Carbon Dioxide and Energy Exchanges in the Coastal Zone of Hudson Bay

by

Glenn J. B. Scott

A Thesis submitted to the Faculty of Graduate Studies of

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in partial fulfilment of the requirements of the degree of

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Lasciate ogni speranza, voi ch'entrate

Nessun maggior dolore Che ricordarsi del tempo felice Nella miseria.

> -Durante Alighieri Divina Commedia, 1308

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Abstract

An eddy covariance system and micrometeorological station was deployed at two locations along the coastline of Hudson Bay during the summers of 2005 and 2006 to document and to understand mass and energy fluxes in high-latitude intertidal and near-shore environments. Despite the proximity of these two zones, it was found that they exhibited distinctly different characteristics. The near-shore zone was a sink for CO₂ with an average uptake of -0.11 µmol·m⁻ 2 ·s⁻¹ and the intertidal zone tended to be a source of CO₂ with an average efflux of 0.04 µmol·m⁻ $^{2}\cdot s^{-1}$ with considerable variability due to the action of the tides. Sensible heat fluxes in the nearshore zone tended to be small and negative and both latent and sensible heat fluxes were significantly enhanced in the intertidal zone. Significantly, increasing wind velocities did not appear to play a role in the enhancement of these fluxes and onshore winds were observed to be unusually dry. As such, key differences were observed that stood in contrast to the results and the conclusions of other flux studies conducted in similar high-latitude coastal-marine environments. It is suggested that these differences could only be understood in the context of the proximity of these areas of living and dead kelp, their respective differences in water depth and the occasional occurrence of a sea-breeze effect that may have implications for the observed fluxes in these areas.

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Glossary

α	-	Albedo		
ABL	-	Atmospheric Boundary Layer		
A _t	-	Total Alkalinity		
AWS	-	Automated Weather and Gas-Analysis Station		
В	-	Bowen Ratio (unitless)		
CaCO ₃	-	Calcium Carbonate (mol)		
CO_2	-	Carbon Dioxide (ppm)		
CST	-	Central Standard Time (UTC -6h)		
delT	-	$(T_a - TSURF)$ (°C)		
DIC	-	Dissolved Inorganic Carbon		
DOC	-	Dissolved Organic Carbon		
ξ	-	Effective Emissivity (unitless)		
e	-	Water Vapour Pressure (kPa)		
EBC	-	Energy Balance Closure (%)		
EC	-	Eddy Covariance (or Eddy Correlation)		
FC	-	Flux of Carbon Dioxide (μ mol·m ² s ⁻¹)		
G	-	Ground Heat Flux (W·m ⁻²)		
GPP	-	Gross Primary Production		
H ₂ O	-	Water (mmol·m ⁻³ or g·m ⁻³)		
Н	-	Flux of Sensible Heat (W·m ⁻²)		
HBL	-	Hudson Bay Lowlands		

IRGA	-	Infrared Gas Analyser
k	-	Gas transfer velocity $(\text{cm} \cdot \text{s}^{-1})$
n	-	Number of samples when referred to in statistical analysis
K _x	-	Turbulent Transfer Coefficient
L	-	Monin-Obukhov length (m)
LE	-	Flux of Latent Heat (W·m ⁻²)
m _a	-	Molecular weight of dry air (mol)
<i>m</i> _v	-	Molecular weight of water vapour (mol)
NEE	-	Net Ecosystem Exchange
OP	-	'Open-Path' (refers to a type of IRGA)
PA	-	Atmospheric Pressure Tendency (kPa)
p_d	-	Vapor pressure of dry air (kPa)
рН	-	Potential Hydrogen (a measure of the acidity or alkalinity of a solution)
p_v	-	Vapor pressure (kPa) – Also referred to as (e).
R	-	Respiration
ρ	-	Density of dry air $(g \cdot m^{-3})$
$ ho_{v}$	-	Density of water vapour $(g \cdot m^{-3})$
r^2	-	Coefficient of Determination (unitless)
Rg	-	Incoming Short Wave Radiation (W·m ⁻²)
Rg_out	-	Outgoing (Reflected) Short Wave Radiation $(W \cdot m^{-2})$
RH	-	Relative Humidity (%)

Rl	-	Incoming Long Wave Radiation (W·m ⁻²)
Rl_out	-	Outgoing Long Wave Radiation (W·m ⁻²)
REBS	-	Radiation Energy Balance Systems
RNET	-	Net Radiation ($W \cdot m^{-2}$)
S	-	Energy Storage (W·m ⁻²)
SST	-	Sea Surface Temperature (°C)
σ	-	Stefan-Boltzmann Constant (5.67·10 ⁻⁸ W·m ⁻² ·K ⁻⁴)
τ	-	Flux of Momentum (kg·m ⁻² ·s ⁻¹)
Та	-	Ambient Air Temperature (°C)
TSURF	-	Surface Temperature (°C)
UST	-	Friction Velocity $(m^2 \cdot s^{-1})$
VPD	-	Vapour Pressure Deficit (kPa)
df	-	Degrees of Freedom (the number of groups or treatments in a statistical analysis)
fCO ₂	-	The Fugacity of Carbon Dioxide
u	-	Lateral wind velocity vector $(m \cdot s^{-1})$
v	-	Cross wind velocity vector $(m \cdot s^{-1})$
W	-	Vertical wind velocity vector $(m \cdot s^{-1})$
WPL	-	Webb, Pearman, Leuning correction
WS	-	Horizontal wind speed $(m \cdot s^{-1})$
x	-	Lateral wind coordinate (m)
У	-	Cross wind coordinate (m)
Z	-	Vertical wind coordinate (m)

- z₀ Roughness Length (m)
- z/L Monin-Obhukov Stability Parameter (unitless)

1 - Introduction

1.1 - Climate change and carbon dioxide

1.1.1 Global Climate Change and Arctic Climatology

Recently, the issue of climate change in Arctic regions has acquired both increased scientific scrutiny and a large element of public awareness, much of which has been promoted with publicity for the International Polar Year [*IPY*, 2007] and through an inherent fixation on the plight of native Arctic fauna such as the polar bear (*Ursus maritimus*) [*Stirling and Parkinson*, 2006]. For the last few years, funding has been made available for various large scale research projects whose goals have been to provide 'integrated assessments' of the effects of climate change on natural, social and ecological systems in northern regions [*Lange et al.*, 2003; *ArcticNet*, 2006; *Barber et al.*, 2007]. The original impetus was driven by an understanding that polar regions are likely to experience climatic change at a much greater rate than the rest of the planet and the observation that many of these environmental changes have already been taking place in these vulnerable areas [*ACIA*, 2005; *IPCC*, 2007].

Historically, Arctic regions have been characterized by very low temperatures, large seasonal variability in the amount of solar radiation received, low precipitation, and a persistent snow and ice cover [*Rouse, 1993*]. However, there are at least two physical peculiarities of the Arctic that could make it naturally susceptible to global climate change:

• Spatial heterogeneity; geographically, this region is unique in the world in the sense that it is an ocean surrounded by landmasses. Thus, it presents a large interface between terrestrial and marine surface types, each with their own susceptibility to external forcings and impact on the local energy balance. As well, the terrestrial and marine components

of the Arctic system are linked not only by the atmosphere, but by riverine input into the Arctic Ocean and there is ample opportunity for feedback between these two areas.

Over the course of a year, any particular area in this region often experiences abrupt changes with regard to surface properties. With the exception of the circumpolar flaw lead [*Barber et al., 2007*] and other polynyas [*Yager et al., 1995*], the Arctic Ocean is covered by sea ice during the winter which acts to insulate the underlying water from significant mass and energy exchanges with the atmosphere. However, once this ice melts and decreases in extent during the Arctic summer, the much lower albedo of water will result in a greater retention of energy in the water column. This represents a considerable amount of energy due to the fact that large parts of this region are subject to 24 hour sunlight during the summer. During the month of June, this region has a potential insolation of 500 W·m⁻² whereas regions located at the equator may receive 450 W·m⁻² [*Rouse, 1993*].

In light of these primary physical factors that could make northern latitudes especially susceptible and vulnerable to climate change, the IPCC [2007], various researchers [*ArcticNet*, 2006] and stakeholders throughout the north [*Woo et al.*, 2007] have noticed and documented substantial climatic changes throughout the Arctic:

- Although there has been a high degree of spatial and temporal variability, average temperatures in the Arctic have increased at nearly twice the global rate over the course of the last 100 years [*IPCC*, 2007].
- The average annual Arctic sea ice extent has been shrinking, is very likely to continue decreasing in thickness [*Barber and Hanesiak, 2004*]. The largest decreases have been

occurring during the summers when the ice extent has decreased by 7.4 % per decade [*IPCC*, 2007].

• In the Canadian Arctic, there was an increase of adverse weather events and increased precipitation [*Hanesiak and Wang*, 2005; *Hanesiak et al.*, 2010]

Compared with other regions on the planet, the Arctic region may indeed have been acting like the 'canary in the coal mine' regarding global climate change. The climate models that are used by the ACIA [2005] and the IPCC [2001, 2007] are based on a number of different natural forcings and parameters and use different social and technological scenarios to predict that there will be further change in store for the Arctic region:

- It is predicted that the mean temperature in the Arctic will have increased by 5 7 °C.
 Summer air temperatures over terrestrial surfaces are projected to have a larger increase than those over marine surfaces due to the inherently slow thermal response of the oceans. Conversely, mean winter air temperatures over marine surfaces are projected to have a larger increase than those over terrestrial surfaces due to the reduced extent of sea ice during winter
- Sea ice in the Arctic is projected to shrink under all emission scenarios. In some projections [*Barber et al. 2007; IPCC 2007*] Arctic late-summer sea ice is projected to disappear almost entirely by the middle of this century.

• Precipitation at high latitudes is projected to increase by 14 % over the Arctic Ocean The accuracy of any model depends on its ability to predict and to parameterize various physical processes with an acute understanding of interactions between the land, the atmosphere and the ocean. There have, unfortunately, been large knowledge gaps in the science of climate change and little research has focused specifically on the interface between terrestrial and marine environments in polar regions.

1.1.2 - Arctic and Sub-Arctic Coastal Zones

At any latitude, the coastal zone represents an open-system interface where the land, the ocean and the atmosphere interact vigorously with one another [*Gattuso et al., 1998; Ducklow and McCallister, 2005*] and have a large impact on global biogeochemical cycles. Globally, coastlines extend over 350,000 km and 37 % of the human race lives within 100 km of a salt water shore. Within the Canadian Arctic, there are approximately 564 communities located on a salt water coastline [*Loring, 2007*] each of which intimately depends on the coastal interface for their livelihoods.

The marine portion of the coastline is shallow (<200 m), exchanges large amounts of matter and energy with the open ocean and is subject to very large inputs of organic matter and nutrients from adjacent shorelines, rivers, run-off and groundwater discharge. This material is subsequently buried in situ, advected to the deep ocean via a 'continental shelf pump' [*Tsunogai et al., 1999*], or is mineralized and frequently lost to the atmosphere as CO_2 [*Gattuso et al., 1998; Ciais et al., 2006*]. Despite significant lateral exchanges of carbon with the open ocean, CO_2 fluxes tend to be more intense per unit area of coastal ocean than those that occur over the deep ocean. The open ocean may have a higher overall impact on the global carbon budget but this is due primarily to its immense surface area [*Borges, 2005*].

Arctic marine coastal zones present at least a few unique differences from their low latitude counterparts:

- Over one third of the high latitude oceans and seas in the northern hemisphere are relatively shallow and are underlain by a continental shelf. This dramatically intensifies the role played by coastal zones relative to the open ocean on the mass and energy budgets of this region.
- With the exception of the circumpolar flaw leads [*Barber et al., 2007*] and various polynyas [*Yager et al., 1995*], a large portion of the Arctic Ocean is covered year-round by a layer of sea ice and the surrounding shelves tend to be covered with ice and fast-ice during the winter and spring. This ice subsequently melts and the underlying water becomes exposed to the atmosphere.
- It is during the brief, but intense polar summers that very high rates of primary production can be sustained in these areas due to the very long photoperiod [*ACIA*, 2005].
- Additionally, it is during the summers that riverine inputs become significantly larger due to melt runoff and summer rainfall over terrestrial surfaces [*ACIA*, 2005]. This results in a significant input of freshwater that tends to carry large quantities of organic and inorganic carbon and nutrients [*Gattuso et al., 1998; Ducklow and McCallister, 2005*] into the coastal zone of the Arctic Ocean and its related seas in the form of buoyant rivers plumes.

1.2 - Hudson Bay and the near-coastal zone

1.2.1 - Hudson Bay

Hudson Bay represents a large extension of the Arctic continental shelf into central Canada and, with an area of almost one million square kilometres, it is the largest inland sea in the world

[Ingram and Prinsenburg, 1998]. It is uniformly shallow with an average depth of 150 m and it exchanges large quantities of water with the Arctic and the Atlantic oceans through the Foxe Basin and Hudson Strait on an annual basis. Its freshwater input comes primarily from fresh water rivers, precipitation and melting ice and the plumes from these rivers are sources of nutrients and organic carbon which support communities of phytoplankton and zooplankton which, in turn support various aquatic fauna such as Arctic cod (Boreogadus saida) [Fortier et al., 1996] and Arctic char (Salvelinus alpinus), and marine mammals such as beluga whale (Delphinapterus leucas), harp seals (Pagophilus groenlandicus), ringed seals (Pusa hispida) and a world-renowned community of polar bears [Stirling and Parkinson, 2006]. Until the development of the ArcticNet project [2006] with its focus on climate change in the Canadian Arctic and sub-Arctic, there had been relatively few studies of the oceanography of Hudson Bay and those that have been done, have been limited in extent [Martini, 1986; Ingram and *Prinsenburg*, 1998; Gagnon and Gough, 2005]. It wasn't even known that the Bay completely froze over during the winter until 1948 [Rouse, 1991] and many of the studies performed since then have received their impetus from the reduction of the flow of water into Hudson Bay due to the development of hydroelectric dams in Manitoba and Quebec [Ingram, 1981].

In many respects, Hudson Bay can be thought of as an extraordinarily large estuary that experiences a full, annual cryogenic cycle [*Gagnon and Gough, 2005a, 2005b*] and is subject to a considerable semi-diurnal tide range that can exceed 12 m on the east side of the Bay [*Saucier et al., 2004*]. In the coastal areas, the ebb and flow of the tides helps to mix the normally well stratified waters of the Bay and can leave very large intertidal areas regularly exposed to the atmosphere. It is in these areas that shorefast ice develops in late October which proceeds to

cover nearly the entire bay with a layer of ice with an average thickness of between 1 and 2 m. This ice begins to break up in May and, due to the overall cyclonic circulation of the Bay, tends to accumulate on the southwestern shore near Churchill, Manitoba and a substantial ice pack can persist into the month of July.

Located immediately onshore of this lingering sea ice pack are the Hudson Bay Lowlands (HBL) which have been gradually emerging from this shoreline since the end of the last glaciation. The HBL are underlain by Silurian and Ordovician carbonate bedrock which is composed largely of limestone and dolostone. Over the course of the Late Cenozoic Ice Ages, this area has been frequently glaciated and the bedrock has been heavily eroded. This is particularly evident in the area near Churchill, Manitoba, where the shoreline is composed of various sized clasts that were once part of the underlying bedrock. The rate of isostatic rebound from the last glaciation in the Churchill area is currently estimated to be 1cm per year [*Braun*, 2006] and may have been more rapid in the past resulting in series of stranded shorelines far inland.

Another notable feature of the coastline are the large kelp beds (*macrocystis sp.*) near the low tide mark. On the shoreward side of the high tide mark and immediately adjacent to the high tide mark, there tend to be large piles of dead, rotting kelp that have been deposited by tidal and wave action.

1.2.2 - Synoptic and mesoscale meteorology

Atmospheric circulation and weather patterns over the HBL can only be understood in the context of the surface properties of Hudson Bay and their effect on the overlying atmosphere. Its

large size, coupled with its persistent ice cover, has a pronounced impact not only on continental scale synoptic meteorology, but on a more local scale through the 'sea-breeze effect' [*McKendry and Roulet*, 1994].

Although there are many diverse source areas for synoptic air masses that advect over the HBL, the dominant source area is the High Arctic and they are driven by the circumpolar westerly vortex [*Rouse*, 1993]. Due to the cold surface temperatures of the Hudson Bay ice pack and generally higher atmospheric pressure, the mean position of the Arctic front from October through April is driven much further south that might otherwise be expected. It is only when the Bay is mostly ice free that the mean position of the front returns to a position that is roughly parallel to the lines of latitude in the Canadian High Arctic. This southward plunge of cold, Arctic air well into the late spring results in the 'winterization of summer' and weather systems frequently follow this trough which tends to pass over the HBL during the spring and fall. The frequency of 'storminess' in the adjacent Hudson Bay is comparable to that in the North Atlantic [Rouse, 1993]. Although the average annual precipitation received in this area is relatively low (402 mm, 181 mm of which is received as snow [Boudreau and Rouse, 1995]), there are many adverse weather events that occur in this area [Hanesiak and Wang, 2005] Depending on the season, this area is prone to events of freezing rain (up to 2 % of the time during the fall and spring), blowing snow (up to 16 % frequency during the winter), fog (up to 10 % frequency during the spring, summer and fall) and low cloud ceilings (up to 25 % frequency during the spring and fall.) Incidents of no significant weather have been decreasing since 1953 due to an increase of precipitation and freezing rain events.

Superimposed on synoptic weather patterns during the melt season in this area is the 'sea-



breeze' phenomenon that has been observed and intensively studied along many global near-shore areas [*Oke, 1987; McKendry and Roulet, 1994; Plant and Keith, 2007*]. When synoptic wind patterns weaken at a coastal interface, a system of breezes may develop in response to gradients of air pressure that may arise due to differential heating of

terrestrial and marine surfaces [*Oke, 1987*]. During the melt season, heating of the land surface during the day can result in strong sensible heat fluxes into the atmosphere. In contrast, due to the large thermal inertia of an adjacent body of water, sensible heat fluxes tend to be suppressed and air temperatures remain cool. This has two mesoscale effects during the summer and fall: horizontally, a pressure gradient is formed over a relatively short distance that will result in an onshore wind during the day (Figure 1.1). Vertically, an expansion of the air column over the terrestrial surface will result in high pressure aloft and a flow at upper levels towards the water where the air mass will subside thus completing a cycle within a large circulation cell. Although

rarely observed due to a reduction of the sensible heat flux during the day, the cooling of the land during the night and the potential phased release of thermal energy from the ocean can result in a sea-breeze with a circulation cell flowing in an opposite direction thus resulting in an offshore surface wind. McKendry and Roulet [1994] researched this phenomenon on the nearby coast of James Bay and observed that during the day, this thermally forced circulation is shallow (<1000 m) but that its effects can extend up to 100 km inland. They observed its occurrence 25% of the time and that the flow in the circulation cells was diverted by the coriolis effect with the implication that such breezes will not necessarily be orthogonal to the shoreline.

