson and Jackson's results (1986), shows that the stratification may result from both fresh water input and heat fluxes in areas where rivers discharge is significant, assuming that $R$ represents freshwater input from river discharge instead of rainfall.

From historical salinity data (Dodimead, 1980), the presence of freshwater is revealed along the mainland coast, especially over the southern part of Queen Charlotte Sound during the summer. Figure 4 shows a tongue of low salinity along a northeast-southwest line just south of Calvert Island that was particularly pronounced from June 29th to July 22nd, 1954. Therefore, the freshwater input will increase the stability of the water column in the vicinity of Cook bank.

Within a time scale of a few days, the surface wind was found to be a dominant factor in cooling surface waters, especially during intense Pacific storms (Large and Crawford, 1994). Their experiment, the "Ocean Storms Experiment", also revealed that the resulting wind stress and entrainment process may cause significant cooling of the SST within a single day by mixing the upper layer and entrainment across the thermocline, depending on the wind speed, the local stratification and the Coriolis parameter. During the summer, wind speeds are generally weak along the west
coast of B.C. due to a ridge of high pressure prevailing over northeastern Pacific, especially in July and August. Therefore, the entrainment process and the cooling of the surface water caused by the vertical mixing are expected to be weaker during that season. However, northwesterly winds generated ahead of the ridge may cause significant cooling of the coastal surface water during the summer by induced upwelling (section 5.3.1). This can be seen from seasonal mean SST (Table 5) showing lower values toward the shore.

Turbulent mixing caused by bottom friction and tidal currents is another mechanism that weakens the stratification by mixing the water column vertically. Simpson and Hunter (1974) proposed a parameter proportional to $H/U^3$ for the separation of mixed and stratified regions, where $H$ is the depth of the water and $U$, the mean speed of tidal current. They assumed the heat flux, density, specific heat, thermal expansion and bottom friction coefficients uniform over an area. Based on their mixing parameter $S = \log_{10}(H/C_d U^3)$ where $C_d$ is the bottom drag coefficient, Jardine et al (1993) produced maps of stratification distribution over the Queen Charlotte Sound and Hecate Strait for spring and neap tides (Fig. 23). Maps show stratified conditions over much of the study area except in the vicinity of Goose Island bank and at the northern end of Vancouver Island.
5.2 Heat Budget (Pickard and Emery, 1982)

The amplitude and character of the SST variations depend on energy exchange processes occurring at the air-sea interface and advective processes occurring below the sea surface. While the energy exchange processes include the solar radiation, effective back radiation, evaporation and conduction of sensible heat, the advective processes are concerned with transport of water resulting from wind-, tidal- and/or density forcing. The net rate of heat flow at the air-sea interface can be stated by the following equation

\[ Q_s + Q_b + Q_h + Q_e + Q_v = Q_T \]

where \( Q_x \) represent the different components of the heat budget as follows:

- \( Q_s \), the rate of inflow of solar energy through the sea surface,
- \( Q_b \), the net rate of heat loss by the sea as long-wave radiation to the atmosphere and space,
- \( Q_h \), the rate of heat loss/gain through the sea surface by conduction,
- \( Q_e \), the rate of heat loss/gain by evaporation/condensation,
- \( Q_v \), the rate of heat loss/gain by the ocean due to advection and
- \( Q_T \) is the resultant rate of gain/loss of heat of the ocean.

During the summer season, the terms, in order of importance, are \( Q_s \) which is always positive, \( Q_b \) and \( Q_e \) which are always negative and \( Q_h \) which is slightly positive. Advection \( Q_v \) may be quite variable over space and time. When \( Q_T \) is positive (negative), there is gain (loss) of heat into (from) the ocean and an increase (decrease) of SST. The different terms of the above equation are discussed individually below.
5.2.1 Solar radiation ($Q_s$)

The main source of energy for a column of water near the surface is the short-wave radiation emitted by the sun. When entering the atmosphere, parts of the incoming short-wave radiation are lost by scattering and absorption due to interaction with the atmosphere's particles and water vapor. However, since the atmosphere re-emits some energy downward (called sky radiation), the loss is reduced. From the energy reaching the sea, a small fraction is reflected at the surface and the rest is used to increase the SST. This energy flux is always positive, however it varies over space and time due to variations in the composition of the atmosphere and the degree of cloudiness. As a reference, Figure 24 (from Pickard and Emery, 1982) shows the daily inflow of solar radiation at the Earth's surface as a function of latitude and time of year, where no cloud and an average atmospheric transmission of 70% have been assumed. Near the latitude of the study area (51°N), the daily rate of short-wave radiation ($Q_s$) varies from 200 W/m² in May to less than 100 W/m² in September. Maximum values estimated at 250 W/m² are found toward the summer solstice. A study by Tabata in the vicinity of Triple Island, B.C. (1958) indicates a summer mean value for $Q_s$ near 230 W/m².

The cloud cover is a major factor for the amount of solar radiation reaching the sea. Tabata (1958) showed that the incident solar radiation at the sea surface under mean cloud cover was approximately half of the incident solar radiation in the absence of clouds. Thus, year-to-year fluctuations of monthly means were mostly associated with variations in cloud cover.

The presence of clouds in the atmosphere reduces parts of the incoming short-wave radiation by scattering and absorption due to interaction with cloud particles. To
take this effect into account, the mean energy which would arrive in the absence of clouds may be multiplied by the following conversion factor: (1-0.09C) where C is the proportion of sky covered by cloud in eighths (oktas) (Pickard and Emery, 1982). It is noteworthy that this method neglects the sunlight scattered by the atmosphere and clouds. However, during the summer a large fraction of $Q_s$ is direct sunlight. A second factor neglected is the reflection at the sea surface which depends upon the elevation of the sun and the sea state. While the reflection on a flat sea can be estimated from a coefficient table, the wave effect is more difficult to estimate.

5.2.2 Long-Wave Radiation ($Q_b$)

The long-wave radiation term $Q_b$, also called back radiation, reflects the net amount of energy lost by the sea as long-wave radiation. The net transfer of long-wave radiation is a function of the sea and air temperatures, the vapour content of the air and the cloud amount and height. The magnitude of $Q_b$ is proportional to the fourth power of the absolute temperature of the emitter; it is always negative, meaning that it contributes to decreasing the SST.

This term can be measured with a radiometer as described in Pickard and Emery (1982). However, when direct measurements are not available, the heat loss by the sea may be estimated from a method taking into account the absolute SST and the overlying water-vapor content. While the SST gives an estimate of the rate of outward flow of energy, the water-vapor content provides a value for the inward flow from the atmosphere since the atmospheric water vapor is the main source of its long-wave radiation. Values for $Q_b$ may be read from Figure 25 when the SST and the relative humidity are known. For example, a relative humidity near 90%
with the SST at 12°C gives an approximate value for $Q_b$ of -92 w/m$^2$, assuming no cloud. In the presence of clouds, the contribution from the atmosphere is increased so that the net loss of heat is reduced. This effect may be taken into account by using the factor $(1-0.1C)$ where $C$ is the amount of sky covered by clouds in oktas (Pickard and Emery, 1982).

The back radiation term is assumed to be relatively constant in time and space because it depends on the absolute temperature, rather than the Celsius temperature, and because the relative humidity does not change much over the sea. The results from Tabata (1958) show a net loss of heat from the effective back radiation near -82 g-cal/cm$^2$/day (-40 W/m$^2$) relatively constant from May to September. Therefore, under clear skies the effect of $Q_b$ on SST variations may be neglected. However, when clouds are present, the loss of energy as long-wave radiation can be sharply reduced. This may result in a noticeable fluctuation in heat flux from one day to the next. Therefore, variations in $Q_b$ could be important for those in the SST.

5.2.3 Conduction ($Q_h$)

Heat may be gained or lost from the sea surface by conduction due to the temperature gradient in the air adjacent to the sea. When the air temperature decreases upward, heat is conducted away from the sea surface. The rate of loss of heat $Q_h$ is proportional to the heat conductivity coefficient $K$, the specific heat of air at constant pressure $C_p$ and the vertical air temperature gradient $dT/dz$ such as:

$$Q_h = -C_p \cdot K \cdot dT/ dz.$$  
When the air is stationary, the heat is transferred by molecular processes as a function of $K$, a molecular conductivity coefficient. However, since
the air is usually in motion and turbulent, the heat is conducted much more
efficiently by air eddies and $K$ is replaced by an eddy conductivity coefficient $A_h$.
Then, uncertainties in $Q_h$ arise because of the lack of information on $A_h$, the latter
being a property of air motions rather than a property of the air itself.

The rate of heat transferred through the sea surface by conduction is generally
positive during the summer time (May to September) since the air temperature is
usually greater than the water temperature. However, values are generally quite
small and relatively unimportant compared with the other terms discussed above.
From annual mean values (Pickard and Emery, 1982) $Q_h$ is near 0 W/m$^2$ over much
of the B.C. coast, which also agrees with results from Tabata (1958). The conduction
term can be more important in coastal areas, especially after a significant upwelling
event has occurred. However, variations in $Q_h$ still remain small compared to those
in $Q_i$ and $Q_s$ and may be neglected in most situations.

5.2.4 Evaporation/Condensation ($Q_e$)

This term is always negative over the ocean because heat is taken away from the
water when evaporation occurs. The rate of heat loss is expressed by $Q_e = F_e \cdot L_l$
where $F_e$ is the rate of evaporation of water and $L_l$ is the latent heat of evaporation.
While $L_l$ can be found in a table, $F_e$ may be approximated by three different
methods. The first of these is called the Pan Method and involves direct
measurements of $F_e$. The latter are usually not very accurate because of the
difficulties in reproducing the real atmospheric conditions and make the method
not very useful.
A second method, called the flow method, is based on the following eddy diffusion-flow formula \( F_e = -A_e \frac{df}{dz} \) where \( A_e \) is the eddy diffusion coefficient for water vapour through the atmosphere and \( \frac{df}{dz} \) is the gradient of water vapour concentration (humidity) in the air overlying the sea surface. Because of difficulties similar to those mentioned in section 5.2.3, this formula is often replaced by a semi-empirical formula such as: \( F_e = 1.4(e_s - e_a) \cdot W \) where \( e_s \) is the saturated vapour over the sea-water and \( e_a \) is the actual vapour pressure in the air measured at 10 m above sea level so that \( (e_s - e_a) \) represents \( \frac{df}{dz} \). \( W \) is the wind speed measured at 10 m height and gives an approximation of the variations of \( A_e \). In the summer the SST is usually lower than the temperature of the overlying air, so that the latter is more stable. As a result, the evaporation rate is reduced compared with winter values. From Tabata (1958), the rate of heat lost by evaporation in the vicinity of Triple Island is near -45 g-cal/cm\(^2\)/day (-21 W/m\(^2\)) between May and September and the annual mean value is near -89 g-cal/cm\(^2\)/day (-43 W/m\(^2\)).

Finally, a third method for estimating \( Q_e \) makes use of the heat budget equation. This method would not be valid for the present work because it requires the assumptions of no advection \( (Q_v = 0) \) and a steady state \( (Q_T = 0) \) which are not true over the spatial and temporal scales involved in this thesis. A description of this method can be found in Pickard and Emery (1982).

In general, variations in the rate of heat loss due to evaporation may be important for the daily SST fluctuations, although the values remain smaller than those associated with the incoming short-wave radiation. From the semi-empirical formula, variations in \( F_e \), and therefore in \( Q_e \), are caused mainly by the variations in the wind speed, which assist the evaporation process. The values for \( (e_s - e_a) \) are
relatively small, especially during the summer because of generally small differences between the air temperature at 10 m height and the SST.

5.2.5 Advection ($Q_v$)

Besides the energy transfer processes described above, the movement of water, i.e. horizontal advection, upwelling and/or mixing, may play an important role in affecting the SST on relatively short spatial and temporal scales. These advection terms are positive when there is inflow of warm water and negative when there is inflow of cold water, mixing and/or upwelling. In particular, significant amount of heat may be advected by entrainment and wind-induced upwelling, especially in coastal areas. In most cases, values are not available due to the lack of measurements. However, an appreciation of the advective term can be done by reference with the other heat budget terms.

5.2.6 Total Heat Gain/Loss ($Q_T$)

The total transfer of heat at the air-sea interface is estimated by the sum of the four heat flux terms across the surface of the ocean ($Q_s$, $Q_b$, $Q_h$ and $Q_e$) plus the advection term ($Q_v$). From a study by Tabata (1958), there is a net gain of heat into the sea near 200 g-cal/cm$^2$/day (97 W/m$^2$) from May to September and the year-to-year deviations are mainly due to modification of solar radiation by cloud cover for much of the summer season.
5.3 Ekman Layer Theory

When the wind blows over the ocean, a frictional force is exerted on the surface. In the atmospheric Ekman layer, the lower boundary condition for the momentum equations requires continuity of velocity and stress at the surface of the ocean, which creates a vertical wind shear. The latter is responsible for a vertical exchange of momentum acting as a driving force for water movements and ocean currents. If $u$ and $w$ are the horizontal and vertical velocities, $\rho$ the density, then the vertical flux of horizontal momentum is given by $\rho u w$ (per unit area). The mean value of this flux over a large area during a long period of time is equal to the mean stress $\tau$ which is related to the wind speed $u$ by the following relationship:

$$\tau = C_D \rho u |u|$$

(5.1)

For light wind speeds, usually taken at 10m above the ocean's surface, there is a linear relationship between $u$ and $C_D$. However, at high winds, a correction is needed to account for high sea states. From Gill (1982), the drag coefficient is given as follows:

$$C_D = \begin{cases} 
1.1 \times 10^{-3} & \text{for } u < 6 \text{ m/s} \\
(0.61 + 0.063u) \times 10^{-3} & \text{for } 6 \text{ m/s} < u < 22 \text{ m/s} 
\end{cases}$$

(5.2)

Figure 26 shows the summer alongshore component of the wind stress at the four buoy stations between 1990 and 1994. Negative wind stress means that the wind is blowing from the northwest and is favorable for upwelling along the west coast of north America according to the theory and previous studies (Allen, 1980, Ikeda et al, 1984 and others). From May to September, more than 70% of the wind stresses are
less than 0.16 N/m² over Queen Charlotte Sound and off northern Vancouver Island. Along coastal areas, this percentage may increase to 90% (Manual Weather Hazards Manual, 1990).

