LOW FREQUENCY RESIDUAL CIRCULATION IN KNIGHT INLET A Fjord of Coastal British Columbia

by

Peter Donald Baker

Diploma in Electrical Engineering Technology, Ryerson Polytechnical Institute, 1975

A THESIS SUBMITTED IN PARTIAL FULFILMENT OF

THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

in

THE FACULTY OF GRADUATE STUDIES

Department of Oceanography

We accept this thesis as conforming

to the required standard

THE UNIVERSITY OF BRITISH COLUMBIA

April, 1992

C Peter Baker, 1992

In presenting this thesis in partial fulfilment of the requirements for an advanced degree at the University of British Columbia, I agree that the Library shall make it freely available for reference and study. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by the head of my department or by his or her representatives. It is understood that copying or publication of this thesis for financial gain shall not be allowed without my written permission.

(Signature)

Department of OCEANOGRAPHY.

The University of British Columbia Vancouver, Canada

Date 29 April 52

Abstract

Many aspects of the low frequency response of a stratified inlet have not been previously observed due to the lack of simultaneous observations of wind, currents and density structure over the entire water column. This thesis describes an experiment that was specifically designed to obtain such observations and the statistical analysis of the sub-tidal response in Knight Inlet, a stratified, high runoff inlet, during the onset of the freshet. Month long observations of currents, temperature and salinity throughout the water column, both outside and inside the sill, were made during the spring of 1988 and the summer of 1989. In addition, simultaneous measurements were made of inlet winds and river runoff data were obtained, giving a complete data set for the analysis of the forcing and the response of the inlet during these time periods.

Diurnal and semi-diurnal tidal energy was removed through harmonic analysis and the dominant residual response was found to be due to the wind, with a coherence of greater than 0.8 in the near surface and some contributions at depth. A transfer function was then derived from the cross spectrum and used to estimate the wind driven currents and density fluctuations throughout the water column. The data records were then dewinded by subtracting the wind response from the detided records and the remaining residual analyzed with respect to the vertically nested thermohaline circulations driven by the surface estuarine process and deep water renewal. The two circulation cells were found to be coupled through conservation of volume with their characteristics dependent on the availability of source water for deep water renewal , the river discharge, and wind. The dynamics of the surface layer was found to be consistent with the work of van der Baaren(1988) who showed that the along inlet balance of forces was between the surface pressure gradient and the interfacial friction.

ü

Acknowledgements	
1. Introduction	
1.1 Coastal Inlets of British Columbia	
1.2 Knight Inlet	
1.3 Thesis Objectives	
2. Low Frequency Residual Dynamics of Fjords	
2.1 Definition of the Low Frequency Residual	
2.2 Energy Sources Driving Inlet Circulation	
2.2.1 Estuarine Flow	
2.2.2 Density Flows	
2.2.3 Wind Driven Flow	
2.2.4 Ital Forcing	
2.3 Previous Observations of the Low Frequency Residual of Inlets	
3. Experiment Descriptions	•••••
3.1.1 Low Frequency Residual Circulation	
3.1.2 Verification of Inlet General Circulation Models	
3.1.3 Modal Response and Dissipation of the Internal Tide	
3.2 Instrumentation	
3.3 Kn88 - Knight Inlet, Spring 1988	
3.4 Kn89 - Knight Inlet, Summer 1989	
3.5 Data Processing	
3.5.1 Instrument Calibration	
3.5.2 C1D Bin Averaging 2.5.2 Cyclesondo Outlier Editing	
3.5.3 Cyclesonde Time Series Interpolation	
3.54 Current Meter and Anemometer Median Averaging	
3.5.5 Inter-calibration	
3.6 Processed Time Series and Spectra	
4. Analysis and Discussion	
4.1 Data Analysis Methodology	
4.2 29.5 Day Mean Response	
4.3 Harmonic Analysis and Computation of the Detided Time Series	
4.4 Detided Inlet Response	
4.5 Cross Spectra of Wind versus Along Channel Current and Density	
4.0 wind Driven Response 4.7 Detided and Dewinded Response	
n / Double und Dominico response	

Table Of Contents

List of Tables

1.1	Charcteristics of some B.C. Fjords	3
3.1	Kn88/Kn89 Instrument Sampling Strategies	41
3.2	Basin Salinities at Station 7 for the Kn88 Experiment	48
4.1	Volume Transport (m^3/s) per metre in the Along Channel Direction for 1988	86
4.2	Volume Transport (m ³ /s) per metre in the Along Channel Direction for 1989	90
4.3	Estimated Surface Layer Depth, Layer Densities (as σ_t), and Density Differences ($\Delta \sigma_t$)	95
4.4	Estimated surface Layer Velocities using Knudsen's Relations	96
4.5	Tidal Constituents used in Harmonic Analysis	106
4.6	Surface and Peak Lag Correlations at depth for Wind and Along Channel Currents	113

List of Figures

1.1 1.2	The Coast of British Columbia Knight Inlet with Oceanographic Stations	2 4
2.1	Energy Paths to the Low Frequency Residual	9
2.2	Wind Response in Alberni Inlet, Surface Layer Thickness and 2m Along Channel Currents	18
2.3	Tidal Kinetic Energy calculated from Depth Mean Velocity Squared In Knight Inlet	22
2.4	Net Current Profiles over the First and Last 25 Hours just Outside the Knight Inlet Sill	28
2.5	Residual Circulation Profiles from Knight Inlet, July and September 1983	30
2.6	Layer Velocities for Knight Inlet; 1986 and 1987	34
3.1	Knight Inlet Profile showing Instrument Moorings	43
3.2	Kn88 River Runoff and Along Channel Wind for Protection and Tomakstum	45
3.3	Kn88 Deployment and Mid-Experiment Cruise CTD Surveys (contours of σ_t)	47
3.4	Kn89 River Runoff and Wind for Protection (Reconstructed) and Tomakstum	50
3.5	Kn89 Deployment and Pickup Cruise CTD Surveys (contours of σ_t)	52
3.6	Instrument Preliminary Processing Data Flow	63
3.7	Kn88 Protection Processed Along Channel Velocity Time Series	65
3.8	Kn88 Tomakstum Processed Along Channel Velocity Time Series	66
3.9	Kn88 Protection Processed Density (as σ_t) Time Series	67
3.10	Kn88 Tomakstum Processed Density (as σ_t) Time Series	68
3.11	Kn88 Tomakstum Raw Spectra	70
3.12	Kn88 Protection Raw Spectra	71
3.13	Kn89 Protection Processed Along Channel Velocity Time Series	73
3.14	Kn89 Tomakstum Processed Along Channel Velocity Time Series	/4
3.15	Kn89 Protection Processed Density (as σ_t) Time Series	75
3.16	Kn89 Tomakstum Processed Density (as σ_t) Time Series	76
3.17	Kn89 Tomakstum Raw Spectra	78
3.18	Kn89 Protection Raw Spectra	79
4.1	Residual Analysis Data Flow	83
4.2	1988 29.5 Day Along Channel Average Velocity (U) and Density (as σ_t) Profiles	85
4.3	1989 29.5 Day Along Channel Average Velocity (U) and Density (as σ_t) Profiles	88
4.4	A Three Dimensional Plot of Knight Inlet Sill	91
4.5	1988 vs 1989 Near Surface Along Channel (U) 29.5 Day Mean Velocity Profiles	93
4.6	1988 vs 1989 Near Surface Density (as σ_t) Profiles	94
4.7	Surface Slope(m/m) vs River Discharge for Knight Inlet	99
4.8	Isobaric Slope Profiles for the 1989 Experiment	101
4.9	1988 Along Channel Wind vs Selected Detided Currents	109
4.10	1989 Along Channel Wind vs Selected Detided Currents	111
4.11	1988 and 1989 Wind Correlations with Depth	112
4.12	1988 Along Channel Wind vs Selected Detided Density (as σ_t)	116
4.13	1989 Along Channel Wind vs Selected Detided Density (as σ_t)	117
4.14	1988 Protection: Power and Cross Sprectra of Wind vs Along Channel Currents	122
4.15	1988 Tomakstum: Power and Cross Sprectra of Wind vs Along Channel Currents	123
4.16	1989 Protection: Power and Cross Sprectra of Wind vs Along Channel Currents	125
4.17	1989 Tomakstum: Power and Cross Sprectra of Wind vs Along Channel Currents	126
4.18	1988 Protection: Power and Cross Spreetra of Wind vs Densities	12/
4.19	1988 Tomakstum. Power and Cross Spreetra of Wind vs Densities	120
4.20	1989 Tomaketum: Power and Cross Spreatra of Wind vs Densities	130
4.22	1988 Along Channel Wind vs Selected Wind Driven Currents	133
4.23	1989 Along Channel Wind vs Selected Wind Driven Currents	134
4.24	1988 Along Channel Wind vs Selected Wind Driven Density Fluctuations (as gr)	136
4.25	1989 Along Channel Wind vs Selected Wind Driven Density Fluctuations (as of)	137
4.45	1707 Hour Chamber while is believed while Diriven Density Hactuations (ds 0[)	1.57

List of Figures(cont'd)

4.26	Wind Driven Variance as a Percentage of Total Detided Residual Variance with Depth	138
4.27	1988 Protection Wind and Wind Driven Along Channel Velocity Contours	140
4.28	1988 Protection Wind and Wind Driven Density Fluctuation Contours	141
4.29	1988 Tomakstum Wind and Wind Driven Along Channel Velocity Contours	143
4.30	1988 Tomakstum Wind and Wind Driven Density Fluctuation Contours	144
4.31	1989 Protection Wind and Wind Driven Along Channel Velocity Contours	146
4.32	1989 Protection Wind and Wind Driven Density Fluctuation Contours	147
4.33	1989 Tomakstum Wind and Wind Driven Along Channel Velocity Contours	149
4.34	1989 Tomakstum Wind and Wind Driven Density Fluctuation Contours	150
4.35	1988 Along Channel Wind vs Selected Dewinded Currents	152
4.36	1989 Along Channel Wind vs Selected Dewinded Currents	154
4.37	1988 Along Channel Wind vs Selected Dewinded Density (as σ_t)	155
4.38	1989 Along Channel Wind vs Selected Dewinded Density (as σ_t)	156
4.39	1988 River Discharge and Tomakstum Dewinded Along Channel Velocity Contours	158
4.40	1988 River Discharge and Tomakstum Dewinded Density Contours	160
4.41	1988 River Discharge and Protection Dewinded Along Channel Velocity Contours	162
4.42	1988 River Discharge and Protection Dewinded Density Contours	164
4.43	1989 River Discharge and Tomakstum Dewinded Along Channel Velocity Contours	166
4.44	1989 River Discharge and Tomakstum Dewinded Density Contours	167
4.45	1989 River Discharge and Protection Dewinded Along Channel Velocity Contours	169
4.46	1989 River Discharge and Protection Dewinded Density Contours	171

Acknowledgements

" 'Did you ever go to a place ... I think it was called Norway?' 'No', said Arthur, 'no I didn't.' 'Pity', said Slartibartfast, 'that was one of mine. Won an award you know. Lovely crinkly edges.' "

-The Hitchhiker's Guide to the Galaxy by Douglas Adams

I'm not sure if this is what first piqued my interest in fjords but the work that this thesis describes certainly provided me with plenty of opportunity to explore our own bit of crinkly coast. While a lot of long hours and hard work went into completing this project, at the end of it all I'm still having fun and this is due to the great number of people and organizations that have lent their support to this research.

Foremost I would like to acknowledge the support and patience of my supervisor Dr. Steve Pond who took a non-physicist under his wing on such a great project. The technical staff at U.B.C. including Denis LaPlante, David Jones, Vivian Lee, Hugh McLean, Pat O'Hara, and Mrigesh Kshatria assisted with the instrumentation and data processing. The officers and crew of the research vessels C.S.S. Tully, C.F.A.V. Endeavour, and C.S.S. Parizeau worked many hours helping us deploy and retrieve moorings. The Institute of Ocean Sciences and Royal Roads Military College graciously lent us some of their equipment and provided assistance in its preparation. This research was funded by the National Science and Engineering Research Council of Canada under Strategic Grants G1820 and G1821 and Operating Grant OGP 0008301 to Dr. S. Pond.

Finally, thanks to snapper@ocgy.ubc.ca for the tireless MIPs and Dr. Susan Allen for the disk space. Thanks also to my office mates and friends in the Department of Oceanography for their infectious energy and enthusiasm. Thanks to Jana, Warren, Anya, Gordon and others for the great diving, ski trips and hikes.

vii

Chapter 1 Introduction

1.1 Coastal Inlets of British Columbia:

The coast of British Columbia, on the west coast of Canada, is characterised by many inlets and sounds that penetrate deeply into the mountains of the coast range. Heavily glaciated during previous ice ages, they exhibit the characteristic 'U' shaped sides, deep basins, and sills of the fjord estuary. Figure 1.1 shows the coast of B. C. and the location of the inlets discussed in this thesis. For more information, Thomson(1981) provides a general introduction to the oceanography of the B.C. Coast and Pickard and Stanton(1980) compare B.C. fjords to others on the Pacific ocean, including those further north in Alaska and in the southern hemisphere in Chile and New Zealand.

During the fall and winter, the coast receives heavy precipitation in the form of rain at lower elevations and snow at higher levels. The resulting runoff and melt water leads to a stratified water column year round in most inlets. This stratification governs the dynamics of the inlet circulation, with a surface outflow above the pycnocline that entrains salt water as it flows seaward and a replenishing inflow below. Tides with ranges of between 4 and 8 m, the seasonal availability of replenishment water of sufficient density to penetrate into the deep basin, and strong winds all modulate this general circulation pattern.

Pickard(1961) describes the physical oceanographic features of British Columbia inlets by such characteristics as depth profile, bottom sediments, runoff, tides, and estuarine circulation. Table 1.1 summarizes for comparison the key topographic parameters of some B.C. inlets. The key factors governing their water properties are river runoff which drives



Figure 1.1: The Coast of British Columbia

the estuarine circulation and proximity to the coastal ocean which determines the characteristics of the deep basin.

Inlet	Length	Sill Depth	Max Depth	Avg ,Runoff
Indian Arm	25 km	20 m	218 m	42 m ³ /s
Knight Inlet	100 km	64 m	540 m	410 m ³ /s
Sechelt Inlet	40 km	14 m	300 m	110 m ³ /s
Observatory Inlet	74 km	46 m	530 m	160 m ³ /s

 Table 1.1:
 Characteristics of Some B.C. Fjords after Pickard(1961)

Pickard(1975) extended the general description to include annual and longer term variability of the deep water properties of the southern coastal inlets, noting that there are variations on a one year period at depths to 100 m or more in most inlets.

1.2 Knight Inlet:

Knight Inlet is on the British Columbia mainland, approximately 300 km north of the city of Vancouver. It was named by Mr. Broughton, commander of Captain Vancouver's tender Chatham, after Admiral Sir John Knight (1748? - 1831) who was a fellow prisoner during the American revolution. It is 100 km long from head to mouth, and opens into Queen Charlotte Strait behind northern Vancouver Island. As shown in Figure 1.2, it is composed of a straight near east-west reach from the mouth to Sallie Point, and then makes an abrupt turn into a sinuous reach with a near north-south alignment to the head. Two sills are present, one at approximately 72 km from the head and with a depth of 68 m and a second at the mouth with a depth of 64 m. Between them an outer basin exists with a maximum depth of 250 m. The inner basin has its greatest depth (540 m) in the sinuous north-south reach just after the bend at Sallie Point. For the purposes of this work the "mouth" of Knight Inlet will be taken as Protection Point in the outer basin and the "sill" will refer to the inner sill. Pickard(1961) classified Knight Inlet as a high runoff inlet and it



Figure 1.2: Knight Inlet with Oceanographic Stations

is stratified throughout the year. The Klinaklini and Franklin Rivers drain a large area of the central coast range and empty into the head of Knight Inlet. These mountains are among the highest in British Columbia, with Mt. Waddington at 4016 m. As a result the bulk of the precipitation, which falls during the winter months as snow, creates a large freshet in the late spring and early summer due to melt water. The Klinaklini, the largest river, has a peak runoff in June at about 800 m³/sec with a smaller second peak in October due to rainfall. The annual average runoff is about 410 m³/sec with a minimum of about 50 m³/sec.

Pickard(1961) describes the summer surface salinity as increasing with distance from the head. Values ranged from near 0 psu at the river delta to values typical of Queen Charlotte Strait (30 psu) at the mouth. Summer surface temperatures of the fresh water at the head are about 10° C and range to a mid inlet maximum before falling again to the outside surface temperatures at the mouth. Pickard (1975) describes the deep water variability as annual with maxima of temperature and salinity in the winter. Oxygen levels in the deepest point in the basin peak in the fall or winter, with values of 3 to 4 ml/l.

Tidal action is predominantly semi-diurnal, with a mean range of 4 m and a large range of 6 m. A strong spring/neap tide inequality is observed and tends to modulate the inlet processes at fortnightly frequencies. During the summer sea breezes develop during the day with weaker land breezes during the night. The up inlet sea breezes tend to be strong due to topographic funnelling from the steep sides of the inlet.

1.3 Thesis Objectives:

The coastal inlets of British Columbia, including Knight Inlet, have been the object of many past studies. Pickard and Rodgers(1959) reported on an early attempt to obtain current meter measurements in Knight inlet, but the limits of available instrumentation

prevented the gathering of large comprehensive data sets and from accurately removing ship motion. While on average their results showed a surface outflow with an inflow underneath consistent with estuarine circulation, the observation period was not long enough to subtract the effects of the wind or investigate the dynamic response of the inlet with respect to the longer period forcing. Farmer(1972), and Farmer and Smith(1978, 1980) investigated the dynamics associated with the tidally driven flows over the sill of Knight Inlet and were primarily interested in the generation and dissipation mechanisms of the internal waves. Webb(1985) in his PhD thesis investigated the propagation of the internal tide around the annular bend in Knight Inlet and as a secondary objective looked at the subtidal residual and record averages of his data. However, he lacked the necessary observations of the wind, and the surface layer above 20 metres to do a comprehensive analysis of the residual response.

Lacking a comprehensive data set, Freeland(1980) investigated the layer thickness dynamics of Knight inlet in the light of Long's(1975) model and concluded that the best fit to existing data was the inviscid case, but noted that the concept of a layered model does not hold true in the vicinity of the sill where mixing has eroded the pycnocline considerably. The studies of Wetton(1981) and van der Baaren(1988) investigated the surface layer dynamics using the dynamic height calculated from CTD surveys and an assumed level of no motion to infer the near surface pressure gradient. While their work gives insight into the dynamics of the sub-tidal response of an inlet, the inverse techniques used suffer inherent uncertainties due to the necessary assumptions and were selected because direct measurements of the response were not available. All previous observations in inlets either lacked the necessary surface layer measurements or did not constitute a sufficiently long time series throughout the water column. They also often lacked simultaneous observations of the wind and runoff so that a conclusive analysis of the low frequency residual could not be performed.

It is the objective of this thesis to first describe experiments that were performed to provide a suitable data set for analysis of the low frequency residual and secondly to use statistical methods to analyse the low frequency response of an inlet and describe it in terms of the forcing caused by runoff, deep water renewal, and wind. The experiments consisted of observations for a period of about a month made throughout the water column of both velocity and density fields, both outside and inside the sill. One experiment occurred during the spring of 1988 during the onset of the freshet and another during the summer of 1989 when the freshet was fully developed. In addition, simultaneous measurements were made of inlet winds and river runoff data were obtained for both years.

The organization of this thesis is as follows; Chapter 2 provides a brief summary of the dynamics important in the low frequency residual response, Chapter 3 describes the experiments performed and presents a sample of the raw time series and spectra, Chapter 4 an analysis of the low frequency residual response of Knight Inlet during these experiments, and Chapter 5 contains the summary and conclusions.

Chapter 2

Low Frequency Residual Dynamics of Fjords

2.1 Definition of the Low Frequency Residual

For the purposes of this work, the low frequency residual is defined as being that portion of the fjord dynamics with a temporal variability longer than the diurnal tide. Classically, the low frequency residual circulation has been portrayed as estuarine with a surface outflow being driven by river discharge at the head. As the water flows towards the fjord mouth, it entrains salt water thus driving a compensating inflow beneath the surface layer. However the natural case is more complicated, with other forces such as the wind and changing boundary conditions at the mouth also driving both surface and deep circulations in the same temporal band. The following section is a summary of the physical processes involved in the low frequency residual dynamics of inlet circulation. For a more comprehensive treatment of the subject of fjord dynamics in general, the reader is directed to Farmer and Freeland(1983) which refers directly to Knight Inlet for many of its examples, Burling(1982) which refers to Indian Arm, and Bo Pedersen(1978) which summarizes current dynamic theories.

2.2 Energy Sources Driving Inlet Circulation

If the water in an inlet is allowed to reach equilibrium it stays at rest unless an energy source is available to do work and move it. However this state is never reached as nature provides various energy sources to do work and cause motion. The principal energy sources are derived from the weather and the tides and their energy paths into the low frequency residual circulation are summarized schematically in Figure 2.1.



Figure 2.1: Energy Paths to the Low Frequency Residual

Surface layer flow is established by fresh water input. In British Columbia it is mainly carried by rivers that empty into the inlet. There is an estuarine circulation with the surface layer entraining salt water as it flows out along the inlet and establishing a compensating flow inwards underneath to maintain salt balance. Winds are topographically confined by the steep sides of the inlet to the along channel directions. Wind stress forces water motion and builds a surface slope whose pressure gradient balances the applied stress. However winds are not always persistent and are often highly variable preventing the surface slope from coming into a complete balance. Surface outflows can be enhanced, reduced, or reversed by this source. The horizontal pressure gradient established by the surface slope will be felt throughout the water column until baroclinic compensation can cancel out the effect in deep water. As this compensation takes some time, there will be some response to the wind at depth as well. Deep water renewal processes are triggered by the availability of salt water at the mouth that is denser than that of the deep basin. Conservation of volume dictates compensating flows above and possibly below the renewal penetration. The result is that the estuarine circulation and the deep water renewal process may form nested vertical thermohaline structures with the water motion at mid-depths a combination of estuarine salt compensation flowing inward and volume compensation from deep water renewal flowing outward.

These first three energy sources are coupled to the weather regime. Rain in the winter and early spring or extended periods of sunny weather can dramatically increase runoff thus changing the estuarine response. Winds are a function of passing storm fronts or sea/land breeze regimes set up by persistent good weather. The availability of salt water of sufficient density to renew the basin deep water is a function of coastal processes that promote upwelling and are themselves a function of the winds established by prevailing meteorological regimes. On the west coast of British Columbia, summers generally bring dominant north westerly winds generated by an Aleutian High. Blowing towards the south

east these winds promote upwelling as the Ekman layer transports water offshore. The upwelled water being cold and saline is available for renewal of the water in the deep basins of inlets, such as Knight Inlet, that are closely coupled to the coastal shelf.

While semidiurnal and diurnal tides have a frequency above the temporal cutoff being considered here, the spring/neap tidal cycle is not and can be expected to modulate the overall response of the inlet. In addition, inlet topography enhances the non-linear interaction of the various tidal constituents enhancing components such as the MS_f , formed through the interaction of the M_2 and S_2 tidal constituents. Low frequency residual dynamics is coupled to higher frequency phenomena via dissipative mixing. This mixing influences the along channel density fields and, through the pressure gradients set up as a result, generates low frequency motions. The following sections discuss each of the forcing mechanisms introduced above and summarizes the corresponding dynamical theory.

2.2.1 Estuarine Flow

Burling(1982) provides a simple explanation of estuarine circulation extended to fjords. Due to the temperate climate and the storage of winter precipitation as snow at higher altitudes, the inlets of the British Columbia coast are positive estuaries. Fresh water input due to precipitation and runoff is in excess of the evaporation throughout the year. This input of fresh water being more buoyant than the underlying salt water flows out on the surface. Interfacial shear leads to entrainment of salt water as the upper layer flows seaward. This progressive entrainment amplifies the surface volume flow and causes a horizontal pressure force which drives a sub-surface inflow of water to conserve salt and volume.

The relations that arise from this conservation of volume and salt in a fjord estuary, are often described as Knudsen's relations and are given as:

$$V_{o} = V_{i} + R + P - E$$

$$V_{o}S_{o} = V_{i}S_{i}$$
where:
$$V_{o}$$
is the volume flow outward in the surface layer
$$V_{i}$$
is the volume flow inward underneath the surface layer
$$R$$
is the volume flow of fresh water input from rivers
$$P$$
is the volume added due to direct precipitation
$$E$$
is the volume loss due to direct evaporation

Assuming direct precipitation and evaporation to be negligible in comparison to the river input, an assumption true for most B.C. inlets including Knight Inlet, the equations for the flow in the two layers can be derived:

$$U_{o} = \frac{V_{o}}{H_{o}B} = \frac{R}{H_{o}B} \left(\frac{S_{i}}{\Delta S}\right) \qquad U_{i} = \frac{V_{i}}{H_{i}B} = \frac{1}{H_{i}B} \left(V_{o} - R\right) = \frac{R}{H_{i}B} \left(\frac{S_{o}}{\Delta S}\right)$$

where:

- U_{o} is the flow out in the surface outflow layer
- U_i is the flow in the inflow layer
- S_o is the salinity in the surface outflow layer
- S_i is the salinity in the inflow layer
- ΔS is the salinity difference $S_i \tilde{S}_o$
- H_o is the depth of the surface outflow layer
- H_i is the depth of the lower inflow layer
- B is the breadth of the inlet

Knudsen's relations assume that the horizontal "diffusion" of salt due to tides and turbulence is small enough to neglect. This is probably a reasonable assumption in fjord estuaries. In the application of these relations practical difficulties do exist however. For example the depth of H_o and H_i are not well known and noise is present in salinity measurements due to wind and tide. Also both U_o and U_i are layer averages and so nothing can be learned about the vertical distribution of velocity within these layers from the application of these relations. Note that an increase in river discharge (R) may not be reflected in a larger surface layer velocity (U_o). The increase in discharge also lowers the surface layer salinity stengthening the pycnocline by increasing the salinity difference term (Δ S). This response is consistent with the increased stability of the water and a resulting reduced entrainment.