Ultimately, both types of atmospheric circulation have had a significant impact on the surrounding region through the redirection of air masses, the forcing of wind direction and their influence on local weather. The spatial and temporal scale of these land-ocean-atmosphere interactions since the end of the last glaciation has had specific implications for the climate of the surrounding region.

1.2.3 - Climatic implications of Hudson Bay

If one were to understand climate as a function only of latitude, then the very existence of the HBL in this geographical location might not be expected. In the same sense that the temperate climate of Western Europe is a latitudinal anomaly due to the existence of the Gulf Stream and the release of sensible heat in the North Atlantic, the harsh and frigid climate of the HBL is anomalous due to the nature and size of Hudson Bay. This has clear implications for the nature of energy exchanges in this region and the potential impact of climate change [*Rouse, 1991, 1993; Ingram and Prinsenburg, 1998; Gagnon and Gough, 2005a, 2005b*].

The effects of climate change have already been observed in Hudson Bay where average annual air temperatures have risen by 0.5 °C per decade over the course of the last thirty years [*Gagnon and Gough, 2005a*] and the sea ice has been breaking up an average of three weeks earlier along the southwestern coast of Hudson Bay [*Stirling and Parkinson, 2006*]. The use of various atmosphere-land-ocean models to project future climate change in this area [*Gagnon and Gough, 2005b*] uniformly indicate that there will be a one month advance of the sea ice melt date and one month delay for the refreezing of Hudson Bay. The zone of continuous permafrost in the Hudson Bay Lowlands is projected to decrease by 35-67 % by 2040-2067 and could entirely disappear by 2090 under a 2X CO₂ scenario. As well, increases in the mean annual air temperature, especially between the month of October and April when the decline of sea ice will be the most noticeable, are projected to be 5 °C higher.

There is, however, a considerable amount of uncertainty in the application of these models due to a limited understanding of how the rates and types of precipitation and the corresponding soil moisture conditions will change in the future. As well, many of these models have a very large spatial resolution and each grid cell can span hundreds of kilometers. Like many Arctic coastal regions, this area is spatially heterogeneous and the biological and physical processes at work within each unique terrain type have not been extensively studied. The impact on vegetation growth dynamics, carbon sequestering and, importantly, the energy balance of this ecosystem will benefit from the development of a process level understanding on a much smaller scale than that which is currently employed in many of these models [*Lynch et al., 1999; Gagnon and Gough, 2005b*] The very existence of the wetlands of the HBL and the livelihoods of the people who depend on the ecosystems may be at risk [*Winter and Woo, 1990; Gilligan, 2007*] and it will be important to be able to produce better projections for change in this vulnerable area. Such models will need to be based on a careful understanding of the nature mass and energy exchanges through the underlying surfaces and how they react to forcing factors over various spatial and temporal scales.

1.3 - Surficial energy and mass exchanges

1.3.1 - Energy and mass exchanges through high latitude, coastal-marine surfaces

Globally, it has been estimated through the upscaling and integration of a large number of individual research projects in different areas that ocean surfaces are a carbon sink for approximately 1.93 PgC·yr⁻¹ [*Borges, 2005*]. In particular, it has been estimated that high latitude coastal areas absorb a highly disproportionate amount of CO₂ in proportion to their surface areas [*Ducklow and McCallister, 2005; Borges, 2005*] and could play a significant role in the global carbon balance. The term, 'near-shore' refers to the zone of water that is immediately adjacent to the low tide mark, is presumably affected by terrestrial or riverine inputs of fresh, nutrient-rich water and extends some distance over the continental shelf. Over the past couple of decades, there have been numerous studies that have focused on CO₂ exchanges over high-latitude continental shelves but, due to various logistical and technical problems with making measurements of energy balances in marine environments [*Fairall et al., 2000*], there have been comparatively few studies that have focused on energy fluxes in these areas.

Mass fluxes in high latitude marine environments are influenced by various physical factors such as wind speed (WS), ice cover, sea surface temperature (SST), total dissolved inorganic carbon (DIC), dissolved organic carbon (DOC), total alkalinity (A_t), salinity and other

aspects of water chemistry. These fluxes are also influenced by biological factors such as gross primary production (GPP) due to phytoplankton and macrophytes such as kelp and respiration (R) by zooplankton and other animals. As such, various studies have modeled mass exchange dynamics in high-latitude coastal areas and have required the development of a process-level understanding that is based on a knowledge of the physical properties of the interface between two liquids – the atmosphere and the ocean. Common to many of the studies of CO_2 exchanges over high latitude continental shelves is the use of the interface/gradient method to make these measurements and the comparatively small, negative fluxes that have been observed (Table 1.1).

The interface/gradient method calculates the flux of CO_2 by determining the relative difference between the partial pressure of CO_2 in the water and its pressure in the overlying atmosphere. This is based on measurements of CO_2 solubility in the water and the calculation of a transfer velocity through the water-air interface:

$$FC = ks(\Delta fCO_{\gamma})$$
^[1]

Where,

 $k = \text{transfer velocity of CO}_2 (\text{m} \cdot \text{s}^{-1})$

s = the solubility of CO₂ at a given water temperature and salinity

 ΔfCO_2 = the air-sea gradient of CO₂ (mmol·m⁻³)

Despite the widespread use of the interface/gradient method, there is still a great deal of uncertainty regarding the exact parameterization of the transfer velocity of CO_2 . Various estimates and parameterization schemes of k, including the cubic relationship of Wanninkhof and McGillis [1999], can differ by up to a factor of two at high wind speeds. Kuss et al. [2004] conducted a series of five cruises into the Baltic Sea (aka. the Gotland Sea) and used both a

Researcher & Technique	Location	Research Year(s)	Time of Year	Surface Type	FC µmol·m ⁻² ·s ⁻¹	Environmental Forcings
Omar et al.	Barents	1990-	Annual	Ocean	-0.05	Wind speed
[2007]	Sea	1999				
Interface gradient						
Yager et al.	Northeast	1992	Summer	Polynya	-0.06	Wind speed
[1995]	Water	100000000			11/1/10	Insolation
Interface gradient	Polynya					GPP
12-101 -1	D.:	4000	C	0	0.000	Storms
Kaltin et al.	Bering -	1993 -	Summer	Ocean/	-0.082	GPP
[2005]	Chukchi Sea	1994		Shelf		TSURF
Cibeen et al	Antoratio	1002	Annual	Constal	0 12 (1at aummar)	CDD
Gibson et al.	Antarctic	1995 -	Annual	Coastal	-0.45 (Tst summer)	GFF Mind anod
Interface gradient	COASI	1555		sea	-0.55 (2nd summer)	Air temperature
Nitiehineky et al	Lantov -	100/	Summor	Ocean/	0.001	Riverine input
[2007]	E Siberian Sea	1004	Juimer	Shelf	0.001	Wind sneed
Interface gradient	E. Olbendir det			oneir		wind speed
Stoll et al	Antarctic	1996	Farly	Ocean	-0 001 to 0 001	Air temperature
[1999]	sea		winter			- in componentino
Interface gradient						
Fransson et al.	Barents -	1996 -	Annual	Ocean	-0.034	Total alkalinity
[2001]	Laptev Sea	1997				
Interface gradient	Contra Co					
McGillis et al.	North	1998	Early	Ocean	-0.15	Wind speed
[2001]	Atlantic		summer			
Eddy covariance						
Interface gradient						
Murata et al.	Beaufort -	1998 -	Summer	Ocean/	-0.038	Water temp.
[2003]	Chukchi Sea	2000		Shelf		
Interface gradient						
Kaltin et al.	Barents	1999	Summer	Ocean	-0.03	Water temp.
[2002]	Sea					
Interface gradient				211 12		
Kuss et al.	Baltic	1999 -	Winter	Shallow	-0.49 (Dec./Jan.)	Wind speed
[2004]	Sea	2000		sea	-0.43 (Jan./Feb.)	
Interface gradient		0000		~	-0.12 (Feb./Mar.)	000
Bates	Chukchi Sea	2002	Summer	Ocean/	-0.1/1	GPP
[2006]				Shelf		
Interface gradient	1 matrix	0005	C +	0	0.0111-0.00	Distanting to the
Semiletov et al.	Laptev	2005	Sept.	Ocean/	-0.014 to 0.02	Riverine input
[2007] Eddy covariance	Sea			Shelf		
Eucy covariance	Hudeon	2005	Cont	Occant	0 226 += 0 101	Divoring input
[2008]	Bay	2005	Oct	Shalf	-0.220 (0 0.191	Rivenne input
[2000]	Day		Oct.	Silell		
intenace gradient						

Table 1.1 Results of previous FC studies performed on comparable high latitude marine surfaces. Although each of these studies was performed either over or close to continental shelves, none of them took place in a 'near-shore' zone.

balance approach and the air-sea flux-gradient approach with various wind speed

parameterizations to determine FC in this area. The balance method involved simply measuring

the difference of DIC in the surface water over discrete time periods and assuming that all changes were the results of vertical exchanges. They found that the balance method tended to provide much larger flux estimates and that the use of different wind speed parameterizations did not make a significant difference in the fluxes that were measured with this method. Yager et al. [*1995*], while studying FC in the surface waters of the Northeast Water (NEW) Polynya, modeled the effects of wind speed on *k* and determined that summer storms may play a significant role in the replenishment of CO₂ after a season of the biological fixation of carbon. They found that 14 days of sustained winds of up to 15 m·s⁻¹ would be sufficient to resaturate the surface waters after a season of strong GPP. The presence of ice cover in the Arctic usually prevents the overlying atmosphere and it associated wind from interacting with the water.

The solubility (*s*) of CO_2 has an inverse relationship between salinity, water temperature and calcium carbonate (CaCO₃) dissolution and a positive one with alkalinity [*Millero*, 2000] and CaCO₃ precipitation [*Gattuso et al.*, 1993]. Of particular importance to the Hudson Bay coastal zone, a 'solubility pump' has been identified wherein the solubility of cool water tends to increase with lower temperatures and absorb CO₂. Due to its increased density by virtue of having cooled, this water will tend to sink and form oceanic bottom water [*Takahashi et al.*, 2002]. This can occur where ocean currents move into higher latitudes and when ice begins to form in polar regions. Ultimately, this CO₂ becomes part of the deep water formation and joins the global oceanic 'conveyor belt' which may upwell at lower latitudes. Consequently, the water will warm, the CO₂ solubility will decrease, it will become supersaturated relative to the atmosphere and the water will evade CO₂ to the atmosphere.

CaCO₃ dissolution and precipitation are related as follows:

$$Ca^{2+} + 2HCO_3 - \Leftrightarrow CaCO_3 + CO_2 + H_2O$$
[2]

Calcification is a source of CO_2 to the water column and atmosphere and the direction and magnitude of the CO_2 flux at the atmosphere-sea interface depends on the difference between the CO_2 partial pressure in these two phases [*Gattuso et at., 1993*]. CaCO₃ in underlying carbonate bedrock could be weathered and dissolved thus reducing the partial pressure of CO_2 in the water column or calcium and bicarbonate ions could be supplied to a water column by previously eroded substrates or by some biological processes. The latter process would increase the partial pressure of CO_2 in the water column through calcification and deposition. Although there are some lower latitude of the effects of calcification on the saturation of CO_2 in the water column [*Dellile et al., 2000*], only the study of Kaltin et al. [2005] has examined the role of calcification or CaCO₃ dissolution on the observed flux of CO_2 in high latitude coastal areas. In this study, CaCO₃ dissolution was not found to play a significant role as a driver of the observed fluxes, but the authors speculated that this would vary on a yearly basis due to the sporadic appearance of CaCO₃ forming plankton.

If the solubility pump was the only process at work in the oceans, then it might be expected that the vertical profile of the partial pressure of CO_2 would consistently show elevated levels at the surface due to the presence of warm, less soluble water or due to the presence of cool, highly soluble water that would be inducing a drawdown of CO_2 from the atmosphere. This, however, is not the case in most regions due to a process known as the 'biologic pump' which is particularly effective in coastal areas due to their high rates of productivity [*Borges et al., 2005*]. Thomas et al. [2005] discussed this in detail while examining biological and thermal forcings on the surface water partial pressure of CO_2 in the North Sea. They noted that coastal seas are the major link between terrestrial and open ocean environments and host a disproportionately large fraction of ocean productivity. Photosynthetic organisms such as phytoplankton induce a drawdown of atmospheric CO_2 into the ocean which is subsequently converted into organic carbon. Depending on the level of mixing on the continental shelf, this organic carbon then sinks due to gravitational sedimentation [*Ducklow and McCallister, 2005*] and joins the formation of deep water or is deposited on the shelf. This is similar to the 'continental shelf' pump proposed by Tsunogai et al. [*1999*] whereby water that is cooled over the continental shelf due to seasonal heat loss is advected through the bottom of the water column to the deep ocean.

Another biological factor that has been shown to influence the partial pressure of CO_2 in sea water is the presence of kelp beds in relatively shallow waters [*Dellile et al., 2000; Delille at al., 2009*]. Kelp reacts strongly to insolation which penetrates the surface layers of the sea water and has been shown to influence FC on both daily and seasonal time scales in high-latitude coastal environments. They tend to exert a drag on water currents and, in the presence of insolation, they tend to increase their GPP in response to fast moving currents. As well, Delille et al. [2000] observed that there tended to be significant rates of calcification within the kelps bed that they were studying near the Kerguela Archipelago in the Southern Ocean.

Carbon dioxide forcings and fluxes were examined in a number of studies that were conducted in coastal areas of the Arctic, the Subarctic and the Antarctic (Table 1.1). Although a few of these studies were conducted in areas that were some distance from the actual coastline and two of them took place in different hemispheres, the majority of them took place over high latitude continental shelves and were subject to many of the same physical influences that would

be expected closer to shore. In general, they tended to be very small compared to their terrestrial counterparts and usually indicated a net drawdown of CO_2 over the course of the various study periods. However, it should be noted that the resultant net fluxes measured during these studies were much more consistent than those that were measured from terrestrial environments. Each study indicated that there was a net drawdown, the strongest of which occurred over the more temperate Baltic Sea and the weakest occurred in the Weddell Sea adjacent the deep ocean. Many of the same primary forcings of FC were examined including wind speed, the biological pump and insolation, and water temperature. Additional forcings that were examined included ice cover, ocean currents, fresh water discharge and storms.

Nearly half of the thirteen studies referred to Table 1.1 specifically consider wind speed to be one of the primary determinants of the flux of carbon dioxide (FC). Taking into consideration the cubic relationship between wind speed and observed fluxes as determined by Wanninkhof and McGillis [*1999*], these studies noted that there were significantly higher fluxes of CO₂ into the water surface during high wind events. During the GasEx-98 project, McGillis et al. [*2001*] compared the use of both direct and indirect methods to estimate FC through the surface of the North Atlantic Ocean. An EC system was deployed in conjunction with an IRGA that analysed surface water samples and the measurements were indirectly calculated using the cubic wind speed relationship. It was found that the measured fluxes were comparable with one another, however the EC estimates exceeded those that used the air-water flux-gradient method and the cubic relationship. Possible reasons for the observation of the enhanced gas transfer velocity as measured with the EC system include the fact that indirect methods cannot discriminate surface process variability such as atmospheric stability, upper ocean mixing, wave

age or wave breaking [*Liss et al., 2004*]. The other studies utilized only the air-water flux gradient method which may be due, in part, to the logistic and methodological difficulties of deploying EC systems on a moving platform [*Kuss et al., 2004*].

Similar to other projects, including one conducted by Gibson and Trull [1999] in the Antarctic, Yager et al. [1995] noted that there is a strong biological drawdown of CO₂ that begins to occur with the advent of spring. A drawdown of up to -0.43 μ molCO₂·m⁻²·s⁻¹ was observed in the East Antarctic Sea and a drawdown of -0.06 μ molCO₂·m⁻²·s⁻¹ in the NEW Polynya both coincided with high levels of chlorophyll-a during the summer when there is a very high level of insolation thus indicating that GPP and the 'biological pump' were significant forcings on mass exchanges in these systems.

A closely related factor is the water temperature. Stoll et al. [1999] conducted a study of carbon exchanges in the Weddell Sea off of the Antarctic coast during a cruise of the *Polarstern*. They concluded that this area is a sink for an average of $0.01 \,\mu\text{molCO}_2 \cdot \text{m}^{-2} \cdot \text{s}^{-1}$ which is due, at least in part, to the cooling of the sea water during the onset of winter and that the 'solubility pump' plays a significant role in determining the magnitude of an observed flux. Although one might expect that this could lead to the water achieving a level of equilibrium with the atmosphere, they found that water continued to be undersaturated beneath the ice cover during the winter due to the removal of this water by ocean currents beneath the shelf. This stands in contrast to the model devised by Yager et al. [1995] and the observations of Miller et al. [1999] which indicated that CO₂ builds up under the ice cover due to the action of respiring organisms (primarily zooplankton) and is outgassed when the ice cover broke up in the spring. However, it was strongly suspected that advection of underlying water masses may have played a significant

role in the observed fluxes of CO_2 over the course of the year at the North Water Polynya [*Miller et al.*, 1999].

Despite the fact that exchanges of heat and momentum play a key role with the coupling of the atmosphere and the ocean, fluxes of sensible (H) and latent heat (LE) have rarely been measured directly in high-latitude oceanic environments [*Bourassa et al., 2009*]. Most of these studies such as the Surface Heat Budget of the Arctic Ocean (SHEBA) [*Ola, et al., 2002*] and the North Pole Environmental Observatory (NPEO) [*McPhee et al., 2003*] that have made long-term observations of H only over sea ice. They have typically been estimated from bulk parameterizations that have been developed in lower latitudes and their applicability to high latitude flux estimations is questionable. In general, the estimated heat fluxes have been very small but are variable and driven by extreme weather events. Serreze et al. [2007] evaluated several H flux estimation projects for the Arctic Ocean that were based on estimations derived from satellite platforms and found that they tended to range from 2.4 to 11 W·m⁻² during the summer period. Studies conducted in coastal zones in other areas of the world have found that H is often influenced by the horizontal advection of temperature and changing surface roughness (z_0), thus making it difficult to obtain valid measurements.

In an eddy covariance (EC) experiment conducted 2 km off of the Danish coast, Mahrt et al. [*1998*] found that there was often no relationship between the thermal and the momentum roughness lengths. Importantly, they found that thermal roughness lengths tended to be significantly enhanced by wave breaks and that horizontal advection during onshore winds was frequently causing a deviation of the flux-gradient relationship from that which would otherwise be predicted by the Monin-Obhukov stability parameter (z/L).