The following table (Table 20) gives simple statistics for the alongshore wind stress components at the buoys for the study period. As mentioned in Chapter 4, the alongshore wind is obtained by rotating the u and v-components by 45°. The summer mean alongshore wind stresses are generally from the northwest, although mean values are relatively small. From minimum and maximum values, southeasterly wind stresses reach generally higher strength than the northwesterly ones. Southeasterly wind stresses are associated with frontal systems and stronger pressure gradients, while northwesterly wind stresses accompany ridges with generally weaker pressure gradients.
### Table 20

<table>
<thead>
<tr>
<th>Year</th>
<th>Buoy sites</th>
<th>Max (N/m$^2$)</th>
<th>Min (N/m$^2$)</th>
<th>Mean (N/m$^2$)</th>
<th>Std (N/m$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990</td>
<td>46204</td>
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<td>-0.13</td>
<td>-0.01</td>
<td>0.04</td>
</tr>
<tr>
<td></td>
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<td>0.17</td>
<td>-0.14</td>
<td>-0.02</td>
<td>0.05</td>
</tr>
<tr>
<td>1991</td>
<td>46204</td>
<td>0.24</td>
<td>-0.14</td>
<td>0.00</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td>46185</td>
<td>0.28</td>
<td>-0.19</td>
<td>0.00</td>
<td>0.06</td>
</tr>
<tr>
<td>1992</td>
<td>46204</td>
<td>0.39</td>
<td>-0.14</td>
<td>-0.01</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
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<td>-0.15</td>
<td>-0.02</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td>46208</td>
<td>0.24</td>
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<td>-0.02</td>
<td>0.08</td>
</tr>
<tr>
<td></td>
<td>40185</td>
<td>0.64</td>
<td>-0.25</td>
<td>-0.02</td>
<td>0.09</td>
</tr>
<tr>
<td>1993</td>
<td>46204</td>
<td>0.21</td>
<td>-0.15</td>
<td>-0.01</td>
<td>0.05</td>
</tr>
<tr>
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<td>-0.16</td>
<td>-0.02</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
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<td>-0.04</td>
<td>0.07</td>
</tr>
<tr>
<td></td>
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<td>0.29</td>
<td>-0.28</td>
<td>-0.03</td>
<td>0.08</td>
</tr>
<tr>
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<td>46204</td>
<td>0.28</td>
<td>-0.11</td>
<td>0.01</td>
<td>0.05</td>
</tr>
<tr>
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<td>0.05</td>
</tr>
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<td>-0.19</td>
<td>0.00</td>
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</tr>
<tr>
<td></td>
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<td>0.26</td>
<td>-0.15</td>
<td>0.00</td>
<td>0.06</td>
</tr>
</tbody>
</table>

Table 20. Maximum, minimum, mean and standard deviation for wind stress data during the summer time (Day 1 (May 1st) to day 153 (September 30th)). 1990 data are missing for the buoys 46185 and 46208, and 1991 data are missing for the buoys 46207 and 46208. Values are in N/m$^2$. 

![Diagram of wind stress](image-url)
There are two sources for the frictional force: one is the molecular friction which may be neglected in most aspects of the dynamics of ocean motions. The second force is related to the degree of instability of the motion: when the motion becomes turbulent, the non-linear terms in the Navier-Stokes momentum equations give rise to terms that have the physical character of friction. In most cases, friction forces achieved through turbulence are much more effective in re-distributing momentum and water properties.

The horizontal stresses $\tau_x$ and $\tau_y$ at the ocean's surface form a vector representing the force per unit area applied on the sea surface. If the ocean is divided into a series of thin layers, this force tends to set a motion to the top layer and thus exert a stress on the layer underneath. If an infinitesimal layer of thickness $\delta z$ is considered, the stress on the layer below can be approximated by:

$$
\left( \tau_x - (\delta z \partial \tau_x / \partial z), \tau_y - (\delta z \partial \tau_y / \partial z) \right)
$$

An equal and opposite stress is exerted on the base of the original layer, and the net force per unit area on that layer is the difference between the stress on the top and bottom as follows:

$$
(\partial \tau_x / \partial z, \partial \tau_y / \partial z) \delta z
$$

The resulting force per unit mass that tends to accelerate the fluid is:

$$
1/\rho(\partial \tau_x / \partial z, \partial \tau_y / \partial z)
$$

and the vertical shear is measured by $\partial \tau / \partial z$ where $\tau = (\tau_x, \tau_y)$. 

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Ultimately, there is a depth where the momentum transmitted by the wind vanishes and so do the frictional forces. Typically, the depth of the oceanic top boundary layer is 10m to 15m for calm summer days. During a study on wind-driven currents over the Oregon continental shelf, Allen (1980) found a surface Ekman layer of about 20m depth under variable alongshore winds up to 10m/s. As a result of the Coriolis force, the wind stress, applied at the ocean surface, forces a surface current to its right (Northern Hemisphere). The latter rotates gradually to the right with increasing depth down to the bottom of the surface Ekman layer. The net depth-integrated transport in the surface Ekman layer is directed at 90° to the right of the wind stress direction. The Ekman transport leads to convergence/divergence of mass and hence, by continuity, to vertical motion of water into or out of the boundary layer called "Ekman pumping." In areas of convergence, a downward vertical motion pushes the isotherms down. In reverse cases, an upward vertical motion responds to divergence of surface waters which rises the isotherms, therefore, cools the upper mixed layer.

The magnitude of vertical velocity \( w_E \) just outside the surface Ekman layer can be found by integrating the continuity equation:

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (5.3)
\]
Integration is made with respect to $z$ across the surface Ekman layer. With the boundary condition of no vertical motion at the ocean's surface, i.e., $w=0$, the integral takes the following form:

\[
\int_{-h}^{0} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) dz = 0 \quad (5.4a)
\]

\[
\frac{\partial}{\partial x} \int_{-h}^{0} udz + \frac{\partial}{\partial y} \int_{-h}^{0} vdz + w(0) - w(-h) = 0 \quad (5.4b)
\]

where $z=-h$ is the bottom of the Ekman layer. $\int_{-h}^{0} udz$ and $\int_{-h}^{0} vdz$ are the Ekman layer volume fluxes which, by integration are equal to \( \frac{\tau_x}{\rho f} \) and \( \frac{\tau_y}{\rho f} \) respectively, where $\rho$ is the water density and $f$ is the Coriolis parameter. It is a valid approximation to integrate the Ekman transport from \((-\infty \text{ to } 0)\) instead of \((-h \text{ to } 0)\) because this transport is assumed to be zero below the level $-h$. Assuming that the variation of $f$ is small compared to the variation of the wind stress and solving 5.4 for $w_e$, the Ekman pumping velocity, gives:

\[
w_e = \frac{1}{\rho f} \left( \nabla \times \bar{\tau} \right) \quad (5.5)
\]

5.3.1 Upwelling Theory

Along the west coast of B.C., under northwesterly winds, the stress transmitted to the ocean's surface produces a net Ekman transport away from the coast, leading to a compensating onshore flow below the surface layer and an upward vertical motion.
near the coast called upwelling. Such a process is observed by the appearance of relatively cold, dense water at the surface near the coastline and the transport of dissolved nutrients from greater depths to the euphotic zone.

The offshore Ekman transport results in an area of low pressure along the coast relative to the pressure offshore. Then, the horizontal pressure gradient drives a current from the high towards the low pressure area (towards the shore) in the region just below the surface Ekman layer. Within this region, called the geostrophic interior, there is a balance between the pressure gradient and the Coriolis force which deflects the motion to the right (northern hemisphere). The resulting southerly flow below the surface Ekman layer is assumed to be geostrophic. The vertical structure of the full water column depth requires another Ekman layer below the geostrophic interior, with a thickness comparable to the surface Ekman layer. In the bottom Ekman layer, the ocean's floor exerts a drag on the current producing a mass flux at $135^\circ$ to the left of the geostrophic current, so that the latter is deflected towards the coast. Then, by continuity, there is an upward vertical motion induced at the coast.

Upwelling is probably the main cause of short-term (two to ten days) SST fluctuations along the west coast of B.C. during the months of July and August. Typical values for the upwelling velocity are found between 1 to 10 m/day (Allen, 1980). The main part of the response is confined to a distance from the coast of the internal Rossby radius (typically 20 km along the coast of B.C.). Off the Oregon coast, the vertical motion has been observed over the continental shelf within a distance of 10-15 km from the shore (Allen, 1980). Over much of Queen Charlotte Sound, upwelling is reduced and cold waters do not always reach the surface. A four month simulation run by a simple two-layer upwelling model over Hecate Strait and
Queen Charlotte Sound has shown the 7°C isotherm rising from 200m to about 130m with the surface temperature simultaneously increasing from 7.8°C to 11.0°C due to surface heating (Ma, 1992). In such a situation, the SST, as measured by buoys may not be appropriate to detect temperature anomalies in fish marine environment below the surface.

From (5.5),

\[ w_k \equiv \frac{\hat{\rho}}{\rho f} (\hat{\nabla} \times \hat{\tau}) \]

an appreciation of upwelling responses can be obtained from the vertical displacement \( \Delta z \) as follows: Assuming upwelling favorable-winds in a two-dimensional Cartesian coordinate system \((x, z)\) with the \(x\)-axis aligned across shore and positive onshore, and the \(z\)-axis vertical, the equation 5.4b leads to:

\[ \frac{\tau}{\rho f} = -wa \]  \hspace{1cm} (5.6)

where \( w \) is the upwelling velocity and \( a \) is the Rossby radius. The alongshore variations of the current have been neglected since it is the cross-shelf mass balance that is of particular concern in upwelling situations.

The Rossby radius \( a \) is defined as follows:

\[ a = \sqrt{\frac{g \Delta \rho}{\rho f} \frac{H}{f}} \]  \hspace{1cm} (5.7)
An approximation of the Rossby radius over the study area can be obtained from survey data of August 23rd, 1950. The following values taken near Kains Island: \( \Delta \rho / \rho = 2.24 \times 10^{-3} \), \( f = 1 \times 10^{-4} \, \text{s}^{-1} \), the thickness of the layer \( H = 78\,\text{m} \) and \( g = 9.8 \, \text{m} / \text{s}^2 \), correspond to a Rossby radius near 13 km. Just south of the buoy 46204, survey data give a Rossby radius near 20 km.

From 5.6, \( w \) is given by:

\[ w = \frac{|\dot{\theta}|}{\rho f a} \quad (5.8) \]

By integrating the relation 5.8 on both sides with respect to time, we get the upward isotherm displacement due to upwelling:

\[ \Delta z = \rho f a \int |\tau| \, dt \quad (5.9) \]

5.3.2 Two-Layer Model

The model described so far represents an ocean where the distributions of water properties such as density, temperature and salinity are continuous through the whole depth of the ocean. Although this model is realistic, it is not computationally easy to solve. However, a solution can be found if the ocean is divided into layers. In this model each layer has a constant density profile, different from one layer to the next so that the variation of the vertical structure of the water column is represented. A two-layer model with constant densities \( \rho_1 \) and \( \rho_2 \) representing the values in the mixed layer and the deep ocean is a good approximation for coastal regions where there is significant river runoff giving rise to an upper layer of low
salinity over a deep layer of much higher salinity, with a sharp halocline between them (Pond and Pickard, 1983). Although this model contains only two vertical modes of oscillation; a barotropic mode and the first baroclinic mode, it still remains a good approximation because in many cases involving long-period hydrostatic motions, most of the energy is usually contained in these first two modes (LeBlond and Mysak, 1978). The barotropic mode is one where the isobars and isopycnals are parallel so that the flow is uniform across the layer, while the first baroclinic mode presents tilted isobars compared with isopycnals so that the fluid in each layer flows in opposite direction. The latter case results in no horizontal net flux.

In mathematical form, a two-layer model is expressed as follows:

By reference to Figure 27, the upper layer has a density is \( \rho_1 \), horizontal velocity components \( u_1 \) and \( v_1 \), an equilibrium depth \( H_1 \), and a surface elevation \( z = \eta(x, y, t) \). Similarly, the second layer is characterized by \( \rho_2 \), \( u_2 \), \( v_2 \), \( H_2 \), and \( z = -H_1 + h(x, y, t) \) where the total depth \( H \) is equal to \( H_1 + H_2 \).

**Approximations for the two-layer shallow water model:**

1. For phenomena of relatively small scale (near 100 km), the Coriolis parameter, \( f = 2\Omega \sin \Phi \), may be assumed constant. The frame of reference used is called: f-plane.

2. When the horizontal velocities are large compared to the vertical velocities and the horizontal scales of motions are greater than the depth of the fluid, then the pressure perturbation is independent of depth and there is a balance between the
vertical pressure gradient and the buoyancy force. This is called the hydrostatic approximation.

3. If the density variations are small, then their effect on the mass of the fluid can be neglected, but their effect on the weight must be retained. This is called the Boussinesq approximation.

With these approximations, the linear shallow water equations with wind forcing are:

\[
\begin{align*}
\frac{\partial u_1}{\partial t} - f v_1 &= -g \frac{\partial \eta}{\partial x} + \frac{\tau_x}{\rho_i H_i} \quad (5.10a) \\
\frac{\partial v_1}{\partial t} + f u_1 &= -g \frac{\partial \eta}{\partial y} + \frac{\tau_y}{\rho_i H_i} \quad (5.10b)
\end{align*}
\]

for the upper layer, and

\[
\begin{align*}
\frac{\partial u_2}{\partial t} - f v_2 &= -g \frac{\partial \eta}{\partial x} - g' \frac{\partial h}{\partial x} \quad (5.11a) \\
\frac{\partial v_2}{\partial t} + f u_2 &= -g \frac{\partial \eta}{\partial y} - g' \frac{\partial h}{\partial y} \quad (5.11b)
\end{align*}
\]

for the lower layer. \(h\) is the upward displacement of the interface and \(g'\) is the reduced gravity for a two layer fluid as follows:

\[
g' = g (\rho_2 - \rho_1) / \rho_2
\]
The continuity equations for the upper and lower layers have the form:

\[ \frac{\partial(\eta + H_1 - h)}{\partial t} + H_1 \left( \frac{\partial u_1}{\partial x} + \frac{\partial v_1}{\partial y} \right) = 0 \]  
\[ (5.12a) \]

\[ \frac{\partial h}{\partial t} + H_2 \left( \frac{\partial u_2}{\partial x} + \frac{\partial v_2}{\partial y} \right) = 0 \]  
\[ (5.12b) \]

The terms in brackets express the divergence of the horizontal transport. From 5.12a, the divergence of the fluid is causing a downward displacement of the free surface, while from 5.12b, it is resulting in an upward displacement of the interface between the two layers.

For the baroclinic mode, the free surface displacement is smaller than the interface displacement so that the rigid lid approximation can be made. Then the equation of continuity 5.12a becomes:

\[ \frac{\partial(-h)}{\partial t} + H_1 \left( \frac{\partial u_1}{\partial x} + \frac{\partial v_1}{\partial y} \right) = 0 \]  
\[ (5.13) \]

The difference in velocity between the two layers (also referred at the amplitude of the baroclinic mode) is obtained by subtracting 5.11 from 5.10 to give:

\[ \frac{\partial \hat{u}}{\partial t} - f\hat{v} = g \frac{\partial h}{\partial x} + \frac{\tau_x}{\rho_1 H_1} \]  
\[ (5.14a) \]

\[ \frac{\partial \hat{v}}{\partial t} + \hat{f}\hat{u} = g \frac{\partial h}{\partial y} + \frac{\tau_y}{\rho_1 H_1} \]  
\[ (5.14b) \]

where \( \hat{u} = u_1 - u_2 \) and \( \hat{v} = v_1 - v_2 \) which can be thought of as the amplitude of the baroclinic mode of a two-layer system.
With the definition for \( \hat{u} \) and \( \hat{v} \), the equation of continuity is given by:

\[
-\left( \frac{1}{H_1} \frac{\partial h}{\partial t} + \frac{\partial \hat{u}}{\partial x} + \frac{\partial \hat{v}}{\partial y} \right) = 0
\]  \hspace{1cm} (5.15)

The equations 5.14 and 5.15 describe the baroclinic response.

