Climate Variability and Change Impacts on Coastal Environmental Variables in British Columbia Canada

Dilumie Saumedaka Abeysirigunawardena
B.Sc. (Eng.), University of Peradeniya, Sri Lanka, 1994
M.Sc.(Coastal Eng.), IHE, Delft, the Netherlands, 1999

A Dissertation Submitted in Partial Fulfillment
of the Requirements for the Degree of

DOCTOR OF PHILOSOPHY

in the Department of Geography

© Dilumie Abeysirigunawardena, 2010
University of Victoria

All rights reserved. This thesis may not be reproduced in whole or in part, by photocopy or other means, without the permission of the author.
Climate Variability and Change Impacts on Coastal Environmental Variables in British Columbia Canada

by

Dilumie Saumedaka Abeysirigunawardena
B.Sc. (Eng.), University of Peradeniya, Sri Lanka, 1994
M.Sc. (Coastal Eng.), IHE, Delft, The Netherlands, 1999

Supervisory Committee

Dr. Dan J. Smith, Department of Geography, University of Victoria
Supervisor

Dr. Chris Houser, Department of Geography, University of Victoria
Member

Dr. Eric Kunze, School of Earth and Ocean Sciences, University of Victoria
Outside Member

Dr. Stephane Mazzotti, School of Earth and Ocean Sciences, University of Victoria
Outside Member
Abstract

Supervisory Committee

Dr. Dan J. Smith, Department of Geography, University of Victoria
Supervisor
Dr. Chris Houser, Department of Geography, University of Victoria
Member
Dr. Eric Kunze, School of Earth and Ocean Sciences, University of Victoria
Outside Member
Dr. Stephane Mazzotti, School of Earth and Ocean Sciences, University of Victoria

The research presented in this dissertation attempted to determine whether climate variability is critical to sea level changes in coastal BC. To that end, a number of statistical models were proposed to clarify the relationships between five climate variability indices representing large-scale atmospheric circulation regimes and sea levels, storm surges, extreme winds and storm track variability in coastal BC. The research findings demonstrate that decadal to interdecadal climatic variability is fundamental to explaining the changing frequency and intensity of extreme atmospheric and oceanic environmental variables in coastal BC. The trends revealed by these analyses suggest that coastal flooding risks are certain to increase in this region during the next few decades, especially if the global sea-levels continue to rise as predicted. The outcome of this study emphasis the need to look beyond climatic means when completing climate impact assessments, by clearly showing that climate extremes are currently causing the majority of weather-related damage along coastal BC. The findings highlight the need to derive knowledge on climate variability and change effects relevant at regional to local scales to enable useful adaptation strategies.

The major findings of this research resulted in five independent manuscripts: (i) Sea level responses to climatic variability and change in
Northern BC. The Manuscript (MC) is published in the Journal of atmospheric and oceans (AO 46 (3), 277-296); (ii) Extreme sea-level recurrences in the south coast of BC with climate considerations. This MC is in review with the Asia Pacific Journal of Climate Change (APJCC); (iii) Extreme sea-surge responses to climate variability in coastal BC. This MC is currently in review in the Annals of the AAG (AN-2009-0098); (iv) Extreme wind regime responses to climate variability and change in the inner-south-coast of BC. This MC is published in the Journal of Atmosphere and Oceans (AO 47 (1), 41-62); (v) Sensitivity of winter storm track characteristics in North-eastern Pacific to climate variability. This manuscript is in review with the Journal of Atmosphere and Oceans (AO (1113)). The findings of this research program made key contributions to the following regional sea level rise impact assessment studies in BC: (i) An examination of the Factors Affecting Relative and Absolute Sea level in coastal BC (Thomson et al., 2008). (ii) Coastal vulnerability to climate change and sea level rise, Northeast Graham Island, Haida Gwaii (formally known as the Queen Charlotte Islands), BC (Walker et al., 2007). (iii) Storm Surge: Atmospheric Hazards, Canadian Atmospheric Hazards Network - Pacific and Yukon Region, C/O Bill Taylor.
Table of Contents

Supervisory Committee ........................................................................................................ii
Abstract ................................................................................................................................iii
Table of Contents ..................................................................................................................v
List of Tables.........................................................................................................................x
List of Figures ......................................................................................................................xiv
Acknowledgments ................................................................................................................xxii

1.0 Introduction .........................................................................................................................1
  1.1 Observed sea level changes and future trends ...............................................................2
  1.2 From Global to Regional Sea level Trends and Uncertainties .....................................3
  1.3 Climate Impacts on Canadian coasts from National to Regional scale ......................4
  1.4 Climate Variability and Change Impacts in coastal British Columbia .....................6
  1.5 The Study Region .........................................................................................................8
  1.6 Research purpose and objectives ................................................................................9
References: ..........................................................................................................................13

2.0 Sea level responses to climatic variability and change in Northern British Columbia Canada ........................................................................................................20
  2.1 Abstract .......................................................................................................................20
  2.2 Introduction ................................................................................................................21
  2.3 Study Region ................................................................................................................27
  2.4 Methods .......................................................................................................................29
    2.4.1 Water level data ...................................................................................................29
    2.4.2 Climatic variability indices ..................................................................................30
    2.4.3 Statistical methods .............................................................................................32
  2.5 Results and Discussion .................................................................................................36
    2.5.1 Long-term sea level (MSL) trends .................................................................36
    2.5.2 Extreme sea level (MAXSL) trends ...............................................................37
### 2.5.3 Return periods of extreme sea level (MAXSL) events

- Page 39

### 2.5.4 Sea level responses to climatic variability events

- Page 41

### 2.5.5 Non-linear relations between sea levels and climate variability indices

- Page 43

### 2.5.6 Sea level response to climate regime shifts

- Page 45

### 2.6 Conclusions

- Page 46

### References

- Page 49

### 3.0 Extreme Sea-level recurrences in the South Coast of British Columbia Canada with Climate Considerations

- Page 74

#### 3.1 Abstract

- Page 74

#### 3.2 Introduction

- Page 74

#### 3.3 Problem definition

- Page 76
  - 3.3.1 Tide Gauge Data
  - Page 76
  - 3.3.2 Data Pre-processing
  - Page 77
  - 3.3.3 Long term sea-level trends
  - Page 78
  - 3.3.4 Climate Variability Indices
  - Page 78

#### 3.4 Methodology

- Page 80
  - 3.4.1 Generalized Extreme-value (GEV) distribution
  - Page 80
  - 3.4.2 Parameter estimation method
  - Page 81
  - 3.4.3 Return Levels (Quantiles)
  - Page 83
  - 3.4.4 Simulate the effect of relative sea-level trends on TWL extremes
  - Page 83
  - 3.4.5 Simulate the effect of climate variability on TWL extremes
  - Page 84

#### 3.5 Results

- Page 86
  - 3.5.1 Extreme TWL recurrences without climate considerations (BaseModel)
  - Page 86
  - 3.5.2 Effect of relative sea-level rise on total water-level extremes
  - Page 87
  - 3.5.3 Effect of Climate Variability on total water-level extremes
  - Page 89
  - 3.5.4 Coastal flooding and extreme sea-surge tide interactions
  - Page 93

#### 3.6 Discussion

- Page 94

### References

- Page 97
4.0 Extreme Sea-Surge Responses to Climate Variability in Coastal
British Columbia Canada.................................................................119
4.1 Abstract ..................................................................................119
4.2 Introduction ..............................................................................119
4.3 Data Pre-processing: .................................................................121
  4.3.1 Tide Gauge Data Quality Assurance ....................................121
  4.3.2 Sea-Surge Computations ....................................................122
  4.3.3 Climate Variability Indices ..................................................123
4.3 Methodology ...........................................................................126
  4.4.1 Generalized Extreme-value (GEV) distribution .................126
  4.4.2 Simulate the effect of climate variability on sea-surge extremes via
covariates ................................................................................128
4.5 Results. ...................................................................................131
  4.5.1 Sea surge dependencies in coastal British Columbia ..........131
  4.5.2 Extreme sea-surge exceedances in coastal British Columbia ..132
  4.5.3 Spatial dependencies of extreme sea surges in coastal British
       Columbia. ............................................................................134
  4.5.4 Coastal flooding and surge- tide interactions .......................136
4.6 Discussion and Conclusions ......................................................137
References .....................................................................................140

5.0 Extreme wind regime responses to climate variability and change in
the inner-south-coast of British Columbia.......................................164
5.1 Abstract ..................................................................................164
5.2 Introduction ..............................................................................164
5.3 Problem Definition .................................................................166
5.4 Data ........................................................................................167
  5.4.1 Directional wind records ....................................................167
  5.4.2 Climate Variability Indices ..................................................168
5.4 Methodology ...........................................................................171
5.5.1 The extreme value approach.............................................................172
5.5.2 Selection of Thresholds.................................................................173
5.5.3 Parameter estimation method .......................................................174
5.5.4 Return Levels (Quantiles).............................................................176
5.5.5 The question of statistical independence.......................................177
5.5.6 Simulate the effect of climate variability on wind extremes via covariates.................................................................179
5.5.7 The question of co-linearity between climate indices .................180
5.5.8 Conditional Extreme Value Models ..............................................183
5.6 Results .............................................................................................184
5.6.1 Extreme non-directional and directional wind recurrences ...........184
5.6.2 Effect of Climate Variability on extreme winds............................186
5.7 Effect of Global Climate Change on extreme winds. .....................189
5.8 Discussion and Conclusions............................................................191
References............................................................................................196

6.0 Influence of Climate Variability and Change on winter-storm track characteristics in North-eastern Pacific..........................225
6.1 Abstract ..........................................................................................225
6.2 Introduction......................................................................................226
6.3 Observational evidence of storm track characteristics in coastal British Columbia..............................................................228
6.3.1 A review on storm track characteristics in coastal British Columbia .. .........................................................................................228
6.3.2 Pineapple-express storms vs. Climate Variability.......................229
6.3.3 The effects of global climate change on storm tracks.................230
6.4 Data.................................................................................................232
6.4.1 NCEP/NCAR re-analysis storm track projections..........................232
6.4.2 Climate variability modes.............................................................233
6.5 Methodology....................................................................................235
<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.5.1 Selecting and characterizing storm tracks</td>
<td>235</td>
</tr>
<tr>
<td>6.5.2 Statistical methods</td>
<td>236</td>
</tr>
<tr>
<td>6.6 Results</td>
<td>238</td>
</tr>
<tr>
<td>6.6.1 Interannual to Decadal changes in Storm track characteristics</td>
<td>238</td>
</tr>
<tr>
<td>6.6.2 Storm track characteristics and Climate Variability</td>
<td>243</td>
</tr>
<tr>
<td>6.6.3 Pineapple-Express Storms and Climate Variability</td>
<td>246</td>
</tr>
<tr>
<td>6.7 Discussion &amp; Conclusions</td>
<td>248</td>
</tr>
<tr>
<td>References</td>
<td>253</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.0 Conclusions</td>
<td>285</td>
</tr>
<tr>
<td>7.1 Introduction</td>
<td>285</td>
</tr>
<tr>
<td>7.2 Major research findings and contributions</td>
<td>287</td>
</tr>
<tr>
<td>7.2.1 Sea-level response trends to long term climate change and short term climate variability in Northern British Columbia</td>
<td>287</td>
</tr>
<tr>
<td>7.2.2 Development of extreme sea level return-periods for the south coast of British Columbia, with climate considerations</td>
<td>289</td>
</tr>
<tr>
<td>7.2.3 The spatial and temporal distribution of extreme sea surge recurrences and its sensitivity to climate variability in British Columbia</td>
<td>290</td>
</tr>
<tr>
<td>7.2.4 Extreme wind regime recurrences and its sensitivity to climate variability in the southern coast of British Columbia</td>
<td>291</td>
</tr>
<tr>
<td>7.2.5 Establishment of winter storm track characteristics and its relationship to various climate variability signals in coastal British Columbia</td>
<td>292</td>
</tr>
<tr>
<td>7.3 Recommendations for future works</td>
<td>294</td>
</tr>
<tr>
<td>References</td>
<td>299</td>
</tr>
</tbody>
</table>
List of Tables

Table 2-1 Pearson’s correlation coefficients for relations between annual average MEI, PDO, ALPI, NOI, and MSL (Prince Rupert) values. .......................................................... 57

Table 2-2 Multiple regression fit statistics for MSL vs MEI, ALPI, NOI and PDO, showing the overall model fit ($R^2$) and the significance of each climate control as predictors. ........................................................................................................................ 58

Table 2-3 Pearson’s correlation coefficients for winter seasonal average (November – March) values of MEI, PDO, NOI and MSL (at Prince Rupert).............. 59

Table 2-4 Pearson’s correlation coefficients for summer seasonal average (March – August) values of MEI, PDO, NOI, and MSL (at Prince Rupert)............ 60

Table 3-1 : 21st century relative sea-level rise trend projections for the Fraser delta (Church, 2002). .................................................................................................................. 101

Table 3-2 : Average climate indices for “strong El Niño”, “strong La Niña” and “neutral” years. The definitions of strong El Niño / La Niña years are based on the Environment Canada (Meteorological Services of Canada-The Green Land) classification scheme. ........................................................................................................ 102

Table 3-3 : Maximum-likelihood fitted parameters for the historical annual maximum total water-levels GEV fit at Point Atkinson (Base Model: B)........... 103

Table 3-4 : Return levels and 95% confidence intervals at Pt. Atkinson. Projections are based on the GEV fit of historical annual maximum total water-levels (Base).................................................................................................................. 104

Table 3-5: Maximum-likelihood fitted parameter values of the GEV fit: (a) the historical annual maximum total water-levels at Point Atkinson in combination with 0.28 cm/yr sea-level rise (B+SLR1) (b) historical annual maximum total water-levels at Point Atkinson in combination with 0.57 cm/yr sea-level rise (B + SLR2).............................................................................................................................................. 105

Table 3-6 : Return levels under assumed relative sea-level rise trends. The return level projections for the base model are indicated for comparison purposes... 106

Table 3-7 : Results of the redundancy test on annual maximum TWL data at Pt. Atkinson. The test results shows that actual physical significance of NOI as a covariate in the location parameter far outweigh the presence of other climate indices in the statistical model. .................................................................................................................. 107
Table 3-8: The extreme total water-level recurrences at Point Atkinson with climate considerations. .......................................................................................................................... 108

Table 3-9: Sea-surge and Astronomical tide occurrences for the 50 highest recorded surges at the 11-tide gauge locations (results expressed to 0 decimals). The results show significant number of extreme sea-surges coinciding with mid to low tides in coastal BC. The highlighted data corresponds to Pt. Atkinson (7795). ........................................................................................................................................ 109

Table 3-10: The surge-tide combination of the highest ever occurred sea surges and highest ever occurred TWL events at 11 tide gauge station in coastal BC. Note that the tide levels are indicated in terms of absolute values and as a percentile with respect to the 2004 astronomical tide levels. The highlighted data corresponds to Pt. Atkinson (7795). ........................................................................................................................................ 110

Table 4-1: Pacific Region tide gauge stations adjacent to the coast of BC with at least 20 years of data. ................................................................................................................................................................................................. 145

Table 4-2: Application of the redundancy test on annual maximum sea-surge data at station 7120................................................................................................................................................................................................................................................................. 146

Table 4-3: Average climate indices for “strong El Niño”, “strong La Niña” and “neutral” years. The definitions of strong El Niño / La Niña years are based on the Environment Canada (Meteorological Services of Canada-The Green Lane) classification scheme. ................................................................................................................................................................................................. 147

Table 4-4: Overall averages of the (a) monthly maximum sea-surges and (b) monthly mean sea-surges recorded to-date at the 11-tide gauge stations in coastal BC, indicating a seasonal preference (October to March). ................................................................................................................................................................................................. 148

Table 4-5: Pearson product-moment correlation coefficients of annual maximum sea surges from 1950-2007 indicating two distinctly different spatial dependencies (demarcated by dashed rectangles) among stations in coastal BC. ................................................................................................................................................................................................. 149

Table 4-6: Projected GEV Coefficients for the Base Model at each station. ..... 150

Table 4-7: Extreme Residual recurrences in coastal BC with no climate considerations (Base Model). ................................................................................................................................................................................................................................................................. 151

Table 4-8: Climate covariate coefficients expressed to the first decimal when applied as individual covariates. The positive/negative (+/-) sign in each cell indicate the relationship (i.e. trend) between the model parameter and the climate covariate. The shaded cells indicate the coefficients that are significant at 95% level). ................................................................................................................................................................................................................................................................. 152
Table 4-9 : Climate covariates coefficients for the final models at each tide-gauge station, selected from the redundancy test. ................................................................. 153

Table 4-10 : Station specific extreme sea surge occurrences having 1% exceedance in a given year with climate considerations. .............................................. 154

Table 4-11 : The sea-surge characteristics and the projected annual percentage exceedance (highlighted columns) observed at each tide gauge station at the time of the strongest ever extreme sea surge event occurrence in southern (16th December 1982) and northern (24th December 2003) BC coasts. Results support the conclusion of weak correlation between the two regions. .............. 155

Table 4-12 : The sea-surge count distribution vs. the percentile astronomical tide levels for the 50-highest surges on record at the 11-tide gauges. Results show significant number of extreme sea-surges coinciding with mid to low tides in coastal BC. ........................................................................................................................................ 156

Table 4-13 : The surge-tide levels of the highest sea surge and highest total water level event on record at each tide gauge station in coastal BC. Note that the tide levels are indicated both in terms of absolute values and as a percentile with respect to the 2004 astronomical tide levels. ......................................................................................................................... 157

Table 5-1 : Dominant directional wind sectors in the in Inner-south-coast of BC. Note that the definitions of the dominant directions are also valid for the Boundary Bay region (Lange, 1998) ........................................................................................................ 202

Table 5-2 : The redundancy analysis process. (Application of redundancy test on non-directional wind data at station YVR) .......................................................... 203

Table 5-3 : Average climate indices for “strong El Niño”, “strong La Niña” and “neutral” years. The definitions of strong El Niño / La Niña years are based on the Environment Canada (Meteorological Services of Canada-The Green Land) classification scheme ........................................................................................................ 204

Table 5-4 : Maximum-likelihood fitted parameters (MLE) for YVR, Sand Heads and Saturna Stations, having fit a GPD models for the extreme non-directional wind events above a threshold u (Base Model: B). ................................................................................................. 205

Table 5-5 : The extreme non-directional wind recurrences at YVR, Sand Heads and Saturna. Projections are based on the GPD fit of all extreme non-directional winds above 40 km/hr threshold for YVR and 50 km/hr threshold for Sandheads and Saturna stations (Base). ....................................................................................... 206

Table 5-6 : Extreme Southerly (S), Northerly (N) and Westerly (W) wind recurrences without climate considerations (Base-directional) at stations YVR,
Sandheads and Saturna. Projections are based on the GPD fit of all extreme directional winds above a threshold $u$ as indicated in the table.

Table 5-7: Maximum-likelihood fitted parameters for YVR, Saturna and Sand Heads stations, having fit a GPD model with climate covariates. The highlighted values indicate significantly improved GPD fits (at 95% level) over the base fit when corresponding climate index was applied as a covariate in the model.

Table 5-8: Maximum-likelihood fitted parameters for YVR stations, having fit a GPD model with climate covariates. The likelihood ratio test between the Base model and the model with covariates indicate significant improvement in the model fit with covariates.

Table 5-9: Extreme non-directional wind recurrences at YVR subjected to climate variability effects. The variation in the return levels of extreme wind speeds are presented under warm, neutral and cold climate conditions.

Table 5-10: Maximum-likelihood fitted parameters for YVR, Sandheads and Saturna, having fit a GPD model with climate covariates to Southerly (S), Northerly (N) and Westerly (W) winds. The ** indicate statistically significant improvements in the GPD fits at 95% level over the base fit, where the base fit is the model without climate covariates.

Table 5-11: Extreme Southerly (S), Westerly (W) and Northerly (N) wind recurrences at YVR under climate variability effects. The variations in the return levels of extreme wind speeds are presented under warm, neutral and cold climate conditions.

Table 6-1: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA and PDO) vs Total January Storm Track (TJST) count in the North-eastern Pacific (1948-2004).

Table 6-2: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA, and PDO) vs. Total January Southerly (TJSST) Storm Track count in the North-eastern Pacific (1948-2004).

Table 6-3: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA, and PDO) vs Total January Northerly (TJNST) Storm Track count in the North-eastern Pacific (1948-2004).

Table 6-4: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA, and PDO) vs Total Storm Tracks with Lifespan greater than 5-days (HECOUNT) in the North-eastern Pacific (1948-2004).
List of Figures

Figure 1-1 : The Study region: The BC Coastal margin. Also indicated in the figure are the tide gauge locations in the vicinity of the BC coast. .................................19

Figure 2-1 : The geographical location of the study region in the Queen Charlotte Islands (Haida Gwaii) offshore from Prince Rupert, BC. ........................................61

Figure 2-2 : Simple linear regression models for annual MSL from 1909-2003 (Thick line) and 1945 to 2003 (Dashed line)  .................................................................62

Figure 2-3 : Annual MSL variation at Prince Rupert and Queen Charlotte City station from 1966-2003. Showing significant similarities in MSL variation.............63

Figure 2-4 : MAXSL at Prince Rupert from 1945-2003. Note that the MAXSL trend is approximately double the MSL trend. .................................................................64

Figure 2-5 : Surge-tide components at Prince Rupert, for annual MAXSL events (Triangles) and annual maximum surges (circles), showing the consistent occurrence of largest surges at low to mid tides and MAXSL events dominated by larger tides.................................................................65

Figure 2-6 : Observed total water level (triangles), predicted tidal water level (circle-dashed), and residual (predicted - observed, squares) surge-generated water level at Queen Charlotte City gauging station during the 24 December 2003 storm event. Note that the peak surge (0.73 m) occurred at low tide and the maximum total water level (8.06 m) occurred approximately 6 hrs later. ..................66

Figure 2-7 : Impacts of the 24 December 2003 storm surge event, eastern Graham Island, BC. Approximately 2.5 m of shoreline was lost at this location along Highway 16 (upper) compromising the road shoulder and bed. Extensive coastal flooding (lower) also occurred, damaging buildings and sending tonnes of drift logs onto nearby roads and properties............................................................67

Figure 2-8 : Extreme sea level recurrence curve for Prince Rupert tide gauge produced using the Extremes toolkit in R.................................................................68

Figure 2-9 : Residual MSL compared with (A) PDO and (B) MEI (ENSO) climate controls.............................................................................................................69

Figure 2-10 : Monthly MSL response to positive and negative (A) MEI, (B) PDO and (C ) NOI climate controls compared with overall monthly averages. .........70

Figure 2-11 : Winter (Nov.- March) SEA results showing the departure of CV signals (MEI, PDO, and NOI) from their corresponding seasonal means (by +/- 1
STDV) during (a) higher and (b) lower than average MSL events (lag = 0). The strength of CV influence is constrained to two seasons prior (lag -2 and -1) and following (lag 1 and 2) the event season. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate variability series.

Figure 2-12: Summer SEA results showing the departure of CV signals (MEI, PDO, and NOI) from their corresponding seasonal means (by +/- 1 STDV) during (a) higher and (b) lower than average MSL events (lag = 0). The strength of CV influence is constrained to two seasons prior (lag -2 and -1) and following (lag 1 and 2) the event season. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate variability series.

Figure 2-13: CumSum graphs for the four climate indices, MEI, PDO NOI (negative) and ALPI, showing the same distinct trends from 1950-1976 and 1976-2003. Also shown is the CumSum curve for MSL indicating a lag response to the major regime shift occurred in 1976.

Figure 3-1: The geographical location of the southern coats of BC, including the lower Fraser delta. Also shown is the tide gauge location at Pt. Atkinson. The two ended arrow indicate the coastal region of interest.

Figure 3-2: Pt. Atkinson mean sea-level anomaly expressed with respect to (1915-2006) mean. Note that the anomalies are consistently negative prior to 1949.

Figure 3-3: Annual maximum hourly total water-levels at Pt. Atkinson. Note that based on a t-test, the increasing trend in extremes is statistically not significant.

Figure 3-4: The (a) historical annual maximum sea-level record at Pt. Atkinson, (b) assumed long-term relative sea-level trend as per Church (2002).

Figure 3-5: Temporal distribution of NOI, PDO and ENSO indicating significant colinearities. Note the 1976 major climate regime shift being captured by all climate indices approximately at the same time.

Figure 3-6: Return level graph for the historical annual maximum sea-levels (Base), calculated from associated GEV distribution (black solid line) with (95%) confidence intervals calculated from the delta (blue solid) and profile-likelihood (dashed–red) methods. Note that the Base model indicating an apparent under-estimation of extremes TWL recurrences when extrapolated beyond 10-20 years.
Figure 3-7: Return level graph based on GEV distribution for (a) A future climate representing strong warm ENSO (El Niño) conditions (red–dashed) (b) A future climate representing Neutral conditions (blue–dashed) (c) A future climate representing strong cold ENSO (La Niña) conditions (green–dashed). For comparison purposes, the mean return level graph for the Base Case with no climate considerations is plotted (black solid line). Note that the average CV index values (i.e. MEI, PDO and NOI) applied for each case is the average index values of all years with strong El Niño, Neutral and strong La Niña years from (1950-2006) respectively. The definition of years was based on Environment Canada Classification.

Figure 3-8: Surge-Tide relationship and the development of the extreme surge event of 16th December 1982 (top panel). Impacts of the storm surge event in southern BC are shown (bottom panels). Extensive coastal flooding occurred in Boundary Bay, Mud Bay and Westham Island and resulted in highest ever occurred water levels in southern BC.

Figure 4-1: The geographical location of coastal of BC, including the Haida Gwaii in northern BC and Vancouver Island in southern BC. Also shown are the Pacific region tide gauge stations and the location of Hacate Strait and the Georgia Strait.

Figure 4-2: Temporal distribution of NOI, PDO and ENSO indicating significant colinearities. Note the 1976 major climate regime shift being captured by all climate indices approximately at the same time.

Figure 4-3: The GEV distribution without climate considerations, 5% and 95% confidence bands together with the annual maxima for each station. The annual maxima fall almost on the GEV curve and are well within the quantile limits.

Figure 4-4: Estimated return levels for residual water levels (sea surges) under Warm ENSO (red-dashed line), Neutral (blue-dashed line) and Cold ENSO (green-dashed line) conditions, from having fit Maximum Annual Residual Water Levels at Station 7120 to a GEV distribution with climate variability effects accounted as covariates. Results for no climate considerations (black-continuous line) and the 95% confidence limits (blue-continuous line) are included for comparison purposes. Similar probability curves have been constructed for each tidal station of the BC coast, (results not shown).

Figure 4-5: Surge-Tide relationship and the development of the extreme surge event of 24th December 2003 (top panel). Impacts of the storm surge event on the eastern Graham Island, BC are shown (bottom panel). Approximately 2.5 m of shoreline was lost at this location along Highway 16 (upper) compromising the road shoulder and bed. Extensive coastal flooding also occurred, damaging buildings and sending tonnes of drift logs onto nearby roads and properties. (Photos courtesy of Mavis Mark).
Figure 4-6: Surge-Tide relationship and the development of the extreme surge event of 16th December 1982 (top panel). Impacts of the storm surge event in southern BC are shown (bottom panels). Extensive coastal flooding occurred in Boundary Bay, Mud Bay and Westham Island and resulted in highest ever occurred water levels in southern BC. (Photos courtesy of Fraser Delta Engineering Department of the BC MOE).

Figure 5-1: The geographical location of the southern coast of BC, including the lower Fraser Delta. Also shown are the three Meteorological stations YVR, Saturna and Sand Heads. The area marked by the rectangle indicates the Boundary Bay region [modified from Lange, 1998].

Figure 5-2: The probability density plot for the extreme winds above 40 km/hr threshold at YVR; the fitted density following the data indicate that the selected GPD model for the data is a satisfactory choice.

Figure 5-3: Maximum likelihood estimates and confidence intervals of the GPD shape (bottom panel) and modified scale parameters (top panel) over a range of thresholds for YVR station. The dashed vertical line indicates the selected threshold for this data base.

Figure 5-4: Temporal distribution of NOI, PDO and ENSO indicating significant colinearities. Note the 1976 major climate regime shift being captured by all climate indices approximately at the same time.

Figure 5-5: Diagnostic quantile plot for (a) YVR, (b) Sand Heads and (c) Saturna stations from fitting all non-directional extreme wind events above a threshold of 40 Km/hr for YVR and 50 km/hr for Sandheads and Saturna to a GPD model. Note the units of both vertical and horizontal axis’s is Km/hr.

Figure 5-6: Return level plots for (a) YVR, (b) Sand Heads and (c) Saturna stations from fitting all non-directional extreme wind events above a wind speed threshold of 40 Km/hr for YVR and 50 km/hr for Sandheads and Saturna to a GPD model. Note the unit of the vertical axis’s (Return levels) is Km/hr and the horizontal axis (Return period) is Years.

Figure 5-7: Scatter plot of annual number of events above 40 km/hr threshold for station YVR. Note the unit of the vertical axis is Km/hr.

Figure 5-8: Extreme non-directional wind recurrence curves at YVR with climate variability effects. Return period curves are presented for warm (long-dashed), neutral (short-dashed) and cold (dotted) climate conditions. Note the unit of the vertical axis is Km/hr.
Figure 5-9: Daily average extreme winds as projected by the CGCM (Canadian Climate Model) averaged over each month for the duration of the record for (a) Base case (1961-2000) (b) Scenario A1B (2046-2065) (c) Scenario A1B (2081-2100) at two locations adjacent to the study region (123.75W, 48.84N; 126.56W, 48.84N).

Figure 5-10: Storm track enhancement (arrows) associated with 700 mb atmospheric pressure anomalies during La Niña (a) and El Niño (b) dominated climate patterns [from Inman and Jenkins, 2003].

Figure 6-1: The shaded area indicates the study region of interest in the northeastern Pacific (30°N - 60°N; 120°W - 150°W). Each line in the figure represents a single storm track that crossed the region during January 2004.

Figure 6-2: Interannual Variability of the overall January storm track Lifespan anomaly in the north-eastern Pacific from 1948-2004.

Figure 6-3: The 5-year running mean of the northerly (continuous line) and southerly (dashed line) January storm track life span anomaly in the north-eastern Pacific from 1948-2004.

Figure 6-4: Interannual Variability of the January (a) Deep and (b) Weak storm track count anomaly in the north-eastern Pacific from 1948-2004.

Figure 6-5: Secular trends in the January storm track count in the north-eastern Pacific from 1948-2004. (i) Continuous black line; Total track count (ii) Long dashed line; Northerly track count (iii) Short dashed line: Southerly track count.

Figure 6-6: Secular trends in the January deep storm track count in the north-eastern Pacific from 1948-2004. (i) Continuous black line; Total deep track count (ii) Long dashed line; Northerly deep track count (iii) Short dashed line: Southerly deep track count.

Figure 6-7: Secular trends in the January weak storm track count in the north-eastern Pacific from 1948-2004. (i) Continuous black line; Total weak track count (ii) Long dashed line; Northerly weak track count (iii) Short dashed line: Southerly weak track count.

Figure 6-8: Secular trends in the mean latitudinal position of the cyclogenesis and the cyclolysis of the overall January storm tracks in the north-eastern Pacific from 1948-2004. (i) Continuous black line; mean cyclogenesis Latitude; (ii) Dashed line: mean cyclolysis Latitude.

Figure 6-9: The cumulative sum (CUMSUM) curves for (i) Continuous black line; mean latitudinal position of the Cyclogenesis; (ii) Short Dashed line: mean...

Figure 6-10: The cumulative sum (CUMSUM) curves for (i) Continuous grey line: mean latitudinal position of the Cyclogenesis; (ii) Light dashed line: mean latitudinal position of the cyclolysis  (iii) Dark dashed line: Deep storm track count; (iv) Continuous black line: Weak storm track count, of the overall January storm tracks in the north-eastern Pacific from 1948-2004.

Figure 6-11: SEA results showing the departure of CV indices from their annual mean during and around the event years (lag = 0). The event years are defined as years with “Total storm track count” anomaly is (i) higher (≥ 1 STDV) (Left-Panel) and, (ii) lower (≤1 STDV) (Right-Panel) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.

Figure 6-12: SEA results showing the departure of CV indices from their annual mean during and around the event years (lag = 0). The event years are defined as years with the average Southerly (SW & SE) storm track count anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤1 STDV) (Right-Panel) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.

Figure 6-13: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Northerly (NW & NE) storm track count anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤1 STDV) (Right-Panel) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.

Figure 6-14: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the Cyclonegnesis Latitude anomaly is (i) higher (≥ 1 STDV)(northerly shift) (Left-Panel) and (ii) (Right-Panel) lower(≤1 STDV) (southerly shift) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-15: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the cycloysis Latitude anomaly is, (i) higher (≥ 1 STDV) (northerly shift) (Left-Panel) and (ii) lower (≤1 STDV) (southerly shift) (Right-Panel) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series. ..... 279

Figure 6-16: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Storm track lifespan anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤1 STDV) (Right-Panel) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series. ................................. 280

Figure 6-17: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track count anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤1 STDV) (Right-Panel) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series. .......................... 281

Figure 6-18: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track path-length anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤1 STDV) (Right-Panel) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series. .................................................................. 282

Figure 6-19: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track southern-limit (Latitude) anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤1 STDV) (Right-Panel) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series. .................................................................. 283
Figure 6-20: SEA results showing the departure of CV indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track west coast-crossing (Latitude) anomaly is, (i) higher (≥ 1 STDV) (Left-Panel) and (ii) lower (≤ 1 STDV) (Right-Panel) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Acknowledgments

I am very grateful to my supervisor Dr. Dan Smith and my mentor Mr. Ben Kangasnaimi from the BC Ministry of Environment, for giving me the opportunity and the resources to conduct climate research in BC. You two are the most understanding and supportive superiors I was privileged to meet in my life time. My committee members, Drs. Eric Kunze (UVic SEOS), Chris Houser (U of Texas A&M) and Stephane Mazzotti (GSC Pacific) are remembered with gratitude for providing much appreciated support and advice. Dr. Eric Gilliland (of NCAR) is respectfully remembered for his valuable support, advice and guidance. Dr. Ian Walker from the Blast research unit of the Department of Geography is acknowledged for the research support provided during the initial stages of my research program. I would also like to thank my external examiner Dr. Audrey Dallimore (RRU) for her enthusiasm on my research.

Thanks goes to Environment Canada and the Department of Fisheries and Ocean (DFO) research and data management teams for providing the data and assistance for my analysis. The University of Victoria Geography Department and the Faculty of Graduate Studies is remembered with gratitude for providing substantial assistance and encouragement.

My family and friends have provided unfaltering support and encouragement over the past few years (and in the case of Mr. & Mrs. Abeysirigunawardena my parents, much longer). My greatest debt and appreciation is to my husband Sisira Kosgoda and my sweet little daughter “Wish”, Wishva Kosgoda for living through this challenge as much as I have and, for keeping the “flame’ alive.

Support for the research and for myself was provided by the Natural Sciences and Engineering Research Council (NSERC), The BC Ministry of Environment (BC MOE), Canadian Climate Impact and Adaptation Program (CCIAP A580), Environment Canada (EC), and the University of Victoria (UVic).
1.0 Introduction

Observational evidence suggests that global climate changes are altering and aggravating coastal environmental drivers such as sea levels, sea-surges, wind-storms and storm tracks around the world (e.g., Cohen et al., 1997; Gommes et al., 1998; Nicholls and Small, 2002; IPCC, 2007). Of these impacts, accelerated sea-level rise has received much attention due to potential additional stresses exerted on low-lying coastal systems by triggering coastal flooding, aggravated erosion and saltwater intrusion. These impacts are not only caused by longer term sea level rise due to changes in global temperature means or norms (i.e. global warming), but also reflect decadal to inter-decadal scale climate variability impacts on sea levels (Gornitz, 1991; Titus et al., 1991; Pernetta, 1992, Bijlsma et al., 1996; Cazenave et al., 1998; Shaw et al., 1998; Storlazzi et al., 2000; Church et al., 2001; Watson et al., 2001; Allan and Komar, 2002; Barrie and Conway, 2002; Schwing et al., 2002; Donnelly et al., 2004; Allan and Komar, 2006; Church and White, 2006). Once a critical mean sea level threshold is exceeded from long-term sea level rise due to global warming, the economic, human and ecological costs of even a small increase in extreme sea levels due to climate variability could be much higher on coastal communities residing in low lying places (IPCC, 2007). Thus any evidence of possible links between extreme sea levels, weather extremes and climate variability would set the bar even higher when it comes to defining acceptable levels of protection against extreme events.
1.1 Observed sea level changes and future trends

Climate change from global warming is mainly caused by human-induced emissions of so-called “greenhouse” gases, which traps long-wave radiation in the upper atmosphere raising the atmospheric temperature. Carbon dioxide (CO2) is the most important of these gases and its atmospheric concentration has exponentially increased since the beginning of the industrial revolution (IPCC 2007). Accumulation of greenhouse gasses would cause the globally averaged surface temperature to rise by 1.4 to 5.8°C over the period 1990 to 2100; while regions at northern high latitudes are very likely to warm more rapidly than the global average (Houghton et al., 2001). The rise in global temperature due to global warming suggests a rise in global mean sea levels at a rate 2 to 4 times the observed rate over the 20th century (Houghton et al., 2001). These changes are due primarily to the thermal expansion of ocean water (11 to 43 cm), followed by contributions from mountain glaciers (1 to 23 cm) and ice caps (−2 to 9 cm for Greenland and −17 to 2 cm for Antarctica) (Church et al., 2001). Even with substantial reductions in future greenhouse gas emissions, global sea level will continue to rise for centuries beyond 2100 due to the response lag of the global ocean system. An ultimate sea level rise of 2 to 4 m has been projected for atmospheric carbon dioxide concentrations that are twice and four times the pre-industrial levels, respectively (Church et al., 2001).

The 20th Century sea level rise estimates are based on world wide tide gauge records that show considerable discrepancies from one study to another (IPCC 2007). For instance, Church and White (2006) estimated the rates to be in
the range of $1.7 \pm 0.3$ mm per year, while in early 1993, when satellite altimetry became available, much higher sea level rise rates of about $3.1 \pm 0.7$ mm/year and $3.2 \pm 0.4$ mm/year were estimated from combined tide gauge and satellite altimetry records (Chambers et al., 2002; Cazenave and Nerem, 2004; IPCC 2007). Holgate (2007) used tide gauge data to estimate the average rate of global sea level change for the 20th Century as $1.74 \pm 0.16$ mm/year (consistent with Church and White, 2006), with the period 1904-1953 experiencing a rate of $2.03 \pm 0.35$ mm/year and the period 1954-2003 a smaller rate of $1.45 \pm 0.34$ mm/year. The uncertainties in the derived trends for the 20th century may be due largely to the interannual oscillations coupled with relatively short records.

1.2 From Global to Regional Sea level Trends and Uncertainties.

A major uncertainty in global sea-level rise projections is how the global projections will manifest themselves on a regional scale (Church et al., 2004). For instance, models analyzed by the IPCC had shown a strongly non-uniform spatial distribution of sea-level rise (Church et al., 2001). This lack of similarity reduces the confidence in projections of regional sea-level changes based on global estimates; although at present global scale projections play a significant role in regional scale sea level rise impact assessment works. One major reason for the regional sea level differences is the localized vertical land movements experienced by many coastal areas due to glacial isostatic adjustment (GIA), tectonics, and sediment compaction processes (e.g., Mitchum, 2000; Peltier, 2001; Douglas and Peltier, 2002; Mazzotti et al., 2003; Lambert et al., 2008).
Such movements could lead to regionally specific land subsidence or uplift resulting in additional rise or fall of the sea levels relative to land, irrespective of absolute sea-level changes.

In addition, various climate variability (i.e. recent trends in El Niño) modes which forms a part of the natural climate system as short term fluctuations (i.e., seasonal, inter-annual to decadal) are likely to exert both direct and indirect impacts on sea levels at regional to local scales and, at temporal scales typically longer than individual weather events (Houghton et al., 2001; Allan and Komar, 2002; Barrie and Conway, 2002; Allan and Komar, 2006; Papadopoulos and Tsimilis, 2006). Studies also suggest that the severity of extreme sea levels and other coastal environmental drivers such as peak winds, storm-surges and storm track characteristics tend to follow the natural climate variability cycles, sometimes reinforcing and sometimes moderating any effects that greenhouse warming might have on extremes (Crawford et al., 1999; Storlazzi et al., 2000; Graham and Diaz, 2001; Allan and Komar, 2002; Schwing et al., 2002; Bond et al., 2003; Brayshaw, 2005; Allan and Komar, 2006). While these responses are likely to be relevant to many coastal locations around the world, they still remain uncertain on scales useful for coastal impact studies.

1.3 Climate Impacts on Canadian coasts from National to Regional scale.

Following the global pattern, Canada is already experiencing numerous climate changes, such as the one degree Celsius increase in the Canadian mean temperature, increasing mean sea levels, melting and retreating glaciers,
increased rainfall intensity and stream flow, increased intensity and frequency of severe winter storms, and wild fire hazards (Thomson and Tabata 1987, 1989; Coulson 1997; Leith and Whitfield 1998; Shaw et al., 1998; Graham and Diaz, 2001; Arendt et al., 2002; Morrison et al., 2002). These changes are all projected to worsen over time, with regionally varying impacts on infrastructure and natural resources.

Despite having the longest coastline in the world, and about 33% of the coastline indicating a moderate to high physical sensitivity to sea-level rise impacts (Shaw et al., 1998), to date Canada has contributed very little research to enhance the knowledge on climate variability and change impacts on its coasts, at scales useful for the coastal planners. This could be a clear challenge to successfully manage the escalating risk of climate impacts on Canadian coasts over the coming decades. In order to overcome these challenges, it is critical to establish the necessary boundary conditions to conduct detailed coastal impacts assessment studied at much finer temporal scales. Understanding the link between sea levels and climate extremes such as extreme winds, storm-track characteristics and their response to climate variability at regional scales becomes a key contributor to any research efforts made towards realizing this objective.