Panin et al. [2005] analysed a number of coastal flux datasets from around the world and found that H and LE estimates in shallow waters are very sensitive to basin depth due to the changing thermal regime of the water and its roughness length. Based on a study that they conducted in the Caspian Sea from 1990 to 1992, they found that the increased drag coefficient of breaking waves in shallow zones resulted in an increased interaction with the atmosphere and an enhancement of H and LE. They developed models for H and LE that were modified by the ratio of wave height to basin depth and found a good correlation between their models and eddy covariance measurements. However, it has been noted in other studies [*Abdella and D'Alessio, 2003*] that positive relationships between roughness length and observed energy fluxes tend to break down at low wind speeds where the influence of 'free convection' will replace the influence of roughness length.

1.3.2 - Energy and mass exchanges in high-latitude tidal flats

The term, 'intertidal flats' refers to the zone between the high and low tide marks that is subject to daily flooding and draining. Most of the factors that influence energy and mass balances over high-latitude near-shore areas can be assumed to be at work in these areas as well, at least when they are flooded. However, very little research has focused on energy fluxes and mass exchanges in these areas and such results have generally been mentioned as a byproduct of other research. This is likely due to the dynamic nature of this tidal environment and the inability of conventional discrete sampling methods such as chamber sampling to resolve CO_2 fluxes on a time scale that would accurately resolve tidal cycles [*Zemmelink et al., 2009*]. Only recently has it been demonstrated that the eddy covariance (EC) technique, which is capable to resolving mass

and energy fluxes on small time scales, can be used in this environment. At this time, there is only one example of published research that deals with CO_2 fluxes in the intertidal areas of high latitude environments.

The most comparable study of CO_2 fluxes that has been conducted in an intertidal environment was performed in the Wadden Sea estuary on the coast of The Netherlands by Zemmelink et al. [2009]. Eddy covariance measurements were conducted during the spring of 2008 that, significantly, did not show any dependence of the flux of CO_2 on tidal stage. They hypothesized that biological productivity in the water column during high tide and in microbial mats that covered the sediments during low tide led to an undersaturation of CO_2 that was strong enough to support the observed fluxes and their consistency between different tidal stages. On average, the tidal flats were a weak sink for CO_2 of 0.5 µmol·m⁻²·s⁻¹ over the course of the sampling period.

Despite the paucity of published research dealing with energy exchanges in high latitude, ice-free intertidal environments, there is one study in particular that presented a very detailed analysis of such fluxes that was based on research performed directly in the intertidal zone of Hudson Bay near Churchill, Manitoba. In August and September 1985, Silis et al. [1989] erected a 10 m tall micrometeorological tower 1.2 km seaward of the high tide mark (Figure 1.6) that was used to measure the energy balance of the intertidal flats. Wind speed and air temperature was measured at multiple heights which allowed the calculation of sensible and latent heat fluxes through the use of the aerodynamic method. The calculation of the other components of the energy balance were measured with soil heat flux plates and water temperature sensors. During the month of August, ambient air temperatures of 10.9 °C were close to the long term average of
11.3 °C for the Churchill area and precipitation was somewhat low at 36.9 mm compared to the long term average for this month of 58.3 mm. They observed that average sea surface temperatures were 8.4 °C and tended to be lower than air temperatures by 1.4 °C at this time of the year. Onshore winds tended to be cooler than offshore winds and were usually close to being saturated. They observed that the albedo of the drained flats was twice what it was when they were inundated at 0.194 and 0.097 respectively. These factors had a direct impact on the energy balance of the area in the vicinity of the station; the ratio of net radiation to incoming short wave radiation was 13% lower due to the larger albedo. This was also influenced by a stronger long wave radiation loss from the exposed flats. They also observed that the ground heat flux into bottom sediments was small with net gains throughout the month of August. During the day, heat storage in the water was large and positive and was small and negative during the night. This occurred even during tide-out conditions due to the presence of ponded water on the intertidal flats. Latent heat fluxes were always positive and the largest during daytime low-tide and nighttime high-tide conditions. Sensible heat fluxes tended to be positive during cooler onshore winds and often negative for warmer offshore winds. In general, Silis et al. [1989] observed that energy storage (S) comprised 60% of net radiation, latent heat fluxes comprised 35% of net radiation and the remainder was divided equally between sensible and ground heat fluxes.

1.3.3 - Present research rationale and objectives

As an integrated assessment of the impacts of climate change on the sensitive ecosystems and the potential vulnerability of the inhabitants of Canada's high latitudes, the ArcticNet Centres of

Excellence project has initiated a large number of studies that have taken into account many of the complex interrelationships between various aspects of the natural and social systems in Hudson Bay [*ArcticNet*, 2006]. This region has already experienced profound climatic change and will likely continue to experience rapid change in the near future due to the complex natural linkages between different components of the land-atmosphere-ocean system. Due to feedback processes, the state of one component influences the mass and energy exchanges of another and these changes will likely have the greatest impact at their interfaces, the most significant of which occurs in the near-coastal zone. To date, there have very few studies that have focused on mass and energy exchanges in high-latitude intertidal and near-shore areas. Therefore, the primary objective of our research is to develop a detailed process-level understanding of mass and energy balances in the coastal zone of Hudson Bay. Specifically, this study intends to:

- <u>document</u> the rates and associated variability of the seasonally evolving air-surface exchange of heat, water vapour and CO₂ in the intertidal and near-shore zones of southwestern Hudson Bay over the course of two consecutive years.
- 2) <u>compare</u> the nature of turbulent exchange within the near-shore and intertidal zones with other studies that have been conducted in similar environments.
- <u>develop</u> a process-level understanding of major atmospheric and oceanic forcings on the mass and energy balances in this area.

Ultimately, we would like to understand how the mass and energy exchanges in this environment will be affected by, and will contribute to climate change in Canada's high latitudes. It is expected that the presence of living kelp in the near-shore zone will result in a significant drawdown of CO_2 that will be counterbalanced by a strong efflux in the intertidal zone due to the

presence of dead kelp. All fluxes, including those of sensible and latent heat will be enhanced when the water is in motion due to enhanced water mixing..

The Methodology section describes in detail the two automated weather stations (AWSs) that were deployed on the coastline of Hudson Bay during 2005 and 2006 in order to measure and document fluxes and forcings in the near-shore and intertidal zones. Shortcomings with each of these stations are described which explain footprinting problems with the Near-Shore Station and the need to develop a net radiation model based on the data from this station in order to be able to estimate the forcings at work in the intertidal zone the following year. A very basic description of eddy covariance theory follows which will help the reader to understand the nature of the corrections that were applied to each dataset. The Results section is divided between the near-shore and the intertidal zone. Within each subsection, forcings and fluxes are described individually as ensemble averages and basic parameterizations have been derived. Each section ends with a focus on the processes that were observed on individual days to help the reader understand the relationships between the individual variables. The Discussion section will focus on three basic facts regarding this environment - mixing within the water column, the presence of dead and living kelp and the existence of the 'sea-breeze' phenomenon. This section will compare and contrast the results with previous studies in similar environments. It is only within the context of these facts that any sense can be made of the parameterizations and a strong suggestion is made regarding the impact of the sea-breeze phenomenon on the observed fluxes. Water chemistry and calcium carbonate dynamics will not be considered in this analysis.

2 - Methodology

2.1 - Sampling locations and equipment

In support of the project objectives, it was necessary to deploy an automated weather and gasanalysis station that measured the fluxes of carbon dioxide, and sensible and latent heat with the use of an open-path (OP) eddy covariance (EC) system. An OPEC was chosen due to its versatility, low maintenance and low power requirements. As well, because the focus of this study was the thaw season, there was no need to deal with inaccurate flux estimations due to instrument heating and be required to use post-processing corrections [*Burba et al., 2008*] that are still not considered to be entirely reliable [*Amiro et al., 2006;Amiro, 2010*]. In addition to this, the stations supported instrumentation that measured other meteorological variables and potential forcing factors such as wind direction and speed, incoming and outgoing radiation and atmospheric and surface temperatures.

The location of the AWS was of particular importance due to footprinting issues. At the time, it was not logistically feasible to locate the instrumentation directly in the middle of the intertidal flats or in the middle of the near-shore area. As such, the only two possible locations to use were onboard a shipwreck located at the low-tide mark and on the coastline near the high-tide mark. The inherent problem would be that an AWS on the shipwreck would be unable to take measurements from the intertidal flats due to flow distortions from the ship itself, and an AWS located at the low-tide mark would not be able to deploy downward facing radiation instrumentation that would include the actual flats in their radiation footprint. Therefore, it was decided to deploy the station on the shipwreck during the first summer of operation to gather flux data from the near-shore area and to carefully monitor radiation exchanges that could be

modeled. During a second summer of operation, the AWS would be redeployed to the adjacent shoreline where the radiation model from the previous year could be applied and used in conjunction with the measured fluxes from the flats in order to fully understand the relationships between the flux estimate and their forcings. However, the shoreline had a low escarpment behind it which, like the shipwreck, would make all flux estimates generated during offshore winds unusable due to flow distortions. Therefore, an implication of this methodology is that all observations will be valid only for onshore winds.

The first station was erected on the coast on the shipwreck of the Ithaca which is situated in the intertidal flats of the coast near the low tide mark and remained in operation from June 11th through August 19th, 2005. This station was subsequently moved to a nearby location at 'the Bluffs' on the coast near the high tide mark and collected data from June 11th through August 31st, 2006. Data were collected by these stations outside of the stated time frames, but were not incorporated into this study due to the presence and formation of sea ice.

2.1.1 - The near-shore zone, 2005

Located in the intertidal flats at 58°46'08.81"N, 93°53'21.44"W, the Ithaca provided an ideal platform from which to sample the near-shore environment to the immediate north (Figure 2.1). The area undergoes extreme changes on both a seasonal and a daily basis; the flats are covered in shorefast ice from November through May and ice floes are deposited by the tide during the months of June and July. For the remainder of the melt season, the flats are flooded by the high tide twice per day and many shallow pools are left behind during low tide. Due to the very gentle slope of the flats, the intertidal area extends an average of 2.2 km into Hudson Bay and the



Figure 2.1 The Near-Shore Station was in operation during 2005 and the Intertidal Station was in operation during 2006.

average water depth where the Ithaca is located is 1.9 m at the maximum height of the average high tide [*Silis et al., 1989*]. The intertidal sediments consist of sand and clay interspersed between large quantities of igneous and sedimentary clasts with an average diameter of 50 cm. There is no organic soil and the biota largely consists of kelp, piles of dead kelp and various crustaceans.

The instruments (refer to Appendix A, '*Hardware*') were mounted on a triangular scaffold (12 " per face) and the system was powered by a grid of 12 V deep-cycle batteries that were recharged with an array of solar panels and a charge regulator. The eddy covariance instrumentation was positioned at a height of 14.8 m above the surface of the intertidal flats. A Campbell Scientific Inc. (CSI) CR5000 datalogger operated the eddy covariance system and was positioned on the upper aft deck of the Ithaca and was oriented due north (Figure 2.2). A Vaisala Inc. temperature and relative humidity probe was placed on the scaffold at a height of 13.3 m. A CSI CR10X datalogger was installed above the bow on the opposite end of the ship to operate and record the data from the Eppley Inc. pryanometers, the Eppley Inc. pyrgeometer and the REBS® pyrradiometer. Each of the instruments were mounted at a height of 9.3 m above the



Figure 2.2 The instrumentation at the Near-Shore Station on board the Ithaca.

unflooded flats and were all facing due south. All EC data were sampled at 20 Hz and all low frequency data were sampled at 0.33 Hz. The data from the CR5000 were transmitted back to a base computer at the Churchill Northern

Studies Centre every 15 minutes through an RF telemetry system to a base computer with LoggerNet© 3.1.3 installed on it. The data gathered by the CR10X were downloaded once per week and copied to the base station computer. Visits to the site took place three times per week over the course of the sampling period during which basic equipment maintenance was performed and dessicant packs were changed.

2.1.2 - The intertidal zone, 2006

After the field season in 2005, all of the equipment was removed from the ship and most of it was placed on a large rectangular scaffold which was erected approximately 1 km to the west of the ship next to the high tide mark at 58°45'55.46"N, 93°56'14.12"W at a site known locally as

'The Bluffs'. The shoreline is oriented along an east-west axis and, with the EC system pointing due north, it provided an ideal location from which to sample the intertidal flats. The EC instrumentation was positioned at a height of approximately 6.8 m. The temperature and relative humidity probe was installed on the scaffold at a height of 5.8 m. Importantly, the pyrradiometer, the pyranometer and the pyrgeometer that were measuring upwelling radiation at the Ithaca could not be used at this site due to the difference between the properties of the terrestrial surface at the base of the scaffold and the marine surface that would be within the fetch of the EC system. However, incoming short wave radiation was still measured and the pyranometer was placed at a height of 6.1 m. This meant that the CR10X was no longer needed, however the data from the CR5000 were collected and weekly maintenance was performed in same manner as it was at the Ithaca.

2.2 - Eddy covariance theory

In order to absolutely determine the small and large scale motion of air and its constituents, one must be able to define and measure all of the factors that could influence the motion and properties of a turbulent eddy. As a second order closure, the eddy covariance (EC) method has greatly improved the accuracy of these measurements, but is still subject to influences that cannot be accounted for with existing equipment or statistical techniques and still suffers from a lack of closure. Although a large number of corrections can be applied in the post-processing of data to continue to make EC systems useful in the determination of the flux of various scalars, there are still issues with homogeneity and stationarity that continue to be difficult to solve.

Because it directly measures the relevant scalars and their fluctuations over a given time

period, the EC technique doesn't depend on knowing what the gradient of that scalar is.

However, a close examination of this method reveals a serious theoretical and practical flaw that must be dealt with by various post-processing corrections.

After a Reynold's decomposition [*Reynolds, 1894*], the equation that describes the flux of any scalar is as follows:

Flux of X =
$$\rho \overline{w} \overline{X} + \rho \overline{w'} X'$$
 [3]

Where,

 ρ = the density of dry air (kg·m⁻³)

 \overline{w} = the mean vertical velocity of air (m·s⁻¹)

 \overline{X} = the mean of the quantity of interest

w' = the mean of the instantaneous fluctuations of the velocity of air (m·s⁻¹)

X' = the mean of the instantaneous fluctuations of the scalar

In order to maintain continuity, it could be assumed that the first term will equal zero and that all fluxes could be calculated with only the second term. However, there are various physical processes that can induce a mean vertical velocity and result in an apparent flux that can be larger than the flux measured by the second term [*Webb et al., 1980; Leuning et al., 1982; Feuhrer et al., 2001; Kyaw Tha Pau U, 2001; van Dijk, 2004*]. Therefore, additional measurements and calculations must be performed to attain an adequate degree of closure if one is trying to determine the flux of an atmospheric constituent such as CO₂. Without this closure, the measured flux will not accurately represent exchanges that are taking place through the underlying surface.

There are various common physical processes that can induce a mean vertical velocity

due to air density and pressure changes at the surface [*Webb et al., 1980; van Dijk, 2004*]. These include heat flux, evaporation, surface friction and sensor-tilt induced mean vertical velocity. It should, however, be strongly noted that if the equipment used in the eddy covariance method measured the mixing ratio and not the density of an atmospheric constituent, that heat and vapour fluxes would not be an issue [*Webb et al., 1980; Kyaw Tha Paw U, 2000; McGillis, 2001*]. After noise, spikes and suspected flow distortions have been removed from the dataset and the above mentioned corrections have been made to the raw data, there may still be issues with homogeneity and stationarity that need to be addressed through proper quality controls measures.

Before any data processing can take place, noise and unrealistic measurements should either be corrected or discarded from the dataset. 'Spikes' in the dataset typically appear as short duration, large amplitude fluctuations in the high frequency range of the spectrum and are often the result of random electronic surges or sonic transducer blockage during precipitation events [*Aubinet et al., 2000; Foken et al., 2004; Heusinkveld et al., 2008*]. Often, spikes can be removed by establishing realistic parameters or by using a standard deviation filter beyond which an individual data point will be removed or replaced. For instance, wind speeds along any axis that are in excess of 50 m·s⁻¹ or density measurements that are negative are either unrealistic or not physically possible. Interpolation methods based on point to point averaging can be used to identify these spikes and replace them with averaged values.

On a larger scale, the distortion of airflow around small, nearby obstacles will need to be corrected or the data may have to be eliminated from the initial dataset. Obstacles may include nearby buildings or abrupt changes in topography and, although EC is generally consider a nonintrusive method of mass and energy flux calculation, by the instrumentation and support

structures themselves [*van Dijk et al., 2004*]. This will result in systematic distortion of the air flow and the turbulence structures that advect past the EC instrumentation and cannot be adequately corrected for with the use of tilt corrections [*Wieringa, 1980*]. Both the obstacle and its wake can be modeled as a 3D ellipsoid and the flow around it can be considered as a timevarying homogenous potential flow for relatively large eddies. However, a less complex approach to this problem is to identify and discard data that have been affected by an obstacle through the use of a friction velocity (UST) analysis. Friction velocity can be used as a simple proxy for surface roughness and the time-average UST from each of the surrounding wind direction sectors (typically 36 sectors – 360 °/10) can be used to identify wind directions that generate consistently high UST values. These can be eliminated from the dataset in order to determine the best acceptance angle. This is the angular distance, centred on the EC array axis of symmetry, where wind approaching the sonic transducer array is either free from blockage by structural elements. Once spikes and flow distortions have been dealt with accordingly, far more fundamental corrections can then be applied to the dataset.