2.2.2 Density Flows

Exchange of water with the outside is not limited to salt compensation inflow of the estuarine circulation. In the absence of mixing and given time to achieve a steady state, the water of the deepest part of an inlet basin will be denser than the water above it and the water column is stabilized by gravity. Over a longer time scale the water properties in the basin will change. In the limit of no mixing, molecular processes will diffuse salt towards the surface where surface flow will carry it seaward. In the more practical case, mixing induced by wind, interfacial shear and tides produce a much larger 'eddy' diffusion. As a result of this diffusion, the density in the deep basin is lowered and gravity flows of denser water from outside the fjord will occur. This inflow will displace the less dense water and lift the isopycnals of the basin, thus raising the potential energy of the water column. Renewal may be to the bottom or to mid-depths depending on the density of the inflowing water and the density of the water resident in the basin.

Density flows are responsible for the oxygen levels that are found in the basin. Bacteria, in the process of breaking down organic matter from the surface, will slowly consume the available oxygen leading to anoxic conditions in the fjord basin if a replenishment source were not available or if downward diffusion is not fast enough. Knight Inlet, as with most most inlets on the B.C. coast is renewed with ventilated water from the outside. Pickard(1961) reported that Knight Inlet rarely has an oxygen level below 3 to 4 ml/l. These values at depths to 500 m or more are higher than in many other B. C. inlets suggesting that some renewal occurs every year in Knight Inlet.

Diffusion is not the only mechanism that controls the renewal of deep basin water.

External forcing can change boundary conditions at the mouth and can cause, enhance, or block these density flows. For example, if the available source water for renewal becomes denser, it will tend to flow over the sill and penetrate into the basin. Such changes are often seasonal and for Knight Inlet, due principally to the upwelling of high salinity water off the west coast in the summer.

Source water must be denser than that in the basin for a density flow to occur. However, if available kinetic energy is sufficient it need not be at or above sill height to penetrate. de Young and Pond(1988) derived a blocking equation to relate the blocking height to the mean flow and the stratification based on the Bernoulli equation. That is, the kinetic energy of the approaching flow must be at least equal to the energy of the water at the top of the sill in order for penetration to take place. Their blocking equation reduces to:

$$h = \frac{u}{N}$$

where:

h is the blocking heightu is the velocity of the penetrating flow

N is the buoyancy frequency, a function of stratification

Enhanced tidal velocities during spring tides add to the kinetic energy available and therefore may trigger renewal. However, if the sill is shallow and extended enough, the enhanced tidal velocities of the spring tide may promote mixing that reduces the density of the source water and block renewal. The Froude number, the ratio of velocity to the internal wave speed gives a way to parameterize the important factors governing these two effects. If the Froude number is greater than 1, turbulent mixing will occur and blocking will result. If it is less than 1, then the sill depth is great enough or stratification strong enough to suppress mixing and renewal will be enhanced during spring tides. Knight Inlet has a relatively deep sill, with strong stratification and Farmer and Freeland(1983) calculate the Froude number at the sill as being much less than 1. Therefore renewal is expected to be enhanced during spring tides. Cases of renewal blocking during spring tides has been reported by Geyer and Cannon(1982) for Puget Sound, de Young and Pond(1988) for Indian Arm, and Griffin and LeBlond(1990) for renewal in Georgia Strait through Boundary Pass.

Other energy sources can modulate renewal. Strong wind mixing deepens the upper layer as it mixes fresh water down and erodes the pycnocline. If deepened to sill depth, this process can also block renewal. Knight Inlet, with its comparatively deep sills, is unlikely to experience this type of control. Burling(1982) suggests that external winds may also cause a response by sloping the surface to balance the wind stress. In order to balance horizontal pressure gradients, the pycnocline tilts in the opposite direction by a factor of about 100 times that of the surface slope and affects the availability of source water at the mouth. Klinck et al.(1981) used a two layer numerical model to investigate the interaction of inlet dynamics with a wind driven coastal regime. They found that such forcing can produce large velocity shears in the fjord thus affecting the eddy diffusion in the basin.

de Young and Pond(1988) examined renewal events in Indian Arm and calculated vertical eddy diffusivities based on the change in basin water properties over the time between events. Their results were consistent with those of Gargett and Holloway(1984) that showed vertical eddy diffusivity to be dependent function of the buoyancy frequency N, itself a function of the stratification. Gargett and Holloway(1984) proposed that this relationship occurred as vertical mixing was principally controlled by internal wave action. However, the exact functions for vertical diffusivity as a function of N varied considerably from that proposed by Gargett and Holloway for energy derived from internal wave fields of a single frequency alone. de Young and Pond suggested that the differences may be due to energy sources other than internal waves contributing to the vertical mixing.

2.2.3 Wind Driven Flow

One source of wind predominant in the summer is the diurnal sea/land breeze where solar heating over the land produces an inflow wind during the day and with the more rapid cooling of the land mass at night an outflow wind. The sea breeze is observed to be stronger producing a net up inlet wind flow. Farmer and Freeland(1983) suggest that this difference is due to the relative thickness of the air mass in each case, with the sea breeze occurring over considerable depth but the land breeze restricted to a thinner nocturnal boundary layer heavily influenced by friction. Due to the strength of the pycnocline and the large stability found in the surface layer of a high runoff inlet it is unlikely that much of the wind energy would mix down into the deep water, but considerable thickening of the surface layer might be expected.

Wind stress accelerates water movement and, in the inviscid homogeneous case with no boundaries, a steady wind would give a steadily accelerating wind-driven layer. However this is not the case in fjords where, due to boundaries, a surface pressure gradient builds to counteract the wind stress and the turbulence generated by shear mixes the momentum down into the surface layer and gives rise to a larger interfacial friction. Large and Pond(1981) give the drag coefficient as observed over the open ocean and thus wind stress may be calculated from the observed winds as:

 $\tau_{w} = \rho_{a} C_{D} u_{w} |u_{w}|$

 τ_{w} is the wind stress where:

is the density of air (1.25 kg/m³)

Pa is the along channel wind at 10 m above the surface uw

is the drag coefficient at 10 m and is 1.2×10^{-3} for $u_w < 12 \text{ m/s}$ and varies linearly with u_a until 4 x 10⁻³ for $u_w = 32$ m/s, the limit of their observations

Direct estimates of the drag coefficient for confined areas such as inlets are limited, but it is likely to behave in a similar fashion. Note that energy input, proportional to $\tau_w |u_w|$, increases as fast as or faster than the cube of the wind speed and that a few extreme wind events may cause a disproportionate amount of mixing especially when stratification is weak in winter.

For the case of a steady wind, and the building of a compensating pressure gradient through a surface slope, the surface outflow may be reversed until the new balance is achieved. If the wind switches direction to an outflow, the enhanced sea surface slope adds to that hydraulic head of the fresh water and a strong outflow occurs. Pickard and Rodgers (1959) in measuring currents in Knight Inlet observed that the wind appeared to enhance or impede the surface outflow. However winds are rarely steady in an inlet and constant forcing from the wind is not likely to be the case. Further, studies by Wetton(1981) and van der Baaren(1985) indicate that the interfacial friction coefficient is also modified by wind further implicating the wind's influence on the dynamics of an inlet.

Farmer(1972) describes the influence of wind upon the surface layer of Alberni Inlet on Vancouver Island. Using cross-spectral analysis, he found that the wind and the longitudinal current were closely coupled in the diurnal band and used the phase angles between the wind and the current to estimate bulk eddy viscosities for the surface layer (1 to 10 cm²/s). He found that strong up inlet winds induce a sudden thickening in the surface layer at the inlet head and the disturbance appeared to propagate back down the inlet from Holm Island to Stamp and Sproat Narrows, suffering an attenuation as it travelled. The return to equilibrium was observed to take several days. He went on to develop a simple frictional model to explain much of what was observed in the change of surface layer thickness on the basis of measured wind speeds. Figure 2.2 shows the response of



Figure 2.2: Wind Response in Alberni Inlet, Surface Layer Thickness and 2 m Along Channel Currents Note Holm Island is closest to the head, Sproat Narrows the closest to the mouth. From Farmer(1972)

both the 2 m current and the surface layer thickness to a strong up inlet wind event in Alberni Inlet.

Buckley and Pond (1976), working with data obtained by surface layer drogue tracking in Howe Sound, were able to attribute horizontal surface layer circulation mainly to the effects of the wind. They found that the time to produce a counter balancing pressure gradient to strong up inlet winds was in the order of 6 to 7 hours. As it would only take about 1 hour for sufficient water to flow and produce the necessary surface slope, they surmised that the wind effect must deepen the surface layer as well. In other words, the response time important in the set up process is the baroclinic and not barotropic response time. Weak density stratification in the surface layer allowed wind momentum to be rapidly mixed down to the pycnocline, thus allowing the surface layer to behave as a slab. The strong pycnocline effectively inhibits turbulent mixing as the buoyancy suppression terms dominate over the mechanical production terms. Thus, the interfacial frictional effects do not seem large. (note due to the small size of Howe Sound, say compared with Knight Inlet, there is very little salt in the surface layer at the mouth, approximately 4 psu). When the up inlet wind dropped, the pressure gradient due to the surface slope dominated over the effect of the fresh water input and produced large down inlet surface velocities. Buckley and Pond concluded that the variation due to the wind effects dominated by a factor of 10 over the contributions due to the tide and the river discharge. They also noted that large lateral shears were present in the surface flows particularly in the top 15 km of the inlet.

2.2.4 Tidal Forcing

An inlet mouth is an open boundary allowing the local barotropic tide to drive an exchange of water between the coastal ocean and the fjord. Knight Inlet is approximately of uniform width and extends 72 km inward from the inner sill which has a depth of 68 m. The mean tidal range is 4 m and on large tides it is 6 m; the tides are predominantly semi-diurnal. The magnitudes of the currents produced by this forcing can be calculated from the tidal prism following Burling(1982) as:

$$u_{T} = \frac{A_{o}}{a_{o}} \left(\frac{h}{t}\right)$$

where:

 $u_{\rm T}$ is the magnitude of the depth averaged tidal current

- A_o is the inlet's area, the base of the 'tidal prism'
- h is the height of the 'tidal prism'
- ao is the cross-sectional area of the inlet mouth
- t is the time of a half tidal cycle, i.e. to flow in or out, about 6 hours or 2.2 x 10⁴ sec for semi-diurnal tides

At the sill for a 4 m tide, u_T is calculated to be in the order of 19 cm/s. u_T is the average current over the tidal cycle with the peak currents expected to be about 57% larger or about 30 cm/s. While the constituent tidal elevations are comparable for both the diurnal and semidiurnal tides in Knight Inlet, the shorter period of the semidiurnal tides results in a faster current flow.

Constrictions at or near the mouth, either a narrowing of the fjord or a sill that reduces the depth, will lessen a_0 and therefore increase u_T for a fjord with a given surface area. As kinetic energy is proportional to u_T^2 , Farmer and Freeland(1983) calculated the along

channel tidal energy from:

$$u_{T}^{2}(x) = \left(\frac{1}{A}\frac{d\zeta}{dt}\int_{0}^{x}Bdx\right)^{2}$$

where:

A(x)is the local cross sectional areaB(x)is the channel breadth ζ is the tidal heightxis the position from the inlet head

For a given inlet size, the kinetic energy of the tidal flow will be larger when the sill is more constricted or for a given sill cross sectional area, the inlet area is large. Figure 2.3 shows the longitudinal distribution of kinetic energy of Knight inlet as determined by Freeland and Farmer. While Knight Inlet's sill is reasonably deep at 68 m, it is one of the larger inlets on the B.C. coast. It can therefore be expected to be highly energetic near the sill.

Sill dynamics are the key generation mechanism for internal waves. The barotropic tide interacting with the sill tends to lift the density surfaces and generate an internal tide, extracting energy from the barotropic tide and coupling it to the internal response of the inlet. During the spring tide with its higher flow velocities, Freeland and Farmer(1980) found more energy extracted from the barotropic tide and coupled into the internal response than during the neap tide. This energy transfer results in a further response at the MS_f, the beat frequency of the M₂ and S₂ periods in addition to the non-linear interactions mentioned in section 2.2. They argue that because more energy is available during the spring than the neap tides, the low frequency residual circulation will be modulated at the MS_f frequency. Freeland and Farmer(1983) suggest a large MS_f component is the signature of an inlet whose mixing is derived from the tide. Thus the higher frequencies and their interaction with topography are likely to have an effect on the low frequency residual circulation.



Figure 2.3: Tidal Kinetic Energy calculated from Depth Mean Velocity Squared in Knight Inlet

Above: Depth of Knight Inlet as a function of distance from the head. Below: Depth mean velocity squared, proportional to kinetic energy due to tidal forcing, subject to zero phase change along the channel. The calculation takes account of cross-sectional area rather than the depth alone, but the figure serves to emphasize the importance of shallow sills, when they are well removed from the head, as locations for strong tidal mixing. From Farmer and Freeland(1983)

2.2.5 Mixing and Diffusion

Knight Inlet is stratified and with a well defined sill geometry. Large tidal velocities interacting with the sill lift the isopycnals and generate an internal tide that propagates away from the sill. If the internal response is large, non-linear terms in the equations of motion may also be important. Farmer and Smith(1980) observed flow separation, stationary lee waves, and an internal hydraulic jump at the Knight Inlet sill with acoustic observations. Freeland and Farmer(1983) point out that non-linear effects give rise to strong fortnightly currents and to tidal harmonics such as the M₄ and M₈ being present.

If the fjord internal structure were two layer, the internal tide would propagate along the interface in a manner analogous to surface waves and would either dissipate against or reflect from the fjord boundaries. Knight Inlet is not a straight channel and has an almost right angle bend just east of Tomakstum at Sallie Point. The theoretical study of Webb(1986) and modelling of Stacey and Pond(1992) showed that there is minimal reflection of the internal tide from this bend, with energy propagating along the inlet towards the head. Stacey and Pond(1992) show that the phase change down the inlet is not constant, however, and undergoes sudden 180 degree changes in phase. This is one of the signatures of a standing wave indicating that some reflection from the head of the inlet does take place.

A two layer model is an approximation only and the pycnocline is really a sharpening of a continuous density gradient with depth and the frequency response is restricted by the buoyancy response at the pycnocline. Therefore, free internal waves are confined to radian frequencies between the local value of the Coriolis parameter f and the maximum of the Brunt Väisälä frequency N:

$$N = \sqrt{-\frac{g}{\rho} \left(\frac{d\rho}{dz}\right) - \frac{g^2}{c^2}} \cong \sqrt{-\frac{g}{\rho} \left(\frac{d\sigma_t}{dz}\right)}$$

where:Nis the buoyancy frequency (rad/s), a function of fluid stability
ggis the acceleration of gravity
 ρ gis the fluid densityzis the depth
ccis the speed of sound
 σ_t is defined as $\rho(s,t,0) - 1000$

Pond and Pickard(1983) state that the approximate expression is suitable except in the deep ocean. At the low frequency limit, $f = 1.13 \times 10^{-4} s^{-1}$ for the latitude of Knight Inlet. At the high frequency end, Webb(1985) calculated monthly mean N² profiles based on CTD and cyclesonde data and gave a maximum for N of about 0.178 s⁻¹. In Knight Inlet, Pickard(1961) reported internal waves with periods ranging from 1 minute (0.105 s⁻¹) to 12 hours (1.405 x 10⁻⁴ s⁻¹) corresponding to these limits. Internal tides being forced Kelvin waves are possible despite these limitations and Farmer and Freeland(1983) note the presence of an internal diurnal tide in Knight Inlet.

Freeland and Farmer(1980) calculated the power withdrawn from the barotropic tide in Knight Inlet by using the observed phase difference in the barotropic tide. Because the sill is short compared to the tidal excursion and deep, they estimated that only ~3% would be lost due to friction with the rest of the energy transferring into the internal processes; internal waves or the non-linear processes documented by Farmer and Smith(1980). Because of the pronounced variation from month to month in the power lost as the stratification changed, they concluded that the energy removed must be coupled into the internal tide.

Stacey (1984) developed a linear model of progressive internal waves to account for energy dissipation of the tide over the sill in Observatory Inlet including seasonal variations. This model was applied to Knight Inlet by Stacey(1985) who found that while non-linear sill

processes as documented by Farmer and Smith(1980) are present, these are generally of less importance than the internal tide as his linear model accurately represented the tides. deYoung and Pond(1989) later estimated the power lost to high frequency internal waves at $\sim 2\%$. Stacey(1985) found that most of the energy removed from the barotropic tide could be accounted for in the first two internal modes with the fourth and above internal modes contributing less than the third.

The energy partition between the first two modes was found by Stacey(1985) to be highly variable and dependent upon the presence of a fresh water surface layer and the density gradient in the deeper water. The mode 1 energy was correlated with deep water renewal events with an amplification in the mode 2 response correlated to the freshening of the 5 - 10 m surface layer from river runoff in the late spring and early summer. A decrease in internal energy occurred during the late fall when the stratification of the entire water column was lessened. He concluded that the surface layer may have a major influence on the circulation in an inlet even when that layer is very thin relative to the total depth of the inlet and is having little apparent influence on the overall velocity field. The presence of the fresh surface layer can have a strong influence on the amount of energy withdrawn from the barotropic tide and, because these motions are dissipated internally within the inlet, influence the internal circulation through the resultant mixing.

Stacey and Pond(1992) applied a modified form of Dunbar's(1985) XZT model to Knight Inlet using data from a cyclesonde near Protection Point to provide boundary conditions. The resultant tidal velocity and density fields compared favourably with the data from three other cyclesondes positioned along the inlet with the exception of the MS_f density fluctuations which appeared to be over represented below the surface layer. This discrepancy was probably due to lack of the mixing of density in the model as such mixing would tend to reduce this variance. By removing the non-linear terms from the momentum

equations they found that the internal motions could still be well represented, the exceptions being the shallow water constituents, the M_4 and MK_3 . These results are consistent with the conclusions of Stacey(1985) and with deYoung and Pond(1989) that little of the total energy lost from the barotropic tide is coupled into non-linear or high frequency internal motions.

Webb and Pond(1986b) used cyclesonde data to determine the M_2 tide and found by modal decomposition that only 40% of the power removed from the barotropic tide could be accounted for at Tomakstum, at least when their observations were made. However this is not necessarily inconsistent with Stacey(1985) as it is possible that most of the energy could be fed into the internal tide and then be removed while it is close to the sill generation region. Stacey and Pond(1992) showed with the XZT model that energy flux rises rapidly to its maximum and then decays a short distance away from the sill. Further, the results of modal decomposition are highly dependent upon the modes fitted and, as surface layer data were not available to Webb and Pond, these results may not be entirely accurate.

2.3 Previous Observations of the Low Frequency Residual of Inlets

Early attempts to measure inlet circulation directly were restricted by the lack of internally recording instrumentation. Measurements had to be done from a ship anchored on station. Costs and the problems of ship movement, made the acquisition of data with sufficiently long time periods to study the low frequency residual impractical as the energetic high frequency tidal and wind 'noise' could not be removed from the lower frequencies with any degree of confidence. Further, the simultaneous acquisition of data from more than one location is necessary if a complete analysis of inlet circulation is to be carried out. Without the ability to do direct measurements, Pickard and Trites(1957) attempted to estimate the residual circulation based on the inlet heat budget.

Pickard and Rodgers(1959) were the first to attempt direct current measurements in Knight Inlet. They measured current profiles at the sill and in the neighbourhood of Tomakstum Island using a combination of Cheasapeake Bay Institute (CBI) drag near the surface and an Ekman Current meter that was lowered to deeper depths. All measurements were done from a ship anchored on station, but with only a single anchor.

They found that correcting for ship motion was a significant problem particularily at deeper depths where the motions of the meter relative to the ship were uncertain. They also found that wire drag on the CBI drag was significant, and despite an attempt to correct for this problem, limited its use to the upper 20 m. Despite these problems, these early measurements were used to describe the mean and the oscillatory currents. An attempt was made to explain their features in terms of the estuarine circulation, the winds, and the tides. Twenty five hour averages were used to subtract tidal components. As observations were labour intensive due to the methods available, only a few days of data were obtained at each station.

Over the sill and in the absence of significant wind, they found a net outflow in the upper half and a net inflow below. They also found the tidal oscillations to be in phase at all depths. In the presence of an up inlet wind the surface flow reversed, with a net outflow occurring below the surface layer and the bottom net inflow being maintained. Figure 2.4 shows the 25 hour mean current profiles at station 3.5 at the Knight Inlet sill and illustrates the switch of the dynamics from a two layer to a three layer system. Near Tomakstum Island, both oscillatory and net currents were observed all the way to the bottom with the net currents showing a three or four layer pattern, rather than the two layer flow previously thought to exist.



Figure 2.4: Net Current Profiles over the First and Last 25 Hours just Outside the Knight Inlet Sill at Station 3.5, 6-8 July 1956. From Pickard and Rodgers(1959)
Pickard and Rodgers found significant discrepancies between calculations of net transport and their measurements and suggested these might be due to:

- 1. Ebb flow preferential to mid-channel
- 2. A horizontal eddy with its down-inlet portion in mid-channel
- 3. Interpolation errors due to widely spaced points in the profile
- 4. The mean velocities significantly smaller than the measured mid-channel values
- 5. Artifacts associated with the presence of internal waves

Pickard and Rodgers also observed significant cross-channel flow that did not alter 180 degrees between the flood and the ebb and had a different direction with depth. They suggested that topography might be the cause.

Webb(1985) investigated the reflection of energy at the bend in Knight Inlet. In addition he examined the residual circulation using two months of cyclesonde data from July through September of 1983. He used harmonic analysis to subtract seven tidal constituents of the the diurnal and higher frequencies from the observations. This residual was then passed through a 24 hour (8 point) running average to smooth the result. A strong MS_f component (1 to 5 cm/s) was found, stronger than what could be accounted for by the barotropic tide alone. This result is consistent with the hypothesis of Freeland and Farmer (1980) that enhanced mixing during spring tides would modify the low frequency residual. In addition to the MS_f, there is energy at frequencies between the fortnightly and diurnal bands perhaps associated with local wind and/or offshore effects. The eight data records from profiling current meters deployed at four stations, Protection Pt., Lull Bay, Tomakstum Island, and Adeane Point (locations are shown in Figure 1.2) during the summer of 1983 were averaged over 29.5 days. These residual profiles, shown in Figure 2.5, were found to be reasonably repeatable from month to month. The Protection Point profile showed outflow in the upper part of the water column and an inflow at depth, although the zero crossing was surprisingly below the pycnocline. At Tomakstum Island three layer flow is observed



i Residual longitudinal velocity profiles at Protection Point. Solid line is July 1983, dashed line is September 1983. Positive is up inlet.



Residual longitudinal velocity profile at Lull Bay July 1983. Positive is up inlet.



ii Residual longitudinal velocity profiles at Tomakstum Island. Solid line is July 1983, long dashed line is Tom-S September 1983, short dashed is Tom-N September 1983. Positive is up inlet.



iv Residual longitudinal velocity profiles at Adeane Point. Solid line is July 1983, dashed line is September 1983. Positive is up inlet.

Figure 2.5: Residual Circulation Profiles from Knight Inlet, July and September 1983. From Webb(1985)

with outflow at the surface and in the deeper water, while at intermediate depths an inflow was present. At Lull Bay he found an outflow at all depths; this result is not consistent with the idea of lateral homogeneity, since mass must be conserved. It should be noted that Webb's data were collected by profiling current meters that due to their sub-surface mooring arrangement were not able to measure the upper 20 m nor the lower part of the water column below 190 m. At Tomakstum (depth ~340 m) and Adeane (depth ~530 m) much of the deep water was not sampled. Further, the surface layer thickness is of order 10 m in Knight Inlet and hence essentially missing from the data set. This was not an experimental oversight but a practical limitation of the current meters available for his study.

Wetton(1981) did a short study of the surface layer dynamics of Knight Inlet using dynamic heights derived from CTD data to estimate the sea surface slope and hence derive the circulation. In doing so a depth of no motion was assumed. The pressure gradients calculated show little evidence of a reverse pressure gradient to drive a compensating inflow, and therefore the depth picked may not have been motionless. The pressure gradient was not found to be balanced by inertial terms and therefore must be balanced by friction. Wetton suggested that wind may also enter into the balance of forces with the up inlet wind stress acting to modify the pressure gradient so as to increase the surface slope required to drive the flow. For one case he estimated the effect of wind to be less than half the pressure gradient established by the river runoff.

van der Baaren (1988) refined the work of Wetton(1981) and investigated the surface layer dynamics of Knight Inlet using a two layer steady state model derived from integrating the 2D momentum equations over each layer. Layer thickness was calculated to achieve the same potential energy and internal wave speed as observed in CTD data from 1986 and

1987. For reference, the vertically integrated momentum equation for the surface layer used by van der Baaren has the following form:



Surface slopes were estimated by calculating dynamic heights from the observed densities. By doing a least squares fit through the dynamic heights the isobaric slope was estimated. The Reynold's stress was determined in terms of both wind stress and interfacial stress between the upper and lower layers. As all quantities but the coefficient of interfacial friction are known in the depth integrated equations for the surface layer it could be estimated from the data. Neglecting the wind stress, she found values of order 5 x 10^{-3} for 1987 and 10^{-2} for 1986 when the runoff was approximately double.

van der Baaren found the interface depth was a constant thickness (7 m in 1986 and 3 m in 1987) inland of the sill but increased rapidly seaward to the mouth. Her estimates of surface layer velocity seem consistent with my observations of average velocity except near the mouth, where her results seem to be overly large. Except near the mouth, inertial terms are small compared to the pressure gradient. Therefore the dynamic balance is between the surface pressure gradient and interfacial friction. At the mouth and sill where the surface layer was moving fastest, the two depth integrated inertial terms were almost 5 times larger than elsewhere; the first term(i) was generally half the second term(ii); their sum is about one half of the pressure gradient. However, the two layer model does not accurately depict the dynamics near the mouth and the velocities calculated from Knudsen's relations are

overly large, probably giving rise to an over estimate of the inertial terms. With no wind, the estimate of the interfacial friction was expected to be low. Wind would increase the pressure gradient and the interfacial stress would then rise to maintain a balance of forces. She found the interfacial friction coefficient increased by about a factor of four when wind was included. A steady state, she concluded, is unlikely to hold unless the wind has been blowing steadily for some time; thus the interfacial friction is somewhere between her values. Figure 2.6 gives van der Baaren's estimates for layer velocities from the interface depth for 1986 and 1987. In conclusion she found that the interfacial friction reacts to tides, wind and runoff.



i Surface layer velocity as a function of distance from the head of for Knight Inlet in 1986 and 1987.



ii Inflowing layer velocity as a function of inflowing layer thickness for Knight Inlet for 1986.