This has been the main motivation behind my research, which aims to better understand the link between the response of coastal environmental variables such as sea level, storm surges, wind-storms, storm tracks and their
relationship to various climate variability modes in coastal British Columbia (BC), Canada. To date similar analysis has not been done for the region.

1.4 Climate Variability and Change Impacts in coastal British Columbia.

Although BC’s coastline is dominated by rocky high-relief coasts, it is not immune to impacts from sea level rise and extreme events. During the past decade, much of BC’s low lying coastlines have been experiencing aggravated coastal flooding hazards due to concurrent occurrences of extreme storm surges on higher than normal mean sea levels (Abeysirigunawardena and Walker, 2008 (Chapter 2); Thomson et al., 2008; Tinis, 2008; Abeysirigunawardena et al., 2009, (Chapter 5)). Several exceptional storms since 2000 in the eastern coast of Graham Island (2003) and the inner south coast of BC (2006), demonstrated that remote low lying communities such as the Graham Islands communities as well as urban population centres in the south coast of BC are equally vulnerable to severe weather and storm surges (Walker et al., 2007; Thomson et al., 2008). Such impacts on natural coastal systems and coastal communities could be more frequent and severe due to major storm occurrences at higher than normal water levels, allowing little opportunity to rebuild natural resilience or to reduce the exposure of property and infrastructure (Forbes, 2004; Forbes et al., 2004; Walker et al., 2007). Under these circumstances, the existing coastal measures along BC’s coastlines may be inadequate to protect coastal infrastructure from climate intensified storm surges over their expected lifetime.
The number of unusually severe floods, wind storms occurred in coastal BC within the past 15 to 20 years suggests that weather extremes are becoming an increasingly serious problem for BC coastal communities (Lambert, 1995). These observations may reflect either a fundamental shift towards a more extreme climate due to global warming and/or temporary climate phenomenon caused by natural climate variability. In order to diminish the uncertainties associated with the response of extremes to a changing climate, it is vital to differentiate the contribution of each climate mode on the present intensification of extremes. It is also important to review and reassess the risks of weather disasters and society’s ability to cope with them under a changing climate. For example, to determine whether an extreme event that now has an estimated probability of occurrence of once in 100 years should be upgraded to, say, a 1 in 50 probability.

A detailed assessment of coastal impacts due to climate variability and change may necessitate the issue be viewed at a much finer temporal scale than the current study. At this scale, the impacts of waves as a coastal environmental variable and the changes to coastal geomorphology become major role players. However, at present, the BC’s coast region does not possess data at sufficiently finer resolution to conduct analyses at the local scale. As such, this study is limited to an investigation of the impacts of climate variability and change on coastal environmental variables (i.e. extreme water levels, winds, storm surges and storm tracks) that vary at regional scale. It is anticipated that the results of this study will eventually contribute boundary conditions to local scale studies in
the future, where the impacts of climate variability and change on the local wave regime and coastal geomorphology will be addressed.

Despite the excellent record of a serious commitment to climate change science in BC, our current understanding of the regional climate change impacts on coastal BC (Lat. 44° N and above) is far from complete as well. The degree of uncertainty in the magnitude and rate of climate change impacts on coasts in the coming decades is still substantial, due to lack of clarity on the regional scale atmospheric and oceanic responses to natural climate variability; particularly the lack of description on the link between the natural climate variability and the frequency and intensity of extreme events. To that end, my research utilizes a network of historical meteorological and oceanographic observational records to describe many aspects of regional scale natural climate variability and climate change impacts on sea levels and many other coastal environmental drivers in BC. More specific aims of this research are two fold: (i) to enhance our understating on the ocean-atmospheric interactions in relation to natural climate variability and anthropogenic climate change and, (ii) to describe the role of natural climatic variability as a fundamental element in explaining the changing frequency and intensity of extreme winds, storm surges and storm tracks).

1.5 The Study Region.

The spatial extent of the study region spans the BC coastal margin between Latitudes 48°N and 55°N (Figure 1.1). Along this coastal belt, the most aggressive impacts from the past extreme events had taken place at two specific
locations: (i) The eastern coast of Graham Island, Queen Charlotte Island in Northern BC where important infrastructure developments, such as the only highway and rail link between Tlell and Masset are located close to the coastline, and being increasingly threatened; and, (ii) the low lying inner south coast of BC where the high demand for waterfront property and growing coastal developments along the low-lying, unprotected coast has significantly increased the vulnerability to extreme events (Figure 1.1). The Geological Survey of Canada has placed these two regions in the top 3% of Canada’s most sensitive coasts to accelerated sea-level rise impacts (Shaw et al., 1998). Thus the case-studies presented in the thesis gives special emphasis on the two study sites mentioned above.

1.6 Research purpose and objectives.

While it is recognized that climate change impacts on coastal zones are prompted by more than global sea-level rise, consideration of natural climate variability as a contributing factor is largely limited in impact assessment studies due to the uncertainties surrounding them. Nonetheless, it is believed that a better interpretation of sea level variations and extreme events in the context of these climate regimes may eventually help separate the contributions of internal variability versus anthropogenically-forced climate change.

The purpose of this study is to describe the dynamical characteristics of five large-scale climate variability regimes and to determine how they have affected coastal environmental drivers in coastal BC over the last 50 years.
Historical observational records of water-levels, storm surges, winds and storm tracks in the study region have been analysed with respect to large-scale atmospheric and oceanic circulation indices. The five climate regimes are described briefly in the following chapters.

The work presented combines and integrates results of a number of different studies. Each study pursues a consistent approach in order to increase our understanding of the key concepts related to impacts of climate variability and change on sea levels in coastal BC. The research has four main objectives:

1. Investigate the linear and non-linear sea level responses to Climate Variability (CV) and Change (CC) signals at multiple temporal scales (inter-decadal to monthly) on northern BC coastline.

Long-term water level records from Prince Rupert (PR: Station 9354), the longest available tide gauge record in the region were statistically analyzed to provide evidence of long-term sea level trends: Linear and non-linear statistical techniques including correlation analyses, multiple regression, Cumulative Sum (CumSum) analysis, and Superposed Epoch Analysis (SEA), were used to explore relationships between sea levels and known CV indices. The study was intended to provide new insights into climate-sea level relationships in coastal BC.

2. Investigate the changes in the severity of sea surge events (i.e. sea surge is defined as the differences between the observed sea heights and the
simultaneous astronomical tide) in response to climate variability in coastal BC, where no previous studies have been done.

In-situ hourly tide gauge data at eleven tide gauge stations located along the BC coast were statistically analyzed to reproduce the observed characteristics of sea surges in the region, while properly accounting for the effects of climate variability. The analysis was designed to provide new insight into the integrated effect of atmospheric parameters, tides, climate variability and change effects on sea-surges in the region.

3. Establish the relationship between extreme winds and various regional and large-scale climate variability modes in coastal BC.

Extreme wind event recurrences and their relationship with ENSO states are investigated. The motive was to understand the role of extreme winds as a contributing factor towards generating extreme sea levels in coastal BC. The study was based on observed independent extreme wind events at three meteorological stations maintained by the Meteorological Service of Canada.

4. Investigate statistically, storm track characteristics in relation to various regional and large-scale climate variability signals in coastal BC.

The study was based on the NCEP/NCAR reanalysis of January storm track projections from 1948-2004 (used as an analog for inter-annual winter storm track variability in BC). The intent was to review current status of knowledge related to the impact of climate variability and change on North-eastern Pacific
storm track characteristics, and to examine whether known climatic variability signals could explain the spatial and temporal variations of the north-eastern Pacific January storm track characteristics and, to explore whether opportunities exists for forecasting storm track variability based on the dynamics of climatic variability events (e.g., inter annual patterns, regime shifts).

The overall results of this dissertation are summarized in a concluding chapter, following the presentation of five independent manuscripts.

1. “Sea level responses to climatic variability and change in Northern BC”. This manuscript which was published in the Journal of Atmospheric and Oceans (Atmosphere-Ocean 46 (3), 277–296) includes the methodology and an evaluation of the sea level responses to climatic variability and change signals at multiple temporal scales (inter-decadal to monthly) on the north coast of BC.

2. “Extreme Sea-level Recurrences in the South coast of BC Canada with Climate Considerations”. This manuscript currently in review in the Asia Pacific Journal of Climate Change (APJCC) presents a statistical model to simulate the prominent features of the impacts of climate variability and change on extreme total water-levels (TWL: total water-level is defined as the combined surge and tidal component at a given time and location) in the southern coast of BC (Point Atkinson).

3. “Extreme Sea-Surge Responses to Climate Variability in Coastal BC Canada”. This manuscript currently in review in the Annals of the Association of American Geographers presents the spatial and temporal changes in the
severity of extreme sea surge events in response to climate variability in coastal BC, Canada.

4. “Extreme Wind Regime responses to climate variability and change in the inner-south-coast of BC Canada”. This manuscript published in the Journal of Atmosphere and Oceans (Atmosphere-Ocean 47 (1), 41–62) evaluates the possible influence of climate variability on extreme wind response. A secondary objective of this study was to demonstrate the use and value of climate information in accurately determining extreme wind recurrences in a region.

5. “Sensitivity of winter storm track characteristics in North-eastern Pacific to climate variability”. The study currently in review in the Journal of Atmosphere and Oceans demonstrates the extent to which the north-eastern Pacific storm track characteristics are manifested by various climate oscillations, using NCEP/NCAR reanalysis January storm track projections from 1948-2004 (used as an analog for winter storm track variability).

References:


Figure 1-1: The Study region: The BC Coastal margin. Indicated in the figure are the tide gauge locations on the BC coast. (Source: the Marine Environmental Data Services of Canada (MEDS)).
2.0 Sea level responses to climatic variability and change in Northern BC Canada.

2.1 Abstract

Sea level responses to climatic variability (CV) and change (CC) signals at multiple temporal scales (inter-decadal to monthly) are statistically examined using long-term water level records from Prince Rupert (PR) on the north coast of BC. Analysis of observed sea level data from PR, the longest available record in the region, indicate a annual average mean sea level (MSL) trend of +1.4±0.6 mm yr⁻¹ for the period (1939-2003), as opposed to the longer term trend of 1±0.4 mm yr⁻¹ (1909-2003). This suggests a possible acceleration in MSL trends during the latter half of the 20th century. According to the results of this study, the causes behind this acceleration can be attributed not only to the effects of global warming but also to cyclic climate variability patterns such as the strong positive PDO phase lasting since mid 1970s'. The linear regression model based on highest sea levels (MAXSL) of each calendar year showed a trend exceeding twice that (3.4 mm yr⁻¹) of MSL. Previous work shows vertical crustal motions on relative sea level as negligible at Prince Rupert.

Relations between sea levels and known climate variability indices are explored to identify potential controls of climate variability phenomena on regional MSL and MAXSL. Linear and non-linear statistical methods including: correlation analyses, multiple regression, Cumulative Sum (CumSum) analysis, and Superposed Epoch Analysis (SEA) are used. Results suggest that ENSO forcing (per MEI and NOI indices) exerts significant influence on winter sea level
fluctuations, while PDO dominates summer sea level variability. The observational evidence at Prince Rupert also shows that, during the period 1939-2003, these cyclic shorter temporal scale sea level fluctuations in response to climate variability, were significantly greater than the longer-term sea-level rise trend, by as much as an order of magnitude and with trends over 2x that of MSL. Such extreme sea level fluctuations related to climate variability events should be of immediate focus for the development of coastal adaptation strategies, as they are superimposed on long term MSL trends, resulting in greater hazard than longer term MSL rise trends alone.

2.2 Introduction.

Long-term sea level rise controlled by climatic change (CC) is regarded as a major threat to the coastal environment, particularly due to erosion and flooding of low lying, sensitive coasts (Gornitz, 1991; Titus et al., 1991; Pernetta, 1992; Bijlsma et al., 1996; Shaw et al., 1998; Church et al., 2001; Watson et al., 2001; Donnelly et al., 2004; Allan and Komar, 2006; Church and White, 2006). These variations are predominantly due to two causes: the steric sea level rise resulting from the thermal expansion of warming oceans, and the eustatic sea level rise due to the export of freshwater from the continents (Munk, 2003). There is also increasing evidence of significant, shorter-term sea level responses to latitude dependent intra- and inter decadal scale climate variability phenomena such as the El Niño Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) (Cazenave et al., 1998; Storlazzi et al., 2000; Allan and Komar, 2002; Barrie and
Conway, 2002; Schwing et al., 2002). For instance, higher than average annual sea levels in the eastern tropical Pacific are associated with increased sea surface temperature (SST) during the warm phase of ENSO. This response is dynamically linked to the Southern Oscillation - an alternating pattern in atmospheric pressures at sea level (SLP) between the eastern and western Pacific. As such, ENSO is a coupled ocean-atmosphere phenomenon that causes global climate variability on inter-annual time scales (Wolter and Timlin, 1993). These sea level fluctuations could introduce significant differences into estimates of century scale sea level trend projections depending on the temporal extent of tide gauge data records (Munk, 2003). Historical tide gauge records have been used widely to understand the sea level responses to complex, non-linear climate variability signals (e.g., ENSO, PDO) (e.g., Allan and Komar, 2002; Barrie and Conway, 2002; Allan and Komar, 2006; Papadopoulos and Tsimpolis, 2006).

During El Niño, weakened easterly trade winds resulting from higher than normal SLP in the western Pacific, cause warm surface ocean waters to migrate eastward as Kelvin waves (Chelton and Enfield, 1986; Wolter and Timlin, 1993, 1998; Goes et al., 2001). This, together with reduced wind-driven up-welling effects, causes SST and hence steric MSL, to rise along northeast Pacific coastal margins. In addition, the strong alongshore southerly winds during warm phase of ENSO could also substantially raise the MSL along the northeast Pacific coastal margin due to geostrophic adjustment of alongshore currents and Coriolis forcing (Ryan and Noble, 2006). For instance, during the major El Niños of 1982-
83 and 1997-98, mean ocean levels from California to Alaska rose by as much as 100 cm above average (Subbotina et al., 2001) and produced extensive coastal erosion and infrastructure damage (e.g., Storlazzi et al., 2000; Allan and Komar, 2002). The same events caused MSL along northern BC coastal margins to reach 10–20 cm above the long term seasonal mean heights (Crawford et al., 1999). Further, during the winter of 1997-98, Hecate Strait experienced a MSL rise of 40 cm from its long term mean and as much as 12 m of coastal retreat on NE Graham Island (Barrie and Conway, 2002).

The ENSO phenomenon has extra-tropical expressions due to the northward deflection of warm equatorial currents from South America, that translate up the western coast of North America. These expressions eventually reach the BC coast with the influence of strong southerly alongshore winds during the warm ENSO phase. Such extra tropical expression of ENSO in the NE Pacific is described by the Northern Oscillation Index (NOI) (Schwing et al., 2002) – a measure of SLP anomalies between the North Pacific High region (NPH) off the coast of California (35°N, 130°W) and Darwin Australia (10°S, 130°E). As these two locations are considered the centres of action for the NE Pacific Hadley-Walker atmospheric circulation, the NOI describes the strength of ENSO events (positive = La Niña, negative = El Niño) experienced in the NPH region (Schwing et al., 2002).

In addition to MSL rise, more intense and more frequent winter (i.e. late Nov to Mar) storms, surges and waves are often associated with warm ENSO events, as a result of an enhanced Aleutian Low (AL) pressure system in the NE
Pacific. Aleutian Low Pressure Index (ALPI) describes the relative intensity of the AL system, which drives regional SLP and, hence, surface winds in the NE Pacific (Beamish et al., 1997). Positive ALPI values reflect a relatively strong winter AL system and, hence, indicate conditions of potentially stronger fall-winter storms that could affect regional MSL and extreme water level events (e.g., storm surges). However when Aleutian Lows are confined to more northern latitudes, accompanying winds tend to blow more strongly from the west toward BC, dropping sea levels in the study region. Thus ALPI on such years may not likely be a useful indicator of winter sea level (Bond et al., 2003).

PDO is a longer-term, ENSO-like pattern of ocean-atmospheric variability defined as the leading principal component of North Pacific monthly SST variability north of 20°N (Mantua et al., 1997). In addition to persistent SST warming, PDO influences the climatic variability of the NE Pacific region by enhancing or suppressing the year-to-year strength of ENSO events at inter-decadal frequencies (Gershunov and Barnett, 1998). For instance, the positive (warm) PDO phase is associated with enhanced El Niño (and weak La Niña) conditions and vice versa.

Abrupt climate regime shifts between warm and cold PDO phases are common in the North Pacific/North America sector, and can last for decades (Gedalof and Smith 2001). For instance, over the 20th century, there were two full PDO cycles with the cold phases dominating from 1890 to 1924 and 1947 to 1976, and the warm phases prevailed from 1925 to 1946 and 1977 to mid 1990s’ (Mantua et al., 1997; Zhang et al., 1997). These rapid (yearly scale) regime shifts
involve reorganization of local climate trends from one relatively stable state to another persistent (decadal scale) state. Regime shifts of the PDO have been highly correlated with localized ocean-atmospheric and biophysical systems (Mantua et al., 1997; Zhang et al., 1997). For example, Pacific Salmon population dynamics and macro zooplankton abundance are closely linked to biophysical oceanographic responses to PDO regime shifts (Beamish et al., 1999). The NE Pacific sea levels also show strong links to PDO regimes (Mantua et al., 1997; Zhang et al., 1997). For example, during the positive PDO phase, this region experiences warmer coastal waters and enhanced AL system (Mantua and Battisti, 1994; Zhang et al., 1997) that translates to local rise in MSL well above long-term trends (e.g., Flick and Cayan, 1984; Crawford et al., 1999; Subbotina et al., 2001; Barrie and Conway, 2002). Despite these findings, the causes behind the PDO cycles and associated regime shifts are still an ongoing scientific discussion (Schneider and Cornuelle, 2005).

To date, the majority of studies on sea level response to climatic variability in the northeast Pacific are: i) confined to regions below 50°N; and, ii) focused on ENSO as the main source of ocean-atmospheric forcing. As the nature and magnitude of sea level response to climate variability signals are latitude dependent, the results of these studies may not completely describe higher latitude responses (e.g. parts of northern BC coastal waters between 53°N to 55°N). There is also a general lack of description on sea level responses to other regionally specific climate variability modes (e.g., PDO, NOI, and ALPI) in the NE Pacific (notable exceptions: Seymour, 1998; Storlazzi et al., 2000; Allan and
Komar, 2002, 2006; Papadopoulos and Tsimilis, 2006). For instance, the significant PDO controls over sea level variability at shorter time scales (i.e., interannual to seasonal) has not been sufficiently recognized. Furthermore, there is a lack of examination of annual maximum sea level events (MAXSL) and seasonal sea level responses to climate variability modes in the study region.

To address these gaps, this study provides a comprehensive analysis of sea level responses to longer-term CC effects and short term known CV phenomena on the northern BC coast between latitudes 53°– 55°N. The goal is to identify significant links between climate variability events as described by several key climate indices, longer-term CC and sea level variations across interdecadal to seasonal time scales. This is done via four research objectives. First, the long-term sea level trend and confidence limits are identified from historic tide gauge data, via simple linear and non-parametric regression models. Second, extreme value analysis is applied on annual MAXSL to develop a representative extreme water level recurrence curve. To place this into context, a case study of a recent extreme storm surge event on 24 December 2003 is examined with respect to the larger MAXSL regime in the region. Third, linear and nonlinear correspondence between sea level variations and major climate variability indices is explored using multiple regression, correlation, and Superposed Epoch Analysis (SEA). Cumulative Sum (CumSum) analysis is also applied to demonstrate correspondence between sea level responses and regime shifts in climate variability events. Fourth, the degree to which different climate variability
indices modulate sea level variability at seasonal to monthly time scales is examined graphically.

To conclude, the study provides a discussion on the influence of long term (CC) and short term (CV) climate controls on sea levels in northern BC between latitude $53^0 \text{N} - 55^0 \text{N}$. Results of this study provide boundary conditions for a larger study of sea-level rise impacts on NE Graham Island, Haida Gwaii (formerly known as Queen Charlotte Islands), BC.

### 2.3 Study Region

The Haida Gwaii is an archipelago of more than 180 smaller islands that lie approximately 80 km off the north coast of BC from PR (Figure 2.1). The Geological Survey of Canada identifies the eastern coastline of the largest island in the archipelago (Graham Island) as one of the top 3% of Canada’s most sensitive coastlines to CC induced sea-level rise, largely due to its low lying, highly erosive, macro tidal shoreline (Shaw et al., 1998; Walker and Barrie, 2007). The prograding north shore of Graham Island fronts onto Dixon entrance, while the eroding eastern shore faces Hecate Strait. These bodies of water provide a coupled sea strait length of approximately 400 km.

Regional winds, waves and, hence, sea level fluctuations are strongly seasonal with most extreme levels occurring during the winter season (Lewis and Moran, 1985; Dobrocky Seatech Ltd., 1987; Maclaren Plansearch Limited. and Oceanweather Inc., 1995; Walker and Barrie, 2007). Wave conditions in northern
Hecate Strait are sheltered from Pacific swell and are dominated by strong SE wind waves (Amos et al., 1995; Barrie and Conway, 1996).

Tides near Graham Island are classified as mixed semidiurnal with M2 tidal amplitude in northern Hecate Strait of almost 2 m that peaks on the east coast of Graham Island (Bowman et al., 1992). Harmonic analysis of long-term tidal records in this region shows that M2, S2, K1, N2, O1 and K2 components comprise about 90% of the total amplitude of semidiurnal constituents (Crawford, 1998). The region experiences the second highest tidal range in BC, with the central portion of the Hecate Strait experiencing a tidal range near 4.5 m and spring tides exceeding 7 m above the Nautical Chart Datum (CD) near PR and Queen Charlotte City (QC-City).

Evidence exists as to mean sea levels in the region being responsive to ENSO events, with seasonal means rising on average by 10 to 20 cm above average seasonal heights (Crawford et al., 1999) to 40 cm during the 1997-98 El Niño (Barrie and Conway, 2002). Such prolonged events cause significant coastal erosion and are superimposed on longer-term sea-level rise. Extreme storms and surges are also common in the region. For instance, the case study examined in this chapter, a high magnitude SE storm on 24 December 2003, generated a surge of about 70 cm that combined with a low spring tide to produce an extreme total water level of 8 m above chart datum (i.e. note that this analysis was based on the hourly water level time series determined by sub-sampling the 15-minute data at the top of each hour at Prince Rupert). This event caused major flooding, 10 to 12 m of erosion, and significant damage to the
coastal highway along the northeastern shores of Graham Island. To date, little is known about the relations between the frequency and magnitude of such storm events and potential relations to CV in the region.

2.4 Methods

2.4.1 Water level data

Time series of monthly, seasonal and annual mean sea levels were compiled from hourly tide gauge data at PR and QC-City (Figure 2.1). These data were sourced from the Marine Environmental Data Service (MEDS) database maintained by the Fisheries and Oceans Canada. According to the data overview of MEDS, the hourly water level time series were determined by taking the hourly levels directly (i.e. no interpolation or averaging) from the digital readings of the water level gauge recorded at 15-minute sampling interval. The data at PR were considered more reliable over QC-City for its longer, more continuous record (back to 1909 vs. 1966). Further, PR being a deep water station did not reflect localized shallow water characteristics, such as the influence of sand bars and shallow coastal shelf as of the QC-City station (M. Foreman, personal communication, 2004). All measured water levels are referenced to a geodetic benchmark and provided as values above a zero level known as nautical chart datum (CD). Years with less than 90% continuous data were eliminated from the analysis, leaving 71 years of hourly observations, most of which occur after 1939.

Obviously the average and extreme sea levels observed at Prince Rupert could differ considerable than those that occur along the eastern coast of
Graham Island due to bathymetric and aspect differences. For instance, Prince Rupert being a deep water port does not experience high extreme surges as those that occur along the shallow banks of the eastern Graham Island. Both extreme and average sea levels measured at Prince Rupert are generally lower than those at Queen Charlotte City. For example the most recent extreme storm event occurred in December 2003 recorded the highest ever occurred extreme water levels at both stations exactly at the same time. However, the associated surge component at Prince Rupert was over 10 cm lower than that measured at Queen Charlotte City station (see also Figure 2.3 for changes in mean sea levels). Despite these differences, the longer temporal scale sea level responses and trends controlled by climate variability and climate change effects (absolute sea level trends) are expected to be comparable at both locations due their close proximity.

2.4.2 Climatic variability indices

Several published climatic variability indices are used in this study. The Aleutian Low Pressure Index (ALPI) which indicates the relative intensity of the winter (Nov to Mar.) AL system, is the mean area (km²) with SLP less than or equal to 100.5 kPa expressed as an anomaly from the 1950-1997 mean (Beamish et al., 1997). A positive index reflects a relatively strong or intense AL. TrenBerth and Hurrell (1995) provide another index of the AL system, the North Pacific Index (NPI), that is virtually the inverse of ALPI (r= -0.86). Both indices indirectly describe the intensity of the AL in the NE Pacific and, thus, the relative strength of wind patterns that affect regional sea levels and surges. Annual ALPI
values are published by the Fisheries and Oceans Canada (www.pac.dfo-mpo.gc.ca/sci/samfpd/downloads/alpi.txt).

An ENSO variation is characterized via the Multivariate ENSO Index (MEI) (Wolter and Timlin, 1993). MEI is the weighted average of a number of tropical Pacific environmental variables including SST, the east-west and north-south components of surface winds, SLP and SST, and cloudiness. Negative MEI values represent the cold ENSO phase (La Niña) while positive MEI values represent the warm ENSO phase. Atmospheric pressure effects are captured in the MEI by the Southern Oscillation Index (SOI) - the SLP anomaly in the equatorial Pacific between Darwin, Australia (10°S, 130°E) and Tahiti (18°S, 150°W). Monthly MEI values are published online at NOAA-CIRES, Climate Diagnostic Center’s Comprehensive Ocean-Atmosphere Data Set (COADS) (www.cdc.noaa.gov/ClimateIndices/List/index.html).

The PDO index characterizes inter annual variability in average north Pacific SST and, as above, reflects NE Pacific regional ocean-atmospheric variability. Monthly PDO index values are published online at the Joint Institute for Study of the Atmosphere and Ocean’s PDO site (www.jisao.washington.edu/pdo) (Mantua et al., 1997).

The Northern Oscillation Index (NOI) is the sea level pressure anomaly between the North Pacific High (NPH, 35°N and 130°W) region of the NE Pacific and the equatorial Pacific near Darwin, Australia, a climatologically low pressure region (Schwing et al., 2002). The NOI is dominated by and, hence, reflects interannual variations of ENSO events, such that large positive (negative) NOI values
are associated with La Niña (El Niño) events. As NOI is partially based in the NE Pacific, it provides a more direct connection between known ENSO variability in the tropical Pacific and climate variability responses in the study region. Monthly NOI values are available online from NOAA’s Pacific Fisheries Environmental Laboratory (www.pfeg.noaa.gov/products/PFEL MODELED/indices/NOIx/noix.html).

2.4.3 Statistical methods

Simple linear and non-parametric regression methods were applied to annual MSL and maximum sea level (MAXSL) data to identify significant long-term linear trends. The reliability of predicted linear regression model was assessed based on the t-test statistic and 95% confidence intervals. The strength of dependence in the simple linear and non-parametric regression models is assessed using the Pearson’s product-moment and Kendall’s Rank correlation coefficients respectively. Auto correlation plots and the Durbin-Watson (DW) test for auto correlation between Ordinary Least Square residuals (between times t and t-1) were used to test for possible residual auto correlations in the dataset.

The strength of shared variance between climate variability indices (independent variables) and sea levels (dependent variable) were investigated via correlation analyses and multiple regressions. In the multiple regression analysis, if a particular climate variability index within the regression model became redundant due to addition of another climate variability index as a predictor, this latter index value is assumed to have more predictive power on sea level fluctuations than the former. Though simple in application, the forecasting ability of multiple regression models is limited when the residuals are
auto correlated. For example, positively auto correlated residuals may make regression tests appear more significant than if non-auto correlated variables are used, which can lead to biased estimates. As such, multiple regression is used here only as a scoping tool to identify potential response relations between climate variability variables and sea levels.

Several non-linear statistical techniques and graphical methods were also used to explore relationships between climatic variability and sea levels. Return Periods of MAXSL were established by fitting a non-linear Generalized Extreme Value (GEV) distribution to ranked annual extreme sea levels (Gumbel 1958). A strong GEV model fit can be obtained via Maximum Likelihood Estimation (MLE) for datasets with more than 50 observations, allowing extrapolation of return periods (R) 3 to 4 times the length of the original data series (i.e., R = 150 to 200 years) with some confidence (Coles and Dixon, 1999; Martins and Stedinger, 2000). Extreme value statistics in this study were derived using the Extreme Value Toolkit developed for extreme weather and climatic variability applications by the National Center for Atmospheric Research (NCAR) ([http://www.isse.ucar.edu/extremevalues/](http://www.isse.ucar.edu/extremevalues/)) (Gilleland et al., 2003).

Graphical analysis was applied to identify the extent to which climate variability signals (MEI, PDO and NOI) modulate the regular monthly average sea level variations. To do so, average monthly MSL values derived form the total record were compared against the average monthly MSL values corresponding to positive and negative climate variability indices values grouped separately. If graphical comparisons of the two groups of MSL values exhibit
large similar deviations from the overall average, it is concluded that forcing from
the corresponding climate variability pattern modulates MSL variations by, either
enhancing or suppressing water levels during the signal period.

Cumulative Sum Analysis (CumSum) was used to compare and illustrate
the correspondence between trends in climate variability indices and their regime
shifts to MSL fluctuations. The approach involved simple addition of data points
in a time series to the sum of all previous points. This allows examination of long
period (decadal scale) trends in the time series (Murdock, 1979). In addition,
trends can also be identified by subtracting the mean of the entire time series
from each data point prior to addition. Thus, a trend indicated in a CumSum
graph usually corresponds to similar trend in the dataset. For instance, a positive
(negative) slope in a CumSum curve corresponds to an increasing (decreasing)
trend in the time series, while the year when the slope changes defines a regime
shift. This technique is applied in fisheries management to study the responses
of fish populations to climatic variability signals and regime shifts (cf. Beamish et
al., 1999).

Superposed Epoch Analysis (SEA), a non-parametric method also known
as “compositing”, is a classification technique that allows testing an association
between a response and an explanatory variable at their extremes (min, max)
(Prager and Hoenig, 1989). SEA has been applied to describe the linkages
between precipitation, drought, stream flow and climatic variability patterns (e.g.,
Kahaya and Dracup, 1993; Kadioğlu et al., 1999; Kahaya and Karabörk, 2001;
Hessl et al., 2004). Despite this, applications to oceanographic responses are
scarce. In this study, SEA was applied to identify current year and lagged nonlinear relationships between climate variability indices and seasonal (i.e., summer or winter) MSL values. The monthly mean MSL value from April to October averages is considered representative of the “Summer” season while the following November to March value was considered as “Winter”. For a given season, significant event years were identified as those with seasonal MSL anomalies exceeding one standard deviation (>1 STDV) from the overall mean. Next, for each CV index (MEI, NOI, PDO), 5-yr epochs were formulated around significant event years (i.e., 2 seasons preceding, 1 during the event season, and 2 seasons following). The five year epoch window was chosen to yield sufficient number of events for the evaluation of multi-seasonal (inter-annual) linkages between climate variability indices and extreme MSL events. Finally, the significance of each climate variability index across the 5-year period was evaluated by comparing the overall mean climate variability values during the event epoch, against the complete climate variability time series, using a randomized Monte Carlo simulation method. This technique randomly picks years, identifies 5-yr windows, calculates expected means, and provides 95% confidence intervals. The computer program “EVENT”, developed as a part of FHX2 fire history program (cf. Grissino-Mayer and Swetnam, 2000), was used to perform the Monte Carlo simulations. The same approach was used to test the association between lower than average seasonal MSL values (i.e., MSL < 1STDV) and the climate variability indices.
2.5 Results and Discussion

2.5.1 Long-term sea level (MSL) trends

Linear best fit calculations based on the method of least squares on the 71 year annual MSL database at PR yield a statistically significant ($p< 0.05$) long-term MSL trend of $1\pm0.4$ mm yr$^{-1}$ (Figure 2.2). Nevertheless, the confidence in this trend is substantially reduced by the presence of significant gaps in the data series, mainly for the period prior to 1940s. A closer, albeit qualitative, assessment of the same time series reveals a possible acceleration in relative MSL since around 1940s, possibly reflecting the global acceleration trends in MSL over the 20th century, as observed in other studies (e.g., Church and White, 2006). For these reasons, a second regression model was derived using MSL data from the latter half of the 20th Century (1939 to 2003). The model fit indicate a statistically significant ($p< 0.05$) trend of $+1.4\pm0.5$ mm yr$^{-1}$ (Figure 2.2). A test for residual auto correlation based on DW test (Durbin Watson) overrules significant auto correlation (DW=1.61). The non-parametric regression model fit to the same annual MSL series indicates a highly significant ($P<0.0001$) trend of $+1.4\pm0.6$ mm yr$^{-1}$, with a Kendall’s rank correlation of 0.4 (Kendall, 1976).

Figure (2.3) compares the MSL variation pattern between PR and QC-City for the period 1966-2003. Though statistically not significant, QC-City shows a negative relative MSL trend. This could likely be an indication of a gradual lowering of relative sea levels along the eastern coastal margin of the Island due to land rebound (Lambert et al., 2008). Despite these differences in relative sea level trends between stations, Figure (2.3) also indicate significant similarities in
the MSL variation pattern around the trend lines, suggesting spatial coherence in absolute sea level variations at PR and the rest of the study region.

Though tectonic uplift signals on relative sea level are considerable in some regions of the BC coast (e.g., southwestern Vancouver Island at Tofino), the vertical geodetic monitoring system in the study region is not sufficiently long enough to provide reliable estimates of crustal movement in the region (Henton et al., 2006; Mazzotti et al., 2007). However, a recent study based on tide gauge record analysis in the northern Pacific, (including PR), shows that the vertical land movements from glacioisostatic and/or tectonic uplift trends in the vicinity of PR is relatively insignificant (< 1 mmyr⁻¹) (Larsen et al., 2003). Therefore the derived relative MSL trend of +1.4±0.6 mmyr⁻¹ at PR is proposed as a more conservative representation of the absolute sea level trend for the study region. However, future verification and refinement of these absolute sea level trend estimates are warranted via accurate quantifications of land subsidence and uplift trends based on longer term geodetic data. To that end, the need to maintain reliable and continuous geodetic monitoring programs in NE Pacific is emphasized.

2.5.2 Extreme sea level (MAXSL) trends

Extreme sea level trends are typically higher than MSL trends and often result from events, such as storm surges, that generate positive residual water levels above predicted tides. These trends may be linked to climatic variability if there is an increasing frequency and/or magnitude of storm events in the region. The simple linear regression model based on yearly maximum sea levels
(MAXSL) at PR from 1945-2003 shows a significant (P< 0.05) trend of +3.4 mm yr⁻¹ (Figure 2.4). From this, it appears that since mid 20th century annual MAXSL events in PR have increased at twice the rate of the relative mean sea level trend. No doubt these increases are due to enhanced storm conditions and the influence of major ENSO events (e.g., 1982-83 and 1997-98) during this period. This agrees well with the findings of Graham and Diaz (2001), who demonstrated remarkably regular trend towards increasing numbers of intense cyclones with associated increase in cyclone related wind speeds in the northeastern Pacific (20⁰- 65⁰ N) for the latter half of the 20th Century. Furthermore, an independent analysis reconfirms these findings by indicating significant (P<0.01) increase in the frequency of surge-related storm winds (i.e., winds from S-SE >50 km hr⁻¹) in the study region since mid 1970s (Abeysirigunawardena and Walker, unpublished results). It is believed that this increasing trend in MAXSL coupled with enhanced storm winds, waves and surges may be responsible for significant beach erosion in the study region over the past 2 decades (Barrie and Conway, 2002; Walker and Barrie, 2007).

In a macro tidal region, the phasing of the astronomical tide relative to the ocean surge is critical in determining coastal hazards due to extreme sea levels. For instance, coincidence of a peak surge with high to moderate tide could pose serious flooding and damage. Considering this possibility, the tide surge relationship was graphically analyzed, by plotting the corresponding surge and tidal components for highest annual total water levels (i.e. one record per year) and highest annual surge events at PR and QC-City (Figure 2.5). Interestingly,
results at both stations (i.e. QC-City not shown here) show highest annual surge events predominantly occurring at intermediate tides, while highest annual total water level events are dominated by astronomical tides. For example, less than 5% of the extreme annual TWL events considered in this analysis were associated with surge events greater than 0.6 meters, a common threshold applied to demarcate extreme surge events. This suggests significant tidal dominance in the MAXWL regime of the study region.

The consistent occurrence of annual maximum surges at intermediate tides signals a possible coupling between astronomical tides and surges at these two stations. However, since this analysis was strictly limited to a study of surge tide relationship at their annual extremes, the conclusions drawn here in should not be generalized. In order to make these findings more generalized, extensive analysis is warranted (i.e. study the surge occurrence at all low and high tides without limiting to extreme events). Such analysis is beyond the scope of this study.

2.5.3 Return periods of extreme sea level (MAXSL) events

Return period (R, in yrs) curves describing the annual recurrence probability (where \( P = R^{-1} \times 100 \), in %) of MAXSL events are useful for predicting and planning for coastal hazards in the face of increasing climate variability and climate change. Here, a case study of an extreme sea level event resulting from a high magnitude storm surge on 24th December 2003 is examined and placed into the context of the larger regime of MAXSL events experienced in the region (Figure 2.6). During this event, an intense storm system on the 24th of December
generated strong ESE winds (reaching 111 km hr-1) and, over 33 hours, generated a maximum surge of 0.73 m above the predicted tide in QC-City. The peak surge occurred at low tide and the maximum TWL of 8.06 m (CD) happened 6 hrs later. This event caused extensive coastal flooding, 10s of meters of erosion, and damage to roads and critical infrastructure on NE Graham Island, BC (Figure 2.7).

An extreme water level recurrence curve for the study region was produced by fitting a nonlinear GEV distribution on annual MAXSL events from PR using the NCAR Extreme Value Toolkit (Gilleland et al., 2003) (Figure 2.8). According to this curve, the 24th December 2003 MAXSL event at PR (7.91 m CD) has a recurrence interval of approximately 100 years (or annual P = 1%). This is the highest water level on record at PR since 1909. Note however, that because the peak surge contributing to this MAXSL event occurred at low tide, a less frequent and lower magnitude surge event occurring on a higher tide stage could result in the same, or greater, MAXSL. It is important to note that MAXSL projections based on the annual maxima extreme value method are entirely a function of progressive and episodic events as recorded in past records. Neither the influence of probable future accelerated mean sea level rise trends, nor the effects of episodic sea level extremes resulting from climate variability (e.g., ENSO, PDO regime shifts) are considered. Thus, the MAXSL curve presented here could be considered an underestimate, should CC increase the frequency and/or magnitude of high water events.
Furthermore, the tide dominance in MAXSL records on this macro tidal coast serves to under represent the potential contribution of surges in the GEV annual maxima method. Such effects are likely to result in higher extreme water levels than projected by this GEV model. In other words, the joint occurrence of tides and surges within the extreme value analysis process should be considered. This is beyond the scope of this study and requires more sophisticated methods such as process-based numerical modeling and Joint Probability modeling (JPM) to develop more representative surge-tide extreme water level recurrence curves (cf. Pugh and Vassie 1980; Flather et al., 1998).

2.5.4 Sea level responses to climatic variability events

Another key objective of this study is to identify whether known climate variability signals can explain sea level variations in northern BC coastal waters on seasonal, annual, and inter-decadal scales and, from this, to explore the possibilities for forecasting sea level variability based on the dynamics of climate variability events (e.g., inter annual patterns, regime shifts). Initial correlations (Table 2.1) show significant ($p< 0.01$) linear correspondence between annual MSL and climate variability indices, some showing strong links between CV and sea level response. As expected, there is a very strong negative correlation ($r = -0.865$) between NOI and MEI (as both relate to the ENSO phenomenon). This signals a strong, regional atmospheric response to ENSO fluctuations which is further confirmed by a significant correlation between ALPI ($r = 0.365$) and MEI. In addition, PDO ($r = 0.732$) is strongly positively correlated with MEI, showing that significant regional ocean-atmosphere response to ENSO fluctuations can
span from seasonal (NOI, ALPI) to inter-decadal (PDO) scales. The strong oceanic (SST) influence on atmospheric sea level pressure (SLP) in the NE Pacific is shown by the moderately strong correlation between PDO and ALPI ($r = 0.564$) (Mantua et al., 1997). All climate variability indices examined are significantly correlated to annual MSL (Table 2.1). PDO shows the highest correlation with annual MSL ($r = 0.650$), followed by NOI ($r = -0.577$), MEI ($r = 0.569$), and ALPI ($r = 0.483$).

Multiple regression models describing the linear correspondence between MSL and various climate variability indices are summarized in Table 2.2. The strength of each model is reflected in the overall $R^2$ value, which increases from 0.387 for model A (MEI only) to 0.605 for model C (MEI, ALPI, NOI, and PDO). Interestingly, of these four climate variability indices, PDO is the only significant predictor in model C, as all others become redundant at 95% confidence level. Thus, it seems that annual MSL variability in northern BC coastal waters is most strongly (linearly) correlated with the PDO.

A better description of potential climate variability controls on MSL variability is attained through seasonal correlation analysis (Tables 2.3 and 2.4). In the previous analyses, PDO showed the strongest control over annual MSL. However, according to the seasonal correlations, during the winter season (November to March) MEI and NOI exert a significantly higher climate control over MSL, combined, accounting for more than 60% of the seasonal MSL variability (Table 2.3). The observed high correlation between winter MEI, NOI and MSL suggests stronger atmospheric controls on sea levels in the winter
season vs. during the summer. Interestingly, during summers (April to October) the strongest correlation can be seen between MSL and PDO, signifying weaker atmospheric controls and stronger oceanic control (perhaps due to persistent SST effects) on summer MSL (Table 2.4).

2.5.5 Non-linear relations between sea levels and climate variability indices

Admittedly, estimating the correspondence among climate variability indices and temporally-averaged seasonal MSL values using simple correlations is a simplistic and crude method. It does not provide information on synchronous changes in the trends (e.g., regime shifts), nor does it provide a measure of likely non-linear dependence between variables. To address this, several non-linear statistical techniques were applied.