Fluxes of heat, water vapour can induce changes in air density at the surface that result in a pushing velocity of air past the height at which the sensor is placed. In the case of a heat flux induced pushing velocity, parcels of cool, relatively dense air travel downward past the sensor towards the surface. At the surface, they are heated, become less dense and rise past the sensor height. Similarly, the addition of water vapour to the air parcel at the surface will result in a pushing velocity by increasing the density of the air. As the air pressure equilibrates with its surroundings, the air will be pushed up in direct proportion to its initial, higher density. For instance, if 6 mm of water were to evaporate, then 6 m of pure water vapour are created and the

atmosphere will rise by a corresponding 6 m [*van Dijk*, 2004]. Fluctuations of density due to changes in temperature or humidity have also been referred to as the 'dilution effect' because increases in moisture or temperature will cause a decrease in the density of an atmospheric constituent such as CO_2 [*Fairall et al., 1999*]. The flux of momentum and pressure has the opposite effect on the vertical flow of air past the sensor height. As air horizontally advects past the sensor, it is affected by surface roughness and a downward flux of momentum is induced. As the air velocity is slowed at the surface, the density of the air is increased and a downward flux of mass is induced. This results in a corresponding downward flux of pressure that becomes increasingly important to the determination of the flux of CO_2 or other quantities with increasing wind velocity. This may be corrected with a modification of the WPL correction [*Webb et al., 1980*; *Massman and Lee, 2002*]:

$$FC = \overline{w'\rho_{c}'} + \overline{\rho_{c}} \left(1 + \overline{\chi_{v}}\right) \left[\delta_{oc} \frac{\overline{w'T'_{a}}}{\overline{T_{a}}} - \frac{\overline{w'p'_{a}}}{\overline{p_{a}}} \right] + \overline{\sigma_{c}\mu_{v}w'\rho'_{v}}$$
[4]

Where,

FC = the flux of CO₂ (μ mol·m⁻²·s⁻¹)

 $\overline{\chi}_{\nu}$ = the volume mixing ratio or mole fraction for water vapour (= $\overline{p}_{\nu} / \overline{p}_{a}$)

 $\delta_{oc} = 1$ for an open path sensor and 0 for a closed path sensor

 p_a = ambient pressure (Pa)

- $\overline{\rho}_{c}$ = the mean density of carbon dioxide (µmol·m⁻³)
- σ_c = the mean mass mixing ratio for CO₂ (= $\overline{\rho}_c / \overline{\rho}_a$)

 μ_v = the ratio of the molecular mass of dry air to water vapour (= m_d / m_v)

This correction is necessary in all forms of micrometeorological measurements whether

they are made using eddy correlation or K-theory [*McGillis et al., 2001*]. The net affect of the application of this correction that is generally observed is to reduce the measured downward flux of CO_2 during the daytime when there tends to be positive sensible and latent heat fluxes. Interestingly, this correction is frequently larger than the original measured flux and indicates the need for its individual components to be carefully measured and included in the final flux estimation.

Instrument tilt relative to the surrounding terrain is another source of mean vertical velocity that must be corrected before the determination of a flux from a surface can be accurately determined. It is essentially impossible to perfectly align a sonic anemometer with terrain-induced wind streamlines that may not be running parallel with the surface. Even if the wind streamlines run perfectly parallel to the surface, it is impractical to expect that there will never be instrument misalignment [*van Dijk et al., 2004*]. Although this kind of misalignment may be extremely small, any tilt induced mean vertical flow can result in a significant flux when the measurements are integrated over long time periods. Therefore, it is always necessary to perform a tilt correction with the use of a coordinate rotation or a 'planar-fit' that will orient the vertical axis of the anemometer with the 'true' vertical direction [*van Dijk et al., 2004*; *Aubinet et al., 2000; Stull, 2000*].

Coordinate rotations begin with yaw and pitch corrections that result in the alignment of the *u* vector with the mean wind velocity [*Aubinet et al., 2000*]. The yaw correction is determined by simply placing the mean horizontal wind direction along the first coordinate axis. Although this may seem like a highly arbitrary frame of reference, the basic laws of turbulent exchange are not violated [*van Dijk et al., 2004*]. This automatically forces mean *v* to zero and

the covariance between *u* and *v* becomes zero as well (uv = 0). Similarly, when the pitch correction is applied there can no longer be any covariance between *u* and *w* (uw = 0). The final rotation is a roll correction around the x-axis which nullifies the final lateral flux density between *v* and *w* (vw = 0).

A practical problem with the use of the eddy covariance method is its need for horizontal homogeneity and statistical stationarity and their absence from most surfaces. Both of these problems affect the degree of closure that is needed by this method in order to determine a flux and can be difficult to correct for. As a result, various quality control techniques such as filtering, and steady state tests must be employed to ensure the best possible estimates of mass and energy exchanges. In addition to this, there are other forms of quality control that will need to take into account issues such as energy balance closure (EBC) analysis, flow distortion and corrections due to night-time stable conditions that may necessitate the use of gap-filling measures which employ methods other than eddy covariance.

The statistical analysis associated with eddy covariance techniques demands that the properties of eddies responsible for any particular flux are a function only of height and time [*Kaimal and Finnigan, 1994*]. This necessitates horizontal homogeneity throughout the area surrounding the instrumentation which is a condition that is almost never satisfied. Even surfaces that appear homogenous at a particular point in time are likely to change in response to short and long term trends and strongly depends on the time averaging period that is used. These trends can be as simple as a change in statistical properties with a change in wind direction and weather. If such a change in wind direction were to occur *in the middle of an averaging period*, the resulting flux calculations may not be valid because the profiles of wind speed, temperature

and other scalars may lag behind the change in wind direction and will not be in equilibrium with the new surface that is represented by the new direction. This is closely related to the concept of statistical stationarity wherein the statistical properties of the variables associated with eddies should not change with time if fluxes are to be measured using eddy covariance [*Panofsky and Dutton, 1984*]. This suffers from the same lack of realism that the assumption of horizontal homogeneity does. Over the course of a day, there will likely be profound changes in the statistical nature of the eddies due to diurnal rhythms and the passage of weather systems. This will make it impossible to calculate fluxes using the eddy correlation method because of the presence of long terms trends in the dataset. Such trends represent a lack of closure and need to be removed from the dataset through the use of high-pass filtering or other trend removal techniques before a flux can be estimated.

Because the eddy covariance method depends on the use of averaging periods, its validity will naturally be compromised by the presence of long term trends in the data. These trends occur in the form of diurnal rhythms, the passage of weather systems or any other frequency component with a period longer than the record length [*Kaimal and Finnegan, 1994*]. It is generally assumed by micrometeorologists (rightly or wrongly) that such slow fluctuations in the signals are not related to transport phenomenon [*van Dijk et al., 2004*] and can be removed. As a consequence, their removal from the spectral curve with the use of high pass filtering will result in a reduction of the measured flux [*Aubinet et al., 2000*].

Another issue concerns the length of the averaging period that is used for the EC calculations and whether or not the data have remained statistically 'stationary' over the course of that period [*Aubinet et al., 2000; Mauder and Foken, 2004*]. A stationarity test developed by

Gurjanov et al. [1984] can be used to compare the statistical parameters determined for an individual averaging period with shorter intervals within this period with the following methodology: First, the averaging interval in question is divided into four to eight equal segments and the covariance of two measured signals (typically w and some other scalar, x) is determined for the averaging interval with the following equation:

$$\overline{w'x'}_{INT} = \frac{1}{M \cdot N - 1} \left[\sum_{i} \left(\sum_{j} w_{j} \cdot x_{j} \right) - \frac{1}{M \cdot N} \sum_{i} \left(\sum_{j} w_{j} \cdot \sum_{j} x_{j} \right) \right]$$
[5]

Where,

M = the number of sub-intervals under consideration within the averaging interval N = the number of measured points or records within the averaging interval (for instance, a 20 Hz system will generate 72,000 records per hour.)

Secondly, the covariances of each of the sub-intervals is determined with the following equation:

$$\overline{w'x'}_{SUB} = \frac{1}{N-1} \left[\sum_{j} w_{j} \cdot x_{j} - \frac{1}{N} \left(\sum_{j} w_{j} \cdot x_{j} \right) \right]$$
[6]

Lastly, a time series is considered to be steady state if the difference between both covariances is lower than 30%:

$$RN_{Cov} = \frac{\left| (w'x'_{SUB}) - (w'x'_{INT}) \right|}{(w'x'_{INT})}$$
[7]

Realistically, sub-intervals with a difference of less than 30 % from the averaging interval are considered to be of high quality and differences of up to 60 % are frequently considered to be of acceptable quality [*Foken et al., 2004*]. If this condition is not satisfied, the dataset will need to be examined and a different averaging interval may need to be considered. For instance, if there appears to be an abrupt step change in the middle of the period, a shorter interval may need to be

considered or the data period may need to be discarded.

Once all of the necessary corrections have been applied and a thorough preliminary analysis of the dataset has been accomplished, it will be possible to reduce the sheer size of the raw data by calculating the half hourly or hourly fluxes. This will produce a manageable dataset that is relatively easy to analyze and from which one can begin to draw conclusions. Unfortunately, there are peculiarities of turbulent exchange, aerodynamics and the instrumentation used that can either complicate the estimation of the flux of CO₂. These issues will continue to necessitate the recalculation of fluxes or the elimination of individual datasets.

Among one of the more pertinent issues of turbulent exchange is the issue of energy balance closure (EBC). According to the first law of thermodynamics, there must be a conservation of both mass and energy as these quantities are exchanged and altered between the surface and the atmospheric boundary layer. This is the basis of the Navier-Stokes equations (see Appendix A, *Flux Estimation Methods*) and the goal of any measurement system is to be able to provide estimates that are as close to this law as possible. Therefore, as long as stationarity and homogeneity can be assumed, the sum of all turbulent fluxes of energy that occur between the sensor height and the surface must equal the sum of the radiative flux of energy that is inputted into this layer and the amount that is stored:

$$RNET = H + LE + G + S$$
Where,
$$RNET = \text{Net radiation (W·m-2)}$$

$$H = \text{Flux of sensible heat (W·m-2)}$$

$$LE = \text{Flux of latent heat (W·m-2)}$$

$$G$$
= Ground heat flux (W·m⁻²)

Both *RNET* and *G* can be measured by relatively simple and reliable sensors that are independent of the EC system and can be used to check the plausibility of the data gathered with the EC system by comparing the amount of radiative energy inputted onto the surface (*RNET-G*) with the amount that is used for turbulent sensible and latent heat transport (H+LE) [*Twine et al.*, 2000; *Aubinet et al.*, 2000]:

$$D = \frac{(H + LE)}{(RNET - G)}$$
[9]

Where,

D = the potential discrepancy between the ratio of (H+LE) and (RNET-G)

If there were perfect energy balance closure, then D would be unity; the amount of energy that was available was used in its entirety and all energy has been accounted for. However, D is typically up to 30% less than unity and the use of an energy balance closure to evaluate EC measurements is open to a variety of criticism including the following:

- The energy storage term is not generally subtracted from the total energy available for turbulent exchange [*Aubinet et al., 2000*]. This may be difficult to measure for a variety of reasons, but could lead to overestimation of the denominator in equation [9] which would result in an underestimation of D that would not reliably constitute conclusive evidence for erroneous turbulent flux measurement.
- Surface inhomogeneities can, and frequently do exist between the surface underlying the radiation measurement sensors and the footprint of the surface that is advecting the turbulent air past the EC system. Each surface may have different properties that will

affect the radiation balance and turbulent exchanges and a comparison between the turbulent exchange from one surface and the radiation balance of the other may not be entirely valid. Although the footprint measured by the EC system may be spatially representative of the surrounding surface by integrating the individual terrain units, the 'fetch' of the radiation sensors is generally up to a couple or orders of magnitude smaller. As such, the radiation measurements will be biased and the size of this bias could be large if surface properties vary on the same scale of the radiative fetch (tens of m²).

Related to problems with the estimation of the storage term is the temporal integration of the measurements that are used to perform EBC. It has been observed [*Malhi et al., 2002*] that extending the averaging period from 1 hour to 4 hours improved the energy balance closure to nearly 100 %. It was surmised that the inclusion of turbulent fluxes from low frequency transport may have solved the EBC problem. However, it should be recognized that the storage of radiative energy may be released on much longer time scales than an hourly averaging period and the use of this energy to melt or refreeze frozen soil may result in this energy being stored and released on seasonal time scales. Thus, the effectiveness of the EBC method may be closely related to the averaging period not only of the EC system itself, but of the averaging period used with the EBC technique.

Another issue with EC system that calls into question the validity of the flux measurements is atmospheric stability. EC systems work because they are able to measure turbulence and its covariance with the fluctuations of a scalar quantity. During the night, aerodynamic conditions tend to be stable and the atmosphere tends to be highly stratified which

results in a significant dampening of the large scale turbulent fluctuations of eddies [Massman and Lee, 2002; Aubinet et al. 2000]. This presents serious limitations to the use of EC systems due to low turbulence issues, the storage of CO₂ below the instrumentation and the presence of unusually large footprints. In the near absence of turbulence, flux estimations may be negligible. Conceptually, this may appear to be an acceptable estimation, but it is important to realize that gas exchanges may still be taking place at the near surface and past the instrument height [Aubinet et al., 2000]. Depending on the surface temperature, respiring organisms may be expelling CO_2 into the atmosphere and, in the absence of turbulent mixing, it will be stored in this layer or will be slowly flushed away from the surface of interest below the instrument height and will not be measured. Additionally, very small, high-frequency eddies may transport CO_2 and other quantities of interest past the instrumentation that cannot be measured due to its insufficient frequency response. It is important to be able to determine what this flux is, especially if one is trying to derive a carbon or energy budget for the environment of interest. However, it may be possible to replace the measurements taken during these periods by deriving other relationships between environmental variables and known fluxes and applying them to night time conditions [Aubinet et al., 2000].

With few exceptions [*Yi et al., 2004*], most researchers go to great lengths to ensure that the estimated fetch of their sensors is less than the distance to any obvious boundaries or inhomogeneities on the surface. However, due to the intrinsic nature of Subarctic environments and the methodological objectives of this particular research project, such homogeneity cannot be assumed and, in fact, surface inhomogeneity is to be expected. Especially in the case of the intertidal environment of the Coastal sites where the surface undergoes a radical transformation

from being flooded with water to being drained on a daily basis, it is important to know that the scale of the measurements is proportional to the scale of the fluxes from the surfaces in question in order to attain our objectives.

2.3 - Application and Data Post-Processing

Many of the previously mentioned corrections and quality control methods were used to ensure that the datasets would be as representative as possible of their respective environments which allowed the data to be integrated and reliably compared.

Once the raw data from each of the stations were saved onto the telemetry computers, the low frequency data were separated from the high frequency EC data and were organized into daily files. The flux estimation program EdiRe[©] was then used to calculate all mass and energy half-hourly fluxes and to record them in daily tables along with measurements of the ambient densities of CO_2 , H_2O and other variables. These tables were then combined with the low frequency data and the entire dataset for each station was combined into a yearly file.

The same noise filters were applied to the datasets from each of the stations. Any raw data values that were beyond three standard deviations of the hourly mean of the dataset were removed and replaced with interpolated values. Additionally, unrealistic measurements were excluded from the half-hourly datasets from each of the sites based on identical parameters. For instance, all orthogonal windspeeds over 50 m·s⁻¹ and all CO_2 density measurements over 1000 mmol·m⁻³ were discarded from the dataset. If more than one quarter of the data from an hourly dataset was missing or had to be discarded, then that half-hour increment would not be used for flux calculations.

Identical density and tilt corrections were applied to each site. Density corrections were applied according to Massman and Lee [2002] with equation [4] and trigonometric rotations were applied around each orthogonal axis according to Stull [2000]. The planar-fit method was not used as a tilt correction at any of the sites due to the rapidly changing roughness and height of the intertidal surface at each site and a standard coordinate rotation was used instead. Despite the height of the EC platforms and the very high sampling rate, spectral corrections were used at the Coastal sites and resulted in an increase in the size of each of the flux estimates. A comparison of spectrally corrected and uncorrected values revealed that there was an average of a 12 % loss of total energy without this correction thus validating its use.

EBC would have been a useful quality control test at the Coastal sites but there were two factors that prevented this: two heat flux plates were deployed at the Ithaca during August, 2005, but the data could not reliably be used because the plates were repeatedly pulled out of the substrate due to tidal action and visits from *Ursus maritimus*. As well, the heat flux plates could not be deployed at either site due to the fact that the surface that was generating the turbulent fluxes (the intertidal flats or the near-shore environment) was radically different from the nature of the surface in the vicinity of the tower and was too far away to place any sensors into.

Although each of the sites presented problems with the acceptance angle, the Ithaca by far presented the greatest challenge. Given the fact that the Ithaca is a very large structure in the middle of an otherwise flat surface environment, it was easily surmised that the ship itself was going to generate flow distortions that must be measured and accounted for. A friction velocity test was performed on all of the data that were collected between June and August, 2005 wherein the friction velocity from each of the 36 sectors ($360 \circ /10 \circ$ sectors) was averaged. A predictable



Figure 2.4 The Ithaca friction velocity test. This test was performed on the dataset from the Near-Shore site and indicated a relatively high average wind shear when the wind was blowing along the length of the Ithaca. Higher than normal values of friction velocity were considered to represent flow distortions generated by the ship. This data was rejected from the further analysis because such turbulence would result in calculated fluxes that would not be indicative of surface processes.

pattern readily emerged in Figure 2.3 which clearly demonstrated that winds that had to blow across the length of the ship before they reached the EC system had nearly twice the shear force as those that did not. These high-friction-velocity sectors were clustered around a 60 ° interval located to the SSW of the EC system and all flux data that was measured from winds that were blowing from this sector were excluded from the dataset. An additional 10 ° were

added to both ends of this interval as a margin of safety and the acceptance angle was comprised of the other 280 °. However, an implication of this was that the tidal flats surface to the south of the Ithaca would only be measureable through a 100 ° acceptance angle that would be divided between easterly and westerly wind directions along the low tide mark. Measurements from these narrow windows were not deemed to be representative of the tidal flats surface as a while and were rejected. Therefore, the rejection angle of the Ithaca dataset comprised the entire 180 ° sector to the south of the ship. The presence of a small (10 - 12 m) escarpment that runs parallel to the shore to the immediate south of the Bluffs site resulted in a similar 180 ° rejection angle.

After all of the corrections had been applied and the quality control measures had

discarded potentially invalid data, the size of the available dataset from each of the sites had been reduced. This was reduced further by telemetry transmission problems. Over the course of the sampling periods in 2005 and 2006, the flux data acquisition success rate was 33 % and 32 % at the Near-Shore and the Tidal Flats site respectively. The low frequency data did not suffer from the same types of problems that the EC systems did and the data acquisition success was often close to 100 % depending on the variable in question.

Gap filling measures were not used for two reasons: Firstly, the intertidal environment made it too difficult to derive any temperature/soil moisture relationships. Secondly, the overall goal of these studies was not to provide a seasonal budget of a near-coastal or intertidal zones, but to document the observed mass and energy fluxes under a variety of forcing factors in order to develop a process-level understanding of their affects on impacts on surface exchanges. Once a better understanding is acquired, it may then be possible to derive the models necessary to be able to effectively fill in data gaps and estimate mass and energy budgets for sub-Arctic near-shore and intertidal surfaces. However, some of the low frequency data such as wind direction, wind speed and air temperature data were used from the nearby 'Churchill A' Environment Canada weather station to fill in these gaps. A comparison of these data with existing data gathered from the each of the sites demonstrated that the measurements taken at the Environment Canada weather station were generally very close to what had been gathered at the research sites used for this project.

2.4 - Modeling net radiation

Due to the absence of upwelling radiation instrumentation at the Tidal Flats Station, RNET

needed to be modeled based on radiation data from the Near-Shore Station that was developed from a combination of previously devised long wave radiation models and their respective estimations of atmospheric emissivity.