### 5.3.2.1 Coastal Upwelling Using The Two-Layer Model

Consider a particular case in which the surface Ekman transport is directed offshore with a vertical wall at the coastline (i.e. at \( y=0 \)). The vertical depth \( H \) and the wind stress \( \tau_x \) are constant, and \( \tau_y = 0 \). In upwelling situations, variations of the surface current in the alongshore direction can be neglected and \( \partial / \partial x \to 0 \). Then 5.14 and 5.15 are combined into a forced shallow-water equation with the following solution:

\[
\hat{v} = -\left( \frac{\tau_x}{\rho f H_1} \right) \left( 1 - \exp \left( \frac{-y}{a} \right) \right)
\]  \hspace{1cm} (5.16a)

\[
h = \left( \frac{c \tau_x}{\rho g' H_1} \right) \left( \exp \left( \frac{-y}{a} \right) \right) t
\]  \hspace{1cm} (5.16b)

\[
\hat{u} = \left( \frac{\tau_x}{\rho H_1} \right) \left( \exp \left( \frac{-y}{a} \right) \right) t
\]  \hspace{1cm} (5.16c)

where \( c \) is the speed of long internal waves and is given by:

\[
c^2 = \frac{g' H_1 H_2}{(H_1 + H_2)}
\]  \hspace{1cm} (5.17)

and \( a = c/f \) is the Rossby radius (Gill, 1982).
Figure 28 illustrates this solution with the coastline at the eastern boundary and a wind blowing towards the equator. The surface offshore transport produces upwelling at the coast. From Gill (1982), a wind stress of 0.1 N/m$^2$, $g'=0.03$ m/s$^2$, $H_1=100$ m, and $H_2 \gg H_1$ result in an upwelling velocity near 5 m/day. In the upper layer, Figure 28 shows a coastal jet which is formed by an increase of the alongshore current. In the lower layer, there is a poleward undercurrent flowing in the opposite direction of the wind (see Gill, 1982 for more details).

While coastal upwelling decrease SST, Gill (1982) mentioned that downwelling cannot warm the surface layer. Therefore, a succession of upwelling and downwelling events still tends to present cold waters along the coast.

An appreciation of the upwelling velocity can be obtained from the present SST data as follows:

Consider at first a calm warm upper layer of temperature $T_1$ and depth $h$ over a cold lower layer of temperature $T_0$ (sketch 1):
Without heat exchange, the thermocline would deepen due to mixing when the wind starts to blow. Then, the temperature of the upper layer becomes $T_2$ and its depth increases to $h + \delta h$ while the temperature of the lower layer remains the same ($T_0$) (sketch 2):

\[ (T_1 - T_0)h = (T_2 - T_0)(h + \delta h) \]  \hspace{1cm} (5.18)

which gives the new sea surface temperature $T_2$ due to wind mixing only:

\[ T_2 = \frac{(T_1h + T_0\delta h)}{(h + \delta h)} \]  \hspace{1cm} (5.19)

Now, consider both upwelling and wind mixing at the same time. Then, the thermocline would move up due to upwelling which results in divergence of the flow near the surface. The vertical displacement is given by 5.9. However, wind mixing would move the thermocline back down towards its original level, as in the above sketch 2, and decrease the vertical displacement. The resulting upwelling
velocity can be obtained by differentiation of 5.19 with respect to time and by substituting \( \partial(\delta h) / \partial t = w \). Therefore,

\[
\frac{\partial T_2}{\partial t} = \frac{(T_0 - T_1)w}{h} \tag{5.20}
\]

as \( \delta h \to 0 \)

where \( w \) is calculated from the equation 5.8.
Figure 23. Distribution of the stratification parameter calculated for the spring tide (left) and the neap tide (right) (Jardine et al, 1993).
Figure 24. Daily short-wave radiation $Q_s$ in watts per square metre (w/m$^2$) received at the surface of the ocean in absence of cloud. From Pickard and Emery (1982).

Figure 25. Long-wave radiation $Q_b$ in watts per square metre (w/m$^2$) from a water surface as a function of surface temperature and the overlying relative humidity in absence of cloud. From Pickard and Emery (1982).
Figure 26. Alongshore wind stress component in Newton per square metre (N/m²) at the buoy stations for the a) summer 1990 (Day 1 (May 1st) to day 153 (September 30th)), b) summer 1991, c) summer 1992. See Table 20 for missing data in 1990 and 1991.
Figure 26. (continued). d) Summer 1993 and e) Summer 1994.
Figure 27. Representation of two superposed shallow homogeneous layers of fluid. $H_1$, $H_2$ are the depths of the layers for a fluid at rest and $H=H_1+H_2$ is the total depth. The $z$ coordinate increases upward where $z = \eta(x,y,t)$ is the surface elevation and $z = -H_1 + h(x,y,t)$ is the position of the interface between the layers. From Gill (1982).

Figure 28. The solution for local upwelling at an eastern boundary. There is a coastal jet created in the upper layer in the same direction of the wind and an undercurrent in the opposite direction. From Gill (1982).
Chapter 6

Comparison Between SST, Synoptic Weather Maps and Wind Stress

From the previous chapter, the main factors responsible for the warming or cooling of the surface of the ocean over the short time scale (a few days to a few weeks) in the study area are the fluctuations in the cloud cover and wind. The cloud cover directly affects the amount of incoming short-wave radiation reaching the surface of the ocean and the net amount of long-wave radiation leaving the water. The wind is important for evaporation, upwelling (or downwelling) and vertical mixing. The emphasis is put on upwelling for the present study.

In this chapter three specific events are described from visual inspection in terms of SST variations and compared with synoptic weather maps, alongshore winds and/or wind stresses and cloud covers. Such analyses help to explain how the pressure patterns over the northeastern Pacific may interact with the upper layer of the ocean through the response of the surface water temperature variations over Queen Charlotte Sound and north of Vancouver Island. The three events, covering 30 to 32 days, have been selected based on an examination of the surface wind, with an emphasis on the direction because of the importance of the induced Ekman transport and vertical motion on the SST (Ikeda and Emery, 1984, Jardine et al, 1993, Fang and Hsieh, 1993, Staples and Hsieh, 1994). The first two events, covering the periods from July 9th to August 8th (day 70 to day 100) of 1990 and from July 3rd to August 4th (day 64 to day 96) of 1992, are dominated by northwesterly winds. The third event, extending from June 29th to July 29th (day 60 to day 90) of 1993, shows variable winds.
To help understand the following analysis, SST, wind stress and cloud cover time series are presented at the end of this chapter while synoptic weather maps can be found in appendices C, D and E. SST is in degrees Celsius and wind stress is in N/m$^2$ with positive values for Southeasterlies and negative values for Northwesterlies. Cloud cover is in oktas. In the following analysis, estimated values of the heat flux have been obtained from climatological data with $Q_r = +230$ W/m$^2$ and $Q_b = -95$ W/m$^2$, both in absence of clouds, $Q_e$ at -25 W/m$^2$ and $Q_h$ near 0 W/m$^2$. The resulting heat flux transferred to the ocean and available to increase the SST is nearly +115 W/m$^2$ in a cloud free environment during the periods selected. However, this value does not include any contributions from the advective term. The effect of clouds on daily $\Delta Q_r$ and $\Delta Q_b$ is considered by using the conversion factors seen in sections 5.2.1 and 5.2.2. The daily variations in latent and sensible heat fluxes ($\Delta Q_e$, $\Delta Q_h$), which depend mainly on changes in the wind speed, are generally much smaller than those involved with the short and long wave radiation and have been neglected for simplification.

6.1 Event 1: July 9th to August 8th of 1990 (day 70 to day 100)

6.1.1 Overview Of Synoptic Weather Situation (Figures. C1 to C31):

At the beginning of the period, a weakening low pressure system was moving into the Gulf of Alaska, forcing cold fronts to move across Queen Charlotte Sound on day 73 (July 12th) and on day 76 (July 15th). Thereafter, a ridge of high pressure dominated over the northeastern Pacific until day 93 (August 1st), while a trough of low pressure established itself along much of the B.C. coast from day 78 (July 17th) to
day 83 (July 22nd). After day 93 (August 1st) and until day 100 (August 8th) a series of frontal waves moved across the B.C. coast.

6.1.2 Comparison Between Synoptic Pressure Pattern, Wind, Cloud and SST Data (Figures 29, 30, 31 and C1 to C31): (Offshore data were available only for the buoys 46204 and 46207 for this event. See Figure 2 for locations.)

Day 70 to day 77 (July 9th-16th):

At the beginning of the event, synoptic maps show south to south-southeasterly winds reaching 15 knots (7.8 m/s) over Queen Charlotte Sound on day 71 (July 10th) ahead of a cold front. Weaker winds (less than 10 knots or 5 m/s) were observed around northern Vancouver Island due to the Californian ridge extending over the island. A weak northwesterly wind component began to develop over Queen Charlotte Sound on day 72 (July 11th) as a ridge of high pressure started to form near the Queen Charlotte Islands. In the wake of the cold front, the northwesterly wind component gained some strength with values reaching just above 15 to 20 knots (7.8 to 10.3 m/s) over Queen Charlotte Sound on day 73 (July 12th). The northwesterly wind speeds were 8.3 m/s and 8.7 m/s at the buoys 46207 and 46204 respectively just behind the front, which correspond to wind stress values near 0.10 N/m² and 0.11 N/m². Meanwhile, the SST remained relatively constant over much of the area during the first 3 days (day 70 to day 73) except for a 0.6°C increase at Kains Island on day 73 (July 12th), which may be explained by a decrease of the cloud cover causing an increase in the heat flux into the water. Behind the front, the SST dropped by 0.02°C and 0.03°C at the buoys 46207 and 46204 respectively and by 0.7°C at Kains Island, suggesting an upwelling response over much of the study area. However, comparing the cooling of the surface waters at coastal (0.7°C) and offshore stations (0.02°C and 0.03°C) suggests that the upwelling was more efficient at the coast. This
difference agrees with the equation 5.16a for the upwelling velocity showing an exponential decay away from the coast.

During the following two days (days 74-75/July 13th-14th) a thermal trough developed over the southern interior of B.C. resulting in a northeasterly flow aloft near the coastal mountains and a lee trough along the west coast of Vancouver Island. Also, there was ridge of high pressure off Vancouver Island extending northward across the Queen Charlotte Islands and into the northern interior of B.C. The resulting pressure gradient along the B.C. coast kept the Northwesterlies relatively constant over Queen Charlotte Sound with no significant variations in the SST until another front moved across Queen Charlotte Sound late on day 75 (July 14th). The cloud cover increased to above 6 oktas in the vicinity of this front, while the northwesterly wind stress values increased slightly to near 0.10 N/m² behind it. This was reflected by a SST drop of 0.4°C at the buoy 46207 and 0.1°C at the buoy 46204 on day 76 (July 15th) and as much as 1.8°C at Kains Island on day 77 (July 16th).

In more detail, the cloud cover increased from near 3 oktas to 6.5 oktas on day 76 (July 15th) at Cape Scott. This corresponds to a reduction in the amount of incoming solar radiation of approximately 35% and a reduction in the net loss of back radiation of approximately 40%. The estimated heat fluxes were +77 W/m² on day 75 (July 14th), down to +37 W/m² the next day. Despite the positive heat flux still transferred to the ocean on day 76 (July 15th), the SST did fall by 1.8°C during that day at Kains Island, which correlates with upwelling-favorable wind stress and an increase in the cloud cover. Wind data are not available for Kains Island; however, by reference with upstream data, it is reasonable to estimate the upwelling-favorable wind stress at nearly 0.10 N/m² as the second front moved
over northern Vancouver Island on day 76 (July 15th). The latter can certainly explain the SST variations via an upwelling response. From the relation 5.8, the upwelling velocity "w" at Kains Island was near 4 m/day on day 76 (July 15th), which gives a cooling of 0.4°C by using the relation 5.20. However, the observed cooling was 1.8°C suggesting other influences such as evaporation and/or mixing. Advection of cold water is also possible, but cannot be confirmed from available data. Also, the relation 5.20, which involves an upper layer well mixed with a uniform temperature "T" over a thermocline at a depth "h", may justify the difference between the observed and calculated SST variation due to an over-estimation of the depth "h" of the upper layer. Further offshore, cloud cover values are not available from buoy data but it is reasonable to assume cloud cover values near 6 oktas as the front moved near the sites of these buoys. Then, similar heat budget arguments can be made for offshore waters, which indicates some upwelling occurring also over Queen Charlotte Sound, but to a lesser degree. The calculated upwelling velocity was near 4.3 m/day at the buoys, giving a cooling near 0.5°C. The latter over-estimates the real cooling at the buoy 46204 by approximately 0.4°C, but is relatively close to the value observed at the buoy 46207.

Day 77 to day 83 (July 16th-22nd):

Behind the second cold front, the ridge was weaker and the Northwesterlies were generally decreasing until day 79 (July 18th) at the buoy 46207 and until day 80 (July 19th) at the buoy 46204. Meanwhile, the sky was clearing and the cloud cover was near 3 oktas at Cape Scott on day 77 and day 78 (July 16th and 17th). Then, the estimated heat flux was +76 W/m² which can explain a general increase of the SST beginning on day 78 (July 17th), except one day earlier at the buoy 46207. Furthermore, a larger increase in SST was initiated on day 80 (July 19th) over much
of the area as the scattered clouds were dissipating. On day 81 (July 20th), there was only 1 okta of cloud reported at Cape Scott while no cloud was observed at McInnes Island. Noteworthy is a trough of low pressure well developed over much of the B.C. coast, especially on day 81 (July 20th), giving northwesterly wind stress values near 0.09 N/m² at the buoy 46207, but somewhat weaker at the buoy 46204 (0.03 N/m²). Despite the upwelling-favorable wind stress, the SST was increasing, especially on day 80 (July 19th), showing a dominant effect from the incoming heat flux, unless there was an intrusion of warm water into the area. However, the SST at Kains Island began to decrease on day 82 (July 21st), which can be explained by coastal upwelling. This cooling phase also correlates with an increase in cloud cover to near 5 oktas at Cape Scott, despite an estimated heat flux near +50 W/m².

Further offshore, despite Northwesterlies, the buoy data show the warming process continuing until day 83 (July 22nd), suggesting the incoming heat flux and/or an intrusion of warm water into Queen Charlotte Sound being dominant. However, weaker warming of the surface waters near the buoy 46207, under stronger Northwesterlies, suggests some effect from upwelling in that area. Other cooling mechanisms are also possible.

**Day 83 to day 87 (July 22nd-26th):**

Between day 83 (July 22nd) and day 86 (July 25th), the Northwesterlies increased slightly, especially at the buoy 46204, as the trough was filling and the ridge building offshore. Also, fog and clouds invaded part of the area, especially along the coast, giving mainly obscured conditions near the northern end of Vancouver Island and along the mainland coast during this three day period. On day 83 (July 22nd), the estimated heat flux was +20 W/m² at McInnes Island and Cape Scott. Further
offshore, synoptic weather maps and cloud cover data for Cape St. James indicate more sunshine for an estimated heat flux near +65 W/m². Meanwhile, upwelling-favorable wind stress were still occurring at the buoys 46204 and 46207 with values between 0.01 N/m² and 0.08 N/m². Then, a cooling episode began at the buoy sites near or just after day 83 (July 22nd), which can be attributed to upwelling but also correlates with an increase in cloud cover from 5 to 8 oktas. At the coast, surface waters were generally cooling, which may also be caused by upwelling and/or reduced incoming solar radiation due to a relatively high cloud cover. This cooling episode ended on day 87 (July 26th) as a weak trough of low pressure moved across the Alaska panhandle. In the wake of this trough, more sunshine and weaker northwesterly wind stresses did promote a warming trend over Queen Charlotte Sound.