Generally, residual sea levels correspond to non-linear fluctuations that cannot be captured by regression models. For example, linear regression models for MSL and MAXSL in this study explain only 35% and 12% (respectively) of sea level fluctuations in the record. Thus, a large percentage of both MSL and MAXSL are unexplained (residuals) resulting from non-linear effects. Graphically, a good temporal correspondence exists between residual MSL at PR and PDO and MEI indices (Figure 2.9).

Monthly sea level modulation curves for MEI, PDO and NOI indices were developed to examine short-term responses of MSL to climate variability (Figure 2.10). Sea level responses to all climate variability patterns are most pronounced during the winter season (November to March). In particular, elevated monthly MSL closely corresponds with positive MEI, positive PDO, and negative NOI
values. Paired sample t-tests indicate that these observed differences are significant at the 99.9% level (p<0.001). The influence of PDO on both positive and negative MSL appears to last longer than other climate controls (i.e., until early summer), suggesting a persistent control of sea level variability by PDO, presumably resulting from persistent inter-annual SST (i.e., thermal expansion) effects. This confirms the earlier suggestion of PDO as the key year around climate variability control on MSL variability in the region, although the specific physical mechanism(s) behind this remain unclear.

Superposed epoch analysis (SEA) was conducted to further explore non-linear relationships between climate variability and extreme high and low sea levels on a seasonal scale (Figures 2.11 and 2.12). Almost all climate variability indices show significant deviations for both high (MSL anomaly ≥ 1 STDV) and low (MSL anomaly ≤ 1 STDV) MSL event seasons (lag = 0) compared to randomized results from 1000 Monte Carlo simulations. Results show that positive MEI, positive PDO, and negative NOI are associated with high sea level event seasons and vice versa. All associations at the event season (0 lag) are significant at the 95% confidence threshold and, in some cases, the 99% confidence limit. This provides statistically significant evidence as to high sea level seasons in Northern BC coastal waters being associated with higher than average SST conditions driven by positive PDO and MEI as well as lower than average SLP at NPH and AL, corresponding to negative NOI and positive ALPI (not shown) respectively.
During winter, higher sea levels show the strongest correspondence with positive MEI and negative NOI (significant at 99% confidence levels, Figure 2.11), while during summers, it is with the PDO (Figure 2.12). Combined with the results of linear correspondence between MSL and climate variability indices, it seems that, the persistent SST warming as shown by the multi decadal PDO signal dominates over the seasonal scale SLP effects during the summer months. Results also indicate that, for both seasons, high or low sea level events are not necessarily preceded by significant climate variability signals (either positive or negative), which suggests lack of serial dependence in extreme (high or low) MSL seasons. Thus, advance detection of extreme sea level events based on the climate variability signals here is very limited.

2.5.6 Sea level response to climate regime shifts

CumSum analyses distinguishes decadal scale climate regime shifts from one relatively stable state to another (Figure 2.13). In particular, regime shifts in the PDO are known to cause major, multi decadal climate variability in the NE Pacific (Mantua et al., 1997, Zhang et al., 1997). Although the four climate variability indices considered here reflect different elements and geographical locations of climatic variability, the CumSum analysis shows that they all capture the late 1970s regime shift at approximately the same time (c. 1976). The CumSum curve for PDO also indicates a minor regime shift in 1998. Such regime shifts in the PDO are known to affect the local climate of the NE Pacific by enhancing or suppressing the strength of ENSO signals (Gershunov and Barnett, 1998) and that enhanced ENSO events may be a manifestation of longer-term
CC (Timmerman et al., 1999). This synchronous change of different climate variability indices also suggests a common event or controlling process may be responsible for climate regime shifts (Beamish et al., 1997), although explanation for this is beyond the scope of this study.

Clearly, sea level fluctuations in northern BC respond significantly to decadal scale climatic oscillations. For instance, approximately 90% of the highest MSL event years (MSL anomaly ≥1 STDV) on record (8 out of 9 in total), have occurred during the post-1976 positive PDO phase, while over 75% of the low MSL event years (i.e. MSL anomaly ≤1 STDV) on record (i.e. 7 out of 9 in totals) occurred during the pre-1976 negative PDO phase. In addition, there is also a close lag of about 1-2 years between regional MSL and the 1976 PDO regime shift. This response lag in MSL partially accounts for some of the non-linear variability in the MSL record and reflects complex ocean-atmospheric interactions occurring in the region.

2.6 Conclusions

This study provides clear, statistically significant evidence that MSL variations on the north coast of BC respond to known climate variability and climate change signals, particularly as reflected in regional SST and SLP fluctuations driven by ENSO and PDO. Over the 20th century, the region has experienced a statistically significant, accelerated MSL trend of +1.4±0.6 mm yr⁻¹ while the MAXSL trend is approximately double in magnitude at +3.4 mm yr⁻¹. According to the results of this study, the causes behind this acceleration can be
attributed not only to the effects of global warming but also to cyclic climate variability patterns such as the strong positive PDO phase lasting since mid-1970s'. Since the climate variability controls are cyclic and future climate warming trends continue to alter dramatically, extrapolation of these trends into the 21st century with the underlying assumption of stable climate may hold serious limitations. Thus, for future adaptation and mitigation works it is always recommended to apply the latest available IPCC (i.e. Intergovernmental Panel on Climate Change) global sea level rise projections based on Global Climate Models (GCM).

Non-linear, statistical analysis of MSL and various CV indices confirm potentially significant climate variability influences on MSL at seasonal, annual to decadal time scales (e.g., ENSO, PDO). However, results do not indicate that climate variability index values from preceding seasons can be used reliably to forecast extreme sea level responses into the future. This may explain the poor performance of forecasting models in predicting the intensity of the 1997-1998 El Niño and associated MSL responses before its onset. This would require a better understanding of the linked ocean-atmosphere processes at work as well as their associated lag times and manifestations in the NE Pacific.

Distinct seasonal MSL responses to climate variability events exist. During winters, MSL responds mainly to ENSO related climate forcing (as captured in MEI and NOI indices) while persistent SST effects (via thermal expansion) related to longer-term PDO signals dominate summer MSL responses. Despite seasonal differences, linear correspondence between annual MSL and PDO
index values shows that SST-driven multi-decadal forcing signals best explain overall inter-annual MSL variability in Northern BC. Therefore, advanced detection of PDO regime shifts could substantially improve forecasting for high and/or extreme sea level fluctuations, and related hazards, for the north coast of BC.

Recent research related to climate controls on coastal systems are mostly limited to El Niño conditions, likely due to the common belief that La Niña conditions are more restrictive in occurrence in the north-eastern Pacific (Cayan and Webb, 1992). For example, impacts due to lower than average sea levels resulting from La Niña conditions, on sensitive coastal ecosystems, such as salt marshes and inter tidal zones, are rarely discussed. However, this study shows that sea levels in Northern BC are equally and significantly affected across the full range (i.e. both positive & negative) of climate events under all three climate indices, and therefore may not be limited in their impacts to the strong positive phase (El Niño).

The results of this study show that, the shorter temporal scale climate variability impacts when superimposed on longer term sea level rise trends can be more hazardous to some coastal systems (i.e. depending on the tidal range and coastal orientation) than the longer term sea level rise trends alone (i.e., global warming). Thus, for sea level studies it is equally important to know whether an area is endangered by flooding due to seasonal, annual, intra and inter decadal scale regional climate variability patterns, rather than limiting the problem to the global climate change perspective. Even a systematic change in
the strength and the direction of local winds due to a regional climate regime shift may increase coastal flooding and associated damages.

References


Table 2-1: Pearson’s correlation coefficients for relations between annual average MEI, PDO, ALPI, NOI, and MSL (Prince Rupert) values.

<table>
<thead>
<tr>
<th></th>
<th>MEI</th>
<th>PDO</th>
<th>ALPI</th>
<th>NOI</th>
<th>MSL</th>
</tr>
</thead>
<tbody>
<tr>
<td>MEI</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson's r</td>
<td>.732**</td>
<td>.365**</td>
<td>-.865**</td>
<td>.569**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.000</td>
<td>.007</td>
<td>.000</td>
<td>.000</td>
<td></td>
</tr>
<tr>
<td>PDO</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson's r</td>
<td>.564**</td>
<td></td>
<td>-.721**</td>
<td>.650**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td></td>
</tr>
<tr>
<td>ALPI</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson's r</td>
<td></td>
<td>-.363**</td>
<td></td>
<td>.483**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td></td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td></td>
</tr>
<tr>
<td>NOI</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson's r</td>
<td></td>
<td></td>
<td></td>
<td>-.577**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td></td>
<td></td>
<td></td>
<td>.000</td>
<td></td>
</tr>
</tbody>
</table>

** Correlation is significant at the 0.01 level (2-tailed)
<table>
<thead>
<tr>
<th>Model</th>
<th>Multiple Regression Model</th>
<th>Model $R^2$</th>
<th>Model Sig.</th>
<th>Significance of climate variability index value</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>MSL = m1(MEI) + C</td>
<td>0.387</td>
<td>.000</td>
<td>MEI: .000, ALPI: .000, NOI: .000, PDO: .383, .006</td>
</tr>
<tr>
<td>B</td>
<td>MSL = m1(MEI) + m2(ALPI) + C</td>
<td>0.504</td>
<td>.000</td>
<td>MEI: .000, ALPI: .014, NOI: .000, PDO: .383, .006</td>
</tr>
<tr>
<td>C</td>
<td>MSL = m1(MEI) + m2(ALPI) + m3(NOI) + m4(PDO) + C</td>
<td>0.605</td>
<td>.000</td>
<td>MEI: .724, ALPI: .501, NOI: .383, PDO: .006</td>
</tr>
</tbody>
</table>

Table 2-2: Multiple regression fit statistics for MSL vs MEI, ALPI, NOI and PDO, showing the overall model fit ($R^2$) and the significance of each climate control as predictors.
Table 2-3: Pearson’s correlation coefficients for winter seasonal average (November – March) values of MEI, PDO, NOI and MSL (at Prince Rupert).

<table>
<thead>
<tr>
<th></th>
<th>MEI</th>
<th>PDO</th>
<th>NOI</th>
<th>MSL</th>
</tr>
</thead>
<tbody>
<tr>
<td>MEI</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson’s r</td>
<td>.654**</td>
<td>-.757**</td>
<td>.621**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td></td>
</tr>
<tr>
<td>PDO</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson’s r</td>
<td></td>
<td>-.528**</td>
<td>.497**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td></td>
<td>.000</td>
<td>.000</td>
<td></td>
</tr>
<tr>
<td>NOI</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pearson’s r</td>
<td></td>
<td></td>
<td>-.684**</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td></td>
<td></td>
<td>.000</td>
<td></td>
</tr>
</tbody>
</table>

** Correlation is significant at the 0.01 level (2-tailed)
Table 2-4: Pearson’s correlation coefficients for summer seasonal average (March – August) values of MEI, PDO, NOI, and MSL (at Prince Rupert).

<table>
<thead>
<tr>
<th></th>
<th>MEI</th>
<th>PDO</th>
<th>NOI</th>
<th>MSL</th>
</tr>
</thead>
<tbody>
<tr>
<td>MEI</td>
<td>Pearson’s r</td>
<td>0.731**</td>
<td>-0.851**</td>
<td>0.474**</td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
</tr>
<tr>
<td>PDO</td>
<td>Pearson’s r</td>
<td>-0.739**</td>
<td>0.648**</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>0.000</td>
<td>0.000</td>
<td></td>
</tr>
<tr>
<td>NOI</td>
<td>Pearson’s r</td>
<td></td>
<td>-0.486**</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td></td>
<td>0.000</td>
<td></td>
</tr>
</tbody>
</table>

Note: **, Correlation is significant at the 0.01 level (2-tailed)
Figure 2-1: The geographical location of the study region in the Queen Charlotte Islands (Haida Gwaii) offshore from Prince Rupert, BC.
**Figure 2-2:** Simple linear regression models for annual MSL from 1909-2003 (Thick line) and 1945 to 2003 (Dashed line).
Figure 2-3: Annual MSL variation at Prince Rupert and Queen Charlotte City station from 1966-2003 showing significant similarities in MSL variation.
Figure 2-4: MAXSL at Prince Rupert from 1945-2003. Note that the MAXSL trend is approximately double the MSL trend.
Figure 2-5: Surge-tide components at Prince Rupert, for annual MAXSL events (Triangles) and annual maximum surges (circles), showing the consistent occurrence of largest surges at low to mid tides and MAXSL events dominated by larger tides.
Figure 2-6: Observed total water level (triangles), predicted tidal water level (circle-dashed), and residual (predicted - observed, squares) surge-generated water level at Queen Charlotte City gauging station during the 24 December 2003 storm event. Note that the peak surge (0.73 m) occurred at low tide and the maximum total water level (8.06 m) occurred approximately 6 hrs later.
Figure 2-7: Impacts of the 24 December 2003 storm surge event, eastern Graham Island, BC. Approximately 2.5 m of shoreline was lost at this location along Highway 16 (upper) compromising the road shoulder and bed. Extensive coastal flooding (lower) also occurred, damaging buildings and sending tonnes of drift logs onto nearby roads and properties. (Photos courtesy of Mavis Mark)
Figure 2-8: Extreme sea level recurrence curve for Prince Rupert tide gauge produced using the Extremes toolkit in R.
Figure 2-9: Residual MSL compared with (A) PDO and (B) MEI (ENSO) climate controls.
Figure 2-10: Monthly MSL response to positive and negative (A) MEI, (B) PDO and (C) NOI climate controls compared with overall monthly averages.
Figure 2-11: Winter (Nov.- March) SEA results showing the departure of climate variability signals (MEI, PDO, and NOI) from their corresponding seasonal means (by +/- 1 STDV) during (a) higher and (b) lower than average MSL events (lag = 0). The strength of climate variability influence is constrained to two seasons prior (lag -2 and -1) and following (lag 1 and 2) the event season. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate variability series.
Figure 2-12: Summer SEA results showing the departure of CV signals (MEI, PDO, and NOI) from their corresponding seasonal means (by +/- 1 STDV) during (a) higher and (b) lower than average MSL events (lag = 0). The strength of climate variability influence is constrained to two seasons prior (lag -2 and -1) and following (lag 1 and 2) the event season. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate variability series.
Figure 2-13: CumSum graphs for the four climate indices, MEI, PDO NOI (negative) and ALPI, showing the same distinct trends from 1950-1976 and 1976-2003. Also shown is the CumSum curve for MSL indicating a lag response to the major regime shift occurred in 1976.
3.0 Extreme Sea-level recurrences in the South Coast of BC Canada with Climate Considerations.

3.1 Abstract

This study investigates statistically the impacts of climate variability and change on extreme total water-levels (TWL: total water-level is defined as the combined surge and tidal component at a given time and location) in the southern coast of BC, Canada (Point Atkinson). The method proposed is primarily based on approximating a Generalized Extreme-Value (GEV) distribution to annual maxima total water-level (TWL) series. The influence of global climate change and regional climate variability effects on extreme sea levels are simulated via covariate coefficients. Results indicate considerable fluctuations in the extreme sea level recurrences in response to shorter temporal-scale climate variability, with the warm ENSO mode significantly increasing the TWL return levels for a given return period. In addition, the historical TWL extremes along the BC coast indicate statistically significant tide dominance, with an apparent scarcity of extreme surge interactions with extreme tides.

3.2 Introduction

Interest in the statistical nature of climate variability and change has heightened within the context of extreme coastal environmental drivers (i.e. sea-levels, wind-storms and sea-surges), particularly when considering how the relative frequency and the intensity of extreme events might change as more conventional statistics, such as the mean or standard deviation of the climate
changes. In this study, a recent development in the extreme-value statistical theory is applied to estimate the changes in extreme total water-level (TWL) recurrences in response to global climate change and regional climate variability in the southern coast of BC, Canada (Point Atkinson). Despite the fact that the relationship between the mean sea-level variability and climate has been studied extensively in the eastern Pacific region (Trenberth and Hurrell, 1995; Crawford et al., 1999; Storlazzi et al., 2000; Subbotina et al., 2001; Allan and Komar, 2002; Abeysirigunawardena and Walker, 2008 (Chapter 2)), studies that relates the statistics of extremes to climate variability and change is rare.

The methodology of this study assumes that an extreme environmental variable (i.e. TWL, sea-surges, wind storms etc.) possesses a probability distribution with a location, scale and shape parameters that are influenced by climate variability and change (Coles and Dixon, 1999; Coles, 2001; Gilleland and Katz, 2005; Butler et al., 2007). As such, the impacts of climate change and climate variability on extreme TWL return periods are simulated by extending the traditional GEV model to incorporate the effects of long term relative sea-level rise trends and shorter temporal scale climate variability phenomena such as ENSO, as covariates in the model coefficients. Incorporation of such effects within the analysis provides more physically meaningful extremes that are vital for development of new design criteria for coastal defence structures, and, for establishing coastal management strategies for a future with changing climate in BC. For the analysis, the R package “extremes” was used because of its ability to
incorporate covariate information into parameter estimates (R Development Core Team, 2007).

3.3 Problem definition.

In recent years, more frequent extreme events, in conjunction with rapid relative sea-level rise has caused significant infrastructure damage to many coastal communities in the Fraser Delta region (i.e. the Boundary Bay, Mud Bay and the Westham Island) of southern BC (Figure 3.1). The Fraser Delta is also subjected to land subsidence with apparent increases in relative sea-levels that far exceed the 21st century sea level rise rates from global warming (Church, 2002). This has raised concerns about the adequacy of the existing hard protections in the region against the ever increasing extreme storm impacts resulting from climate variability and change. To this end, this research attempts to apply the statistics of extremes to better describe the relationship between sea-level extremes and climate in the BC south coast, with particular emphasis on the region encompassing the Fraser River tidal flats.

3.3.1 Tide Gauge Data.

This study considers the annual maximum TWL data at Point Atkinson (CHS station 7795; 49.34 N, 123.25W). This station was chosen for the analysis due to its relatively long data record (1915- to date) and proximity to the Boundary Bay region (Figure 3.1). The tide gauge data were sourced from the Marine Environmental Data Service archives of Fisheries and Oceans Canada (MEDS). The TWL measurements in the MEDS archive is expressed with
respect to Chart Datum (CD), a reference datum used on Canadian Hydrographic Service nautical charts to indicate an elevation that water-levels are unlikely to fall below.

3.3.2 Data Pre-processing

First, the mean-sea level anomaly plots were studied to identify and reject clearly anomalous tide gauge data resulting from changes to instrumentation, measurement techniques or observational practices. The approximate 7 cm shift in the annual average MSL anomaly at Pt. Atkinson prior to and after 1949 suggest either a change due to a natural phenomena or a possible shift in the instrumentation location around late 1940s’ (Figure 3.2). In support of the latter, the gauge history report of Pt. Atkinson indicated more than one gauge movements prior to 1949 and no gauge movement (i.e. fixed location) since 1949. Since the exact error components associated with each gauge movement prior to 1949 was unknown, all data prior to 1949 were eliminated from the analysis. This reduced the total available database to 51 years of record (Figure 3.3). A further step of pre-processing consisted of identifying and eliminating years with greater than 10% missing data from the analysis. However, since the extremes generally occur during fall and winter seasons (i.e. October – March), certain years with 100% data coverage from October to March were considered for the analysis even when the annual data coverage was less than 90%.
3.3.3 Long term sea-level trends in southern British Columbia.

The historical data at Pt. Atkinson illustrates an annual mean sea-level trend of +1 mm/yr (at 95% confidence level). However, with projected relative sea-level rise, the rate rise will be much higher due to the effects of local land subsidence and, possible accelerations in the absolute sea level rise as projected by the IPCC (2007) (Church, 2002; Henton et al., 2006; IPCC 2007; Mazzotti et al., 2008). For instance, Church (2002) projects much higher 21st century sea level rise trends for the Fraser river flood plain, a deltaic region in close proximity to Pt. Atkinson (Table 3.1). Such trends signal that the projections of extreme TWL recurrences based on historical data may not exactly show what the future holds for the coastal communities of southern BC. Hence this study attempts to incorporate the projections of future sea level rise trends in extreme value analysis to arrive at more realistic projections of future TWL recurrences. The sea level rise rates projected by Church (2002) were adopted for this purpose. Although, on going studies suggest that relative sea-level rise in this region could be even higher than those projected by Church (2002); these estimates were not available at the time of this analysis (Lambert et al., 2008; Mezzotti et al., (in review)).

3.3.4 Climate Variability Indices.

Climate variability occurs due to systematic variations in the global and regional scale atmospheric pressure, temperature, and sea surface temperature patterns lasting from a year or two to multiple decades (Trenberth and Hurrel
Observational evidence confirms the existence of climate regimes in the NE Pacific linked to climate variability (Mantua et al., 1997; Zhang et al., 1997). These climate excursions can have major impacts on coastal processes, modifying the sea levels, wave climate, coastal erosion and inundation.

This study investigates the relationship between extreme TWL’s and a number of dominant regional and global scale climate variability indices in the NE pacific: (i) The Aleutian Low Pressure Index (ALPI), which is a North Pacific regional index describing the relative intensity of the controlling pressure system in the NE Pacific (i.e. the Aleutian Low) (Beamish et al., 1997); (ii) the Multivariate ENSO Index (MEI) characterizing the full regime of ENSO variations (i.e. El Niño, Neutral and La Niña) in the tropical Pacific with dominant spectral peaks at about 3 and 6 plus years (Wolter and Timlin, 1993, 1998); (iii) the Pacific Decadal Oscillation (PDO), a sea surface temperature pattern characterizing the inter-annual variability in average North Pacific sea surface temperature (Mantua et al., 1997); (iv) the Pacific/North American teleconnection (PNA) describing the low frequency atmospheric variability in the Northern Hemispheric extra-tropics during winters (Leathers and Palecki, 1992; Van den Dool et al., 2000); and, (v) the Northern Oscillation Index (NOI) describing the sea-level pressure anomaly between the North Pacific High (NPH, 35°N and 130°W) region of the NE Pacific and the equatorial Pacific near Darwin, Australia, a climatologically low-pressure region (Schwing et al., 2002).
3.4. Methodology.

The methodology is based on statistical modeling of extremes via maximum-likelihood (ML) estimation in the presence of covariates (Coles and Dixon, 1999; Coles, 2001).

3.4.1 Generalized Extreme-value (GEV) distribution.

The GEV distribution is described as the limiting distribution (Eq. 3.1) of the maxima of a series of independent and identically distributed observations which has a non-degenerate limiting distribution (Leadbetter et al., 1983; Coles, 2001).

\[
G(z; \mu; \sigma; \xi) = \exp\left[-\left\{1 + \xi (z - \mu)/\sigma\right\}_+^{1/\xi}\right]
\]  

(3.1)

where, $-\infty < \mu < \infty, \sigma > 0, -\infty < \xi < \infty$ are the location, scale and shape parameters respectively.

Depending on the sign of the shape parameter ($\xi$), three extremal type distributions exist. The GEV distribution corresponding to $\xi > 0$, has a heavy tail with a polynomial decay so that higher values of maxima are obtained with greater probability (Fréchet). For $\xi < 0$, the distribution is of Weibull type with a bounded upper tail, meaning that there is a finite value which the maximum cannot exceed. The distribution corresponding to $\xi = 0$ is Gumbel type with an unbounded thin tail. In a Gumbel distribution, although the maxima can take on infinitely high values, the probability of obtaining such levels becomes small exponentially (Leadbetter et al., 1983; Coles, 2001).
3.4.2 Parameter estimation method

Different statistical methods are available for performing parameter estimate in the GEV distribution including: Method of Moments Estimation, Probability Weighted Moments or equivalent L-Moments, Maximum-Likelihood Estimation, and Bayesian methods (Hosking and Wallis, 1987). In this study, the Maximum-Likelihood Estimation method is applied exclusively because of its ability to easily incorporate covariate information into the model parameter estimates. Another advantage of the Maximum-Likelihood-Estimation method is that, the approximate standard error for estimated parameters and return levels can be automatically produced, either via the information matrix or through the profile likelihood estimation method (Gilleland and Katz, 2005). Nevertheless, the performance of the Maximum-Likelihood Estimation method is known to be unstable and can give unrealistic estimates for the shape parameter when the sample size is small (i.e. \( n \leq 25 \)) (Hosking and Wallis, 1987; Coles and Dixon, 1999). However, when enough data are available (\( n > 30 \)), Maximum-Likelihood Estimation is comparable in performance to other more advanced estimation techniques (Martins and Stedinger, 2000; Gilleland and Katz, 2005). The theory that underlies maximum likelihood estimation is presented below.

Let \( x \) be a continuous random variable with the probability density function (pdf) of,

\[
f(x; \theta_1, \theta_2, \ldots, \theta_k)
\]

(3.2)

Where \( \theta_1, \theta_2, \ldots, \theta_k \) are \( k \) unknown constant parameters which needs to be estimated via maximum likelihood estimation,
Let \( x_1, x_2, \ldots, x_N \) be a random sample of \( N \) independent observations (i.e. a sample of extremes in the current analysis). Then, the likelihood function is given by the following product:

\[
L(x_1, x_2, \ldots, x_N | \theta_1, \theta_2, \ldots, \theta_k) = L = \prod_{i=1}^{N} f(x_i; \theta_1, \theta_2, \ldots, \theta_k) \quad i = 1, 2, \ldots, N
\]  

(3.3)

Once data have been collected and the likelihood function given by the data determined, the parameter values of the desired probability distribution that maximize the likelihood function can be established. According to MLE principle, this is the distribution that is most likely to have generated the observed data.

For computational convenience, the MLE estimate is obtained by maximizing the log-likelihood function (Eqn 3.4). This is because the two functions, \( \ln L(x | \theta) \) and \( L(x | \theta) \), are monotonically related to each other so that the same MLE estimate can be obtained by maximizing either one.

The logarithmic likelihood function is given by:

\[
\Lambda = \ln L = \sum_{i=1}^{N} \ln f(x_i; \theta_1, \theta_2, \ldots, \theta_k)
\]

(3.4)

Assuming that the log-likelihood function is differentiable, if \( \theta_{\text{MLE}} \) exists, it must satisfy the following partial differential equation (Eqn.3.5), because the definition of maximum or minimum of a continuous differentiable function implies that its first derivatives vanish at such points.

\[
\frac{\partial(\Lambda)}{\partial \theta_j} = 0 \quad j = 1, 2, \ldots, k
\]

(3.5)
However an additional condition should also be satisfied to ensure that $\ln L(x|\theta)$ is a maximum and not a minimum, which the first derivative cannot reveal. To be a maximum the shape of the log-likelihood function should be convex (i.e. a peak when approaching the $\theta_{\text{MLE}}$). This is indicated by a negative value for the second derivative of the log-likelihood.

### 3.4.3 Return Levels (Quantiles)

The extremes of a random variable are typically expressed as the return levels (quantiles) of an extreme event ($Z_p$, in Eq. 3.6 & 3.7), such that, there is a probability of $p$ that $Z_p$ is exceeded in any given year. Alternatively, the same can be expressed in terms of return period of an event, defined as the level that is expected to be exceeded on average once every $1/p$ years (where $1/p$ is referred to as the return period). For example, the 100-year return level at a given location has a 1% (1/100=0.01) chance to occur in any given year.

The return levels of extremes ($Z_p$) are derived by setting the cumulative distribution function in equation (3.1) equal to the desired probability/quantile, 1-$p$, and then, solving for the return levels following equations (3.6 or 3.7),

\[
Z_p = \mu - \frac{\sigma}{\xi} \left[ 1 - \{-\log(1 - p)\}^{-\xi} \right] \quad \text{for} \quad \xi \neq 0 \quad (3.6)
\]

\[
Z_p = \mu - \sigma \log\{-\log(1 - p)\} \quad \text{for} \quad \xi = 0 \quad (3.7)
\]

### 3.4.4 Simulate the effect of relative sea-level trends on TWL extremes.

The effect of longer-term sea-level trend on extremes was simulated by adding a linearly increasing sea-level trend component to the annual maximum
TWL record at Point Atkinson. Two cases with added sea level trend component equivalent to the upper (0.57 cm/yr) and the lower (0.28 cm/yr) bounds of the best available longer term relative sea-level trend projections. was considered (Church, 2002) (Figure 3.4). Return periods of extreme TWLs with future sea level rise trends was simulated by applying GEV models to the modified TWL series, where the effect of long term sea-level rise on extremes was presented as a linear trend covariate (in time) at the location parameter (Eq. 3.8).

\[ \mu(t) = \mu_0 + \mu_1 t, \]  

(3.8)

Here \( t \) denotes time (e.g. in units of days or years) and the location parameter (\( \mu \)) represents the mean of the extremes. Projections of extreme TWL return levels with long-term sea-level rise effects can be done by making the location parameter in equation (3.8) conditional on the return periods of interest.

The rationale in this approach assumes that the historical annual maximums TWL time series at Point Atkinson would mirror into the future in conjunction with the projected future relative sea-level trends. As a result, each annual maximum TWL event in the new series now becomes higher (or lower) than the original event by an amount equivalent to the corresponding relative sea-level trend.

3.4.5 Simulate the effect of climate variability on TWL extremes.

The scope of this task was limited to an investigation of simple linear relationships between climate excursions and extreme TWLs. As such, the dependence between the extremal TWLs and various climate variability indices
(X) were related to the GEV model coefficients (µ, σ and ξ) as linear covariates (Eqn. 3.9).

\[
\begin{align*}
\mu(X) &= \mu_0 + \mu_1X, \\
\ln(\sigma(X)) &= \sigma_0 + \sigma_1X, \\
\xi(X) &= \xi
\end{align*}
\]

(3.9).

Initially each candidate climate indices (i.e. MEI, PDO, PNA, ALPI and NOI) was applied individually as a parameter covariate in the GEV model. The resulting fit diagnostics were tested against those of the “Base model” (i.e. which is the model without climate covariates) for significant improvements (at the 95% level). Any climate covariate that does not generate statistically significant improvement in the modified GEV model is considered superfluous. Those that significantly improve the modified GEV model fit are carried forward to the next level of analysis.

Evidently many climate variability indices have shared patterns (i.e. colinearities) that may lead to compatible responses when applied as individual covariates within a GEV model. For example, the climate pattern that triggered the mid 1976 climate regime shift was captured by all climate indices considered in this analysis (Figure 3.5). Such relationships may cause more climate covariates to qualify as influential parameter covariates when applied as individual entities in the modified GEV model. Therefore, a systematic approach was developed to isolate the climate covariates from the group that best describes the final GEV model with climate considerations. This method successively fits a series of increasingly more complex GEV models to all selected climate indices. In the process, the fit diagnostics of a subsequent
model is cross-compared with the previous model via a likelihood ratio test, to
determine whether the more complicated model is significantly better than the
former. Climate indices having significant contributions towards improving the
model fit at all levels (i.e. without becoming redundant) are chosen for the final
GEV model that simulates the climate effects on extreme TWL recurrences. This
approach essentially avoids repetition of climate patterns that are common to
many climate indices within the GEV models and by that simplify the final GEV
model while satisfactorily accounting for climate variability.

In order to demonstrate the effects of climate variability on extremes and,
to make meaningful inferences of return levels, the finalized GEV model with
randomly varying climate covariates had to be reduced to that of a time varying
function. This was achieved by making the randomly varying climate covariates
conditional on a range of assumed climate variability indices values
corresponding to three distinct climate states as per the Environment Canada
climate classification scheme: (i) the extremely warm (strong El Niño), (ii)
extremely cold (strong La Niña) and (iii) Neutral. Table (3.2) summarizes the
value of each climate indices averaged over the range of years classified under
these three climate states.

3.5 Results.

3.5.1 Extreme TWL recurrences without climate considerations (Base Model).

The maximum-likelihood estimate of -0.231 for the shape parameter ($\xi$) in
the base-model indicates a bounded tail (Table 3.3). This is about 3 standard
errors farther from a typical Gumbel model ($\xi=0$), making it a much weaker case
if a priori fit was constrained to Gumbel. The projected 100-year TWL return level is 560.1 cm (CD) with levels increasing gradually for higher return periods (Table 3.4). Beyond about 100 years, the confidence bounds are notably wide, reflecting the inherent uncertainty associated with making inferences far beyond the range of the data. Further, the TWL return level projections based on the base-model indicate a notable under-estimation when extrapolated beyond 10-20 years (Figure 3.6). The observed under-prediction of the largest extremes by the base-model approach reflects its inability to capture the tail end of the distribution accurately. This may be largely due to the underlying assumption of stable future climate in the conventional GEV approach, where as, the actual observations reflects non-stationarity resulting from decadal to inter-decadal scale climate variability effects.

### 3.5.2 Effect of relative sea-level rise on TWL extremes

The extent to which statistics of extremes change with relative sea level rise is demonstrated via two examples. Each example modifies the same historical extreme-event sequence applied in the base-model by adding a linear trend equivalent to a relative sea-level trend of 0.28 mm/yr and 0.57 mm/yr respectively (Figure 3.4). When the new series are fitted to a GEV distribution with a linear trend covariate in the location parameter ($\mu = \mu_0 + \mu_1 t$), the coefficient of the location parameter successfully captured the positive sea-level trend superimposed on the historical database (Table 3.5). The positive covariate coefficients in the location parameter ($\mu$) ensure a gradual increase in extremes at higher return periods by raising the mean level accordingly. This facilitates the
extension of sea-level trends beyond the limits of the data record by making the linear trend covariate conditional on the return period of interest. Accordingly, if the relative sea-levels rise at a rate of 0.28 and 0.57 mm/yr, respectively, the 100 year return level of 560.1 cm above CD would become 588.1 cm and 617.1 cm, 100 years from now (Table 3.6). This is equivalent to a simple superposition of the expected sea level rise on top of the initial estimates (i.e. with no climate considerations), a common approach many groups use today to model the impacts of long term relative sea-level rise on extremes.

The above two models were applied to assess the adequacy of the existing dike elevations along the Fraser Delta (i.e. Boundary Bay) under anticipated relative sea level rise rates (see Figure 3.1 for the location). The most exposed dikes along this coastline are designed to a height of 664 cm above CD, of which, over 100 cm accounts for the wave set-up, run-up and the freeboard at the dikes. The balance elevation of 564 cm is approximately the 200 year return level of the base- model, suggesting that the existing dike heights are capable of protecting the hinterland form an event with a 200 year return level, provide that the future climate remains stable (Table 3.6). However, when relative sea-level rise is considered as a factor of influence on extremes, the recurrence levels become a function of time and as a result the return levels for a given return period will continue to increase with time. In this example the 200-yr return level could reach as high as 619 cm and 677 cm, 200 years from now, if the sea-levels rise at rates of 0.28 and 0.57 mm/yr, respectively, suggesting more frequent over
3.5.3 Effect of Climate Variability on TWL extremes

The effect of climate variability on extremes was examined by fitting a GEV distribution to the historical annual maximum TWL record at Point Atkinson with climate variability indices as covariates. Of all individual climate variability indices considered, only MEI and NOI as location parameter (μ) covariates indicated significant improvement in the tail distribution of the modified GEV model, compared to the base-model fit (Table 3.7). However, the systematic elimination approach that followed made MEI obsolete in the presence of NOI, suggesting that the actual physical significance of NOI far outweigh the presence of MEI in the statistical model (Table 3.7). The fact that certain climate covariates became redundant in the presence of others made apparent that multiple climate indices described the same climatic phenomenon. The final GEV model coefficients obtained through this process is expressed in equation (3.10).

$$\mu = 523 - 3.2(NOI) ; \sigma = 12.4 ; \xi = -0.229$$

The change in extreme TWL return level projections due to climatic variability is demonstrated by making the modified GEV model’s covariate coefficients as indicated in equation (3.10) conditional on warm ENSO, cold ENSO and Neutral climate states. The average climate index values corresponding to the three distinct climatic states were computed based on the
classification of years with warm ENSO, cold ENSO and Neutral conditions as per Environment Canada climate classification scheme (Table 3.2).

The response of the tail distribution to natural climate transitions between warm (red dashed) and cold (green dashed) ENSO conditions shows warm ENSO (i.e. El Niño) conditions shifting the curve above the base-curve towards higher return levels. Cold ENSO (i.e. La Niña) conditions shift the curve below the base-curve, where the base-curve is defined as the TWL return-period curve without climate considerations (Figure 3.7). The conditional model corresponding to warm ENSO conditions (red dashed) is the "best-fit" GEV distribution consistently following the highest four-TWL records in the historical record (Figure 3.7). Essentially this implies more frequent extreme TWL occurrences (i.e. shorter return periods) during warm ENSOs and less frequent occurrences (i.e. longer periods) during cold ENSOs. This outcome strongly agrees with the findings from similar studies conducted in the region where more frequent occurrences of extreme events have been reported under warm ENSO conditions (Crawford et al., 1999; Storlazzi et al., 2000; Subbotina et al., 2001; Allan and Komar, 2002; Barrie and Conway, 2002).

The likelihood of extreme TWL occurrences in any given year in the face of climate variability indicates a consistently decreasing trend from Warm ENSO towards Cold ENSO conditions (Table 3.8). For instance, TWL elevations corresponding to 1% exceedance in a given year with no climate considerations (560.1 cm), has a 2% change to exceed during a warm ENSO year while the percentage exceedance will reduce to 0.2 % during a cold ENSO year. These
changes are likely due to changes in residual sea levels linked to thermal and salinity variations in the water body and, changes in the magnitude and frequency of storm activity in response to the North Pacific Climate Variability patterns (Storlazzi et al., 2000; Graham and Diaz, 2001; Allan and Komar, 2002; Barrie and Conway, 2002; Schwing et al., 2002; Abeysirigunawardena et al., 2009 (Chapter 5); Abeysrigunawardena et al., (in review),(Chapter 4)). Since these climate variability patterns are short lived (i.e. decadal to inter decadal scale) and it is unclear whether the climate excursion that trigger such changes will continue in to the future, the extreme TWL projections of this nature cannot be extrapolated progressively into the future. Due to this, the changes in the exceedance probabilities of extreme sea surges due to climate variability should always be presented as a percentage exceedance in a given year under a prevailing climate state, as oppose to return periods.

I conclude this section with a real life example, where the highest ever total water-level event experienced by the coastal communities in the Lower Fraser delta region is placed within the context of this analysis. The event not only caused severe damages to the physical infrastructure in the Fraser Delta but also resulted in significant flooding that lasted for days. On December 16th 1982, a powerful southeaster with extreme winds and very low pressures in combination with high tides generated the highest observed sea-level in southern BC, recorded at 560 cm above chart at Point Atkinson.

This event caused extensive damage to a far wider region in southern BC, inundating low-lying areas like the Boundary Bay, Mud Bay and Westham Island
(Figure 3.8). The surge development was calculated as the difference between the predicted and measured tides on the day of the event, where the predicted tides were computed based on 8-major tidal constituents (M2, S2, N2, K2, K1, O1, P1, and Q1). Apparently, the measured extreme total-water-level did not exactly overlap with the peak surge during the event. In fact, the peak surge was leading the highest astronomical tide by approximately 2-3 hours and occurred at rising tide of approximately 306 cm. Had the peak storm surge of 110 cm lagged by 4 hours and overlapped with the 460 cm peak tide, the resulting total water level would have been 572 cm, almost 100 cm higher than the current historical high.

In addition to recording the highest extreme total-water-level, 1982 was also the year with the highest annual average mean sea-level on record at Pt. Atkinson. Pt. Atkinson has had its three highest mean sea-levels' in 1982, 1992 and 1998, the three years with strongest El Niño events. The analysis based on historical data alone (Base case) shows that a flood of comparable height in the southern BC coasts environs should occur about once every 100-years on average (Table 3.4), while rising seas due to global warming could make such disastrous storm events more frequent with time (Table 3.6). The percentage exceedance of a similar event during a typical warm-ENSO year could double in magnitude (i.e. from 1% exceedance to 2% exceedance) while a dominantly cold- ENSO phase could reduce the percentage exceedance by 5-times (i.e from 1% exceedance to 0.2% exceedance) (Table 3.8).
3.5.4 Coastal flooding and extreme sea-surge tide interactions.

Positive surges occurring at high tide often produce coastal flooding. In macro-tidal areas, the impacts of a surge on the coast will depend on the tide level at the moment of the surge peak. Impacts will be at a maximum for positive surges occurring at spring-high-tide, whereas even exceptionally high surges may escape unnoticed when they occur at lower tide levels. Pugh and Vassie (1979) have shown that surges and tides are statistically independent. However, in shallow waters, non-linear tide-surge interactions due to the effects of friction and advection could decrease the surge levels at high water and increase the surge levels on a rising tide (Prandle and Wolf, 1978).

In order to understanding the surge and astronomical tide relationship in Coastal BC, the astronomical tide levels (expressed as percentiles) of the fifty-highest sea surge events at eleven tidal stations (i.e. 550 events in total) along the coastal margin were analysed (Table 3.9). Just 5% of the extreme sea-surge events coincided with an astronomical tide equal to or higher than the 95th percentile tide-level, while 75-80% of the extreme surge events coincided with astronomical tides below the 75th percentile level. This finding suggests a scarcity of extreme sea-surge occurrences at the time of high astronomical tides. To strengthen these findings, the surge-tide combinations of the highest ever occurred sea surge event and highest ever occurred TWL events at each tide gauge station were compared (i.e. 24 events in total). Results showed the selected extreme-sea surge events consistently coinciding with an astronomical tide below the 50th percentile level while at all stations the highest ever occurred TWL events always coincided with astronomical tides exceeding the 95th
percentile level (Table 3.10). Thus it is likely that most of the exceptionally high historical surges in coastal BC have escaped unnoticed due to their occurrence at lower tide levels. The fact that almost 80% of the 50 highest surge events at all stations has coincided with astronomical tides below 75th percentile suggest that serious coastal flooding in BC is governed by extreme astronomical tide levels and not by extreme surges. Yet, an in-depth understanding on the response of extreme sea surges to climate is equally important as this is the water level component that transforms an extreme TWL event from a deterministic to a stochastic process.