Incoming and outgoing long wave radiation (Rl and Rl_out respectively) are critical components of the radiation balance of any surface:

$$RNET = (Rg - Rg _ out) + (Rl - Rl _ out)$$
[10]

Where,

 $Rg = Incoming Short Wave Radiation (W \cdot m^{-2})$

Rg_out = Outgoing (Reflected) Short Wave Radiation $(W \cdot m^{-2})$

 $Rl = Incoming Long Wave Radiation (W \cdot m^{-2})$

Rl_out = Outgoing (Reflected) Long Wave Radiation $(W \cdot m^{-2})$

If Rl or Rl_out aren't directly measured with pyrgeometers, then they can be calculated with the use of the Stefan-Boltzmann equation and knowledge of the emissivity and the temperature of the surface or volume in question:

$$Rl = \xi \sigma T^{4}$$
 [11]

Where,

 ξ = The emissivity of the object or surface in question (unitless)

 σ = The Stefan-Boltzmann Constant (5.67·10⁻⁸ W·m⁻²·K⁻⁴)

T = The temperature of the object or surface or volume in question, often surface temperature (TSURF) and ambient air temperature (T_a) (° K)

 T_a and TSURF are commonly measured and readily available but the emissivity requires a knowledge of the inherent properties of the surface or volume. Surfaces that are composed of

water or a muddy substrate have readily documented emissivities that can be applied to complete equation [11] [*Oke*, 1987]. However, the emissivity of a volume of air requires measurements of water vapour pressure (e), cloud cover fraction and height, cloud properties, trace gas content and concentrations of aerosols [*Flerchinger et al.*, 2009; Jin et al., 2006; Best , 1998]. Often, only measurements of e, cloud cover fraction and T_a are available from which numerous models have been devised that have been applied to various locations with varying degrees of success.

The development of a model that could be used to generate reliable estimates of RNET at the Bluffs began with developing a regressed relationship between TSURF and T_a for each of the four primary tidal stages: 'PEAK', 'FLOW', 'EBB' and 'TROUGH' and day or night conditions. Each general tidal stage was based on data for this part of Hudson Bay that were generated by the Canadian Hydrographic Service [2010]. The 'PEAK' stage was considered to have occurred when the water was near its maximum height for a given tidal cycle and the 'TROUGH' stage occurred when a given tidal cycle was near its minimum. The 'EBB' stage was considered to have occurred when the water was flowing from the 'PEAK' to the 'TROUGH' stage and the 'FLOW' stage occurred when the water was going in the opposite direction. Since both the

Diurnal and Tidal State	Regression of Ta to TSURF
DAY_EBB	y = 0.943x + 2.669 (r ² : 0.95, SE: 0.0001)
DAY_FLOW	y = 0.983x + 1.989 (r ² : 0.94, SE: 0.0001)
DAY_PEAK	y = 0.947x + 2.608 (r ² : 0.93, SE: 0.0001)
DAY_TROUGH	y = 0.963x + 2.267 (r ² : 0.94, SE: 0.0001)
NIGHT_EBB	y = 1.001x + 0.536 (r ² : 0.86, SE: 0.0001)
NIGHT_FLOW	y = 0.967x + 0.930 (r ² : 0.97, SE: 0.0001)
NIGHT_PEAK	y = 0.961x + 0.832 (r ² : 0.80, SE: 0.0001)
NIGHT_TROUGH	y = 0.955x + 1.113 (r ² : 0.95, SE: 0.0001)
Table 2.1 Surface (T regressions generated)	SURF) and air temperature (Ta) from data from the Near-Shore Station.

'PEAK' and 'TROUGH' stages are discrete points along the tidal cycle (the water does not conveniently stay at its maximum or minimum for a definable

time interval,) all daily 30

minute sampling intervals were divided equally among each of the stages based on their proximity to the maximum or minimum water height and the water flow direction. The regressive relationships that were developed based on data from the Near-Shore Station were then used to estimate TSURF for the tidal flats at the Bluffs from which Rl_out was calculated using equation [11] (Table 2.1). It was found that there was always a very close and positive relationship between T_a and TSURF but that these regressed relationships did not change appreciably with tidal stage or the time of day. Four different models of atmospheric emissivity

	Application of Clear Sky Models									
	MBE	RMSE	m	b	r ²					
Xin et. al. [2006]	4.7	50.4	1.2059	-27.8	0.987					
Niemela et al. [2001]	10.0	52.2	1.2100	-23.2	0.986					
Idso [1980]	2.7	50.5	1.2056	-29.8	0.986					
Angstrom [1918]	-34.7	123.3	1.2074	-67.4	0.987					
	Applicatio Based on	n of Cloud Clear Sky	y Sky Mode Model: Xir	els n et. al. [20	0061					
	Applicatio Based on	n of Cloud Clear Sky	y Sky Mode Model: Xir	els net. al. [20	006]					
	Applicatio Based on MBE	n of Cloud Clear Sky I RMSE	y Sky Mode Model: Xir m	els n et. al. [20 b	006] r ²					
Sugita & Brutsaert [1993]	Applicatio Based on MBE 13.1	on of Cloud Clear Sky RMSE 59.1	y Sky Mode Model: Xir m 1.2243	els n et. al. [20 b -26.0	006] r ² 0.986					
Sugita & Brutsaert [1993] Keding [1989]	Applicatio Based on MBE 13.1 24.8	n of Cloud Clear Sky RMSE 59.1 63.34	y Sky Mode Model: Xir m 1.2243 1.2258	els net.al.[20 b -26.0 -14.5	006] r ² 0.986 0.986					
Sugita & Brutsaert [1993] Keding [1989] Brutsaert [1982]	Applicatio Based on MBE 13.1 24.8 48.1	n of Cloud Clear Sky RMSE 59.1 63.34 75.7	y Sky Mode Model: Xir <u>m</u> 1.2243 1.2258 1.2276	els n et. al. [20 b -26.0 -14.5 8.5	006] r ² 0.985 0.985 0.985					
Sugita & Brutsaert [1993] Keding [1989] Brutsaert [1982] Jacobs [1978]	Applicatio Based on MBE 13.1 24.8 48.1 55.4	n of Cloud Clear Sky RMSE 59.1 63.34 75.7 80.8	y Sky Mode Model: Xir 1.2243 1.2258 1.2276 1.2283	els n et. al. [20 -26.0 -14.5 8.5 15.6	006] r ² 0.986 0.986 0.986 0.985					

Table 2.2 Clear sky emissivity (ξ clr) and cloudy sky emissivity (ξ cld) models. Clear sky models of emissivity are used as inputs into cloudy sky models along with estimates of cloud fraction (c). The coefficients in each of these equations were empirically derived in the respective studies. The implication of this is that different models of clear sky emissivity can be used in conjunction with different models of cloudy sky emissivity.

for clear sky conditions were evaluated (Table 2.2). By calculating RNET using these models and comparing the results to the measured RNET during clear sky conditions, it was found that the models of Idso [*1980*] and Xin et al. [*2006*] had the closest fit based on a combination of low mean bias errors (MBE) and root mean square error (RMSE) (Table 2.3). The emissivity

calculated during clear sky conditions is generally used as the primary input into models that describe emissivity during cloudy sky conditions. As such, both of these models were then used as potential inputs into five models of emissivity empirically formulated for cloudy sky conditions. The other primary input, cloud fraction, had to be estimated based on hourly data from the 'Churchill A Station'. Because this station is 19 km away from the study sites and provides only very general categories to estimate cloud cover fraction, this could be a large

Application of Clear Sky Models										
	MBE	RMSE	m	C	r ²					
Xin et. al. [2006]	4.7	50.4	1.2059	-27.8	0.987					
Niemela et al. [2001]	10.0	52.2	1.2100	-23.2	0.986					
Idso [1980]	2.7	50.5	1.2056	-29.8	0.986					
Angstrom [1918]	-34.7	123.3	1.2074	-67.4	0.987					
	Applicatio Based on	n of Cloud Clear Sky	y Sky Mode Model: Xir	els n et. al. [20	006] r ²					
Sugita & Brutsaert [1993]	13.1	59.1	1 2243	-26.0	0.985					
Kedina [1989]	24.8	63.34	1.2258	-14.5	0.985					
Brutsaert [1982]	48.1	75.7	1.2276	8.5	0.985					
Jacobs [1978]	55.4	80.8	1.2283	15.6	0.985					
Maykut & Church [1973]	27.4	65.2	1.2264	-12.0	0.983					
000 (4 880) / 000 (/ 1 6 4 6 6 7 6 7 8 7 6 7 7 8 7 6 7 7 8 7 6 7 7 8 7 6 7 7 8 7 6 7 7 8 7 6 7 7 8 7 7 8 7 8	Based on MBE	Clear Sky RMSE	Model: Ids m	o [1980] c	r ²					
Sugita & Brutsaert [1993]	10.9	58.7	1.2242	-28.1	0.985					
Keding [1989]	22.6	62.48	1.2256	-16.7	0.985					
Brutsaert [1982]	45.7	74.2	1.2274	6.1	0.985					
Jacoba [4070]	52 9	79.1	1.2281	13.2	0.985					
Jacobs [1976]										

Table 2.3 Development of an algorithmic model for RNET at the Ithaca. Incoming long wave radiation was estimated for clear sky conditions based on four different models of clear-sky emissivity. RNET was then calculated and compared against measured RNET in order to evaluate the accuracy of each of the models in this environment. The models by Xin et al. [2006] and Idso [1980] proved to be the most accurate. These were then used as inputs into five different models for cloudy sky emissivity, each of which was then used to calculate RNET for a comparison with actual RNET.

these models. However, reasonably good estimations of RNET were generated by using a combination of the Idso [1980] emissivity model for clear sky conditions and the Sugita and Brutsaert [1993] emissivity model for cloudy sky conditions with a MBE of 10.9 W·m⁻² and a RMSE of 58.7 W·m⁻². A recurring problem with each of the

potential source of error in

model combinations were the slopes (m) of the linear regressions; they very consistently

averaged 1.22 (Table 2.3) thus indicating that there could always be an overestimation of RNET using these models depending on the offset (c). To complete the calculation of the energy balance for the Bluffs using equation [**9**], the average albedo for each tidal stage and clear or cloudy sky conditions was used to estimate Rg_out. These average albedo estimates were then used to calculate Rg-out at the Near-Shore Station by multiplying it with the measured Rg.

Despite the sources of error, reasonable and coherent estimates of RNET were formulated for the Tidal Flats Station using only Ta, e, Rg, tidal stage and cloud cover fraction and models that had been developed based on the observed radiation balance at the Near-Shore Station. Thus, an analysis of the forcings and fluxes from the intertidal flats in front of the Bluffs site could proceed.

3 - Results

3.1 - The Near-Shore Zone, 2005

3.1.1 - Seasonal meteorology and forcings

3.1.1.1 – *Air and Surface Temperatures*

The time series of Ta, TSURF and temperature differences (delT) were generally continuous and contained very few gaps. Because there was a very close relationship between the temperature data from the AWS on the Ithaca and the data from the 'Churchill A Station', this data was used to fill in missing values. Ta averaged 11.2 °C for the sampling period with average daily temperatures only 1.1 °C warmer than the nights (Table 3.1). Maximum daily temperature rose

NEAR-SHORE, 20	05 - KEY	FORCING	AVERAG	GES and ST	ABILITY	PARAMET	ER MEDI	AN	
	Ta	TSURF	delT	RG	VPD	UST	WS	z/L	zo
	(°C)	(°C)	(°C)	(W·m⁻²)	(kPa)	(m·s⁻¹)	(m·s⁻¹)	(median)	(m)
AVERAGE	11.2	12.1	-1.6	340.9	0.51	0.25	4.94	0.01	0.034
DAY	11.5	12.8	-1.9	353.5	0.58	0.27	5.06	0.00	0.037
NIGHT	10.4	10.3	-0.6	2.6	0.34	0.20	4.63	0.05	0.023
CLEAR	10.0	11.2	-1.7	317.9	0.39	0.20	4.36	0.00	0.040
CLOUDS	11.7	12.5	-1.6	350.8	0.57	0.28	5.20	0.01	0.031
FLOW	11.2	12.2	-1.5	322.2	0.51	0.20	5.01	0.01	0.020
PEAK	11.3	12.2	-1.7	319.3	0.55	0.28	4.85	-0.01	0.035
EBB	11.4	12.3	-1.6	371.0	0.53	0.27	4.96	0.01	0.034
TROUGH	10.9	11.7	-1.5	355.4	0.47	0.26	4.94	0.02	0.045

Table 3.1 Near-Shore Station, 2005 - Key Forcing Averages. Each of the columns are averages with the exception of z/L which is a median value.

from 5 °C to 25 °C by DOY 192. There appeared to be a heat wave from DOY 182 to 192 with a difference of 10 °C between daily maximums and nightly minimums. The air temperatures were variable for the remainder of the sampling period but averaged between 10 and 15 °C. Notably, cloudy conditions were warmer than clear conditions during both the day and the night (Table

3.2).

Over the course of the sampling period, TSURF averaged 12.1 °C and tended to be 1.6 °C

warmer than the overlying air. This difference was confirmed through measurements taken by two different instruments, a precision infrared pyrgeometer and an infrared transducer (see

NEAR-SHORE, 2005 - KEY FORCING AVERAGES and STABILITY PARAMETER MEDIAN									
	Ta	TSURF	delT	RG	VPD	UST	WS	z/L	zo
	(°C)	(°C)	(°C)	(W·m⁻²)	(kPa)	(m·s⁻¹)	(m·s⁻¹)	(median)	(m)
DAY_CLEAR	10.2	11.8	-2.0	322.6	0.42	0.25	4.50	-0.02	0.040
DAY_CLOUDS	12.0	13.2	-1.9	367.2	0.64	0.28	5.29	0.00	0.036
NIGHT_CLEAR	9.6	9.7	-0.8	7.7	0.31	0.06	4.03	0.08	0.039
NIGHT_CLOUDS	10.7	10.6	-0.5	1.9	0.36	0.26	4.93	0.04	0.017

Table 3.2 Near-Shore Station, 2005 – Day/Night, Clear/Cloud Forcing Averages. Each column is an average with the exception of z/L which is a median value.

Appendix A, '*Hardware*'). Fluctuations of TSURF appeared to closely parallel fluctuations in Ta and the r^2 of the regressed relationship between Ta and TSURF ranged between 0.89 and 0.94. There were, however, minor variations between Ta and TSURF based on the time of day and delT was the smallest during the night (Table 3.1). A t-test that was performed between the TSURF observed during each tidal stage and the TSURF observed during all tidal stages indicated that there were no significant differences with tidal state at p>0.10.

3.1.1.2 – Radiation Components

Data acquisition success rates for Rg was close to 100 % and contained very few gaps. During the daytime, there was a very close relationship between RNET and Rg ($r^2=0.88$ to 0.97). Although RNET was not used for further analysis, it averaged 150 W·m⁻² and was highest during the daytime with an average of 218.0 W·m⁻² and lowest during the nighttime with an average of -33.7 W·m⁻². Average Rg was 317.9 and 350.8 W·m⁻² during clear and cloudy sky conditions respectively (Table 3.1). These counter-intuitive results were carefully scrutinized and will require further explanation. Small positive values of Rg during nighttime conditions are due to post-sunset and pre-dawn conditions.

3.1.1.3 – *Humidity*

Relative humidity (RH) measurements over the course of the sampling period were very coherent and continuous with almost no gaps. RH averaged 89% over the summer and nightly maximums remained consistently high. Daily minimums were highly variable and gradually rose throughout the sampling period. However, VPD varied considerably on a diurnal basis and in response to sky conditions; it was greatest during the daytime at 0.58 kPa and lowest during the night at 0.34 kPa (Table 3.1). Interestingly, cloudy sky conditions resulted in a higher average VPD than clear sky conditions at 0.57 and 0.39 kPa respectively (Table 3.2). The high VPD during cloudy conditions was especially pronounced during the day; daytime cloudy conditions produced the higher VPD of 0.64 kPa.

3.1.1.4 – Wind and Stability

The stability parameter (z/L) indicated that neutral or stable conditions prevailed throughout most of the sampling period and under nearly any combination of forcing conditions (Table 3.1). Conditions tended towards instability only during average daytime and clear sky conditions. In general, z/L did not vary much with the time of day. The average friction velocity (UST) was $0.25 \text{ m}\cdot\text{s}^{-1}$ throughout the sampling period and dropped to $0.20 \text{ m}\cdot\text{s}^{-1}$ during clear sky conditions at night. Wind speeds (WS) were generally very high throughout the sampling period with an average of 4.94 m·s⁻¹ (Table 3.1). WS tended to be the highest during the day time and during cloudy sky conditions; when both conditions occurred at the same time, average WS peaked at 5.29 m·s⁻¹ (Table 3.2). Roughness lengths (zo) varied between 0.023 m and 0.45 m under a range of conditions and averaged 0.034 m over the course of the sampling period. There were significant differences (t-test p-values <0.10) between day and night conditions with average roughness lengths of 0.037 and 0.023 m respectively. As well, zo tended to be very low (0.020 m) during the 'FLOW' tidal stages and were very high (0.045 m) during the 'TROUGH' tidal stage.

3.1.2 - Seasonal Flux Estimates, Energy Partitioning and Uncertainty

It should be noted that due to the 180 ° rejection angle of the EC platform and the exclusion of all

NEAR-SHORE, 2005 - FLUXES AND STORAGE											
		Н			LE			FC		S	
	(W·m ⁻²)	±	(p-value)	(W·m ⁻²)	±	(p-value)	(µmol·m ^{-z} ·s)	±	(p-value)	(W·m ⁻²)	
AVERAGE	-3.4			23.1			-0.11			*	
DAY	-1.7	1.0	0.241	26.0	1.1	0.019	-0.22	0.02	0.000	*	
NIGHT	-7.8	1.0	0.000	15.6	1.2	0.000	0.17	0.05	0.000	*	
CLEAR	-0.4	1.2	0.045	26.3	1.3	0.021	-0.13	0.08	0.645	*	
CLOUDS	-4.8	1.0	0.210	21.6	1.1	0.417	-0.11	0.03	0.937	*	
FLOW	-4.2	1.5	0.566	22.9	1.7	0.957	-0.21	0.04	0.026	*	
PEAK	-2.6	1.7	0.743	28.6	1.8	0.003	-0.18	0.04	0.119	*	
EBB	-0.9	1.7	0.199	22.9	1.8	0.955	0.02	0.04	0.008	*	
TROUGH	-5.9	1.4	0.077	18.0	1.6	0.009	-0.07	0.05	0.549	*	

Table 3.3 Near-Shore Station, 2005 – Mass and energy fluxes. Values for energy storage have been omitted (*) because they could not be accurately estimated at this station.

southerly winds, data acquisition success was as low as 33 % for most of the estimated fluxes. An analysis of the standard error of the mean indicated that H and FC were not significantly different from zero during the 'EBB' stage (Table 3.3). As well, a t-test performed between the mean estimates of H and LE for each tidal stage and the entire sample indicated that there were very few significant differences due to tidal stage within a reasonable degree of certainty (p>0.10). An exception to this was the 'TROUGH' stage; estimates of H and LE during this stage were significantly different (p<0.10). FC during clear or cloudy conditions could not be statistically distinguished from one another (p>>0.10). However, when the data was sorted for both day versus night and clear versus cloudy conditions, each grouping of FC estimates could be readily distinguished from one another (p<0.10).