Day 88 to day 90 (July 27th-29th):

During this period the ridge re-built off Vancouver Island, across the Queen Charlotte Islands and over the northern interior of B.C. while a lee trough developed just west of Vancouver Island. As a result, the pressure gradient increased over much of the study area during that period; this is reflected by the increasing upwelling-favorable wind stress values at the buoy sites. However, the SST was increasing over much of the area, despite the upwelling-favorable wind stress between 0.02 N/m² and 0.06 N/m². The SST increased by 0.8°C at Kains Island and by 0.2°C at the buoy 46207 with an estimated heat flux near +88 W/m², suggesting a dominant influence from the incoming heat flux. On the other hand, the SST decreased slightly at the buoy 46204, (0.1°C), which may result from upwelling.
Day 90 to day 96 (July 29th-August 4th):

On day 90 (July 29th), the northwesterly wind reached relatively strong values giving northwesterly wind stresses up to 0.13 N/m², while the cloud cover increased to almost 8 oktas at Cape Scott for an estimated heat flux near +20 W/m². The SST, dropping by 1.6°C at Kains Island on that day, suggests an upwelling response. The calculated upwelling velocity is near 5.5 m/day which underestimates the real cooling by 1°C. Noteworthy is the correlation between the cloud cover increase and the beginning of the cooling phase. Further offshore the SST decreased only very slightly at the buoy 46204 and remained nearly stationary at the buoy 46207 indicating that the upwelling did not reach the surface over Queen Charlotte Sound.

By day 92 (July 31st), the northwesterly wind component began to decrease with the approach of a deep low pressure system from the Pacific. The alongshore wind shifted to the southeast on day 94 (August 2nd) with maximum speed near 4.9 m/s at buoy 46207 and 6.0 m/s at buoy 46204 on day 95 (August 3rd). The corresponding wind stress values were near 0.03 N/m² at the buoy 46207 and 0.05 N/m² at the buoy 46204. Meanwhile, the sky was mainly cloudy with between 5 and 8 oktas reported at Cape Scott, McInnes Island and Cape St.James for an estimated heat flux between +20 W/m² and +55 W/m². Despite Southeasterlies and incoming heat into the surface waters, a cooling phase was occurring at Kains Island between day 92 (July 31st) and day 95 (August 3rd). Noteworthy is a sharper cooling on day 95 (August 3rd) at Kains Island and the buoy 46204 which coincides with an increase in cloud cover from below 6 oktas to near 8 oktas on the previous day. Between day 92 (July 31st) and day 95 (August 3rd), the SST data for the buoy 46207 indicates a warming phase occurring further offshore, which suggests a dominant effect from the
incoming heat flux and/or the advective term by intrusion of warm water over Queen Charlotte Sound.

Thereafter, a series of frontal waves moving over northeast Pacific kept a dominant southeasterly wind component with speeds up to 5.9 m/s at the buoys until the end of the event. The corresponding wind stress values reached 0.05 N/m². Meanwhile the amount of cloud remained generally above 6 oktas for relatively low values of heat flux between $+20 \text{ W/m}^2$ and $+42 \text{ W/m}^2$. This was reflected by a cooling phase after day 97 (August 5th). The evaporation and mixing processes were probably important in this case.

6.1.3 Summary

During this event, the heat flux remained positive, even under cloudy skies. However, it appears that variations in cloud cover have some effect in regulating the SST variations. By looking at cloud cover data, this may be seen on days 76 (July 15th) and on day 90 (July 29th), where a cooling of the surface waters was coincident with a significant increase in the cloud cover. Such correlation can also be seen from a general cooling episode initiated on day 83 (July 22nd) when the sky became completely obscured with clouds. However, an upwelling response is shown by looking at small differences in the SST variations between the sites compared with the wind stress. For example, between day 77 (July 16th) and day 83 (July 22nd), the warming process was weaker at the buoy 46207 where the northwesterly wind stress was in fact stronger, therefore more upwelling-favorable. It is noteworthy that the cooling effect of the evaporation process may help to further understand the small SST variations.
From the atmospheric pressure patterns, the wind stress was upwelling favorable behind cold fronts and with a ridge of high pressure across the Queen Charlotte Islands. With cold fronts, the northwesterly wind stress reached similar values at both buoys as the fronts moved through. The SST response was stronger at Kains Island and weaker offshore especially at the site of the buoy 46204. The mean SST was nearly 1.6°C cooler at the coast compared with those further offshore (Table 21). Such upwelling distribution results from an exponential decay of the response away from the coast. With a ridge extending across the Queen Charlotte Islands, Northwesterlies tended to remain longer over the area, which could have been reflected in stronger upwelling event. This could not be confirmed from the data, especially because of interferences from other cooling mechanisms having. Table 21 just below gives the mean, maximum and minimum SST with the standard deviation for this event.

Maximum upwelling-favorable wind stresses developed with the presence of a ridge extending across the Queen Charlotte Islands and a lee trough. The strongest values were observed offshore, as indicated by the data of the buoy 46207, due to the position of the trough. In general, the troughs of low pressure tended to be centered near the coast so that the atmospheric pressure gradient was stronger offshore. Therefore, Northwesterlies tended to be stronger over offshore waters as compared with nearshore areas.

<table>
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<th>Stations</th>
<th>SST(mean)</th>
<th>SST(max)</th>
<th>SST(min)</th>
<th>STD</th>
</tr>
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<td>14.8</td>
<td>12.2</td>
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</tr>
<tr>
<td>46204</td>
<td>15.5</td>
<td>16.6</td>
<td>14.6</td>
<td>0.5</td>
</tr>
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<td>46207</td>
<td>15.3</td>
<td>16.5</td>
<td>14.7</td>
<td>0.4</td>
</tr>
</tbody>
</table>

Table 21. Mean, maximum and minimum SST (°C) with the standard deviation during event 1.
6.2 Event 2: July 3rd to August 4th of 1992 (day 64 to day 96)

6.2.1 Overview Of Synoptic Weather Situation (Figures D1 to D33):

At the beginning of the period, the northeastern Pacific was under the influence of a broad ridge of high pressure located near 140°W. However, after day 67 (July 6th) a series of frontal waves moved across the B.C. coast. On day 77 (July 16th) the offshore ridge re-built in the wake of the last frontal wave, while a lee trough developed along much of the coast. Except for a weak front that moved across the B.C. coast on day 82 (July 21st), a ridge remained anchored over northeast Pacific until day 84 (July 23rd). Thereafter, a low pressure system moved into the Gulf of Alaska, sending a series of frontal waves across the B.C. coast until day 90 (July 29th). Then, a ridge re-built and remained over the northeast Pacific until day 96 (August 4th).

6.2.2 Comparison Between Synoptic Pressure Pattern, Wind, Cloud and SST Data (Figures 32, 33, 34 and D1 to D33):

Day 64 to day 67 (July 3rd-6th):

During this first period, the pressure gradient ahead of the ridge was relatively weak so that the upwelling-favorable wind was generally less than 5 m/s at the buoys, slightly stronger at buoy 46208. This corresponds to wind stress values near 0.03 N/m², except near 0.05 N/m² at the buoy 46208. The cloud cover was above 6 oktas at Cape Scott, McInnes Island and Cape St. James during those three days giving an estimated heat flux between +20 W/m² and +40 W/m². Meanwhile, the SST increased during the first day and then decreased from day 65 to day 67 (July 4th-6th).
In particular, the SST increased by 2.3°C at Kains Island on the first day while there were 8 oktas of clouds (estimated heat flux near +20 W/m²) with northwesterly wind stress near 0.01 N/m². The latter appears relatively weak and probably not sufficient to initiate an upwelling response, which would give the incoming heat flux more importance. However, the SST at Kains Island began to drop on day 66 (July 5th) when the northwesterly wind stress reached near 0.04 N/m² at the buoy 46207 with no significant changes in the cloud cover. The next day, the cloud cover decreased to near 7 oktas for an estimated heat flux near +30 W/m² and the upwelling-favorable wind stress increased further to above 0.04 N/m². This was reflected in a 2.3°C temperature drop at Kains Island suggesting an upwelling response. Further offshore, the SST behaved in a similar fashion at the buoy 46207 but with smaller temperature variations. At the other sites (buoys 46204, 46185 and 46208), there is insufficient evidence to give satisfactory explanations for the SST variations.

**Day 68 to day 77 (July 7th-16th):**

On day 68 (July 7th), an upper trough moved eastward across Queen Charlotte Sound and northern Vancouver Island early in the day. In its wake, the offshore ridge gained more strength, as well as the upwelling-favorable winds. Maximum speeds between 6.2 m/s and 8.0 m/s were recorded at the buoy sites near day 69 (July 8th), which correspond to upwelling-favorable wind stress values between 0.05 N/m² and 0.09 N/m². Also, the cloud cover decreased to near 6 oktas giving an estimated heat flux near +43 W/m². Meanwhile, the SST increased at Kains Island and the buoy 46204, which may be caused by more heat coming into the sea surface and/or by an intrusion of warm water around northern Vancouver Island. Further offshore, the surface water was slightly warming at buoy 46208. However, the SST
was dropping at the buoys 46207 and 46185, which correlates with upwelling-favorable wind stress.

The first of a series of frontal waves moved across the area late on day 70 (July 9th) increasing the cloud cover to above 6 oktas at McInnes Island and to near 8 oktas at Cape Scott and Cape St. James. This increase correlates with a short cooling episode at Kains Island on that day, although the estimated heat flux was still positive (near +20 W/m$^2$). On the other hand, stronger upwelling-favorable wind stress and coastal upwelling can be assumed for a few hours in the wake of the wave and could also justify this cooling episode. Meanwhile, the SST was increasing at most buoys until day 71 (July 10th) and after, especially the buoy 46204, suggesting a dominant response to the incoming heat flux, while the upwelling-favorable wind stress was below 0.05 N/m$^2$ and decreasing. However, further offshore the SST was decreasing at buoy 46207 during that period, which could be explained by somewhat stronger Northwesterlies (wind stress values between 0.05 N/m$^2$ and 0.06 N/m$^2$) in that area, making an upwelling response more likely. Also, more evaporation due to stronger winds was probably contributing to the cooling phase.

On day 72 (July 11th), a second frontal wave moved across the area and the northwesterly winds dropped to near 0 m/s at most buoy sites. The sky was mainly cloudy as the wave moved through with above 6 oktas of cloud reported at Cape Scott, McInnes Island and Cape St James on that day. The estimated heat flux was between +20 W/m$^2$ and +43 W/m$^2$. These factors: weak upwelling-favorable winds (if any) and incoming heat, supported a warming trend for all stations. After day 74 (July 13th), a weak ridge built temporarily along the coast of B.C. and the cloud cover began to decrease. The increasing heat flux into the sea surface supported the warming phase until day 76-77 (July 15th-16th). Meanwhile the upwelling-favorable
winds (if any) were generally weak until the last frontal wave moved through the area early on day 77 (July 16th). Just before this wave moved across the area, the estimated heat flux was between +30 W/m² and +56 W/m² while the northwesterly wind stress (if any) was below 0.02 N/m².

**Day 77 to day 84 (July 16th-23rd):**

After day 77 (July 16th), the offshore ridge built up over the northeastern Pacific giving further clear skies to the area. The estimated heat flux increased to +110 W/m² between day 77 and day 79 (July 16th-18th). By day 79 (July 18th), the ridge had rotated clockwise and was aligned through the Queen Charlotte Islands and the northern interior of B.C. Further east, there was a thermal trough over the southern interior of B.C. resulting in a lee trough along the B.C. coast. The intensifying pressure gradient generated upwelling-favorable winds with maximum values reaching between 8.7 m/s and 12.0 m/s at the buoy sites near day 79 and day 81 (July 18th-20th). The corresponding wind stress values were between 0.12 N/m² and 0.25 N/m². Also, there was a low pressure system moving into the Gulf of Alaska on day 79 (July 18th), squeezing the offshore ridge between 135°W-140°W so that the pressure gradient and associated northwesterlies were relatively strong over Queen Charlotte Sound and northern Vancouver Island. As a result of upwelling-favorable wind stresses, a cooling phase began near or just before day 77 (July 16th) at most stations despite mainly sunny skies. Noteworthy is a rapid cooling that occurred on day 80 (July 19th) at Kains Island (3.7°C) as the coastal trough reached its maximum development. The upwelling-favorable wind stress values reached between 0.10 N/m² and 0.23 N/m², suggesting an upwelling response. The calculated upwelling velocity was near 5 m/day which under-estimates the real cooling by as much as 3°C. The estimated heat flux near +90 W/m² at Cape Scott,
makes the upwelling process more evident. However, at the buoy sites the
temperature decrease was much smaller (less than 0.4°C), which suggests an
upwelling response much weaker offshore as predicted by equation 5.16a. The
calculated upwelling velocity was near 5 m/day which over-estimates the cooling by
as much as 94%.

On day 81 (July 20th), a weak front moved across the northern part of the B.C. coast,
bringing clouds into the study area. In particular, the cloud cover increased to near 5
oktas at Cape Scott, and McInnes Island on day 81 (July 20th), decreasing the
estimated heat flux to near +54 W/m². Also, the upwelling-favorable winds
decreased in behind the front, especially in the northern part of Queen Charlotte
Sound. However the northwesterly wind stress remained relatively strong until
day 84 (July 23rd), with values above 0.6 N/m² due to the offshore ridge being still
relatively strong. Meanwhile, the SST generally decreased suggesting a dominant
upwelling response.

Day 84 to day 90 (July 23rd-29th):

After day 84 (July 23rd), a low pressure system moving into the Gulf of Alaska sent a
series of frontal waves across the B.C. coast until day 90 (July 29th). As a result, the
upwelling-favorable winds stopped and the clouds moved in. The estimated heat
flux varied between +20 W/m² and +43 W/m² from day 84 to day 90 (July 23rd-29th)
except up to +65 W/m² on day 86 (July 25th). Meanwhile, the SST generally
increased between 0.5°C and 1.0 °C, which suggests a dominant influence of the
incoming heat flux. It is noteworthy that the increase in SST began later (on day
86/July 25th) at Kains Island, which correlates with a relative minimum in the
cloud cover. Thereafter, the SST began to fall on or just after day 90 (July 29th) with increasing Northwesterlies.

Day 90 to day 96 (July 29th-August 4th):

On day 90 (July 29th) the last frontal wave moved across the area and the ridge built offshore. On days 91 and 92 (July 30th-31st), the ridge was coupled with a thermal trough over the southern interior of B.C. and Washington State. There also was a weak lee trough just west of Vancouver Island extending along the eastern boundary of Queen Charlotte Sound. This increased the pressure gradient and the Northwesterlies over much of the study area. The upwelling-favorable wind stress reached between 0.07 N/m\(^2\) and 0.17 N/m\(^2\) on day 92 (July 31st). The SST dropped by 1.1°C at Kains Island, 0.7°C at the buoy 46204 and 0.8°C at the buoy 46185 which suggests an upwelling response. Noteworthy is the clearing during these days which re-enforced a dominant upwelling response: more heat was available, but the SST did fall. The estimated heat flux reached near +110 W/m\(^2\) on day 92 (July 31st).

After day 92 (July 31st), some clouds moved over the study area, especially in the southern portion. Also, the lee trough was relatively weak which is reflected by a decrease in the Northwesterlies for the next following two days. Meanwhile, the SST generally increased which may be explained by an estimated heat flux between +40 W/m\(^2\) and +65 W/m\(^2\). However the SST was temporarily falling at Kains Island on day 94 (August 2nd) suggesting an intrusion of cold water. By day 95 (August 3rd), the Northwesterlies generally increased due to the offshore ridge, especially in the southern part of Queen Charlotte Sound and around the northern end of Vancouver Island. Then the SST dropped at Kains Island and the southern most buoys (46204 and 46207) suggesting an upwelling response. Further north, the
SST increased slightly at the buoys 46185 and 46208, which may be explained by sunny breaks.