3.6 Discussion

This study shows that decadal to inter decadal climatic variability is a fundamental element to explain the changing frequency and intensity of extreme sea levels in coastal BC. These trends make the flooding risk even more important during the next decades, especially if the global sea-levels continue to rise. However, long-term absolute sea-level rise trends would be difficult to predict, as it is likely that the scientists will not develop a consensus on its rate. One example is the wide range of sea level rise noted in IPCC (2007).

Society needs to live with this uncertainty and choose sea level rise rates that include reasonable uncertainties for future developments. When projections are made for actual design purposes, it is advisable to adhere to the absolute global sea level rise rates published by the most recent report of the IPCC. However, the current IPCC projections can have uncertainty and can be low, as it
excludes possible contributions from future-rapid-dynamical-change in ice flow on global sea level rise (IPCC, 2007). Further, the extrapolation of global sea level trends from the 20th century out to 200-years may not be appropriate as it is likely that the climate will be even warmer after 2100 (IPCC, 2007). In future, if the projections of global rise per century exceed 1m, the local variation (i.e. land subsidence due to tectonic effects) of lower order of magnitudes (i.e. millimetres-centimetres) may even become less important when predictions are made at 200-year level. Therefore if the province of BC does need 200-yr extreme total water-level predictions for actual design purposes, a more appropriate higher rate of sea level rise for the 22nd century should be agreed upon.

Knowledge on the response of extremes to climate variability and change is essential in determining how to seriously consider specific adaptation measures against climate change and variability impacts. Thus any progress toward producing an acceptable statistical model that could simulate the effects of climate variability and change impacts on extremes would be a substantial contribution. The goal was to contribute to research efforts that deal with assessing potential impacts of climate variability and change on society. To that end, this study attempted to modify the widely applied conventional extreme value statistical technique based on the assumption of stable climate, to a more physically meaningful approach. This was accomplished by incorporating the effects of trends and cycles resulting from global climate change and local climate variability as covariates.
The results show a significant increase in the frequency of occurrence of extreme TWLs due to longer-term sea-level rise and warm ENSO conditions. Cold ENSO regimes appear to relax such impacts to a certain extent. It is worth noting that possible lead/lag relationships between climate excursions and extreme TWL events were not considered in this study. Such analysis is recommended as it could provide useful information related to the role of climate variability in shifting a prevailing pattern of extreme TWLs from one dominant state to another.

The historical TWL extremes along the BC coast indicate statistically significant tide dominance, with an apparent scarcity of extreme surge interactions with extreme tides. Thus simple extreme value statistical models based on extreme TWLs, with naturally occurring surge-tide associations (i.e. extreme events with statistically significant surge-tide coincidences) are more likely to produce reliable extreme sea-level projections for practical coastal constructions and set back lines.

In summary, this study raises awareness of the possible combined impacts of climate variability and climate change on extremes, particularly as regarded at the level of decision-making. It is anticipated that the approach used here will contribute to the development of realistic adaptation strategies against climate variability and change for many low-lying coastal communities in BC. To that end the accuracy of global to regional climate change and variability projections at local scales is critical and, in fact, turns out to be important in improving the projections of extreme sea level occurrences at local scale. The
challenge that still remains is to find ways to narrow the focus of climate variability and sea-level rise trends to a finer local scale while maintaining the accuracy.

References


Table 3-1: 21st century relative sea-level rise trend projections for the Fraser delta (Church, 2002).

<table>
<thead>
<tr>
<th>Factor</th>
<th>Low Rate (mm/year)</th>
<th>High Rate (mm/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Increase of seawater volume by glacier melt</td>
<td>1.1</td>
<td>2.2</td>
</tr>
<tr>
<td>Thermal expansion of seawater</td>
<td>1.0</td>
<td>2.0</td>
</tr>
<tr>
<td>Land sinking (net effect of sediment + tectonics)</td>
<td>0.7</td>
<td>1.5</td>
</tr>
<tr>
<td>Total</td>
<td>2.8</td>
<td>5.7</td>
</tr>
</tbody>
</table>
**Table 3-2**: Average climate indices for “strong El Niño”, “strong La Niña” and “neutral” years. The definitions of strong El Niño / La Niña years are based on the Environment Canada (Meteorological Services of Canada-The Green Land) classification scheme.

<table>
<thead>
<tr>
<th>MAXSL (CD-cm)</th>
<th>PDO</th>
<th>MEI</th>
<th>NOI</th>
<th>PNA</th>
<th>ALPI</th>
</tr>
</thead>
<tbody>
<tr>
<td>strong El Niño (Warm ENSO) years: 1957, 1982, 1991, 1992, 1998</td>
<td>536</td>
<td>0.219</td>
<td>0.928</td>
<td>-1.486</td>
<td>0.105</td>
</tr>
<tr>
<td>strong La Niña (Cold ENSO) years: 1973, 1974, 1988, 1989, 1999</td>
<td>521</td>
<td>-0.370</td>
<td>-0.688</td>
<td>1.185</td>
<td>-0.081</td>
</tr>
</tbody>
</table>
**Table 3-3**: Maximum-likelihood fitted parameters for the historical annual maximum total water-levels GEV fit at Point Atkinson (Base Model: B).

<table>
<thead>
<tr>
<th>Parameter (B)</th>
<th>Maximum-Likelihood Estimate</th>
<th>Stand. Err</th>
<th>Lower Bound</th>
<th>Upper Bound</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\mu$</td>
<td>522.50</td>
<td>2.06</td>
<td>520.43</td>
<td>524.55</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>13.30</td>
<td>1.44</td>
<td>11.83</td>
<td>14.71</td>
</tr>
<tr>
<td>$\xi$</td>
<td>-0.231</td>
<td>0.09</td>
<td>-0.33</td>
<td>-0.14</td>
</tr>
</tbody>
</table>

Negative log-likelihood: 205.67
Table 3-4: Return levels and 95% confidence intervals at Pt. Atkinson. Projections are based on the GEV fit of historical annual maximum total water-levels (Base).

<table>
<thead>
<tr>
<th>Return Period (Years)</th>
<th>Return Level cm (CD)</th>
<th>Lower bound cm (CD)</th>
<th>Upper bound cm (CD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>527.2</td>
<td>523.0</td>
<td>531.4</td>
</tr>
<tr>
<td>5</td>
<td>539.3</td>
<td>534.9</td>
<td>544.4</td>
</tr>
<tr>
<td>10</td>
<td>545.8</td>
<td>541.0</td>
<td>552.8</td>
</tr>
<tr>
<td>15</td>
<td>549.0</td>
<td>544.0</td>
<td>557.6</td>
</tr>
<tr>
<td>20</td>
<td>551.0</td>
<td>545.9</td>
<td>560.6</td>
</tr>
<tr>
<td>25</td>
<td>552.5</td>
<td>547.2</td>
<td>562.7</td>
</tr>
<tr>
<td>50</td>
<td>556.6</td>
<td>550.7</td>
<td>569.4</td>
</tr>
<tr>
<td>75</td>
<td>558.7</td>
<td>552.4</td>
<td>573.2</td>
</tr>
<tr>
<td>100</td>
<td>560.1</td>
<td>553.4</td>
<td>575.8</td>
</tr>
<tr>
<td>150</td>
<td>561.9</td>
<td>554.7</td>
<td>579.4</td>
</tr>
<tr>
<td>200</td>
<td>563.1</td>
<td>555.5</td>
<td>581.9</td>
</tr>
</tbody>
</table>

*B – Base: based on historical annual events assuming no climate change and constant relative sea-level*
Table 3-5: Maximum-likelihood fitted parameter values of the GEV fit: (a) the historical annual maximum total water-levels at Point Atkinson in combination with 0.28 cm/yr sea-level rise (B+SLR1) (b) historical annual maximum total water-levels at Point Atkinson in combination with 0.57 cm/yr sea-level rise (B+SLR2)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Maximum-Likelihood Estimate (cm)</th>
<th>Stand. Err (cm)</th>
<th>Lower Bound (cm)</th>
<th>Upper Bound (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(B + SLR1)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\mu_0$</td>
<td>523.12</td>
<td>3.94</td>
<td>519.18</td>
<td>527.06</td>
</tr>
<tr>
<td>$\mu_1$ : (Year)</td>
<td>0.28</td>
<td>0.13</td>
<td>0.15</td>
<td>0.41</td>
</tr>
<tr>
<td>$\sigma_0$</td>
<td>13.24</td>
<td>1.44</td>
<td>11.8</td>
<td>14.68</td>
</tr>
<tr>
<td>$\xi_0$</td>
<td>-0.229</td>
<td>0.095</td>
<td>-0.324</td>
<td>-0.132</td>
</tr>
<tr>
<td>Negative log-likelihood: 205.66</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Maximum-Likelihood Estimate (cm)</th>
<th>Stand. Err (cm)</th>
<th>Lower Bound (cm)</th>
<th>Upper Bound (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(B+SLR2)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\mu_0$</td>
<td>523.12</td>
<td>3.94</td>
<td>536.25</td>
<td>2.36</td>
</tr>
<tr>
<td>$\mu_1$ : (Year)</td>
<td>0.57</td>
<td>0.13</td>
<td>0.44</td>
<td>0.70</td>
</tr>
<tr>
<td>$\sigma_0$</td>
<td>13.24</td>
<td>1.44</td>
<td>11.8</td>
<td>14.68</td>
</tr>
<tr>
<td>$\xi_0$</td>
<td>-0.229</td>
<td>0.095</td>
<td>-0.324</td>
<td>-0.134</td>
</tr>
<tr>
<td>Negative log-likelihood: 205.66</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* B+SLR1 – Base with 2.8 mm/yr future sea-level rise
* B+SLR2 – Base with 5.7 mm/yr future sea-level rise
Table 3-6: Return levels under assumed relative sea-level rise trends. The return level projections for the base model are indicated for comparison purposes.

<table>
<thead>
<tr>
<th>Return Period (Yr)</th>
<th>Mean Return Level cm (CD)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>B</td>
</tr>
<tr>
<td>2</td>
<td>527.2</td>
</tr>
<tr>
<td>5</td>
<td>539.3</td>
</tr>
<tr>
<td>10</td>
<td>545.8</td>
</tr>
<tr>
<td>15</td>
<td>548.9</td>
</tr>
<tr>
<td>20</td>
<td>551.0</td>
</tr>
<tr>
<td>25</td>
<td>552.5</td>
</tr>
<tr>
<td>50</td>
<td>556.6</td>
</tr>
<tr>
<td>75</td>
<td>558.7</td>
</tr>
<tr>
<td>100</td>
<td>560.1</td>
</tr>
<tr>
<td>150</td>
<td>561.9</td>
</tr>
<tr>
<td>200</td>
<td>563.1</td>
</tr>
</tbody>
</table>

* B – Base: historical annual events assuming no climate change and constant relative sea-level rise
* B+SLR1 – Base Case with 2.8 mm/yr future sea-level rise
* B+SLR2 – Base Case with 5.7 mm/yr future sea-level rise
Table 3-7: Results of the systematic elimination of climate covariates in the extreme value analysis of annual maximum TWL data at Pt. Atkinson with climate considerations. The test results shows that actual physical significance of NOI as a covariate in the location parameter far outweigh the presence of other climate indices in the GEV statistical model.

<table>
<thead>
<tr>
<th>Model</th>
<th>Parameter</th>
<th>Negative log-likelihood</th>
<th>Significance wrt. Chi-Square Value for 0.05 significance level</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>BASE</td>
<td>-30.3</td>
<td></td>
</tr>
<tr>
<td>2.1</td>
<td>BASE + CV (µ 1 = NOI)</td>
<td>-33.0</td>
<td>5.4&gt;3.84</td>
</tr>
<tr>
<td>2.2</td>
<td>BASE + CV (µ 2 = MEI)</td>
<td>-32.8</td>
<td>5.0&gt;3.84</td>
</tr>
<tr>
<td>3.1</td>
<td>BASE + CV (µ 1 = NOI ; µ 2 = MEI )</td>
<td>-33.2</td>
<td>0&lt;3.84</td>
</tr>
</tbody>
</table>

SELECTED FINAL MODEL

BASE + CV (µ 1 = NOI )
Table 3-8: The extreme total water-level recurrences at Point Atkinson with climate considerations.

<table>
<thead>
<tr>
<th>Mean Return Level (cm) (CD)</th>
<th>Return Period (Year)</th>
<th>Base (B)</th>
<th>B+W</th>
<th>B+N</th>
<th>B+C</th>
</tr>
</thead>
<tbody>
<tr>
<td>527.2</td>
<td>2</td>
<td>50.0</td>
<td>&gt;50</td>
<td>&gt;50</td>
<td>38.9</td>
</tr>
<tr>
<td>539.3</td>
<td>5</td>
<td>20.0</td>
<td>33.0</td>
<td>18.2</td>
<td>11.8</td>
</tr>
<tr>
<td>545.8</td>
<td>10</td>
<td>10.0</td>
<td>17.4</td>
<td>8.3</td>
<td>4.7</td>
</tr>
<tr>
<td>551.1</td>
<td>20</td>
<td>5.0</td>
<td>9.2</td>
<td>3.7</td>
<td>1.8</td>
</tr>
<tr>
<td>556.7</td>
<td>50</td>
<td>2.0</td>
<td>4.0</td>
<td>1.3</td>
<td>0.5</td>
</tr>
<tr>
<td>560.1</td>
<td>100</td>
<td>1.0</td>
<td>2.1</td>
<td>0.5</td>
<td>0.2</td>
</tr>
<tr>
<td>562.0</td>
<td>150</td>
<td>0.7</td>
<td>1.5</td>
<td>0.3</td>
<td>0.1</td>
</tr>
<tr>
<td>563.1</td>
<td>200</td>
<td>0.5</td>
<td>1.2</td>
<td>0.2</td>
<td>0.05</td>
</tr>
</tbody>
</table>

*B – Base: based on historical annual events assuming no climate change and constant relative sea-level*

*B+W – Base with extreme warm ENSO (i.e. strong El Niño conditions)*

*B+N – Base with neutral ENSO*

*B+C – Base with extreme cold ENSO (i.e. strong La Niña conditions)*
Table 3-9: Sea-surge and Astronomical tide occurrences for the 50 highest recorded surges at the 11-tide gauge locations (results expressed to 0 decimals). The results show significant number of extreme sea-surges coinciding with mid to low tides in coastal BC. The highlighted data corresponds to Pt. Atkinson (7795).

<table>
<thead>
<tr>
<th>Tide-gauge Station</th>
<th>Astronomical tide level (AT) Percentile with respect to CD based on 2004 data (cm)</th>
<th>50-highest recorded sea-surge distribution vs. percentile Astronomical tide level (count)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>95&lt;sup&gt;th&lt;/sup&gt;</td>
<td>90&lt;sup&gt;th&lt;/sup&gt;</td>
</tr>
<tr>
<td>7120</td>
<td>274</td>
<td>262</td>
</tr>
<tr>
<td>7277</td>
<td>331</td>
<td>319</td>
</tr>
<tr>
<td>7735</td>
<td>444</td>
<td>428</td>
</tr>
<tr>
<td>7795</td>
<td>446</td>
<td>430</td>
</tr>
<tr>
<td>8074</td>
<td>411</td>
<td>398</td>
</tr>
<tr>
<td>8408</td>
<td>454</td>
<td>431</td>
</tr>
<tr>
<td>8545</td>
<td>322</td>
<td>303</td>
</tr>
<tr>
<td>8615</td>
<td>336</td>
<td>317</td>
</tr>
<tr>
<td>8976</td>
<td>447</td>
<td>423</td>
</tr>
<tr>
<td>9354</td>
<td>619</td>
<td>584</td>
</tr>
<tr>
<td>9850</td>
<td>634</td>
<td>600</td>
</tr>
<tr>
<td>Average</td>
<td>429</td>
<td>409</td>
</tr>
</tbody>
</table>

% with respect to the total events considered (50): 5 4 12 26 38 6 8
Table 3-10: The surge-tide combination of the highest ever occurred sea surges and highest ever occurred TWL events at 11 tide gauge stations in coastal BC. Note that the tide levels are indicated in terms of absolute values and as a percentile with respect to the 2004 astronomical tide levels. The highlighted data corresponds to Pt. Atkinson (7795).

<table>
<thead>
<tr>
<th>Tide-gauge Station</th>
<th>Max Surge Surge vs Astronomical Tide (AT)</th>
<th>Max Total Water-level Surge vs Astronomical Tide (AT)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Max Surge (cm)</td>
<td>AT (cm)</td>
</tr>
<tr>
<td>7120</td>
<td>85</td>
<td>231</td>
</tr>
<tr>
<td>7277</td>
<td>84</td>
<td>240</td>
</tr>
<tr>
<td>7735</td>
<td>124</td>
<td>91</td>
</tr>
<tr>
<td>7795</td>
<td>104</td>
<td>336</td>
</tr>
<tr>
<td>8074</td>
<td>99</td>
<td>145</td>
</tr>
<tr>
<td>8408</td>
<td>75</td>
<td>109</td>
</tr>
<tr>
<td>8545</td>
<td>94</td>
<td>222</td>
</tr>
<tr>
<td>8615</td>
<td>128</td>
<td>191</td>
</tr>
<tr>
<td>8976</td>
<td>97</td>
<td>207</td>
</tr>
<tr>
<td>9354</td>
<td>98</td>
<td>343</td>
</tr>
<tr>
<td>9850</td>
<td>124</td>
<td>272</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td><strong>101</strong></td>
<td><strong>217</strong></td>
</tr>
</tbody>
</table>
Figure 3-1 : The geographical location of the southern coast of BC, including the lower Fraser delta. Also shown is the tide gauge location at Pt. Atkinson. The two ended arrow indicate the coastal region of interest.
Figure 3-2: Pt. Atkinson mean sea-level anomaly expressed with respect to (1915-2006) mean. Note that the anomalies are consistently negative prior to 1949.
Figure 3-3: Annual maximum hourly total water-levels at Pt. Atkinson. Note that based on a t-test, the increasing trend in extremes is statistically not significant.
Figure 3-4: The (a) historical annual maximum sea-level record at Pt. Atkinson, (b) assumed long-term relative sea-level trend as per Church (2002).
Figure 3-5: Temporal distribution of NOI, PDO and ENSO indicating significant colinearities. Note the 1976 major climate regime shift being captured by all climate indices approximately at the same time.
Figure 3-6: Return level graph for the historical annual maximum sea-levels (Base), calculated from associated GEV distribution (black solid line) with (95%) confidence intervals calculated from the delta (blue solid) and profile-likelihood (dashed–red) methods. Note that the Base model indicating an apparent under-estimation of extremes TWL recurrences when extrapolated beyond 10-20 years
Figure 3-7: Return level graph based on GEV distribution for (a) A future climate representing strong warm ENSO (El Niño) conditions (red–dashed line) (b) A future climate representing Neutral conditions (blue–dashed line) (c) A future climate representing strong cold ENSO (La Niña) conditions (green–dashed line). For comparison purposes, the mean return level graph for the Base Case with no climate considerations is plotted (black solid line). Note that the average climate variability index values (i.e. MEI, PDO and NOI) applied for each case is the average index values of all years with strong El Niño, Neutral and strong La Niña years from (1950-2006) respectively. The definition of years was based on Environment Canada Classification.
Figure 3-8: Surge-tide relationship and the development of the extreme surge event of 16th December 1982 (top panel). Impacts of the storm surge event in southern BC are shown (bottom panels). Extensive coastal flooding occurred in Boundary Bay, Mud Bay and Westham Island and resulted in highest ever occurred water levels in southern BC. (Photos courtesy of Fraser Delta Engineering Department of the BC MOE)
4.0 Extreme Sea-Surge Responses to Climate Variability in Coastal British Columbia Canada.

4.1 Abstract

This chapter presents a statistical investigation of the spatial and temporal changes of extreme sea surges in response to climate variability in coastal BC. The study was based on an investigation of in-situ hourly tide gauge data spanning the interval from 1950 to 2007 at eleven tide gauges in the region. The characteristics of the recorded extreme sea surges were parametrically modeled by the Generalized Extreme Value distribution while treating for trends and dependence via climate covariate coefficients. The study confirms that decadal to inter-decadal climatic variability is a fundamental element in explaining the changing frequency and intensity of sea surges in coastal BC. For instance, the sea-surge response to climate variability impacts on sites along the BC coast were found to be fairly synchronous, with an increase (decrease) in the magnitude of extreme sea surges in response to warm (cold) ENSO conditions. These trends make the flooding risk even more important during the next decade, especially if the global sea-levels continue to rise as predicted by climate models.

4.2 Introduction

Extreme sea-surges are one of the greatest natural threat to coastal communities, in terms of property devastation and lives lost (Murty, 1988). Typically, “sea surges” (also referred to as the residual sea levels) consist of two
meteorological components namely the “wind setup” and “barometric setup” and, a third concurrent factor resulting from the thermal expansion (contraction) of water due to warmer (colder) than normal climatic conditions (i.e. ENSO conditions). Once a critical mean sea level threshold is exceeded, the economic, human and ecological costs of even a small increase in extreme sea surges due to climate variability could be much higher on low lying coastal communities, (IPCC, 2007). Given this, any evidence of possible links between extreme sea surges and decadal to inter-decadal scale climate variability modes would set the safety bar even higher when it comes to defining acceptable levels of protection against extreme events.

Recent occurrences of extreme sea-surge events in conjunction with progressive mean sea-level rise are resulted in significant infrastructure damage to many coastal communities in BC (Walker et al., 2007; Thomson et al., 2008). These events have raised concerns about our lack of understanding on the relations between extreme sea surges and the changing climate in BC. This study is primarily aimed at developing a better understanding of the temporal and spatial distribution of extreme sea surges and the response to various regional climate variability modes by statistically exploring eleven historical long-term water level records in coastal BC (Figure 4.1).

The study methodology follows a recent development in the extreme-value statistical theory, where the conventional Generalized Extreme Value distribution (GEV) (Jenkinson, 1969; Leadbetter et al., 1983; Coles, 2001) is applied to parametrically model the multi-year annual maximum sea surge events while
accounting for trends due to climate variability via climate covariate coefficients (Gilleland and Katz, 2005). The sea surge magnitudes applied for the statistical analysis are computed by taking the concurrent differences between the observed in-situ hourly tide gauge data and the predicted astronomical tidal levels.

4.3. Data Pre-processing:

4.3.1 Tide Gauge Data Quality Assurance.

Hourly water level data from eleven tide gauge sites with data covering at least twenty years or more were considered for analysis (Figure 4.1). The data originated from the Marine Environmental Data Service archives of Fisheries and Oceans Canada (MEDS). The station name coordinates and acronyms are indicated in Table (4.1).

Potential problems due to large data gaps were eliminated by discarding years with greater than 10% missing data. Since extremes generally occur during fall and winter seasons (i.e. from October to March of the following year), years that satisfied 100% monthly data coverage from October to March were included in the analysis, even when the annual data coverage was less than 90%. The presence of in-homogeneities and data abnormalities were identified by comparing the mean sea level anomaly plots against the gauge history reports. Appropriate corrections were applied when sufficient information was available to justify possible heterogeneity resulting from gauge shifts. Some stations indicated ambiguous shifts in the annual MSL anomalies suggesting possible temporal offsets associated with instrument problems or site relocations. Most of
the shifts could be linked to gauge movement or gauge corrections via the gauge history information. Accordingly, the data for Point Atkinson (7795), Tofino (8615), Vancouver (7735) and Bamfield (8545) indicated possible offsets associated with the Mean Sea level (MSL) linked to gauge shifts, while Patricia Bay (7277), Campbell River B.C (8074) Bella Bella, B.C. (8976) indicated significant offsets between estimated MSL (i.e. Z0 in the tidal constituents) and the calculated MSL based on observed data.

4.3.2 Sea-Surge Computations

The observed water level at any given station has a deterministic (i.e. astronomical tides) and, a stochastic (sea-surge) component. Astronomical tidal variations are deterministic, provided that all the amplitude and phase values of the gravitationally induced constituent tides are known at the location (Foreman, 1977). On the other hand the sea-surges or the residuals water level component could vary stochastically depending on the prevailing atmospheric pressure, winds and ENSO effects.

For the current work, the sea surge (residual) time series at each site was computed by subtracting the predicted astronomical tidal level from the measured total-water-levels at the tide gauge. The station specific astronomical tides were based on the tidal constituents developed at the Department of Fisheries and Oceans (DFO) (Foreman, 1977). To eliminate unrealistic sea-surge anomalies (i.e. events that appear or disappear suddenly, independently of meteorological events) from the analysis, each annual maxima sea-surge event
was compared against the corresponding residual water levels at the closest tide
gauge stations, and with meteorological station data across two tidal cycles.

4.3.3 Climate Variability Indices.

Climate variability in the NE Pacific is characterized by a number of climate
variability indices, such as the Multivariate ENSO Index (MEI), Pacific Decadal
Oscillation (PDO), Northern Oscillation Index (NOI), Aleutian Low Pressure Index
(ALPI) and Pacific North American teleconnection (PNA). Previously, climate
variability modes such as the ENSO and PDO have been directly linked to inter-
annual sea-level variations along the eastern Pacific coastal margin (Trenberth
and Hurrell, 1995; Crawford et al., 1999; Storlazzi et al., 2000; Subbotina et al.,
2001; Allan and Komar, 2002; Abeysirigunawardena and Walker, 2008 (Chapter
2)). Given this, the extreme sea surges recurrences along coastal BC are
expected to shift in conjunction with these climate variability modes as well.

This study investigates the relationship between extreme sea surges in
coastal BC and a number of dominant climate variability modes in the NE Pacific.
The Aleutian Low Pressure Index (ALPI) is a NE Pacific regional index which
indicates the relative intensity of the controlling pressure system in the NE Pacific
and therefore indirectly describes the relative strength of regional wind patterns.
The index is calculated as the mean area (km$^2$) with sea-level pressure less than
or equal to 100.5 kPa expressed as an anomaly from the 1950-1997 mean
(Beanish et al., 1997). Positive ALPI index values reflect a relatively strong or
intense Aleutian Low while negative values indicate the opposite. ALPI values
are published online by the Canadian Department of Fisheries and Oceans <www.pac.dfo-mpo.gc.ca/sci/sa-mfpd/downloads/alpi.txt>.

The classification of ENSO events in the present study is based on the National Oceanic and Atmospheric Administration’s (NOAA) Multivariate ENSO Index (MEI). (Wolter and Timlin, 1993, 1998). MEI is defined as the weighted average of a number of tropical Pacific environmental variables including sea surface temperature, east-west (i.e. zonal) and north-south (i.e. meridional) surface winds, sea-level pressure , and cloudiness. Negative MEI values represent the cold ENSO phase (La Niña) while positive MEI values represent the warm ENSO phase (El Niño). Monthly MEI values are published online at NOAA-CIRES Climate Diagnostic Center’s Comprehensive Ocean-Atmosphere Data Set (COADS) <www.cdc.noaa.gov/people/ klaus.wolter/MEI/>.

The PDO index characterizes interannual variability in average North Pacific sea surface temperature and, reflects NE Pacific regional climatic variability (Mantua et al., 1997). PDO is well correlated with many records of North Pacific and Pacific Northwest climate and ecology, including sea-level pressure, winter land–surface temperature and precipitation, and stream flow (Mantua et al., 1997; Zhang et al., 1997; Hessl et al., 2004; Schneider, and Cornuelle, 2005). Predominantly positive or negative sea–surface temperature anomalies along the Pacific coast of North America characterise the phase of the PDO pattern as being in a "warm" or "cold" state respectively. Monthly PDO index values are published online at the Joint Institute for Study of the Atmosphere and Ocean’s PDO site <www.jisao.washington.edu/pdo> (Mantua et al., 1997).
The Northern Oscillation Index (NOI) is the sea-level pressure anomaly between the North Pacific High (NPH, 35°N and 130°W) region of the NE Pacific and the equatorial Pacific near Darwin, Australia, a climatologically low-pressure region (Schwing et al., 2002). The NOI is dominated by the interannual variations of ENSO, such that large positive (negative) NOI are associated with La Niña (El Niño) (Schwing et al., 2002). As NOI is partially based in the NE Pacific, it provides a more direct connection between various climate processes in the NE Pacific and remotely teleconnected global climate events. The NOI time series is available online from NOAA's Pacific Fisheries Environmental Laboratory. <http://www.pfeg.noaa.gov>.

The Pacific/North American teleconnection (PNA) describes one of the most dominant modes of low frequency variability in the Northern Hemisphere extra-tropics during North Hemisphere winter (Van den Dool et al., 2000). The PNA index is constructed through a linear combination of normalized 700-mb height anomalies nearest to the nominal centers of the PNA pattern (Leathers and Palecki, 1992). The PNA pattern occurs as a result of an amplification or damping of the mean flow configuration over the unique geography of the Pacific Basin and North America (Leathers and Palecki, 1992). PNA is strongly influenced by the mode of ENSO, where the positive phase of PNA tends to be associated with warm ENSO and the negative phase with cold ENSO.
4.3 Methodology

The methodology of this study is based on the extreme value theory, a widely applied approach for drawing inferences about the extremes of a stochastic process based on independent extremes of a process (Coles and Dixon, 1999; Coles, 2001; Butler et al., 2007). Two main approaches are proposed in the literature: the Block Maxima models (ex. Annual maxima) and the Peaks-over-Threshold models (POT). A summary of these techniques can be found in Palutikof (1999). In this work, a GEV distribution is fitted to annual maxima sea surge data at each site using the Block (i.e. Annual) Maxima models. The impacts of climate variability on extreme sea-surge recurrences are simulated by incorporating the dominant climate variability indices as model parameter covariates. For the analysis, the R (R Development Core Team, 2007) package “extremes” was used because of its ability to incorporate covariate information into parameter estimates (Gilleland, and Katz, 2005).

4.4.1 Generalized Extreme-value (GEV) distribution.

The extreme value theorem describes the Generalized Extreme-value (GEV) distribution as the limit distribution of maxima of a sequence of independent and identically distributed random observations. Thus the GEV distribution can be used as an approximation to model the maxima of long (finite) sequences of random variables. (Leadbetter et al., 1983; Coles, 2001) (Eq. 4.1).
\[ G(z; \mu; \sigma; \xi) = \exp \left[ - \left\{ 1 + \frac{\xi(z - \mu)}{\sigma} \right\}^{\frac{1}{\xi}} \right] \]  
(4.1)

Where \(-\infty < \mu < \infty, \sigma > 0, -\infty < \xi < \infty\) are the location, scale and shape parameters respectively.

The sign of the shape parameter (\(\xi\)) determines the GEV distribution type (Leadbetter et al., 1983; Coles, 2001).

(i) \(\xi > 0\) corresponds to a Fréchet type distribution having a heavy tail with a polynomial decay so that higher values of maxima are obtained with high probability.

(ii) \(\xi < 0\), corresponds to a Weibull type distribution with a thin unbounded tail, meaning that there is a finite value which the maximum cannot exceed.

(iii) \(\xi = 0\) corresponds to a Gumbel type with an unbounded thin tail

In this study, the GEV parameter estimation was based on the Maximum-Likelihood Estimation method (MLE), because of its ability to easily incorporate covariate information into the model parameter estimates (Hosking, and Wallis, 1987). Another advantage of the Maximum-Likelihood Estimation method is that, approximate standard errors for estimated parameters and return levels can be automatically produced, either via the information matrix or through the profile likelihood estimation method (Gilleland and Katz, 2005). Nevertheless, the performance of the Maximum-Likelihood Estimation method is known to be unstable and can give unrealistic estimates for the shape parameter when the sample size is small (i.e. \(n \leq 25\)) (Hosking and Wallis, 1987; Coles and Dixon,
1999). However, when enough data are available \((n > 30)\), Maximum-Likelihood Estimation is comparable in performance to other more advanced estimation techniques (Martins and Stedinger, 2000; Gilleland, and Katz, 2005).

The magnitude of extremes are often expressed as the return levels of an extreme event \((Z_p)\), such that there is a probability \((p)\) that \(Z_p\) is exceeded in any given year. \(Z_p\) is obtained by inverting the fitted distribution function in equation (1) and solving for the return levels following equations (4.2.1 or 4.2.2),

\[
Z_p = \mu - \frac{\sigma}{\xi} \left[ 1 - \left(\frac{-\log(1 - p)}{\xi}\right)^{-\frac{1}{\xi}} \right] \quad \text{for} \quad \xi \neq 0 \tag{4.2.1}
\]

\[
Z_p = \mu - \sigma \log\left(\frac{-\log(1 - p)}{\xi}\right) \quad \text{for} \quad \xi = 0 \tag{4.2.2}
\]

The probability \((p)\) is typically expressed as \((1/p)\) years, where \((1/p)\) is referred to as the return period. For example, the 100-year return level \((Z_{100})\) at a given location has a 1\% \((1/100=0.01)\) chance to occur in any given year.

4.4.2 Simulate the effect of climate variability on sea-surge extremes via covariates.

Since this study is the first to investigate the relationship between extreme sea surge recurrences and climate variability in Costal BC, the scope is limited to simple linear relationships between climate variability indices and extreme sea surges. As such, the climate-variability indices \((Y)\) are related to the GEV model coefficients \((\sigma, \xi)\) as linear covariates (Eqn 4.3):

\[
\mu(X) = \mu_0 + \mu_1 Y, \quad \ln \sigma(X) = \sigma_0 + \sigma_1 Y, \quad \xi(X) = \xi_0 + \xi_1 Y \tag{4.3}
\]
4.4.2.1 The dependencies between climate indices

Initially the candidate climate indices (i.e. MEI, PDO, PNA, ALPI and NOI) are included as individual covariates in the GEV model. The fit diagnostics of each model was tested for significant improvements (i.e. at the 95% confidence level) against the “base model” (i.e. the corresponding model without climate covariates). Any individual climate covariate that did not show statistically significant improvements in the modified GEV model fit against the “base model” fit was considered superfluous, while those that significantly improve the model fit were carried forward to the next level of analysis.

4.4.2.2 Assessing the comparative performance of the GEV models with climate covariates.

Evidently many large-scale climate regimes described by various climate variability indices are related. For example, the strong climate regime shift that occurred in mid 1976 was captured by all climate indices considered in this analysis at approximately at the same time, suggesting that the climatic phenomena that triggered the regime shift has affected a wide range of climate indices (Figure 4.2). Due to these interdependencies, climate indices when applied as individual covariates within a GEV model may show very similar responses. On the other hand, when extreme surge climatology in a region is significantly affected by climate variability characteristics unique to a climate index, they should indicate comparatively better GEV model performance over others. Following this, the “Likelihood ratio test” was applied to test the comparative performance of different GEV models in the presence of correlation...
between the explanatory variables (i.e. climate variability indices). In this approach, the GEV models were made more complex systematically by increasing the number of explanatory variables in successive models.

This approach aids in choosing the GEV model that best describes the data without having more parameters than necessary. Each time the fit diagnostics of each successive model is cross-compared with the former via a likelihood ratio test, in order to determine whether the new model is significantly better than the previous one. The combination of climate indices that best describes the model fit are chosen for the final GEV model to simulate the climate effects on extreme sea surge recurrences. This approach avoids repetition of patterns that are common to many climate indices within the GEV models and simplifies the final GEV model. Table (4.2) demonstrates the application of this test on extreme sea surge data at station 7120.

4.4.2.3 Inferences under different climate states

In order to make meaningful extreme sea surge return level inferences from a GEV model with randomly varying climate variability covariates, (see Eqn. 4.3) the models had to be reduced to a time dependent function by setting the climate variability indices (i.e. covariates) to a set of fixed average values conditional on three distinct climate states. The climate states were chosen based on the Environment Canada climate classification scheme as: (i) the extremely warm (strong El Niño); (ii) extremely cold (strong La Niña); and, (iii) Neutral. Table 4.3 summarizes the corresponding climate indices averaged over the range of years classified under these three climate states.
4.5 Results.

4.5.1 Sea surge dependencies in coastal British Columbia

Table 4.4 demonstrate the overall average monthly maximum and monthly mean positive sea-surges (i.e. residual water levels) along the BC coast since 1950. The statistics are comparable with less than 2 cm standard deviations between the eleven stations. None of the station records indicated extreme residuals exceeding the 2 m mark during this period. The annual cycle of the mean and maximum sea-surges at all stations indicate a 40-50% higher peak surge magnitudes during the fall and winter seasons (i.e. from October to March) compared to those occurred from April to September, while December being the month with the strongest sea surges.

Correlation analysis of annual maximum sea surges from 1950-2007 based on the Pearson product-moment and the Sperman’s rank correlation coefficients indicate two groups of tide gauges having very similar extreme sea-surge variability patterns (Table 4.5). For instance, tide gauges located on Vancouver Island and in the Strait of Georgia showed significant (at 99% confidence level) correlations (correlation between 0.7 and 0.95). Such strong spatial dependencies can lead to flooding occurring simultaneously along entire reaches from a single event. The correlation does however seem to decrease with increasing latitudinal distance from the southern most stations. Stations 9850, 9354 and 8976 located in proximity to Hecate Strait indicate only moderate to weak correlation (0.3-0.4) with the southern Vancouver Island stations. Understanding these spatial dependencies would be beneficial towards improved reallocation of monitoring sites.
4.5.2 Extreme sea-surge exceedances in coastal British Columbia

4.5.2.1 GEV model approximations without climate covariates.

The GEV approximations without climate covariates (hereafter referred as the Base Model) indicated an upper limit for the extreme sea-surge magnitudes for all but one station (i.e. station 9850) (Table 4.6). Station 9850 has been identified by DFO as one with significant shallow water effects. Although observed annual maxima sea surges fall well within the 95% confidence limits in all the GEV approximations, there is an apparent under-estimation of the GEV approximations at the tail end. The extreme sea surge return level projections shows just three stations (i.e. stations 8615, 8074 and 9850) exceeding the 1 m mark for the 100 year return level (Figure 4.3). The return level magnitudes at stations located in close proximity (i.e. 7120 & 7277, 7735 & 7795) are comparable to each other (Table 4.7).

4.5.2.2 GEV model approximations with climate covariates.

Certain stations indicated significant improvement in the GEV approximations with individual climate covariates, compared to their base fits (Table 4.8). Over 55-percent of the stations showed statistically significant model improvement when PNA and NOI are applied as location parameter covariates. Both the indices are known to describe large scale pressure patterns directly affecting the NE Pacific SST and storm frequency. The PNA has a strong influence on the strength and location of the jet stream which influence storminess in coastal BC. The MEI which describes the remotely teleconnected oceanic and atmospheric patterns appears to have significant influences on scale
and shape parameter while PDO shows the least amount of impact on extreme sea surges in coastal BC.

Generally speaking, the location parameter which defines the mean of the extreme residuals within the GEV model is the most affected by climate variability (i.e. over 60-percent of the stations). The least affected is the scale parameter that describes the skewness of the models. As far as the sign of the climate covariate coefficients are concerned, there is a notable similarity in the location parameter at all stations, suggesting spatially consistent climate variability impacts on extreme sea surges in coastal BC. A strong presence of ALPI, NOI, PNA and MEI is seen among the final extreme value statistical models (GEV) at each station (Table 4.9). The fact that certain climate covariates became redundant in the presence of others suggests that multiple climate indices describe the same climatic phenomenon.

4.5.2.3 Extreme sea-surge exceedances under ENSO conditions.

Using the modified GEV approximations and the average climate index values for warm, neutral and cold ENSO conditions (Table 4.3), the sensitivity of the exceedance probability of extreme sea surges to climate variability was examined. For illustration final results for station 7120 are shown (Figure 4.4). According to the results, for a given return period, nine out of the eleven stations (> 80%) indicated higher (lower) residual water level occurrences under Warm (cold) ENSO conditions, compared to the base case (i.e. projections without climate considerations). The station specific extreme sea surge magnitudes with
1% probability of exceedance reaching 1 m mark would increased from 33% in the base model to 56% under warm ENSO conditions (Table 4.10).

Even though climate variability modes are known to be short lived, they generally prevail long enough (i.e. decadal to inter decadal scale) to impose significant impacts on vulnerable coastal margins and infrastructure. This study clearly demonstrates how an apparent underestimation of extreme sea surge return-levels could result, when climate variability impacts are not included in the GEV approximations, particularly under warm ENSO phase.

4.5.3 Spatial dependencies of extreme sea surges in coastal BC.

A strong spatial dependency of extreme sea surges can lead to flooding occurring simultaneously along entire reaches of coastline from a single event. Two surge events were chosen to demonstrate the risk of concurrent occurrence of extreme sea surge events in coastal BC. The spatial dispersion of the two events was assessed via a comparison of the return levels of the maximum sea surges and the duration of the events.

The first case study is a high magnitude sea surge event that occurred on December 24th 2003 (hereinafter “12/24”). This event resulted in extensive damage to the eastern coast of Graham Island Queen Charlotte Island. During the 12/24 event, an intense storm system resulted in strong south-easterly winds reaching 111 km hr-1, generated a maximum surge of 0.73-m above the predicted tide at tide gauge station 9850. The peak surge occurred at low tide and the maximum Total Water level (TWL) of 8.06-m (CD) happened five hours later (Figure. 4.5). This event caused extensive coastal flooding, tens of meters
of coastal erosion, and damage to roads and critical infrastructure. The second case study is based on an event that resulted in the highest extreme total water-level event on the inner south coast of BC. The event occurred on the December 16th 1982 (hereinafter “12/16”), when a powerful southeaster with average winds of 50 km/hr, and high tides generated a high water mark of 5.6-m (CD), at station 7795 (Figure 4.6). The surge developed at each of these stations was examined with respect to the day prior to the event, the event day and the day after the event (i.e. 72 hours).

The low annual probability of exceedance (<10%) at stations 7120, 7735, 7795 indicates that an event typical of 12/16 is capable of generating exceptionally high surges over the southern BC coastline, while the surges generated by the same event further north was weak with annual probability of exceedance above 70-percent (i.e. 8976, 9850 and 9354). Thus an event typical of 12/16 may not cause wide spread flooding along the entire BC coast.

Even though the winds were blowing from the southeast sector during both events, the extreme surge exceedance corresponding to the 12/24 event indicated a different spatial distribution of surges compared to 12/16 event. For instance the event indicated a gradual transition in the annual percentage exceedances of the maximum surge elevations from low (i.e. rare intense events), moderate to high (i.e. weak frequent events) values towards the south.