3.1.2.1 – Sensible Heat Fluxes

In general, H was negative throughout the summer and the study area was a small sink for sensible heat (Table 3.3). H began strongly negative and remained steady throughout the summer but there is an indication that the study area may have been returning to being a source of

NEAR-SHORE, 2005 - FLUXES AND STORAGE											
		Н			LE			FC		S	
	(W·m ⁻²)	±	(p-value)	(₩·m ⁻²)	±	(p-value)	(µmol·m ^{−z} .s)	±	(p-value)	(W·m ⁻²)	
DAY_CLEAR	2.3	1.5	0.001	29.2	1.6	0.001	-0.20	0.04	0.027	*	
DAY_CLOUDS	-3.6	1.3	0.800	24.5	1.4	0.279	-0.23	0.03	0.001	*	
NIGHT_CLEAR	-7.3	1.3	0.002	18.9	1.9	0.052	0.09	0.09	0.033	*	
NIGHT_CLOUDS	-7.8	1.3	0.003	14.2	1.6	0.000	0.21	0.06	0.000	*	

Table 3.4 Near-Shore Station, 2005 – Mass and energy fluxes. Day/Night, Clear/Cloud Averages Values for energy storage have been omitted (*) because they could not be accurately estimated at this station.

sensible heat by the end of the study period. There was very little diurnal variation in H and it tended to remain negative for most conditions and was more negative at night. The only conditions under which H was positive were daytime clear sky conditions (Table 3.4). H formed only a minor part of the energy balance during day time conditions. However, the contribution to the energy balance increased to 20-24% of available energy during nighttime conditions.

3.1.2.2 – Latent Heat Fluxes

Throughout the sampling period, the cumulative LE comprised 15% of the energy budget for all conditions (Table 3.3). In general, LE tended to be higher during daytime clear sky conditions. LE comprised a very significant part of the energy balance during nighttime conditions (44-51%) but the average values were only 14.2-18.9 W·m⁻² (Table 3.4). Despite the relatively low magnitude of H and LE, the average Bowen Ratio (B) exhibited a diurnal variation and sensible heat fluxes tended to increase B to 1.0 and beyond during the night but fell to an average of 0.8 during the day.

3.1.2.3 – The Flux of Carbon Dioxide

Over the course of the sampling period, this area was a sink for CO_2 with an average carbon dioxide flux of -0.11 µmol·m⁻²·s⁻¹ (Table 3.3). However, the rate of accumulation dropped to approximately zero by DOY 204 when the average daily efflux began to balance the average daily uptake. Seasonally, there was only a very weak diurnal pattern with an average range of -0.22 to 0.17 µmol·m⁻²·s⁻¹. There was no significant relationship (p-value >>0.10) between FC and cloudy or clear sky conditions during the day (Table 3.3). However, cloudy sky conditions at night appeared to enhance the efflux of CO_2 .

3.1.3 – Tidal and Diurnal Variation of Flux Components

3.1.3.1 – Diurnal Variation

On a seasonal basis, the three flux estimates under consideration did not exhibit significant diurnal variation (Figure 3.1). However, LE tended to rise slightly from 15h00 to 22h00 (CST).

3.1.3.2 – *PEAK Stage*

The 'PEAK' stage exhibited more diurnal variation on a seasonal basis among each one of the fluxes under consideration than any of the other stages (Figure 3.2). Much of this variation tended to occur late in the afternoon; there was a noticeable trough with FC, a slight peak with H and a pronounced peak with LE between the hours of 15h00 and 20h00 (CST).

3.1.3.3 – TROUGH Stage

During the 'TROUGH' stage, the diurnal fluxes tended to look much like the seasonal/diurnal average (Figure 3.3); there was no observable variation with FC and H, but LE exhibited a slight peak towards the end of the afternoon.

3.1.3.4 – FLOW Stage

There was a noticeable drawdown of CO_2 during the late afternoons between 16h00 and 20h00 that was accompanied by a small increase in LE during the same time. H appeared to be largely flat during this tidal stage.

3.1.3.5 – EBB Stage

Diurnal variation of fluxes appeared to be similar to the 'FLOW' stage with the exception that FC appeared to peak during the early mornings. There was no noticeable late afternoon drawdown of CO_2 . H remained flat throughout the day and there continued to be a late afternoon peak with LE late in the afternoons.


Shore Station, 2005 for all tidal stages. The 'box and whisker' format for each of the hourly bins represent quartile ranges of value ranks from 0.10 to 0.25, 0.25 to 0.50, 0.50 to 0.75 and 0.75 to 0.90th percentiles. (\circ) represents outliers.



0.25, 0.25 to 0.50, 0.50 to 0.75 and 0.75 to 0.90th percentiles. (\circ) represents outliers.



0.25, 0.25 to 0.50, 0.50 to 0.75 and 0.75 to 0.90th percentiles. (\circ) represents outliers.



Shore Station, 2005 for the 'FLOW' tidal stage. The 'box and whisker' format for each of the hourly bins represent quartile ranges of value ranks from 0.10 to 0.25, 0.25 to 0.50, 0.50 to 0.75 and 0.75 to 0.90th percentiles. (\circ) represents outliers.



Figure 3.5 Diurnal variation of fluxes from the Near-Shore Station, 2005 for the 'EBB' tidal stage. The 'box and whisker' format for each of the hourly bins represent quartile ranges of value ranks from 0.10 to 0.25, 0.25 to 0.50, 0.50 to 0.75 and 0.75 to 0.90th percentiles. (\circ) represents outliers.

3.1.4 – Diurnal Variation on Representative Days

Three key periods were chosen based on their representativeness of the general rise in air temperatures throughout the sampling period and by data availability. Stable conditions dominated throughout the first two periods and unstable conditions prevailed only during the final period. H exceeded LE in terms of its strength, but is decidedly negative. It is only by Period Three when unstable conditions dominate that H became slightly positive at 5.9 W·m⁻² and LE became strongly positive at 36.4 W·m⁻². During Period Two, LE appeared to be responsive to average vapour pressure deficit and became as low as -4.7 W·m⁻² when the average VPD dropped to 0.27 kPa. When average atmospheric conditions became unstable during period three, both H and LE were became positive and LE became a significant component of RNET at 36.4 W·m⁻². Each of the three periods were characterized by observations of forcings and fluxes on DOYs 163, 201 and 223 respectively.

3.1.4.1 – Forcings on Representative Days (DOYs 163,201,223)

On DOY 163, cloud cover between 7h00 and 18h00 (CST) coincided with a distinctly high TSURF next to the station in the later part of the day (Figure 3.6). There was no significant variation of UST over the day, but WS and z/L fluctuated frequently with a slight tendency towards windy, stable conditions late in the afternoon (Figures 3.7 and 3.8). Notably, z/L displayed a strong spike which coincided with a spike in VPD and T_a at 17h00 (CST). However, z/L was in excess of unity which suggests that this air mass may have been outside of the atmospheric boundary layer (ABL).

On DOY 201, WS, z/L and UST declined throughout the afternoon occurred as clouds



Figure 3.6 Ta, TSURF and VPD on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.



Figure 3.7 z/L and UST on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.



Figure 3.8 zo and WS on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.

cover increased and T_a , TSURF and VPD rose. z/L once again indicated that these measurements may have been made outside of the ABL. However, a sharp spike towards stability at 19h00 (CST) coincided with a sharp rise in VPD, T_a , TSURF. WS and zo appeared to covary throughout the first half of the day, but a strong divergence took place in the afternoon when WS dropped and zo rose significantly.

By DOY 223, periods of unstable atmospheric conditions corresponded with a TSURF that was consistently higher than Ta. TSURF peaked in the early evening with the onset of cloud cover and declined thereafter WS decreased shortly before noon, but increased throughout the rest of the afternoon as unstable conditions began to prevail. zo displayed the opposite trend and increased shortly before noon and then decreased during the first part of the afternoon.

3.1.4.2 – Fluxes on Representative Days (DOYs 163,201,223)

During DOY 163, there were still ice floes in Hudson Bay and the observed fluxes and forcings were typical of these background conditions. There were two tidal peaks, one in the early morning and one in the middle of the afternoon (Figure 3.9). Despite the very large difference between TSURF and Ta, both H and LE remained flat throughout the day and did not appear to be responsive to diurnal variation or to the tidal stage. As well, LE did not appear to react to the strong VPD that developed in the middle of the afternoon. The flux of carbon dioxide showed no diurnal pattern but a considerable amount of variation between the hourly ensembles.

By DOY 201, atmospheric forcings appeared to have an observable affect on the estimated fluxes. Although the signal fluctuated considerably, FC was largely negative and there was a brief, but intense drawdown by the middle of the afternoon. TSURF becomes increasingly



Figure 3.9 H, LE and FC on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.

larger than Ta during the second half of the day with the introduction of cloud cover. LE becomes somewhat positive towards late afternoon when VPD became very large. Variations in this signal appeared to closely correspond with variations in H. However, H was still small and tended to be negative. Also, a sharp spike towards stability at 19h00 (CST) coincided with a sharp rise in VPD, Ta, TSURF, LE and FC and a sudden drop of H.

Although there was a great deal of variability on DOY 223, FC demonstrated a diurnal pattern with efflux during the night and drawdown during the sunny day. H displayed no diurnal variation but was generally positive for the first time in this sampling period. However, LE became much more pronounced towards the late afternoon which coincided with a high VPD, intense instability and the onset of cloudy conditions.

3.2 - The Intertidal Zone, 2006

3.2.1 - Seasonal Meteorology and Forcings

3.2.1.1 - Air and Surface Temperatures

The Ta time series was largely continuous and, similar to the Near-Shore Station, missing values were filled in with data from the 'Churchill A Station'. Because TSURF was modeled based on Ta, this time series had the same coherence and exhibited the same fluctuations. Over the course of the sampling period, Ta was 12.1 °C with average nightly temperatures only 1.8 °C cooler than the days (Table 3.5). Although there were numerous fluctuations, the average maximum Ta was steady throughout the summer whereas the nightly average minimum temperatures rose over the course of the sampling period from just above 0 °C to approximately 10 °C. Clear sky conditions during both the day and night were warmer than cloudy conditions but the delT tended to be

much lower during the night (-0.6 °C).

TIDAL FLATS, 2006 - KEY FORCING AVERAGES and STABILITY PARAMETER MEDIAN										
	Ta	TSURF	delT	RG	VPD	UST	WS	z/L	zo	
	(°C)	(°C)	(°C)	(W·m⁻²)	(kPa)	(m·s⁻¹)	(m·s⁻¹)	(median)	(m)	
AVERAGE	12.1	13.6	-1.5	209.2	0.91	0.27	4.45	-0.04	0.028	
DAY	12.9	14.7	-1.8	307.6	1.14	0.27	4.53	-0.12	0.030	
NIGHT	10.1	10.7	-0.6	9.4	0.41	0.27	4.25	0.14	0.024	
CLEAR	13.0	14.4	-1.4	242.1	0.59	0.26	3.99	-0.01	0.031	
CLOUDS	11.6	13.1	-1.5	191.4	1.15	0.27	4.71	-0.06	0.026	
FLOW	12.0	13.5	-1.5	206.3	0.51	0.26	4.45	0.01	0.033	
PEAK	12.0	13.4	-1.5	229.9	0.53	0.27	4.55	-0.01	0.025	
EBB	12.3	13.8	-1.5	211.6	2.33	0.27	4.25	-0.08	0.026	
TROUGH	12.1	13.5	-1.4	187.5	0.52	0.27	4.46	-0.10	0.026	

Table 3.5 Intertidal Station, 2006 – Key Forcing Averages & Stability Parameter Median (A). Each of the columns are averages with the exception of z/L which is a

TIDAL FLATS, 2006 - KEY FORCING AVERAGES and STABILITY PARAMETER MEDIAN										
	Ta	TSURF	delT	RG	VPD	UST	WS	z/L	zo	
	(°C)	(°C)	(°C)	(W·m⁻²)	(kPa)	(m·s⁻¹)	(m·s⁻¹)	(median)	(m)	
DAY_CLEAR	14.4	16.2	-1.8	400.1	0.73	0.26	4.09	-0.13	0.036	
DAY_CLOUDS	12.1	14.0	-1.9	265.7	1.40	0.27	4.74	-0.11	0.027	
NIGHT_CLEAR	10.4	10.9	-0.6	11.6	0.36	0.26	3.81	0.17	0.023	
NIGHT_CLOUDS	9.9	10.5	-0.6	7.7	0.46	0.28	4.61	0.11	0.025	

Table 3.6 Intertidal Station, 2006 – Key Forcing Averages & Stability Parameter Median (B). Each of the columns are averages with the exception of z/L which is a median value.

3.2.1.2 - Radiation Components

All of the estimates of RNET were based on an algorithmic, modeled relationship with RNET,

Ta, cloud cover fraction and water vapour pressure (e) as observed at the Near-Shore Station in

2005. The resulting time series was generally coherent and the range of values was within

expectations. Over the course of the sampling period the daily average maximum was 119.9

 $W \cdot m^{-2}$ and the average minimum was steady throughout the season at -52.1 $W \cdot m^{-2}$ (Table 3.7).

The daily maximum tended to decrease slightly by the end of the summer. Clear sky conditions

favoured a higher RNET of 142.7 W·m⁻² compared to 109.7 W·m⁻² for cloudy conditions (Table

TIDAL FLATS, 2006 - ENERGY PARTITIONING									
	RNET	S/RNET	H/RNET	LE/RNET	В				
	(W·m⁻²)	(unitless)	(unitless)	(unitless)	(unitless)				
AVERAGE	119.9	0.60	0.19	0.37	1.11				
DAY	196.2	0.64	0.19	0.28	0.91				
NIGHT	-52.1	1.04	0.36	-0.25	1.83				
CLEAR	142.7	0.44	0.19	0.30	1.30				
CLOUDS	109.7	0.70	0.18	0.41	0.99				
FLOW	103.1	0.69	0.20	0.39	0.97				
PEAK	136.4	0.82	0.16	0.35	1.19				
EBB	133.7	0.44	0.22	0.35	1.29				
TROUGH	107.9	0.28	0.18	0.39	1.03				

Table 3.7 Tidal Flats Station, 2006 – Energy Partitioning (A).

3.8). Despite being based on a modeled relationship from the near-shore zone, RNET appeared to vary with tidal state and rose to 136.4 and $133.7 \text{ W} \cdot \text{m}^{-2}$ for 'PEAK' and 'EBB' conditions respectively. Rg was somewhat low for the sampling period at 209.2 W \cdot m⁻² (Table 3.5) with clear

TIDAL FLATS, 2006 - ENERGY PARTITIONING										
	RNET	S/RNET	H/RNET	LE/RNET	В					
	(W·m ⁻²)	(unitless)	(unitless)	(unitless)	(unitless)					
DAY_CLEAR	269.4	0.63	0.19	0.22	1.10					
DAY_CLOUDS	168.7	0.66	0.17	0.32	0.80					
NIGHT_CLEAR	-55.2	1.03	0.34	-0.25	1.78					
NIGHT_CLOUDS	-50.1	1.02	0.36	-0.25	1.89					
Table 3.8 Tidal Flats Station, 2006 – Energy Partitioning (B).										

conditions raising the average Rg to 242.1 W·m⁻² compared to 191.4 W·m⁻² for cloudy conditions. Daytime clear conditions resulted in an average Rg of 400.1 W·m⁻² compared to 265.7 W·m⁻² for cloudy conditions.

3.2.1.3 – *Humidity*

The RH time series was highly fragmented and featured a large gap in the middle of the season thus complicating further analysis. The nightly maximums varied but remained relatively constant throughout the summer at approximately 90%. The daily minimums displayed a great deal of variability and varied between 5 and 75%. VPD was greatest during the day and smallest during the night at 1.14 and 0.41 kPa respectively (Table 3.5). Similar to the observations from the Near-Shore Station, clear sky conditions had lower VPDs than cloudy sky conditions at 0.59 and 1.15 kPa respectively.

3.2.1.4 - Wind and Stability

Throughout the sampling period, z/L indicated that most conditions were slightly unstable or neutral (-0.04) and that most of the measurements that were taken at this station were well within the boundary layer (Table 3.5). z/L exhibited a distinct diurnal pattern, and stable conditions tended to occur during the night in contrast to the largely unstable conditions during the day. Estimates of UST indicated that there was almost no variability and remained close to the period average of 0.27 m·s⁻¹ during almost any combination of conditions (Table 3.5). WS for the sampling period was 4.45 m·s⁻¹ and slightly higher wind speeds were observed during the day than the night at 4.53 and 4.25 m·s⁻¹ respectively. Wind speeds were greater during cloudy sky conditions (4.71 m·s⁻¹) than clear sky conditions (3.99 m·s⁻¹). This difference was statistically significant at p<<0.10. The average value of surface roughness at the Tidal Flats Station was 0.028 m and tended to be lower than those observed at the Near-Shore Station for almost all combinations of diurnal, sky conditions and tidal stages (Tables 3.5 and 3.6). Similar to the Near-Shore Station, zo tended to be higher during the day time, clear sky conditions (0.036 m). As well, zo was lower during the 'TROUGH' tidal stage at 0.026 m.

3.2.2 - Seasonal Flux Estimates, Energy Partitioning and Uncertainty

Data acquisition success for estimated turbulent fluxes over the course of the sampling period was as low as 32% which meant that many of the time series were highly fragmented and

TIDAL FLATS, 2006 - FLUXES AND STORAGE										
		Н			LE			FC		S
	(W·m ⁻²)	±	(p-value)	(W·m ⁻²)	±	(p-value)	(µmol·m ^{-z} .s)	±	(p-value)	(W·m ⁻²)
AVERAGE	22.5			44.1			0.04			72.1
DAY	36.5	1.7	0.000	55.3	1.7	0.000	-0.45	0.19	0.000	125.9
NIGHT	-18.6	1.9	0.000	13.1	2.1	0.000	1.42	0.46	0.000	-54.1
CLEAR	27.6	2.4	0.097	43.1	2.0	0.719	0.04	0.15	0.959	63.3
CLOUDS	19.3	1.7	0.247	44.8	1.9	0.724	0.04	0.29	0.948	76.3
FLOW	20.4	2.3	0.571	39.8	2.5	0.135	-0.29	0.30	0.048	70.7
PEAK	22.0	3.0	0.950	47.8	2.8	0.222	-0.10	0.39	0.514	111.5
EBB	29.0	3.1	0.083	47.2	2.9	0.304	0.42	0.44	0.057	58.8
TROUGH	18.9	2.8	0.364	42.2	3.2	0.589	0.21	0.33	0.267	29.7

Table 3.9 Tidal-Flats Station, 2006 – Fluxes & Storage (A).