6.2.3 Summary

During this event, two sequences are worthy of attention. The first one occurred between day 72 and day 77 (July 11th-16th) while the SST generally increased at all stations. During this short period, a series of frontal waves moved across the B.C. coast so that the alongshore wind stress was not favorable for upwelling. As a result, the SST was more vulnerable to the incoming heat flux across the surface of the ocean which increased from +20 W/m² on day 72 (July 11th) to near +100 W/m² on day 77 (July 16th). Also, as each wave approached the area, the winds shifted temporarily to being from a more southerly direction, making advection of warm waters from the south possible. These wind variations are not shown by the daily wind data since they may occur within less than 24 hours, but are inferred from work experience as a meteorologist.

The second interesting sequence followed immediately upon the first one and shows a dominant effect from the advective term via the upwelling process. Between day 77 and day 84 (July 16th-23rd), the pressure pattern was dominated by a ridge of high pressure offshore and a trough along the B.C. coast giving Northwesterlies to Queen Charlotte Sound. In particular, the ridge extended across or just northwest of the Queen Charlotte Islands and into the northwestern interior of B.C. from day 79 to day 81 (July 18th-20th), generating relatively strong upwelling-favorable winds over Queen Charlotte Sound and in the vicinity of the northern end of Vancouver Island. This sequence was reflected in the SST by a general
decrease for several days at all stations. In particular, the SST dropped by 2.9°C at Kains Island, by 1.2°C and 1.6°C at the buoys 46204 and 46207, and by near 4.0°C at the buoys 46185 and 46208. Less cooling occurred offshore which is in agreement with the exponential form of the upwelling velocity (equation 5.16a). The upwelling-favorable wind stress was stronger in the northern part of Queen Charlotte Sound, justifying a stronger cooling phase near the buoys 46185 and 46208. The following table (Table 22) gives the mean, maximum and minimum SST during this event with the standard deviation.

<table>
<thead>
<tr>
<th>Stations</th>
<th>SST(mean)</th>
<th>SST(max)</th>
<th>SST(min)</th>
<th>STD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kains Island</td>
<td>13.3</td>
<td>15.9</td>
<td>11.8</td>
<td>1.0</td>
</tr>
<tr>
<td>46204</td>
<td>14.3</td>
<td>15.0</td>
<td>13.5</td>
<td>0.4</td>
</tr>
<tr>
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<td>13.5</td>
<td>0.4</td>
</tr>
<tr>
<td>46185</td>
<td>14.2</td>
<td>15.9</td>
<td>11.3</td>
<td>0.9</td>
</tr>
<tr>
<td>46208</td>
<td>13.2</td>
<td>14.4</td>
<td>9.8</td>
<td>0.9</td>
</tr>
</tbody>
</table>

Table 22. Mean, maximum and minimum SST (°C) with the standard deviation.
6.3 Event 3: June 29th to July 29th of 1993 (day 60 to day 90)

6.3.1 Overview Of Synoptic Weather Situation (Fig E1 to E31):

During the first two days of this period (day 60-62/June 29th-July 1st) much of the B.C. coast was under the influence of a weak low pressure system. Then, a ridge started to develop offshore, especially in the wake of a weak frontal wave that moved across the B.C. coast on days 63-64 (July 2nd-3rd). This ridge remained anchored over the northeastern Pacific until day 81 (July 20th) and was coupled with a trough of low pressure along much of the B.C. coast between day 69 and day 77 (July 16th). Thereafter, a series of relatively weak frontal waves moved across the B.C. coast from day 82 to day 90 (July 21st-29th).

6.3.2 Comparison Between Synoptic Pressure Pattern, Wind, Cloud and SST Data (Fig. 35, 36, 37 and E1 to E31): (The following analysis does not include the area near the site of the buoy 46208 because cloud cover data were not available for Cape St.James during that period.)

Day 60 to day 64 (June 29th-July 3rd):

On days 60 and 61 (June 29th-30th) a weakening low pressure system gave light and variable winds to Queen Charlotte Sound and the northern end of Vancouver Island. Meanwhile the sky was mainly cloudy with an estimated heat flux between +20 W/m² and +40 W/m². As a result, the SST was increasing at all stations.

On day 62 (July 1st), the offshore ridge began to build so that Northwesterlies started to develop over Queen Charlotte Sound and the northern end of Vancouver Island. Northwesterly winds between 2 and 5 m/s were reported at the buoy sites on days 62-63 (July 1st-2nd), corresponding to upwelling-favorable wind stress values.
between 0.01 N/m$^2$ and 0.04 N/m$^2$. Meanwhile, the SST decreased at buoys 46204 and 46185, which reflects a dominant upwelling response. The SST increased slightly at Kains Island and the buoy 46207, which can be explained by the heat flux (estimated at +40 W/m$^2$) being dominant.

Between day 63 and day 64 (July 2nd-3rd), a weak frontal wave moved across Queen Charlotte Sound and Vancouver Island so that the upwelling-favorable winds stopped. Instead, there was a southeasterly wind component reported at most buoys with speeds near 3 m/s for wind stress values near 0.02 N/m$^2$. Also, there were near 6.5 oktas of clouds reported at Cape Scott and McInnes Island on day 64 (July 3rd) for an estimated heat flux near +37 W/m$^2$. Meanwhile, the SST increased slightly at most stations, which can be due to the heat flux into the surface waters and/or an intrusion of warm water.

**Day 65 to day 82 (July 4th-21st):**

On day 65 (July 4th), a broad high pressure system developed over the northeastern Pacific, as well as Northwesterlies over Queen Charlotte Sound and around Vancouver Island. Also, clouds were dissipating so that the incoming heat flux was increasing from an initial value near +37 W/m$^2$. On day 66 (July 5th), the heat flux was estimated between +55 W/m$^2$ to +70 W/m$^2$. Meanwhile, a cooling phase had started at most stations near day 66-67 (July 5th-6th), showing clearly an upwelling response. The cooling phase started one day earlier at Kains Island, which suggests that upwelling is more efficient near the coast. Some cooling from evaporation and/or mixing could also help starting this cooling phase.
Later, the offshore ridge continued to build along a northwest-southeast line and reached its maximum development on day 68 (July 7th). As a result, the estimated heat flux reached maximum values between $+60 \, \text{W/m}^2$ in the south and $+110 \, \text{W/m}^2$ in the north. The upwelling-favorable wind stress was also relatively strong with values near $0.14 \, \text{N/m}^2$ in the south to $0.20 \, \text{N/m}^2$ in the north on that day. This was reflected by further upwelling at all stations. In particular, the SST dropped by $2.1^\circ\text{C}$ at Kains Island, $1.7^\circ\text{C}$ at the buoy 46204, $0.3^\circ\text{C}$ at the buoy 46207 and as much as $3.1^\circ\text{C}$ at the buoy 46185 between day 65 and day 68 (July 4th-7th). The calculated upwelling velocity, between $6 \, \text{m/day}$ and $8 \, \text{m/day}$, generally approximates the observed cooling. However, the cooling is largely under-estimated at the buoy 46185 and over-estimated at the buoy 46207. Noteworthy is the estimated heat flux largely positive on day 68 (July 7th) which re-enforced a dominant upwelling response. On day 69 (July 8th) the offshore ridge started to migrate westward and a trough developed along the coast. Then, the upwelling-favorable winds began to decrease. On day 70 (July 9th) the values decreased and were between $0.09 \, \text{N/m}^2$ and $0.14 \, \text{N/m}^2$. Therefore, there was less upwelling and a warming phase began at most stations. Meanwhile the cloud cover was increasing, but the estimated heat flux remained positive with values near $+30 \, \text{W/m}^2$ at Cape Scott and near $+92 \, \text{W/m}^2$ at McInnes Island. This supported the warming process.

After day 70 (July 9th), the trough started to fill along the coast and the pressure gradient continued to weaken over Queen Charlotte Sound and around Vancouver Island. Then, the Northwesterlies continued to decrease until day 79 (July 18th). Also, more clouds invaded the area, giving an estimated heat flux between $+20 \, \text{W/m}^2$ and $+40 \, \text{W/m}^2$. Meanwhile, the warming trend continued at the stations located close to the shore, especially at the buoys 46204 and 46185. Noteworthy are short cooling episodes at Kains Island between day 73 and day 75 (July 12th-14th) and
on day 78 (July 17th) embedded in a general warming phase. These cooling episodes are correlated with the passage of two weak fronts, which suggests an upwelling response from a band of stronger Northwesterlies in the wake of each front. Part of the cooling may also be caused by the evaporation and/or mixing processes. Further offshore, the SST oscillated at the buoy 46207 with temperature variations generally less than 0.5°C until the warming phase started on day 76 (July 15th).

From day 79 (July 18th) to day 81 (July 20th), the offshore ridge and the Northwesterlies decreased further. The northwesterly wind stress was between 0.01 N/m² and 0.02 N/m², except near 0.04 N/m² further offshore (buoy 46207). Meanwhile, the sky remained mainly cloudy giving an estimated net heat flux between +20 W/m² and +40 W/m² at Cape Scott and McInnes Island. However, the SST was generally falling at most stations suggesting a dominant effect from the advective term, most likely through an intrusion of cold water since the upwelling-favorable wind stress was relatively small.

**Day 82 to day 86 (July 21st-25th):**

On day 82 (July 21st), a deepening low pressure system moving from the west forced the ridge onto the coast. Then, clouds began to dissipate. The estimated heat flux near +20 W/m² on day 82 (July 21st) increased to near +30 W/m² on day 83 (July 22nd). The alongshore winds shifted from northwesterly to southeasterly on day 82 (July 21st) and were relatively strong. The southeasterly wind stress was up to 0.16 N/m² during those two days. Meanwhile, the SST decreased at all stations, especially on day 83 (July 22nd) when the wind speed was relatively strong. This
suggests a dominant effect from wind mixing, although a part of the cooling may be attributed to the evaporation process.

On day 84 (July 23rd), the offshore ridge re-developed in the wake of the low pressure system. As a result, the sky continued to clear and the upwelling-favorable winds re-developed over Queen Charlotte Sound and near Vancouver Island. Maximum heat flux near +70 W/m$^2$ occurred on day 85 (July 24th). Maximum upwelling-favorable winds were recorded at the buoy sites on day 86 (July 25th) when a lee trough developed just west of Vancouver Island. The corresponding wind stress values ranged between 0.07 N/m$^2$ and 0.11 N/m$^2$ from the northwest. Meanwhile, the SST was generally increasing, suggesting a dominant effect from the incoming heat flux.

**Day 87 to day 90 (July 26th-29th):**

On day 87 (July 26th) a cold front approached Queen Charlotte Sound and the northern end of Vancouver Island spreading clouds to the area. The cloud cover increased to near 8 oktas on that day, which correlates with the onset of a cooling phase at Kains Island. From synoptic weather maps, winds were generally light and variable near the front. The northwesterly wind stress was below 0.01 N/m$^2$ at the buoy sites and probably not strong enough to initiate an upwelling response. In fact, the SST increased at all buoy sites. After day 87 (July 26th), the cloud cover decreased to between 4 and 7 oktas, which supported the warming process offshore almost until day 90 (July 29th). Meanwhile, the SST continued to decrease at Kains Island until day 89 (July 28th) and then increased.
6.3.3 Summary

The most important sequence during this period occurred between day 65 and 68 (July 4th-7th) while a ridge was building just off the coast of B.C. The relatively strong Northwesterlies ahead of the ridge were reflected by a cooling phase at all stations due to upwelling. Noteworthy is the temperature drop being larger near the coast, especially at the site of the buoy 46185. Later, when a trough developed along the B.C. coast, the upwelling-favorable wind stress continued, but at decreasing strength. This was reflected by a general warming phase of the surface waters, showing a direct response to the wind stress. Finally, a temperature drop near 0.7°C recorded at the buoy sites on day 83 (July 22nd) reflected a dominant effect from vertical mixing due to southeasterly winds up to 25 knots (12.8 m/s) for wind stress values near 0.25 N/m². Some cooling from the evaporation process was also probably occurring. Table 23 gives the mean, maximum and minimum SST during this event with the standard deviation.

<table>
<thead>
<tr>
<th>Stations</th>
<th>SST(mean)</th>
<th>SST(max)</th>
<th>SST(min)</th>
<th>STD</th>
</tr>
</thead>
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</tr>
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<td>46207</td>
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<td>14.7</td>
<td>12.9</td>
<td>0.4</td>
</tr>
<tr>
<td>46185</td>
<td>13.6</td>
<td>14.7</td>
<td>11.0</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Table 23. Mean, maximum and minimum SST (°C) with the standard deviation.
Figure 29. Wind stress time series for event 1: July 9th-August 8th (day 70-100) of 1990. Wind stress values are in Newton per square metre (N/m²).
Figure 30. SST time series for event 1: July 9th-August 8th (day 70-100) of 1990. SST values are in degree Celsius (°C). The dotted lines correspond to the summer mean SST.
Figure 31. Cloud cover time series for event 1: July 9th-August 8th (day 70-100) of 1990. Cloud cover values are in oktas.
Figure 32. Wind stress time series for event 2: July 3rd–August 4th (day 64-96) of 1992. Wind stress values are in Newton per square metre (N/m²).
Figure 33. SST time series for event 2: July 3rd-August 4th (day 64-96) of 1992. SST values are in degree Celsius (°C). The dotted lines correspond to the summer mean SST.
Figure 34. Cloud cover time series for event 2: July 3rd-August 4th (day 64-96) of 1992. Cloud cover values are in oktas.
Figure 35. Wind stress time series for event 3: June 29th-July 29th (day 60-90) of 1993. Wind stress values are in Newton per square metre (N/m²).
Figure 36. SST time series for event 3: June 29th-July 29th (day 60-90) of 1993. SST values are in degree Celsius (°C). The dotted lines correspond to the summer mean SST.
Figure 37. Cloud cover time series for event 3: June 29th-July 29th (day 60-90) of 1993. Cloud cover values are in oktas.
Chapter 7

Satellite Remote Sensing of Sea Surface Temperature

The previous chapters have explored the relationship between buoy data and information from lighthouse stations. By looking at specific events in terms of SST, cloud cover and wind stress, a relatively good correlation was found between Kains Island and the buoy sites, although the daily SST tends to vary more at Kains Island than at the buoy sites due to coastal effects. Now, to go a step further in this study, it would be interesting to find out how representative of the whole study area are those buoys in terms of SST. This is done in this chapter by using Advanced Very High Resolution Radiometer (AVHRR) satellite imagery. The latter will give a larger picture of the SST distribution over the study area and will allow a further understanding of the relationship between coastal and buoy stations. However, it was not possible to get a time series for any events analyzed in the previous chapter because of frequent fog and clouds over Queen Charlotte Sound and the northern end of Vancouver Island. Nevertheless, useful information could be obtained from a few AVHRR infrared satellite images by visual inspection of the radiance distribution and by plotting transects. Analyses of these transects are presented in section 7.7.

7.1 Data Source

Satellite images have been provided by the advanced very high resolution radiometer (AVHRR) on board the NOAA 10 and NOAA 11 satellites. These
satellites are in sun-synchronous orbits 870 km above the earth so that a fixed geographical location on the earth is always observed at the same local solar time on each pass every day. This can be achieved by a suitable choice of the orbit's inclination (near 100°) so that the orbit plane is fixed relative to the sun as the earth travels around the sun once a year. For the NOAA satellites, a complete orbit takes approximately 103 minutes so that they are repeated 14 times a day. As the earth rotates underneath, the satellites scan from north to south over one face of the earth and south to north over the other face giving a sampling rate of two images a day for a given location.