The two events signify distinctly different spatial dependencies of storm surge occurrences in southern and northern BC coastlines. This in combination with the existing weak correlation between the surges in northern and southern
coastal stretches suggest that the possibilities of simultaneous flooding occurring along the entire BC coastal reach from a single sea surge event may be rare. The location of the jet stream may be a determining factor of the location and the spatial extent of damage due to surges in coastal BC. As demonstrated above, a combination of maximum sea surge return levels corresponding to the same event at a number of tide gauge stations that are spatially apart provides excellent indications as to the spatial dispersion and dependencies of these events. Similar knowledge on many events may facilitate improvements in spatial predictions and the planning and designing of sea level monitoring networks.

4.5.4 Coastal flooding and surge-tide interactions.

Positive surges occurring at high tide often produce coastal flooding. In macro-tidal areas, the impacts of a surge on a coast will depend upon the tide level at the time of the surge peak. Impacts will be at a maximum for positive surges occurring at spring high tide, whereas even exceptional surges may escape unnoticed when they occur at low tide levels. Pugh and Vassie (1979) have shown that surges and tides are statistically independent. Tide-surge interactions can be non-linear due to the effects of friction and advection. In an area with shallow waters and large tidal range such interactions could decrease the surge levels at high water whilst increasing surge levels on the rising tide (Prandle and Wolf, 1978).

An analysis of the astronomical tide levels corresponding to the 50-highest independent sea surge events on record at the 11-tidal stations (550 events in total) indicated that, on average, just 5% of the highest fifty surges coincided with
an astronomical tide level equal to or higher than the 95th percentile level. On the other hand about 75-80 percent of extreme surges coincided with astronomical tides below the 75th percentile level. This suggests a scarcity of high surges occurring at the time of high astronomical tides in coastal BC (Table 4.12).

While the threat of an extreme surge coinciding with a spring high tide is ever present (5% probability), the fact that almost 80% of the 550 highest surge events analysed coincided with astronomical tides below 75th percentile level suggest that serious coastal flooding in coastal BC is governed by extreme astronomical tide levels rather than by extreme sea surges. For instance, the astronomical tide corresponding to the highest ever occurred sea surge event at each tide gauge in coastal BC (11 events) indicated tidal levels below the 50th percentile, while the highest ever occurred total water level event (11 events) coincided with astronomical tidal levels exceeding the 95th percentile level. Thus it is likely that most of the exceptionally high historical surges in coastal BC have escaped un-noticed due to their occurrence at lower tide levels (Table 4.13).

4.6 Discussion and Conclusions

The key focus of this chapter was to establish the extreme sea surge climatology in coastal BC by properly accounting for climate variability and change signals embedded in the observational data. The methodology provided a flexible and statistically rigorous approach for quantifying changes in the extremal properties of sea-surge processes by incorporating more physical information into the extreme value analysis. In particular, I have attempted to
reduce the uncertainties in the parameter estimates by explicitly incorporating knowledge of linear dependencies to climate variability and change. The results of this study suggest that the benefits of an analysis of this nature could be substantial within an oceanographic context. One key practical issue revolves around the fact that climate patterns are short-lived (i.e. decadal to inter decadal scale) and, as a result, it is unclear how the climate excursions that trigger changes in extreme sea-surges will continue. Since this is the case, projections cannot be extrapolated indefinitely into the future. Thus, in practical applications, the outcomes should be expressed as percentage exceedance in each year linked to a prevailing climate state (i.e. warm, cold or Neutral ENSO’s).

The overall strategy for the final model selection with climate covariates may seem simple, due to its initial model selection being performed separately for each of the three GEV parameters (location, scale and shape). Admittedly, these parameters are quite highly correlated, so that a certain covariate included in a model parameter is likely to impact on which covariates to be included in the other model parameters. Acknowledging this limitation, it is recommended to develop more advanced approach where different combinations of covariates are simultaneously incorporated into different GEV parameters using the likelihood ratio tests. In this approach the number of possible models would become large and the process of testing would need to be automated.

The historical data shows an increase of extreme sea-surge exceedances under warm ENSO conditions. Sea-surge signals in contrast, appear to be weak during cold ENSO’s. The most pronounced changes to the sea surge climatology
across coastal BC are caused by the ALPI, NOI, PNA and MEI climate excursions. These observed changes may be largely attributed to changes in residual sea levels resulting from warmer oceans and changes in storm activity due to shifts in the Pacific storm track in response to the north pacific climate variability patterns (Trenberth and Hurrell, 1995; Crawford et al., 1999; Storlazzi et al., 2000; Subbotina et al., 2001; Allan and Komar, 2002; Abeysirigunawardena and Walker, 2008 (Chapter 2)). No significant and consistent long term sea-surge trends were apparent.

Sea surge occurrences in coastal BC show two distinct spatial dependencies among tide gauges located in the Hecate Strait and Georgia Strait regions of BC. The results suggest possible coastal flooding occurring simultaneously along a wider coastal reach within the two regions, while the observed weak dependencies between the two regions suggest rare possibilities of simultaneous flooding occurring along the entire BC coastal reach from a single sea surge event. Knowledge on such spatial dependencies help determine to what extent different stations could be used to describe extreme sea surge conditions and facilitate reallocation of monitoring sites to improve spatial representation of sea-level measurements in the region.

Apparently, serious coastal flooding in BC is governed by extreme astronomical tide levels and, not by extreme surges. It is therefore highly likely that most of the exceptionally high historical surges in coastal BC have escaped unnoticed due to their occurrence at lower tide levels. Nevertheless, it is argued that the extreme surge coincidence with low to mid tides in coastal BC may be
partially due to chance. This diminishes the risk of getting very high sea levels as a result of extreme sea surges. However, the fact that extreme sea surges do still, infrequently, occur in conjunction with high tides means simply that the risk has not been eliminated completely. Thus the threat of extreme flooding due to such events is ever present. For instance if the greatest storm-surge measured at station 9850 (124 cm) had occurred at the time of the stronger astronomical tide of 12/24 event (741 cm), coastal flooding could have been as much as 60-cm higher than any recorded level in northern BC. Similarly if the greatest storm-surge measured at station 7795 (104 cm) had occurred at the time of the stronger astronomical tide of 12/16 event (470 cm), coastal flooding could have been as much as 14-cm higher than any recorded level in southern BC. Thus the risk of occurrence of such random events in the near future should be of concern.

References


Table 4-1: Pacific Region tide gauge stations adjacent to the coast of BC with at least 20 years of data.

<table>
<thead>
<tr>
<th>No</th>
<th>Station Number</th>
<th>Station Name</th>
<th>Station Co-ordinates</th>
<th>Duration (yrs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>7120</td>
<td>Victoria Harbour</td>
<td>48.42 123.37</td>
<td>68</td>
</tr>
<tr>
<td>2</td>
<td>7277</td>
<td>Patricia Bay</td>
<td>48.65 123.45</td>
<td>25</td>
</tr>
<tr>
<td>3</td>
<td>7735</td>
<td>Vancouver</td>
<td>49.29 123.11</td>
<td>51</td>
</tr>
<tr>
<td>4</td>
<td>7795</td>
<td>Point Atkinson</td>
<td>49.34 123.25</td>
<td>52</td>
</tr>
<tr>
<td>5</td>
<td>8074</td>
<td>Campbell River</td>
<td>50.04 125.25</td>
<td>33</td>
</tr>
<tr>
<td>6</td>
<td>8408</td>
<td>Port Hardy</td>
<td>50.72 127.49</td>
<td>39</td>
</tr>
<tr>
<td>7</td>
<td>8545</td>
<td>Bamfield</td>
<td>48.84 125.14</td>
<td>36</td>
</tr>
<tr>
<td>8</td>
<td>8615</td>
<td>Tofino</td>
<td>49.15 125.91</td>
<td>54</td>
</tr>
<tr>
<td>9</td>
<td>8976</td>
<td>Bella Bella</td>
<td>52.16 128.14</td>
<td>46</td>
</tr>
<tr>
<td>10</td>
<td>9354</td>
<td>Prince Rupert,</td>
<td>54.32 130.32</td>
<td>57</td>
</tr>
<tr>
<td>11</td>
<td>9850</td>
<td>Queen Charlotte City</td>
<td>53.25 132.07</td>
<td>39</td>
</tr>
</tbody>
</table>
Table 4-2: Application of the redundancy test on annual maximum sea-surge data at station 7120.

<table>
<thead>
<tr>
<th>Model</th>
<th>Parameter</th>
<th>Negative log-likelihood</th>
<th>Significance compared with the Chi-Square Value for 0.05 significance level</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>BASE</td>
<td>214.8</td>
<td></td>
</tr>
<tr>
<td>2.1</td>
<td>BASE + CV (µ 1 = MEI)</td>
<td>211.6</td>
<td>6.4 &gt;3.84</td>
</tr>
<tr>
<td>2.2</td>
<td>BASE + CV (µ 2 = NOI)</td>
<td>210.8</td>
<td>8 &gt;3.84</td>
</tr>
<tr>
<td>2.3</td>
<td>BASE + CV (µ 3 = PNA)</td>
<td>211.2</td>
<td>7.2 &gt;3.84</td>
</tr>
<tr>
<td>3.1</td>
<td>BASE + CV (µ 1 = NOI ; µ 2 = PNA)</td>
<td>208.7</td>
<td>4.2 &gt;3.84</td>
</tr>
<tr>
<td>3.2</td>
<td>BASE + CV (µ 1 = NOI ; µ 2 = MEI)</td>
<td>210.7</td>
<td>&lt;3.84</td>
</tr>
<tr>
<td>3.3</td>
<td>BASE + CV (µ 1 = PNA ; µ 2 = MEI)</td>
<td>209.8</td>
<td>&lt;3.84</td>
</tr>
<tr>
<td>4.1</td>
<td>BASE + CV (µ 1 = NOI ; µ 2 = PNA ; µ 3 = MEI)</td>
<td>208.7</td>
<td>0 &lt;3.84</td>
</tr>
</tbody>
</table>

SELECTED FINAL MODEL

BASE + CV (µ 1 = NOI ; µ 2 = PNA)
Table 4-3: Average climate indices for “strong El Niño”, “strong La Niña” and “neutral” years. The definitions of strong El Niño / La Niña years are based on the Environment Canada (Meteorological Services of Canada-The Green Lane) classification scheme.

<table>
<thead>
<tr>
<th>MAXSL (Chart cm)</th>
<th>PDO</th>
<th>MEI</th>
<th>NOI</th>
<th>PNA</th>
<th>ALPI</th>
</tr>
</thead>
<tbody>
<tr>
<td>strong El Niño (Warm ENSO) years: 1957, 1982, 1991, 1992, 1998</td>
<td>536</td>
<td>0.219</td>
<td>0.928</td>
<td>-1.486</td>
<td>0.105</td>
</tr>
<tr>
<td>strong La Niña (Cold ENSO) years: 1973, 1974, 1988, 1989, 1999</td>
<td>521</td>
<td>-0.370</td>
<td>-0.688</td>
<td>1.185</td>
<td>-0.081</td>
</tr>
</tbody>
</table>
Table 4-4: Overall averages of the (a) monthly maximum sea-surges and (b) monthly mean sea-surges recorded to-date at the 11-tide gauge stations in coastal BC, indicating a seasonal preference (October to March).

<table>
<thead>
<tr>
<th>Month</th>
<th>Overall Average Monthly Extreme Sea Surges (cm)</th>
<th>Overall Average Monthly Mean Positive Sea Surges (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>STDV</td>
</tr>
<tr>
<td>JAN</td>
<td>88.7</td>
<td>11.4</td>
</tr>
<tr>
<td>FEB</td>
<td>82.5</td>
<td>10.7</td>
</tr>
<tr>
<td>MAR</td>
<td>76.7</td>
<td>6.4</td>
</tr>
<tr>
<td>APR</td>
<td>63.4</td>
<td>9.1</td>
</tr>
<tr>
<td>MAY</td>
<td>51.2</td>
<td>10.2</td>
</tr>
<tr>
<td>JUN</td>
<td>52.6</td>
<td>14.6</td>
</tr>
<tr>
<td>JUL</td>
<td>50.1</td>
<td>13.6</td>
</tr>
<tr>
<td>AUG</td>
<td>43.8</td>
<td>10.1</td>
</tr>
<tr>
<td>SEPT</td>
<td>50.9</td>
<td>9.3</td>
</tr>
<tr>
<td>OCT</td>
<td>73.3</td>
<td>14.0</td>
</tr>
<tr>
<td>NOV</td>
<td>86.5</td>
<td>12.6</td>
</tr>
<tr>
<td>DEC</td>
<td>95.1</td>
<td>16.7</td>
</tr>
</tbody>
</table>
Table 4-5: Pearson product-moment correlation coefficients of annual maximum sea surges from 1950-2007 indicating two distinctly different spatial dependencies (demarcated by dashed rectangles) among stations in coastal BC.

<table>
<thead>
<tr>
<th>Station</th>
<th>7120</th>
<th>7277</th>
<th>7735</th>
<th>7795</th>
<th>8074</th>
<th>8408</th>
<th>8545</th>
<th>8615</th>
<th>8976</th>
<th>9354</th>
<th>9850</th>
</tr>
</thead>
<tbody>
<tr>
<td>7120</td>
<td>1.0</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.7**</td>
<td>0.6**</td>
<td>0.3</td>
<td>0.3*</td>
</tr>
<tr>
<td>7277</td>
<td>0.9**</td>
<td>1.0</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>0.6**</td>
<td>0.4*</td>
<td>0.4*</td>
</tr>
<tr>
<td>7735</td>
<td>0.9**</td>
<td>0.9**</td>
<td>1.0</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.7**</td>
<td>0.5**</td>
<td>0.3*</td>
<td>0.3*</td>
</tr>
<tr>
<td>7795</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>1.0</td>
<td>0.9**</td>
<td>0.7**</td>
<td>0.9**</td>
<td>0.6**</td>
<td>0.6**</td>
<td>0.3*</td>
<td>0.4*</td>
</tr>
<tr>
<td>8074</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>1.0</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>0.7**</td>
<td>0.5**</td>
<td>0.4*</td>
</tr>
<tr>
<td>8408</td>
<td>0.8**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>0.7**</td>
<td>0.8**</td>
<td>1.0</td>
<td>0.8**</td>
<td>0.7**</td>
<td>0.9**</td>
<td>0.5**</td>
<td>0.6**</td>
</tr>
<tr>
<td>8545</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.9**</td>
<td>0.8**</td>
<td>1.0</td>
<td>0.9**</td>
<td>0.6**</td>
<td>0.5**</td>
<td>0.4*</td>
</tr>
<tr>
<td>8615</td>
<td>0.7**</td>
<td>0.8**</td>
<td>0.7**</td>
<td>0.6**</td>
<td>0.8**</td>
<td>0.7**</td>
<td>0.9**</td>
<td>1.0</td>
<td>0.6**</td>
<td>0.3*</td>
<td>0.4*</td>
</tr>
<tr>
<td>8976</td>
<td>0.6**</td>
<td>0.6**</td>
<td>0.5**</td>
<td>0.6**</td>
<td>0.7**</td>
<td>0.9**</td>
<td>0.6**</td>
<td>0.6**</td>
<td>1.0</td>
<td>0.6**</td>
<td>0.7**</td>
</tr>
<tr>
<td>9354</td>
<td>0.3</td>
<td>0.4*</td>
<td>0.3*</td>
<td>0.3*</td>
<td>0.462**</td>
<td>0.5**</td>
<td>0.5**</td>
<td>0.3*</td>
<td>0.6**</td>
<td>1.0</td>
<td>0.7**</td>
</tr>
<tr>
<td>9850</td>
<td>0.3*</td>
<td>0.4*</td>
<td>0.3*</td>
<td>0.4*</td>
<td>0.391*</td>
<td>0.6**</td>
<td>0.4*</td>
<td>0.4*</td>
<td>0.7**</td>
<td>0.7**</td>
<td>1.0</td>
</tr>
</tbody>
</table>

Note: **. Correlation is significant at the 0.01 level
* Correlation is significant at the 0.05 level.
Table 4-6: Projected GEV coefficients for the base model at each station.

<table>
<thead>
<tr>
<th>Station</th>
<th>Data Coverage (Yr)</th>
<th>Location Parameter (µ)</th>
<th>Scale Parameter (σ)</th>
<th>Shape parameter (ξ)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Mean</td>
<td>Std. Error</td>
<td>Mean</td>
</tr>
<tr>
<td>7120</td>
<td>68</td>
<td>56.01</td>
<td>1.49</td>
<td>10.99</td>
</tr>
<tr>
<td>7277</td>
<td>25</td>
<td>60.12</td>
<td>2.73</td>
<td>12.53</td>
</tr>
<tr>
<td>7735</td>
<td>51</td>
<td>66.5</td>
<td>1.97</td>
<td>12.69</td>
</tr>
<tr>
<td>7795</td>
<td>52</td>
<td>67.3</td>
<td>1.86</td>
<td>12.42</td>
</tr>
<tr>
<td>8074</td>
<td>33</td>
<td>71.5</td>
<td>2.73</td>
<td>14.18</td>
</tr>
<tr>
<td>8408</td>
<td>39</td>
<td>54.8</td>
<td>2.12</td>
<td>11.99</td>
</tr>
<tr>
<td>8545</td>
<td>36</td>
<td>57.9</td>
<td>2.80</td>
<td>14.57</td>
</tr>
<tr>
<td>8615</td>
<td>54</td>
<td>61.6</td>
<td>2.19</td>
<td>14.71</td>
</tr>
<tr>
<td>8976</td>
<td>46</td>
<td>59.6</td>
<td>2.00</td>
<td>14.47</td>
</tr>
<tr>
<td>9354</td>
<td>57</td>
<td>64.2</td>
<td>1.40</td>
<td>10.27</td>
</tr>
<tr>
<td>9850</td>
<td>39</td>
<td>69.7</td>
<td>2.47</td>
<td>13.29</td>
</tr>
</tbody>
</table>
Table 4-7: Extreme Residual recurrences in coastal BC with no climate considerations (Base Model).

<table>
<thead>
<tr>
<th>Station</th>
<th>Data Coverage (Yr)</th>
<th>2</th>
<th>5</th>
<th>10</th>
<th>20</th>
<th>25</th>
<th>50</th>
<th>100</th>
</tr>
</thead>
<tbody>
<tr>
<td>7120</td>
<td>68</td>
<td>59.8</td>
<td>69.5</td>
<td>74.5</td>
<td>78.3</td>
<td>79.4</td>
<td>82.3</td>
<td>84.7</td>
</tr>
<tr>
<td>7277</td>
<td>25</td>
<td>64.3</td>
<td>72.8</td>
<td>76.0</td>
<td>78.1</td>
<td>78.6</td>
<td>79.8</td>
<td>80.6</td>
</tr>
<tr>
<td>7735</td>
<td>51</td>
<td>70.9</td>
<td>81.6</td>
<td>86.7</td>
<td>90.6</td>
<td>91.7</td>
<td>94.5</td>
<td>96.7</td>
</tr>
<tr>
<td>7795</td>
<td>52</td>
<td>71.7</td>
<td>82.6</td>
<td>88.2</td>
<td>92.6</td>
<td>93.8</td>
<td>97.1</td>
<td>99.8</td>
</tr>
<tr>
<td>8074</td>
<td>33</td>
<td>76.3</td>
<td>87.8</td>
<td>93.1</td>
<td>97.0</td>
<td>98.0</td>
<td>100.7</td>
<td>102.7</td>
</tr>
<tr>
<td>8408</td>
<td>39</td>
<td>58.8</td>
<td>67.2</td>
<td>70.5</td>
<td>72.6</td>
<td>73.2</td>
<td>74.4</td>
<td>75.3</td>
</tr>
<tr>
<td>8545</td>
<td>36</td>
<td>62.9</td>
<td>75.0</td>
<td>80.8</td>
<td>85.1</td>
<td>86.2</td>
<td>89.3</td>
<td>91.6</td>
</tr>
<tr>
<td>8615</td>
<td>54</td>
<td>66.9</td>
<td>82.2</td>
<td>91.4</td>
<td>99.6</td>
<td>102.1</td>
<td>109.4</td>
<td>116.3</td>
</tr>
<tr>
<td>8976</td>
<td>46</td>
<td>65.9</td>
<td>77.0</td>
<td>82.6</td>
<td>87.1</td>
<td>88.3</td>
<td>91.7</td>
<td>94.4</td>
</tr>
<tr>
<td>9354</td>
<td>57</td>
<td>67.8</td>
<td>78.1</td>
<td>84.1</td>
<td>89.3</td>
<td>90.9</td>
<td>95.4</td>
<td>99.4</td>
</tr>
<tr>
<td>9850</td>
<td>39</td>
<td>74.9</td>
<td>90.0</td>
<td>99.7</td>
<td>108.8</td>
<td>111.7</td>
<td>120.4</td>
<td>129.0</td>
</tr>
</tbody>
</table>
Table 4-8: Climate covariate coefficients expressed to the first decimal when applied as individual covariates. The positive/negative (+/-) sign in each cell indicate the relationship (i.e. trend) between the model parameter and the climate covariate. The shaded cells indicate the coefficients that are significant at 95% level).

<table>
<thead>
<tr>
<th>Stn.</th>
<th>Data yrs. (Yr)</th>
<th>Location Parameter (μ)</th>
<th>Scale Parameter (σ)</th>
<th>Shape Parameter (ξ)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Linear Trend</td>
<td>ALPI</td>
<td>MEI</td>
<td>NOI</td>
</tr>
<tr>
<td>7120</td>
<td>68</td>
<td>0.1</td>
<td>0.7</td>
<td>4.8</td>
</tr>
<tr>
<td>7277</td>
<td>25</td>
<td>0.3</td>
<td>1.8</td>
<td>1.5</td>
</tr>
<tr>
<td>7735</td>
<td>51</td>
<td>0.2</td>
<td>1.4</td>
<td>2.5</td>
</tr>
<tr>
<td>7795</td>
<td>52</td>
<td>-0.1</td>
<td>0.3</td>
<td>2.5</td>
</tr>
<tr>
<td>8074</td>
<td>33</td>
<td>-0.1</td>
<td>1.0</td>
<td>4.2</td>
</tr>
<tr>
<td>8408</td>
<td>39</td>
<td>-0.2</td>
<td>0.6</td>
<td>1.4</td>
</tr>
<tr>
<td>8545</td>
<td>36</td>
<td>0.2</td>
<td>1.9</td>
<td>7.9</td>
</tr>
<tr>
<td>8615</td>
<td>54</td>
<td>-0.2</td>
<td>1.2</td>
<td>4.3</td>
</tr>
<tr>
<td>8976</td>
<td>46</td>
<td>-0.2</td>
<td>1.2</td>
<td>3.5</td>
</tr>
<tr>
<td>9354</td>
<td>57</td>
<td>0.2</td>
<td>1.5</td>
<td>4.0</td>
</tr>
<tr>
<td>9850</td>
<td>39</td>
<td>-0.0</td>
<td>0.8</td>
<td>3.6</td>
</tr>
</tbody>
</table>
Table 4-9: Climate covariates coefficients for the final models at each tide-gauge station, selected from the redundancy test.

<table>
<thead>
<tr>
<th>Stn.</th>
<th>Location Parameter ( (\mu) )</th>
<th>Scale Parameter ( (\sigma) )</th>
<th>Shape Parameter ( (\xi) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>7120</td>
<td>( \mu = 57.5 - 2.28(\text{NOI}) + 7.04(\text{PNA}) )</td>
<td>( \sigma = 10.5 )</td>
<td>( \xi = -0.368 )</td>
</tr>
<tr>
<td>7277</td>
<td>( \mu = 60.1 )</td>
<td>( \sigma = 12.5 )</td>
<td>( \xi = -0.568 )</td>
</tr>
<tr>
<td>7735</td>
<td>( \mu = 68.6 + 1.37(\text{ALPI}) )</td>
<td>( \sigma = 11.8 )</td>
<td>( \xi = -0.265 )</td>
</tr>
<tr>
<td>7795</td>
<td>( \mu = 68.5 )</td>
<td>( \sigma = 13.1 )</td>
<td>( \xi = -0.511 + 0.331(\text{MEI}) )</td>
</tr>
<tr>
<td>8074</td>
<td>( \mu = 71.5 - 2.8(\text{NOI}) )</td>
<td>( \sigma = 13.5 )</td>
<td>( \xi = -0.421 )</td>
</tr>
<tr>
<td>8408</td>
<td>( \mu = 57.8 + 5.9(\text{PNA}) )</td>
<td>( \sigma = 11.5 )</td>
<td>( \xi = -0.555 )</td>
</tr>
<tr>
<td>8545</td>
<td>( \mu = 59.1 - 4.3(\text{NOI}) )</td>
<td>( \sigma = 13.1 )</td>
<td>( \xi = -0.441 )</td>
</tr>
<tr>
<td>8615</td>
<td>( \mu = 62.4 )</td>
<td>( \sigma = 13.4 )</td>
<td>( \xi = -0.128 + 1.94(\text{PNA}) )</td>
</tr>
<tr>
<td>8976</td>
<td>( \mu = 59.4 + 1.2(\text{ALPI}) - 2.1(\text{NOI}) )</td>
<td>( \sigma = 11.5 )</td>
<td>( \xi = -0.272 - 0.109(\text{ALPI}) )</td>
</tr>
<tr>
<td>9354</td>
<td>( \mu = 65.3 + 1.0(\text{ALPI}) + 2.9(\text{MEI}) )</td>
<td>( \sigma = 9.3 )</td>
<td>( \xi = -0.114 )</td>
</tr>
<tr>
<td>9850</td>
<td>( \mu = 69.7 )</td>
<td>( \sigma = 13.3 )</td>
<td>( \xi = -0.017 )</td>
</tr>
</tbody>
</table>
**Table 4-10**: Station specific extreme sea surge occurrences having 1% exceedance in a given year with climate considerations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Extreme sea-surge projections with 1% exceedance probability in each year (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Base</td>
</tr>
<tr>
<td>7120</td>
<td>84.9</td>
</tr>
<tr>
<td>7735</td>
<td>96.7</td>
</tr>
<tr>
<td>7795</td>
<td>99.8</td>
</tr>
<tr>
<td>8074</td>
<td>102.7</td>
</tr>
<tr>
<td>8408</td>
<td>75.3</td>
</tr>
<tr>
<td>8545</td>
<td>95.1</td>
</tr>
<tr>
<td>8615</td>
<td>116.3</td>
</tr>
<tr>
<td>8976</td>
<td>94.4</td>
</tr>
<tr>
<td>9354</td>
<td>100.6</td>
</tr>
</tbody>
</table>
Table 4-11: The sea-surge characteristics and the projected annual percentage exceedance (highlighted columns) observed at each tide gauge station at the time of the strongest ever extreme sea surge event occurrence in southern (16th December 1982) and northern (24th December 2003) BC coasts. Results support the conclusion of weak correlation between the two regions.

<table>
<thead>
<tr>
<th>Station</th>
<th>16th December 1982 event</th>
<th>24th December 2003 event</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TWL (m)</td>
<td>Tide (m)</td>
</tr>
<tr>
<td>7120</td>
<td>3.2</td>
<td>2.3</td>
</tr>
<tr>
<td>7735</td>
<td>5.6</td>
<td>4.7</td>
</tr>
<tr>
<td>7795</td>
<td>5.6</td>
<td>4.6</td>
</tr>
<tr>
<td>8074</td>
<td>5.2</td>
<td>4.3</td>
</tr>
<tr>
<td>8408</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>8545</td>
<td>3.4</td>
<td>2.6</td>
</tr>
<tr>
<td>8615</td>
<td>3.3</td>
<td>2.5</td>
</tr>
<tr>
<td>8976</td>
<td>4.4</td>
<td>3.9</td>
</tr>
<tr>
<td>9354</td>
<td>6.2</td>
<td>5.7</td>
</tr>
<tr>
<td>9850</td>
<td>6.5</td>
<td>6.0</td>
</tr>
</tbody>
</table>
Table 4-12: The sea-surge count distribution vs. the percentile astronomical tide levels for the 50-highest surges on record at the 11-tide gauges. Results show significant number of extreme sea-surges coinciding with mid to low tides in coastal BC.

<table>
<thead>
<tr>
<th>Tide-gauge Stn.</th>
<th>50-th &lt;AT</th>
<th>95-th &gt;AT&gt;90-th</th>
<th>90-th &gt;AT&gt;75-th</th>
<th>75-th &gt;AT&gt;50-th</th>
<th>50-th &gt;AT&gt;10-th</th>
<th>10-th &gt;AT&gt;5-th</th>
<th>5-th &gt;AT</th>
</tr>
</thead>
<tbody>
<tr>
<td>7120</td>
<td>2</td>
<td>1</td>
<td>3</td>
<td>15</td>
<td>21</td>
<td>4</td>
<td>4</td>
</tr>
<tr>
<td>7277</td>
<td>3</td>
<td>4</td>
<td>1</td>
<td>18</td>
<td>19</td>
<td>4</td>
<td>1</td>
</tr>
<tr>
<td>7735</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>10</td>
<td>11</td>
<td>3</td>
<td>23</td>
</tr>
<tr>
<td>7795</td>
<td>3</td>
<td>1</td>
<td>9</td>
<td>16</td>
<td>18</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>8074</td>
<td>6</td>
<td>4</td>
<td>4</td>
<td>6</td>
<td>26</td>
<td>4</td>
<td>0</td>
</tr>
<tr>
<td>8408</td>
<td>5</td>
<td>3</td>
<td>6</td>
<td>10</td>
<td>18</td>
<td>5</td>
<td>3</td>
</tr>
<tr>
<td>8545</td>
<td>2</td>
<td>1</td>
<td>13</td>
<td>12</td>
<td>20</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>8615</td>
<td>2</td>
<td>1</td>
<td>6</td>
<td>12</td>
<td>24</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>8976</td>
<td>2</td>
<td>3</td>
<td>11</td>
<td>11</td>
<td>18</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>9354</td>
<td>2</td>
<td>2</td>
<td>6</td>
<td>21</td>
<td>14</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>9850</td>
<td>0</td>
<td>2</td>
<td>9</td>
<td>12</td>
<td>24</td>
<td>0</td>
<td>3</td>
</tr>
<tr>
<td><strong>Average Events</strong></td>
<td><strong>3</strong></td>
<td><strong>2</strong></td>
<td><strong>6</strong></td>
<td><strong>13</strong></td>
<td><strong>19</strong></td>
<td><strong>3</strong></td>
<td><strong>4</strong></td>
</tr>
</tbody>
</table>

% with respect to the total events considered (50)

|                  | 5 | 4 | 12 | 26 | 38 | 6 | 8 |
Table 4-13: The surge-tide levels of the highest sea surge and highest total water level event on record at each tide gauge station in coastal BC. Note that the tide levels are indicated both in terms of absolute values and as a percentile with respect to the 2004 astronomical tide levels.

<table>
<thead>
<tr>
<th>Tide-gauge Station</th>
<th>Surge-Tide levels of the Highest Surge events</th>
<th>Surge-Tide levels of the Highest Total Water-level event</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Surge (cm)</td>
<td>AT (cm)</td>
</tr>
<tr>
<td>7120</td>
<td>85</td>
<td>231</td>
</tr>
<tr>
<td>7277</td>
<td>84</td>
<td>240</td>
</tr>
<tr>
<td>7735</td>
<td>124</td>
<td>91</td>
</tr>
<tr>
<td>7795</td>
<td>104</td>
<td>336</td>
</tr>
<tr>
<td>8074</td>
<td>99</td>
<td>145</td>
</tr>
<tr>
<td>8408</td>
<td>75</td>
<td>109</td>
</tr>
<tr>
<td>8545</td>
<td>94</td>
<td>222</td>
</tr>
<tr>
<td>8615</td>
<td>128</td>
<td>191</td>
</tr>
<tr>
<td>8976</td>
<td>97</td>
<td>207</td>
</tr>
<tr>
<td>9354</td>
<td>98</td>
<td>343</td>
</tr>
<tr>
<td>9850</td>
<td>124</td>
<td>272</td>
</tr>
<tr>
<td>Average</td>
<td>101</td>
<td>217</td>
</tr>
</tbody>
</table>
Figure 4-1: The geographical location of coastal of BC, including the Haida Gawii in northern BC and Vancouver Island in southern BC. Also shown are the Pacific region tide gauge stations and the location of Hecate Strait and the Georgia Strait.
Figure 4-2: Temporal distribution of NOI, PDO and ENSO indicating significant colinearities. Note the 1976 major climate regime shift being captured by all climate indices approximately at the same time.
Figure 4-3: The GEV distribution without climate considerations, 5% and 95% confidence bands together with the annual maxima for each station. The annual maxima fall almost on the GEV curve and are well within the quantile limits.
Figure 4-4: Estimated return levels for residual water levels (sea surges) under Warm ENSO (red-dashed line), Neutral (blue-dashed line) and Cold ENSO (green-dashed line) conditions, from having fit Maximum Annual Residual Water Levels at Station 7120 to a GEV distribution with climate variability effects accounted as covariates. Results for no climate considerations (black-continuous line) and the 95% confidence limits (blue-continuous line) are included for comparison purposes. Similar probability curves have been constructed for each tidal station of the BC coast, (results not shown).
Figure 4-5: Surge-Tide relationship and the development of the extreme surge event of 24th December 2003 (top panel). Impacts of the storm surge event on the eastern Graham Island, BC are shown (bottom panel). Approximately 2.5 m of shoreline was lost at this location along Highway 16 (upper) compromising the road shoulder and bed. Extensive coastal flooding also occurred, damaging buildings and sending tonnes of drift logs onto nearby roads and properties. (Photos courtesy of Mavis Mark)
Figure 4-6: Surge-Tide relationship and the development of the extreme surge event of 16th December 1982 (top panel). Impacts of the storm surge event in southern BC are shown (bottom panels). Extensive coastal flooding occurred in Boundary Bay, Mud Bay and Westham Island and resulted in highest ever occurred water levels in southern BC. (Photos courtesy of Fraser Delta Engineering Department of the BC MOE)
5.0 Extreme wind regime responses to climate variability and change in the inner-south-coast of British Columbia.

5.1 Abstract

The long-term hourly wind speed data maintained by the Meteorological Service of Canada are used to evaluate the possible influence of climate variability on extreme wind response in the inner-south-coast of BC (Lat 48-49° N, Long 123° W). Preliminary results suggest significantly different extreme wind responses to warm and cold ENSO modes, with a tendency for high extreme winds to occur during negative (i.e. cold) ENSO phase. The methodology is primarily based on approximating a Generalized Pareto distribution (GPD) to extreme winds in the presence of climate variability covariates. This study demonstrates how the information on climate variability can be of significant use and value in understanding and improving the wind forecasting skills in a region.

5.2 Introduction

Climate variability such as the ENSO cycles has been shown to have significant impacts on various atmospheric and oceanic parameters and phenomena over global and regional scales. The present study is motivated by a desire to better understand the relationship between various climate excursions (i.e. ENSO) and extreme environmental variables in the Pacific Northwest. Here the focus is on extreme winds; more specifically, on its relationship to various climate variability cycles.
From an impact point of view, it is the extreme windstorms and associated storm surges that pose the greatest natural threat to coastal communities, particularly in low lying deltaic regions. Such events have been named as the world’s foremost natural hazard in terms of property damages and lives lost (Murty, 1988). Potential consequences of windstorms on coastal systems increase when persisting storm patterns last for more than decades in response to climate variability and change. In addition a moderate windstorm can also become hazardous as a result of its occurrence on higher than average sea levels due to thermostatic expansion of sea-water during warm ENSO episodes. Such conditions could occur as a result of systematic variations in the global and regional atmospheric and oceanic characteristics that could last from a year-or-two to multiple decades (Cavazos and Rivas, 2004; Bromirski et al., 2005).

This work is an attempt to investigate the relationship between extreme winds (i.e. as an analog to extreme storminess) and regional and large-scale climate variability signals in the Boundary Bay region of the Inner-south-coast of BC (Figure 5.1). The analysis is based on observed independent extreme wind events at three meteorological stations located in the vicinity of the Boundary Bay region (Figure. 5.1). The extreme wind records above a pre-defined threshold \( u \) are fitted to a Generalized Pareto (GPD) distribution with climate variability as parameter covariates. Using this, extreme wind event recurrences and their relationship with ENSO states are investigated. One major objective of this work is to test the validity of the common notion of ENSO related storminess trends in the Pacific Northwest defined as, “a general increase in storminess during warm
ENSO phase.” Even though this has been proven true for the southern California region (Graham and Diaz, 2001; Chang et al., 2002; Cavazos and Rivas, 2004; Bromirski et al., 2005), comparable studies have not been completed in northern latitudes, along Canada’s coastal regions.

5.3 Problem Definition

Extreme events occurring in conjunction with rapid relative sea-level rise are causing significant infrastructure damage in the Boundary Bay region of southern BC, Canada (Figure 5.1). This damage has raised concerns about the adequacy of the existing dyke system to continue to withstand extreme event impacts resulting from climate variability and change. Understanding the local-scale extreme wind responses to climate variability and change is key towards a realistic assessment of future storm damage in the Boundary bay region. Even though previous much research has examined the influence of climate variability on extreme windstorm properties in many parts of the world (e.g. Leathers and Palecki, 1992; Roger, 1997; Pirazzoli, 2000; Gupta et al., 2003; Storlazzi et al., 2003) very little comparable research has been completed in southern BC. One reason may be the relatively poor network density of directional wind measurements over this coastal region (Barton, 2005). The current work attempts to apply the statistics of extremes to better understand the response (i.e. sensitivity) of extreme windstorms in this region to Climate Variability (CV) and Change (CC) (Figure 5.1).
5.4  Data.

5.4.1 Directional wind records

In order to establish accurate extreme wind recurrences, meteorological records extending over 50 years or more are recommended (Coles and Dixon, 1999; Coles, 2001). In this instance, the longest available record of about 50 years (1953-2006) of hourly wind measurements from Vancouver International Airport (YVR: CS1108447) (Figure 5.1).

Observation based studies limited to a single station do not however provide a complete description of the changing wind response to long-term climate variability patterns, as it is rare that a single site would sample the full range of weather patterns. Therefore, in addition to YVR record, two stations in close proximity to YVR having shorter records (1993-2006) were chosen for spatial validation (Sand Heads (CS 1107010) and Saturna Island (CS 1017101), (Figure 5.1). The hourly directional wind data records were obtained from the Meteorological services of Canada digital archive.

Compared to Saturna, the YVR and Sand Heads stations are less exposed to certain wind directions (i.e. weak Northerly exposure due to higher land to the North). Despite this, these records can still be used to assess the general trends and characteristics of the wind regime (Simmons, 1975). Although, the accuracy of the wind instrument, poor instrument exposure and incomplete understanding of the effects of changes in the instrument's surroundings are key factors that result in erroneous analysis. Data errors have not previously been detected in YVR (e.g., Tuller, 2004; Barton, 2005)
Lange (1998) shows that the general wind pattern across the Inner-south-coast of BC is characterised by significant wind energy fluctuations from diurnal, synoptic (i.e. days) seasonal to annual, with dominant winds blowing from Juan de Fuca Strait-Puget Sound along a southerly direction. Westerly winds from the Strait of Georgia are the strongest and most dangerous over some parts of southern BC close to Vancouver. Depending on the aspect of the Boundary Bay coastline, certain segments are exposed to winds blowing from northerly, westerly and southerly directions. Consequently extreme winds from these sectors become potentially damaging to the coastal margin. Following the sector classification described by Lange (1998), three dominant wind sectors were chosen for the study: northerly (N), westerly (W) and southerly (S) (Table 5.1). Note that each wind sector overlaps to a certain degree with adjacent sectors, and that this is likely to account for the uncertainty resulting from the wind dispersion at the boundaries.

5.4.2 Climate Variability Indices.

Observational evidence supports the existence of distinct climate regimes in the NE Pacific that develop in response to natural climate variability ranging in periodicity from several years to decades (Trenberth and Hurrell, 1995; Crawford et al., 1999; Allan and Komar, 2002). Such variations are attributed to global and regional oceanic and atmospheric circulation patterns represented by distinct climate variability indices such as the Multivariate ENSO Index (MEI), Pacific Decadal Oscillation (PDO), Northern Oscillation Index (NOI), Aleutian Low Pressure Index (ALPI) and Pacific North American teleconnection (PNA) Index.
Some climate variability patterns such as the ENSO and PDO have been directly linked to both average and extreme directional wind regime variations in the Pacific (Trenberth and Hurrell, 1995; Crawford et al., 1999; Allan and Komar, 2002), suggesting that shifts in the frequency and the strength of extreme winds in the study region may occur in conjunction with climate variability episodes.

Several published climatic variability indices are used in this study. The Aleutian Low Pressure Index (ALPI) is a North Pacific regional index which indicates the relative intensity of the controlling pressure system in the NE Pacific and, therefore, indirectly describes the relative strength of regional wind patterns. The index is calculated as the mean area (km$^{-2}$) with sea-level pressure less than or equal to 100.5 kPa expressed as an anomaly from the 1950-1997 mean (Beamish et al., 1997). Positive ALPI index values reflect a relatively strong or intense Aleutian Low, while negative values indicate the opposite. ALPI values are published online by the Canadian Department of Fisheries and Oceans <www.pac.dfo-mpo.gc.ca/sci/sa-mfpd/downloads/alpi.txt>.

The classification of ENSO events used in this study are defined by the National Oceanic and Atmospheric Administration's (NOAA) Multivariate ENSO Index (MEI) (Wolter and Timlin, 1993, 1998). MEI is defined as the weighted average of a number of tropical Pacific environmental variables including sea surface temperature, east-west (i.e. zonal) and north-south (i.e. meridional) surface winds, sea-level pressure and temperature, and cloudiness. Negative MEI values represent the cold ENSO phase (La Niña), while positive MEI values represent the warm ENSO phase (El Niño). Monthly MEI values are published
The PDO index characterizes inter-annual variability in average North Pacific sea surface temperature and, as above, reflects NE Pacific regional climatic variability (Mantua et al., 1997). The PDO is well correlated with many records of North Pacific and Pacific Northwest climate and ecology, including sea-level pressure, winter land–surface temperature and precipitation, and stream flow (Mantua et al., 1997; Zhang et al., 1997; Hessl et al., 2004; Schneider and Cornuelle, 2005). Predominantly positive or negative sea–surface temperature anomalies along the Pacific coast of North America characterise the phase of the PDO pattern as being in a "warm" or "cold" phase respectively. Monthly PDO index values are published online at the Joint Institute for Study of the Atmosphere and Ocean’s PDO site <www.jisao.washington.edu/pdo> (Mantua et al., 1997).