Figure 2.6: Layer Velocities for Knight Inlet for 1986 and 1987 From van der Baaren(1988)

Chapter 3.0

Experiment Descriptions

3.1 Experiment Design

Previous data sets of inlet circulation have lacked the detailed observations of the surface layer necessary to analyse the low frequency residual response. As a result, new experiments were designed that would span the entire water column at both the mouth and in the inlet basin. Particular attention was paid to including observations through the surface layer and the wind forcing. While the key question being investigated was the residual response, the cost of making such a complex set of measurements dictated that care be taken to address other needs as well. These other areas of investigation while not directly related to the goals of this thesis were the verification of inlet general circulation models and the modal response and dissipation of the internal tide. These later objectives are discussed briefly here in so much as they impacted the experimental design. I would not like to leave the reader with the impression that the design of these experiments was my own. Steve Pond at the Department of Oceanography, U.B.C. and the supervisor of my work has amassed a great deal of experience in making observations of this kind, and was the chief architect of these experiments. In that the quality of the data is largely due to the initial experimental design, and that the field program occupied a great deal of my time at U.B.C. it is discussed in detail here.

3.1.1 Low Frequency Residual Circulation

In order to fully examine the low frequency (subtidal) residual circulation in a fjord it is necessary to have high vertical and temporal resolution of both velocity and scalar data throughout the the water column including the surface layer. Vertical resolution must be adequate to resolve the anticipated high shears in the surface layer and to gauge the high frequency responses that are concentrated near the pycnocline. These responses, although containing relatively high energy levels, will fluctuate over a corresponding smaller vertical scale due to the larger restoring forces in this region. Data are also necessary below the pycnocline, though at a lower spatial and temporal resolution. The dynamics of the surface and deep water are intimately tied together through the entrainment and the fact that other processes such as deep water renewal and wind mixing may modulate the response. Data records of the relevant forcing must also be obtained for open boundary conditions including the surface meteorology. Observation records must be of sufficient length so that forcing such as the tides and wind can be analysed and subtracted through statistical methods. Previous attempts by Pickard and Rodgers(1959) and Webb(1985) either were of insufficient length to subtract interfering signals from other low frequency energy inputs or lacked the direct measurements in the surface layer.

3.1.2 Verification of Inlet General Circulation Models

Numerical models such as those by Dunbar(1985) and Nowak(in prep) have been developed. With further refinement they may be able to predict the water properties, oxygen content, and the movement of contaminants and pollutants in these economically and ecologically important bodies of water. In order to refine these models and verify their usefulness as practical predictive tools, comprehensive data sets of velocity and scalar fields of representative fjords must be collected. Knight Inlet represents a high runoff, highly energetic inlet and one objective of these experiments was to provide data sets adequate for this purpose. Previous data though less comprehensive were gathered by U.B.C. Oceanography for Indian Arm, a low runoff inlet. A similar experiment has also been recently completed in Sechelt Inlet, a small inlet featuring multiple side branches and important to the mariculture industry.

For model calculations, the data set must contain the initial conditions for the inlet at the start of the experiment. The density distribution calculated from a CTD survey is required as pressure gradients are responsible for establishing the velocity field. The data set must also describe the forcing at the mouth where an open boundary couples the local external oceanographic conditions to the internal dynamics of the inlet. Because wind has been shown to be important in driving the surface layer dynamics, air temperature and wind speed must be recorded to allow for the inclusion of surface heat flux and wind stress. River runoff establishes a pressure gradient that drives the estuarine circulation and so it too must be obtained for the duration of the experiment. In order for the verification to have a reasonable degree of confidence, several key sites must be observed in the interior of the inlet.

3.1.3 Modal Response and Dissipation of the Internal Tide

The internal response of an inlet is multi-modal. The lack of observations throughout the entire water column, especially in the surface layer and through the pycnocline, has prevented the definitive determination of the modal composition of the internal response to external forcing. While previous work such as that by Stacey(1984, 1985) and Webb(1985) has attempted to resolve this question, limitations in previous current meter design and lack of sufficient equipment and ship time have prevented data being gathered at sufficient resolution and through the entire water column to resolve this question completely. Freeland(1984) speculates that the modal response with the highest energy is the one with a zero crossing at sill depth. Stacey(1984) observed that modal response in Observatory inlet varied with oceanographic conditions. Considerable work remains to be done to resolve these questions completely.

Knight Inlet has a near right angle bend at Sallie Point and while the question of whether the internal tide was transmitted through or reflected from this bend was answered by Webb(1985), a great deal remains to be explained. For example, the exact nature of the high dissipation rate in the straight reach, particularily in the immediate vicinity of the sill still needs to be explained. Simultaneous measurements outside the sill, just inside the sill, and at the end of the straight reach are necessary to investigate this dissipation and how it affects the mixing of the surface layer and the more saline water below the pycnocline.

3.2 Instrumentation

In order to meet the objectives a great deal of instrumentation of various types had to be employed. Three stations were picked for investigation; outside the inner sill at Protection Point, just inside the inner sill, and at Tomakstum Island near the end of the straight east/west reach (see Figure 1.2 for the locations). In 1989 more instrumentation was available and a fourth station at Axe Point in the sinuous north/south reach and nearer the head of the inlet was also included. Due to the narrowness of the channel and the limited number of suitable instruments lateral homogeneity was assumed and a single mooring in the centre of the channel at each station was used. Each station was composed of a vertical array of current meters that measured and internally recorded the vector horizontal velocity, and scalar temperature and salinity (through conductivity) fields from which density could be calculated to establish a complete picture of the water column. Each experiment lasted approximately 31 days, the limit of the recording capacity for most of our instrumentation and the minimum required to resolve the longer period monthly tidal constituents by harmonic analysis. All internally recorded data was later retrieved from the instruments at the end of the experiment and processed on our mainframe and workstations at the U.B.C. Department of Oceanography.

Each station was sampled by a cyclesonde profiling current meter described by van Leer et al.(1974). This unique instrument can be programmed to vertically profile through the water column by inflating or deflating a helium gas bladder to adjust its buoyancy by about 1000 grams. As it moves up or down, it samples at fixed time intervals which approximately determines the vertical sampling resolution. A pressure sensor is included so that the depth of each sample is known and to allow data processing to remove the vertical velocity component due to profiling from the measured current vector. Currents are measured by two savonius rotors on either side of the instrument. The rotors are on a horizontal axis and have opposite direction of rotation. When the two rotor speeds are averaged the slight angle of attack sensitivities of the rotors effectively cancel. Savonius rotors normally have a stall speed of about 2.5 cm/s, however when the cyclesonde is profiling the rotors are biased by its vertical movement of approximately 10 cm/s. Alternate up and down profiles were set for every three hours with a one minute sampling rate and consequent vertical sampling resolution of about 4 - 10 metres from 15 to 190 metres depth. The profiling speed changes as the instrument uses up its gas supply. It falls at about 10 metres per minute and rises at about 4 metres per minute at the start. By the end of the deployment these rates are reversed. Below the surface layer, this sampling strategy was judged adequate to resolve the important higher frequency tidal constituents such as the M_4 . When not profiling the cyclesonde was set to sample at least every 5 minutes to provide an inter-comparison record for instrumentation placed either above or below its profiling range. The depth limitation is a function of helium gas storage capacity and the number of profiles to be performed. Where depths exceeded 190 metres, Anderaa RCM-4 and RCM-7 current meters were deployed at 40 metre intervals and sampling at 10 minute intervals to the bottom of the water column.

At two stations, outside the inner sill at Protection Point and in the straight reach inside the inner sill at Tomakstum Island, an additional surface mooring was deployed nearby with

InterOcean S4 electromagnetic current meters through the surface layer at 2, 4, 6, 9, and 12 metres. Vertical resolution was concentrated in the surface layer where there are few previous observations and none for longer than two to three days. These current meters, described by Lawson et al.(1983), operate by generating a magnetic field in the water surrounding the instrument and allowing the motion of conductive sea water to generate a voltage by Faraday's Law of Induction. It is then measured and recorded by an internal microcomputer that performs true vector averaging of the velocity data based on samples taken every half second. These instruments are ideal for surface layer work because they are immune to the errors induced by rotor pumping and directional alignment with surface wave trains as documented by Kollstad and Hansen(1985) as well as other authors. In addition, they have the ability to be programmed for a variety of burst or continuous sampling strategies and sufficient memory capacity to allow reasonable sampling rates in the region of the highest buoyancy frequency.

Storage capacity and power limitations prevented the sampling strategy employed for these instruments from resolving the highest frequencies expected in this area, and so alternate instruments were set to burst sample to provide an estimate of how much energy might be aliased into the other measurements. Additional storage capacity was added to the current meters for the 1989 experiment which allowed the sampling intervals to be halved for this experiment. Sampling rates were set as shown in table 3.1. The 15 metre cyclesonde data and the 12 metre S4 data provided the ability to inter-compare instrument performance near the bottom of the surface layer. Table 3.1 summarizes the instrument sampling strategies used for both the 1988 and 1989 experiments.

Geodyne toroid buoys were located at the top of each surface mooring and equipped with a J-Tech or Anderaa meteorological station recording air temperature, wind speed, and wind direction at 15 minute sampling intervals at 4 metres above the water's surface.

Depth (metres)	Instrument	Mode	Kn88 Sampling Rate	Kn89 Sampling Rate
-4	Anemometer	Cont	1 sample / 15 min	1 sample / 15 min
2	S4	Cont	2 min avg / 10 min	1 min avg / 5 min
4	S4	Burst	9, 2 min avg / hr	18, 1 min avg / hr
6	S4	Cont	2 min avg / 10 min	1 min avg / 5 min
9	S4	Burst	9, 2 min avg / hr	18, 1 min avg / hr
12	S4	Cont	2 min avg / 10 min	1 min avg / 5 min
15 - 190	Cyclesonde	Cont	~ 1 sample / 3 hr	$\sim 1 \text{ sample} / 3 \text{ hr}$
230	Anderaa	Cont	1 sample / 10 min	1 sample / 10 min
270	Anderaa	Cont	1 sample / 10 min	1 sample / 10 min
310	Anderaa	Cont	1 sample / 10 min	1 sample / 10 min
350	Anderaa	Cont	(not used)	1 sample / 10 min
390	Anderaa	Cont	(not used)	1 sample / 10 min

 Table 3.1:
 Kn88/Kn89
 Instrument
 Sampling
 Strategies

As a significant objective of these experiments was to investigate the estuarine component of the low frequency residual, the experiments took place in the spring and summer during the melt water freshet. Runoff in Knight inlet is principally controlled by the Klinaklini River and can be expected to double when either heavy rainfall or periods of intense hot weather persist and significantly speed snow melting at higher altitudes. Fresh water runoff was obtained through stream gauges routinely deployed and maintained by Inland Waters of Environment Canada and provided as daily mean discharges in m³/s.

Finally, surveys using a Guildline Model 8705 CTD were done at the start and end of each experiment. In addition periodic samples of water were taken with a General Oceanics rosette both to confirm the CTD readings of Salinity with a Guildline AutoSal laboratory salinometer and to establish dissolved oxygen levels via Winkler titration. These deployment and pickup cruise surveys provided both the initial and final water property distribution and were used as an intermediate standard to perform insitu calibration and inter-calibration of the conductivity and temperature sensors of the other instrumentation. Part way through each experiment, additional CTD surveys from small vessels were

performed as an additional check on instrument performance. These surveys took place over a number of days and in 1989 were only available to 200 metres.

Figure 3.1 shows the oceanographic stations used in the CTD surveys and the location of the moorings and associated instrumentation used in the Knight 1988 and 1989 experiments distributed along the inlet centre line profile. The time base of all data presented is expressed in decimal Julian Days, where day 1.0 is midnight December 31st, pacific standard time of year previous to the experiment. This definition makes Julian Day 1.5 equivalent to 12:00 a.m. pst January 1st of the year in which the experiment started and a convenient sequential time base for displaying the observations. The reader is cautioned that other definitions of the Julian Day exist, and this particular time base was adopted as a convenience for processing and analysis. While wind was recorded using the standard meteorological convention (that is the direction from which the wind is blowing) all wind data presented in this thesis will be rotated to have a directional orientation consistent with the currents (that is the direction towards which the current is flowing).



Figure 3.1: Knight Inlet Profile showing Instrument Moorings

3.3 Kn88 - Knight Inlet, Spring 1988

The Kn88 experiment took place in the spring from March 22 (Julian Day 82) through April 26 (Julian Day 117) when the inlet was in a low runoff condition ($\sim 50 \text{ m}^3/\text{s}$) to moderate runoff caused by the beginning of a freshet due spring melting ($\sim 180 \text{ m}^3/\text{s}$). The average runoff throughout the 35 day experiment was 98 m³/s.

At both Protection in 1988 and at Tomakstum in 1989, an examination of the wind records indicate that the steep sides of the inlet essentially confine the winds to the along channel (either up or down) direction. The anemometer direction sensor failed at the Tomakstum mooring and therefore the vector wind record was reconstructed using the speed from Tomakstum and the direction from Protection as detailed in section 3.5.5. Weather was dominated by a succession of storm fronts with their associated winds peaking at about 10 m/s with a 2.5 to 3 day period and with a trend towards a net outflow during the first half of the experiment. The meteorological regime then settled into a clear sunny regime with a dominant high pressure starting on Julian Day 101 continuing during the second half with a fairly steady up inlet wind with a speed of 5 m/s. The wind fields show a small down inlet mean wind at Protection and a small up inlet wind at Tomakstum in the record average. Figure 3.2 shows the composite time series of the river runoff and the hourly averaged along channel wind as measured at Protection and as estimated (according to the procedure outlined in section 3.5.5) at Tomakstum during the Kn88 experiment.

Three CTD surveys were completed for the Kn88 experiment; a deployment survey at the start of the experiment on Julian Day 82, a series of mid-experiment surveys near Julian Day 99, and a pickup survey on Julian Day 115. Unfortunately the CTD data logger malfunctioned during the pickup cruise and those data were not recoverable. Considerable



Figure 3.2: Kn88 River Runoff and Along Channel Wind for Protection and Tomakstum

noise from internal tides existed in the CTD records obtained near the sill, but not near the head.

The results of the first two surveys were used to compute contours of density that provide snapshots of the initial and mid-experiment oceanographic conditions in the inlet. Contours of σ_t computed for the Kn88 deployment cruise and mid-experiment cruise CTD data are shown in Figure 3.3. Some deep water renewal did indeed take place in the spring of 1988 shown as the rise of the 24.0 and 24.1 σ_t isopleths. The availability of water of sufficient density to penetrate to these levels can be seen just outside the sill in the mid-experiment survey, with sufficient density to penetrate into the interior of the basin. With 24.3 source water outside the sill but not quite at sill depth by the middle of the experiment. Stronger flows due to spring tides may complete the energy requirements required for penetration.

While CTD data were not available from the pickup cruise, salinity samples were obtained at standard depths during both the deployment and pickup cruises. These salinity data are compared with that of the mid-experiment CTD survey in Table 3.2. The increase of salinity with time is consistent with renewal continuing through to the end of the experiment. Dissolved oxygen levels obtained by Winkler titration during the deployment cruise were typical for Knight Inlet, with 3.26 ml/l at 500 m in the deepest part of the basin. These values are consistent with those reported by Pickard(1961). Dissolved oxygen in the deep water at the mouth was 5.10 ml/l.

The deployment cruise surface salinities were 21.03 psu at station 11 near the head, 28.83 at station 5 near Tomakstum Island, and 30.04 at station 3 near Protection Point. In this situation, Knight Inlet appears to fit the classical estuarine case with the steady entrainment of salt water as surface water proceeds to the mouth.



1988 Deployment and Mid-Experiment Ctd Surveys, Contours of Density (as στ) CTD Station #

Figure 3.3: Kn88 Experiment CTD Surveys, Contours of Density (as σ_t) Density as σ_t calculated from temperature, conductivity, and depth plotted as along channel contours. Solid black contours are the deployment cruise isopleths of σ_t , dotted are the mid-experiment cruise isopleths.

Depth (metres)	Deployment psu	Mid-Experiment psu	Pick Up psu
50	30.60	30.69	30.78
75	30.65	30.78	30.88
100	30.75	30.84	31.02
150		30.97	31.18
200	31.12	31.19	31.34
300	31.31		31.39
400	31.35		31.38
500	31.36		31.38

Table 3.2:

e 3.2: Basin salinities at Station 7 from the Kn88 Experiment

Values for the deployment and pickup cruises were obtained by lab salinometer measurements from salinity samples taken at standard depths. Values for the mid-experiment were taken from the CTD data at the same depths.

Data records from the moored instruments were retrieved intact, with the exception of the direction sensor of the Tomakstum anemometer mentioned previously and the conductivity sensor on the Tomakstum cyclesonde. Unfortunately the S4 current meter deployed at 12 m at this location also lacked a conductivity cell and thus an unfortunate 'blind spot' exists at Tomakstum in the density data records from 12 to 190 m.

The Knight 1989 experiment took place during the early summer from June 19 (Julian Day 170) through July 25 (Julian Day 206) when the inlet was in a high runoff condition from about 400 m³/s to 750 m³/s. Average runoff for the period of the 36 day experiment was 569 m³/s and the time series shows that the river can very nearly double its discharge and then drop back within a period of about 5 days. Figure 3.4 shows a composite time series of the river runoff and hourly averaged wind as estimated at Protection (according to the procedure outlined in section 3.5.5) and as measured at Tomakstum for the Kn89 experiment.

The anemometer at Protection failed during Kn89 due to a faulty power cable; and as a result the along channel component of the Protection wind was reconstructed from the Tomakstum record based on the relationship of wind speed at the two locations obtained from the Kn88 experiment and as detailed in section 3.5.5. Except for a brief sunny period from Julian Day 174 through 178, weather at the beginning of the experiment was warm but overcast . The sunny period is characterised by a switch to a net up inlet wind with a velocity of about 5 m/s and the overcast periods by weaker outflow winds. By Julian Day 188 clear skies again predominated through Julian Day 196 and the river runoff responds accordingly. At the end of the experiment, weather again cooled and the river runoff decreased. Record averages for both the Tomakstum and estimated Protection along channel winds show a net up inlet wind during the experiment. The large variations in runoff indicate that at least in the early summer, runoff response to changes in the weather can be quite rapid as the amount of insolation affects the snow melt. Up inlet winds were generally stronger than in 1988 indicating the relative strength of the barometric pressure differences driving the summer winds.



Figure 3.4: Kn89 River Runoff and Wind for Protection (Reconstructed) and Tomakstum

Densities computed as ot from the 1989 deployment and pickup cruise CTD surveys are shown in Figure 3.5. Deep inner basin densities are slightly lower than in 1988, suggesting that deep water renewal is taking place later in 1989 or that perhaps vertical eddy diffusion was greater during the previous winter than the 1987/88 winter. Farmer and Freeland(1983) give an example of how strong winter storms and weak stratification can cause deep mixing and thus vertical eddy diffusion may vary from year to year. Water of a density sufficient to completely renew the deep water is available at the mouth, at or near sill level. The pickup cruise CTD data shown in Figure 3.5 show a rise in the 24.4 isopleth indicating renewal occurred during the 1989 experiment. CTD survey data was sparser from the mid-experiment cruise in Kn89 and due to instrument limitations was only available to 200 m. As a result, these data were used only for inter-calibration of the moored instruments temperature and conductivity sensors. Oxygen levels in the deep basin were slightly higher in 1989 at 3.33 ml/l. Oxygen levels in the renewal source water were similar to 1988 at 5.10 ml/l.

Surface water is extremely fresh in 1989, with a surface salinity of 0.29 at station 11 near the head, 3.38 at station 5, and only 8.17 psu at station 3 near Protection Point during the deployment cruise. These extremely low surface salinities are the result of the high runoff. While the river discharge has increased, salt entrainment has actually been reduced and salinity is still extremely low at the mouth. Increased stability will also limit the effects of wind induced mixing. The strong pycnocline existing along the inlet may effectively decouple the wind from the lower layers, as suggested in Buckley and Pond(1976).

Due to instrument malfunctions, data records were incomplete for the Protection and Sill cyclesondes for Kn89 with only about 2 weeks of data retrieved from each. In addition, the Tomakstum cyclesonde skipped a number of profiles at the beginning and the end of the deployment.



Figure 3.5: **Kn89 Experiment CTD Surveys, Contours of Density (as** σ_t) Density as σ_t calculated from temperature, conductivity, and depth plotted as along channel contours. Solid black contours are the deployment cruise isopleths of σ_t , dotted contours are the pickup cruise isopleths.

3.5 Data Processing

The data records retrieved from the instruments were at various sampling rates and were recorded in a variety of proprietary formats particular to each instrument type. Further, most of the instruments recorded data as time series at a particular depth whereas the cyclesondes recorded data as a time series of profiles at irregular depths depending on their speed of ascent or descent. In order to render all of the observations into a consistent data base suitable for statistical analysis the following processing was performed.

3.5.1 Instrument Calibration

Pre-experiment lab calibrations provided initial processing coefficients for temperature and conductivity records for all instruments, including the CTD, the cyclesondes, and the Anderaa RCM instruments. Previous experience with these instruments has shown that unless major repairs have been done to the sensors or circuitry of these instruments, their calibrations are stable for an extended period of time. However, experience with S4 current meters was limited and so the calibration coefficients for these instruments were based on a set of extensive pre and post calibrations. In addition, the velocity zeros of these instruments is sensed relative to local magnetic north. The calibration routines therefore rotate all directions to be relative to true north by adding the current local value of the magnetic deviation (24 degrees for Knight Inlet).

Further adjustments were made to the temperature and conductivity records of the instruments according to inter-calibration procedures discussed in section 3.5.6. Briefly these adjustments consisted of adjusting the conductivity records of the instruments so that

they agreed with the CTD measurements made during the deployment, mid-experiment, and pickup cruises and then ensuring that the calculated density increased with instrument depth.

3.5.2 CTD Bin Averaging

The Guildline Model 8705 CTD logs conductivity, temperature, and pressure data as the instrument is lowered and then raised again through the water column. Data is collected into bins with a 10 cm vertical resolution and averaged to improve the estimate of the insitu value. During the cast, extraneous data are rejected. The bin averaged data are then decimated to provide an estimate for each 1 m of depth to 50 m and each 5 m of depth thereafter by taking the binned value with the depth desired. While data from both up casts and down casts are available, down cast data were preferentially used as the up cast was frequently halted to trigger bottle samples at standard depths. Inlet CTD surveys were performed at the beginning, during, and at the end of each experiment. These data provided the initial and final density fields of each experiment and were also used to inter-calibrate the less accurate temperature and conductivity sensors incorporated in the current meters. The inter-calibration will be discussed in section 3.5.6.

3.5.3 Cyclesonde Data Outlier Editing

Cyclesonde data records are stored as a serial synchronous data stream on magnetic tape. In the process of reading these tapes, data dropouts occasionally occur leading to either an incomplete data record, missing data records or both. As a result, all records are scanned and repaired as necessary. Where the number of bits missing is small, the data record can often be recovered by shifting the data back into data field alignment. Where records do not conform at all to the specified data format, the data records are deleted and the frame

numbers of sequential records checked to ensure that they increment with time. This procedure ensures that garbled data records do not contaminate the final time series and that the time base of the observations can be accurately recovered.

Data records are then converted to engineering units by applying polynomial calibration coefficients to the integer data fields. The cyclesonde's vertical profiling speed as determined from the change in pressure with depth is removed from the resulting speed by vector subtraction. These converted files were then scanned for anomalous values detected by looking for values inconsistent with the physical oceanography of fjords. For example, scalar data were compared with a fit to a local T/S curves and anomalies (usually in conductivity) were interpolated out after individual examination. Velocity data were obtained by averaging the port and starboard rotor speeds whenever their rotor counts were within 5 cm/s, otherwise the lowest value was taken on the assumption that an overspeed caused by contact bounce from the reed switch contacts in the rotor sensors had occurred. This appears to be the case in only a few percent of the records. Where excessively low speeds were found, values interpolated from the previous and next readings were substituted on the assumption that the rotor had stalled.

Incomplete profiles at the start and the end of each experiment were trimmed from the final data records. Conductivity data near the beginning of each set of observations were also trimmed when the cell showed signs of an exponential settling to a consistent value in the deep water. This behaviour is believed due to air being trapped in the conductivity cell during deployment and preventing accurate readings from being obtained until it diffuses away in a period with the order of one to two days.