The AVHRR sensor is an optical instrument, providing multi-spectral imaging by sensing reflected solar radiation in two visual or near visual wavebands (channels 1 and 2) and by sensing thermal emissions in three infrared wavebands (channels 3, 4 and 5) as follows:

<table>
<thead>
<tr>
<th>Band</th>
<th>Wavelength (μm)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Band 1</td>
<td>0.58 - 0.68</td>
<td>Visible</td>
</tr>
<tr>
<td>Band 2</td>
<td>0.73 - 1.10</td>
<td>Near infrared</td>
</tr>
<tr>
<td>Band 3</td>
<td>3.55 - 3.93</td>
<td>Thermal infrared</td>
</tr>
<tr>
<td>Band 4</td>
<td>10.3 - 11.3</td>
<td>Thermal infrared</td>
</tr>
<tr>
<td>Band 5</td>
<td>11.5 - 12.5</td>
<td>Thermal infrared</td>
</tr>
</tbody>
</table>

The main waveband used in this chapter corresponds to Band 4 in the infrared (IR) region. This is the best channel to measure the SST because of a relatively good balance between the peak of the thermal emission and the atmospheric transparency in the 10.3 - 11.3 μm wavelength range. On the other hand, Band 1 is used for cloud detection and land references.

The spatial resolution of AVHRR imagery is 1.1 km at the sub-satellite point and decreases to near 7 km at the edge of the swath that is 2580 km wide. The calibration
of IR images is relatively important to get the most accurate absolute values of the SST. It is achieved with the use of a blackbody cavity at a known temperature carried on board the satellite. The radiometer resolution is measured by a noise equivalent difference temperature (NEΔT) parameter where ΔT refers to the apparent change in temperature. For typical SST with values between 270 and 300 °K, the resolution is near 0.13 - 0.10 °K for Bands 4 and 5. In practice, the value of NEΔT is generally higher due to calibration errors, atmospheric effects (absorption, re-emission, scattering and reflection). Large absolute errors may also be introduced by sub-pixel (i.e. less than 1.1 km) cloud contamination (Robinson, 1994). To remove these errors, further calibration has been done in this thesis by using in situ SST measurements from buoys.

A summary of the basic principle underlying the operation of IR sensors to get SST values is presented in the following section.

### 7.2 Characteristics Of Infrared Radiation

Infrared radiation is emitted by the sea surface as a result of vibrational and rotational excitations within the molecules of water. The spectral range of infrared radiation includes Band 3, 4 and 5. Based on Planck's theory and the emission of a perfect emitter at 300 °K (average temperature of the sea surface), the peak of the ocean radiation spectrum is at 10-12 μm (Fig. 38). The spectral emittance for a body at temperature T is described by Planck's relationship as follows:

\[
M(\lambda) = \frac{C_i}{\lambda^3[\exp(C_2 / \lambda T) - 1]} 10^{-6} \text{ W/m}^2/\mu\text{m} \tag{7.1}
\]
where $\lambda$ is the wavelength ($\mu$m), $C_1$ and $C_2$ are constants with the following values: $C_1 = 3.74 \times 10^{-16} \text{W/m}^2$ and $C_2 = 1.44 \times 10^{-2} \text{m}^\circ \text{K}$. $M(\lambda)$ is the emittance (also called exitance) for a particular $\lambda$ which expresses the radiant flux density per unit bandwidth centered at $\lambda$, leaving a unit area, regardless of the direction.

Equation 7.1, based on ideal thermodynamic laws, is valid only if the source of emission is a perfect emitter (i.e. blackbody). For the ocean surface, which is not a perfect blackbody, the emittance is described in term of its spectral emissivity, $e(\lambda)$. The latter is the ratio of the emittance from a real body at temperature, $T$, compared to a blackbody at the same temperature, $T$, and wavelength, $\lambda$:

$$e(\lambda) = \frac{M(\lambda)}{M(\lambda)_{BBT}} \quad (7.2)$$

The spectral emissivity is almost constant with respect with temperature, but varies significantly with wavelength.

The integration of equation 7.1 over all wavelengths results in the total exitance of a blackbody, $M$, which is the Stephan-Boltzman Law as follows:

$$M = \sigma T^4 \text{ W/m}^2 \quad (7.3)$$

where $\sigma = 5.669 \times 10^{-8} \text{ W/m}^2\text{K}^4$ is Stephan's constant.

The wavelength, $\lambda_{\text{max}}$, corresponding to the peak emission is given by Wien's displacement law:

$$\lambda_{\text{max}} T = 2897 \text{ (\mu m}^\circ \text{K)} \quad (7.4)$$
and is shown in Figure 38 where the peak of the emission for a surface at 300 °K such as the ocean occurs near 10 μm. The Wien's displacement also shows that by making use of sensors detecting the range covered by Band 4 (10.5 - 11.3 μm), it is possible to measure sea surface temperature.

7.3 Satellite Measurements

The AVHRR is a sensor measuring the radiant flux, ϕ, in watts (w) per unit solid angle, Ω, in steradian (sr), per unit normal area (m) per unit bandwidth (μm).

By definition, the radiant flux per unit solid angle leaving a point source in a specific direction is the radiant intensity:

\[ I = \frac{d\Phi}{d\Omega} \text{ w/sr} \tag{7.5} \]

with the assumption that the sea surface is Lambertian, i.e. emits isotropically (Robinson, 1985). If the sea surface is divided into small areas, the radiant intensity leaving a unit area normal to the direction of the remote sensing is called the radiance, L, and is given by the following:

\[ L = \frac{dI}{d(\text{Acos} \theta)} \text{ w/sr/m}^2 \tag{7.6} \]
By using the equation 7.1, since the emitted radiation is independent of the direction, the radiance for a given wavelength is given by:

\[ L(\lambda) = \frac{M(\lambda)}{\pi} \text{ W/m}^2/\text{sr/\mu m} \]  

(7.7)

In other words, \( L(\lambda) \) is the radiant flux, \( \Phi \), in watts (w) per unit solid angle, \( \Omega \), in steradian (sr) per unit normal area (m) per unit bandwidth (mm) for a blackbody.

Then, it is possible to find the temperature, \( T \), by using (7.7) and inverting (7.1). This temperature, \( T \), is the brightness temperature for a particular wavelength assuming that the source is a perfect emitter. If the emissivity is known, then the true temperature can be found from the radiance of the blackbody using the following expression:

\[ M(\lambda)_{BB} = \frac{\pi L(\lambda)_{measured}}{\epsilon(\lambda)} \times 10^{-6} \text{ W/m}^2/\mu\text{m} \]  

(7.8)

For the sea surface, the emissivity is approximately 0.98. It varies little with temperature, wavelength and surface roughness. However, the value may be affected by surface slicks or debris (Robinson, 1985).

The AVHRR sensor measures \( L(\lambda) \), and the apparent temperature of the sea surface is obtained by using equations 7.8 and 7.1. However, the IR radiation measured by the satellite, \( L(\lambda) \), is emitted by a very thin layer of molecules at the surface of the ocean. For the wavelengths of channel 4 and 5 (10.5 to 12.5\( \mu \)m), the thickness of that layer is less than 0.1 mm (Grassl, 1976). Therefore the satellite remote sensing derived sea surface temperature is a skin temperature. Typical values of the
temperature deviation between the skin temperature and the bulk sea surface temperature measured by buoys and lighthouse stations within the top metre are near 0.01 °K to 0.05 °K with the skin of the ocean usually being cooler than the layer just underneath (Schluessel, 1990). Useful interpretation of SST measured by the AVHRR can be made if the thermal vertical structure of the near surface layer is known. The next section presents an overview of a typical upper layer thermal structure, followed by some limitations it may have on the satellite remote sensing of SST.

7.4 Upper Layer Thermal Structure

On the basis of Planck's theory, the sun emits radiation like a blackbody at 6000 °K. The spectral irradiance reaching the sea level covers mainly the visible region and a small part of the infrared. The full spectrum falls approximately within 0.30 µm to 2.5 µm with the peak emission near 0.50 µm (Fig. 39). Most of this energy is absorbed by the ocean within the first few metres (Fig. 39) and used to heat the surface layer or to promote photosynthesis (Thurman, 1994). The first wavelengths to be absorbed are the shorter ones, i.e., the infrared, followed by the red while the blue light can been seen down to 100 m due to scattering by water molecules and particles. At 100 m, only 1% of the energy incident on the surface remains.

The incoming solar energy is transmitted into the ocean by conduction and turbulent mixing due to winds and waves. As mentioned in section 5.2.3, conduction is a very slow process and allows only a fraction of the heat to be transferred downwards. However, turbulent mixing is much more efficient in redistributing the energy downward. Therefore, this process controls the
temperature of the mixed layer depth which can be as thick as 200-300 m at mid-latitudes in the open oceans but as little as 10 m in protected coastal areas during the summer. Below the mixed layer, the temperature decreases rapidly to approximately 1000 m. This layer forms the permanent thermocline. Then, from 1000 m to the bottom of the ocean, the temperature decreases only very slowly to 0°C to 3°C (Fig. 40).

During the summer, due to an increase of the incoming solar radiation and generally less turbulent mixing, a seasonal thermocline usually forms above the permanent one. It is not unusual to observe a seasonal thermocline in May near 20 m with a mixed layer above it. As the summer progresses, the surface mixed layer deepens so that seasonal thermoclines reach near 40 m in September (Fig. 41). The actual depth of the summer upper mixed layer varies with time and space as it is closely related to local wind forcing (section 5.3.2) and solar radiation. Furthermore, during warm sunny days without much wind stress, a diurnal thermocline can also develop on top of the seasonal one. Such a thermocline, occurring generally at a few metres deep with a typical temperature difference near 1-2°C (Open series, 1991), is more localized and is destroyed at night due to heat losses to the atmosphere (Robinson, 1985) (Fig. 42). On the other hand, the higher the surface temperature, the more stable is the upper layer due to the non-linear relationship between the density \( \rho \) and temperature \( T \): A change in density \( \Delta \rho \) is larger for a unit change in temperature \( \Delta T \) under warm conditions than cool ones. Hence, warm waters tend to be more stable than colder waters in similar environments.
7.5 Limitations Of IR Sensing Of The Ocean Surface

The first and main limitation of satellite-sensed SST depends on the difference between the sea skin temperature and the bulk sea surface temperature because oceanographers are more interested in the latter when studying ocean dynamics and water properties.

7.5.1 Characteristics Of The Skin Layer

The skin layer has a thickness always less than 1 mm (Grassl, 1976) which depends on the vertical heat flux through the air-sea interface. Its temperature is typically a few tenths of a degree Kelvin lower than the temperature measured just a few centimetres below (between 0.1 °K and 0.5 °K) (Schluessel, 1990) as a result of the long-wave radiation emitted from the upper few micrometres of the ocean. The transfer of latent and sensible heat between the ocean and the atmosphere generally produces further cooling of the skin layer. The relative importance of these fluxes depends on the air-sea temperature and water vapor mixing ratio differences. It is also related to the surface wind speed which creates turbulent exchanges of heat and momentum. Only exceptionally the sea skin temperature was found to be higher than the near surface temperature just below. The sharpest temperature gradient within the skin layer normally persists at wind speeds up to 10 m/s, but it is destroyed by breaking waves at wind speeds above 10 m/s. However, studies have shown that it takes only 10 to 12 s for the skin layer to redevelop after the wind has decreased below the 10 m/s limit (Schluessel, 1990). Differences between the surface skin temperature and the bulk temperature measured approximately 1 metre below the surface are found between -1.0 °K to 1.0 °K (Robinson, 1985, Schluessel, 1990).
Correction for the skin effect is done in this thesis by making comparisons between the bulk temperatures measured from buoys with radiance values from AVHRR after the images have been processed. This method is the most common because of its simplicity (Schluessel, 1990). On the other hand, a study by Tabata (1981) off Vancouver Island showed good agreement between AVHRR measurements of the SST and the *in situ* SST measurements from ships.

The following section gives details on the image processing.

### 7.6 Image Processing

The first requirement when using AVHRR satellite data is that images be relatively cloud-free so that ocean dynamic process can be observed. Therefore, the images have been checked at the Institute of Ocean Sciences and/or the University of British Columbia for cloud contamination and only those with a low cloud content (generally less than 20%) have been kept. Then, available images within the events of chapter 6 have been further checked for cloud contamination over the study area by visual inspection of Band 1 images and by using threshold values on enhanced Band 4 images. From 11 images, 3 have been selected for further processing: One image for July 31st (day 92), 1992 during event 2 and two others for July 7th and 8th (day 68, 67), 1993 during event 3. Each image was taken during the afternoon pass so that the SST measurements can be compared.
Further processing was performed on the images by applying a low-pass filter to block the high spatial frequency details. The $3 \times 3$ weighted-filter used is the following:

\[
\begin{bmatrix}
1 & 1 & 1 \\
1 & 4 & 1 \\
1 & 1 & 1
\end{bmatrix}
\]

This type of filter is called an "unequal weighted-filter" and was developed by Wang et al. (1983). While a simple smoothing operation blurs the images, especially at the edges, this type of filter helps to reduce blurring.

Then, the following high-frequency filter developed by Pratt in 1978 was applied to the images to accentuate or sharpen edges:

\[
\begin{bmatrix}
-1 & -1 & -1 \\
-1 & 9 & -1 \\
-1 & -1 & -1
\end{bmatrix}
\]

The image processing has been done by using the NIH Image program for the Macintosh. Further information about processing and/or this program can be found in the user's guide provided by Macintosh.
7.7 AVHRR Observations

The next following three sub-sections (7.7.1, 7.7.2 and 7.7.3) present the descriptions of satellite images. Each sub-section begins with a synoptic situation describing the sea-level pressure pattern over the northeastern Pacific and western North America within approximately 1 hour of the time the images was taken (a). This is followed by a general description of clouds and winds from the synoptic weather map and/or the satellite image (b). Also included are daily averaged in situ SST with anomalies (SSTA) and wind stress from the buoy data for the day the image was taken. Then, the image description follows in "c", before an analysis of transects in "d". These transects, taken from the images, show the spatial SST distribution between the stations. There is also a table in sub-section "d" including the mean, maximum and minimum calibrated SST with the standard deviation for each transect. Unless otherwise mentioned, SST refer to calibrated SST values in this section. The calibration of the radiance values has been made with in situ buoy data.
7.7.1 Image 1

The first image (N11-19845) shown on Figure 43 was taken on day 92 (July 31st) of 1992 at 22:19 Pacific daylight time (PDT).

a) Synoptic Situation:

The sea-level pressure pattern over northeastern Pacific at 23:00 PDT on day 92 (July 31st) was dominated by a ridge of high pressure near 135-140°W (Fig. 44). Further east, there was a thermal trough over the interior of British Columbia and Washington State.

b) Clouds, Winds and *in situ* SST:

The sky was generally sunny, except for a band of cirrus, approximately 55 km wide, extending southeastward from north of Calvert Island (Fig. 43). Also, there was a narrow band (near 10 km wide) of low clouds from southern Calvert Island extending into Queen Charlotte Strait. Northwesterly winds near 10 knots (5 m/s) were reported over much of the study area except for northerly 20 knots (10.3 m/s) at Cape Scott (Fig. 44). The buoy data indicate upwelling-favorable wind stress near 0.07 N/m² in the South and up to 0.17 N/m² in the North, while the averaged daily *in situ* SST were generally below normal (Table 24).
c) Image Description (Fig. 43):

Over the cloud-free area, the radiance distribution shows many variations that reflect a complicated flow pattern over the study area as described in Crawford et al. (1995). In particular, there is evidence of cold upwelled-water (deep blue color) just off Aristazabal Island and further north along the mainland coast, which has also been observed by Jardine et al. (1993). Then, a cold water plume (between 4 to 8 km wide) is seen approximately 20 km east of the buoy 46185 and southward between Middle and Goose Island banks. Further south, the signature of the plume is hidden by the cirrus clouds. The SST is below 12°C along that plume and as low as 10.6°C just off Aristazabal Island. Then, there is a large (near 70 km in diameter) area of warmer water (light blue/white color) over Middle bank with SST up to 15.7°C. This feature has been found to be relatively persistent in time by Jardine et al. (1993) and was called the "Moresby Eddy". To the west of this eddy, the image suggests a flow of cooler water (SST near 14°C) into the Moresby Trough. Then, 10 km southeast of Cape St. James, there is a small cold core (5 km in diameter) with SST near 11.9°C. Another cold area (SST near 12.4°C) appears 15 km southwest of Cape St. James.