TIDAL FLATS, 2006 - FLUXES AND STORAGE										
		Н			LE			FC		S
	(W·m ⁻²)	±	(p-value)	(₩·m ⁻²)	±	(p-value)	(µmol·m ^{−z} ·s)	±	(p-value)	(W·m ⁻²)
DAY_CLEAR	51.2	3.1	0.000	58.0	2.7	0.000	-0.56	0.11	0.00	169.1
DAY_CLOUDS	28.9	1.9	0.010	53.9	2.2	0.000	-0.39	0.28	0.01	111.8
NIGHT_CLEAR	-19.0	2.3	0.000	13.6	2.3	0.000	1.24	0.39	0.00	-57.0
NIGHT_CLOUDS	-18.1	2.9	0.000	12.7	3.3	0.000	1.60	0.82	0.00	-51.1

Table 3.10 Tidal-Flats Station, 2006 – Fluxes & Storage (B).

difficult to interpret. Similar to the situation at the Near Shore Station, the 180 ° rejection angle meant that all offshore winds had to be rejected from this analysis. Other sources of data loss were persistent power failures caused by faulty batteries and the loss of a solar panel due to a high wind event. Each one of the flux estimates could be statistically distinguished with a reasonable degree of certainty by dividing them based on day or night conditions (p<<0.10) (Table 3.9). Clear or cloudy skies could not generally be distinguished from one another based on the estimates of any of the fluxes (p>>0.10). However, when each of the fluxes was sorted for

day versus night and clear versus cloudy conditions, they could be readily distinguished from each other with a very high degree of certainty (p<<0.10) (Table 3.10). Estimates of H and LE during each tidal stage could not be statistically distinguished from one another (p>0.10). However, 'FLOW' and 'EBB' conditions appeared to be statistically different from one the entire sample of FC estimates (p>0.10).

3.2.2.1 - Sensible Heat Fluxes

H underwent a continuous cumulative rise throughout the sampling period and the sampling area was a significant source of sensible heat (Table 3.9). On average, H was 22.5 W·m⁻² and rose to an average of 36.5 W·m⁻² during the day and fell to -18.6 W·m⁻² at night. Nights were the only time when H was negative and night-time sky conditions did not appear to have a significant influence on these estimates. Conditions that appeared to favour lower values of H included cloudy skies and the 'TROUGH' tidal stage at 19.3 and 18.9 W·m⁻² respectively (Table 3.9). There was a marked difference between values of H during daytime clear and cloudy conditions; H peaked at 51.2 W·m⁻² during clear conditions and fell to an average of 28.9 W·m⁻² during cloudy conditions (Table 3.9). As a proportion of RNET, H tended to be very large at night but the absolute values were relatively low (Table 3.6).

3.2.2.2 - Latent Heat Fluxes

LE experienced a steady rise throughout the season and was almost always positive. The period average was 44.1 W·m⁻² and featured an average diurnal range of 55.1 to 13.1 W·m⁻² (Table 3.9). Notably, LE did not decline appreciably during the 'TROUGH' tidal stage. Although there was

no significant absolute difference (p-value >>0.10) between clear and cloudy conditions during any time of the day (table 3.8), LE tended to comprise a larger proportion of RNET during cloudy conditions as opposed to clear conditions during the day at 41 and 30% respectively (Table 3.6). Bowen ratios display a distinct diurnal variation with high values occurring during the night (1.83) and low values occurring during the day (0.91). Higher values of B were favoured during clear sky conditions (1.30) as opposed to cloudy conditions (0.99). However, when sorted for both diurnal and sky conditions, B tends to be the lowest during cloudy conditions during the day time (0.80) (Table 3.7).

3.2.2.3 - Energy Storage

Although there were notable gaps in this time series, the S was the primary component of the energy budget for this area at 60% of available RNET (Table 3.6) with a period average of 72.1 $W \cdot m^{-2}$. S actually exceeded available RNET during night-time conditions by 4% and this ratio was different for cloudy and clear sky conditions at 0.70 and 0.44 respectively. S exhibited a strong diurnal variation with average values of -54.1 $W \cdot m^{-2}$ during the night and 125.9 $W \cdot m^{-2}$ during the day (Table 3.9). The ratio of S to RNET was the highest during 'FLOW' and 'PEAK' tidal stages and dropped to a low of 0.28 during 'TROUGH' stages.

3.2.2.4 - The Flux of Carbon Dioxide

Again, various gaps in the time series of FC due to periodic power system malfunctions made it difficult to analyse. FC displayed a noticeable diurnal pattern with an average nighttime efflux of $1.42 \ \mu mol \cdot m^{-2} \cdot s^{-1}$ and a daytime drawdown of -0.45 $\mu mol \cdot m^{-2} \cdot s^{-1}$ (Table 3.9). For the period, this

area was a small source of CO₂ with an average FC of 0.04 μ mol·m⁻²·s⁻¹. During the daytime, clear sky conditions coincided with the greatest draw down of -0.56 μ mol·m⁻²·s⁻¹ and night time cloudy conditions coincided with the greatest efflux of 1.6 μ mol·m⁻²·s⁻¹ (Table 3.10).

3.2.3 – Tidal/Diurnal Variation

3.2.3.1 - Diurnal Variation

The seasonal diurnal variation of H and FC during each of the stages was similar to its ensemble diurnal variation (Figure 3.10). However, the diurnal variations of LE in response to tidal stage differed somewhat from the seasonal diurnal variation. On an average diurnal basis, LE exhibited the same pattern as H with a gradual rise throughout the day until approximately 17h00 (CST) when a sharp decline was observed.

3.2.3.2 – 'PEAK' Stage

Both RNET and S tended to be higher during 'PEAK' stage at 133.7 W·m⁻² and 111.5 W·m⁻² respectively (Table 3.6). Both H and LE were the highest during 'PEAK' stage late in the day between 14h00 and 17h00 (CST) (Figure 3.11). FC exhibited a distinct diurnal pattern with significant drawdown occurring during the middle of the day shortly before the peaks of H and LE.

3.2.3.3 – 'TROUGH' Stage

RNET was lower during 'TROUGH' stage at 107.9 W·m⁻² (Table 3.6) and B remained low at 1.03. S was actually less than LE during this stage at 29.7 W·m⁻² (Table 3.9). H exhibited a











strong diurnal pattern during this stage and rose and declined very sharply during the early mornings and late afternoons respectively (Figure 3.12). However, the average value of H was the lowest during this stage at 18.9 W·m⁻² and LE remained relatively high at 42.2 W·m⁻² (Table 3.9). FC displayed a diurnal pattern and an average efflux of 0.21 μ mol·m⁻²·s⁻¹.

3.2.3.4 – 'FLOW' Stage

RNET and B were the lowest during the 'FLOW' stage at 103.1 W·m⁻² and 0.97 respectively (Table 3.6). Both LE and H peaked later in the day between 14h00 and 17h00 with a gradual rise followed by a sharp decline early in the evening (Figure 3.13). This stage tended to favour the largest drawdown of CO₂ with an FC of -0.29 μ mol·m⁻²·s⁻¹.

3.2.3.5 – '*EBB*' Stage

RNET and B tended to be higher during the 'EBB' stage at 133.7 W·m⁻² and 1.29 respectively (Table 3.6). S was relatively low during this stage at 58.8 W·m⁻² and mean H was the highest during this stage at 29.0 W·m⁻² (Table 3.6). Both H and LE tended to rise sharply in the morning and peak close to noon (Figure 3.14). FC exhibited the strongest efflux at 0.42 μ mol·m⁻²·s⁻¹.

3.2.4 - Diurnal Variation on Representative Days

Three key periods were selected for a more detailed analysis due primarily to data availability and because they reflected the gradually rising Ta and TSURF over the course of the sampling period. This served to highlight many of the trends that were noted in the more general analysis of the key forcings and variables. The average net radiation declined from a high of 134.1 W·m⁻²



percentiles. (\circ) represents outliers.



0.25, 0.25 to 0.50, 0.50 to 0.75 and 0.75 to 0.90th percentiles. (\circ) represents outliers.

during Period One to a low of 75.5 W·m⁻² during Period Three. Neutral conditions prevailed during Periods One and Three, but slightly unstable conditions dominated during Period Two. This paralleled a general trend in S with very low average values recorded during the first and last periods and a high average value (114.9 W·m⁻²) was observed during Period Two. delT was always negative and H was always positive. However, H declined to a very low average value (2.0 W·m⁻²) during the final period. It was also during Period Three that there was a relatively high VPD of 0.81 kPa observed which was coincident with an extremely large relative LE of 35.0 W·m⁻². LE was at its largest during the unstable Period Two at 66.5 W·m⁻². Each of the three periods were characterized by observations of forcings and fluxes on particular day and 177, 210 and 240 were selected for a detailed analysis.

3.2.4.1 - Observed Forcings on Representative Days (DOYs 177,210,240)

Although clouds were present on DOY 177, delT was especially large during day light hours and TSURF and VPD reached their maximums by 19h00 (CST) (Figure 3.15). This was accompanied by a gradual decline in WS and z/L throughout the day both of which reached their lowest values between 14h00 and 18h00 (CST) (Figures 3.16 and 3.17).

During DOY 210, delT was the largest during the day and, similar to DOY 177, was always negative. However, it was observed that WS and UST decreased throughout the day. At the same time, zo increased as unstable, cloudy conditions began to prevail.

By DOY 240, Ta, TSURF and VPD continued to display the same pattern apparent in the previous two representative days with late afternoon highs for each of these variables and an especially high value for VPD. As well, there was a trend towards instability and low WS in the



Figure 3.15 Ta, TSURF and VPD on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.



Figure 3.16 z/L and UST on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.





afternoon. This sharp increase occurred despite the arrival of cloud cover between 11h00 and 15h00 (CST).

3.2.4.2 - Observed Fluxes and Energy Storage on Representative Days (DOYs 177,210,240)

On DOY 177, both H and LE appeared to covary, but LE generally exceeded H throughout this day (Figure 3.18). Despite the fact that delT was negative during the night time, H became slightly negative and appeared to be running 'counter-gradient'. S was very large and by far the dominant component of the energy balance and spiked during the mid-morning but dropped sharply by midday and became negative by 15h00 (CST). H and LE did not decline as rapidly and persisted with positive values into the early evening. Small or negative values of S appeared to coincide with the tidal 'TROUGH' stage that occurred between 13h00 and 16h00 (CST). FC displayed a rough diurnal pattern with strong a strong efflux that occurred well into the evening.

Later in the summer, DOY 210 displayed many of the same trends: S spiked in the midmorning, but was matched or exceeded by H and LE by the mid-afternoon when the tide was out. In general, H and LE appeared to covary but LE was consistently larger and was almost always positive. A sudden downward spike of FC occurred at the same time that WS and UST experienced a similar downward spike. At the same time, zo increased as unstable conditions began to prevail.

By the end of the sampling period, the events of DOY 240 demonstrated the potential impacts of daytime cloudy conditions on the measured fluxes. The time series for each of the flux and storage estimates began with a number of spikes early in the morning. However, each of them began to develop coherence throughout a clear late morning and S rose sharply whereas



Figure 3.18 H, LE and FC on representative days. Hourly time series of key variables are plotted against a background that represent general sky conditions (clear or cloudy) and tidal state on this particular day. The height of the tide is measured from the low tide mark.

LE and H did not appear to change. Cloudy sky conditions developed in the early afternoon and S began to experience some downward fluctuations whereas both LE and H began to peak. The clouds disappeared by 15h00 (CST) but S continued to fall to negative values as tidal trough conditions occurred. At the same time, both H and LE remained close to zero. There was a trend towards instability and low WS in the afternoon. This occurred at the same time that VPD, TSURF, T_a rose sharply and coincided with the peaks of LE and H. This sharp increase occurred despite the arrival of cloud cover between 11h00 and 15h00 (CST).

4 - Discussion

Despite various sources of error in each of the datasets and the absence of key measurements within the water column, estimates of fluxes and forcings were developed that enabled comparisons to be drawn with previously published data from research projects in other, similar environments. However, the presence of 'cloud-bias' during 2005 made the interpretation of the relationships between these variables difficult. The available evidence suggests that energy and mass fluxes in the near-shore and intertidal zones may have been influenced by a sea-breeze effect and by the gradually decreasing average water depth towards the shoreline. As well, the average estimates of FC in each area and their enhancement during tidal stages when the water is in motion indicate that the presence of living and dead kelp are a significant forcing of CO_2 exchanges along this coastline.

4.1 - The Near-Shore Zone, 2005

Differences between radiation and flux footprints and the different nature of the underlying surfaces that generated the observed measurements was a very serious methodological problem that meant that it was not possible to do a complete analysis of the energy budget nor was it possible to do a critical evaluation of the role that surface temperatures played on the observed turbulent fluxes in this particular area. The radiative and turbulent components of the energy budget could not be coupled with any degree of confidence which limited the number of forcings that could be critically evaluated as drivers of the observed fluxes. Another factor that complicated this analysis was the existence of a 'cloud-bias' during the day time. However, several useful observations and estimations were made that would prove useful to the analysis of

processes at work in the Tidal Flats during 2006 and to an overall understanding of both the Near-Shore and Intertidal zones.

The height of the station on board the Ithaca proved advantageous in the sense that it meant that a very large flux footprint could be sampled. However, it also meant that, depending on stability and wind speed, there was approximately a 1.5 km difference separating the radiation footprint at the base of the Ithaca and the flux footprint some distance offshore. The underlying surfaces are fundamentally different in each of these footprints; within the radiation footprint was shallow water or exposed sediments depending on the tidal stage and within the flux footprint was deep water. The analysis of Panin and Foken [2005] from the Caspian Sea clearly indicated that there are going to be significant differences between the forcings influencing each environment and the observed fluxes. As a consequence of this, it was not possible to include heat storage in the energy balance which has been observed to be the largest component of the energy balance in this area [*Silis et al., 1989*]. As well, it was not possible to properly evaluate the influence of surface temperature on the observed fluxes thus making it difficult to draw any decisive conclusions regarding the existence of a 'solubility' or 'biological pump' and their influence on the flux of carbon dioxide in the near-shore zone.

A significant anomaly was observed wherein higher average values of incoming short wave radiation and net radiation occurred during cloudy as opposed to clear conditions. When averaged over the course of the period, incoming short wave radiation was 350.8 W·m⁻² whereas during clear conditions it averaged 317.9 W·m⁻² (Tables 3.1). This obviously counter-intuitive result made all subsequent observations and variable parameters questionable due to the possibility of instrument or sampling error and the suspicion that this dataset was biased and/or

highly inaccurate. However, a close examination of the dataset revealed that cloud occurrence over the course of the day was not random. In general, the occurrence of cloudy skies ranged from 40 – 50 % over the course of an average day. During the early mornings until 08h00 (CST), average cloud occurrence tended to be close to 40 %. 'Cloud occurrence' refers to the probability during a particular hour of the day that there would be cloudy sky conditions as opposed to clear. This tended to increase up to 50% as solar noon approached and there was a very small peak close to 17h00 (CST). Although the difference between cloud occurrences over the course of the day were not profound, they were large enough that early mornings, when the sun was at its lowest elevation, tended to be clear whereas afternoons, when the sun was at its highest elevation, tended to be cloudy. As a result of this, incoming shortwave appeared to be the most intense during cloudy sky conditions and was weakest during clear skies thus making it seem brighter when there were clouds. Because there tends to be a close relationship between net radiation and incoming short wave radiation during the day, this anomaly also affected observations and calculations of net radiation in the context of cloudy or clear sky conditions.

Despite these problems, observations and parameterizations of key variables highlighted the importance of the flux of latent heat to the energy budget, the role of the vapour pressure deficit and the small and consistently negative flux of sensible heat. Congruent with the observations of Silis et al. [1989], it was expected that the flux of latent heat would dominate the partitioning between H and LE and this proved to be true; B was very low under almost all combinations of forcings and the LE generally comprised 14 - 19 % of RNET but was still generally small with an average seasonal value of only 23.1 W·m⁻² (Table 3.3). In contrast to the observations of Silis et al. [1989], onshore winds appeared to be associated with increases of

VPD which became more intense when there was cloud cover. Although the period averages didn't indicate it (Tables 3.1 and 3.3), there frequently appeared to be a relationship between VPD and LE as observed during the late afternoons of DOY 201 and 223 (Figures 3.6 and 3.9); VPDs increased at the same time as there was a sharp increase of Ta on both days. This appeared to have resulted in small increases in the flux of latent heat on both days. Notably, the same pattern of a late afternoon increase of air temperature and vapour pressure deficit occurred earlier in the season on DOY 163 (Figure 3.6) but there was no corresponding influence on the observed energy fluxes (Figure 3.9). This was likely due to the accumulation of offshore sea ice that was observed approximately 1 - 3 km offshore that is known to impede turbulent interactions between the sea-water and the atmosphere [*Serreze et al., 2007*].

Although it has been noted that the flux of sensible heat is often small and negative in marine environments due to the role of horizontal advection from nearby coastlines [*Mahrt et al., 1998*], it was not expected to be so consistently negative under various combinations of conditions (Tables 3.3 and 3.4) such that the Near-Shore area was observed to be a sink of sensible heat over the course of the sampling period. It should be noted that the study conducted by Mahrt et al. [*1998*] only observed this tendency with the sensible heat flux during offshore winds whereas this study considered only onshore winds

Tides were not observed to exert any significant influence on any of the energy fluxes but did appear to have a small influence on the flux of carbon dioxide. In general, the near-shore zone was a CO_2 sink with an average flux of -0.11 µmol·m⁻²·s⁻¹ which was comparable to other observations from studies in similar environments (Table 1.1). However, this was almost 100X more intense than the observations made by Else et al. [2008] during the ArcticNet, Leg II
expedition in 2005. Similar to the results generated by this study, the waters of Hudson Bay in the near-shore zone were a sink for CO₂ but the downward flux would be incongruent with their observation that coastal sea waters tended towards supersaturation. This would indicate that there could be intense biological productivity taking place in these coastal waters and/or that the partial pressure of CO₂ changes in these coastal waters with the time of year; importantly, the study of Else et al. [2008] took place in the mid- to late fall of 2005 whereas this study was conducted during the summer of 2005. Both are distinct possibilities given that the nutrient-rich plume of the Churchill River discharges 20 km away from these coastal sites and that the temperature of Hudson Bay begins to decrease rapidly during the fall until ice-cover forms. Importantly, there tended to be a noticeable efflux of CO₂ during morning 'EBB' tidal stage (Table 3.3) which seemed to be countered by a late afternoon/early evening drawdown during 'FLOW' and 'PEAK' conditions (Figure 3.2 and Figure 3.4). This late-day drawdown was not apparent during 'EBB' conditions (Figure 3.5). However, there appeared to be a significant efflux occurring during the early mornings. It must be noted that the water surface temperatures tended to be lower during the nights than the days at 10.3 and 12.6 °C respectively. In general, there was an efflux of CO₂ during the nights of 0.17 μ mol·m⁻²·s⁻¹ and a drawdown of -0.22 μ mol·m⁻²·s⁻¹ during the days. Thus, if the solubility of CO₂ were the primary determinant of its surface flux, then it should be expected that the reverse would be true but this is not the case and it may be assumed that there is significant primary production taking place in these waters and that the 'biological pump' is dominating the 'solubility pump'. This needs to be considered in the context of the direction of water movement wherein the surface waters were moving away from the shore during the 'EBB' stage and towards the shore during the 'FLOW' and part of the

'PEAK' stages. The water during the 'EBB' stage is coming from the tidal flats and presumably has been recently influenced by that environment. As such the flux of CO_2 is different and the observed efflux stands in contrast to the drawdown of the other tidal stages.