3.5.4 Cyclesonde Time Series Interpolation

As a cyclesonde profiles at approximately 5 to 10 m/min and the sampling rate during profiling was set at 1 min, profile data were then interpolated to standard depths with 10 m vertical separation. A cyclesonde measures speed by counting rotor revolutions between samples and therefore represents the average velocity magnitude over the part of the water column through which the instrument has travelled. Direction is sampled with scalar data (conductivity, temperature, and pressure) by clamping a magnetic compass needle to a potentiometer. Calibration software rotates all directions from relative to local magnetic north to true north. The rectangular components of velocity were determined using:

$$u = \frac{V_n(\sin\theta_n + \sin\theta_{n-1})}{2}$$
$$v = \frac{V_n(\cos\theta_n + \cos\theta_{n-1})}{2}$$

where:

u is the east-west component with positive flow from east to west v is the north-south component with positive flow from south to north V_n is the speed obtained from the rotor counts at sample n θ_n is the true direction recorded as a spot direction recorded at sample n

These velocity components are then sorted into bins at fixed 10 metre depth intervals, and data in each bin averaged. Time of the velocity time series is adjusted to reflect the averaging process in each bin. Scalar data were sorted by depth but not averaged. To provide data at a fixed depths, a linear interpolation is then applied between data points to provide a data point at the requested depth.

To provide smoother profiles near the surface where changes with depth are larger, this procedure was repeated twice, interpolating at 10 metre intervals starting at 20 metres to the bottom and starting at 15 metres and proceeding to 35 m. The data from the interpolated profiles were then sorted by depth into a set of times series spanning all profiles. The

sampling interval of these time series was approximately 3 hours, but timing differences due to the asymmetry in the instruments rise and fall rates are present. This procedure produced a data set from the cyclesondes that was consistent in structure with the time series produced from the other current meters and allowing a single processing stream to be used in later analysis.

3.5.5 Current Meter and Anemometer Median Averaging

S4 current meters vector average all velocity data internally based on samples taken every 0.5 seconds. As mentioned in section 3.2, two of the instruments on each mooring at 4 and 9 m depth were configured to 'burst' mode and recorded 9 two minute velocity averages centred on the hour in 1988 and 18 one minute averages every hour centred on the hour in 1989. The rest of the instruments were set for 'continuous' mode and recorded a 2 minute average every 10 minutes in 1988 and a 1 minute average every 5 minutes in 1989. The burst mode instruments were used to investigate possible aliasing of high frequency energy into the other data records, but although a few spikes that may be associated with the high frequency phenomena can be seen in the record they do not contribute significantly to the total energy of the record.

Anderaa current meters record a rotor count that represents an average speed between two sampling periods. Direction is a spot reading recorded at each sampling interval. Therefore the same method of estimating the rectangular components was applied to these instruments as the cyclesondes (section 3.5.4) before data were subject to further processing.

In order to reduce the raw time series to hourly estimates, velocity data for one hour centred around the hour were binned, the higher and lower values removed and the remaining data points averaged. For example, the averaging bin for the 'continuous' mode data sampled at

10 minute intervals in 1988 included three samples before the hour, the sample on the hour and three samples after the hour for a total of seven samples. The two highest and two lowest values were rejected and the remaining three samples averaged. For the 'continuous' mode data sampled at 5 minute intervals 1989, a total of thirteen samples were binned, the four highest and lowest rejected and the remaining five samples were averaged. Burst mode data was treated in a similar manner with the averaging window and rejection count altered appropriately. This procedure effectively removed any contributions from higher frequency energy from the resulting time series.

Spikes associated with high frequency energy were more noticeable in the unaveraged spot readings scalar data of the S4 current meters, but still represented relatively small contributions to the record. To reduce the scalar time series of the S4 and Anderra current meters to hourly values the averaging process discussed above was applied. The time values of the vector velocity and scalar data sets were then adjusted to account for any shifts due to instrument timing errors based on the correct time at the beginning and the end of the record, although the worst drift was slightly less than a minute.

Anemometers were mounted on the Geodyne surface buoy of the S4 current meter moorings at Protection and Tomakstum in both years. The anemometers record a rotor count that represents an average speed between two sampling periods and a direction that is a spot reading in a manner similar to the Anderaa current meters. Therefore the same method of estimating the rectangular components was applied to these instruments as the cyclesondes and Anderaa current meters (section 3.5.4) before data were subject to further processing. The sampling interval of these instruments was 15 min.

The raw wind records show a considerable fluctuating cross channel component. The direction of the wind is taken as a spot reading and may not represent an average wind

direction. Therefore at least some of the cross channel component is likely due to buoy motion induced by wind waves and the dampening characteristics of the instrument's directional vane. In order to reduce this directional noise, the raw wind data were subjected to the same averaging processing used for the current meters, with the averaging window expanded to two hours (9 data points) and the three highest and lowest rejected before averaging.

As mentioned in sections 3.3 and 3.4, the direction sensor failed at Tomakstum in 1988 and the instrument at Protection totally failed in 1989 due to a faulty power cable. At both Protection in 1988 and Tomakstum in 1989 the average hourly magnitude of the along channel component was close to (within 10% at Protection 1988 and 7% at Tomakstum 1989) and well correlated with the hourly average speed (0.98 in both cases) indicating that the wind field was essentially confined along channel. Therefore the Tomakstum 1988 wind record was reconstructed by using the hourly average speed from the Tomakstum and the hourly average direction from the Protection wind records. In addition, the speed records from both anemometers in 1988 were separated into up and down channel groups based on the along channel sense of direction at Protection. The correlation between the speeds was reasonable (0.74 for the up channel and 0.77 for the down channel) and the down inlet winds were twice as strong at Protection as at Tomakstum, while up inlet winds were just slightly stronger (1.1 x) at Protection. This relationship was then used to estimate the along channel component of the wind at Protection from the along channel component of the hourly averaged wind record at Tomakstum for 1989. If one was to assume that the wind was confined completely along channel and that any cross channel component was due strictly to residual vane flop caused by buoy motion, this method of estimation of the along channel components might be in error by at most the 10% difference noted. However, the examination of the actual topography of Knight Inlet reveals the opening of Tribune channel in the proximity of the Protection mooring and the lowering of the

sidewalls at Glendale Cove near Tomakstum. Therefore the difference between the magnitude of the along channel component and the speed is likely some combination of directional 'noise' and true cross channel variability.

Time series from the anemometers, S4s, cyclesondes, and Anderaas were then merged into separate vector velocity(uv) and property scalar(ts) time series data bases for each station. Because of the averaging the time for the velocity and scalar samples differ.

3.5.6 Inter-calibration

Velocities records were compared as a final check on instrument operation. The 12 metre S4 and 230 metre Anderaa current meters agreed well with the 15 metre and 190 metre interpolated cyclesonde records, respectively. However conductivity sensor drift and to some extent temperature sensor shifts required that inter-calibration be performed on the scalar data records from the current meters.

Inter-calibration of temperature and conductivity records used the in situ values provided by the CTD surveys as a secondary calibration standard. The goal of this procedure was to ensure that the temperature and conductivity records from the current meters matched the observed density field at the beginning and end of each experiment and were consistent with density increasing with depth. A check on the CTD accuracy was made by comparing the salinities with those obtained by lab salinometer from water samples taken at standard depths. The mean difference was slightly less than 0.01 psu. The CTD casts nearest in time to the beginning and end of each instruments data record were then used for inter-calibration procedure listed below:

1. Temperature records were compared to the CTD and adjusted if necessary.

- 2. Salinity records were compared to the CTD and an equivalent conductivity offset computed from the salinity difference
- 3. An offset and trend in temperature (if necessary) and conductivity was applied continuously along each instruments calibrated data record.
- 4. The corresponding salinity and density from the corrected temperature and conductivity were recomputed.

Adjustments were applied to the calibrated but unprocessed data records so that the unaveraged/uninterpolated data would benefit from this procedure for future use. All data records were then reprocessed using the Median Averaging technique described in section 3.5.5 for S4 and Anderaa records and the interpolation described in section 3.5.4 for cyclesonde profiles. As a final check on the inter-calibration, composite profiles were extracted from the adjusted data records and plotted with the nearest CTD profiles to ensure that the adjustments had been applied correctly.

S4 temperature records required little adjustment with only three instruments requiring adjustment in 1988 and one in 1989. More of the conductivity records were adjusted as it was found that the conductivity electrodes appeared to shift characteristics if mishandled. In 1988 two conductivity records were shifted based on CTD data. However CTD data from the surface layer are noisy and the large gradients involved make the exact determination of the final adjustment difficult. As a result the data records of three others had to be shifted in order to ensure that the resulting density increased with the depth of the instrument. In 1989 only one such conductivity shift was necessary due to the more careful handling of the instruments, whose electrodes have a soft titanium oxide coating.

Cyclesonde temperature records were checked and as is normal for these instruments did not require adjustment to agree with the CTD within $\pm 0.02^{\circ}$ C. In 1988, one cyclesonde

required no adjustment in conductivity, one had a failed conductivity sensor, and one required a small adjustment. In 1989 three of the four cyclesondes required adjustment to their conductivity records.

All Anderaa current meters were adjusted to the CTD in both years. In 1988 some of the instruments had a temperature sensor with limited resolution and hence some of these records are of limited value. In 1989 the Anderaa current meters at Axe point showed a trend towards slightly lower salinities in the middle of the record with a subsequent rise at the end. This characteristic was present in three of the instruments spaced between 230 and 310 metres. The Axe cyclesonde record did not show any corresponding change at 190 metres and it is assumed, because Axe Point is near the head of Knight Inlet, that these mid record lower salinities must be an artifact of sensor fouling perhaps caused by the rain of glacial till input by the river.

Figure 3.6 shows the data flow for the preliminary processing of the data acquired by the instruments deployed during the experiments.



Figure 3.6: Instrument Preliminary Processing Data Flow

3.6 Processed Time Series and Spectra

The time series produced by the preliminary processing were used in the analysis presented in Chapter 4. Because data are available over the entire water column only at the Protection and Tomakstum sites, the data from these stations were the focus of this work and a representative subset are presented here to describe their general character. It is these records that are truly unique in that they include wind and near surface velocity and scalar values from the S4 current meters as well as throughout the rest of the water column. A complete presentation of the data from all stations may be found in the data report for the Kn88 and Kn89 experiments (currently in preparation), available from the Department of Oceanography at the University of British Columbia.

Figures 3.7 and 3.8 present the time series of the along channel velocity from the Protection and Tomakstum moorings of the Kn88 experiment for selected depths. Various temporal scales can be seen in the processed time series, including diurnal and semi-diurnal tidal components. However, in the the along channel velocity records of the surface layer (2 to 12 m) it can be seen that a component with an approximately 3 day period contributes significantly to the variance. When visually compared with the wind record, the major features are similar. This is the influence of the wind on the surface layer and it decreases rapidly with depth. Deeper in the water column the tidal periods dominate the variance and the spring/neap tidal cycle can be clearly seen.

Figures 3.9 and 3.10 present the density (as o_t) time series from the Protection and Tomakstum moorings. Tidal fluctuations can be seen in the surface layer along with a response to the wind characterised by increased densities coincident with up inlet winds. This is due to the wind reversing or slowing the flow of the surface layer and preventing the outward flow of fresher water. Down inlet winds restore the lower densities as the


Kn88 Protection Wind and U Velocities (cm/s) at Selected Depths

Figure 3.7: Kn88 Protection Processed Along Channel Velocity Time Series



Kn88 Tomakstum Wind and U Velocities (cm/s) at Selected Depths

Figure 3.8: Kn88 Tomakstum Processed Along Channel Velocity Time Series



Kn88 Protection Wind(cm/s) & Sigma-t(kg/m^3) at Selected Depths

Figure 3.9: Kn88 Protection Processed Density (σ_t) Time Series Note that in the above composite plot, σ_t increases upward and each division is 5 kg/m³. The range for each record is from 20 to 25 kg/m³.



Kn88 Tomakstum Wind(cm/s) & Sigma-t(kg/m^3) at Selected Depths

Figure 3.10: Kn88 Tomakstum Processed Density (σ_t) Time Series Note that in the above composite plot, σ_t increases upward and each division is 5 kg/m³. The range for each record is from 20 to 25 kg/m³ except for the 2 m record which is 15 to 25 kg/m³.

surface outflow is re-established. Response in the density field is more pronounced at Tomakstum than Protection, because the vertical density gradient in the surface layer is stronger. The surface layer density exhibits an overall trend towards lower values with time at both stations consistent with the increase in runoff.

The raw spectra computed for the wind, 2 m and 12 m (9 m for Tomakstum 1988 as the 12 m instrument stopped recording 5 days before the end of the experiment) detrended and demeaned velocity time series are presented in Figures 3.11 and 3.12. 'Raw spectra' have not had any band averaging applied. Note that these spectra are plotted in power preserving form so that the area under the curve is proportional to the total variance. The energy in the 2 m spectrum at Tomakstum is dominated by time scales of about 3 days at the same frequencies as the wind energy. The spectrum is consistent with observations by Farmer(1972) and others. At Protection significant energy also exists at these time scales, but the ratio of the current response to the wind is lower (Note that the scales differ as the tidal peaks are higher at Protection). The stronger response at Tomakstum is probably due to the stronger vertical density gradient at this station confining the wind driven response closer to the surface. The 12 m spectrum at Protection and the 9 m spectrum at Tomakstum still show some energy at these temporal scales, but now the semidiurnal tides dominate. Diurnal energy is present but it may not entirely be tidal as sea/land breezes also contribute energy in this temporal scale. At Tomakstum peaks can also be seen at a period of 0.25 days corresponding to the M_4 constituent. Its relative absence at Protection is probably due the fact that it is enhanced by the non-linear interactions in the region of the sill, between these two moorings.



Kn88: Tomakstum Raw Spectra



Kn88 Tomakstum Raw Spectra Raw spectra have not had any band averaging applied. Note that these spectra are plotted in power preserving form so that the area under the curve is proportional to the total variance.



Kn88: Protection Raw Spectra

Figure 3.12:

Kn88 Protection Raw Spectra

Raw spectra have not had any band averaging applied. Note that these spectra are plotted in power preserving form so that the area under the curve is proportional to the total variance. Selected along channel wind and velocity time series for the Kn89 experiment are shown in Figures 3.13 and 3.14. The character of the wind record is considerably different in 1989, with a prominent fortnightly component instead of the 2 to 3 day disturbances seen in 1988. Wind influence can be clearly be seen in the surface layer along with a diurnal and semidiurnal tides. The cyclesonde at Protection stopped profiling approximately half way through the Kn89 experiment resulting in the loss of data at 15 m and below for the second half of the experiment. At Tomakstum, the cyclesonde skipped profiles intermittently between Julian Day 173 and 180 and again after Julian Day 193.

The corresponding density time series are shown in Figures 3.15 and 3.16. Lower densities in the surface layer are observed at both Protection and Tomakstum than in 1988, due to the higher river discharge during the Kn89 experiment. The stronger density gradient in the surface layer results in a strong wind response in the density field at both stations. Here the up inlet/down inlet wind regime is of a longer period than in 1988 and the switch to an up inlet wind regime brings a corresponding raising of surface layer density as in 1988. During down inlet winds, the density suddenly lowers and then partially recovers after 2 to 3 days (eg: between Julian Day 179 through 185). If the initial lowering of the surface layer density is caused by the outflow driven by the combined forcing of runoff and the pressure gradient built by the previous up inlet winds, this recovery may be explained by the release of the stored 'fresh' water in the head of the inlet. Once this pool of 'fresh' water has been released a thinner surface layer is reestablished.

The raw spectra of the wind and surface currents are shown in Figures 3.17 and 3.18. It is to be noted that the 2m semi-diurnal tidal energy at Protection is twice that of 1988. This change in energy is probably due to a change in the modal composition of the internal tidal energy and not a change in the barotropic tidal response which is primarily controlled by topography. A larger difference is seen in the 2m spectra for Tomakstum. The 1989



Kn89 Protection Wind and U Velocities (cm/s) at Selected Depths

Figure 3.13: Kn89 Protection Processed Along Channel Velocity Time Series



Kn89 Tomakstum Wind and U Velocities (cm/s) at Selected Depths

Figure 3.14: Kn89 Tomakstum Processed Along Channel Velocity Time Series



Kn89 Protection Wind(cm/s) & Sigma-t(kg/m^3) at Selected Depths

Figure 3.15: **Kn89 Protection Processed Density** (σ_t) **Time Series** Note that in the above composite plot, σ_t increases upward and each division is 5 kg/m³. The range for each record is from 20 to 25 kg/m³ except for the 2 m record which is 5 to 20 kg/m³ and the 6 m record which is 10 to 25 kg/m³.



Kn89 Tomakstum Wind(cm/s) & Sigma-t(kg/m^3) at Selected Depths

Figure 3.16: **Kn89 Tomakstum Processed Density** (σ_t) **Time Series** Note that in the above composite plot, σ_t increases upward and each division is 5 kg/m³. The range for each record is from 20 to 25 kg/m³ except for the 2 m record which is 5 to 20 kg/m³ and the 6 m record which is 10 to 20 kg/m³.

semidiurnal energy at 2m is approximately 7.5 times that of 1988. However this peak is considerably lower at 12 m suggesting this change is also due to the modal composition of the internal tide. The examination of the tidal results is not included in this thesis as the focus is on the lower frequency residuals. Webb and Pond(1986b) have examined the tides in Knight inlet before and shown that changes in the modal composition do occur. As the data from these experiments is more complete (in that they include the near surface) the tidal results will be examined and the results reported elsewhere.

The difference in the character of the wind is clearly seen in the wind spectra with the bulk of the energy at the temporal scales of approximately 15 days as previously noted and at 1 day. The energy at this latter time scale appears to be due to the sea/land breeze regime set up during periods of better weather. The wind forcing contributes approximately the same amount to the total energy of the 2 m velocity in both years. An integration over the wind band from the 1 cycle per month to 1 cycle per day gives a ratio of 2 m current speed to wind speed of approximately 3.8 and 4.1 percent at Protection and 6.1 and 5.5 percent at Tomakstum for the 1988 and 1989 experiments respectively. This is in reasonable agreement with Burling(1982) who gives the wind driven current as approximately the wind speed/30. At 12 m the wind influence appears to be negligible at least at the longer periods and the semidiurnal tides dominate.



Kn89: Protection Raw Spectra



these spectra are plotted in power preserving form so that the area under the curve is proportional to the total variance.



Kn89: Tomakstum Raw Spectra



18: Kn89 Tomakstum Raw Spectra

Raw spectra have not had any band averaging applied. Note that these spectra are plotted in power preserving form so that the area under the curve is proportional to the total variance.

Chapter 4 Analysis and Discussion

4.1 Data Analysis Methodology

As seen in the spectra presented in section 3.6, it appears that wind energy dominates the forcing in the surface layer while the semi-diurnal tides dominate over the rest of the water column. The analysis presented in this chapter examines the major components of the low frequency (sub-diurnal) response of an inlet by first examining the monthly mean (29.5 day) response and then examining the records using multivariate statistical methods. The contributions to the variance that could be statistically attributed to known forcing such as the tides and the wind were then sequentially removed and the residual response examined.

An analysis of the monthly mean response is presented in section 4.2. This method was used by Webb(1985) to examine cyclesonde data as mean profiles from 20 to 190 m for July and September 1983 in Knight Inlet. The mean of each time series was calculated over 29.5 days as this length is roughly equal to that of the longest tidal constituent (M_m) that can be resolved in the time series. It assumes that the wind forcing in this average is also zero which, as discussed in sections 3.3 and 3.4, is approximately true for both experiments. The use of this method allows direct comparison with the vertical structures of Webb's mean profiles (presented in figure 2.5) and has been expanded to include density (as σ_l) as well as along channel currents and near surface data to 2 m. The structures were examined for consistency with estuarine circulation forced by river runoff and deep water renewal evidenced in the raising of along channel isopycnals with time during each experiment (figures 3.3 and 3.5).

The contributions to the total record energy from the diurnal and higher frequency tidal constituents seen in the raw spectra (presented in figures 3.11, 3.12, 3.17 and 3.18) are significant and dominate much of the water column. In order to best determine the influence of other forcing, the high energy diurnal and higher frequency tides were removed. Detiding by harmonic analysis is presented in section 4.3 and was accomplished by fitting the well known astronomical forcing frequencies of the tides, in the least squares sense, to the record variations and determining an amplitude and relative phase for each constituent. The tidal contributions from the diurnal and higher forcing frequencies were then calculated from these parameters and subtracted from the original record to give the detided residual. Harmonic analysis is a standard method of tidal analysis and prediction. Godin(1972) is an excellent general reference on tidal analysis. Section 4.4 discusses the detided response.

Anemometer records from the Protection and Tomakstum moorings allowed the detided residual records to be dewinded by cross spectral techniques and the method and discussion of results is presented in section 4.5. The contributions of the wind to the low frequency circulation occurs over a wide range of frequencies beginning with diurnal sea/land breezes, to forcing of periods of 2 to 3 days representing the passage of storm fronts, to lower frequency near fortnightly periods representing changes in the prevailing weather. These contributions are significant and Pickard and Rodgers(1959) reported that the wind forcing was sufficient to reverse the direction of flow in the surface layer.

Cross spectra of the wind and along channel velocity and the wind and density records were used to estimate the wind influence by converting the cospectra and quadrature spectra into a spectral coherence and phase. The coherence squared represents, for a linear system, the contribution to the variance of one signal (density/current) from the other (wind) as a

function of discrete frequency bands. The coherence was used to estimate the wind driven response in the along channel currents and density records. Farmer(1972) successfully used this method on surface layer data from Alberni Inlet. A discussion of the wind driven response is presented in section 4.6. Having estimated the influence of the wind throughout the water column, the dewinded residual was then determined by subtracting the wind driven response from the detided time series. A discussion of the dewinded residual response is presented in section 4.7.

Observational time series are only approximately stationary and continuous. For example, the amplitude and phase of a tidal constituent at a given depth is not fixed (Stacey 1985) and varies with the modal composition, in turn a function of the stratification and which changes with time. Therefore the statistical methods employed such as harmonic analysis which calculate fixed amplitude and phase parameters that best describe the response over the whole record are inexact and lead to a small amount of the higher frequency 'noise' being left in the residual. Therefore the detided and dewinded time series are presented after application of a post processing low pass spectral filter with a cutoff equal to the period of the O_1 tidal constituent (0.93 cpd).

Figure 4.1 shows the processing data flow used in the analysis of the low frequency residual presented in this chapter, with data flowing from the hourly averaged time series on the left to the monthly mean, detided, wind driven, and dewinded time series on the right. The ovals represent each software routine that operated on the data in turn as it moves from left to right in the diagram.

Residual Analysis Data Flow Diagram



Figure 4.1: Residual Analysis Data Flow Diagram

Data flows from the hourly averaged time series on the left to the monthly mean, detided, wind driven, and dewinded time series on the right. The ovals represent the software routines that operated on the data as it moves from left to right. Note that a low pass spectral filter (fc = 0.93 cpd) was applied to all processed data before final presentation.

4.2 29.5 Day Mean Response

Figure 4.2 shows the 29.5 day mean along channel velocity and density (as σ_t) profiles for the 1988 experiment. Surface layer data at Protection and Tomakstum show a net outflow with considerable vertical shear. At Tomakstum an inflow just below the surface layer, at 12 to 15 m, may be interpreted as at least partial salt replacement forced by entrainment in the surface outflow. The CTD surveys for 1988 showed evidence of deep water renewal taking place in the outer basin and the inner basin at mid depths. The mean velocities corresponding to a net inflow in the lower third of the water column at Protection (below 110 m) are consistent with the renewal process in the outer basin. An inflow at Tomakstum from sill depth to 180 m suggests renewal at mid-depths in the inner basin, with return flows above and beneath the penetration to conserve volume. At the sill mooring (just inside the inner sill), this penetration can also be seen, although spread through more of the water column. It is possible that more vigorous mixing due to sill dynamics distributes the momentum vertically in this area. At Protection the mid-depths through to the surface all have outflow, likely due to the fact that in the outer basin the renewal water travels inward near the bottom and therefore all volume compensation must be above the penetration. However, the velocity minimum at 35 m may be a manifestation of the salt replacement inflow maximum. These structures correspond well to the concept of nested thermohaline circulation presented in Chapter 2.0, with estuarine forcing the surface layer and deep water renewal the lower part of the water column. However, as noted in the theoretical discussion, linear superposition may be somewhat simplistic. There is also a small local minimum at 12 m at Protection. This is within instrument error, although the larger value from the cyclesonde at 15 m when compared to the 12 m S4 might be due to wave induced pumping of the cyclesonde rotors.



Kn88: 29.5 Day Average Along Channel Velocities and Sigma-t



By conservation of volume, the vertical integration of these profiles as a net volume transport should flow out of the inlet and be equal to the mean river discharge for the averaging period. At both Protection and Tomakstum moorings, where data existed for the entire water column, the profiles were separated into arbitrary layers corresponding to the outflowing and inflowing portions of the profile and the integrations performed. The vertical integration interval (dz) for each data point began at a point one half way between the depth of the data and the depth of the next shallowest data and extended to a point half way between the depth of the data and the depth of the next deepest data. The results are presented in table 4.1 below as m³/m width.

	Protectio	n	Tomakstum					
Layer	Thickness	Volume	Layer	Thickness	Volume			
		Transport			Transport			
	(m)	(m ³ /s per m)		(m)	$(m^3/s \text{ per } m)$			
1	105.0	-1.26	1	10.5	-0.55			
2	75.0	0.91	2	7.0	0.04			
			3	57.5	-0.76			
			4	110.0	1.72			
			5	155.0	-4.21			
Total	180.0	-0.35	Total	340.0	-3.76			
River		-0.04	River		-0.04			

Table 4.1: Volume Transport (m³/s) per metre width in the along channel direction for 1988. Note that 'layers' are arbitrary and refer to the thicknesses of inflowing and outflowing water masses as depicted in figure 4.2. Mean river discharge (98 m³/s) is also shown after being corrected for the channel width local to the mooring.