The Northern Oscillation Index (NOI) represents the sea-level pressure anomaly between the North Pacific High (NPH, 35°N and 130°W) region of the NE Pacific and the equatorial Pacific near Darwin, Australia, a climatologically low-pressure region (Schwing et al., 2002). NOI is dominated by the interannual variations of ENSO, such that large positive (negative) NOI are associated with La Niña (El Niño) (Schwing et al., 2002). As NOI is partially based in the NE Pacific, it provides a more direct connection between various climate processes in the eastern north Pacific and remotely teleconnected global climate events.
The NOI time series is available online from NOAA’s Pacific Fisheries Environmental Laboratory. <http://www.pfeg.noaa.gov>.

The Pacific/North American teleconnection (PNA) describes one of the most dominant modes of low frequency variability in the extra-tropics during the North Hemisphere winter (Van den Dool et al., 2000). The PNA index is constructed through a linear combination of normalized 700-mb height anomalies located nearest to the nominal centers of the PNA pattern (Leathers and Palecki, 1992). The PNA pattern occurs as a result of an amplification or damping of the mean flow configuration over the unique geography of the Pacific Basin and North America (Leathers and Palecki, 1992). The PNA is strongly influenced by the mode of ENSO, where the positive phase of PNA tends to be associated with warm ENSO and the negative phase with cold ENSO.

5.4 Methodology

This study is based on extreme value theory, a widely applied tool for drawing inferences about the extremes of stochastic processes based upon independent extreme values (Coles and Dixon, 1999; Coles 2001; Butler et al., 2007; R Development Core Team, 2007). Two main approaches are presented in the literature: the Block Maxima models and the Peaks-over-Threshold models (POT). A summary of these techniques can be found in Palutikof et al. (1999). This work adopts the latter; the parametric POT models based on the Generalized Pareto Distribution (GPD) which is a statistical technique commonly applied for extreme wind analysis. One advantage of this model is that, it
properly accounts for the effects of natural climate variability in extremes (Pickands, 1975; Hosking and Wallis, 1987). The GPD makes use of all data over a chosen threshold; the obvious benefit being that this approach allows use of more than just annual maxima.

5.5.1 The extreme value approach

A GPD extreme wind analysis focuses on the high-end tail, by selecting those independent events above a high threshold value \( u \). The GPD has a cumulative distribution function (cdf) with the scale parameter \( \sigma \), the shape parameter \( \xi \), and the location parameter \( \mu \) is the chosen threshold (\( u \)) (5.1),

\[
F(x; \sigma, \xi) = 1 - \left(1 - \frac{\xi}{\sigma}(x-u)\right)^{\frac{1}{\xi}}
\]  

(5.1)

Where \( u \leq x \leq u + \frac{\sigma}{\xi} \) for \( \xi > 0 \), and \( u \leq x \leq \infty \) for \( \xi \leq 0 \).

For the special case where \( \xi = 0 \),

\[
F(x; \sigma) = 1 - \exp\left(-\frac{x-u}{\sigma}\right)
\]  

(5.2)

Where \( u \leq x \leq \infty \) for \( \xi \leq 0 \). The derivative of the cdf in (5.1) gives the probability density function (pdf):

\[
F(x; \sigma, \xi) = \frac{1}{\sigma}\left(1 - \frac{\xi}{\sigma}(x-u)\right)^{\frac{(1-\xi)}{\xi}}
\]  

(5.3)

Figure (5.2) demonstrates the probability density function and the histogram for extreme wind data at YVR by fitting a GPD model; the fitted density follows the shape of the data suggesting that the GPD model is a good choice for these data.
5.5.2 Selection of Thresholds.

The choice of the threshold \((u)\) is critical to the GPD analysis on POT selected data. Too high of a threshold \((u)\) could result in too few exceedances and high variance estimators. On the other hand, too small \(u\) values could provide biased estimators and the approximation to a GPD would not be feasible. In this study a graphical approach is applied to determine the ideal wind speed threshold for each station, where a number of GPD models are fitted to the data using a range of thresholds and then the parameter estimates are graphed along with their variability. The appropriate threshold is chosen as the one with lowest possible wind speed such that any higher threshold would result in similar parameter estimates.

This approach is developed based on a unique characteristic of the GPD, termed the “threshold stability property” that makes it suitable for extreme value analysis (Hosking and Wallis, 1987). Accordingly, if a random variable \(X\) is GPD distributed, then the conditional distribution of \((X-u)\) where \(X \geq u\) is also a GPD with the similar shape parameter \(\xi\). That is, for a simple wind climate described by a single GPD distribution, \(\xi\) should not change significantly with the appropriate choice of the threshold (Figure 5.3) demonstrates the selection of the threshold for YVR extreme winds using this approach. As shown, the lowest possible threshold wind speed, such that any higher threshold would result in similar parameter estimates, is 40 km/hr.

It should be emphasised that the choice of threshold is purely based on statistical practicality and not on any physically meaningful level. Once a statistically appropriate threshold is found, then the GPD can be fit, and
information involving any physically meaningful threshold can subsequently be
determined from this model. That is, once $u$ is chosen, it is then possible to
determine, for example, $Pr \{ X > x \mid x > u \}$, where $x$ is a physically important level.

5.5.3 Parameter estimation method

Different statistical methods are available for performing parameter
estimations in the GPD distribution including: Method of Moments Estimation,
Probability Weighted Moments or equivalent L-Moments, Maximum-Likelihood
Estimation, and Bayesian methods (Hosking and Wallis, 1987). In this study, the
Maximum-Likelihood Estimation method is applied exclusively because of its
ability to easily incorporate covariate information into the model parameter
estimates. Another advantage of the Maximum-Likelihood Estimation method is
that approximate standard errors for estimated parameters and return levels can
be automatically produced, either via the information matrix or through the profile
likelihood estimation method (Gilleland and Katz, 2005). Nevertheless, the
performance of the Maximum-Likelihood Estimation method is known to be
unstable and can give unrealistic estimates for the shape parameter when the
sample size is small (i.e. $n \leq 25$) (Hosking and Wallis, 1987; Coles and Dixon,
1999). However when sufficient data are available ($n > 50$), Maximum-Likelihood
Estimation is comparable in performance to more advanced estimation
techniques (Martins and Stedinger, 2000; Gilleland and Katz, 2005). Maximum
likelihood estimation is derived by letting $x$ be a continuous random variable with
the probability density function ($pdf$),
\[ f(x; \theta_1, \theta_2, ..., \theta_k) \quad (5.4) \]

Where \( \theta_1, \theta_2,..., \theta_k \) are \( k \) unknown constant parameters which need to be estimated via maximum likelihood estimation. Let \( x_1, x_2, ..., x_N \) be a random sample of \( N \) independent observations (i.e. a sample of extremes in the current analysis). Then, the likelihood function is given by the following product:

\[ L(x_1, x_2, ..., x_N | \theta_1, \theta_2, ..., \theta_k) = L = \prod_{i=1}^{N} f(x_i; \theta_1, \theta_2, ..., \theta_k) \quad i = 1, 2, ..., N \quad (5.5) \]

Once data have been collected and the likelihood function of a model given the data is determined, the parameter values of the desired probability distribution that maximize the likelihood function can be established. According to MLE principle, this is the population distribution that is most likely to have generated the observed data.

For computational convenience, the MLE estimate is obtained by maximizing the log-likelihood function (Eqn 5.6). This is because the two functions, \( \ln L (x | \theta) \) and \( L (x | \theta) \), are monotonically related to each other so that the same MLE estimate can be obtained by maximizing either one. The logarithmic likelihood function is given by:

\[ \Lambda = \ln L = \sum_{i=1}^{N} \ln f(x_i; \theta_1, \theta_2, ..., \theta_k) \quad (5.6) \]

Assuming that the log-likelihood function is differentiable, if \( \theta_{MLE} \) exists, it must satisfy the following partial differential equation (Eqn.5.7), because the definition of maximum or minimum of a continuous differentiable function implies that its first derivatives vanish at such points.
\[
\frac{\partial (\Lambda)}{\partial \theta_j} = 0 \quad j = 1,2,\ldots,k
\]  

(5.7)

However an additional condition should also be satisfied to ensure that \(\ln L(x|\theta)\) is a maximum and not a minimum, which the first derivative cannot reveal. To be a maximum, the shape of the log-likelihood function should be convex (i.e a peak when approaching the \(\theta_{MLE}\)). This is indicated by a negative value for the second derivative of the log-likelihood.

### 5.5.4 Return Levels (Quantiles)

A useful tool that is easily obtained from extreme value distributions are return levels (quantiles) for an extreme event. These values give the level of the variable of interest such that there is a probability of \(p\) that \(Z_p\) is exceeded in any given year. Alternatively, the same can be expressed in terms of the return period of an event, defined as the level that is expected to be exceeded on average once every \(1/p\) years (where \(1/p\) is referred to as the return period). For example, the 100-year extreme wind return level at a given location has a 1% (1/100=0.01) chance to occur in any given year.

By assuming that the extreme wind speed exceedance follows a GPD distribution, estimates of the GPD quantiles can be made with the definition of event crossing rate (\(\lambda\)). Where, \(\lambda\) is the expected number of peaks above the threshold \(u\) per year. Accordingly the expression for the number of exceedances in \(t\) years is,

\[
\lambda_s = \lambda_t (1 - F)
\]

(5.8)
Substituting for \((1-F)\) from (1) and setting \(\lambda_x\) to unity, and \(t\) to \(T\), that is, one exceedance in \(T\) years, solutions for the quantiles \(X_T\) can be established,

\[
X^{GPD}_T = u + \frac{\sigma}{\xi} \left[ 1 - \left( \frac{\lambda T}{\lambda T + 1} \right)^{-\frac{1}{\xi}} \right]
\]

(5.9)

Now the problem of estimating quantiles is reduced to estimating the GPD parameters \((\sigma \text{ and } \xi)\) for a given threshold \((u)\).

5.5.5 The question of statistical independence

In extreme value analysis, selected extremes must satisfy independence (i.e. Poisson distributed) (Harris, 1999). When several data points are considered from each year, as in the case of POT analysis, there may well be several clustered hourly maximum wind speeds above the threshold \(u\) from a single storm. Such events are unlikely to be statistically independent. An approach frequently employed to remove dependent events before proceeding with a statistical analysis is to decluster the data based on a minimum time separation between selected events and utilizing only a summary of each cluster (e.g., the maxima) (Cook, 1982; Gusella, 1991; Coles and Walshaw, 1994).

In this study the removal of dependent extreme events from the databases was carried out by following the method proposed in Cook (1982) for the identification of independent extreme storms from a continuous meteorological data set. Essentially this method is applied to isolate storm peaks of independent storm events from a data base where hour-by-hour wind records are available in a computer readable form. The minimum requirement is a collection of complete
years with sufficient continuity within the years. The method can be demonstrated via the following steps (Harris, 1999):

(i) The continuous hourly wind records are first subjected to a simple block-averaging low-pass filter technique with a period of 10 hr. The purpose is to eliminate any shorter temporal scale maxima resulting from meteorological mechanisms such as thunderstorms.

(ii) The average records are searched to establish the times at which they fall below an arbitrary lower bound threshold (i.e. 5 m/s in this case). Cook (1982) shows that, as this threshold value used to define “lulls” is raised, the apparent number of storms per year decreases. Thus the threshold should be kept sufficiently low to ensure that the annual storm rate per year is greater or equal to the number of years (i.e. ensure at least one record per year). On the other hand, setting a higher threshold would help to eliminate any storm maxima from other meteorological causes which survived the filtering process in (step (i)).

(iii) Each downward crossing in the 10 hr averaged series defines the start of “lulls” in the record, and by definition between each pair of lulls, there is an independent storm.

(iv) By searching the original hourly data base between “lulls”, the maximum wind speed occurring in each independent storm can be established.

(v) These independent storm maxima can then be subjected to extreme value analysis to arrive at extreme wind recurrences. For instance in
the POT extreme value analysis all independent wind maxima above a pre-determined threshold will be subjected to extreme value analysis.

A computer program developed in “Perl” used here to implement the method described above is presented in appendix (5-A).

5.5.6 Simulate the effect of climate variability on wind extremes via covariates.

Many studies suggest a relationship between climate excursions (i.e. ENSO) and the NE Pacific wind frequency and intensity (Cavazos and Rivas, 2004; Tuller, 2004; Bromirski et al., 2005). In this study, the statistical dependence of the extreme wind distribution and various climate variability patterns is studied by incorporating climate variability indices as covariates within the extreme-value model parameters. The analysis relates storm forced winds with speeds of 40-50 km/hr or higher and the monthly climate variability indices. Upon applying the extreme wind speed criterion on the original data bases, the study essentially reduces to that of a fall-winter seasonal extreme wind analysis with corresponding circulation indices. This results, because, according to the Environment Canada wind classification, the winds in the BC region are weak during summers and cannot be considered as storm forced.

As the first step towards developing extreme wind recurrence models with climate considerations, the study focus is limited to simple linear relationships between climate excursions and the extreme wind regime. As such the climate-variability indices \(Y\) are related to the GPD model coefficients \(\sigma\) and \(\xi\) as linear covariates (Eqn 5.10).
In other words, possible lag/lead relationships are not investigated in this study. However, the importance in exploring possible lag/lead relationships between extreme winds and various climate patterns within the context of extreme values analysis is recognized, particularly when the climate excursion leads the extreme wind regime. Such analysis could provide useful information on dominant climate variability patterns that trigger a prevailing extreme wind regime transformation from one dominant state to another. Therefore analysis including circulation indices with proper lags that are known to have affected the BC’s regional climate is warranted.

5.5.7 The question of co-linearity between climate indices

Initially the candidate climate indices (i.e. MEI, PDO, PNA, ALPI and NOI) are applied in the GPD model parameters as individual covariates. The results of the fit diagnostics are tested for significance (at the 5% level) against a “base model”, which is the model without climate covariates. Any individual climate covariate that does not show statistically significant improvements in the GPD model fit against the “base model” are considered superfluous, while those that significantly improve the model fit are carried forward to the next level of analysis.

Evidently many large-scale climate regimes described by various climate variability indices are related. For example, the strong climate regime shift that occurred in mid-1976 was captured by all of the climate indices considered in this analysis at approximately the same time. The outcome suggests that the climatic
phenomena that triggered the 1976-regime shift equally and dramatically affected a wide range of climate indices (Figure 5.4). Due to these shared patterns (i.e. colinearities), climate indices may show compatible responses when applied as individual covariates within a GPD model. On the other hand, when extreme wind climatology in a region is significantly affected by climate characteristics unique to a given climate excursion, the corresponding climate indices should dominate in terms of statistical significance over others, in a GPD model that combines more than one climate covariate. Following this, a systematic approach (i.e. "Redundancy test") is developed to eliminate the co-linearity issues between various climate variability indices.

5.5.7.1 The Likelihood ratio test

The likelihood ratio test is used to infer whether a parameter-rich model of sequence evolution (i.e. "alternative model with climate covariates") fits an extreme wind data set significantly better than a simpler model which has fewer parameters, (i.e "the base model"). The test can be performed as follows.

Consider comparing two models, a simple (Base) model that has \( N_0 \) free parameters and a parameter-rich (Alternative) model with \( N \) free parameters (where \( N > N_0 \)).

(i) The simple model is fitted to the data and its (maximal) log-likelihood is recorded (denoted \( \text{ln}L_0 \)).

(ii) The parameter-rich model is fitted to the data and its log-likelihood is recorded (\( \text{ln}L \)).
(iii) Twice the difference in negative log-likelihoods \( (nLL \text{ and } nLL_0) \) is computed - this is the test statistic: \( D = 2(nLL_0 - nLL) \). Here the parameter-rich model will always have a better fit, due to the extra parameters and will therefore have the highest log-likelihood, so this is a positive number.

(iv) The test-statistic can be used to find the corresponding p-value by comparing it to a chi-squared distribution with \( (N - N_0) \) degrees of freedom (i.e. if the one model has 2 free parameters and the subsequent alternative model has 3, then the statistic should be compared to a chi-squared distribution with 3 - 2 = 1 degrees of freedom). This way one can test whether a parameter-rich model is significantly better than a simpler model. One important condition that has to be fulfilled to apply the likelihood ratio test is that the simpler model must be a constrained version of the parameter-rich model.

5.5.7.2 The Redundancy test

The idea behind this test is to choose the GPD model that best describes the data without having more parameters than strictly necessary. This is achieved by fitting a series of increasingly more complex GPD models to all selected climate indices successively. Each time the fit diagnostics of each successive model is cross-compared with the former via a likelihood ratio test, to determine whether the more complicated model is significantly better than the previous model. The Climate indices that have a significant contribution towards improving the model fit at all levels (i.e. without becoming redundant) are chosen
for the final GPD model to simulate the climate effects on extreme storm surge recurrences. This approach essentially avoids repetition of climate patterns that are common to many climate indices within the GPD models and by that simplify the final GPD model while satisfactorily accounting for climate variability. Table (5.2) demonstrates the application of this test on extreme wind data at station YVR.

5.5.8 Conditional Extreme Value Models

In order to make meaningful inferences from GPD models with random variables as covariates (i.e. climate covariates), the models must be reduced to that of a time varying function (see Eqn. 5). This procedure is accomplished by making the GPD model with climate covariates conditional on distinct climate states, in which case the climate state is represented via a range of values assumed by the climate variability indices. Based on the Environment Canada climate classification scheme, three distinct climate states were selected to demonstrate the effects of climate variability on extremes: (a) the extremely warm (strong El Niño), (b) extremely cold (strong La Niña) and (c) Neutral. Table (5.3) summarizes the corresponding climate indices averaged over the range of years classified under the three climate states.
5.6 Results.

5.6.1 Extreme non-directional and directional wind recurrences

5.6.1.1 Extreme non-directional wind recurrences (non-directional base-model).

All non-directional independent wind events above the threshold of 40 km/hr for YVR and 50 km/hr for Sandheads and Saturna are considered. The threshold wind speed $u$ for fitting GPD models at each station is chosen from graphs typical of Figure 5.3; where again, this threshold is selected in order to ensure that the assumptions for the appropriateness of the GPD are met while minimizing the variability in the estimates. The station specific fit diagnostics and maximum-likelihood fitted parameters in Table 5.4 indicate negative shape parameters for all three stations, suggesting bounded upper tails. The diagnostic quantile plots of each station indicate that the underlying assumption of the GPD distribution is reasonable for these data (Figure 5.5).

Table 5.5 summarise the extreme wind return levels corresponding to non-directional base-models (i.e. without climate covariates) of YVR, Saturna and Sand Heads. Clearly the rate of extreme wind recurrence at Saturna is much higher than the other two stations. For instance, Saturna experiences storm-forced winds (> 85 km/hr as per Environment Canada storm warning scale) at least once every year; while the other two stations (Sand Heads and YVR) experience storm forced winds only once every 15 and 50 years, respectively. Further, the 50-year return level at Saturna is about 12% higher than the corresponding YVR level. These differences in return level projections are almost certainly due to the changes in the station exposure resulting from the local topography.
All the non-directional base models satisfactorily follow the extreme wind speed tail distribution, with almost all measurements falling within the uncertainty bounds. There is however an apparent under estimation of extreme wind recurrences when these are extrapolated to longer recurrences, particularly at YVR and Sandheads stations (Figure 5.6). The failure to capture the extreme tail end could be attributed to the presence of non-stationarity and mixed wind climates in the data sets resulting from decadal to inter-decadal scale climate variability effects and/or lack of directional resolution, respectively. Using longer extreme wind time series to stabilize extreme wind quantiles and, accounting for the directionality of the wind regime by developing separate GPD models for dominant wind directions, may help eliminate issues related to non-stationarity and improve the tail predictions substantially.

5.6.1.2 Extreme directional wind recurrences (directional base-model).

This section analyses the extreme directional wind recurrences (i.e. southerly (S), westerly (W) and northerly (N) directional sectors), where directional wind sectors are categorized following Lange (1998); (Table 5.1). Table (5.6) summarises the station specific extreme wind return level projections without climate considerations (i.e. base models for directional winds). Non-directional wind recurrences are also indicated for comparison purposes. The results indicate southerly wind dominance at Saturna and Sand Heads, while at YVR the westerly winds are the strongest. Note, however, that the southerly wind regime at Saturna is quite close in compression to the westerly winds due to its greatest exposure to winds blowing from all directions in the Georgia Strait. Overall the
southerly winds are the strongest winds in the Boundary Bay region, while northerly winds remain the weakest at all three stations. Saturna has a much stronger Northerly wind regime when compared to the other two stations. This finding suggests that the region encompassing YVR and Sand Heads stations, including the Boundary Bay region, is protected from Northerly winds, likely due to the presence of higher land to the North. It is also evident from the results that the strongest wind regime in the Georgia Strait gradually transforms from a more southerly (S-SE) to westerly (SW-W) regime when blowing towards the north along the strait.

5.6.2 Effect of Climate Variability on extreme winds

A qualitative assessment of the scatter plot of annual number of independent extreme wind events above 40 km/hr threshold at YVR indicates a clear early 1960s and 70s dominance with respect to both magnitudes and frequency of occurrence (Figure 5.7). Further, a decreasing trend is seen towards the early 1980’s before the number of events and their strengths start to increase by the mid 1990s. Clearly the scatter plots indicate strong variations in the wind extremes that occur with time scales that are typically on the order of 2-7 years, while changes are also visible on an inter-decadal scale.

The observational evidence at all three stations do not indicate a trend towards increasing winds during the early to mid 1980’s. In fact, the early to mid 1970s seem to be in general “rougher” than the 1980s. Despite this, during the strong El Niño period of December 1982, which was the most severe storm surge on record occurred in the Boundary Bay region (Abeysirigunawardena et al., (in
This observation, and the fact that extreme winds in southern BC were not overly strong in the early 1980s, suggests a possible out of phase relationship between extreme winds and extreme Total Water-levels in the region. It also indicates that local winds may not be the primary factor controlling extreme surges in the region.

5.6.2.1 Effect of Climate Variability on extreme non-directional wind recurrences.

To discern the effect of climate variability on extreme non-directional winds, more complex extreme-value models with climate variability information as covariates were considered. Initially, MEI, NOI, PDO, PNA and ALPI indices were selected as natural candidates to represent the effect of climate variability in the GPD distribution. The significance of the individual climate indices within the GPD model was estimated via the negative log likelihood estimate against the base-fit. The results in Table 5.7 indicate that, MEI in the shape parameter ($\xi$) and MEI, PDO and NOI in the scale parameter ($\sigma$) produce significant model improvements at station YVR compared to the corresponding base-model. The lack of significance in climate covariates at the Sand Heads and Saturna stations is attributed to insufficient data coverage.

A notable resemblance exists between the climate covariate coefficient's trends (i.e. sign) of all three extreme non-directional wind models, particularly for the MEI, NOI and PDO indices. For example, the MEI coefficient corresponding to the scale and shape parameters was negative for all three stations, while significant model improvement was limited to YVR. The fact that many climate coefficients follow similar trends at more than one station in the vicinity of the
study region indicates spatial consistency in the controls of extreme wind regime in the region. The final non-directional extreme wind model for station YVR obtained through the redundancy analysis is summarized in Table (5.8) and (Eq. 5.11.1 – 5.11.3). The reduced log-likelihood compared to the Base-model in Table (5.8) confirms a significantly improved GPD tail prediction with climate considerations.

\[
\begin{align*}
\mu &= 40.0 \\
\xi &= -0.096 \\
\sigma &= 8.97 -0.586(MEI)-0.463)(PDO)+(0.209)(NOI)
\end{align*}
\]

The effect of climate variability on extreme wind return-level projections at YVR was demonstrated by making equation (5.11.3) conditional on warm, neutral and cold climate episodes (Table 5.9 and Figure 5.8). Accordingly the neutral state is almost equivalent to the base case while extreme winds during warm ENSO conditions generally remained weaker than its cold counterpart.

5.6.2.2 Effect of Climate Variability on extreme directional wind recurrences

The effect of Climate Variability on directional winds was investigated by fitting different GPD models with climate covariates into directionally categorized extreme winds. Table (5.10) summarises the corresponding covariate coefficients and their significance for the three stations. Station YVR has the highest number of significant climate variability covariates while Saturna and Sandheads show the least, likely due to lack of temporal extent in the two latter series. Nonetheless, a close resemblance exists between the covariate trends (i.e. sign) at YVR and the other two stations, particularly those that are significant at YVR.
This observation indicates that spatial coherence exists in extreme directional winds response to climate variability in the study region.

At the YVR station, MEI, PDO and NOI exert significant controls on the westerly (W) winds, while the PDO, NOI and ALPI influence northerly (N) winds and NOI the southerly(S) winds. Return levels are projected for YVR by setting the covariates in the final GPD model, conditional on warm, neutral and cold climate states (Table 5.11). As seen, the wind recurrences corresponding to all three directions were much stronger under cold ENSO states while extreme winds during warm episodes generally remained weaker than the cold counterpart. The most significant difference in directional wind responses with respect to warm vs cold episodes was seen in the Westerly wind regime, while the least differences are observed in the Northerly wind regime.

5.7 Effect of Global Climate Change on extreme winds.

As the greenhouse gas levels in the atmosphere change, extreme storm frequency and intensity may change as well. Even though this may be the case, the exact connection between climate change and extreme events has not been established adequately to reach any conclusive results for such responses. One major shortfall is a lack of resolution in the Global and Regional Climate Models (GCMs & RCMs) to accurately represent extreme events. Nevertheless, the rapid development of climate modeling capabilities in concert with increased computer resources has resulted in recent improvements in the model resolution, simulation errors and parameterisation of processes, making such studies more
credible. However, results are still mixed and highly uncertain (Meehl et al., 2000). For example, studies on climate change and storminess states conflicting effects of climate change on extratropical storms. While some models project storms under climate change to be fewer in number but stronger in intensity (Lambert, 1995), other models project not only more but more intense storms (Carnell and Senior, 1998; Frei et al., 1998). Despite these discrepancies, GCMs remain the best tool for accounting for the effect of greenhouse-gas accumulations but need to be refined to make local scale inferences.

As a step towards this, Timmermann et al. (1999) applied a high resolution version of a GCM to simulate ENSO under conditions of warming. The results indicated tropical Pacific climate systems undergoing systematic changes when greenhouse gas concentration doubled. In particular, their results suggest a world with average conditions similar to typical El Niño conditions while La Niña episodes were projected to become more intense with strong inter-annual variability.

In this study, the differences in the GCM model projections are demonstrated by comparing the monthly averaged daily maximum wind speeds for existing and changing climate states as simulated by the Canadian Global Climate Model (CGCM3.1/T63). Here, the future climate state is chosen as the A1B scenario of the IPCC 20th Century Experiment (i.e. "Balanced" progress across all resources and technologies from energy supply to end use; IPCC, 2007), while the current conditions are represented by the real-time daily extreme wind projections made by the same model for the period 1961-2000 (base
model). Results are presented at two adjacent grids in the vicinity of the study region (126.56W, 48.84N and 123.75W, 48.84N) (Figure 5.9). The patterns associated with each location indicate satisfactory seasonal extreme wind variability projections by the CGCM model. The model also projects a consistent and significant reduction in the extreme daily summer (May-September) winds with increased greenhouse gas emissions at both grids, while projections for the winter months (October – March), which are also the stormiest months of the year, remain mixed between the adjacent grids. For instance, at grid 123.75W, 48.84N (Figure 5.9-top panel), which is predominantly covered by the landmass, projects significant increases in extreme winter winds in response to increasing greenhouse gas emissions, while the differences in the adjacent grid that is mostly covered by the water body (126.56W, 48.84N: Figure 5.9- bottom panel) are smaller and not as persistent from month to month. This makes it nearly impossible to make a conclusive interpretation on the effects of long term climate change on the winter wind regime in the study region.

5.8 Discussion and Conclusions

This study examined the effects of climate variability and change on extreme wind speed recurrences at three meteorological stations located within the inner south coast of BC. The methodology proposed dealt with three issues in extreme value analysis: the importance in event independence, selection of appropriate thresholds and the incorporation of climate variability effects in the extreme value distribution. Shifts in the distribution of extreme winds were
identified in association with the broad-scale atmospheric (MEI, NOI) and oceanic (PDO) circulation modes. The results of these analyses suggest strong atmospheric and oceanic controls on the prevailing extreme wind regime in the study region. They highlight a dominant increase in BC’s inner south coast extreme winds associated with the ENSO cold phase (i.e. La Niña) signal, particularly in the fall and winter. Physical mechanisms that connect southern BC extreme winds to ENSO may be related to changes in storm direction and/or frequency; however, this verifying connection awaits further analysis.

The suggested relationship of increased (or decreased) extreme winds associated with ENSO cold (warm) phases is not limited to the current study; past research has also supported similar relationships. For example, Bromirski et al. (2005) investigated the trends in the extreme storms in North Pacific via the variation in the distribution of long period waves. They attributed the NE Pacific extreme storm characteristics to that of high-amplitude long period waves, as these waves are generated only by high sustained wind speeds over a large fetch. Their results indicate more northerly displaced storminess and heightened wave energy during La Niña episodes. The same study reports observed upward trend in long period wave energy during La Niña episodes at a buoy offshore of Washington, which is in close proximity to the Inner-south-coast of BC. Further, Schwing et al. (2002) demonstrated greater pole ward wind stress accompanied with cooler SSTs (i.e. cold ENSO characteristics) north of 44°N, in locations off Oregon and Washington also in close proximity to southern BC. Harper (2005) documents dominant cold phase signal in the Northern Great-plains November
wind speeds, where a relationship was suggested between La Niña (i.e. Cold ENSO) and more frequent occurrence of higher median wind speeds. While it appears a relationship exists between cold ENSOs and strong winds in the inner south coast of BC (North of 44° N), many studies suggest that extreme winds are opposite in nature in more southern latitudes of the Pacific coastal margin (i.e. Southern California), where more active storms are generally observed during warm ENSO episodes (e.g. Seymour, 1984; Storlazzi et al., 2000; Graham and Diaz, 2001; Allan and Komar, 2002; Chang et al., 2002; Bromirski et al., 2005).

Previous studies established that ENSO is well-connected to upper-level air flow over North America, and to storm tracks that characterize warm ENSOs, when the jet stream extends equator-ward and much farther downstream than during ‘normal’ winters (e.g. Horel and Wallace, 1981; Wallace and Gutzler, 1981; Trenberth and Hurrell, 1994; Bromirski et al., 2005). This trajectory could bring more active land falling winter storms to southern latitudinal locations such as California (Graham and Diaz, 2001; Chang et al., 2002; Chang and Fu, 2003) than to more northern locations such as the south coast of BC during a warm ENSO episode (see Figure 5.10). On the other hand during cold ENSO events, the jet stream shifts northwards (e.g. Hoerling et al., 2001; Rodionov and Assel, 2001; Enloe et al., 2004) bringing heavy wind storms to southern BC. These shifts indicate that the signals associated with ENSO and extreme winds lack spatial consistency over the extent of the eastern Pacific coastal margin. Consequently using a single wind time series to represent climate variability on a wider spatial scale could give an incomplete, if not inaccurate, view of the actual
relationship between extreme winds and climate variability on a much larger spatial scale.

Residual water-levels (Total water level–Tide) can be used as a measure of storminess in a region and have been shown to vary in association with ENSO phases. For instance, the Storm Surge Almanac for the SW BC Fall/Winter 2007-2008 (Tinis, 2008) suggests that, during strong El Niño events a combination of low pressure over the Pacific Ocean and thermal expansion of warm water causes residual water levels along the eastern Pacific Ocean to rise by 20 to 30 cm. A similar magnitude depression characterizes strong La Niña events (Tinis, 2008). This observation indicates that a contrasting relationship exists between residual water levels and prevailing extreme winds in the region (Abeysirigunawardena et. al (in review), (Chapter 4)). Arguably the most important conclusion arising from this observation is the important role played by controls related to warm ENSO events. Notable are the thermosteric effects on residual water levels in the study region that result in substantially elevated residuals during warm ENSOs, even under moderately extreme winds.

Evidently, natural climate variability continues to exert an important influence on the wind regime of the study region even as the changes induced by rising concentrations of green house gases begin to be felt. These natural variations in climate complicate the detection of the greenhouse effects. It was not possible to reach a conclusive interpretation of the effects of CC on storminess in southern BC.
Climate excursions are a major perturbation of the Earth’s climate system that involves large-scale changes in wind and surface pressure. Clearly North American weather patterns are also affected. Understanding the relationships between environmental variables and climate excursions, and finding ways to simulate such effects statistically, is important to longer-term response forecasting. The objective of this study was to provide an initial assessment of the extreme surface wind responses that typically occur under different climate states. Knowledge about these response patterns is needed in order to improve the wind forecasting skills. However, limitations in the data sets over both space and time make it possible that this study has not identified all the important aspects of such a response.

For Canada as a whole, several aspects of extreme weather events have not been fully assessed due to unavailability of long period data. Few comprehensive studies exist that focus on severe weather events at regional to smaller spatial scales. There is a need to continue monitoring trends of extreme events, particularly for assessing the response of such events to climate variability and change on the regional and smaller spatial scales. Further more the economic and societal impacts of weather extremes and its linkage to climate variability and global warming needs to be more accurately assessed for Canada. Improved understanding on the association between climate variability and extreme winds, coupled with the increased capacity for ENSO forecasting may lead to increased safety awareness.
References


Asian southwest monsoon during the Holocene and their links to the North


Harper, B., 2005. ENSO’s effect on the wind energy production of South Dakota.
Significant Opportunities in Atmospheric Research and Science (SOARS)
program of the University Corporation for Atmospheric Research. 14 p.

of Wind Engineering and Industrial Aerodynamics, 80: 1-30.


climate response to ENSO’s extreme phases. Journal of Climate, 14:
1277–1293.

associated with the Southern Oscillation, Monthly Weather Review, 109:
813–829.

generalized Pareto distribution. Technometrics, 29:339–349

from Encyclopedia of Coastal Science. The Earth Science Encyclopedia
Online (www.eseo.com)

IPCC 2007. Intergovernmental Panel on Climate Change (IPCC), Climate
Change 2007: The Physical Basis. Summary for Policymakers (Geneva:
World Meteorological Organization (WMO) and UN Environment Program


Table 5-1: Dominant directional wind sectors in the Inner-south-coast of BC. Note that the definitions of the dominant directions are also valid for the Boundary Bay region (Lange, 1998)

<table>
<thead>
<tr>
<th>Basic Winds</th>
<th>Dominant Wind Directional Sectors</th>
<th>Regional Patterns (Directional Range)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northerly</td>
<td>N, NE</td>
<td>340-80</td>
</tr>
<tr>
<td>Easterly</td>
<td>E, SE</td>
<td>30-240</td>
</tr>
<tr>
<td>Southerly</td>
<td>S, SW</td>
<td>130-240</td>
</tr>
<tr>
<td>Westerly</td>
<td>W, NW</td>
<td>210-30</td>
</tr>
</tbody>
</table>
Table 5-2: The redundancy analysis process. (Application of redundancy test on non-directional wind data at station YVR.)

<table>
<thead>
<tr>
<th>Model</th>
<th>Parameter</th>
<th>Negative log-likelihood</th>
<th>Significance compared with the Chi-Square Value for 0.05 significance level</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>BASE</td>
<td>2409.5</td>
<td></td>
</tr>
<tr>
<td>2.1</td>
<td>BASE + CV (σ 1 = MEI)</td>
<td>2401.3</td>
<td>16.4&gt;3.84</td>
</tr>
<tr>
<td>2.2</td>
<td>BASE + CV (σ 2 = NOI)</td>
<td>2401.9</td>
<td>15.3&gt;3.84</td>
</tr>
<tr>
<td>2.3</td>
<td>BASE + CV (σ 3 = PDO)</td>
<td>2404.3</td>
<td>10.6&gt;3.84</td>
</tr>
<tr>
<td>2.4</td>
<td>BASE + CV (ξ 1 = MEI)</td>
<td>2407.1</td>
<td>4.8&gt;3.84</td>
</tr>
<tr>
<td>3.1</td>
<td>BASE + CV (σ 1 = MEI; σ 2 = NOI)</td>
<td>2399.2</td>
<td>4.2&gt;3.84</td>
</tr>
<tr>
<td>3.2</td>
<td>BASE + CV (σ 1 = MEI; σ 3 = PDO)</td>
<td>2400.4</td>
<td>1.8&lt;3.84</td>
</tr>
<tr>
<td>3.3</td>
<td>BASE + CV (σ 1 = MEI; ξ 1 = MEI)</td>
<td>2401.0</td>
<td>0.6&lt;3.84</td>
</tr>
<tr>
<td>3.4</td>
<td>BASE + CV (σ 2 = NOI; σ 3 = PDO)</td>
<td>2399.1</td>
<td>4.4&gt;3.84</td>
</tr>
<tr>
<td>3.5</td>
<td>BASE + CV (σ 2 = NOI; ξ 1 = MEI)</td>
<td>2401.7</td>
<td>&lt;3.84</td>
</tr>
<tr>
<td>3.6</td>
<td>BASE + CV (σ 3 = PDO; ξ 1 = MEI)</td>
<td>2403.8</td>
<td>&lt;3.84</td>
</tr>
<tr>
<td>4.1</td>
<td>BASE + CV (σ 1 = MEI; σ 2 = NOI; σ 3 = PDO)</td>
<td>2397.0</td>
<td>4&gt;3.84</td>
</tr>
<tr>
<td>4.2</td>
<td>BASE + CV (σ 1 = MEI; σ 2 = NOI; ξ 1 = MEI)</td>
<td>2399.1</td>
<td></td>
</tr>
<tr>
<td>4.3</td>
<td>BASE + CV (σ 1 = MEI; σ 3 = PDO; ξ 1 = MEI)</td>
<td>2399.9</td>
<td></td>
</tr>
<tr>
<td>4.4</td>
<td>BASE + CV (σ 2 = NOI; σ 3 = PDO; ξ 1 = MEI)</td>
<td>2399.2</td>
<td></td>
</tr>
<tr>
<td>5.1</td>
<td>BASE + CV (σ 1 = MEI; σ 2 = NOI; σ 3 = PDO; ξ 1 = MEI)</td>
<td>2397.0</td>
<td>&lt;3.84</td>
</tr>
</tbody>
</table>

SELECTED FINAL MODEL

BASE + CV (σ 1 = MEI; σ 2 = NOI; σ 3 = PDO)
Table 5-3: Average climate indices for “strong El Niño”, “strong La Niña” and “neutral” years. The definitions of strong El Niño / La Niña years are based on the Environment Canada classification scheme (Meteorological Services of Canada-The Green Land).

<table>
<thead>
<tr>
<th>MAXSL (Chart cm)</th>
<th>PDO</th>
<th>MEI</th>
<th>NOI</th>
<th>PNA</th>
<th>ALPI</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>strong El Niño (Warm ENSO) years: 1957, 1982, 1991, 1992, 1998</strong></td>
<td>536</td>
<td>0.219</td>
<td>0.928</td>
<td>-1.486</td>
<td>0.105</td>
</tr>
<tr>
<td><strong>strong La Niña (Cold ENSO) years: 1973, 1974, 1988, 1989, 1999</strong></td>
<td>521</td>
<td>-0.370</td>
<td>-0.688</td>
<td>1.185</td>
<td>0.081</td>
</tr>
</tbody>
</table>
Table 5-4: Maximum-likelihood fitted parameters (MLE) for YVR, Sand Heads and Saturna Stations, having fit a GPD models for the extreme non-directional wind events above a threshold $u$ (Base Model: B).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>YVR</th>
<th>Sandheads</th>
<th>Saturna</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>MLE</td>
<td>Stand. Error</td>
<td>MLE</td>
</tr>
<tr>
<td>(Threshold ) $u$</td>
<td>40</td>
<td>-</td>
<td>50</td>
</tr>
<tr>
<td>(Scale) $\sigma$</td>
<td>9.21</td>
<td>0.441</td>
<td>11.13</td>
</tr>
<tr>
<td>(Shape) $\xi$</td>
<td>-0.091</td>
<td>0.032</td>
<td>-0.218</td>
</tr>
</tbody>
</table>
Table 5-5: The extreme non-directional wind recurrences at YVR, Sand Heads and Saturna. Projections are based on the GPD fit of all extreme non-directional winds above 40 km/hr threshold for YVR and 50 km/hr threshold for Sandheads and Saturna stations.

<table>
<thead>
<tr>
<th>Return Period (Yr)</th>
<th>Return Level based on POT method (Km/hr)</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(YVR)</td>
<td>(SAND HEADS)</td>
<td>(SATURNA)</td>
</tr>
<tr>
<td>2</td>
<td>66.31</td>
<td>76.5</td>
<td>89.3</td>
</tr>
<tr>
<td>5</td>
<td>72.31</td>
<td>80.9</td>
<td>92.2</td>
</tr>
<tr>
<td>10</td>
<td>76.53</td>
<td>83.7</td>
<td>93.7</td>
</tr>
<tr>
<td>15</td>
<td>78.88</td>
<td>85.2</td>
<td>94.3</td>
</tr>
<tr>
<td>20</td>
<td>80.49</td>
<td>86.1</td>
<td>94.7</td>
</tr>
<tr>
<td>25</td>
<td>81.71</td>
<td>86.8</td>
<td>95.0</td>
</tr>
<tr>
<td>30</td>
<td>82.69</td>
<td>87.4</td>
<td>95.2</td>
</tr>
<tr>
<td>40</td>
<td>84.21</td>
<td>88.2</td>
<td>95.5</td>
</tr>
<tr>
<td>50</td>
<td>85.36</td>
<td>88.8</td>
<td>95.7</td>
</tr>
<tr>
<td>75</td>
<td>87.38</td>
<td>89.9</td>
<td>96.0</td>
</tr>
<tr>
<td>100</td>
<td>88.78</td>
<td>90.5</td>
<td>96.2</td>
</tr>
</tbody>
</table>

Note: Threshold: YVR= 40 km/hr; Sandheads & Saturna = 50 km/hr
Table 5-6: Extreme Southerly (S), Northerly (N) and Westerly (W) wind recurrences without climate considerations (Base-directional) at stations YVR, Sandheads and Saturna. Projections are based on the GPD fit of all extreme directional winds above a threshold u as indicated in the table.