Table 24. Daily in situ SST and SSTA in degree Celsius (°C), and wind stress (N/m²) for day 92 (July 31st) of 1992.

<table>
<thead>
<tr>
<th>Stations</th>
<th>Daily SST (°C)</th>
<th>Daily SSTA (°C)</th>
<th>Wind Stress (N/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kains Isld.</td>
<td>11.8</td>
<td>-1.9</td>
<td>----</td>
</tr>
<tr>
<td>46204</td>
<td>13.7</td>
<td>-0.5</td>
<td>-0.07</td>
</tr>
<tr>
<td>46207</td>
<td>14.8</td>
<td>+0.1</td>
<td>-0.07</td>
</tr>
<tr>
<td>46185</td>
<td>14.2</td>
<td>-0.5</td>
<td>-0.17</td>
</tr>
<tr>
<td>46208</td>
<td>13.7</td>
<td>-0.1</td>
<td>-0.17</td>
</tr>
</tbody>
</table>

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Close to the mainland coast and south of Aristazabal Island, there is no evidence of cold upwelled-water and temperatures reach near 15.5°C just off Calvert Island. Warmer waters appear just north of Vancouver Island with values up to 16.6°C near Cook bank and east of it. An area of cold water (deep blue and purple) is evident off Cape Scott with SST as low as 10.7°C. This feature was found to be the result of strong tidal mixing (Jardine et al. 1993). Finally, a large warm eddy (near 45 km in diameter) is shown 25 km west of Kains Island with SST up to 16.3°C.

d) SST Distribution:

1. Transect #1: Buoys 46185 - 46204 (Figure 45a)
(The transect extends from 10 km north of the buoy 46185 to 6 km south of the buoy 46204.)

The SST is near 13.5°C at the initial point (0 km) where the satellite image suggests upwelling. It increases to 14.1°C near the site of the buoy 46185 and to near 15.7°C at the northeastern edge of the Moresby eddy. Then, the SST gradually decreases to just below 12°C in the cold plume before increasing again. Further south, pixels with radiance values above 155 (i.e., SST < 11°C) have been ignored because of cloud contamination. The satellite image suggests that cold upwelled-waters appear mainly north and east of the buoy 46185 and that the SST at the site of the buoy is more representative of those over the Moresby eddy. Further south, warmer SST is evident near the site of the buoy 46204 suggesting stratified conditions. The SST at the site of the buoy 46204 is near 15.2°C.
2. Transect #2: From the buoy 46207 to 46204 (Fig. 45b)
(The transect extends from the site of the buoy 46207 to 7 km past the buoy 46204.)

The SST distribution is relatively uniform between the two buoy sites. Moreover, the SST is the same at the sites of the two buoys (15.2°C). However, there is a band (18 km wide) with higher SST 20 km from the site of the buoy 46207. Also, there is a band (25 km wide) of cooler water approximately 20 km from the site of the buoy 46204 that looks like tidally induced upwelling. This phenomena has been observed in previous studies (Jardine et al, 1993).

3. Transect #3: From Kains Island to the buoy 46207 (Figure 45c)
(The transect begins 2.5 km south of Kains Island and ends at the site of the buoy 46207.)

The SST distribution within the first 105 km from Kains Island reflects cold waters from coastal upwelling and tidal mixing with temperature near 13.9°C just off Kains Island in Quatsino Sound and as low as 11.2°C 76 km from there. However, there is a band of warmer waters between distances of 33 km and 40 km from Kains Island which may be associated with a frontal area between the coastal upwelling and tidal mixing area. On the other hand, it may reflect the water temperatures of a warm eddy as mentioned in c). Further offshore (beyond 100 km from Kains Island), the SST reflect the water temperatures around the site of the buoy 46207 where the mixing seems weaker. Values increase to near 15.2°C at the site of that buoy, which is more than 3°C warmer than the waters just off Kains Island.

<table>
<thead>
<tr>
<th>Transects</th>
<th>Mean SST(°C)</th>
<th>Max. SST(°C)</th>
<th>Min. (°C)</th>
<th>STD (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>46185-46204</td>
<td>----</td>
<td>15.8</td>
<td>----</td>
<td>----</td>
</tr>
<tr>
<td>46207-46204</td>
<td>15.2</td>
<td>15.9</td>
<td>14.4</td>
<td>0.4</td>
</tr>
<tr>
<td>Kains I.-46207</td>
<td>13.8</td>
<td>15.7</td>
<td>11.2</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Table 25. Standard deviation and mean, maximum and minimum SST (°C) along the transects.
7.7.2 Image 2

This image (N11-24662) shown on Figure 46 was taken on day 68 (July 7th) of 1993 at 23:57.

a) Synoptic Situation:

On day 68 (July 7th) of 1993 at 01:00 PDT, the pressure pattern over the northeastern Pacific was dominated by a strong high pressure system centered near 50°N/146°W (Fig. 47). Further east, there was a thermal low over southeastern British Columbia.

b) Clouds, Winds and in situ SST:

The sky was generally sunny, except for traces of thin clouds far off Cape Scott and a small area of low cloud half way between Cape St. James and the mainland coast (Fig. 46). The pressure gradient was relatively strong over Queen Charlotte Sound and near the northern end of Vancouver Island with northwesterly winds up to 25 knots (12.8 m/s) corresponding to upwelling-favorable wind stress up to 0.25 N/m² (Fig. 47). The daily in situ SST were below normal at all stations (Table 26).

<table>
<thead>
<tr>
<th>Stations</th>
<th>Daily SST (°C)</th>
<th>Daily SSTA (°C)</th>
<th>Wind Stress (N/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kains Isld.</td>
<td>11.4</td>
<td>-2.2</td>
<td>----</td>
</tr>
<tr>
<td>46204</td>
<td>13.6</td>
<td>-0.7</td>
<td>-0.15</td>
</tr>
<tr>
<td>46207</td>
<td>13.1</td>
<td>-0.9</td>
<td>-0.14</td>
</tr>
<tr>
<td>46185</td>
<td>11.0</td>
<td>-2.9</td>
<td>-0.20</td>
</tr>
<tr>
<td>46208</td>
<td>----</td>
<td>----</td>
<td>-0.23</td>
</tr>
</tbody>
</table>

Table 26. Daily in situ SST and SSTA in degree Celsius (°C), and wind stress (N/m²) for day 68 (July 7th) of 1993.
c) Image Description (Fig. 46):

The radiance distribution is relatively uniform over much of Queen Charlotte Sound with a mean SST near 13-14°C. However, there are significant variations from offshore to coastal areas. In particular, there is clear evidence of upwelling (deep blue and purple colors) along much of the mainland coast, especially near Aristazabal Island and northward where coldest waters are found. A cold water plume extends offshore from Aristazabal Island as far as the site of the buoy 46185 and 60 km to the South. The SST near the site of that buoy is near 12°C but decreases to near 10.6°C within 15 km to the Northeast. The warm Moresby eddy shows again over Middle and Goose Island banks, but covers a larger area (up to 100 km wide). The highest SST (near 14.2°C) appears over Goose Island bank, approximately 60 km off the northern end of Aristazabal Island. Further west, there is a sharp temperature gradient (1.7°C within 3.5 km) near the 1000 m depth contour. Also, a large cold eddy (50 km in diameter) is evident approximately 55 km south of Cape St. James with SST near 11°C.

Over the southeastern area, cold upwelled waters extend from the coast to approximately 20 km offshore and southward. There is a cold plume (SST near 11°C) extending from Calvert Island to approximately 35 km east of the buoy 46204. Cold water is also evident off Cape Scott and Kains Island (purple and deep blue) where the SST is near 2-3°C cooler than the surrounding waters. Cold waters extend within approximately 35 km off Kains Island due to upwelling and up to 80 km off Cape Scott due to tidal mixing (Jardine et al, 1993). Further offshore, a large area of warmer waters (near 14°C) is evident beyond 50 km off Kains Island where a small eddy (10 km in diameter) can be seen.
d) SST Distribution:

1. Transect #1: From the buoy 46185 to 46204 (Fig. 48a)

The SST is near 12°C at the initial point (10 km north of the buoy 46185) and near 12.3°C at the buoy 46185. The satellite image shows clearly cold-upwelled waters reaching the site of that buoy. Then, the SST decreases to a minimum of 11.2 °C towards the centre of the plume (approx. 10 km southeast of the buoy 46185). Further south, the SST increases gradually to reach a maximum value (14.0°C) over Goose Island bank (approx. 120 km southeast of the buoy 46185) and then decreases slightly towards the site of the buoy 46204. From the satellite image, it seems that cold upwelled water near the mainland coast flows southward into the Goose Island trough and towards the site of the buoy 46204. However, the SST at the site of the buoy 46204 is warmer (13.5°C), suggesting less or no influence from upwelling.

2. Transect #2: From the buoy 46207 to 46204 (Fig. 48b)

As for Image 1, the SST distribution is relatively uniform between the two sites of these buoys suggesting similar dynamics and/or heating process along the transect. However, the SST is slightly lower near the site of the buoy 46204 which can be attributed to an intrusion of cold upwelled waters from the mainland coast. The SST is 13.7°C at the site of the buoy 46207 and 13.4°C at the site of the buoy 46204.

3. Transect #3: From Kains Island to the buoy 46207 (Fig. 48c)
Relatively cold upwelled-waters are indicated within the first 76 km from Kains Island. The coldest waters (near 10°C) are found within approximately 15 km of the coast. Then, the SST increases to near 12°C. Beyond 76 km, the SST increased by 1.7°C. An area of cold waters (near 12.8°C) appears between a distance of 78 km to 125 km due to tidal mixing. Then, the SST rises to reach 13.7°C near the site of the buoy 46207 where water conditions are more stratified.

<table>
<thead>
<tr>
<th>Transects</th>
<th>Mean SST(°C)</th>
<th>Max. SST(°C)</th>
<th>Min. (°C)</th>
<th>STD (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>46185-46204</td>
<td>12.9</td>
<td>14.0</td>
<td>11.2</td>
<td>0.8</td>
</tr>
<tr>
<td>46207-46204</td>
<td>13.5</td>
<td>13.9</td>
<td>13.0</td>
<td>0.2</td>
</tr>
<tr>
<td>Kains I.-46207</td>
<td>12.5</td>
<td>13.8</td>
<td>10.3</td>
<td>0.7</td>
</tr>
</tbody>
</table>

Table 27. Standard deviation and mean, maximum and minimum SST (°C) along the transects for day 68 (July 7th) of 1993 at 23:57 PDT.
7.7.3 Image 3

This image (N11-24676) shown on Figure 49 was taken on day 69 (July 8th) of 1993 at 23:45.

a) Synoptic Situation:

On day 69 (July 8th) of 1993 at 01:00 PDT, the high pressure centre had moved further to the northwest from the previous day and was located near 54°N/149°W (Fig. 50). Further east, there was a cold front from the just east of the Queen Charlotte Islands extending to southeastern British Columbia. Also, there was a weak lee trough just along the west coast of Vancouver Island.

b) Clouds, Winds and SST:

The sky was generally sunny over the study area, except for a few traces of cirrus clouds (Fig. 49). Also, there was an area of low clouds around the northern end of Vancouver Island. The winds were generally from the northwest with speeds near 20 knots (10.3 m/s) (Fig. 50). The daily in situ SST were still below normal at most stations.
<table>
<thead>
<tr>
<th>Stations</th>
<th>Daily SST (°C)</th>
<th>Daily SSTA (°C)</th>
<th>Wind Stress (N/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kains Isld.</td>
<td>11.1</td>
<td>-2.5</td>
<td>----</td>
</tr>
<tr>
<td>46204</td>
<td>12.9</td>
<td>-1.5</td>
<td>-0.09</td>
</tr>
<tr>
<td>46207</td>
<td>13.5</td>
<td>-0.6</td>
<td>-0.15</td>
</tr>
<tr>
<td>46185</td>
<td>11.2</td>
<td>-2.8</td>
<td>-0.14</td>
</tr>
<tr>
<td>46208</td>
<td>----</td>
<td>----</td>
<td>-0.19</td>
</tr>
</tbody>
</table>

Table 28. Daily in situ SST and SSTA in degree Celsius (°C), and wind stress (N/m²) for day 69 (July 8th) of 1993.

c) Image Description (Fig. 49):

The radiance distribution is still relatively uniform over Middle and Goose Island banks with SST comparable to those of the previous day (13-14°C). The cold upwelled-water reaching the site of the buoy 46185 has not progressed much further offshore, while its temperature has remained relatively constant off Aristazabal Island and northward. However, the cold water plume is now reaching approximately 10 km further south than the previous day. The core of that plume flows 20 km east of the buoy 46185 with SST values between 10.4°C off Aristazabal Island and 12.4°C further south. The highest SST (near 14.5°C) still appears over Goose Island bank. Further west, the sharp temperature gradient near the 1000 m depth and the cold eddy south of Cape St. James do not show significant change from the previous day.

Over the eastern shore of Queen Charlotte Sound, the satellite image shows a larger upwelling area than the previous day with a few filaments of cold waters extending southward from the mainland coast. Near Calvert Island, the upwelling area has expanded by approximately 5 km further west with the coldest waters (11°C) just north of the Island. Also, the cold upwelled-water is reaching much further south...
with its intrusion near the buoy 46204 evident. The SST is 12°C 9.5 km northeast of that buoy and 13.5°C at the site. Further south, there is an area of low clouds around the northern end of Vancouver Island hiding the SST distribution just off Kains Island. The cold waters resulting from tidal mixing are still visible off Cape Scott and are now spreading towards the buoy 46207. In fact, the SST has decreased by 1°C at the buoy from the previous day. Further south and approximately 50-60 km off Kains Island, an area of warm water is still evident, but the SST has decreased by near 1°C from the previous day.

d) SST Distribution:

1. Transect #1: Buoy 46185-46204 (Fig. 51a)

The SST is just below 12°C at the initial point (10 km north of the buoy 46185) and near 12.2°C at the buoy 46185, which is similar to the previous day. Further south, the SST is relatively low (between 11.5°C and 12.4°C) in the cold plume. Noteworthy is a sharp increase 70 km south of the buoy 46185 which is the southern limit of that plume. Then, the SST oscillate around 13.8°C over a distance of approximately 60 km. The warmest SST (14.4°C) is reached at a distance of 115 km from the buoy 46185. Further south, the SST decreases again towards the buoy 46204 due to cold upwelled-waters reaching the area.