The total volume transports were of correct sign but were an order of magnitude greater than the mean river discharge, corrected for the surface channel width at each mooring. Because the channel width varies by at most ~ 10 % with depth at both moorings these results are a reasonable representation of the total volume transport. Being as conservation of inlet volume must not be violated, one is lead to suspect that cross channel variability may be more important than previously hoped. Figure 2.5 presents Webb's(1985) mean profiles from two cyclesondes moored across the channel at Tomakstum in September 1983. These moorings to the North and South of the channel centre line show differences in along channel transport at mid-depths and suggest even larger differences above 30 m. Further, despite similar inflows at the bottom of the Protection Mooring for both July and September 1983, the single centre channel mooring at Tomakstum has a mid depth inflow with a peak of twice the velocity over the two off centre moorings for September. Although the penetration into the inner basin in September 1983 compared to July 1983 could be weaker in spite of the penetrations in the outer basin being similar, this result supports the evidence from the vertically integrated volume transports that cross channel variations in flow do exist and may be significant. Although we do not have near surface data from the sill mooring, its profile suggests a serious volume imbalance suggesting more cross channel variability near the sill. Webb's(1985) results for Lull Bay just outside the inner sill also show a large volume imbalance is likely there.

The relative depths of the 24.0 σ_t isopleth in the mean profiles at the Protection and Sill moorings indicates a positive potential energy difference exists above sill depth between them. Renewal is likely taking place on a more less continuous basis throughout the experiment. The lack of a mean density profile between 9 and 230 m at Tomakstum is due to the failure of the cyclesonde conductivity cell during the deployment. The mean density profile at the Protection and Sill moorings were used to compute an estimated buoyancy period profile, expressed in minutes/radian. Values were comparable to those presented in Farmer and Freeland(1983) for the spring, with deep water values of 3 to 10 minutes/radian and near surface values of about 0.5 minutes/rad.

The 29.5 day mean along channel velocity and density (as σ_t) profiles for the 1989 experiment are shown in Figure 4.3. Because only 2 weeks of data were available from the Protection and Sill cyclesondes, averages were computed over a 14.7 day period for these





Figure 4.3: **1989 29.5 Day Mean Along Channel Current (U) and Density (as** σ_t) **Profiles** U profiles are plotted on a scale of 5 cm/s per division with a solid vertical line representing the zero for each station. σ_t profiles are plotted on a scale of 1 kg/m³ per division with a solid vertical line representing 24.5 kg/m³ for each station. Location of the 24.0 isopycnal is represented with a circle on each profile. Dashed horizontal lines represent the depth of the 12 m S4, and the deepest cyclesonde data; 170 m at the Protection and Sill moorings, 190 m at the Tomakstum and Axe moorings.

instruments. Fortunately the mean wind velocity over this period at Protection and Tomakstum was quite small.

At Protection, a surface outflow and some compensating inflow underneath are present as well as deep inflow consistent with the evidence of deep water renewal from the CTD surveys presented in Chapter 3.0. As with the 1988 experiment, the inflow just below the surface may be interpreted as providing at least a partial salt balance for entrainment despite the significantly lower surface salinities in 1989. The peak inflow velocities at about 100 m correspond to the higher density water being available for renewal at this depth during the 1989 experiment. In 1988 the peak inflow velocities at Protection were deeper (~150 m) and weaker. The positive velocities at the bottom of the Tomakstum mooring indicate that the renewal in the inner basin was deeper than in 1988 but evidence from the Axe mooring suggests this penetration did not extend to the deepest part of the basin and is consistent with the movement in the 24.4 isopycnal presented in figure 3.5. As deep basin densities (below 200 m) are lower and source water in the outer basin higher in density than in 1988, deep water renewal is expected. However observed mean penetration velocities of deep water are smaller than than in 1988.

While an inflow just below the surface layer is present at Protection, an inflow does not seem to be present at Tomakstum although a minimum in the outflowing velocity does occur. As the salinity of the surface layer remained much fresher along the whole length of the channel from the mouth to the sill in 1989 compared with 1988, less entrainment was taking place. Less entrainment would require a somewhat smaller salt compensation flow. Mixing caused by the shear with volume compensation from the deep water renewal might explain the presence of this minimum or perhaps the compensating salt water inflow is being advected seaward by the outward flow of water at mid-depths in response to deep water renewal.

At the Sill mooring, a disturbing result is the net outflow at virtually all depths. This behaviour is inconsistent with conservation of volume and may be due to topographic steering causing cross inlet variability in this region. The location of the Sill mooring was changed for the 1989 experiment placing it shallower and closer to the sill. The sill bottom topography is presented as a three dimensional plot in figure 4.4. It can be seen that the path presenting the smallest energy barrier for renewal water at or deeper than sill depth is off to the south side of the channel. This assymetry might explain the lack of inflow seen when the mooring was moved closer to the sill in 1989. Also the east-west reach realigns slightly at the sill and inertial effects may carry shallower inflows to the north side of the channel again missing the mooring. Webb(1985) presents a similar mean profile from a cyclesonde mooring at Lull Bay, just on the outside of the sill, with outflow at all depths.

Volume transports were calculated at Protection and Tomakstum by vertically integrating the mean profiles as for the 1988 data. The results are shown in table 4.2. As with 1988,

[Protectio	n	Tomakstum					
Layer	Thickness	Volume	Layer	Thickness	Volume			
		Transport	·		Transport			
	(m)	$(m^3/s per m)$		(m)	(m ³ /s per m)			
1	7.5	-0.50	1	165.0	-5.05			
2	10.0	+0.13	2	175.0	+1.51			
3	57.5	-1.94						
4	105.0	1.74						
Total	180.0	-0.57	Total	340.0	-3.54			
River		-0.24	River		-0.25			

Table 4.2: Volume Transport (m³/s) per metre width in the along channel direction for 1989. Note that 'layers' are arbitrary and refer to the thicknesses of inflowing and outflowing water masses as depicted in figure 4.3. Mean river discharge (569 m³/s) is also shown after being corrected for the channel width local to the mooring.



Figure 4.4: A Three Dimensional Plot of the Knight Inlet Sill This plot represents the bottom topography of the inner sill of Knight Inlet as digitized from chart number 3578 of the Canadian Hydrographic Service. The view is from the inner basin towards the sill with the south shore on the left and the north shore on the right. Location of the 1989 Sill mooring is shown in 180 m of water. The 1988 mooring was also centre channel in 340 m of water closer to the viewer.

they show a net outflow but larger by an order of magnitude than the average river transport. These results confirm the need to look more closely at the assumption of lateral homogeneity in inlets.

The mean density profiles for 1989 are also shown in Figure 4.3. Again a potential energy difference exists between the Protection and Sill moorings, above sill height. Therefore sufficient energy is present to drive a penetrating density flow throughout the experiment. Also of note though, is a similar difference between the Sill mooring and the Tomakstum and Axe moorings. The density profile at Protection is characterised by a noticeably larger vertical gradient with depth from about 40 m to 150 m caused by the very dense water (σ_t > 24.7) in the outer basin. Buoyancy period estimates were calculated for all moorings and were consistent over all depths, with the exception of Protection where the larger vertical density gradient mentioned above gave a corresponding shorter period. Deep water periods were in from 5 to 10 min/rad, in the same order as for 1988. Surface layer periods were much shorter than in 1988, extending to approximately 0.12 min/rad.

To allow closer inspection of the surface layer response, Figure 4.5 shows the 29.5 day (14.7 day at Protection in 1989) near surface mean velocities to 35 metres at both the Protection and Tomakstum moorings where S4 current meters were placed through the surface layer starting at a depth of 2 m. Figure 4.6 shows the 29.5 day (14.7 day at Protection in 1989) near surface mean density profiles. Note that although the runoff was on average 5.7 larger in 1989 compared to 1988, the surface layer volume transport did not increase substantially. However the vertical velocity shear is much larger. The surface salinity was extremely low along the entire length of Knight Inlet in 1989 and the increased stability of the surface layer would suppress entrainment, thus preventing the volume amplification that might otherwise be expected.



Kn88/89: 29.5 Day Average Along Channel Near Surface Velocities

Figure 4.5: **1988 vs 1989 Near Surface Along Channel (U) 29.5 Day Mean Velocity Profiles** Dashed horizontal lines show the level of the observations; S4 electromagnetic current meters to 12 m and cyclesonde profiling current meters below.



Kn88/89: 29.5 Day Average Near Surface Density

Protection

1988 vs 1989 Near Surface 29.5 Day Density (as σ_t) Profiles Dashed horizontal lines show the level of the observations; S4 electromagnetic current meters to 12 m and cyclesonde profiling current meters below. Figure 4.6:

Table 4.3 summarizes the outflowing surface layer depth, layer densities, and density differences estimated from the data presented in figures 4.5 and 4.6. The depth of the surface outflowing layer was estimated from the velocity profile as the depth of the first zero crossing or, where a near surface velocity minimum existed, the depth of the upper

Experiment	Protection				Tomakstum			
	Depth	Inflow	Outflow	$\Delta \sigma_t$	Depth	Inflow	Outflow	$\Delta \sigma_t$
	(m)	${\sigma_t}$			(m)		σ _t	
1988	12	23.7	23.2	0.5	10	23.3	21.8	1.5
1989	8	22.7	16.6	6.1	6	22.4	14.8	7.6

Table 4.3: Estimated Surface Layer Depth, Layer Densities (as σ_t), and Density Differences ($\Delta \sigma_t$) across the surface outflow and near surface inflow layers. σ_t is expressed in units of kg/m³.

inflexion point of the minimum. The average density of the outflowing surface layer was estimated as the average density of the 4 m record as the volume transport is greatest in the upper part of the surface layer. The average density of the inflowing layer was estimated to be the average of the record from just below the surface layer. The average change in density between the upper and lower layers was then calculated. The reason for the deeper depths of the surface layer for 1988 is not certain, but may be caused by wind mixing as the surface layer density is considerably higher during this experiment leading to small values of $\Delta \sigma_t$ and resulting restoring buoyancy forces.

The conservation of salt equation given in section 2.2.1, $V_0S_0 = V_1S_i$ is approximated by $V_0\sigma_{to} = V_i\sigma_{ti}$ in an estuarine situation where density is controlled by salinity. Therefore the ratio of the salt compensation inflow to the estuarine outflow should be:

$$\frac{V_i}{V_o} = \frac{\sigma_{to}}{\sigma_{ti}}$$

For 1988 the inflow volume flux should be approximately 98% of the outflow at Protection and 94% of the outflow at Tomakstum. For 1989 the inflow volume should be approximately 73% of the outflow at Protection and 66% of the outflow at Tomakstum. Clearly the volume flux of the inflow layer just below the outflowing surface layer (when there is one) is insufficient to provide the salt balance in itself. One may surmise that volume compensations due to other inflows such as deep water renewal may be advecting the salt compensation inflow out of the inlet and that this layer is really much thicker than would first appear.

The volume flow out of the inlet can be calculated using Knudsen's relations (given in section 2.2.1). Again substituting σ_t for salinity, the volume flow out of the inlet becomes a function of river discharge and the ratio of the inflowing layer density to the density difference:

$$V_{o} = R \frac{S_{i}}{\Delta S} = R \frac{\sigma_{u}}{\Delta \sigma_{t}}$$

Using the inflowing layer density, layer thickness and density differences given in table 4.3, the mean river discharge over the length of each experiment given in sections 3.3 and 3.4, and an approximate channel width of 2.5 km an estimate of the surface layer velocity can be therefore calculated by dividing the volume flow by the layer thickness and the channel width. The velocity estimates (U_0) from Knudsen's Relations and the observed average velocities vertically integrated over the surface layer (U_{avg}) are shown in table 4.4.

Experiment	Protection				Tomakstum					
{ -	<u> o</u> ti	R	Depth	Uo	Uavg	<u> o_{ti}</u>	R	Depth	Uo	Uavg
	∆σ _t _	<u>(</u> m ³ /s)	_(m)_	(cm/s	(cm/s)	∆σ _{t_} _	<u>(m³/s</u>)_(m)_	(cm/s)(cm/s)
1988	47.4	- 98	12	16	4.0	15.5	- 98	10	6	6.0
1989	3.72	569	8	11	7.0	2.95	569	6	11	14.9

Table 4.4: Estimated Surface Layer Velocities using Knudsen's relations (U_o) versus the observed integrated surface layer velocity (U_{avg}) . A linear profile was assumed from 2 m to the surface for the integration. The value used for R is the average river discharge during the experiment, the density and layer thicknesses are from table 4.3, and an approximate channel width of 2.5 km was used. $\Delta \sigma_t$ for Protection is quite uncertain and could be larger if the surface layer thickness were increased to lower the estimate of U_o .

At Tomakstum the velocities obtained from Knudsen's relations appear to give a reasonable estimate of the vertically integrated surface layer velocity, however at Protection this does not appear to be the case. In both years a deceleration of the surface layer was observed from Tomakstum to Protection possibly due to the vertical transfer of momentum due to the mixing processes over the sill. These same processes make it difficult to estimate the thickness of the surface layer and likely contribute to the error. For example if the surface layer depth at Protection in 1988 were taken as 25 m and the value of $\Delta \sigma_t$ adjusted accordingly, the estimated and average velocities would be in reasonable agreement. The average velocity profile still shows outflow at this depth and perhaps the interpretation that this outflow is completely due to volume compensation due to deep water renewal is incorrect. The observed deceleration from Tomasktum to Protection is consistent with the thickening of the surface layer, perhaps due to vertical momentum redistribution as a result of sill processes. For 1989 the depth would have to be 12 m. While this seems unlikely looking at the average velocity profile, it is perhaps illustrative of the types of problems that can arise in approximating a continuous system as two layers.

In both years the volume flux out and the density of the inflowing layer are approximately the same, while the ratio of the 1989 to 1988 density difference (12 at Protection and 5 at Tomakstum) increases to the same order as ratio of the 1988 to 1989 record average river discharge(5.8). This result indicates that the balance in Knudsen's relations comes not from an increased surface volume flux with discharge, but from a change in the entrainment leading to the larger density differences observed with increased runoff.

Wetton(1981) used dynamic height calculations to estimate the surface slope and showed that it increased with runoff. van der Baaren(1988) also estimated surface slopes in the same way during high runoff in 1986 and 1987. Pond and Pickard(1983) provide a general

introduction to the calculation of dynamic heights through the numerical integration of specific gravity anomaly with pressure. Of course Wetton and van der Baaren had to assume that the pressure gradient was zero at some fairly deep level (200 m for Wetton and 100 m for van der Baaren). While it might not have been zero, as we have seen that the velocity is not zero at depth, it will be small compared with the surface pressure gradient. Therefore any pressure gradient that existed at the level of no motion would be much less than the surface slope and the error in the surface slope estimation will be small. van der Baaren's programs were recovered from tape archive and used to estimate the surface slopes for both experiments using data from the 1988 deployment and 1989 deployment and pickup cruise CTD surveys. Dynamic height profiles for casts taken inside the inner basin (stations 4 through 11) were calculated at 1 m intervals to 50 m and at 5 m intervals thereafter, from 1 m to the level of no motion . The level of no motion was taken as 180 m for the 1988 experiment and 170 m for 1989 experiment and was selected from the Tomakstum 29.5 day mean velocity profile presented in figures 4.2 and 4.3 by estimating the depth of the deepest zero crossing.

Following van der Baaren(1988) surface slope was then estimated by applying a linear least squares fit through the dynamic heights at 1 m, the shallowest level available. A composite plot of all calculated surface slopes (including those of Wetton and van der Baaren) versus river discharge is presented in figure 4.7. Although this plot is noisy, there is a clear relation between increasing river discharge and increasing surface slope. The three isobaric surface slopes calculated for the 1988 and 1989 experiments are consistent with those of Wetton, while somewhat under those of van der Baaren. However, a CTD survey of an inlet that is about 100 km long requires a great deal of time and CTD casts were not taken during the same phase of the tide. As the typical slope of say 10⁻⁶ corresponds to about a 10 cm rise over 100 km noise in these results is hardly surprising. Three of Wetton's surface slope estimations had pressure gradients of reverse sign sloping back towards the



Knight Inlet Surface Slope vs River Discharge

Figure 4.7: Surface Slope (m/m) vs River Discharge for Knight Inlet Note that surface slopes has been multiplied by -1 so that slopes above zero on the y axis are numerically negative. This plot incorporates the results of Wetton(1981) [open circles], van der Baaren(1988) [dots], and those from the 1988 deployment and 1989 deployment and pickup cruises [solid squares]. The level of no motion used were 200 m, 100 m, 180 m and 170 m respectively.

head of the inlet and driving surface flow back towards the river mouth. Because these three slopes were based on data from cruises in November, December and January, strong outflow winds in conjunction with the low river discharge may be responsible. Winds and the difficulty in estimating a level of no motion without velocity data contribute to the noise observed in the relationship between surface slope and river discharge.

Following Wetton(1981) and van der Baaren(1988) I also estimated the isobaric slope profiles by repeating the least squares fit at each level through to the level of no motion. Unlike those of Wetton and van der Baaren whose isobaric profiles with only one exception do not cross zero, the profiles calculated for the 1988 and 1989 data set show zero crossings in the pressure gradient consistent with the character of the mean velocity profiles. The 1989 pickup cruise isobaric slope profile resembles a detided (see section 4.3 for method) velocity profile from the Tomakstum mooring from a few days earlier, before the cyclesonde stopped profiling. The 1989 isobaric slope profiles and the velocity profile are presented in figure 4.8.

The reason for the difference between these results and those of Wetton and van der Baaren is not certain, but may simply be due to offset errors arising from selecting a level of no motion where small velocities are present. The 1988 and 1989 data sets suggest that selecting a level of no motion is difficult without velocity data. Also, the correlation in the least squares fit while high at the surface (0.7 to 0.8) falls off rapidly as slopes are estimated through the interior of the fluid. Except near the surface where the slopes are large, dynamic height calculations are noisy and that noise could easily mask the character of the internal pressure gradients.

The balance of forces for the surface layer in Knight Inlet was shown by van der Baaren(1988) to be between the surface pressure gradient and the interfacial friction except


Kn89: Isobaric Profiles and Representative Tomakstum U Velocity Profile

Figure 4.8: Isobaric Slope Profiles for the 1989 Experiment

Isobaric slope profiles as estimated from the 1989 deployment and pickup cruises and a representative velocity profile from Tomakstum. Slopes are in m/m. The level of no motion used was 170 m. These two isobaric slope profiles clearly show reverse pressure gradients driving inflows at some depth. Note the maximum slope at the surface is off scale on these profiles. The pickup cruise isobaric slope profile resembles the detided velocity profile at Tomakstum from a few days earlier.

perhaps seaward of the sill where inertial terms were larger. It appears that, compared with the observed 29.5 day mean velocities from the 1988 and 1989 data sets, the estimates of surface velocity calculated using Knudsen's relations shown in table 4.4 and those of van der Baaren shown in figure 2.6, were high outside the sill. In fact there is a small deceleration in the surface layer from Tomakstum to Protection in both years, consistent with a thickening of the surface layer due to increased mixing and the resultant loss of momentum. Therefore it is likely that the balance between the surface pressure gradient and interfacial friction holds true all along the inlet.

Referring again to figure 4.5, a much larger vertical velocity shear can be seen across the surface layer during the 1989 experiment suggesting that greater turbulent friction is balancing the larger pressure gradient associated with the increased runoff. van der Baaren(1988) estimated coefficients of interfacial friction in Knight Inlet for the late spring of two years, 1986 and 1987. The runoff during 1986 was much higher and her calculated frictional coefficients were an order of magnitude higher for 1986 than 1987. She also states that these coefficients are not static values, and likely vary with and on the same temporal scales as the tides and the wind.

4.3 Harmonic Analysis and Computation of the Detided Time Series

Detiding was accomplished through harmonic analysis, the least squares fit of well known astronomical tidal forcing frequencies to the fluctuations of each data record. This procedure estimates a fixed amplitude and relative phase representing the contributions of each tidal constituent to the data record. Once the amplitude and relative phase of the constituents has been determined, the tide can be reconstructed and subtracted from the original data record to create a detided residual time series.

The harmonic analysis program used solves an over determined system expressed as a linear matrix equation

where:

[A] [X] = [D]
[A] is the coefficient matrix to be determined
[X] is the matrix of orthogonal tidal constituents
[D] is the matrix of observational records

[D] is a matrix of observational records and for the purpose of this analysis consisted of either vector velocity (u, v) or scalar (t, s, σ_1) time series from a single depth. [X] is a matrix of orthogonal constituents for each tidal forcing frequency expanded as sine and cosine terms and computed along the same time base as the observational data in matrix [D]. The mean and the trend were also included as zero order and first order polynomials. [A] is the relative amplitudes of the sine and cosine components of each constituent and is determined by the following process. First, for the solution of [A], the equation is rearranged to the form:

$$[A] = [X]^{-1} [D]$$

Second, the inverse matrix form of [X] is computed and the problem is reduced to decomposing the equation to find the triangular form of matrix [A]. Third, the solution to

[A] can obtained through back substitution. As each vector velocity component or scalar property time series at each depth held simultaneous observations, the manipulations to obtain a triangular matrix need only be performed once and then a back substitution made into each of u and v or temperature, salinity and σ_t to determine the coefficients for each. The amplitudes of the cosine and sine components of each trigonometric constituent where then converted to an amplitude and relative phase to describe the constituent contributions. For the purposes of tidal prediction and comparison to other observations, phase is usually expressed relative to Greenwich Mean Time. However, since only relative phases were required for detiding, the phase calculated was left relative to Julian Day 1.0, midnight (PST) January 1 of the year of each experiment.

The method of singular value decomposition is recommended by Press et al.(1988) for reducing [A] to triangular form as it produces the best fits in the presence of 'noise', in this case fluctuations not attributable to the tidal constituents alone. It also has an advantage in that this method was chosen for analysis in general inlet circulation models under development by others and procedural compatibility was desirable for future work. As the lower frequency monthly and fortnightly tidal constituents fall well inside the bandwidth of the low frequency residual as defined in this work, only the diurnal and higher frequency tidal constituents were subtracted from the data record as part of the detiding procedure. However the mean, trend and lower frequency monthly and fortnightly tidal constituents were included in the analysis as including them considerably improved the fit of the higher frequency constituents to the data records.

The selection of the astronomical tidal constituents to include in the harmonic analysis requires care as harmonic analysis does not conserve variance. The high frequency limit was determined from the Nyquist frequency, one over twice the sampling rate of the data. For example, as the processed time series had a sampling rate of at least every 3 hours, the

 M_4 quadiurnal harmonic was included. The lower limit is determined by the ability to resolve a full cycle of the constituent over the data record and therefore all trigonometric constituents included in the analysis must have a frequency greater than one over the record length. For example, the observational records were generally for greater than 30 days, and therefore the lunar monthly constituent M_m was included. The exception was the 1989 Protection cyclesonde mooring where premature instrument failure allowed only the lunar solar fortnightly constituent M_S_f , to be resolved.

Between these limits, care must also be taken in specifying the constituents so that their frequencies are far enough apart that they can be clearly distinguished by the fitting procedure. If two constituents fit the data equally well, ambiguities may lead to computed coefficients with large magnitudes that are delicately and unstably balanced to cancel out. To determine if two tidal constituents were too close in the frequency domain, the Rayleigh criterion specified by Foreman (1977, 1979) was applied. This states that the record length must be greater than the reciprocal of the difference between the two frequencies to be resolved. For example, applying this criterion to the principal lunar M₂ and principal solar S₂ constituents, a record length of greater than 14.9 days is required. And so both of these were included. Applying this criterion to the MS_f and M_f a record length of 182 days is required. In the latter case, the MS_f was included over the M_f because a strong non-linear interaction between the M₂ and S₂ was reported by Freeland and Farmer(1983) in Knight Inlet. Similarly the Q_1 , P_1 , and K_2 constituents were left out of the analysis. For both 1988 and 1989, with the exception of data from the 1989 Protection and Sill cyclesonde time series, the constituents chosen for analysis are given in Table 4.5. In the two cases cited above the M_m was not included.

#	Name	Description	Period	Frequency
			(nours)	(cpa)
	Z_0	- Mean, zero order polynomial		
2	Z_1	- Trend, 1st order polynomial		
3	М _т	- Lunar Monthly	661.3	0.0363
4	MSf	- Lunar Solar Fortnightly	354.4	0.0677
5	O ₁	- Principal Lunar Diurnal	25.82	0.930
6	\mathbf{K}_{1}	- Luni-Šolar Diurnal	23.93	1.003
7	N_2	- Large Lunar Elliptic	12.66	1.900
8	M_2	- Principal Lunar	12.42	1.932
9	S_2	- Principal Solar	12.00	2.000
10	MK ₃	- Lunar Solar Tridiurnal	8.18	2.93
11	M4	- Quadiurnal Harmonic	6.12	3.92

Table 4.5: Tidal Constituents used in Harmonic Analysis

Once the contributions of each constituent to the data had been computed, those constituents with periods equal to the O_1 and shorter were reconstructed from the determined amplitude and phase, and subtracted from the original time series to form the detided residual. The mean, trend and low frequency (M_m and MS_f) trigonometric constituents were left in, as these have frequencies in the same band as the desired residual and subtracting them would also subtract some of the response from other forcing at these frequencies. In effect the detiding procedure, has the same effect as a low pass filter for the observations, selectively removing only diurnal and higher frequency tidal energy.

Close inspection of the spectrum of a time series before and after detiding revealed that the tidal constituents are significantly reduced by this process but not totally eliminated. It is likely that the tidal response is not strictly invariant over the length of the observational record. Stacey(1985) noted a change in the modal response of the tide with stratification in Knight Inlet which implies that the constituent parameters may be continuous functions of river runoff and wind mixing. Other non-tidal high frequency variance also remains in the record although it has little energy. For the purposes of presentation only, a low pass spectral filter with a half power cutoff equal to 0.9 times the absolute cutoff frequency and a discretized cosine taper was used to remove remaining higher frequency fluctuations. For

the purposes of displaying the data records in the following sections, the absolute cutoff was set to equal to the lowest tidal frequency (O_1) removed.