4.2 - The Tidal Flats, 2006

Although there is no way to conclusively evaluate the net radiation model that was developed based on data from the Near-Shore Station from the previous year, the results were within expectations and were comparable to the results from the study conducted by Silis et al [1989]. As such, it has been possible to more critically evaluate the relationships between surficial forcings and the estimated turbulent fluxes given that the radiative, thermal and flux footprints coincide with one another. However, observations of the other components of the energy balance and key forcings from the Tidal Flats Station demonstrated some key differences with previous studies that had been conducted in similar environments.

Surface temperatures and the net radiation of the tidal flats were estimated based on a model derived from the data from the Near-Shore Station with the assumption that the processes responsible for these two variables would be largely the same in 2006. With estimates of net radiation in hand, the turbulent energy budget was rearranged to estimate surface heat storage as a residual of the energy balance equation (Equation 8). Unfortunately, there was no way to double-check these estimates with the exception of comparing them to the results of the study conducted by Silis et al. [*1989*] and arbitrarily determining whether or not they looked reasonable. Net radiation in the Tidal Flats was estimated to range from an average of -52.1 $W \cdot m^{-2}$ during the night to 196.2 $W \cdot m^{-2}$ during the day with a period average of 119.9 $W \cdot m^{-2}$

(Table 3.6). This was directly comparable to the results of Silis et al. [1989] where net radiation during onshore winds had a similar diurnal range. Importantly, their estimation of surface heat storage, based on direct measurements of water temperature and the heat flux into the bottom sediments were nearly identical to those that were measured in the tidal flats during the summer of 2006; in both studies, surface heat storage comprised an average of 60% of net radiation. Despite indications that this model was overestimating net radiation by 22% (Table 2.3), this would appear to lend some credibility to the use of the net radiation model that was used for this study. As well, the average surface heat storage was 111.5 W·m⁻² during the 'PEAK' tidal stage and fell to an average of 29.7 W·m⁻² during the 'TROUGH' stage which is consistent with the thermal inertia of the two substrates; water has a much higher specific heat capacity than the mineral sediments of the drained tidal flats.

There were, however, some key differences involving TSURF and H that were in sharp contrast to the results of Silis et al. [1989]. TSURF during any tidal stage in the tidal flats were consistently higher than ambient Ta by an average of 1.6 °C whereas Silis et al. [1989] observed that average TSURF was 1.4 °C *cooler* than Ta. However, these contrasting measurements from each study don't necessarily indicate that either one of the studies was in err. There was a key difference between the two methodologies that were used to estimate surface temperature; Silis et al. [1989] were measuring the temperature of the water column and the substrate with shielded thermocouples and thermisters whereas this study was inferring surface temperature by measuring the long wave radiation from the surfaces in question. In this study, only the temperature at the immediate interface between the atmosphere and the water or substrate was known. Assuming that both estimates of surface temperature could be applied to either study,

this indicates that the surface of the water was responsive to radiative and turbulent inputs of energy and that the underlying water remained cool because the water column is not particularly well-mixed at the low-tide mark and is thermally stratified.

The sensible heat fluxes measured in the tidal flats during the summer of 2006 were markedly different from those measured in the study by Silis et al. [*1989*]. In the previous study, H was very small and composed only a marginal proportion of the energy budget (<5 %). In this study, they ranged from 19 to 36 % of net radiation depending on the time of day (Table 3.7). Importantly, this proportion was also observed to vary with tidal stage; when the tidal flats were drained during the 'TROUGH' stage or when the tidal flats were completely flooded during the 'PEAK' stage, sensible heat only composed 16 to 18 % of the turbulent energy budget. However, when the water was in motion over this surface, the proportion of the energy budget allotted to H rose to 20 to 22 % of RNET (Table 3.6). This was driven largely by negative fluxes during the night when H actually exceeded LE thus resulting in negative Bowen Ratios that were well above unity (Table 3.8). This suggests that the motion of the water travelling over the tidal flats during the inflow and outflow stages ('FLOW' and 'EBB' respectively) is mixing otherwise stratified water.

Average LE in the tidal flats during 2006 was comparable to the results of Silis et al. [1989] and was estimated to comprise 35 to 39 % of RNET. However, there did not appear to be any significant differences observed between the tidal stages. This is congruent with the observation that, even when the tidal flats are drained during the 'TROUGH' stage, that there are significant quantities of ponded water and the underlying substrate tends to remain saturated.

Importantly, a similar diurnal pattern of fluxes and forcings that was observed at the

Near-Shore Station in 2005 on DOYs 201 and 223 was also observed at the Tidal Flats Station in 2006 on DOYs 177, 210 and 240. On each of these days, there was cloud cover present during the mid to late afternoons. Average wind speed decreased into the late afternoon and was accompanied by a steep increase of ambient air temperature, surface temperature and vapour pressure deficit. There was a trend towards instability and small peaks in both the sensible and latent heat fluxes in the middle of each afternoon (Figure 3.16 and Figure 3.18).

The average flux of CO₂ was a single order of magnitude less than other studies performed in similar high-latitude marine environments (Table 1.1) but clearly indicated that the tidal flats were a source of CO₂ over the course of the sampling period with an average efflux of $0.04 \,\mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$. This is consistent with the findings of Else et al. [2008] where it was found that the efflux of CO₂ tends to increase with decreasing distance to the shoreline of Hudson Bay. Although Else et al. [2008] did not sample such shallow waters as the ones sampled during this study, they suggested that this increasing efflux is due to the proximity of nearby river plumes that tend to be supersaturated with CO₂. However, there was a very considerable range of flux estimates from the tidal flats; there was large daytime drawdown of CO₂ of -0.45 μ mol·m⁻²·s⁻¹ and a very large nighttime efflux of 1.42 μ mol·m⁻²·s⁻¹. Therefore, the tidal flats are a source of CO₂ primarily due to night time efflux but there is a considerable amount of primary productivity occurring during day time conditions. In contrast to the findings of Zemmelink et al. [2009], the flux of CO₂ exhibited large variations in response to tidal stage. Both 'FLOW' and 'PEAK' stages were sinks for CO₂ whereas 'EBB' and 'TROUGH' stages were sources of CO₂ (Table 3.9). These fluxes tended to be the largest in either direction when the water had the highest velocity during the 'FLOW' and 'EBB' stages at -0.29 and 0.42 μ mol·m⁻²·s⁻¹ respectively.

Similar to the flux of sensible heat it is suggested that the flux of CO_2 , positive or negative, is enhanced by the rapid mixing of the water column.

4.3 - The Coastal System

A comparison of the datasets from the Near-Shore and the Tidal Flats Stations indicates that turbulent fluxes were more intense in the Tidal Flats due to the mixed state of the water column, not due to increased wave height. Carbon dioxide fluxes were likely being driven by intense biological activity and primary production. Primary production is favoured in the near-shore, whereas night time respiration appears to be marginally favoured in the intertidal zone. Water is being transferred over a relatively short distance between the intertidal and the near-shore zones. As well, the observations of fluxes and forcings at each site may be consistent with the frequent development of meso-scale circulation cells due to the sea-breeze effect.

Waves are known to increase in height and steepen as they approach a shore due to the decreasing depth of the water column. Panin and Foken [2005] demonstrated that there is a positive correlation between the ratio of wave height (*h*) to water depth (H) and turbulent fluxes. They suggest that as waves become more 'abrupt' [*sic*] in the shallow zone, the surface roughness length increases and turbulent exchanges with the atmosphere are enhanced due to breaking waves. Although this explanation is easy to conceptualize and is well documented in their own research on the Caspian Sea, this assertion is not supported by the measurements of roughness lengths that were made in this study. Average roughness lengths were slightly higher in the near-shore zone in 2005 (0.034 m) than in the tidal flats in 2006 (0.028 m) despite the comparable average wind speeds from each year (Tables 3.1 and 3.5). Therefore, the higher

average turbulent fluxes measured in the tidal flats when they were inundated with water cannot be explained by wave action alone and that conditions within the water column may have been playing a primary role with the generation of fluxes.

This could also be supported by the observation that fluxes tended to be more intense in the tidal flats and in the near-shore zone when the water was actively moving between the high and low tide marks during the 'EBB' and 'FLOW' stages. Water has been measured moving as fast as $1 \text{ m} \cdot \text{s}^{-1}$ over the assemblage of rocks in the flats and can be assumed to be well-mixed when it's moving [Silis et al., 1989]. That water column conditions are a key forcing of turbulent fluxes is further supported by the observation that as water moves between the near-shore and intertidal zones, it tends to bring the properties of the previous zone into its new location. Tables 3.1 and 3.5 indicate that drawdown was favoured in the near-shore zone and efflux was favoured in the intertidal zone. The notion that this could be due to the presence of living kelp in the nearshore zone and dead kelp in the intertidal zone is supported by the fact that there is a net drawdown during the day and efflux during the night. When water moved from the near-shore zone to the intertidal zone during the 'FLOW' stage, it continued to be a sink for CO₂. Once in the intertidal zone, the water presumably equilibrated with its new surroundings and, at least during the night time, became net heterotrophic. Likewise, when this water migrated out of the intertidal zone to the near-shore zone during the 'EBB' stage, it continued to be a source of CO₂. Of course, this kind of interaction depends on the assumption that both the near-shore and intertidal zones were behaving in the same way during both 2005 and 2006.

The observations from the DOYs 201 and 223 from 2005 and DOYs 177, 210 and 240 from 2006 and consistent with the development of coastal meso-scale circulation cells that are

the result of the 'sea-breeze' effect which is frequently superimposed on synoptic scale weather patterns. As the nearby land surface heats up over the course of the day, a pressure gradient builds up between the warm land and the adjacent cooler sea. This will result in onshore winds that will warm and ascend thus creating high pressure aloft. Because there has been a steady movement of air next the surface from the sea to the land, there will be low pressure aloft above the sea. Therefore, there will be a high altitude wind that will move the rising air mass from the land to the sea (Figure 1.5). As it rises, there is often a release of latent heat into the air mass as the moist air condenses [*Kraus et al., 1990*] and low level clouds form. It is suggested here that if any precipitation occurs over the course of this process, then the air mass will be both comparatively warmer and drier as it cools and descends back over the sea. This would be consistent with the following observations from this study:

- The frequent occurrence of cloud cover during the afternoons this occurs due to condensation occurring in rising, moist air masses from the land that are then blown over the adjacent sea
- Late afternoon onshore winds that are warm and have high vapour pressure deficits if any precipitation has occurred, then the air mass will be comparatively dry as it descends over the sea
- The frequent occurrence of negative sensible heat fluxes in the near-shore zone warm air over cool water will result in a turbulent transfer of heat from the air mass to the water
- The frequent occurrence of positive spikes in the flux of sensible heat due to the rapid warming of the upper layer of stratified water in the near-shore zone
- Late afternoon positive spikes in the flux of latent heat dry air masses will encourage

the transfer of latent heat from the water

- The suppression of lateral winds due to the presence of descending air masses
- A switch from stable to unstable boundary layer conditions due to reduced horizontal wind speeds and the presence of convection

5 - Summary and Conclusions

What would otherwise have been a set of dubious estimations with some counter-intuitive results and varying degrees of certainty has become understandable in the context of the sea-breeze phenomenon coupled with an understanding of the role of water depth, water-mixing and biological activity in the both the Near-Shore and the Intertidal zones. Although no direct measurements were made that would have properly documented the role of these three effects, the datasets are congruent with these suggestions.

This study has added to a developing body of knowledge regarding turbulent exchanges of mass and energy in high-latitude intertidal and near-shore zones. Basic estimations of fluxes and forcings have been formulated through direct observations which improve our understanding of the processes at work in Hudson Bay. Specifically, the observations of Else et al. [2008] regarding the carbon budget of Hudson Bay and the influence of the proximity of the coastline and its associated river plumes has been extended to the shoreline. The weak sink of CO₂ that was observed in the waters of Hudson Bay has been shown to intensify through the near-shore zone and become a net source in the intertidal flats. Key differences with the study of Zemmelink et al. [2009] were highlighted and tidal stages have been shown to exert a distinct influence on the observed carbon fluxes. In terms of energy budgets, this study has added to the observations of Mahrt et al. [1998] and Serreze et al. [2007] with parameterizations of sensible and latent heat in coastal environments during onshore winds. As well, observations of the energy budget in the intertidal flats have significantly augmented those of Silis et al. [1989] and have highlighted differences in the role played by sensible heat.

This study has shown that the intertidal and the near-shore zone are fundamentally

different due to the change in water depth and the presence of dead and living kelp in this area. They are, however, linked by the orthogonal movement of tidally-driven water masses and by meso-scale circulation cells that are unique to coastal environments. The proximity of these two zones highlights the importance of surficial heterogeneity in subarctic regions and the need to develop a process-level understanding in order to understand these complex environments.

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Appendix A

Hardware

The Near-Shore Station, 2005										
			Variable(s)							
Sensor	Height (m) Description	Measured	Units	Manufacturer					
LI7500	14.8	Infrared Gas Analyser	ρCO ₂	mmol·m ⁻³	LiCor					
			ρH₂O	mmol∙m ⁻³						
			PA	kPa						
CSAT3 Anemometer	14.8	3-D Sonic Anemometer	u',v',z',Ts	m⋅s ⁻¹	CSI ²					
Wind Monitor	16.2	RMYoung Wind Monitor	WS	m⋅s⁻¹	RMYoung ³					
		5	WDIR	0	Ū					
HMP45C	13.3	T.& RH Probe	Ta	°C	Vaisala⁴					
			RH	%						
Pyrradiometer	9.3	Net Radiation	RNET	W∙m⁻²	REBS⁵					
Prec. Spec. Pyranometer	9.3	Inc. Short Wave Rad.	Rg	W⋅m⁻²	Eppley ⁶					
Prec. Spec. Pyranometer	9.3	Ref, Short Wave Rad.	Rg_out	W⋅m⁻²	Eppley ⁶					
Prec. Infrared Radiometer	9.3	Long Wave Rad. Out	RI_out	W⋅m⁻²	Eppley ⁶					
IR Transducer	9.3	Surface Temperature	TSURF	°C	Apogee ⁷					
The Intertidal Station, 2006										
			Variable(s)							
Sensor	Height (m) Description	Measured	Units	Manufacturer					
LI7500	6.8	Infrared Gas Analyser	ρCO ₂	mmol⋅m⁻°	LiCor'					
			ρH₂O	mmol∙m ⁻³						
			PA	kPa						
CSAT3 Anemometer	6.8	3-D Sonic Anemometer	u',y',z',Ts	m⋅s⁻¹	CSI ²					
Wind Monitor	7.1	RMYoung Wind Monitor	WS	m⋅s⁻¹	RMYoung ³					
			WDIR	0						
HMP45C	5.8	T.& RH Probe	T _a	°C	Vaisala⁴					
			RH	%						
Prec. Spec. Pyranometer	6.1	Inc. Short Wave Rad.	Rg	W⋅m⁻²	Eppley ⁶					
¹ Campbell Scientific Incorporated, Logan, Utah, USA										
² Licor Instruments, Lincoln, Nebraska, USA										
³ RMYoung Company, Traverse City, Minnesota, USA										
Vaisala Incorporated, Woburn, Massachusetts, USA										
Radiation Energy Datatice Systems, Seattle, Washington, USA										
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Appendix **B**

Julian Day/Calendar Converter

	Regular Years - (2002, 2003, 2005, 2006, 2007, 2009 etc.)												
		Jan	Feb	Mar	Apr	Мау	Jun	Jul	Aug	Sep	Oct	Nov	Dec
	1	1	32	60	91	121	152	182	213	244	274	305	335
	2	2	33	61	92	122	153	183	214	245	275	306	336
	3	3	34	62	93	123	154	184	215	246	276	307	337
	4	4	35	63	94	124	155	185	216	247	277	308	338
	5	5	36	64	95	125	156	186	217	248	278	309	339
	6	6	37	65	96	126	157	187	218	249	279	310	340
	7	7	38	66	97	127	158	188	219	250	280	311	341
	8	8	39	67	98	128	159	189	220	251	281	312	342
	9	9	40	68	99	129	160	190	221	252	282	313	343
	10	10	41	69	100	130	161	191	222	253	283	314	344
	11	11	42	70	101	131	162	192	223	254	284	315	345
	12	12	43	71	102	132	163	193	224	255	285	316	346
_	13	13	44	72	103	133	164	194	225	256	286	317	347
nth	14	14	45	73	104	134	165	195	226	257	287	318	348
Day of Mo	15	15	46	74	105	135	166	196	227	258	288	319	349
	16	16	47	75	106	136	167	197	228	259	289	320	350
	17	17	48	76	107	137	168	198	229	260	290	321	351
	18	18	49	77	108	138	169	199	230	261	291	322	352
	19	19	50	78	109	139	170	200	231	262	292	323	353
	20	20	51	79	110	140	171	201	232	263	293	324	354
	21	21	52	80	111	141	172	202	233	264	294	325	355
	22	22	53	81	112	142	173	203	234	265	295	326	356
	23	23	54	82	113	143	174	204	235	266	296	327	357
	24	24	55	83	114	144	175	205	236	267	297	328	358
	25	25	56	84	115	145	176	206	237	268	298	329	359
	26	26	57	85	116	146	177	207	238	269	299	330	360
	27	27	58	86	117	147	178	208	239	270	300	331	361
	28	28	59	87	118	148	179	209	240	271	301	332	362
	29	29		88	119	149	180	210	241	272	302	333	363
	30	30		89	120	150	181	211	242	273	303	334	364
	31	31		90		151		212	243		304		365