<table>
<thead>
<tr>
<th>Return Period (Yr)</th>
<th>YVR</th>
<th>Sandheads</th>
<th>Saturna</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Non-Dir. Winds &gt;40 Km/hr</td>
<td>S &gt; 40 Km/hr</td>
<td>N &gt; 30 Km/hr</td>
</tr>
<tr>
<td>2</td>
<td>66.3</td>
<td>53.9</td>
<td>36.3</td>
</tr>
<tr>
<td>5</td>
<td>72.3</td>
<td>59.3</td>
<td>40.1</td>
</tr>
<tr>
<td>10</td>
<td>76.5</td>
<td>63.0</td>
<td>42.8</td>
</tr>
<tr>
<td>15</td>
<td>78.9</td>
<td>65.0</td>
<td>44.4</td>
</tr>
<tr>
<td>20</td>
<td>80.5</td>
<td>66.3</td>
<td>45.4</td>
</tr>
<tr>
<td>25</td>
<td>81.7</td>
<td>67.3</td>
<td>46.2</td>
</tr>
<tr>
<td>30</td>
<td>82.7</td>
<td>68.1</td>
<td>46.9</td>
</tr>
<tr>
<td>40</td>
<td>84.2</td>
<td>69.3</td>
<td>47.9</td>
</tr>
<tr>
<td>50</td>
<td>85.4</td>
<td>70.2</td>
<td>48.7</td>
</tr>
</tbody>
</table>
Table 5-7: Maximum-likelihood fitted parameters for YVR, Saturna and Sand Heads stations, having fit a GPD model with climate covariates. The highlighted values indicate significantly improved GPD fits (at 95% level) over the base fit when corresponding climate index was applied as a covariate in the model.

<table>
<thead>
<tr>
<th>Climate Indices</th>
<th>Shape Parameter ($\xi$) Covariates</th>
<th>Scale Parameter ($\sigma$) Covariates</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>YVR</td>
<td>SATURNA</td>
</tr>
<tr>
<td>MEI</td>
<td>-0.07**</td>
<td>-0.021</td>
</tr>
<tr>
<td>PDO</td>
<td>-0.10</td>
<td>-0.022</td>
</tr>
<tr>
<td>NOI</td>
<td>0.01</td>
<td>0.002</td>
</tr>
<tr>
<td>PNA</td>
<td>0.02</td>
<td>0.027</td>
</tr>
<tr>
<td>ALPI</td>
<td>-0.003</td>
<td>0.010</td>
</tr>
</tbody>
</table>

*Note: ** - Significant at 95% confidence level*
### Table 5-8: Maximum-likelihood fitted parameters for YVR stations, having fit a GPD model with climate covariates. The likelihood ratio test between the Base model and the model with covariates indicate significant improvement in the model fit with covariates.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>MLE</th>
<th>Stand. Error</th>
<th>Lower Bound</th>
<th>Upper Bound</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\sigma$</td>
<td>8.970</td>
<td>0.442</td>
<td>8.530</td>
<td>9.410</td>
</tr>
<tr>
<td>$\sigma_1$(MEI)</td>
<td>-0.586</td>
<td>0.461</td>
<td>-1.047</td>
<td>-0.125</td>
</tr>
<tr>
<td>$\sigma_2$(PDO)</td>
<td>-0.463</td>
<td>0.323</td>
<td>-0.786</td>
<td>-0.140</td>
</tr>
<tr>
<td>$\sigma_3$(NOI)</td>
<td>0.209</td>
<td>0.104</td>
<td>0.105</td>
<td>0.313</td>
</tr>
<tr>
<td>$\xi$</td>
<td>-0.096</td>
<td>0.031</td>
<td>-0.127</td>
<td>0.065</td>
</tr>
</tbody>
</table>

Model name: **YVR gpd.fit: with covariates**

Likelihood-ratio test statistic for models: M0 = Base Case and M1 = gpd.fit: with covariates is: 23.1244 > 9.4877 = 1 - 0.05 quantile of a Chi-square with 4 degrees of freedom.

**p-value = 0.00012**
Table 5-9: Extreme non-directional wind recurrences at YVR subjected to climate variability effects. The variation in the return levels of extreme wind speeds are presented under warm, neutral and cold climate conditions.

<table>
<thead>
<tr>
<th>RETURN PERIOD</th>
<th>NO CV EFFECTS (BASE CASE)</th>
<th>RETURN LEVEL AT YVR (KM/HR)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>WARM CLIMATE</td>
<td>NEUTRAL CLIMATE</td>
</tr>
<tr>
<td>2</td>
<td>66.3</td>
<td>62.7</td>
</tr>
<tr>
<td>5</td>
<td>72.3</td>
<td>67.8</td>
</tr>
<tr>
<td>10</td>
<td>76.5</td>
<td>71.4</td>
</tr>
<tr>
<td>15</td>
<td>78.8</td>
<td>73.4</td>
</tr>
<tr>
<td>20</td>
<td>80.5</td>
<td>74.8</td>
</tr>
<tr>
<td>25</td>
<td>81.7</td>
<td>75.8</td>
</tr>
<tr>
<td>30</td>
<td>82.7</td>
<td>76.6</td>
</tr>
<tr>
<td>40</td>
<td>84.2</td>
<td>77.9</td>
</tr>
<tr>
<td>50</td>
<td>85.3</td>
<td>78.9</td>
</tr>
<tr>
<td>75</td>
<td>87.3</td>
<td>80.6</td>
</tr>
<tr>
<td>100</td>
<td>88.7</td>
<td>81.7</td>
</tr>
</tbody>
</table>
Table 5-10: Maximum-likelihood fitted parameters for YVR, Sandheads and Saturna, having fit a GPD model with climate covariates to Southerly (S), Northerly (N) and Westerly (W) winds.

<table>
<thead>
<tr>
<th>Climate Indices</th>
<th>Shape Parameter (ξ)</th>
<th>Scale Parameter (σ)</th>
<th>Non-Directional Winds</th>
<th>S</th>
<th>N</th>
<th>W</th>
<th>Non-Directional Winds</th>
<th>S</th>
<th>N</th>
<th>W</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>YVR Covariates</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MEI</td>
<td>-0.07**</td>
<td>-0.05</td>
<td>-0.03</td>
<td>-0.07**</td>
<td>-1.39**</td>
<td>-0.53</td>
<td>-0.43</td>
<td>-1.52**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PDO</td>
<td>-0.10</td>
<td>-0.03</td>
<td>-0.20**</td>
<td>-0.02</td>
<td>-0.91**</td>
<td>-0.14</td>
<td>-1.04**</td>
<td>-0.99**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NOI</td>
<td>0.01</td>
<td>0.06**</td>
<td>0.08**</td>
<td>0.01</td>
<td>0.35**</td>
<td>0.28</td>
<td>0.20</td>
<td>0.39**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PNA</td>
<td>0.02</td>
<td>0.06</td>
<td>-0.15</td>
<td>0.01</td>
<td>-0.09</td>
<td>0.72</td>
<td>-0.14</td>
<td>-0.31</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ALPI</td>
<td>-0.00</td>
<td>-0.02</td>
<td>-0.05**</td>
<td>-0.001</td>
<td>-0.15</td>
<td>-0.10</td>
<td>-0.38**</td>
<td>-0.21</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sandheads Covariates</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MEI</td>
<td>-0.04</td>
<td>-0.04</td>
<td>-0.09</td>
<td>-0.06</td>
<td>-0.48</td>
<td>-0.51</td>
<td>-1.26</td>
<td>-1.02</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PDO</td>
<td>-0.05</td>
<td>-0.04</td>
<td>-0.18</td>
<td>-0.12</td>
<td>-0.10</td>
<td>-0.20</td>
<td>-1.32</td>
<td>-1.08</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NOI</td>
<td>0.01</td>
<td>0.01</td>
<td>0.32</td>
<td>0.01</td>
<td>0.14</td>
<td>0.12</td>
<td>0.50</td>
<td>0.44</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PNA</td>
<td>0.09*</td>
<td>0.09</td>
<td>-0.08</td>
<td>0.13</td>
<td>1.02</td>
<td>0.87</td>
<td>-2.34**</td>
<td>1.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ALPI</td>
<td>-0.01</td>
<td>-0.00</td>
<td>-0.06</td>
<td>-0.04</td>
<td>-0.11</td>
<td>0.26</td>
<td>-0.68</td>
<td>-0.46</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Saturna Covariates</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MEI</td>
<td>-0.021</td>
<td>-0.023</td>
<td>-0.088</td>
<td>-0.226</td>
<td>-0.540</td>
<td>-0.602</td>
<td>-0.56</td>
<td>-3.965</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PDO</td>
<td>-0.022</td>
<td>-0.022</td>
<td>-0.41**</td>
<td>-1.03**</td>
<td>-0.525</td>
<td>-0.518</td>
<td>-5.63</td>
<td>-4.663</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NOI</td>
<td>0.002</td>
<td>0.002</td>
<td>0.04</td>
<td>0.04</td>
<td>0.008</td>
<td>0.059</td>
<td>-0.129</td>
<td>0.565</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PNA</td>
<td>0.027</td>
<td>0.028</td>
<td>-0.09</td>
<td>0.311**</td>
<td>0.697</td>
<td>0.589</td>
<td>-1.83</td>
<td>-0.668</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ALPI</td>
<td>0.010</td>
<td>0.011</td>
<td>-0.12</td>
<td>-0.060</td>
<td>0.281</td>
<td>0.359</td>
<td>-1.00</td>
<td>1.013</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: ** - indicate statistically significant improvements in the GPD fits at 95% level over the base fit, where the base fit is the model without climate covariates.
Table 5-11: Extreme Southerly (S), Westerly (W) and Northerly (N) wind recurrences at YVR under climate variability effects. The variations in the return levels of extreme wind speeds are presented under warm, neutral and cold climate conditions.

<table>
<thead>
<tr>
<th>Return Period (Yr)</th>
<th>YVR Southerly Winds (S)</th>
<th>YVR Westerly Winds (W)</th>
<th>YVR Northerly Winds (N)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>WARM CLIMATE</td>
<td>NEUTRAL CLIMATE</td>
<td>COLD CLIMATE</td>
</tr>
<tr>
<td>2</td>
<td>52.4</td>
<td>53.5</td>
<td>54.2</td>
</tr>
<tr>
<td>5</td>
<td>56.8</td>
<td>58.9</td>
<td>60.4</td>
</tr>
<tr>
<td>10</td>
<td>59.4</td>
<td>62.5</td>
<td>64.7</td>
</tr>
<tr>
<td>15</td>
<td>60.8</td>
<td>64.4</td>
<td>67.1</td>
</tr>
<tr>
<td>20</td>
<td>61.7</td>
<td>65.7</td>
<td>68.7</td>
</tr>
<tr>
<td>25</td>
<td>62.4</td>
<td>66.7</td>
<td>69.9</td>
</tr>
<tr>
<td>30</td>
<td>62.9</td>
<td>67.5</td>
<td>70.9</td>
</tr>
<tr>
<td>40</td>
<td>63.6</td>
<td>68.6</td>
<td>72.5</td>
</tr>
<tr>
<td>50</td>
<td>64.1</td>
<td>69.5</td>
<td>73.6</td>
</tr>
</tbody>
</table>
Figure 5-1: The geographical location of the southern coast of BC, including the lower Fraser Delta and the Boundary Bay region. Also shown are the three Meteorological stations YVR, Saturna and Sand Heads. [Figure modified from Lange, 1998].
Figure 5-2: The probability density plot for the extreme winds above 40 km/hr threshold at YVR; the fitted density curve following the data indicate that the selected GPD model for the data is a satisfactory choice.
Figure 5-3: Maximum likelihood estimates and confidence intervals of the GPD shape (bottom panel) and modified scale parameters (top panel) over a range of thresholds for YVR station. The dashed vertical line indicates the selected threshold for this data base.
Figure 5-4: Temporal distribution of NOI, PDO and ENSO indicating significant colinearities. Note the 1976 major climate regime shift being captured by all climate indices approximately at the same time.
Figure 5-5: Diagnostic quantile plot for (a) YVR, (b) Sand Heads and (c) Saturna stations from fitting all non-directional extreme wind events above a threshold of 40 Km/hr for YVR and 50 km/hr for Sandheads and Saturna to a GPD model. Note the units of both vertical and horizontal axis’s is km/hr.
Figure 5-6: Return level plots for (a) YVR, (b) Sand Heads and (c) Saturna stations from fitting all non-directional extreme wind events above a wind speed threshold of 40 Km/hr for YVR and 50 km/hr for Sandheads and Saturna to a GPD model. Note the unit of the vertical axis's (Return levels) is Km/hr and the horizontal axis (Return period) is Years.
Figure 5-7: Scatter plot of annual number of independent wind events above the 40 km/hr threshold for station YVR. Note the unit of the vertical axis is Km/hr.
**Figure 5-8**: Extreme non-directional wind recurrence curves at YVR with climate variability effects. Return period curves are presented for warm (long-dashed), neutral (short-dashed) and cold (dotted) climate conditions. Note the unit of the vertical axis is Km/hr.
Figure 5-9: Daily average extreme winds as projected by the CGCM (Canadian Climate Model) averaged over each month for the duration of the record for (a) Base case (1961-2000) (b) Scenario A1B (2046-2065) (c) Scenario A1B (2081-2100) at two locations adjacent to the study region (top panel: 123°.75W, 48°.84N, bottom panel: 126°.56W, 48°.84N).
Figure 5-10: Storm track enhancement (arrows) associated with 700 mb atmospheric pressure anomalies during La Niña (a) and El Niño (b) dominated climate patterns [from Inman and Jenkins, 2003].
223
Appendix (5-A)
Computer implementation of independent extreme wind event
identification.
(Language : Perl)
#!/usr/bin/perl -w
use strict;
if($#ARGV < 2) {
print("Usage: ./find_independent_storms.pl <file> <ws> <wt>\n");
exit(1);
}
# Filename, averaging timestep, and wind speed threshold
my($filename, $ws_time, $wt) = @ARGV;
my($fn, @lull);
# Step one: Find the LULLs
open($fn, $filename) or die("Couldn't open file " . $filename . "!\n");
# Skip header row
my($header);
$header = readline($fn);
# Print header
print(join(",", "start_off", "start_year", "start_month", "start_day", "start_hour",
"end_off", "end_year", "end_month", "end_day", "end_hour", "off", $header));
# Go through the file sequentially computing averages and maximums
between lulls
my($i, $ws_sum, $ws_sum_num, $ws_avg, $old_ws_avg, $ws_avg_num) =
(0, 0, 0, 0, 0, 0);
my($max_ws, $max_rec, $prev_rec) = (0, "", "");
my($year, $month, $day, $hour, $ws, $slp, $wd);
while(<$fn>) {
chomp();
my($line) = $_;
($year, $month, $day, $hour, $ws, $slp, $wd) = split(/,/, $line);
if($ws ne "NaN") {
$ws_sum += $ws;
$ws_sum_num++;
# Check if current ws is max ws; store record if so
if($ws > $max_ws) {
$max_ws = $ws;
$max_rec = join(",", $i, $line);


# If this is the first row, store this as the first time record
if($i == 0) {
    $prev_rec = join(",", $i, $year, $month, $day, $hour);
}

# If we're done summing up...
if($i != 0 && $i % $ws_time == 0) {
    # Only consider records if more than 75% of data is present
    if($ws_sum_num / $ws_time > 0.75) {
        # Store old average, compute new average
        $old_ws_avg = $ws_avg;
        $ws_avg = $ws_sum / $ws_time;

        # If previous ws_avg is GT and current ws_avg is LT
        if($ws_avg <= $wt && $old_ws_avg > $wt) {
            # We're at the end of a storm; print a summary record
            print(join(",", $prev_rec, $i, $year, $month, $day, $hour,
            $max_rec) . "\n");

            # Store this time record
            $prev_rec = join(",", $i, $year, $month, $day, $hour);

            # Reset maximums
            ($max_ws, $max_rec) = (0, "");
        }
    }
}

$ws_sum = $ws_sum_num = 0;

$i++;
}
close($fn);

# If last record is GT, then there will be an LT in future
if($ws_avg > $wt) {
    print(join(",", $prev_rec, $i - 1, $year, $month, $day, $hour, $max_rec) .
    "\n");
}
6.0 Influence of Climate Variability and Change on winter-storm track characteristics in the North-eastern Pacific

6.1 Abstract

The north-eastern Pacific winter storm track responses to climate variability and change was examined based on 56 years of January storm tracks derived from the 6 hourly NCEP–NCAR reanalysis data for the period of 1948 to 2004. The approach presented here differs from many, by the fact that storm track directionality (i.e. direction of propagation) is introduced in to the analysis. Results shows secular and decadal-scale changes in storm track frequency, intensity, life time and deepening rates. The north eastern Pacific storm track variability is found to be significantly correlated with regionally dominant climate variability modes such as ALPI, PNA and PDO, where the tracks from the northern quadrant strengthens (weakens) in response to cold (warm) ENSO like conditions, while the tracks from the southern quadrant strengthens (weakens) in response to warm(cold) ENSO like conditions. In general, the characteristic changes in the northerly (i.e. tracks with net equator ward displacement) and southerly (i.e. tracks with net poleward displacement) winter storm track characteristics in response to climate variability demonstrated opposite tendencies. Further, in the early 1970s a major transition from a weak to a much stronger storm track structure is indicated in the north eastern Pacific. In general, the decadal mean storm track intensity during the 1990s is substantially stronger than that of the late 1960s and early 1970s. It is shown that, the net northward shift of the southerly tracks (closer to the BC coastal stretch) and, the occurrence of stronger than normal northerly tracks could result in high wind-storms in
coastal BC region during a cold ENSO phase. Warm ENSO episodes even with winds at moderate strength could cause severe storm damages to the region, due to the increased probability of storm-high-tide interactions resulting from longer than normal storm track life spans.

6.2 Introduction

North eastern Pacific winter storm tracks are a key element of the winter weather and climate in the region. Evidence suggests systematic changes in the geographical location, intensity, and frequency of these storm tracks in response to climate variability (Reitan, 1979; Zishka and Smith, 1980; Parker et al., 1989; Changnon et al., 1995; WASA Group, 1998; Bengtsson et al., 2006; McDonald, 2006). For example, the well known climate mode ENSO (El Niño Southern Oscillation) is known to exert an important impact on the maintenance and development of Pacific storm tracks by extending (withdrawing) the tracks southward (northward) from their mean position during its warm (cold) phase (Zhang et al., 1997; Weijun and Zhaoboo, 1999).

The sharp transition of the interannual storm track variability during the early 1970s and the subsequent upward trend during the past few decades has been linked to the Pacific Decadal Oscillation, also an ENSO-like inter-decadal climate variability mode (Mantua et al., 1997; Zhang et al., 1997). Similarly, climate variability patterns described by the Pacific North American (PNA) teleconnection and the Aleutian Low Pressure Index (ALPI) have also been linked to the eastern Pacific cyclone frequencies during the last two decades (Gulev, 2001). Such systematic changes in the storm track
characteristics in response to climate variability may have longer term implications on regional weather and climate patterns. If this occurs, there are likely to be expenses due to high winds, precipitation and cloudiness they create, along with secondary impacts such as flooding, extreme waves and storm surges (Chang and Fu, 2002; Brayshaw, 2005).

Predictions of climate variability and change impacts in coastal BC are increasingly focussed on changes in variability and extremes, of which storm tracks are an important component. Despite recent evidence of serious storm damage to the coastal zones of BC (Abeysirigunawardena and Walker, 2008 (Chapter 2); Abeysirigunawardena et. al, (in review),(Chapter 4), changes in storm track characteristics in response to climate variability have not been studied to a great extent in BC.

In this study, the secular, decadal and interannual variability of the north-eastern Pacific winter storm tracks are investigated, using 56 years (1948-2004) of north-eastern Pacific (Lat 40°N- 60°N; Lon 120°W – 150°W) January storm track data as a proxy to the winter season. The main objective of the study was to investigate the extent to which north-eastern Pacific winter storm track variations are manifested by various climate oscillations such as the Multivariate ENSO Index (MEI), Pacific Decadal Oscillation (PDO), Northern Oscillation Index (NOI), Aleutian Low Pressure Index (ALPI) and Pacific North American teleconnection (PNA).
6.3 Observational evidence of storm track characteristics in coastal British Columbia.

6.3.1 A review on storm track characteristics in coastal British Columbia

According to the west coast monthly storm track composite maps developed by Environment Canada (i.e. based on synthesized climatological observations from 1957-1983), the majority of the storm tracks having direct impacts on coastal BC are initiated west of the international dateline and dissipate near Gulf of Alaska. A very few severe storms penetrate farther inland before dissipating. Generally, the highest winds in the north eastern Pacific correspond to tracks from the south sector (SSE-SSW).

Storms landing on the lower Vancouver Island almost always tend to produce intense winds on southern Vancouver Island and the south coast of BC down to Oregon. Southern storm tracks with landfalls from North Vancouver Island to the Queen Charlottes are further travelled thus may not reach the extreme range similar to the south coast of BC. However, there were a very few exceptions where the storm tracks with cyclone’s path located further north along the coastline of BC has produced a good winter gale (i.e. 2003 Christmas storm event in Queen Charlotte; Abeysirigunawardena and Walker, 2008 (Chapter 2)). When the storm tracks move further south from the southern Vancouver Island, the area of damaging winds moves away from the BC coast.

Many studies have linked the characteristics of the north-eastern Pacific storm track to climate variability and change (Trenberth and Hurrel, 1995; Seymour, 1998; Graham and Dias 2001; Allan and Komar, 2002). Besson and Neil (2002) studied storm track variability along coastal BC. Using
the pressure fall data and average significant wave height as a proxy for storminess, they found an increase in the frequency of extreme storms over Canada's West Coast during the past 50 years, with the sharpest increasing trend occurring over the last decade.

6.3.2 Pineapple-Express Storms vs. Climate Variability

Pineapple-express storms generate disastrous winter floods in Pacific Northwest coastal rivers. These storms are referred to as Pineapple-express storms because they draw heat and vapour from the subtropics or tropics near Hawaii and route that heat and vapour directly to the west coast of North America (Dettinger, 2004). More than 90% of the overall poleward water-vapour transport in the extratropics is due to Pineapple-express events (Zhu and Newall, 1998). “Pineapple-express” storm magnitudes show a modest maximum near 45° N latitude in close proximity to the Fraser Delta region of the southern coast of BC (Dettinger, 2004).

Historically, the most intense Pineapple-Express seasons have tended to occur when PDO was positive and ENSO was near neutral (Mantanu et al., 1997; Dettinger, 2004). Four of five winters with the most vigorous Pineapple-Express storms in western North America have occurred under positive PDO conditions, when the storm tracks across the North Pacific tended to be farther south than in other decades (Dettinger, 2004). During this phase of PDO, the Aleutian Low pressure system over the North Pacific is typically enhanced and storm tracks are more likely to be shifted southward, making it easier for cold fronts to draw heat and moisture northeast-ward from the subtropics and tropics. According to Dettinger
(2004), every El Niño winter from 1948 to 1999 yielded at least one Pineapple-Express circulation pattern that impacted the west coast of North America. In contrast, four of nine La Niña winters had no pineapple-express days. Storms in La Niña winters are shown to arrive more often from the west-southwest; while transports during El Niño storms arrive about half the time from the south-southwest.

6.3.3 The effects of global climate change on storm tracks.

In general the effect of global warming on the storm track activity is reported to be globally significant in the southern hemisphere than in the northern hemisphere (Konig et al., 1993). However, according to IPCC Third Assessment report (Houghton et al., 2001) as well as many other related publications, “there is no general agreement among climate models concerning future changes in mid-latitude storms and, there remains uncertainty with respect to the governing mechanisms”. For example, (i) some models predict a reduction in the number of cyclones but with an increase in the number of more intense cyclones (Lambert, 1995; Zhang and Wang, 1997; Lambert and Fyfe, 2006); (ii) some indicate a poleward shift in the storm tracks (Yin, 2005; Bengtsson et al., 2006), which has been linked to the sea surface temperature (SST) gradients (Inatsu et al., 2002; Inatsu and Hoskins, 2004; Bengtsson et al., 2006); (iii) some indicate no apparent change in the geographical distribution of the storm tracks (Lambert and Fyfe, 2006); and, (iv) some studies suggest no significant change in any cyclone characteristics at all (Bengtsson et al., 2006; Bengtsson et al., 2007; Bengtsson et al., 2008).
The uncertainties and mixed responses of storm tracks to climate change in Global Climate Models (GCMs) can be partly attributed to the difficulties in representing the noise introduced by the shorter temporal scale (decadal to inter decadal) natural climate variability within a GCM (Brayshaw, 2005; Johns et al., 2006; Ringer et al., 2006; Greeves et al., 2007). For example, a certain phase of natural climate variability can superimpose itself over the long-term climate signals leading to periods with decreasing storminess despite a long term increasing trend (Brayshaw, 2005). These uncertainty levels become more prominent on regional level, making it difficult to determine climate change signals at that scale. For example, while there has been an increasing tendency of the north-eastern Pacific storminess in the latter half of the 20th century, it is still unclear whether this is a natural climate variability phenomenon or a long term climate change signal. Thus, from a model evaluation perspective or when planning adaptation measures, it is important to account for the natural climate variability in a climate system and to distinguish this from the long term climate change signal (e.g. Ringer et al., 2006). To ensure a good representation of storm tracks in a coupled GCM, the ENSO signal needs to be properly accounted for, particularly in the eastern Pacific where the storm tracks and their characteristics are predominantly controlled by climate variability (Johns et al., 2006; Greeves et al., 2007). A better description of the historical storm track responses to various climate variability modes may provide ample information for this purpose.
6.4 Data

6.4.1 NCEP/NCAR re-analysis storm track projections.

The interannual variability of the north-eastern Pacific winter storm tracks was studied using 56-years (1948-2004) of north-eastern Pacific (Lat 40°N- 60°N; Lon 120°W – 150°W) January storm track data, produced by the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP/NCAR) reanalysis project (Serreze, 1995; Kalnay et al., 1996; Serreze et al., 1997; Kistler et al., 2001; Paciorek et al., 2002). NCEP/NCAR combines a fixed state-of-the-art global data assimilation, a model, and a comprehensive observational database, to construct continuous multi-decadal observational analyses of the atmosphere (Kalnay et al., 1996; Kistler et al., 2001; Paciorek et al., 2002). This approach allows comparison of climate information derived from the same dataset over a large spatial area (Paciorek et al., 2002). The NCEP system has been widely applied to establish trends and relationships between climate and various atmospheric and oceanic parameters in the Northern Hemisphere (Geng and Sugi, 2001; Gulev et al., 2001; Hoskins, and Hodges, 2002; Paciorek et al., 2002; Chan, 2005). Data is accessed via the NCAR reanalysis web page of NCEP at <http:// wesley.wwb.noaa.gov/Reanalysis.html>.

The storm tracks analysed in the current study were derived using NCEP/NCAR 6-hourly Northern Hemispheric Sea level Pressure (SLP) data, interpolated to a 250 x 250 km² lower resolution grid, and an algorithm developed by Serreze (1995) and Serreze et al. (1997). The algorithm essentially detects the storm tracks based on a local minimum threshold of sea level pressure, and tracks the event’s life cycle for the most recent 30
days. The total allowable distance a cyclone can move between 6 hourly time steps is restricted to 800 km (133 km/hr) to allow for "center jumps" to be tracked.

Each new storm track is ascribed a unique number that is maintained throughout the life history of the system from cyclogenesis to cyclolysis. The output variables include, the position of each cyclone, the cyclone number, the year, month day and hour of each observation, central pressure, the SLP at the cyclone center (a measure of intensity), pressure tendency (as determined between subsequent central pressure values) and, whether the system represents a cyclogenesis or cyclolysis event based on the first and last observations. All cyclone numbers are resets at 1 January 0000Z of each year.

6.4.2 Climate variability modes

Several published climatic variability indices are used in this study. The ALPI is a North Pacific regional index which indicates the relative intensity of the controlling pressure system in the NE Pacific and, therefore, indirectly describes regional storm track characteristics. The index is calculated as the mean area (km$^2$) with sea-level pressure less than or equal to 100.5 kPa expressed as an anomaly from the 1950-1997 mean (Beamish et al., 1997). Positive ALPI index values reflect a relatively strong or intense Aleutian Low while negative values indicate the opposite. ALPI values are available at <www.pac.dfo-mpo.gc.ca/sci/sa-mfpd/downloads/alpi.txt>.

ENSO events are defined by the MEI (Wolter and Timlin, 1993, 1998). MEI is defined as the weighted average of a number of tropical Pacific
environmental variables including sea surface temperature, east-west (i.e. zonal) and north-south (i.e. meridional) surface winds, sea-level pressure and temperature, and cloudiness. Negative MEI values represent the cold ENSO phase (La Niña), while positive MEI values represent the warm ENSO phase (El Niño). Monthly MEI values are published online at <www.cdc.noaa.gov/people/klaus.wolter/MEI/>.

The PDO index characterizes inter-annual variability in average North Pacific sea surface temperature and, as above, reflects NE Pacific regional climatic variability (Mantua et al., 1997). PDO is well correlated with many records of North Pacific and Pacific Northwest climate and ecology, including sea-level pressure, winter land–surface temperature and precipitation, and stream flow (Mantua et al., 1997; Zhang et al., 1997; Hessl et al., 2004; Schneider and Cornuelle, 2005). Predominantly positive or negative sea–surface temperature anomalies along the Pacific coast of North America characterise the phase of the PDO pattern as being in a "warm" or "cold" phase respectively. Monthly PDO index values are online at <www.jisao.washington.edu/pdo> (Mantua et al., 1997).

The Northern Oscillation Index (NOI) is the sea-level pressure anomaly between the North Pacific High (NPH, 350N and 1300W) region of the NE Pacific and the equatorial Pacific near Darwin, Australia, a climatologically low-pressure region (Schwing et al., 2002). NOI is dominated by the interannual variations of ENSO, such that large positive (negative) NOI are associated with La Niña (El Niño) (Schwing et al., 2002). As NOI is partially based in the NE Pacific, it provides a more direct connection between various climate processes in the eastern north Pacific and remotely teleconnected
global climate events. The NOI time series are available at <http://www.pfeg.noaa.gov>.

The Pacific/North American teleconnection (PNA) describes one of the most dominant modes of low frequency variability in the Northern Hemisphere extra-tropics during North Hemisphere winter (Van den Dool et al., 2000). The PNA index is constructed through a linear combination of normalized 700-mb height anomalies nearest to the nominal centers of the PNA pattern (Leathers and Palecki, 1992). The PNA pattern occurs as a result of an amplification or damping of the mean flow configuration over the unique geography of the Pacific Basin and North America (Leathers and Palecki, 1992). PNA is strongly influenced by the mode of ENSO where the positive phase of PNA tends to be associated with warm ENSO and the negative phase with cold ENSO.

6.5 Methodology

6.5.1 Selecting and characterizing storm tracks

All the storm tracks that lasted for at least 18-24 hours; travelled further than 200 km; with majority of their life cycle spent within Latitudes 30°N to 60°N and Longitude 120°W to 150°W considered for analysis. Figure (6.1) demonstrates the selected storm tracks for the month of January, 2004. Each vector represents a single storm track that intercepted the region in that year. Similar January storm track data base was developed for each year from 1948-2004. The selected storm tracks were characterised as:
(i) Deep and Weak Tracks: The tracks with central pressure < 990 hPa are characterized as “Deep” tracks while the tracks with central pressure > 990 hPa is characterized as “weak” tracks.

(ii) The Lifetime of Tracks: The lifetime of tracks is estimated as the number of 6 hourly steps between the cyclone generation and decay. Since the minimum temporal resolution is 6 hours, the estimated lifetime of a track can have a ±6 hour error margin.

(iii) Directional Tracks: The storm tracks were categorized based on their direction of propagation. Here the tracks with a net displacement towards northern latitudes (i.e. polarward) are designated as “southerly” tracks and those with a net southerly displacement (i.e. equator wards) are designated as “northerly” tracks. Accordingly, the Southerly tracks contain storm tracks that propagate from SE, SW directions, while the northerly tracks are those that propagate from NE, NW directions.

(iv) Latitudinal variability of the cyclogenesis: Here the storm tracks are characterized based on the latitudinal variability of the cyclogenesis.

6.5.2 Statistical methods

Significant long-term trends of inter-annual January storm track characteristics were identified via the simple linear regression method. The statistical significance of the linear trend was assessed based on the t-test statistic and 95% confidence intervals. The strength of dependence was assessed via the Pearson’s product-moment correlation coefficients.
Cumulative Sum Analysis (CumSum) was used to compare and illustrate the
decadal scale temporal trends in the storm track characteristics. The
approach involved simple addition of data points in a time series to the sum of
all previous points. This methodology allows for examination of long period
(decadal scale) trends in the time series (Murdock, 1979). In addition, trends
can also be identified by subtracting the mean of the entire time series from
each data point prior to addition. Thus, a trend indicated in a CumSum graph
usually corresponds to similar trend in the dataset. For example, a positive
(negative) slope in a CumSum curve corresponds to an increasing
(decreasing) trend in the time series, while the year when the slope changes
define a regime shift. This technique has been applied to study the sea level
variability and the responses of fish populations to climatic variability signals
and regime shifts (cf. Beamish et al., 1999; Abeysirigunawardena and Walker,
2008 (Chapter 2)).

Superposed Epoch Analysis (SEA), was applied to identify current year
and lagged nonlinear relationships between climatic variability and the storm
track characteristics (i.e. Shift of Cyclogenesis latitude, Track Lifespan,
number of Deep and Weak tracks). SEA, also known as “compositing”, is a
classification technique that allows for testing of an association between a
response and an explanatory variable at their extremes (Prager and Hoenig,
1989). SEA has been widely applied to describe the linkages between
precipitation, drought, stream flow, sea levels and climatic variability patterns
(e.g., Kadioğlu et al., 1999; Kahaya and Karabörk, 2001; Hessl et al., 2004;
Abeysirigunawardena and Walker, 2008 (Chapter 2)). In the SEA method,
first, significant “event years” are identified as those with storm track
characteristic anomalies exceeding one standard deviation (>1 STDV) from the overall mean (i.e. from 1948 to 2004). For each climatic variability index (MEI, NOI, PDO, PNA AND ALPI), 5-yr epochs were formulated around “event years” (i.e., 2 years (January) preceding, 1 during the event year, and 2 years (January’s) following). The five year epoch window was chosen to yield a sufficient number of events for evaluation of inter-annual linkages between climatic variability indices and storm track characteristics. Finally, the significance of each climatic variability index across the 5 year epoch was evaluated by comparing the overall mean climatic variability values during the event epoch, against the complete climatic variability time series, using a randomized Monte Carlo simulation method. This technique randomly picks years, identifies 5-yr windows, calculates expected means, and provides 95% confidence intervals. The computer program “EVENT”, developed as a part of FHX2 fire history program (cf. Grissino-Mayer and Swetnam, 2000), was used to perform the Monte Carlo simulations. The same approach was used to test the association between lower than average storm track characteristics (i.e., MSL < 1STDV) and the climatic variability indices.

6.6 Results

6.6.1 Interannual to Decadal changes in the Storm track characteristics.

6.6.1.1 Storm track Lifespan.

The average life time of the NE Pacific January storm tracks is about 2-3 days, with just over 10% having a lifetime longer than 5-days (i.e. long-lived). Based on directionality, the southerly tracks are comparatively long lived (2.8 days) than the northerly tracks (1.9 days), with an average lifespan
almost 50% higher than the northerly tracks. The overall storm track lifespan anomaly reveals three significant positive peaks in mid 1950s, 1960s and 1990s (Figure 6.2). A general decreasing trend in the storm track life span from the mid 1970s to the late 1980s shows the absence of long-lived tracks in the NE Pacific during this period. The opposite relationship between the 5-year running means of the directional (i.e. northerly and southerly) storm track life span anomalies in Figure 6.3 indicates that the external oceanic and (or) atmospheric phenomena causing the pronounced longevity in the southerly tracks are at the same time responsible for reducing the average lifespan of the northerly tracks. This opposing relationship has apparently become more prominent after the climate regime shift of mid 1970s.

6.6.1.2 Storm track count

52% of the January storm tracks from 1948-2004 were deep events with central pressures less than 990 hPa. However, the interannual variations of the January deep and weak storm track count anomalies from the mid-1960s to early-1980s indicate a relatively calmer period with a predominant positive anomaly in the “weak storm track count” compared to its opposite counter part (Figures 6.4a and 6.4b). Since the early 1980s, however, the deep storms tracks have become more frequent, suggesting a relatively intense stormy period since then.

Secular changes in the overall January storm track count indicate a weak increasing trend from 1948-2004 (Figure 6.5). In contrast, the trends became more distinct and apparent when storm track directionality was accounted for (Figure 6.5). For instance, the frequency of occurrence of
southerly storm tracks (i.e. track from the South (SW and SE)) were clearly in opposition to the northerly tracks (i.e. tracks from the North (NW & NE)), where the former indicated an upward trend of about 8-9 tracks per decade, while the latter demonstrated a downward trend of about 5-6 tracks per decade (Figure 6.5).

6.6.1.3 Storm track Central Pressure Variability (CP): (Deep vs Weak Track Count).

Statistically significant positive trends were evident in the overall and southerly deep track counts (i.e. intense events or events with CP< 990 hPa), while the northerly deep track counts indicated a statistically non-significant negative trend (Figure 6.6). On the other hand, the overall weak track count (i.e. events with CP > 990 hPa) indicated a statistically significant negative trend, and a statistically non-significant negative trends for both southerly and northerly weak track count (Figure 6.7). In essence, these trends suggest an apparent increase in the deep track count in the north eastern Pacific from 1948-2004, with the most significant contributions coming from the southerly tracks. These results are in agreement with the findings of Gulev et al., (2001) and Sickmoeller et al., (2000).

6.6.1.4 Latitudinal displacement of the Storm tracks.

The North-eastern Pacific winter storm track vectors generally transit between latitudes 30°N-50°N, centered on 42°N (Weijun, and Zhaoboo, 1999; Salathé, 2006). However, global climate model (GCM) simulations that contributed to the IPCC Fourth Assessment Report, projected possible poleward shifts of the North Pacific mid-latitude storm tracks for the 21st century under climate change (Yin, 2005; Salathé, 2006). Evidence has also
been presented for more intense and poleward northern hemisphere storm tracks during the 20th century (McCabe et al., 2001). The inter-annual mean latitudinal variation of the cyclogenesis and the cyclolysis points of the current analysis closely follow the range indicated by Salathé (2006) (Figure 6.8). In addition the apparent long-term negative trend in the mean overall cyclogenesis latitude and a slightly positive trend in the mean overall cyclolysis latitude suggest a gradual southward shift of the storm track initiation point, and a northerly shift of the track termination point over the period considered (Figure 6.8). The respective opposite trends of the mean cyclogenesis and the cyclolysis latitudes also imply a gradual widening of the storm track span over the 56 years examined. Figure 6.8 also shows strong decadal trends with a significant southerly shift in cyclogenesis latitude during 1980s and 1990s and a clear northward shift of the cyclolysis latitude during early-to mid-1970s. The fact that the overall mean cyclogenesis latitude is consistently lower than the cyclolysis latitude over the 56 year period suggests a dominant presence of southerly tracks (i.e. northward propagating tracks) in the NE Pacific.

Cumulative sum analysis technique (CUMSUM) was applied to further elucidate the impact of variable track positions on storm track characteristics. To this end, Figures 6.9 and 6.10 present the respective CUMSUM curves of the storm track life span, and the deep and weak storm track count anomalies for the 56 years considered. Common to both are the CUMSUM curves of the mean cyclogenesis and cyclolysis latitude anomalies, as well as the 5-year running mean for each variable. Figure 6.9 shows a decadal cycle with a marked northward shift in the cyclogenesis point and a corresponding
southward movement of the cyclolysis point from mid 1970s to early 1980s. This is suggestive of not only the prevalence of narrow storm tracks, but also tracks with shorter lifespan. The same trend is also shown by the negative shape of these two variables from mid-1970s to early-1980s in Figure 6.9. The concurrent increase in weak track count anomaly during the same period suggests that these narrow tracks may be weaker in strength (Figure 6.10).

In contrast, a positive trend was evident in the storm track lifespan and the deep track count anomaly from early 1990s to date, suggesting an apparent increase in the storm track strength since 1990s (Figures 6.9 and 6.10). This observation is in agreement with Salathé (2006), suggesting an increase in track lifetime since 1990s. It is also interesting to observe the concurrent negative and positive trends of the mean (CG) and (CC) latitudinal anomalies during the same period, which suggest an overall widening of the storm tracks. These results are reflective of the observed increase in long lasting storms in the region since 1990.

In conclusion, the results depict a gradually narrowing (northward movement of CG and a southward movement of the CC) of storm track vectors up until mid-1970s and subsequent widening of the track vectors (southward movement of CG and a northward movement of CC) since then. However there is a notable lead/lag correspondence between the year at which the trend transition for CC latitude (from negative to positive) and the CG latitude (from positive to negative) took place. Here the southerly shift of the CG latitude leads the northerly movement of the CC latitude by about 7-10 years; likely an indication that the latter movement may have been triggered as a feedback effects of the former process, which needs further research for
verification. A systematic shift in storm track characteristics as presented above, naturally suggests a similar shift in the band of intense precipitation over the study region, since peak rainfall generally occurs along the southern margin of a storm track (Salathé, 2006). Another significant impact of such a displacement would be a change in the mean pressure field off the coast, which controls a variety of climate impacts including various coastal ocean processes such as the coastal storm surge climatology in the region (Abeysirigunawardena et. al, (in review),(Chapter 4)).