Note that cyclesonde data are slightly irregular in time since the instrument alternately profiles up and then down through the water column and each profile takes approximately one half hour. In order to apply a filter or do spectral analysis, data equally spaced in time were required whereas the harmonic analysis can be and was done using the original data. As the data needed adjustment over only a small fraction ($\leq 1/5$) of the sampling interval, linear interpolation was used to grid the detided data onto a regular sampling interval. Being as the bulk of the higher frequency variance has been removed by the detiding procedure, the interpolation should have little effect, if any, on the lower frequencies. The cyclesonde at Tomakstum in 1989 failed to profile at regular intervals, especially at the start and the end of the data records. In order to allow these data records to be subject to the same spectral analysis and filters as the others, the detided records for this instrument were linearly interpolated at 3 hour intervals between existing data points and there is a bit more uncertainty in these results.

Note that further processing and analysis utilized the unfiltered detided data so as to not to adversely affect the analysis of contributions at frequencies near the filter cutoff. For consistency in presentation, where wind and detided data are presented together, the wind data were subjected to the same low pass filter.

4.4 Detided Inlet Response

The results of the detiding process for the along channel currents at selected depths are shown in Figures 4.9 and 4.10. Note that depths selected for display are representative of the surface layer and those deeper depths where peaks in correlation of wind (shown in Figure 4.11) to velocity were found. Because the tidal fluctuations have been removed, it was necessary to change the velocity scales with depth in order to adequately portray the response.

Figure 4.9 shows that the 1988 2 to 12 m along channel currents at Protection closely resemble the wind with a decrease in this response with depth. At 12 m the current still appears positively correlated with the wind, but with a larger lag and about half the peak values of the 2 m current. At 25 m the wind driven current is smaller still and appears to be of opposite phase with the wind for the last two thirds of the record, while at 150 m the wind response appears larger than that observed at 120 m but with what appears to be a more complex phase relationship than that observed at shallower depths. The peak lag correlation at 150 m is negative while at 120 m it is positive. There clearly appears to be wind influence throughout the water column at the Protection mooring in 1988. In contrast, the wind response at Tomakstum is more confined to the surface layer, with much less response at 12 m, probably due to the greater stratification on the inside of the sill and buoyancy forces suppressing the vertical transfer of momentum. Correlations at depth are small but fluctuations with the same character as the wind's lower frequencies dominate the along channel currents as deep as 270 m.



1988 Filtered Wind and Detided Along Channel Velocities

Figure 4.9: **1988 Along Channel Wind vs Selected Detided Currents** Note both wind and detided currents have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. Depths shown are representative of the surface layer and of those depths where significant correlations with the wind were found.

Figure 4.10 shows that the 1989 2 and 4 m along channel current at Protection and to a lesser extent at Tomakstum also resemble the wind. At 12 m the current appears to be negatively correlated with the wind at both moorings. Stratification is stronger in 1989 than in 1988 and is confining the direct wind response nearer to the surface. Unfortunately the Protection cyclesonde failed half way through the experiment and the dominant forcing frequencies of the wind (10 to 15 days) are much lower, making it difficult to place a lot of faith in the correlations at depths greater than 12 m. At Tomakstum the cyclesonde only intermittently profiled at the beginning and end of the experiment and occasionally skipped profiles during the rest of the deployment. To allow the data records to be used in further analysis and to be filtered with the same spectral filter, a linear interpolation between data points in the detided records were used as a basis to estimate values for the missing data points. The 15, 70 and 120 m filtered time series suggest a much stronger wind response at depth in the inner basin than in 1988. At 15 m the currents appear to be negatively correlated with the wind and at 70 and 120 m positively correlated. At 310 m a strong negative correlation appears. However, one must be careful about inferring that these correlations are all due to baroclinic response to the wind. The frequencies of the wind are near fortnightly and it is likely that some of the motion may be due to the fortnightly tides or deep water renewal modulated by the spring/neap tidal cycle.

Figure 4.11 shows the lagged cross correlations of the wind and along channel currents for both experiments. These values were plotted using the sign of the correlation coefficient (R) but with a magnitude of R², the coefficient of determination. Roughly this value estimates the intensity of association between two variables that appear correlated and can be interpreted as the percentage of the energy in the along channel currents that might be explained by the energy in the wind.



Figure 4.10: **1989 Along Channel Wind vs Selected Detided Currents** Note both wind and detided currents have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. Note that prior to filtering, where the cyclesonde failed to profile continuously, a linear interpolation between the detided samples was used to provide an estimate of a continuous time series. Depths shown are representative of the surface layer and of those depths where significant correlations with the wind were found.



Wind/Along Channel Velocity Lag Correlations With Depth



Peak correlations are indicated with circles for Protection and squares for Tomakstum. Correlation was plotted as R^2 (with the sign of R) as it gives an estimate of the contribution of the wind driven current energy to the total current energy at that depth. Table 4.6 gives the lags associated with the surface layer correlations and the peak lag correlations at depth.

Table 4.6 gives the lags associated with wind and along channel velocities for the surface layer and for the peaks in the correlation profiles at depth as highlighted by the circles and squares on figure 4.11.

Depth	1988		1989	
(m)	Prt	Tom	Prt	Tom
	R^2 , Lag(h)	R^2 , Lag(h)	R^2 , Lag(h)	R^2 , Lag(h)
2	0.81, 7	0.72, 4	0.83, 1	0.71, 2
4	0.81, 7	0.56, 5	0.66, 1	0.15, 0
6	0.76, 8	0.37, 6	0.07, 0	-0.24, 30
9	0.72, 8	0.17, 10	-0.35, 30	-0.26, 30
12	0.50, 9	0.03, 11	-0.56, 14	-0.27, 25
15	-0.29, 39	-0.10, 63	-0.52, 15	-0.64, 45
25	-0.52, 33	•	,	,
35		-0.37, 3		
50		-	0.26, 48	
70			•	0.55, 63
80			-0.45, 0	,
120	0.55, 39			0.69, 72
140			-0.35, 39	,
150	-0.37, 69			
180		0.10, 81		
270		-0.10, 81		
310		• -		-0.72, 63

Table 4.6:

.6: Surface and Peak Lag Correlations at depth (as R² with the sign of R) for the Wind and Along Channel Currents

Through the surface layer the lag times appear to describe the time scales associated with the vertical transfer of momentum. For example, in 1988 the peak correlation at 9 m is weaker in the inner basin at Tomakstum and the peak correlation occurs at a greater time lag (10 hours) than at Protection (8 hours) although at 2 m the situation is reversed with a smaller lag time in the peak correlation at Tomakstum. In 1989 the lower density of the surface layer results in a rapid (1 to 2 hours) vertical transfer of momentum at both moorings, although the time taken for this transfer at Tomakstum is higher. It appears that the stratification is suppressing vertical momentum transfer as shallow as 4 m in the inner basin in 1989.

At depth, peak correlations occurred with the current lagging the wind with temporal scales of 40 to 80 hours. As the speed of a mode 1 internal wave based on the densities and layer thickness given in table 4.3 at Tomakstum is about 0.4 m/s (1988) and 1 m/s (1989). It would take about 83 hours in 1988 or 33 hours in 1989 for a baroclinic wave to travel the approximately 60 km up to the head and return. To balance the pressure gradients associated with the surface slope built to balance the wind stress, the isopycnals at depth near the head would have to be depressed. This baroclinic adjustment would require an internal wave to reach the head and return. If these peaks in the lag correlation are due to the wind these temporal scales support the work of Buckley and Pond(1976) who found that the baroclinic adjustment controlled the wind response in Howe Sound.

Wetton(1981) raises the possibility of quarter wave internal resonance for wind forcing in a channel open at one end. The period required for a mode 1 wave to traverse a 100 km long inlet would be about 69 hours in 1988 and 28 hours in 1989. A standing wave in a channel with a length equal to one quarter wave length, requires a forcing period of 11.5 days in 1988 and 4.6 days in 1989. As these are within the temporal scales of the wind forcing for the 1988 and 1989 experiments, perhaps some of the wind response at depth is due to internal seiches. As the pressure gradient response to the wind forcing will vary from being 180 degrees out of phase at frequencies much lower than resonance to being in phase at the resonant frequency this process may account for some of the apparent change in the phase with frequency noticed in the deeper along channel velocity time series.

At 80 m at Protection in 1989 a reasonably strong anticorrelation with the wind is found at a lag of 0 hours. However, the time scales of the wind record in 1989 are longer than in 1988, from 10 to 15 days and the shortness of the along channel current record due to instrument failure makes the interpretation of the correlations difficult. Also, the time scales of the wind are close to the time scales associated with the spring/neap tidal cycle in 1989 and the possible fortnightly tidal modulation of deep water renewal in the inner and outer basins also makes an interpretation of the correlations difficult.

Figures 4.12 and 4.13 present the detided density (as σ_t) time series at Protection and Tomakstum for the 1988 and 1989 experiments. As noticed earlier there are large fluctuations in the surface layer density field that resemble the wind. With up inlet winds, a corresponding raising of the isopycnals occurs as estuarine outflow is retarded or reversed. With down inlet winds, a lowering of density is observed consistent with the reestablishment of the estuarine outflow. In 1989, where winds are of a much longer period the up inlet winds at Tomakstum starting at day 186 are followed by an increase in the 2 m density for a period of about 4 days followed by a sudden drop despite continuing up channel winds. This result may be explained by the fact that the surface pressure gradient must come into balance with the wind at some point if the forcing is steady. At this time the estuarine outflow will re-assert itself with the fresher water lowering the observed density again. The time scale of this adjustment (4 days or 96 hours) also suggests that the wind response is controlled by the baroclinic and not barotropic adjustment. The density fluctuations are larger at Tomakstum than at Protection in both years because of the stronger pycnocline on the inside of the sill. The 2 m detided near surface density records show a trend to decreasing density with time consistent with the increase in river discharge towards the end of each experiment.

In order to improve the analysis further, a cross spectral analysis between the wind and the along channel currents is presented in section 4.5. The coherence spectra may be used to estimate the contributions of the wind to the current in each frequency band and calculate the wind driven flow and the phase spectra will allow the investigation of mechanisms controlling the wind response in the along channel currents at depth.



1988 Filtered Wind and Detided Densities (as Sigma t)





1989 Filtered Wind and Detided Densities (as Sigma t)

Figure 4.13: **1989 Wind vs Selected Detided Density (as** σ_t) Note in the above composite plot, σ_t increases upward and each division is 5 kg/m³. Both wind and densities have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. The wind influence on the density field is larger than in 1988 and more confined to the surface layer due to the stronger pycnocline.

4.5 Cross Spectra of Wind versus Along Channel Current and Density

The lag correlations presented earlier indicated that the dominant forcing in the surface layer was wind, and that considerable effects were discernible at other depths in the water column. Further, the raw spectra of the along channel currents presented in section 3.6, showed energy in the same bands as the wind spectra. A cross spectrum presented in the form of coherence and phase spectra can provide information similar to the correlations presented in the previous section, but as a function of discrete frequency bands rather than as a single number representing all of the variance in the two records.

The original data records were not sampled at the same rates at all depths and the data at 15 m through 190 m was not sampled on a strictly repeated sampling interval. Cyclesondes profile through a given depth starting alternately from either the top or the bottom of their profiling range making the sampling interval at a specific depth vary by as much as 40 minutes in the worst case. As the detided observations have had all diurnal and higher frequency removed, all data was gridded by linear interpolation to a fixed sampling interval of 3 hours. This places the Nyquist folding frequency at 4 cycles per day and above any significant remaining energy in the detided residual data records. It also has the advantage of placing all of the spectral estimates on the same frequency scale.

The frequency spectra of the wind, current, and density were computed by demeaning and detrending the time series and by applying the fast fourier transform algorithm. This procedure determines a series of discrete fourier coefficients, a_k and b_k , that represent the amplitude of two sinusoids in quadrature at each specific frequency in the discrete fourier series that represents the original data.

The discrete fourier series representing the transformed data can be expressed as follows:

$$X(t) = \sum_{k=1}^{\frac{N}{2}} \left(a_k \cos(2\pi f_k t) + b_k \sin(2\pi f_k t) \right)$$

where: $f_k = 1/(2N \Delta t)$, represents the discrete frequencies in the transform N = total number of samples in a time series $\Delta t = the sampling interval of the data record. Note the frequency range$ over which the transform is performed is between the limits: $<math>f(1) = 1/(N \Delta t)$, one cycle in the record length $f(N) = 1/(2 \Delta t)$, the Nyquist Frequency, one cycle in 2 samples

From the coefficients in the discrete fourier series, the power can be estimated at each discrete frequency as follows:

$$P(f_k) = \frac{1}{2} [a_k^2 + b_k^2]$$

Because the sampling function is a discrete boxcar whose frequency domain representation is a sinc function, the spectrum is the convolution of the original signal's spectrum and this function. If the original data contains a single sinusoid at precisely one of the discrete fourier frequencies, it will be correctly represented as a single spectral line. If the original data contains the representation of a single sinusoid at a frequency other than a fourier frequency, energy is spread amongst the nearest discrete fourier frequencies. Observational data contains many characteristic frequencies and therefore to improve the accuracy of the fft power estimation, band averaging is routinely applied. All power spectra presented in this section have been averaged using a Daniel window spanning seven discrete fourier frequencies.

The cross spectrum can be computed from the discrete fourier coefficients for any two time series that are coincident on the same time domain, X(t) and X'(t). It consists of two orthogonal spectra the co-spectrum $Co(f_k)$, and the quadrature spectrum $Q(f_k)$. The cospectrum $Co(f_k)$ is defined as that part of the two spectra that are in phase, while the quadrature spectrum $Q(f_k)$ is defined as the part of the two spectra that are out of phase by 90 degrees. These spectra are computed from fourier coefficients of the two time series, a_k b_k and $a_k' b_k'$, as follows:

$$Co(f_k) = \frac{1}{2} [a_k a_k' + b_k b_k']$$
$$Q(f_k) = \frac{1}{2} [a_k b_k' + a_k' b_k]$$

An alternative and often more useful representation of the cross spectrum for analysis are the coherence $C(f_k)$ and phase spectra $F(f_k)$. These are computed from the co and quadrature spectra as follows:

$$C^{2}(f_{k}) = \frac{Co^{2}(f_{k}) + Q^{2}(f_{k})}{P(f_{k})P'(f_{k})}$$
$$F(f_{k}) = Arc \tan\left(\frac{-Q(f_{k})}{Co(f_{k})}\right)$$

Calculation of the coherence and phase spectra must be made over bands of discrete frequencies, for without it the coherence is identically 1 as any discrete sinusoid is coherent with any other of the same frequency. Through experimentation it was found that averaging over seven discrete fourier frequencies produced reasonable estimates of coherence and phase through the wind band while maintaining as much spectral resolution as possible. The coherence squared can be interpreted as that fraction of variance of one signal that is related to the other in each frequency band regardless of phase. The phase gives the amount by which the fluctuations in one time series leads or lags the other for each discrete frequency band.

Figures 4.14 through 4.17 show representative band averaged power, coherence squared and phase spectra of the wind and along channel currents at each mooring. The power spectra of the wind are scaled and plotted with the 2 m power spectrum with a grey line for reference. As with the raw spectra presented earlier, the vertical axis is plotted as $f\Phi(f)$ so that the area under the curve is proportional to the total variance of the original record on a semi-logarithmic plot. Note that while there is more energy in the wind at Protection than at Tomakstum, there is more energy in the 2 m current at the peak frequencies of the wind for Tomakstum in 1988. A comparison of the power spectra to 12 and 15 m suggests that the stronger stratification at Tomakstum is confining the direct effects of the wind more closely to the surface.

In 1988 the along channel velocities are generally highly coherent with the wind (where coherence squared is > 0.5) in the low frequency end of the spectrum with periods generally greater than 2 and less than 7 days. At Protection there is clearly wind energy to 150 m near the bottom of the mooring, while there is considerably less at 25 m particularily at the frequencies corresponding to the peak of the wind energy. At Tomakstum the deep currents are not very coherent, at least at the dominant 2 to 3 day period of the wind.

The 1988 phase spectra for the near surface are consistent with the lag times of the peak correlations presented earlier in section 4.5. At 25 m depth and greater at Protection, the phase of the energy at the frequencies close to the peak of the wind energy is close to being 180 degrees out of phase with the surface currents (and thus slightly more than 180 degrees with respect to the wind). At Tomakstum, a similar phase reversal can be found but shallower at 15 m. Below approximately 50 m the coherence squared falls below 0.5 and the phase spectra become more difficult to interpret. Except in the near surface, the phase is a function of frequency and it appears that the contribution of the wind to the variance of the deeper records is more complex than can be determined with a simple lag correlation. This phase variation may explain some of the higher correlations found at depth in section 4.5. At the high frequency end of the spectra, the coherence squared is generally low and



1988 Protection Cross Spectra of Wind vs Along Channel Currents at Selected Depths



4: 1988 Protection: Power and Cross Spectra of Wind vs Along Channel Currents at Selected Depths

Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the currents are lagging the wind, negative phases denote that the currents are leading the wind.



1988 Tomakstum Cross Spectra of Wind vs Along Channel Currents at Selected Depths

Figure 4.15: **1988 Tomakstum: Power and Cross Spectra of Wind vs Along Channel Currents at Selected Depths**

Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the currents are lagging the wind, negative phases denote that the currents are leading the wind.

unstable with a corresponding phases that have large variations. Priestley(1981) states that extremely large variations in the phase may occur when coherence is low.

For 1989 the highly coherent part of the signal is also in the lower end of the spectrum, but generally with periods about 1.5 to 30 days. The broader band of the 1989 coherence squared spectra is due to the shift in the wind regime noted earlier. As will be shown later in Figure 4.26, the energy coherent with the wind is more confined to the surface in 1989 probably due to the stronger stratification, with significant wind energy penetrating deeper to perhaps 20 m at Protection. The near surface velocity phases appear to be consistently leading the wind in 1989. This result is somewhat contrary to expectations and was also found in the lower frequencies of the wind response in Alberni Inlet by Farmer(1972). He proposed that offshore disturbances may propagate as surface gravity waves in advance of the wind.

Figures 4.18 through 4.21 show representative band averaged power and cross spectra of the wind and density fluctuations (as σ_t) at each mooring. In general much less of the variance in the density field is coherent with the wind at the peak frequencies of the wind forcing. The density field appears to respond more to the lower frequencies in the wind energy, probably due to the periods of net up and down inlet wind stress reversing the flow of the surface layer and affecting the density field. In 1988 at Protection and Tomakstum the phase spectra of the 2 m density field is at ~90 degrees indicating the wind driven density is fluctuating in quadrature with the velocity. Thus the near surface density is responding to the time rate of change of the velocity. This response can be taken as evidence that a pressure gradient does build to balance the wind stress. In 1989 the wind driven density is close to being in phase with the velocities indicating the stronger stratification is inhibiting mixing and the fluctuations in surface density are likely the result of the surface layer behaving as a 'slab' flow.



1989 Protection Cross Spectra of Wind vs Near Surface Along Channel Currents

Figure 4.16: **1989 Protection: Power and Cross Spectra of Wind vs Along Channel Currents at Selected Depths**

Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the currents are lagging the wind, negative phases denote that the currents are leading the wind.



1989 Tomakstum Cross Spectra of Wind vs Near Surface Along Channel Currents

Figure 4.17: **1989 Tomakstum: Power and Cross Spectra of Wind vs Along Channel Currents at Selected Depths**

Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the currents are lagging the wind, negative phases denote that the currents are leading the wind.



1988 Protection Cross Spectra of Wind vs Selected Densities (as or)

Figure 4.18: **1988 Protection: Power and Cross Spectra of Wind vs Densities at Selected Depths** Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the densities are lagging the wind, negative phases denote that the densities are leading the wind.



Figure 4.19: **1988 Tomakstum: Power and Cross Spectra of Wind vs Densities at Selected Depths** Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the densities are lagging the wind, negative phases denote that the densities are leading the wind.



1989 Protection Cross Spectra of Wind vs Near Surface Densities (as or)

Figure 4.20: **1989 Protection: Power and Cross Spectra of Wind vs Densities at Selected Depths** Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the densities are lagging the wind, negative phases denote that the densities are leading the wind.



1989 Tomakstum Cross Spectra of Wind vs Near Surface Densities (as or)

Figure 4.21: **1989 Tomakstum: Power and Cross Spectra of Wind vs Densities at Selected Depths** Note the grayed power spectra at 2 m is the wind power spectra for reference. All spectra have been averaged over 7 adjacent discrete frequencies and have been computed from the unfiltered detided time series. Positive phases denote that the densities are lagging the wind, negative phases denote that the densities are leading the wind.

4.6 Wind Driven Response

The evidence from the cross spectra presented in section 4.5 suggest that the wind response may be estimated using the coherence from the cross spectra of the wind versus the along channel currents and density fluctuations. As the coherence spectrum can be interpreted as that fraction of fluctuations of one signal that are related to the other, it can serve as a basis to calculate the contribution of the wind to the fluctuations in the along channel currents or density records. Farmer(1972) used this technique to remove the variance due to the winds from the near surface along channel currents in Alberni Inlet. The spectral transfer function he used is defined as:

$$T(f_{k})_{w \to u} = \sqrt{\frac{P(f_{k})_{u}C^{2}(f_{k})_{wu}}{P(f_{k})_{w}}}$$

where:	$T(f_k)_{w->u}$ $P(f_k)_u$	is the amplitude transfer function describing the contribution of the wind to the along channel current or scalar property. is the power spectrum of the along channel current or scalar
	$C^2(f_k)_{wu}$	is the coherence squared between the wind and the along
	$P(f_k)_w$	is the power spectrum of the wind.

Hence in this application, the fourier series representing the wind driven component of a data record can be estimated by multiplying the fourier series representing the along channel current or density fluctuations by the coherence from the cross spectrum. The time series representing the wind driven component can then be computed by applying an inverse fourier transform to the discrete fourier series. Because the mean and the trend were removed before the calculation of the fourier series, all of the mean and the trend are left out of the wind driven component as determined by this method. The estimation of the wind driven component of a data record uses the coefficients of the discrete fourier transform of

the original current or density data record complete with its phase information intact and hence the phase spectrum is not used in the calculation.

The fourier transform and its inverse preserve only a time base relative to the start of the data record, and therefore the original time base of the data record was recreated by applying a time offset equal to the time of the first sample in the original detided time series. To remove high frequency noise, the same cosine tapered spectral filter with a cutoff frequency equal to 0.929 cpd used in the presentation of the detided data was applied before plotting the wind driven time series.

The wind driven velocity and density time series at selected depths for 1988 and 1989 are shown in figures 4.22 through 4.25. The depths chosen for display correspond to those of the detided time series presented in figures 4.9, 4.10, 4.12 and 4.13. As the cyclesonde at Protection in 1989 stopped profiling approximately half way through the experiment the wind driven currents were calculated using a shorter 14 day time series. The cyclesonde at Tomakstum in 1989 profiled only intermittently at the beginning and the end of the experiment while profiling continuously in the middle of the record. Therefore a linear interpolation was made between data points in the detided (but not filtered) time series and the interpolated data subject to the same spectral processing as the others.

In both years the 2 m wind driven currents shown in figures 4.22 and 4.23 closely resemble the 2 m detided currents and clearly have approximately the same amplitudes. In 1988, it is clear that a good proportion of the variance at Protection is due to the wind at most depths however this is much less true at Tomakstum, where the down inlet flow at



1988 Filtered Wind and Wind Driven Along Channel Velocities

Figure 4.22: **1988 Along Channel Wind vs Selected Wind Driven Currents** Note both wind and wind driven currents have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. Depths shown correspond to those chosen for the detided times series presented in figure 4.9.



1989 Filtered Wind and Wind Driven Along Channel Velocities



270 m in particular (seen in figure 4.9) does not appear to be wind related. In 1989 the magnitude of the wind driven current below the surface at Protection is less than in 1988 but the record is short. The reverse is true at Tomakstum, where considerable motion in the deep basin (120 and 310 m) seems to be attributed to the wind. Perhaps it is the baroclinic response to the surface setup caused by the longer period wind forcing in 1989 or it may be associated with deep water renewal with a period similar to the wind.

The near surface wind driven density field in figures 4.24 and 4.25 appears to follow the fluctuations in the detided data, but clearly not all of these fluctuations can be attributed to the wind. The density fluctuations are also a function of river runoff and whatever mixing processes (particularily those in the vicinity of the sill) that water being returned up inlet may have been subject to. The wind driven fluctuations at depth shown in figures 4.28, 4.30, 4.32, and 4.34 are also small. In the basin the horizontal gradients of density are known to be small and even large flows due to baroclinic compensation for surface setup would not result in a large change in density.

Figure 4.26 shows the portion of the detided variance (as percentage) that may be attributed to the wind. It was calculated by finding the variance of the wind driven time series for all depths and dividing by the variance of the corresponding detided time series. At Protection, from 35 to 50 % of the detided along channel velocity variance is attributed to the wind below 100 m in 1988, while this value drops to approximately 20 % in 1989. At Tomakstum in 1988 below 100 m these contributions are also small, but surprisingly they seem larger in 1989. Correspondingly large values (35 % at Protection and as high as 80 % at Tomakstum) in the density field for 1989 seem anomalously high. The broad peak reaching 80% at Tomakstum occurs from 120 to 190 m and may not be due to the wind. The longer periods of the wind forcing in 1989 are close to that of the fortnightly tides and the separation of the wind driven component from the deep water renewal is difficult.