2. Transect #2: Buoy 46207-46204 (Fig. 51b)

Once again, the SST distribution is relatively uniform between the two buoys. However, the SST northeast of the buoy 46204 are decreasing due to the intrusion of
cold upwelled-waters towards that buoy as mentioned in c). Lower SST are also seen at the buoy 46207 due to tidally mixed waters spreading towards that buoy.

3. Transect #3: Kains Island-Buoy 46207 (Fig. 51c)

The SST have been ignored for the first 45 km from Kains Island to eliminate any cloud contamination. The main feature is what appears as a front at a distance of 66 km from Kains Island. Noteworthy is that front now being 10 km closer to Kains Island compared with the previous day. An area of relatively cooler waters appears between distances of 88 km and 116 km from Kains Island, which is associated with the core of mixing area due to the action of the tide. Then, the SST slowly increases towards the buoy 46207.

<table>
<thead>
<tr>
<th>Transects</th>
<th>Mean SST(°C)</th>
<th>Max. SST(°C)</th>
<th>Min. (°C)</th>
<th>STD (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>46185-46204</td>
<td>12.5</td>
<td>13.9</td>
<td>11.2</td>
<td>0.8</td>
</tr>
<tr>
<td>46207-46204</td>
<td>13.2</td>
<td>13.6</td>
<td>12.5</td>
<td>0.2</td>
</tr>
<tr>
<td>Kains I.-46207</td>
<td>12.5</td>
<td>13.6</td>
<td>11.3</td>
<td>0.5</td>
</tr>
</tbody>
</table>

Table 29. Standard deviation and mean, maximum and minimum SST (°C) along the transects for day 69 (July 8th) of 1993 at 23:45 PDT.
7.7.4 General Observations In Upwelling Conditions

The satellite images show clearly that with upwelling-favorable winds the SST at Kains Island does not reflect those over much of Queen Charlotte Sound. Upwelling is seen near the northern end of Vancouver Island with SST near 2 °C colder than that 20 km to 30 km offshore. Further west, the transects between Kains Island and the buoy 46207 show a consistent temperature change at distances of 65 km to 75 km from Kains Island which is the limit of the mixing area. Then, beyond approximately 80 km off Kains Island, the SST increases and is generally a few degrees warmer at the site of the buoy 46207. The latter represents the conditions over a large part of Queen Charlotte Sound. North of Vancouver Island, the SST is relatively constant between the buoys 46207 and 46204, however, it generally decreases past the buoy 46204. There are a few cold water plumes extending within approximately 20 km from the mainland coast, which may occasionally reach as far as the site of the buoy 46204. Further north, the SST are relatively uniform beyond the upwelling area, with maximum SST generally reached over Goose Island bank, 30 km to 40 km north of the buoy 46204. However, the upwelling covers a much wider area with cold water plumes spreading up to 70 km off the mainland coast. In particular, upwelling seems a dominant feature near the site of the buoy 46185 where the SST is similar to that near Kains Island, but can be more than 1 °C cooler than that near the buoy 46204. Finally, the information is limited near the site of the buoy 46208 due to cloud cover. However, upwelling is seen within approximately 25 km off the west coast of the Queen Charlotte Islands, but does not seem to have much influence on the water properties near the site of the buoy 46208.
Figure 38. Emission spectra at three different temperatures. From Robinson (1985).

Figure 39. Energy-wavelength spectrum of solar radiation at the surface of the ocean and at different depths (nm=nanometre=10^{-9}m). (From Brown et al, 1989).
Figure 40. Typical mean temperature profile for mid-latitudes in the open oceans.

Figure 41. Temperature profiles showing successively the growth (solid lines) and decay (broken lines) of a seasonal thermocline in the Northern hemisphere.
Figure 42. Typical near-surface temperature profiles showing the diurnal thermocline during calm, slightly-mixed and night-time conditions. (From Robinson, 1985).
Figure 43. AVHRR satellite image for day 92 (July 31st) 1992, 22:19 PDT.
Figure 44. Synoptic weather maps for July 31st, 1992 at 01:00 am PDT. From the Pacific Weather Centre, Environment Canada.
Figure 45. Satellite sensed sea surface temperature (°C) distribution for day 92 (July 31st), 1992 at 22:19 PDT from a) 10 km north of the buoy 46185 to 6 km south of the buoy 46204, b) the buoy 46207 to 7 km east-northeast the buoy 46204 and c) Kains Island to the buoy 46207. The star signs marked the location of the sites along the transects. Distances are in kilometre (km).
Figure 46. AVHRR satellite image for day 68 (July 7th) 1993, 23:57 PDT.
Figure 48. Satellite sensed sea surface temperature (°C) distribution for day 68 (July 7th), 1993 at 23:57 PDT from a) 10 km north of the buoy 46185 to 6 km south of the buoy 46204, b) the buoy 46207 to 7 km east-northeast the buoy 46204 and c) 2.5 km south of Kains Island to the buoy 46207. The star signs marked the location of the sites along the transects. Distances are in kilometre (km).
Figure 51. Satellite sensed sea surface temperature (°C) distribution for day 69 (July 8th), 1993 at 23:45 PDT from a) 10 km north of the buoy 46185 to 6 km south of the buoy 46204, b) the buoy 46207 to 7 km east-northeast the buoy 46204 and c) 2.5 km south of Kains Island to the buoy 46207. The star signs marked the location of the sites along the transects. Distances are in kilometre (km).
Chapter 8

Conclusion

8.1 Summary and Conclusions

Summer daily SST over Queen Charlotte Sound and near the northern end of Vancouver Island have been studied using buoy and lighthouse data. From a visual inspection, the most evident cycle of SST anomalies time series occurs at 2 to 3 days due to the influence of weather systems. The statistical analysis does not show much similarity between the stations, especially between coastal and offshore stations. The correlation analysis indicates that most r-values are not significantly different from zero. However, it shows a significant correlation between the buoys 46204 and 46207 in 1992 and 1993. On the other hand, the wind time series of the buoy stations are all significantly correlated with r-values above 0.60. Finally, the spectral analysis shows generally more energy at low frequencies. In general, the statistical analysis has not been very successful because the time series present many uncorrelated fluctuations due to local effects.

In chapter 6, an event-by-event analysis is done mainly by visual inspection: by looking at SST, windstress and cloud cover data along with synoptic weather maps. This analysis shows that the main cause of SST variations during the summer is upwelling associated with a high pressure centre offshore, a ridge extending in the vicinity of the Queen Charlotte Islands and/or further north and a lee trough along the west coast of Vancouver Island. The Northwesterlies and upwelling response can be enhanced over Queen Charlotte Sound with the approach of a low pressure system from the west. In other situations, upwelling can be triggered behind a cold
front. However, relatively strong Northwesterlies do not persist over a long period of time and so neither does the upwelling. The induced vertical velocity is near 4 m/day to 5 m/day during the peak of the events. However, the response is stronger near the coast, i.e. at Kains Island and the buoy 46185 and decreases at the offshore buoys as predicted by the theory. A second cause of SST variations is the changes in the incoming heat flux due to fluctuations in the cloud cover. The influence of clouds can be observed mainly when the upwelling-favorable winds are relatively weak or non-existent. However, the lack of consistency between cloud cover and SST time series does not permit conclusions with certainty. The variations in latent and sensible heat fluxes have been neglected in the estimated heat flux compared with those involved in the incoming and back radiation. However, it appears that the daily variations of the latent heat flux from the evaporation process could help to further explain the SST variations, especially when upwelling is not dominant. Finally, a cooling effect due to vertical mixing could be occasionally detected with strong south to southeasterly winds which are not frequent during the summer time.

In general, the SST is lower at Kains Island compared with the offshore buoys due to a stronger upwelling response. A visual inspection of the SST time series for specific events shows that the SST variations among the stations agree relatively well, especially between the buoys 46204 and 46207. On the other hand, the SST at Kains Island shows more short term variations, while that at the buoy 46185 manifests episodic strong cooling. The AVHRR satellite images confirm the difference between coastal and offshore SST during upwelling-favorable situations in which case the SST at Kains Island does not represent that over much of Queen Charlotte Sound. In fact, the most representative station of that area is the buoy 46207 which does not appear to be influenced much by upwelling. Then, it would be
interesting to compare the fish migration route and the northern diversion with the SST at that buoy. However, the fish may respond to factors other than SST!

8.2 Future Work

In order to further investigate the SST variations, a more complete heat budget should be completed by looking at the air temperature. This would allow us to consider the evaporation term which may be relatively important during upwelling events. Also, a more accurate heat budget may explain the SST variations when upwelling-favorable winds are weak or non-existent. On the other hand, atmospheric pressure data from the buoys could be used to make an index related to the northern diversion.


Crawford, W.R. Physical Oceanography of the Waters around the Queen Charlotte Islands. Submitted for publication in Ecology of marine and shoreline birds of the Queen Charlotte Islands, British Columbia.


Large, W.G., Crawford, G.B. Observations and simulations of upper ocean response to wind events during the Ocean Storms Experiment. Accepted by JPO for special Ocean Storms issue. 1994.


Appendix A

Decoding of Synoptic Report

The model used on weather synoptic maps is the following:

\[
\begin{array}{c}
\text{C}_H \\
\text{CM} \\
T_T \\
\text{VV} \\
\text{ww} \\
\text{ppp} \\
\text{ppa} \\
\text{wR}_t \\
\text{RR} \\
\text{CL/NH} \\
\text{h} \\
\end{array}
\]

where

- ppp is the atmospheric pressure,
- ppa the pressure tendency within the last 3 hours,
- wR_t is the past weather conditions,
- RR is the precipitation in tenth of millimetre,
- CL, NH and h refer to the type, coverage and height of low clouds,
- T_d T_d is the temperature of the dew point,
- VV and ww are the visibility and atmospheric conditions,
- TT is the air temperature measured 10 m above the ground and
- N is the cumulated amount of cloud (in oktas).

For ship and/or buoy reports, the SST is indicated within brackets and below the temperature of the dew point T_d.
The wind is added to the reports with the following convention:

- 1 to 2 knots
- 5 +/- 2 knots
- 10 +/- 2 knots
- 15 +/- 2 knots
- 20 +/- 2 knots
- 25 +/- 2 knots
- 30 +/- 2 knots
- 35 +/- 2 knots
- 40 +/- 2 knots
- 45 +/- 2 knots

The direction is given by the orientation of the arrow and indicates where the wind is coming from.

For the purpose of this thesis, only the wind, the cumulated amount of cloud N and the SST are important. The amount of cloud N is represented as follows:

- No cloud
- One okta or less
Two oktas

Three oktas

Four oktas

Five oktas

Six oktas

Seven oktas

Completely overcast

Sky completely obscured by fog, haze, smoke

...
Figure B1. Example of synoptic weather map over Northeast Pacific and Western Canada. Queen Charlotte Islands, Queen Charlotte Sound and Vancouver Island are indicated by QC1, QC2, and VRI respectively. The synoptic code is given in Appendix A. The number in the upper left corner indicates the day number with May 1st as a reference for day 1.
Appendix C

Synoptic Weather Maps For Event 1

July 9 to August 8 (day 70 to day 100), 1990
Figure C1. Synoptic weather map for July 9, 1990 at 5 am PDT. From the Pacific Weather Centre (PWC), Environment Canada.
Figure C2. Synoptic weather map for July 10, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C3. Synoptic weather map for July 11, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C5. Synoptic weather map for July 13, 1990 at 5 am PDT. From the PW/C, Environment Canada.
Figure C7. Synoptic weather map for July 15, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C9. Synoptic weather map for July 17, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C10. Synoptic weather map for July 18, 1990 at 5 am PDT. From the PWIC, Environment Canada.
Figure C11. Synoptic weather map for July 19, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C12. Synoptic weather map for July 20, 1990 at 5 am PDT. From the PWG, Environment Canada.
Figure C14. Synoptic weather map for July 22, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C1.6. Synoptic weather map for July 24, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C18. Synoptic weather map for July 26, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C19. Synoptic weather map for July 27, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C21. Synoptic weather map for July 29, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C23. Synoptic weather map for July 31, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C.6. Synoptic weather map for August 3, 1990 at 5 am PDT. From the PWU, Environment Canada.
Figure C27. Synoptic weather map for August 4, 1990 at 5 am PDT. From the PWG Environment Canada.
Figure C30. Synoptic weather map for August 7, 1990 at 5 am PDT. From the PWC, Environment Canada.
Figure C.1. Synoptic weather map for August 8, 1990 at 5 am PDT. From the PWG, Environment Canada.
Appendix D

Synoptic Weather Maps For Event 2

July 3 to August 4 (day 64 to day 96), 1992
Appendix D

Synoptic Weather Maps For Event 2

July 3 to August 4 (day 64 to day 96), 1992
Figure D1. Synoptic weather map for July 3, 1992 at 5 am PDT. From the Pacific Weather Centre (PWC), Environment Canada.
Figure D2. Synoptic weather map for July 4, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D3. Synoptic weather map for July 5, 1992 at 5 am PDT. From the PWG Environment Canada.
Figure D4. Synoptic weather map for July 6, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D5: Synoptic weather map for July 7, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D7: Synoptic weather map for July 9, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D8. Synoptic weather map for July 10, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D9. Synoptic weather map for July 11, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D11. Synoptic weather map for July 13, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D12. Synoptic weather map for July 14, 1992 at 5 am PDT. From the PW/C, Environment Canada.
Figure D14. Synoptic weather map for July 16, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D15. Synoptic weather map for July 17, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D18. Synoptic weather map for July 20, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D19. Synoptic weather map for July 21, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D21. Synoptic weather map for July 23, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D22. Synoptic weather map for July 24, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D24. Synoptic weather map for July 26, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D.5. Synoptic weather map for July 27, 1992 at 5 am PDT. From the FWC, Environment Canada.
Figure D27. Synoptic weather map for July 29, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D29. Synoptic weather map for July 31, 1992 at 5 am PDT. From the PWC, Environment Canada.
Figure D31. Synoptic weather map for August 2, 1992 at 5 am PDT. From the PWC, Environment Canada.
Appendix E

Synoptic Weather Maps For Event 3

June 29 to July 29 (day 60 to day 90), 1993
Figure E1. Synoptic weather map for June 29, 1993 at 5 am PDT. From the Pacific Weather Centre (PWC), Environment Canada.
Figure E2. Synoptic weather map for June 30, 1993 at 5 am PDT. From the PWG, Environment Canada.
Figure E3: Synoptic weather map for July 1, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E4. Synoptic weather map for July 2, 1988 at 5 am PDT. From the FWC, Environment Canada.
Figure E5. Synoptic weather map for July 3, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E7: Synoptic weather map for July 5, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E8. Synoptic weather map for July 6, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure B9. Synoptic weather map for July 7, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E11. Synoptic weather map for July 9, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E12. Synoptic weather map for July 10, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E14. Synoptic weather map for July 12, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E16. Synoptic weather map for July 14, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E1.8. Synoptic weather map for July 16, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E20. Synoptic weather map for July 18, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E21. Synoptic weather map for July 19, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure 22: Synoptic weather map for July 20, 1993 at 5 am PDT. From the FWC, Environment Canada.
Figure E23. Synoptic weather map for July 21, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E25. Synoptic weather map for July 23, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E26. Synoptic weather map for July 24, 1993 at 5 am PDT. From the PWC, Environment Canada.
Figure E28. Synoptic weather map for July 26, 1993 at 5 am PDT. From the PWIC, Environment Canada.
Figure B30. Synoptic weather map for July 28, 1993 at 5 am PDT. From the PVC Environment Canada.
Figure E31. Synoptic weather map for July 29, 1993 at 5 am PDT. From the PWC, Environment Canada.