6.6.2 Storm track characteristics and Climate Variability.

A key objective of this study was to explore the relationships between known climatic variability modes and, storm track characteristics in the north-eastern Pacific (e.g., inter annual patterns, regime shifts). Even though the “Total Storm Track Count” did not show significant correlations (p< 0.05) with the climatic variability indices considered (Table 6.1), statistically significant relationships between climatic variability and “Storm Track Count” became apparent when the storm track directionality was considered (Tables 6.2 and 6.3). Accordingly, all climatic variability indices considered were significantly correlated to the Southerly Storm Track Count (TOTALS), with PNA (0.618) and ALPI (0.614) indicating the highest positive correlations, followed by PDO (0.551) and MEI (0.438) (Table 6.2). Because of this relationship, the southerly storm track count is expected to increase substantially under the positive phase of MEI and PDO (i.e. higher than normal Pacific SST), and, when the Aleutian low is significantly strong. Significant negative correlations were seen between the Northerly track count (TOTALN) and climatic
variability, with the strongest being the PNA (-0.695) and PDO (-0.529) followed by ALPI (-0.518) and MEI (-0.342). These suggest a substantial increase in the northerly track count when the Pacific SST is below normal and when the Aleutian low is significantly weak.

Table 6.4 demonstrates the linear correspondence between climatic variability and the “Extreme storm track count”, where the “Extreme storm tracks” are defined as a track that lasted for more than 5-days. Accordingly, NOI (-0.415) shows the strongest relationship with the extreme storm track count, followed by MEI (0.278). This suggests that the storms are likely to last longer when NOI is in a negative phase. The remainder of the climatic variability indices considered did not show any significant correspondence with the extreme storm track count.

The results of the SEA did not show significant relationships between high (anomaly $\geq 1$ STDV) and low (anomaly $\leq 1$ STDV) “Total storm track count” event years (at lag = 0) and various climatic variability modes (Figure 6.11). However, significant relationships between the storm track event count and climate variability became apparent when storm track directionality was considered (Figure 6.12 & 6.13). For instance, the results shown in Figure 6.12(a) suggest strong positive modes of ALPI, PNA and, strong negative mode of NOI supporting high southerly storm track counts event years (i.e. all associations at the event year (0 lag) are significant at the 99% confidence threshold). Notably, the above three climate variability modes represent dominant atmospheric controls in the NE Pacific. Figure 6.12(b) suggests that southerly storm track frequency could be significantly low (significant, at 99% level) when PNA is at its strongest (significant at the 99% confidence
threshold) negative phase, while MEI and NOI seems to have no significant control in reducing the southerly storm track count.

The relationship between climate variability and the Northerly storm track count indicated that strong negative phases of PNA, PDO and ALPI were associated with higher than normal northerly event count years (0 lag), while positive phase of ALPI, PNA, PDO and MEI were associated with lower than normal northerly event count years (0 lag) (Figure 6.13). According to this analysis, the strongest control on both northerly (higher as well as lower) and southerly (higher as well as lower) storm track count were PNA and ALPI, where both are regionally specific climate modes. Thus it is speculated that, the ability to successfully project the characteristics of PNA and ALPI years in advance may facilitate successful forecasting of the response of storm track characteristics to climate variability in coastal BC (e.g., inter annual patterns, regime shifts).

Figures 6.14 and 6.15 demonstrate the association between climatic variability and the latitudinal storm track position. Here a high latitude event years corresponds to years with significant northward shifts (anomaly ≥ 1 STDV) either in the mean cyclone genesis latitude (Figure 6.14(a)) and/or the cyclolysis latitude (Figure 6.15(a)), where as the low latitude event years corresponds to years with significant southerly shift (anomaly ≤ 1 STDV) of the same (Figure 6.14(b) and Figure 6.15(b)). Results show cyclone genesis latitude shifting northward from its norms (anomaly ≥ 1 STDV) under strong negative PDO, PNA and positive NOI phases (Figure 6.14(a)). The most significant relationship is indicated with the negative PDO phase (relationship significant at the 99% confidence threshold). On the other hand, the cyclone
genesis latitude shift southward from its norms (anomaly ≤ 1 STDV) when MEI, ALPI, PNA are at its strong positive modes and NOI is at its strong negative phase. Here the strongest relationship was seen with ALPI and NOI (relationship significant at the 99% confidence threshold). The results do not suggest any statistically significant climate controls exist over the cyclolysis latitudinal positions (Figure 6.15).

Finally, Figure 6.16 demonstrates the associations between climatic variability and the storm track lifespan, where high event years corresponds to years with longer than normal average storm track lifespan (anomaly≥1 STDV) and vice versa. The results suggest that only MEI exerts a strong influence on the January storm track life span in the region, with significantly positive MEI (significant at 95% confidence level) modes resulting in long-lasting tracks (Figure 6.16(a)). Significantly negative MEI modes cause the tracks to have a shorter duration than normal (Figure 6.16(b)).

6.6.3 Pineapple-Express Storms and Climate Variability.

Dettinger (2004) presented a study on the “Pineapple Express storm” events, using the NCEP/NCAR Reanalysis daily water-vapour transport pathways. The study indicates 206 days of Pineapple Express events in the Pacific Northwest from 1948 to 1999 (approximately 4-events per year). These events were distributed between the months of October and April. Most events occur during January and February.

Superposed Epoch Analysis (SEA) was applied on the Pineapple Express storm characteristics extracted from Dettinger, (2004). The purpose was to identify current year and, lagged nonlinear relationships between
climate variability and Pineapple-Express storm characteristics such as: (i) the storm count (CO); (ii) southerly latitudinal limit, defined as the southernmost latitude that connects with the south-westerly back trajectory or 20ºN (whichever was reached first) (SL); (iii) west coast crossing latitude, defined as the latitude at which the trajectory with largest transport rates crosses 120ºW (WL); and, (iv) path length defined as the number of 2.5º grid cells traversed along the back trajectory to reach either southern terminus or 20ºN (whichever was reached first) (PL). The significant event years in the SEA analysis were defined as those with Pineapples Express storm track characteristic anomalies that were higher (or lower) than one standard deviation (≥1 STDV ( or ≤ 1 STDV)) from the overall mean.

Figure 6.17 illustrates the relationship between the Pineapple Express storm track count anomaly and various climate variability indices. As seen, even though MEI and PDO, remains consistently positive (negative) during high (low) event count years (i.e. year 0), the anomalies are not statistically significant. This shows that, positive but near neutral ENSO phase is likely to prevail during years with higher than average Pineapple Express event count years and vice versa. The most significant relationships exist between the above average Pineapple Express event count years and positive ALPI mode and negative NOI mode, signifying lower than average pressures conditions along the BC coast favouring Pineapple Express storm conditions.

As shown, the longer (shorter) path lengths are supported by near neutral but negative (positive) ENSO mode (Figure 6.18). As a consequence, the storm tracks may weaken due to the extra distance travelled. Thus, it is more likely for a Pineapple Express storm to be weaker (stronger) when
ENSO is in its negative (positive) mode. Figure 6.19 demonstrates the relationship between climatic variability and southern latitude limit of the Pineapple Express storm tracks. Here a positive anomaly of the southern latitude limit implies more northerly connection of the south-westerly back trajectory and vice versa. From the results it is evident that the southerly limit of the Pineapple Express storm tracks are likely to shift northward (i.e. located north of the mean southerly limit) and remain narrow and strong, when PDO and ALPI are significantly positive (i.e. statistically significant at 95% level). This in combination with Figure 6.20, which demonstrates the link between climatic variability and the track’s west coast crossing latitude anomaly suggests that the largest transport rates of the Pineapple Express storm tracks are more likely to intercept the west cost of North America at a higher latitude when PDO and ALPI are significantly positive (i.e. when above average Sea Surface Temperatures and below average Aleutian Pressures prevailing in the Pacific), possibly causing higher impacts on the BC coast.

6.7 Discussion & Conclusions.

The inter annual, decadal trends and nonlinear relationships between climate variability and the North-eastern Pacific winter storm track characteristics were simulated using the NCEP–NCAR North-eastern Pacific January storm track projections. The approach presented here differs from many, by the fact that it introduces the storm track directionality in to the analysis. Interestingly, when directionality is accounted for, the results indicated significant trends and relationships between climate variability and storm tracks that were otherwise not apparent (i.e. statistically not significant).
As noted this study was based only on January storm track characteristics. While the chances are highly likely that these interpretations provide an analog to winter time storm characteristics in the region, continuing this analysis for the entire winter season (DJF) is recommended prior to using the information for practical applications.

The results indicated significant opposite correspondence between northerly and southerly storm track characteristics, with southerly (northerly) tracks becoming stronger (weaker) during the warm ENSO states. The tracks also indicated a net equatorward shift during El Niño years, likely in response to local enhancement of the Hadley circulation over the eastern Pacific (Trenberth and Hurrell, 1994; Straus, and Shukla, 1997), while La Niña events were marked with opposite shifts. This shows that warmer than normal eastern Pacific favour southerly tracks with more southerly cyclone genesis latitude and long life spans. Colder than normal eastern Pacific favour more northerly cyclone genesis latitude and shorter storm track lifespan.

The fact that southerly tracks shifts northward (closer to the BC coastal stretch) and, northerly tracks strengthen when the eastern Pacific is under a cold phase (La Niña like regime) increases the possibility of short-lived but high wind-storms impacting the BC coast (Lat 48°–52° N ) during a cold climate regime. The storm tracks corresponding to warm ENSOs are likely to generate much weaker winds on the BC coast than their cold counterpart due to longer than normal excursions they traverse from their point of origin in order to reach the BC coast (i.e. more southerly cyclone genesis latitude). This finding is in agreement with those of Abeysirigunawardena et al. (2009), (Chapter 5), who report stronger winds in southern BC during cold ENSO
years. However, the results of this study also suggest a significant increase in the storm track life span during warm ENSOs, which in turn could increase the probability of storm-high-tide interactions substantially. This may (likely) be the reason for severe damages from storm surges in coastal BC during warm ENSO periods, even when the winds associated with the storm tracks are at moderate strengths.

Geng and Sugi (2001), Graham and Diaz (2001), McCabe et al. (2001) and Paciorek et al. (2002), have shown that the frequency and intensity of extreme storms over the Pacific basins has increased over the second half of the twentieth century. However, Chang and Fu (2002) found a marked inter-decadal variability in the intensity of storm tracks where the tracks became 40% stronger during the 1990s than during 1960s. The results of the current study are in general agreement with these previous findings but specifically show that the winter storm track intensity in the North-eastern Pacific region decreased until the mid-1970s, but have intensified since then. This behaviour is likely a response to the climate regime change in mid 1970s over the Pacific basin (Hurrell, 1995; Mantua et al., 1997; Deser et al., 1999; Abeysirigunawardena and Walker, 2008 (Chapter 2)). This study also suggests significant influences of regional climatic variability modes (i.e. PNA and ALPI) on the NE Pacific winter storm track characteristics, which show the need to account for the impacts of natural climate variability on storm track characteristics in any regional planning and adaptation measures and, avoid assuming that the changes will be monotonic or linear. For instance, in preparing measures to manage the impacts of increased storminess in north-eastern Pacific, it is important to note that a decrease in storminess could
occur over the coming decades while still being consistent with a longer term increase (or decrease) in storminess due to anthropogenic climate change.

Detecting the climate change impacts on storm track characteristics is a complex issue due to the noise introduced by the natural climate variability. The current study suggests that this may become more prominent on regional scale, making it even more difficult to distinguish the climate change signals at that scale. For instance, a certain phase of natural climate variability can superimpose itself over the long-term climate signals leading to periods with decreasing storminess despite a long term increasing trend. Even the IPCC TAR admits the presence of significant trends towards one particular phase of natural climate oscillations during the last 30 years (page 153). Consequently it is not quite clear whether the recent changes in the storm track characteristics are signals of anthropogenic climate change or simply occurring by natural variations alone. It is expected that implementing continuous and reliable monitoring programs of atmospheric and oceanic parameters would eventually provide a better understanding on these differences.

Although the data assimilation system of the NCEP/NCAR reanalysis is fixed, assimilation inputs have changed during the last four decades. As a result, climate impact researchers have a mixed view on the reliability of the reanalysis datasets. For instance, some studies report that the storm track intensities in the reanalysis dataset are biased low over data spares regions such as the eastern Pacific, before early 1970s (Chang and Fu, 2002; Harnik and Chang, 2003). Nevertheless, other authors who examined radiosonde observations along the storm tracks found that the observations are largely
consistent with the NCEP/NCAR re-analysis projections, with the exception of over Japan (Chang, and Fu, 2002; Harnik and Chang, 2003).

The Northern Hemispheric NCEP/NCAR reanalysis projections are believed to be reliable even in 1950s, because of the models proven skill in re-forecasting 4-days in advance the first “storm of the century” of November 1950 and the North Sea gale of February 1953 in Europe (Smith 1950; Phillips 1958; Salathé, 2006). In addition, the NCEP-NCAR re-analysis data is also in fairly good correspondence to the “climate shift in the mid-to-late 1970s” as noted in the literature (Trenberth and Hurrell, 1994; Mantua et al., 1997, Zhang et al., 1997). Both the mid-winter suppression in north Pacific storm track activity, as well as the more recent finding that the Pacific storm track is significantly stronger during the late 1980s and early 1990s than during the early 1980s, are also confirmed by NCEP reanalysis, but, with a slightly stronger intensification than the actual observations (Chang, 2003; Harnik and Chang, 2003). The predictions of interannual variability in the NCEP/NCAR reanalysis are also found to be well correlated with independent observations (Kistler et al., 2001). However, in view of the uncertainties, it is vital to recognise that, storm track characteristics assessed from the NCEP/NCAR re-analysis may be influenced by time dependent biases inherent in numerical weather prediction products. Thus it is recommended additional efforts be made to better determine the variability of the storm tracks (location and intensity) over the oceanic regions and, to see whether the trends observed in NCEP/NCAR reanalysis data are consistent with other in-situ observations as well(i.e. waves, water levels, winds and weather maps etc.).
References


Table 6-1: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA and PDO) vs Total January Storm Track (TJST) count in the North-eastern Pacific (1948-2004).

<table>
<thead>
<tr>
<th></th>
<th>ALPI (YEAR)</th>
<th>MEI (JAN)</th>
<th>NOI (JAN)</th>
<th>PNA (JAN)</th>
<th>PDO (JAN)</th>
<th>TJST</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pearson's r</td>
<td>.436**</td>
<td>-.447**</td>
<td>.711**</td>
<td>.605**</td>
<td>.104</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.001</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td>.442</td>
<td></td>
</tr>
<tr>
<td>N</td>
<td>55</td>
<td>57</td>
<td>55</td>
<td>57</td>
<td>57</td>
<td>57</td>
</tr>
<tr>
<td>Pearson's r</td>
<td>-.661**</td>
<td>.344*</td>
<td>.437**</td>
<td>.046</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.000</td>
<td>.010</td>
<td>.001</td>
<td>.741</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
</tr>
<tr>
<td>Pearson's r</td>
<td>-.271*</td>
<td>-.304*</td>
<td></td>
<td>.322</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.045</td>
<td>.022</td>
<td>.322</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>N</td>
<td>55</td>
<td>55</td>
<td>57</td>
<td>57</td>
<td>55</td>
<td>55</td>
</tr>
<tr>
<td>Pearson's r</td>
<td>.736**</td>
<td></td>
<td>.100</td>
<td></td>
<td>.017</td>
<td></td>
</tr>
<tr>
<td>Significance</td>
<td>.000</td>
<td>.466</td>
<td>.903</td>
<td></td>
<td></td>
<td>57</td>
</tr>
<tr>
<td>N</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>57</td>
</tr>
</tbody>
</table>

** Correlation is significant at the 0.01 level (2-tailed)

* Correlation is significant at the 0.05 level (2-tailed)
Table 6-2: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA, and PDO) vs. Total January Southerly (TJSST) Storm Track count in the north eastern Pacific (1948-2004).

<table>
<thead>
<tr>
<th></th>
<th>ALPI (YR)</th>
<th>MEI (JAN)</th>
<th>NOI (JAN)</th>
<th>PNA (JAN)</th>
<th>PDO (JAN)</th>
<th>TJSST</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pearson’s r</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Significance</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>N</strong></td>
<td>55</td>
<td>57</td>
<td>55</td>
<td>57</td>
<td>57</td>
<td>57</td>
</tr>
</tbody>
</table>

- ALPI (YEAR)
- MEI (JAN)
- NOI (JAN)
- PNA (JAN)
- PDO (JAN)
- TJSST
Table 6-3: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA, and PDO) vs Total January Northerly (TJNST) Storm Track count in the north eastern Pacific (1948-2004).

<table>
<thead>
<tr>
<th></th>
<th>ALPI (YEAR)</th>
<th>MEI (JAN)</th>
<th>NOI (JAN)</th>
<th>PNA (JAN)</th>
<th>PDO (JAN)</th>
<th>TJNST</th>
</tr>
</thead>
<tbody>
<tr>
<td>ALPI (YEAR)</td>
<td>Pearson’s r</td>
<td>.436**</td>
<td>-.447**</td>
<td>.711**</td>
<td>.605**</td>
<td>-.518**</td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.001</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>57</td>
<td>55</td>
<td>57</td>
<td>57</td>
</tr>
<tr>
<td>MEI (JAN)</td>
<td>Pearson’s r</td>
<td>-.661**</td>
<td>.344*</td>
<td>.437**</td>
<td>-.342*</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.000</td>
<td>.010</td>
<td>.001</td>
<td>.011</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
</tr>
<tr>
<td>NOI (JAN)</td>
<td>Pearson’s r</td>
<td>-.271*</td>
<td>-.304*</td>
<td>.164**</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.045</td>
<td>.022</td>
<td>.005</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>55</td>
<td>57</td>
<td>57</td>
<td>57</td>
</tr>
<tr>
<td>PNA (JAN)</td>
<td>Pearson’s r</td>
<td>.736**</td>
<td>-.695**</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.000</td>
<td>.000</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>55</td>
<td></td>
<td>57</td>
<td></td>
</tr>
<tr>
<td>PDO (JAN)</td>
<td>Pearson’s r</td>
<td>-.529**</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td></td>
<td></td>
<td></td>
<td>57</td>
<td></td>
</tr>
<tr>
<td>TJNST</td>
<td>Pearson’s r</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 6-4: Pearson’s correlation coefficients for relations between climate variability (ALPI, MEI, NOI, PNA, and PDO) vs Total Storm Tracks with Lifespan greater than 5-days (HECOUNT) in the north eastern Pacific (1948-2004).

<table>
<thead>
<tr>
<th></th>
<th>ALPI (YEAR)</th>
<th>MEI (JAN)</th>
<th>NOI (JAN)</th>
<th>PNA (JAN)</th>
<th>PDO (JAN)</th>
<th>HECOUNT</th>
</tr>
</thead>
<tbody>
<tr>
<td>ALPI (YEAR)</td>
<td>Pearson’s r</td>
<td>.436**</td>
<td>-.447**</td>
<td>.711**</td>
<td>.605**</td>
<td>.120</td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.001</td>
<td>.000</td>
<td>.000</td>
<td>.000</td>
<td>.374</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>57</td>
<td>55</td>
<td>57</td>
<td>57</td>
</tr>
<tr>
<td>MEI (JAN)</td>
<td>Pearson’s r</td>
<td>-.661**</td>
<td>.344*</td>
<td>.437**</td>
<td>.278*</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.000</td>
<td>.010</td>
<td>.001</td>
<td>.040</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>55</td>
<td>57</td>
</tr>
<tr>
<td>NOI (JAN)</td>
<td>Pearson’s r</td>
<td>-.271*</td>
<td>-.304*</td>
<td>-.415**</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.045</td>
<td>.022</td>
<td>.001</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>57</td>
<td>57</td>
<td></td>
<td></td>
</tr>
<tr>
<td>PNA (JAN)</td>
<td>Pearson’s r</td>
<td>.736**</td>
<td>-.102</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td>.000</td>
<td>.460</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>55</td>
<td>55</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PDO (JAN)</td>
<td>Pearson’s r</td>
<td>-.021</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td></td>
<td>.876</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td></td>
<td>57</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HECOUNT</td>
<td>Pearson’s r</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>57</td>
</tr>
</tbody>
</table>
Figure 6-1: The shaded area indicates the study region of interest in the NE Pacific (Latitude: 30°N - 60°N; Longitude: 120°W - 150°W). Each line in the figure represents a single storm track that crossed the region during January 2004.
Figure 6-2: Interannual Variability of the overall January storm track Lifespan anomaly in the north eastern Pacific from 1948-2004.
Figure 6-3: The 5-year running mean of the northerly (continuous line) and southerly (dashed line) January storm track life span anomaly in the north eastern Pacific from 1948-2004.
Figure 6-4: Interannual Variability of the January (a) Deep and (b) Weak storm track count anomaly in the north eastern Pacific from 1948-2004.
Figure 6-5: Secular trends in the January storm track count in the northeastern Pacific from 1948-2004. (i) Continuous black line; Total track count (ii) Long dashed line; Northerly track count (iii) Short dashed line: Southerly track count.
Figure 6-6: Secular trends in the January deep storm track count in the north-eastern Pacific from 1948-2004. (i) Continuous black line; Total deep track count (ii) Long dashed line; Northerly deep track count (iii) Short dashed line: Southerly deep track count.
Figure 6-7: Secular trends in the January weak storm track count in the north-eastern Pacific from 1948-2004. (i) Continuous black line; Total weak track count (ii) Long dashed line; Northerly weak track count (iii) Short dashed line: Southerly weak track count.
Figure 6-8: Secular trends in the mean latitudinal position of the cyclogenesis and the cyclolysis of the overall January storm tracks in the north-eastern Pacific from 1948-2004. (i) Continuous black line; mean cyclogenesis Latitude;(ii) Dashed line: mean cyclolysis Latitude.
Figure 6-9: The cumulative sum (CUMSUM) curves for (i) Continuous black line; mean latitudinal position of the cyclogenesis; (ii) Short Dashed line: mean latitudinal position of the cyclolysis (iii) Long Dashed line: Storm track lifespan of the overall January storm tracks in the north-eastern Pacific from 1948-2004.
Figure 6-10: The cumulative sum (CUMSUM) curves for (i) Continuous grey line: mean latitudinal position of the cyclogenesis; (ii) Light dashed line: mean latitudinal position of the cyclolysis (iii) Dark dashed line: Deep storm track count; (iv) Continuous black line: Weak storm track count, of the overall January storm tracks in the north-eastern Pacific from 1948-2004.
Figure 6-11: SEA results showing the departure of climate variability indices from their annual mean during and around the event years (lag = 0). The event years are defined as years with “Total storm track count” anomaly is (a) higher (≥ 1 STDV) and, (b) lower (≤ 1 STDV) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-12: SEA results showing the departure of climate variability indices from their annual mean during and around the event years (lag = 0). The event years are defined as years with the average Southerly (SW & SE) storm track count anomaly is, (a) higher (≥ 1 STDV), and (b) lower (≤1 STDV) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-13: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Northerly (NW & NE) storm track count anomaly is, (a) higher ($\geq 1$ STDV), and (b) lower ($\leq 1$ STDV) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-14: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the cyclone genesis latitude anomaly is (a) higher (≥ 1 STDV) (northerly shift), and (b) lower (≤ 1 STDV) (southerly shift) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-15: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the cycloysis Latitude anomaly is, (a) higher (≥ 1 STDV) (northerly shift), and (b) lower (≤ 1 STDV) (southerly shift) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively, derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-16: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Storm track lifespan anomaly is, (a) higher (≥ 1 STDV), and (b) lower (≤ 1 STDV) than the (1948-2004) overall mean. The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-17: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track count anomaly is, (a) higher (≥ 1 STDV), and (b) lower (≤ 1 STDV) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-18: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track path-length anomaly, (a) higher (≥ 1 STDV) and, (b) lower (≤1 STDV) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-19: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track southern-limit (Latitude) anomaly is, (a) higher (≥ 1 STDV) and, (b) lower (≤ 1 STDV) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
Figure 6-20: SEA results showing the departure of climate variability indices from their annual mean around the event years (lag = 0). The event years are defined as years with the average Pineapples express storm track west coast-crossing (Latitude) anomaly is, (a) higher (≥ 1 STDV), and (b) lower (≤1 STDV) than the (1948-1998) overall mean (Dettinger 2004). The assessment is constrained to two years prior (lag -2 and -1) and following (lag 1 and 2) the event year. The horizontal solid and dashed lines are the 99% and 95% confidence intervals respectively derived from 1000 Monte Carlo simulations performed on the entire climate series.
7.0 Conclusions

7.1 Introduction

The goal of the research presented in this dissertation was to determine whether climate variability is critical to understanding the documented changes in coastal environmental variables in BC. To that end, a number of statistical models were proposed to clarify the relationships between five climate variability indices representing large-scale atmospheric circulation regimes and sea levels, storm surges, extreme winds and storm track variability in coastal BC. By clarifying the concepts in statistical terms, I attempted to provoke a different way of thinking about climate impacts on extremes in a region, where similar analyses have not been undertaken. Overall, the study showed that many of the impacts of climate variability and change on coastal environmental variables are readily detectable over time, through deliberate observation of current events and by data mining of past records.

The research program employed a step-by-step approach, whereby each finding stimulated a new line of investigation and information gathering. Each step was designed to investigate the climate-driven response of a coastal environmental variable that was assumed to have either a direct or an indirect impact on regional sea level. The research involved five distinct components: The establishment of:

i. sea-level response trends to long-term climate change and short-term climate variability in northern BC;

ii. an extreme sea level recurrence curve for the south coast of BC, with climate considerations;
iii. an understanding of the spatial and temporal distribution of extreme sea surge recurrence, and it’s sensitivity to regionally dominant climate variability patterns in coastal BC;

iv. an understanding the extreme wind recurrence and it’s sensitivity to climate variability for the southern coast of BC; and,

v. an investigation of winter storm track characteristics and their relationship to various climate variability signals in coastal BC.

The outcomes of the study indicate there is a need to look beyond climatic means when completing climate impact assessments. In addition, my research demonstrated that climate extremes are currently causing the majority of weather-related damage along coastal BC.

The lack of data availability at a much finer resolution (local scale) in most parts of coastal BC prevented this research from assessing the geomorphologic impacts on the BC coast due to changing climate. This is the same scale at which the dynamics of local wave regimes and their impacts become important. However, I anticipate that the results of my research will eventually provide boundary conditions (i.e. extreme water levels, winds, storm surges and storm tracks etc.) for future local scale studies that will be completed when sufficient data becomes available.

Thus far, my research has contributed substantially to the next level of climate impact research in BC, where investigations on how environmental parameters such as waves and coastal geomorphology would respond to climate change and variability will be assessed at local scale. The results at local scale are crucial for providing answers to questions such as: (i) which coastal systems, populations and infrastructure will be affected by climate
variability and change; (ii) what specific coastal impacts are of concern in BC; (iii) are all BC coastal communities equally affected; and, (iv) how much of the BC population resides in at risk areas. Investigations at local scale will have to be conducted via mathematical modeling, as the phenomena that governs the dynamics of a coastal system at this scale is quite complex.

7.2 Major research findings and contributions.

The research findings reported in this dissertation demonstrate that decadal to inter-decadal climatic variability is fundamental to explaining the changing frequency and intensity of extreme atmospheric and oceanic environmental variables in coastal BC. The trends revealed by these analyses suggest that coastal flooding risks are certain to increase in this region during the next few decades, especially if the global sea-levels continue to rise as predicted.

A major contribution of my research is the innovative application of a number of statistical approaches to estimate the potential changes of extremes due to natural climate variability. These applications required modification of a widely applied conventional extreme value statistical technique based on the assumption of a stable climate, to a more physically meaningful approach that incorporates the effects of local climate trends and cycles as covariates. Using the observations of historical changes in extreme sea levels, storm surges and winds in coastal BC, I was able to demonstrate that the changes in the distributional tails of coastal environmental variables may not occur in proportion to changes in the mean. This discovery highlighted that simply changing the mean variable may not be statistically
defensible for some of the extreme environmental variables considered, especially where the role of variability is critical.

In my research I demonstrated with examples that extreme events need to be addressed more consistently in climate change impact assessments. Based on the historic data examined, I demonstrated that historic coastal inundation events were a product of either a moderate high tide plus a “very high” residual water level, or were a consequence of an exceptionally high tide plus a “very low” residual water level. The timing of such an event depends as much on the random confluence of events, as on background changes in sea level driven by anthropogenic global climate change. The following sections of this chapter summarise the major outcomes of the dissertation and summarize the findings described in the previous chapters.

7.2.1 Sea-level response trends to long term climate change and short term climate variability in Northern British Columbia

Linear and non-linear statistical techniques were applied on the longest available water level records in northern BC to establish the long-term sea level trends and sea level responses to climatic variability in the region. Critical to this research was the development of linear regression models that would provide long-term sea level trends for this region. These models were based on the observed annual mean and annual maximum sea level data at Prince Rupert.

The analyses indicated that over the 20th century, this region experienced a statistically significant, accelerated, mean sea level rise of $+1.4 \pm 0.6$ mm
yr\(^{-1}\). It was demonstrated that over the same period the MAXSL trend was approximately twice this rate at +3.4 mm yr\(^{-1}\). My analyses demonstrated that the observed accelerated long-term sea level trend during the latter half of the 20\(^{th}\) century was, not only due to global warming, but also due to natural cyclic climate variability patterns.

This study statistically linked the trends of extreme sea levels and storm surges to major cyclical variations in climate, including El Niño, and the Pacific Decadal Oscillation. The research showed that shorter temporal scale climate variability impacts, when superimposed on longer term sea level rise trends, can be more hazardous to some coastal systems in BC than the longer term sea level rise trends alone. Thus, the extreme sea level fluctuations related to climate variability events should be of immediate focus for the development of coastal adaptation strategies in coastal BC.

7.2.2 Development of extreme sea level return-periods for the south coast of BC, with climate considerations

A sea-level return period curve was developed for the south coast of BC using extreme value analysis statistical technique in combination with climate covariates and historical extreme total water-levels (TWLs) at Point Atkinson. The results were utilized to illustrate warm (cold) ENSO’s significantly increasing (decreasing) the extreme TWL return levels in southern BC. It was shown that the historical TWL extremes along the BC coast were dominated by high astronomical tides, with an apparent scarcity of extreme surge-extreme tide interactions. These findings suggest that the sea level return period projections based on naturally occurring extreme TWLs are
more relevant for impact and adaptation studies in coastal BC, than return period projections based on more complex extreme value analysis techniques that consider rare surge and tide associations.

7.2.3 The spatial and temporal distribution of extreme sea surge recurrences and its sensitivity to climate variability in British Columbia.

Coastal areas in BC appear to be at risk from climate-change augmented storm surges when warm ENSO events and/or severe winter storms coincide with high tides. By examining the hourly tide gauge data recorded at 11-tide gauges in this region, I developed a model of extreme storm surge climatology for coastal BC. The model and its relationship to natural climate variability were established by parametrically examining the annual maximum extreme sea-surge data, while treating for trends and dependencies via climate covariate coefficients.

The sea-surge responses to climate variability impacts on sites along the BC coast were found to be fairly synchronous, with an increase (decrease) in the magnitude of extreme sea surges in response to warm (cold) ENSO conditions. It was shown that the most pronounced changes to the sea surge climatology were caused by the regionally dominant climate excursions such as ALPI, NOI, PNA and MEI. No significant long term extreme sea-surge trends were apparent along the BC coastline.

This study was the first to illustrate the spatial probability distribution of the largest surge events in coastal BC. The analysis showed a scarcity of high surges occurring at the time of high astronomical tides, suggesting that serious coastal flooding is governed by extreme astronomical tide levels and
not by extreme surges. Extreme sea surge event occurrences in the northern and southern regions of coastal BC showed weak dependencies, suggesting that the possibility of simultaneous flooding along the entire BC coastal reach from a single extreme sea surge event is rare. These historical storm characterizations are a critical first step in understanding future storms and their impacts under rising sea levels.

7.2.4 Extreme wind regime recurrences and its sensitivity to climate variability in the southern coast of British Columbia.

Three long-term hourly wind speed data records were used in combination with extreme value statistical techniques to assess the impact of climate variability on extreme wind speed recurrences in the inner south coast of BC. Shifts in the distribution of extreme winds in association with the broad-scale atmospheric and oceanic circulation modes indicated the existence of strong atmospheric and oceanic controls over the prevailing extreme wind regime in the region.

It was shown that high extreme winds in this region favour cold ENSO conditions. An important finding was that the signals associated with ENSO and extreme winds in the northern latitudes of the eastern Pacific coastal margin are opposite in nature than those of more southern latitudes where, more active winter storms were associated with warm ENSOs. This observation suggests that, using a single wind regime characteristic to represent a much wider spatial scale in the eastern Pacific region could give an incomplete, if not inaccurate, view of the actual relationship between extreme winds and climate in the region. This is an important finding that
strongly emphasizes the need to include regionally specific extremes, for regional-scale projections of climate variables.

Some of the main relationships affecting the extremes wind speeds and climate along eastern Pacific coastal margin were explained as follows: 1) during warm ENSOs, the Pacific storm track extends equator-ward and much farther downstream than during ‘normal’ winters. This brings more extreme winter storms to southern latitudinal locations; and, 2) during cold ENSOs, the jet stream shifts northwards bringing heavy storms to southern BC. The findings of this research also suggest that exceptionally high residual water levels along coastal BC during warm ENSO episodes may be due to other significant controls of warm ENSOs, such as the thermosteric effects and fresh water influx. Interestingly, the contribution of extreme winds towards elevating residual water levels during warm ENSO episodes along the southern coast of BC may be marginal.

7.2.5 Establishment of winter storm track characteristics and its relationship to various climate variability signals in coastal British Columbia.

By using NCEP/NCAR January storm track characteristics as a proxy for the winter storm track variability, in concert with linear and non-linear statistical techniques, it was shown that the north-eastern Pacific northerly and southerly winter storm track responses to natural climate variability was opposite in nature. The approach used in this study differed from many, in that storm track directionality was introduced into the analysis.

Storm track variability was shown to be significantly correlated with regionally dominant natural climate variability modes such as ALPI, PNA and PDO, where the tracks from the northern quadrant strengthens (weakens) in
response to persisting cold (warm) climatic conditions, while tracks from the southern quadrant strengthens (weakens) in response to persisting warm (cold) climatic conditions. It was shown that the northward shift of the southerly tracks and stronger than normal northerly tracks during cold climatic conditions were responsible for observed high wind-storms in coastal BC during cold ENSO periods. It was also shown that longer life span and, longer than normal storm track excursions during warm ENSOs were likely to cause winds of moderate strength lasting over periods of days in coastal BC, compared to their cold counter part. A major transition in the storm track structures from weak to a strong storm track regime was detected during early to mid-1970s. The winter storm tracks were probably responding to the major climate regime shift of mid-1970s which transmitted a shift in the ocean-atmosphere conditions along the Pacific coast through a series of climatic forcing mechanisms in the Pacific Northwest.

By integrating the results from the “extreme wind response analysis” and the results of this research component, it was shown that the frequent occurrences of cold ENSO related high wind-storm events in coastal BC were likely due to the net northward shift of the southerly storm tracks (closer to the BC coastal stretch) and, stronger than normal northerly storm tracks during the cold ENSO periods. Following this analysis, I was able to show that regional climate characteristics are fundamental in assessing the characteristic changes to storm tracks in coastal BC. These findings of these analyses are useful for modelling the future behaviour of storm tracks to naturally and anthropogenically-driven climate variability, and for assessing
the associated impacts when formulating regional planning and adaptation measures.

7.3 Recommendations for future works.

This dissertation describes potential regional scale changes in the physical coastal environmental variables of BC that are likely to result from the influence of climate variability and change. Several high impact events were situated within the context of this work to illustrate potential applications of the research results and the methodology. My results emphasize that the BC coast will be physically-impacted by higher sea levels over the next century.

Due to data scarcity in coastal BC, detailed local scale investigations on the response of the near shore wave dynamics and coastal geomorphology in response to Climate Change and Variability were not addressed within this study. Such studies should be conducted via complex mathematical modeling tools, as the phenomena that governs the coastal dynamics at local scale is quite complex. The results of such studies are crucial for a serious second look at many zoning laws, building codes, and engineering standards, as well as various plans and capabilities for dealing with emergencies due to climate impacts. Based upon the findings of my research, the following recommendations for future research are suggested:

i) There is a need to spatially expand the existing meteorological and oceanographic observation network along coastal BC, and a need to develop indicator time series data.
ii) Further research should be undertaken to distinguish climate variability (natural) impacts from climate change (human induced) impacts on BC’s coastal systems and, to identify thresholds in coastal systems in relation to past changes in climate.

iii) Enhance predictive capabilities of climate variability and change impacts on coastal BC by improving the: (a) quantitative assessment methods and, predictive models; (b) scale (i.e. resolution) of the investigation, and, (c) Improved data quality.

iv) Conduct detailed investigations on the contribution of steric sea level changes on sea level rise in coastal BC.

v) Re-evaluate the existing design specifications of coastal infrastructure to ensure they incorporate the physical impacts of future climate change and variability.

Development of robust and systematic coastal and ocean monitoring programs that ensure the availability of reliable long term homogenous time series data on a wider spatial scale is fundamental to: (i) better understand how climate change is driving changes in the coastal ecosystems; (ii) better establish the connectivity of large-scale regimes across the region; (iii) successfully differentiate the variation trends in climate over different temporal scales; and, (vi) identify consequences of climate change as they emerge. Some of the sea level rise and coastal flood related regional climate indicators that should be monitored include, but not limited to; mean sea-level, high water-level, extreme winds, storm track characteristics and shoreline erosion. It is also expected that systematic monitoring of climate indicators can play a
critical role in refining future projections and reducing uncertainties of climate variability and change impacts on coastal BC.

Considering the importance of the monitoring programs for future climate impacts works in BC, the federal government of Canada as well as the provincial government of BC needs to sustain existing monitoring programs and ensure their longevity by providing continuous and guaranteed funding. To that end, a dialog should be initiated among climate researchers on what aspects of BC's natural coastal environment should be monitored into the future as indicators.

During my research a high level of uncertainty about the potential responses of both meteorological (winds and storm tracks) and oceanographic (i.e. sea levels) variables to climate change became apparent. There is a significant knowledge gap that inhibits useful prediction of future sea level changes. In order to reduce these uncertainties, an integrated scientific program of sea level studies that combines the historic data with ongoing monitoring programs of physical and environmental changes is needed for the coastal BC region. Further research is also needed to make a clear distinction between the extent to which the present intensification of extremes is due to natural climate phenomenon and/or due to warmer climate from anthropogenic climate change.

Current research related to climate variability and change impacts in coastal BC largely emphasizes simple tracking of changes, which only allows for a reactive response to any negative impact. In order to be proactive it is necessary to determine what lies ahead. Application of predictive mathematical models with better representations of real-world settings and
processes facilitate these assessments for decision making. Significant opportunities still exist to improve predictions of coastal response to sea-level rise via mathematical modeling applications in many parts of coastal BC. For example, our understanding of the processes controlling the rates of sediment flux in both natural and, especially in human-modified coastal systems in BC is still limited at the regional scale. In addition, further research and regional scale numerical modeling is needed to improve the forecast and predictions of major storm events in BC, especially in the context of anticipated changes in frequency, intensity and direction in response to the changing climates. Development of new quantitative assessment methods and the refinement of existing techniques for advance predictions of climate variability and change impacts on coastal BC is another research area that has substantial research opportunities.

Even though the need for regional-scale climate impact analysis has been strongly voiced in BC, much remains to be done to devise appropriate methods for simulating future climate at regional level. Regionality is important, as the effects of global warming on climates and particularly on extreme weather are likely to differ substantially at regional scale. This regional scale (i.e. high resolution) mathematical modeling is fundamental to accurately simulate the implications of climate on sea level elevations and extreme events. High resolution bathymetric data, especially in the near-shore is fundamental for a complete topographic-bathymetric model to predict with greater confidence wave and current actions, inundation, coastal erosion, sediment transport, and storm effects.
Clearly, having the data to conduct a consistent analysis in a regional scale model in BC will not be easy, unless convenient access to high frequency data is made available by the government authorities responsible for maintaining such data. Therefore, the province of BC should actively invest in enhancing the availability of baseline data (i.e. elevation, bathymetric data, and shoreline position) for the coastal zone of BC at higher spatial resolution collected over appropriate time spans. Appropriate measures should also be taken to provide free and open access to all publicly-funded physical and social scientific data bases concerning the coastal and ocean regions of BC for the research community. More effort should also be made towards implementing lidar surveys with high horizontal and vertical accuracy for the development of high resolution bathymetric data in coastal BC. While some of these mapping data are being collected now, there are substantial areas along the BC coast that still needs higher quality data.

My research revealed many quality issues in the historical meteorological and oceanographic observational records that were made public via the Environment Canada as well as the DFO data bases. Many of these problems were linked to changes in the instrumentation over time and due to the changing physical characteristics of weather stations and their surroundings. Since these factors could introduce false trends into long-term climate analyses, it is recommended that they should be detected and corrected prior to releasing them to the public.

My research did not specifically address the contributions of halosteric and thermosteric effects on regional sea level rise. However, the findings of my research indirectly suggest that steric sea level changes in response to
changing climate may be an important contributor towards sea level variability in BC. Clearly more research and data is needed to verify this hypothesis. It is expected that, our understanding on the steric contributions to sea levels will continue to improve with the expanding time series of CTD profile data in the vicinity of BC (ex. cabled observatory instrumentation of NEPTUNE Canada). However, spatial expansion of such projects to cover the entire BC coastal region is recommended.

In order to ensure effective adaptation to current and future climate variability and change impacts, it is important that climate variability and change is considered as a criterion in coastal structural designs. Currently, the vulnerability of different types of infrastructure and, the potential impacts on the existing and new engineering codes and standards will have from climate change are largely unknown for the BC coastal region. In fact, one of the lessons learned from the recent 2006 storm surge was that the authorities to need to reassess their estimate of the risks of weather disasters and our ability to cope with them. This includes but not limited to, reviewing the calculations of the probabilities of different kinds of sea level and weather extremes, factoring sea level rise in to the planning of new infrastructure designs and setting up coastal set backs. Substantial research opportunities still exist in areas where reassessments, upgrade and regular updates of the design values are needed, particularly to ensure the reliability and the safety of coastal infrastructure.

References