Figure 4.24: **1988 Wind vs Selected Wind Driven Density Fluctuations (as** σ_t) Note in the above composite plot, σ_t increases upward and each division is 5 kg/m³. Both wind and densities have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. At Tomakstum the 12 m S4 was not equipped with a conductivity cell and the conductivity sensor of the cyclesonde failed. Depths shown correspond to those chosen for the detided times series presented in figure 4.12.


1989 Filtered Wind and Wind Driven Denisities (as or)

Figure 4.25: **1989 Wind vs Selected Wind Driven Density Fluctuations (as** σ_t) Note in the above composite plot, σ_t increases upward and each division is 5 kg/m³. Both wind and densities have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. The wind influence on the density field is larger than in 1988 and more confined to the surface layer due to the stronger pycnocline. Depths shown correspond to those chosen for the detided times series presented in figure 4.13.



Wind Driven Variance as a Percentage of Total Detided Residual Variance with Depth

Figure 4.26: Wind Driven Variance as a Percentage of Total Detided Residual Variance with Depth Solid lines are values for 1988, speckled for 1989. Note that no density information was available from the cyclesonde at Tomakstum in 1988 due to a failure in its conductivity cell.

However, if the wind is enhancing deep water renewal by helping complete the energy requirements to overcome sill blocking, a large part of the observed density change might be attributed to the wind.

Figures 4.27 and 4.28 show the wind driven along channel velocities and density (as σ_t) fluctuations contoured on axes of depth versus time for Protection during the 1988 experiment. The along channel wind driven velocity shown in figure 4.27 clearly reflects the character of the wind in the near surface during the periodic 2 to 3 day wind forcing from the beginning to about Julian Day 102. Velocities as large as 5 cm/s can be attributed to the wind as deep as 20 to 25 m during this time. During the second half of the record after Julian Day 102, the wind forcing is steadier and is generally up inlet. The near surface layer currents clearly are a reflection of the wind whenever there is a substantial change in the wind however from Day 107 through 109 there is only a small up inlet current despite persisting strong up inlet winds. Perhaps this change is due to the fact that a surface pressure gradient has built up to balance (or nearly balance) the wind stress. During the second half of the record the penetration by the wind is clearly more limited with wind driven velocities of greater than 5 cm/s no deeper than 12 to 15 m. A possible explanation is that the net up inlet winds have retarded the flow of fresher water along the surface lowering the density and buoyancy forces are suppressing the vertical exchange of momentum. Also on Day 105 a sharp increase in river discharge occurs, possibly enhancing the suppression during this time.

At depth the wind response is weaker, but velocities as high as 2.5 cm/s exist throughout the water column particularly during the period of periodic forcing at the beginning of the record. The response is often three layer to up inlet winds and but occasionally appears to be single layer or nearly so (very weak return flows at mid depth if any) during periods of strong down inlet winds such as occur at Julian Days 86 and 88. Persisting down inlet



1988 Protection Wind and Wind Driven Along Channel Current





1988 Protection Wind and Wind Driven Density (as on)



winds such as those seen from Day 96 to Day 101 seem to generate a 3 or perhaps 4 layer response. There is clearly much less wind response in the second half of the record at depth, perhaps because of the stronger stratification perhaps too because a surface pressure gradient has built to balance the wind stress and little baroclinic compensation is being driven as a result.

Figure 4.28 shows that the strongest density response is in the surface layer as expected, but that the response increases in depth from 10 to 12 m at the beginning of the record to 12 to 20 m during the second half of the record. This is likely due to the thickening of the surface layer due to the net up inlet wind stress. Smaller changes are seen at depth perhaps due to the tilting of the isopycnals due to baroclinic compensation.

Figures 4.29 through 4.30 show the wind driven along channel velocities and density (as σ_i) fluctuations contoured on axes of depth versus time for Tomakstum during the 1988 experiment. Figure 4.29 shows that while wind response is present at Tomakstum, it is much more confined to the near surface with the depth of the 5 cm/s contours rarely greater than 10 m. Stratification is stronger at Tomakstum likely due to the fact that it is on the inside of the sill and well away from its associated mixing processes. At depth, in contrast to Protection, there are only weak flows (< 2.5 cm/s) although the shape of the zero velocity contours might suggest that there are similar vertical structures in response to the wind as observed at Protection. Figure 4.30 shows the wind driven density contoured over the first 9 m. The 12 m S4 lacked a conductivity cell in 1988 and the cyclesonde conductivity cell failed. Here, due to the stronger stratification, the wind driven response is much greater than at Protection as noted in the discussion of the time series presented in figure 4.24. However the structure of the wind response in the near surface is similar to that at Protection.



1988 Tomakstum Wind and Wind Driven Along Channel Current

Figure 4.29: 1988 Tomakstum Wind and Wind Driven Along Channel Velocity Contours vs Depth and Time
Wind, wind driven velocity in the near surface and wind driven velocity over the entire water column. Note that the contours are at 0 and +-5, 10, 20, and 30 cm/s in the near surface plot and at 0, +-2.5, 5, 10, 20 and 30 cm/s in the plot of the entire water column. Light shading depicts positive velocities (inflow), dark shading negative velocities (outflow).



1988 Tomakstum Wind and Near Surface Wind Driven Density (as or)

Figure 4.30: **1988 Tomakstum Wind and Wind Driven Density Fluctuations (as** σ_t) **Contours vs Depth and Time** Wind, wind driven density in the near surface to 9 m. The 12 m S4 lacked a conductivity cell and the cyclesonde conductivity cell failed. Note that the contours are at 0 and +-0.2, 0.4, and 0.6 kg/m³ in the near surface plot. Light shading depicts density increases, dark shading density decreasing.

Figures 4.31 through 4.32 show the wind driven along channel velocities and density (as a) fluctuations contoured on axes of depth versus time for Protection during the 1989 experiment. The river discharge is much larger during the 1989 experiment than during the 1988 experiment and low surface salinities noted in section 3.4 result in a pronounced pycnocline even at Protection. The wind forcing is a bit stronger in magnitude than in 1988 and of much longer periods. The wind driven velocity shown in figure 4.31, shows the wind response essentially confined to the near surface (above 25 m) consistent with the observations from 1988 that showed the wind response at Tomakstum much more confined to the surface than at Protection where stratification was weaker. Strong changes in the wind such as the one that occurs at about Julian Day 177 indicate some response (3 layer) as deep as 160 m. Unfortunately the cyclesonde record is short due to an instrument malfunction and whether this is typical over the entire record is not certain. Much of the deeper response however seems to be above sill depth and might be due to baroclinic response to the surface slope setup in the inner basin.

During the period of approximately constant up inlet winds from Julian Day 187 through 196, surface wind driven currents are strong (> 30 cm/s) and then appear to taper off (to 20 cm/s). Unfortunately the winds are not completely steady and a small drop in the magnitude of the up inlet wind forcing at about Day 194 is coincident with near zero outflow, likely due to the surface surface slope adjusting to rebalance with the lower wind stress. The approximate steady down inlet winds from Julian Day 177 through 187 shows a nearly steady decline in the surface wind driven current from approximately 30 cm/s at the start to about 5 cm/s at the end. This decline may be the surface slope coming into balance with the wind stress but the wind also decreases after day 183. Unfortunately no period of precisely steady winds is likely to occur in nature to allow a better interpretation of this behaviour.



1989 Protection Wind and Wind Driven Along Channel Current





1989 Protection Wind and Wind Driven Density (as ot)

Figure 4.32: 1989 Protection Wind and Wind Driven Density Fluctuations (as σ_t) Contours vs Depth and Time Wind, wind driven density in the near surface and wind driven density over the entire water column. Note that the contours are at 0 and +-0.2, 0.5, 1.0, and 2.0 kg/m³ in the near surface plot and at 0 and +- 0.01, 0.02, 0.05, 0.1, 0.2, 0.5, 1.0, and 2.0 kg/m³ for the plot of the entire water column. The lack of data below 12 m in the latter half of a record is due the failure of the cyclesonde to profile during this period. Light shading depicts density increases, dark shading density decreasing.

Figure 4.32 shows the wind driven density fluctuations. Due to the strong stratification, the signal is large with fluctuation s in excess of 2 kg/m^3 in the near surface. What signal there is in the deeper water seems to be mainly confined to above sill depth indicating it may be due to the tilting of the isopycnals along the entire inlet and not just in the outer basin.

Figures 4.33 and 4.34 show the wind driven along channel velocities and density (as σ_t) fluctuations contoured on axes of depth versus time for Tomakstum during the 1989 experiment. Figure 4.33 shows the wind driven velocity is confined to the near surface in a similar manner to Protection, perhaps to a slightly greater extent. The deeper depths exhibit much less vertical structure as seen by the more random meandering of the zero velocity contours. An interesting exception are the inflows and outflows near sill depth at Julian Days 180, 186, 195 and 201. These movements might be wind related, perhaps part of the baroclinic adjustment to the longer period wind forcing in 1989. Wind effects might also enhance or retard the deep water renewal process by helping to complete energy requirements to overcome sill blocking. However these flows are in close antiphase with the spring/neap tidal cycle (peak spring tides occur near Days 185 and 201) and separation of the wind effects from the fortnightly modulation of deep water renewal is not possible without a longer observational period.

Figure 4.34 shows similar and stronger density fluctuations with the wind in the near surface at Tomakstum in 1989 as at Protection. However there is even less wind response in the density fluctuations at depth, although some does appear as deep as 50 m. This type of response was expected in the inner basin but unfortunately we do not have a density record from the previous year at Tomakstum for comparison.



Figure 4.33: 1989 Tomakstum Wind and Wind Driven Along Channel Velocity Contours vs Depth and Time Wind, wind driven velocity in the near surface and wind driven velocity over the entire water column. Note that the contours are at 0 and +-5, 10, 20, and 30 cm/s in the near surface plot and at 0, +-2.5, 5, 10, 20 and 30 cm/s in the plot of the entire water column. Light shading depicts positive velocities (inflow), dark shading negative velocities (outflow).



Figure 4.34: **1989 Tomakstum Wind and Wind Driven Density Fluctuations (as \sigma_t) Contours vs Depth and Time** Wind, wind driven density in the near surface and wind driven density over the entire water column. Note that the contours are at 0 and +-0.2, 0.5, 1.0, and 2.0 kg/m³ in the near surface plot and at 0 and +- 0.01, 0.02, 0.05, 0.1, 0.2, 0.5, 1.0, and 2.0 kg/m³ for the plot of the entire water column. Light shading depicts density increases, dark shading density decreasing.

4.7 Detided and Dewinded Response

The dewinded residual was calculated by subtracting the wind driven component as determined in section 4.6 from the detided residual time series. Given time it is assumed that a surface pressure gradient will build to balance the wind stress and as a result all of the record mean and the trend were preserved in the dewinded residual. Although this technique reduces the variance due to the wind dramatically, it probably does not eliminate all of the energy due to the fact that the original time series are neither continuous or stationary causing errors in the spectral estimates. To remove high frequency noise, the same cosine tapered spectral filter with a cutoff frequency equal to 0.929 cpd used in the presentation of the detided data was applied before plotting the dewinded residual time series.

The results of the dewinding process for the 1988 and 1989 moorings are shown in figures 4.35 through 4.38. These are time series plotted at the same depths and scales as the detided and wind driven time series and against the filtered wind time series for reference. Figure 4.35 clearly shows a mean outflow at 2 m during the 1988 experiment with the Tomakstum mooring showing a continuous outflow over the entire record length but Protection dropping to a near zero outflow in the second half of the record. Both records show a trend of decreasing outflow with time despite the river discharge increasing at around Day 105. A sudden decrease in outflow at Protection and Tomakstum at Day 102 appears to be correlated with the wind. It is worth noting that the runoff and the wind are coupled, at least at the lower frequencies below a period of 7 to 10 days, through the prevailing meteorology and have a correlation of 0.7 for 1988 and 0.6 for 1989 with a lag of 2 to 3 days. As a result, the dewinding process may also be removing part of the estuarine velocity response at these lower frequencies. At 12 m at Protection and 15 m at Tomakstum the two negative peaks (stronger outflows) do occur, one around Julian Day



1988 Filtered Wind and Dewinded Along Channel Velocities

Figure 4.35: **1988 Along Channel Wind vs Selected Dewinded Currents** Note both wind and dewinded currents have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. Depths shown correspond to those chosen for the detided times series presented in figure 4.9.

95 and another at Day 110. These increases are coincident with the spring tides and may be attributed to the volume compensation for the deep water renewal that is enhanced by the additional available kinetic energy. The presence of these effects at 12 metres, well into the surface layer, is surprising. At Protection inflow is initially seen at 150 m and then at 120 m towards the end of the record, likely the signature of the renewal process in the outer basin. At 180 m at Tomakstum, an inflow is seen towards the end of the record, and an increasing return flow is present at 270 m. This may be the signature of the deep water renewal process in the inner basin and a compensating volume flow underneath.

Figure 4.36 show the dewinded velocity time series for the 1989 experiment. The near surface flow at 2 m is outward over the entire record and and with a larger velocity than in 1988, as noted earlier. The river runoff has peaks at around Julian Day 176 and 195. Slight increases in surface outflow can be seen at Tomakstum several days afterward although the signal is very small. It would take approximately 2.7 days for water to travel the 60 km from the head to the Tomakstum mooring at 25 cm/s.

Figure 4.37 and 4.38 show the dewinded density time series for the two experiments. Despite having subjected the records to the dewinding process approximately 50 to 80 percent of the detided variance remains in the 2 m records. In 1988 a trend towards decreasing densities with time is present, but as a sharp increase in river runoff does not occur until Julian Day 105 the estuarine signal may at least be partially obscured by the thickening of the surface layer caused by the net up inlet wind stress. In 1989 decreases in the 2 m density are roughly coincident (when the lag of 2.7 days for along inlet transport is taken into account) with the sharp increases in river discharge at days 175 and 192. The









1988 Filtered Wind and Dewinded Densities (as or)

Figure 4.37: 1988 Wind vs Selected Dewinded Density (as σ_t)

Note in the above composite plot, σ_t increases upward and each division is 5 kg/m³. Both wind and densities have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. At Tomakstum the 12 m S4 was not equipped with a conductivity cell and the conductivity sensor of the cyclesonde failed. Depths shown correspond to those chosen for the detided times series presented in figure 4.12.



1989 Filtered Wind and Dewinded Denisities (as or)

Figure 4.38: 1989 Wind vs Selected Dewinded Density (as σ_t)

Note in the above composite plot, σ_t increases upward and each division is 5 kg/m³. Both wind and densities have been filtered with a spectral filter with a cutoff frequency with a period of 0.929 days. The wind influence on the density field is larger than in 1988 and more confined to the surface layer due to the stronger pychocline. Depths shown correspond to those chosen for the detided times series presented in figure 4.13.

mean river discharge and its variations are much larger in 1989 perhaps making the estuarine signal much easier to detect.

Figures 4.39 and 4.40 present the dewinded velocity and density data contoured against depth and time for the 1988 Tomakstum mooring. Figure 4.39 shows that a constant outflow is present in the near surface and extends to approximately 5 or 6 m and sometimes deeper. An inflow just below the outflow appears centred at about 15 m except for two short periods centred on Julian Days 95 and 110. This surface outflow with an inflow immediately below is certainly suggestive of classical estuarine circulation, with fresher water flowing seaward on the surface and an inflow of salt water underneath to compensate for the entrainment of salt water as the surface flow moves seaward. These depths confirm that the thickness of the surface layer as estimated in section 4.2 (10 m) from the 29.5 day mean velocity was reasonable. Given the trend towards decreasing density with time even at 10 m perhaps the estimate of the density of the inflowing layer water is also reasonable when figure 4.40 is consulted. The velocity of the surface layer at Tomakstum in 1988 as estimated from Knudsen's relations agreed with the 29.5 day mean observations (6.0 cm/s) also suggesting these estimates were representative of the record mean values. There does not appear to be any clear signal of the river discharge in the surface outflow and this is consistent with the results we obtained from our analysis of the 29.5 day mean circulation presented in section 4.2.

At depth from 70 m (approximately sill depth) to 150 m a band of inflow occurs with stronger inflow centred on Days 95 and 110. These stronger flows are approximately in phase with the spring tides and probably show the enhancement of the deep water renewal process by the added kinetic energy of these tides. Given a deep water buoyancy period at Protection of 5 minutes/radian (from section 4.2) and a spring to neap M_2 tidal variation of approximately 12.5 cm/s peak and by applying the blocking formula (h = u / N) of



1988 River Discharge and Tomakstum Dewinded Along Channel Current

Figure 4.39: **1988 River Discharge and Tomakstum Dewinded Along Channel Velocity Contours vs Depth and Time** River discharge, dewinded velocity in the near surface and dewinded velocity over the entire water column. Note that the contours are at 0 and +-5, 10, 20, and 30 cm/s in the near surface plot and at 0, +- 2.5, 5, 10, 20 and 30 cm/s in the plot of the entire water column. Light shading depicts positive velocities (inflow), dark shading negative velocities (outflow).

de Young and Pond(1988) from section 2.2.2, it appears that the added kinetic energy of the peak spring tides may be sufficient to lift renewal water by as much as 38 m.

There appear to be flows back towards the sill both above and below the penetration. These are likely displacements due to the penetration of the new water. The enhancement of deep water renewal on the spring tides also provides a possible explanation for the interruption of the inflow of salt water seen in the estuarine circulation. Due to the enhanced inflow of renewal water during the spring tides and conservation of volume, the salt inflow may be advected seaward (and hence appear as an outflow). The concept of two vertically nested thermohaline cells presented in section 2.0 provides a reasonable interpretation of these observations with conservation of volume serving as a coupling mechanism between them.

Figure 4.40 shows the river discharge and dewinded density at Tomakstum to 9 m. Unfortunately the 12 m S4 lacked a conductivity cell and the cyclesonde conductivity cell failed during this deployment. The trend over the record is towards decreasing salinity in the surface layer, but while this is not inconsistent with increased river discharge, the trend towards lower density appears in advance of the increase of runoff on Day 105. One is left to conclude that the lowering of the density of the surface water is therefore due to the wind. At the beginning of the record a steady decrease in the 22.5 contour may be due to wind mixing from the oscillatory winds during the first half of the record. Due to the vertical exchange of salt, mixing should also raise the density near the surface. There is no evidence of this in the observations, however the upper record is at a depth of 2 m is perhaps too deep to show this effect. During the second half of the record, strong up inlet winds may thicken the upper layer as they retard surface outflow and build a surface slope whose pressure gradient balances the wind forcing. Perhaps the added fresh water from the river also is playing a part in this effect towards the end of the record.



1988 Runoff and Tomakstum Dewinded Near Surface Density (as or)

Figure 4.40: **1988 River Discharge and Tomakstum Dewinded Density** (as σ_t) Contours vs Depth and Time River discharge, dewinded density in the near surface to 9 m. The 12 m S4 lacked a conductivity cell and the conductivity cell on the cyclesonde failed. The 22.5 contour drawn with a heavier line for reference. Figures 4.41 and 4.42 present the dewinded velocity and density data contoured against depth and time for the 1988 Protection mooring. Figure 4.41 shows that the surface outflow does not persist for the entire dewinded record at Protection. This result is somewhat puzzling, particularily as the dewinding process appeared to be reasonably successful at Tomakstum. Surface velocities are larger at Tomakstum than at Protection and perhaps the sudden change in the wind regime is not handled as well by the technique used. To the north of Protection Point, Tribune Channel connects Knight Inlet to other sources of fresh water such as Kingcome Inlet. One might speculate that fresher water entering via Tribune Channel is flowing towards the head of Knight Inlet, with the 'surface'' layer of Knight Inlet being more dense, sinking below. However, there is a general outflow on the surface and an inflow underneath between 15 and 50 m. As at Tomakstum, this inflow appears to disappear periodically likely also due to return flows from penetration of water due to deep renewal in the outer and inner basins.

The depth of the surface layer estimated from the 29.5 day mean velocity in section 4.2 was 12 m. This appears to be an underestimate based on these results and perhaps 25 m would be a better layer thickness estimate. The velocity of the surface layer estimated from Knudsen's relations for a surface layer thickness of 12 m was 16 cm/s considerably larger the observed 29.5 day mean of 4.0 cm/s. Correcting the surface layer thickness for the observed 25 m and using the dewinded density of the inflowing water (as σ_t) of 23.9 (from figure 4.42) would give a surface layer velocity estimate of 5.4 cm/s, close to that observed in the 29.5 day mean. It appears that the major difficulty with using Knudsen's relations for such estimates is in determining an appropriate surface layer thickness to use. At Protection the mixing due to sill processes likely breaks down the initial stratification allowing the wind to mix the surface of the water column more easily. The surface layer thickness can then be expected to vary with time and such difficulties are understandable particularily when data from a single CTD cast are used for the estimate.





At 170 m, there appears to be strong inflows centred on Julian Day 87 and 100 roughly coincident with the neap tides. Not much movement is observed towards the inner sill on Day 95 but movement is seen from 100 to 150 m centred on Day 110. Perhaps the latter is the movement towards the sill caused by the deep water renewal in the inner basin, although the lack of a strong flow on Day 95 is somewhat disappointing. If these movements on the spring tides can be explained by renewal of the inner basin, then perhaps the stronger inflows on the neap tides are penetrations over the outer sill into the outer basin as part of the deep water renewal there. The outer sill is shallow and extended and it is likely that the enhanced mixing due to the energy of the spring tides inhibits and not enhances deep water renewal in the outer basin as found in Indian Arm by de Young and Pond(1988) and Puget Sound by Geyer and Cannon(1982). It seems that Knight Inlet has, at least during part of the year, two distinct partitions in its deep water exchange cycle with the outer basin receiving deep water on the neap tides and the inner basin on the spring tides.

Figure 4.42 shows the dewinded density at Protection in 1988. As with Tomakstum, there is a trend towards a lower density with time in the near surface. However the timing of the decrease suggests that as with the Tomakstum record that it is mostly due secondary wind effects and not due to the direct effects of increased river discharge at the end of the experiment. At depth the rising of the 24.0 and greater isopycnals indicate that renewal is taking place in the outer basin and that greater potential energy is available in the deep water to drive renewal into the inner basin. This increase in energy towards the end of the record is consistent with the larger flow noted on the spring tide in the velocity record for Protection and the evidence of deep water renewal occurring deeper with time in the dewinded record at Tomakstum.



1988 Runoff and Protection Dewinded Density (as or)



Figures 4.43 through 4.44 present the dewinded velocity and density data contoured against depth and time for the 1989 Tomakstum mooring. Figure 4.37 shows near surface outflow for the entire record length to a depth of 5 to 7 m. Just underneath this outflow to a depth of perhaps 15 m (23 m at Julian Day 179) is an inflow again suggestive of classical estuarine circulation and a salt water inflow to compensate for the entrainment of salt water.

While the surface layer thickness of 6 m used in the estimates of surface layer velocity in section 4.2 looks reasonable, the estimate of 11 cm/s from Knudsen's relations and the observed 14.9 cm/s 29.5 day mean are not in close agreement. An inspection of the dewinded density field near 15 m reveals a strong density gradient in this error and perhaps the original estimate of the inflowing layer density (as σ_t) of 22.4 is slightly too large. For example decreasing the inflowing density to 19.8 produces an estimate for U_o from Knudsen's relations of 15 cm/s and this is certainly reasonable when looking at the dewinded densities in figure 4.44. It appears that when stratification is strong producing large density gradients in the near surface deciding on the 'correct' density to represent the inflowing layer is not particularily straight forward.

Between Days 190 and 198 there is a deep outflow although there are two occurrences of velocity zeros, one at day 191 and one at day 197. From 15 m to 150 m there is generally outflow, sometimes stronger than 5 cm/s. These stronger outflows are coincident with the spring tides at Julian Day 172, 185, and 201. The first centred at Day 175 and 70 m depth has only a small volume inflow at 250 to 300 m and perhaps the bulk of the deep water renewal is occurring below the deepest Anderaa at 310 m or cross channel effects are taking the bulk of the inflow to one side of the mooring. At Day 185 another region of outflow occurs at 50 to 100 m and again only a small inflow at 250 to 300 m. At the end of the record another mid depth outflow occurs paired with and preceded by a much larger volume of inflow at depth. If the deep water renewal was initially occurring deeper than the



1989 River Discharge and Tomakstum Dewinded along Channel Currents

Figure 4.43: **1989 River Discharge and Tomakstum Dewinded Along Channel Velocity Contours vs Depth and Time** River discharge, dewinded velocity in the near surface and dewinded velocity over the entire water column. Note that the contours are at 0 and +-5, 10, 20, and 30 cm/s in the near surface plot and at 0, +- 2.5, 5, 10, 20 and 30 cm/s in the plot of the entire water column. Light shading depicts positive velocities (inflow), dark shading negative velocities (outflow).



1989 River Discharge and Tomakstum Dewinded Density (as on)

Figure 4.44: **1989 River Discharge and Tomakstum Dewinded Density** (as σ_t) Contours vs Depth and Time River discharge, dewinded density in the near surface and dewinded density over the entire water column. The 22.0 contour is drawn with a heavier line on both density contour plots for reference.

310 m Anderaa (the actual depth at the Tomakstum mooring was 340 m), then as renewal continued the depth of the penetration might be expected rise as the density of the inner basin increased, thus explaining why the large outflow seen at Day 175 has almost no corresponding inflow while the more modest outflow at the end of the record has a larger corresponding inflow. It may be that renewal is happening on a continuous basis during the 1989 experiment due the high potential energy of the water available at the depth of the inner sill as shown in figure 3.5. The spring tides would then be simply enhancing the flow at these times and accounting for the general outflow over most of the record at 15 to 150 m.

Figure 4.44 shows the dewinded density for Tomakstum. The depth of the 15.0 isopycnal appears to be correlated with the river runoff, becoming shallower with lower discharge and becoming deeper again when the river runoff increases at the end of the experiment. The depth of the 22.0 through 23.5 isopycnals however appear to decrease over time, indicating that the water below about 10 m is becoming more dense. If deep water renewal is occurring one would expect that all of the isopycnals would lift as the deep inflows displace deep water and lift it towards the surface raising the potential energy of the basin. However both the 24.0 and 24.3 isopycnals are almost flat, perhaps due to an error in the inter-calibration of the cyclesonde which profiled only intermittently during the start of the record making conductivity adjustments difficult. This picture is however consistent with the density contours from the deployment and pickup cruise surveys presented in figure 3.4. Most of the renewal appears to occur deeper than the Tomakstum 310 m Anderaa. The maximum depth at this mooring site was 340 m while the deepest part of the inner basin is slightly over 500 m.

Figures 4.45 through 4.46 present the dewinded velocity and density data contoured against depth and time for the 1989 Protection mooring. The second part of the record is



1989 River Discharge and Protection Dewinded Along Channel Current

Figure 4.45: **1989 River Discharge and Protection Dewinded Along Channel Velocity Contours vs Depth and Time** River discharge, dewinded velocity in the near surface and dewinded velocity over the entire water column. Note that the contours are at 0 and +-5, 10, 20, and 30 cm/s in the near surface plot and at 0, +- 2.5, 5, 10, 20 and 30 cm/s in the plot of the entire water column. Light shading depicts positive velocities (inflow), dark shading negative velocities (outflow).

missing below 12 m due to the failure of the cyclesonde to profile during the last half or the experiment. Figure 4.45 shows a continuous near surface outflow between 5 an 10 m at Protection. The thickness of the outflow does vary and might possibly be correlated with the river discharge, with increased discharge giving rise to a thicker surface layer. A reexamination of the Tomakstum record reveals that it also exhibits some of this behaviour, although any such relationship is obscured in the 1988 data perhaps due to the weaker estuarine forcing and the greater mixing due to weaker stratification. Underneath between 10 and 20 m there is generally inflow. Again this structure is consistent with what we would expect from estuarine circulation. The estimates of the surface outflow calculated with Knudsen's relations would almost match the 29.5 day mean observations (7 cm/s) if 10 m instead of 8 m had been taken as the surface layer depth in section 4.2.

At around 100 m there is enhanced flow towards the head of the inlet coincident with both the spring and the neap tides and it is not inconsistent with the analysis from the 1988 data of renewal of the outer basin occurring or being enhanced on the neap tide while renewal of the inner basin occurs or is enhanced on the spring tide. During the neap tide at around Julian Day 180 the inflows appears confined above 120 m and there is some return volume compensation underneath. If this water were penetrating into the outer basin after some renewal had already taken place this behaviour would be expected. Between 25 and 70 m there is an outflow that is likely volume compensation due to deep water renewal.

Figure 4.46 shows the depth of the 15.0 isopycnal may be a function of river runoff as discussed for Tomakstum. The 23.0, 24.8, and 24.85 isopycnals appear to rise with time consistent with rise in the potential energy of the water column due to renewal. However the lack of movement by the intermediate contours is puzzling. The fall in the 24.0 isopycnal is also puzzling but appears to agree with the CTD data. One would expect that the volume compensation outflows occurring at these depths would slowly raise these



1989 Runoff and Protection Dewinded Density (as or)



isopycnals with time as the inner basin water is renewed. Perhaps from sill depth to the bottom of the estuarine circulation vertical shear is mixing some of the surface water down into the deeper water lowering its density as it flows seaward.
Chapter 5

Summary and Conclusions

The analysis presented in Chapter 4.0 attempts to describe the low frequency residual response of a high runoff inlet in terms of the longer period forcing of river runoff, deep water renewal, and wind. Month long observations taken throughout the water column both inside and outside the inner sill of Knight Inlet during the onset of the freshet in 1988 and 1989 were analysed. Both the 29.5 day mean and the detided and dewinded records show vertical structures consistent with two vertically nested thermohaline circulation cells, an estuarine outflow at the surface with an inflow of salt water directly underneath and deep water renewal with flow back towards the inlet mouth both above and below the penetration. These two structures are coupled through conservation of volume and possibly through vertical mixing induced by shear, sill processes, and the wind.

5.1 Estuarine Circulation

The estimated velocity of the surface outflow from Knudsen's relations agreed well with the observed 29.5 day mean velocity vertically integrated over the surface layer providing a representative depth for the surface layer and the density of the inflowing salt water could be determined. Outside the sill, mixing initially from sill processes and then enhanced by the wind, makes the determination of the depth of the surface layer difficult. Inside the sill, at least during high runoff conditions, strong density gradients in the region of the salt water inflow requires careful determination of the the density of the inflowing salt water. The detided and dewinded records show that in simplifying an estuarine system to a two layer representation care must be taken to understand that the depth of the surface layer and the densities of the two layers vary with time. It appears that wind forcing is one of the key contributors to this variance with wind mixing deepening the surface layer by eroding the pycnocline and by the surface setup in response to longer period forcing.

Values for the surface layer velocity determined from Knudsen's relations by van der Baaren(1988) appear reasonable when compared with the observations with the exception of those outside the sill. Also the along inlet surface slopes of Wetton(1981), van der Baaren(1988) and myself determined using dynamic height calculations are consistent with the corresponding river discharge. Therefore van der Baaren's conclusion that the balance of forces along the inlet is between the the surface pressure gradient and the interfacial friction appears to be valid. My 29.5 day observations show a deceleration in the surface outflow from inside to outside the sill. This observation is consistent with the thickening of the surface layer and the vertical exchange of momentum by mixing. It is likely that the higher velocities outside the inner sill estimated by van der Baaren using Knudsen's relations are in error due to the difficulty in estimating surface layer thickness from a single CTD cast. Therefore it is likely that the inertial terms do not play a major role in the balance of forces there and that the balance remains between the surface pressure gradient and the interfacial friction all along the inlet.

Isobaric slopes determined from dynamic height calculations using CTD survey data and using a level of no motion determined from the 29.5 day mean observations show a surface slope in the order of 10 cm/100 km driving the estuarine outflow and small reverse pressure gradients just underneath the surface, consistent with the inflow of salt water being driven by entrainment.

The 29.5 day mean velocity and density profiles suggest that the primary signature of the estuarine forcing is in the density and not the velocity fields. With the river discharge during the 1989 experiment an average of 5.7 times the river discharge during the 1988

174

experiment, the change in the 29.5 day mean density differences between the surface water and the water below the velocity minimum at the Tomakstum mooring was 6.6. While greater surface velocities were observed at both Tomakstum and Protection during 1989, greater shear in the surface layer itself prevented the volume transport from being substantially different. This behavior is consistent with the higher density gradient in 1989 reducing the entrainment of salt through the increased stability in the water column. This result supports the interfacial friction coefficients calculated by van der Baaren(1988) which showed an order of magnitude difference between a low runoff and high runoff regime.

The dewinded velocity time series show no clear correlation between the runoff and surface layer transport during either experiment. If the balance of forces is between the along inlet pressure gradient and the interfacial friction as proposed by van der Baaren this result is not surprising. In both experiments river discharge increased towards the end of the record and the near surface isopycnals deepen with time. However the timing of the variations suggest that secondary effects of the wind such as mixing and the deepening of the surface layer by setup certainly contribute to a large part in this observation. However, during the 1989 experiment when stratification is high and wind mixing suppressed, a fluctuation following the variability in the runoff signature is visible close to the surface. Many of the isopycnals just below the first instruments rising in what appears to be a response to the deep water renewal of the inner basin.

5.2 Deep Water Renewal

The 1988 deployment and mid-experiment and the 1989 deployment and pickup CTD surveys reveal that water of sufficient potential energy to displace inner basin water at least at some depths is present outside the inner sill in both years. The inner basin densities are slightly lower and densities outside the inner sill higher during the 1989 experiment. The

movement of the isopycncals between the two cruises suggest that deep water renewal is taking place in both years, with replenishment at deeper depths in 1989. The 29.5 day mean velocity profiles suggest that deep water renewal is indeed taking place and that return flows due to volume compensation are taking place both above and below the penetrations.

An examination of the dewinded residual shows that deep water renewal in the inner basin of Knight Inlet appears to be triggered or enhanced by the additional kinetic energy of the spring tides. This is consistent with the blocking equation of deYoung and Pond and the control of the renewal process for well stratified inlets with deep sills (i.e. with a Froude number < 1). Inflow movements during the neap tides at Protection during both years suggest that the outer sill which is shallow and extended inhibits renewal on the spring tides. The outer sill, having larger tidal velocities and a lower internal wave speed from reduced stratification (i.e. with a Froude number > 1) has enhanced mixing during the spring tides which reduces the density of the water available for penetration into the outer basin. Thus deep water renewal in the outer basin is favoured during the lower velocity neap tides. The renewal of the deep water of Knight Inlet is a two step process, with the outer basin renewing on the neap tides and the inner basin renewing once the outer basin has filled with suitable source water, most likely on the spring tides.

The dewinded velocity also shows inflow just below the surface layer more or less continuously. Interruptions tend to be synchronous with the enhanced inward flows of deep water renewal on the spring tides suggesting that when the inflow disappears it is actually being advected out of the inlet by the volume compensation due to the deeper penetrations. This behaviour illustrates the coupling that exists between the two vertically nested circulation cells. It is likely that mixing due to vertical shear, sill processes, and even wind also serve to couple these two circulation cells.

176

5.3 Wind

The influence of the wind dominates over all other processes in the surface layer of a stratified water column. At depth the wind response is a function of stratification and it can contribute to significant motion throughout the water column when stratification is weaker during modest runoff. The influence of other mixing processes that break down stratification, such as those in the vicinity of the sill, can cause a major change in wind response. During 1988 at Protection, this wind response has a multi-layered structure but the exact nature of the structure changes with time indicating changes in phase response. Wind driven currents greater than 2.5 cm/s were observed at the bottom of the water column. In 1989, an extremely sharp pycnocline effectively limits the direct influence of the wind to the surface layer and the vertical structure of the wind response at depth is much less defined. The baroclinic response to wind setup is consistent with Buckley and Pond(1976) who proposed that response time to wind stress in a stratified water column is dominated by the baroclinic adjustment and not the barotropic adjustment of sea surface slope alone.

At Tomakstum in 1989 there is a large contribution by the wind to the detided density variance between 150 and 200 m. It is likely that the wind is aiding the deep water renewal process through the tilting of the isopycnals. However the time scales of the 1989 wind forcing and the deep water renewal are similar and the certainty of the analysis could be improved by longer records

van der Baaren(1988) found that interfacial friction was a function of wind. She also found that the along inlet balance of forces was between the surface pressure gradient and the interfacial friction. With winds of a shorter period than the baroclinic adjustment time, mixing and the work done to erode the pycnocline may be a reasonable explanation for this

177

dependance. For winds which persist for a longer period of time, the balance of forces may be adjusting to the wind setup in a manner similar to that observed with the estuarine forcing.

Bibliography

- Anderaa Instruments 1983. Operating Manual for RCM 4/5 Recording Current Meter, Technical Description 119. 69 pp.
- Buckley, J. R. 1971. The Currents Winds and Tides of Northern Howe Sound., PhD. Thesis, The University of British Columbia. 228 pp.
- Buckley, J. R. and Pond, S. 1976. Wind and the Surface Circulation of a Fjord., Journal of the Fisheries Research Board of Canada, 33(10), pp. 2265-2271.
- Burling, R.W. 1982. Deep Circulations in Inlets: Indian Arm as a Case Study., in Marine Tailings Disposal, D. V. Ellis ed., Ann Arbour Science, Ann Arbour Michigan, pp. 85-132.
- Canadian Hydrographic Service Marine Science Branch 1969. Chart 3578 Knight Inlet, British Columbia. Department of Energy, Mines, and Resources, Ottawa, Canada.
- Canadian Hydrographic Service Marine Science Branch 1973. Chart 3596 Port Harvey to Queen Charlotte Strait, British Columbia. Department of Energy, Mines, and Resources, Ottawa, Canada.
- Canadian Tide and Current Tables, Pacific Coast, Volume 6 1988. Department of Fisheries and Oceans, Ottawa, Canada. 73 pp.
- Canadian Tide and Current Tables, Pacific Coast, Volume 6 1989. Department of Fisheries and Oceans, Ottawa, Canada. 73 pp.
- de Young, B. and Pond, S. 1987. The Internal Tide and Resonance in Indian Arm, British Columbia. Journal of Geophysical Research. 92(C5), pp. 5191-5207.
- de Young, B. and Pond, S. 1988. The Deepwater Exchange Cycle in Indian Arm, British Columbia., Estuarine and Coastal Shelf Science, 26, pp. 285-308.
- de Young, B and Pond, G. S. 1989. Partition of Energy Loss From the Barotropic Tide in Fjords. Journal of Physical Oceanography, 19(2), pp. 246 252.
- **Dunbar, D. S., 1985.** A Numerical Model of Stratified Circulation in a Shallow Silled Inlet. Ph.D. Thesis, University of British Columbia. 273 pp.
- **Dunbar, D. S. and Burling, R. W. 1987.** A Numerical Model of Stratified Circulation in Indian Arm, British Columbia. Journal of Geophysical Research. 92(C12), pp. 13,075-13,105.
- Farmer, D. M. 1972. The Influence of Wind on the Surface Waters of Alberni Inlet. Ph.D. Thesis, University of British Columbia. 92 pp.
- Farmer, D. M. 1976. The Influence of the Wind on the Surface Layer of a Stratified Inlet. Part II. Analysis. Journal of Physical Oceanography, 6(6), pp. 941-952.
- Farmer, D. M. and Huppert, H. E. 1979. The Oceanography of Fjords. Nature 280, 26 July 1979, pp. 273 280.

- Farmer, D. M. 1982. Stratified Flow Over Sills. In: Coastal Oceanography. Gade, Edwards, and Svendsen eds. pp. 337 362.
- Farmer, D. M. and Freeland, H. J. 1983. The Physical Oceanography of Fjords. Progress in Oceanography. 12(2), pp. 147-219.
- Farmer, D. M. and Smith, J. D. 1980a. Tidal Interaction of Stratified Flow with a sill in Knight Inlet. Deep Sea Research. 27A, pp. 239-254.
- Farmer, D. M. and Smith, J. D. 1980b. Generation of Lee Waves over the Sill in Knight Inlet. In: Fjord Oceanography, H. J. Freeland, D. M. Farmer, and C. D. Levings, eds., Plenum Press, New York, pp. 259 269.
- Foreman, M.G.G. 1977. Manual for Tidal Heights Analysis and Prediction. Pacific Marine Science Report 77-10, Institute of Ocean Sciences, Patricia Bay, B. C. 101 pp.
- Foreman, M.G.G. 1979. Manual for Tidal Currents Analysis and Prediction. Pacific Marine Science Report 78-6, Institute of Ocean Sciences, Patricia Bay, B. C. 101 pp.
- Freeland, H. J. 1980. The Hydrography of Knight Inlet, B.C., in the Light of Long's Model of Fjord Circulation. In: Fjord Oceanography, H. J. Freeland, D. M. Farmer, and C. D. Levings, eds., Plenum Press, New York, pp. 271 - 276.
- Freeland, H. J. and Farmer, D. M. 1980. Circulation and Energetics of a Deep, Strongly Stratified Inlet. Canadian Journal of Fisheries and Aquatic Sciences. 37(9), pp. 1398-1410.
- Gargett, A. E. 1980. Turbulence Measurements through a Train of Breaking Internal Waves in Knight Inlet, B. C. In: Fjord Oceanography, H. J. Freeland, D. M. Farmer, and C. D. Levings, eds., Plenum Press, New York, pp. 277 - 281.
- Gargett, A. E. and Holloway G. 1984. Dissipation and Diffusion by Internal Wave Breaking. Journal of Marine Research, 42, pp. 15 27.
- Gartner, J. W. and Oltmann, R. N. 1985. Comparison of Recording Current Meters used for Measuring Velocities in the Shallow Waters of San Francisco Bay, California. MTS/IEEE Conference Record of Ocean Engineering and the Environment. II, pp. 731-737.
- Geyer, W. R. and Cannon, G. A. 1982. Sill Processes Related to Deep Water Renewal in a Fjord., Journal of Geophysical Research, 87(C10), pp. 7985-7996.
- Godin, G. 1972. The Analysis of Tides. University of Toronto Press, Toronto, Canada. 264 pp.
- Griffin, D. and LeBlond, P. H. 1990. Estuary/Ocean Exchange Controlled by Spring-Neap Tidal Mixing., Estuarine, Coastal and Shelf Science, 30, pp. 275-297.
- Guildline Instruments. 1979. Model 8705 Digital CTD Probe and Model 87102 Control Unit, Technical Manual, Vol. 1. 117 pp.
- Hamming, R. W. 1983. Digital Filters, 3rd Edition. Prentice-Hall, Englewood Cliffs, N.J. 284 pp.

- InterOcean Systems Inc. 1987. User's Manual for the S4 Current Meter., San Diego, 52 pp.
- Jamart, B. M. and Winter, D. F. 1980. Finite Element Computation of the Barotropic Tides in Knight Inlet, B. C., In: Fjord Oceanography, H. J. Freeland, D. M. Farmer, and C. D. Levings, eds., Plenum Press, New York. pp. 283 - 289.
- Kanwisher, J, and Lawson, K. D. 1975. Electromagnetic Flow Sensors. Limnology and Oceanography. 20(2) March 1975. pp. 174-182.
- Karl, J. H. 1989. An Introduction to Digital Signal Processing. Academic Press, San Diego. 341 pp.
- Kollstad, T. F. and Hansen, S. E. 1985. An Investigation of the Performance of 4 Different Current Meters When Applied in the Wave Zone. In: Advances in Underwater Technology and Offshore Engineering Vol. 4; Evaluation, Comparison and Calibration of Oceanographic Instruments. Proceedings of an International Conference (Ocean Data). Society for Underwater Technology, London, England. pp. 223-245.
- Klinck, J. M., O'Brien, J. J., and Svendsen H. 1981. A Simple Model of Fjord and Coastal Circulation Interaction. Journal of Physical Oceanography. 11(12), pp. 1612-1626.
- Large, W.G. and Pond, G.S., 1981. Open Ocean Flux Measurements in Moderate to Strong Winds, Journal of Physical Oceanography 11(3), pp. 324-336.
- Lawson, K. D. et al. 1983. The Development of a Spherical, Electromagnetic Current Meter. In: Proceedings of Oceans 1983, Vol. 1 Technical Papers. Institute of Electrical and Electronics Engineers, Piscataway, New Jersey. pp. 187-193.
- Long, R. R. 1975b. Circulations and Density distributions in a Deep Strongly Stratified, Two Layer Estuary. Journal of Fluid Mechanics, 71, pp. 529 - 540.
- Long, R. R. 1980. The Fluid Mechanical Problem of Fjord Circulations. In: Fjord Oceanography, H. J. Freeland, D. M. Farmer, and C. D. Levings, eds., Plenum Press, New York, pp. 67 116.
- Marine Profiles Inc. 1979. Operation Manual, Cyclesonde PCM-1 Profiling Current Meter. 117 pp.
- Miles, J. W. 1961. On the Stability of Heterogeneous Shear Flows. Journal of Fluid Mechanics, 10, pp. 496- 508.
- Nowak, L. P. (in prep). A Numerical Model of Stratified Circulation in Inlets Using a Simple Barotropic Baroclinic Mode Splitting Scheme. Manuscript Report, Department of Oceanography, University of British Columbia.
- Pearson, C. E. and Winter, D. F. 1978. Two-Layer Analysis of Steady Circulation in Stratified Fjords. In: Hydrodynamics of Estuaries and Fjords, J. Nihoul ed., Elsevier Oceanographic Series, pp. 495 - 514.

- Pederson, F. B. 1978. A Brief Review of Present Theories of Fjord Dynamics. In: Hydrodynamics of Estuaries and Fjords, J. Nihoul ed., Elsevier Oceanographic Series, pp. 407 - 422.
- Pickard, G. L. 1953. Oceanography of British Columbia Mainland Inlets. II. Currents. Fisheries Research Board of Canada, Pacific Progress Report, 97, pp 12-13.
- Pickard, G. L. 1954. Oceanography of British Columbia Mainland Inlets. III. Internal Waves. Fisheries Research Board of Canada, Pacific Progress Report, 98, pp. 13-16.
- Pickard, G. L. 1956. Physical Features of British Columbia Inlets. Transactions of the Royal Society of Canada. L(III), pp. 47-58.
- Pickard, G. L. and Trites, R. W. 1957. Fresh Water Transport Determination from the Heat Budget with Applications to British Columbia Inlets. Journal of the Fisheries Research Board of Canada. 14(4), pp. 605-616.
- Pickard, G. L. and Rodgers, K. 1959. Current Measurements in Knight Inlet, British Columbia. Journal of the Fisheries Research Board of Canada. 16(5), pp. 635-678.
- Pickard, G. L. 1961. Oceanographic Features of Inlets in the British Columbia Mainland Coast. Journal of the Fisheries Research Board of Canada. 18(6), pp. 907-999.
- Pickard, G. L. 1975. Annual and Longer Term Variations of Deepwater Properties of the Coastal Waters of Southern B. C., Journal of the Fisheries Research Board of Canada. 32, pp. 1561-1587.
- Pickard, G. L. and Stanton, B. R. 1980. Pacific Fjords A Review of Their Water Characteristics, In: Fjord Oceanography, H.J. Freeland, D.M. Farmer, and C.D. Levings, eds., Plenum Press, New York, pp. 1 - 51.
- Pickard, G. L. and Emery, W. J. 1982. Descriptive Physical Oceanography. Pergammon Press, Oxford. 249 pp.
- Pond, G. S. and Pickard, G. L. 1983. Introductory Dynamical Oceanography, 2nd Edition. Pergammon Press, Oxford. 329 pp.
- Press, W. H., B. P. Flanners, S. A. Teukolsky, and V. T. Vetterling 1988. Numerical Recipes in C, The Art of Scientifc Computing. Cambridge University Press. 735 pp.
- Priestley, M. B. 1981. Spectral Analysis and Time Series Vol. 1 and 2. Academic Press, San Diego. 890 pp.
- Rattray, M. Jr. 1967. Some Aspects of the Dynamics of Circulation in Fjords., In: Estuaries, G. H. Lauff, ed., American Association for the Advancement of Science, Publication No. 83, pp. 52-63.
- Robinson, E. A. and Treitel, S. 1980. Geophysical Signal Analysis. Prentice Hall, Englewood Cliffs, New Jersey. 466 pp.

- Smith, J. D. and Farmer, D. M. 1977. Nonlinear Waves and Internal Hydraulic Jumps in a Fjord, Geofluidynamical Wave Mathematics: Research Contributions., Applied Mathematics Group, University of Washington, Seattle
- Smith, J. D. and Farmer, D. M. 1980. Mixing Induced by Internal Hydraulic Disturbances in the Vicinity of Sills. In: Fjord Oceanography, H. J. Freeland, D. M. Farmer, and C. D. Levings, eds., Plenum Press, New York, pp. 251 257.
- Stacey, M. W. 1984. The Interaction of the Tides with the Sill of a Tidally Energetic Inlet. Journal of Physical Oceanography, 14, pp. 1105-1117.
- Stacey, M. W. 1985. Some Aspects of the Internal Tide in Knight Inlet, B. C., Journal of Physical Oceanography, 15, pp. 1652-1661.
- Stacey, M. W. and Pond, S. (1992). A Numerical Model of the Internal Tide in Knight Inlet, British Columbia. Atmosphere-Ocean (in press).
- Surface Water Data, British Columbia, 1988. Inland Waters Directorate, Water Resources Branch, Water Survey of Canada, Ottawa, Canada. 300 pp.
- Surface Water Data, British Columbia, 1989. Inland Waters Directorate, Water Resources Branch, Water Survey of Canada, Ottawa, Canada. 300 pp.
- Thomson, R. E. 1981. The Oceanography of the British Columbia Coast, Canadian Special Publication of Fisheries and Aquatic Sciences 56, Department of Fisheries and Oceans, Canada, 291 pp.
- University of British Columbia, 1988. Department of Oceanography, British Columbia Inlets Cruises, Data Report #57. 54 pp.
- University of British Columbia, 1989. Department of Oceanography, British Columbia Inlets Cruises, Data Report #58. 54 pp.
- University of British Columbia, (in prep). Department of Oceanography, Knight Inlet 1988/89 Experiments, Data Report.
- van der Barren, A., 1988. Surface Layer Dynamics and a Study of Finestructure in Knight Inlet, British Columbia., M.Sc. Thesis, The University of British Columbia, 107 pp.
- van Leer, J. et al. 1974. The Cyclesonde: an Unattended Vertical Profiler for Scalar and Vector Quantities in the Upper Ocean. Deep Sea Research. 21 pp. 385-400.
- van Leer, J. 1980. Profiling Devices, in Air/Sea Interaction. Dobson, Masse, and Davis eds. GC 190 5A37 1980. pp. 701-724.
- Webb, A. J. 1985. The Propagation of the Internal Tide around a Bend in Knight Inlet, B. C., Ph.D. Thesis, The University of British Columbia. 160 pp.
- Webb, A. J. and Pond S. 1986a. The Propagation of a Kelvin Wave Around a Bend in a Channel. Journal of Fluid Mechanics. 169, pp. 257-274.

- Webb, A. J. and Pond S. 1986b. A Modal Decomposition of the Internal Tide in a Deep, Strongly Stratified Inlet: Knight Inlet, British Columbia. Journal of Geophysical Research. 91(C8), pp. 9721-9738.
- Wetton, B., 1981. Dynamics of the Surface Layer: Knight Inlet, British Columbia. unpublished report of the Physics Co-op Work Project, University of Victoria, pp. 68.
- Winter, D. F. 1973. A Similarity Solution for Steady-State Gravitational Circulation in Fjords. Estuarine and Coastal Marine Science. I, pp. 387-400.