

**Spatial and temporal changes of photosynthetically available radiation,
temperature, and salinity beneath a variable sea ice cover**

by

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ABSTRACT

Melt ponds greatly increase the transmission of solar radiation through sea ice relative to snow covered or bare ice. This rise in transmittance has the potential to enhance water column heating and primary production. I examine how spatially variable sea ice surfaces control the under-ice salinity, temperature and photosynthetically active radiation (PAR) and provide estimates of solar heating and primary production during melt. Conductivity, temperature and PAR profiles were measured in the Canadian Arctic under snow covered ice, leads, bare ice and melt ponds. The under-ice light field to a depth of 10 to 13 m was highly variable, controlled by increased transmission under melt ponds and shading by bare ice. Below, the light field became relatively homogeneous showing the depth the surface heterogeneity had an effect on transmitted PAR. Furthermore, one water column profile is not representative of the PAR, salinity or temperature under a spatially heterogeneous surface.

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CHAPTER 1 INTRODUCTION

1.1 MOTIVATION

The principal motivation of my thesis is to examine the spatial variability of light under sea ice during freezing and melting in the Arctic. These data are required to understand the spatial and temporal scales of photosynthetically available radiation (PAR) beneath and ice and also for calculations of the influence of shortwave radiation on thermal heating beneath ice. In turn, this is important because of the essential role that sub-ice producers play in the overall productivity of the Arctic marine system. Previous researchers usually measure light above the ice and in the ice and then make references to what the light field looks like under the ice and at depth. Traditionally in situ temperature and salinity measurements start at 5 to 10 m below the ice and just assume the top layer is homogeneous. As you will see in this thesis this assumption is not always appropriate as the light environment 2 to 5 m under the ice can be significantly different than in nearby locations at the same depth and also throughout the rest of the water column.

1.2 THESIS OUTLINE

There are seven chapters included in this thesis. The first chapter describes the motivation for doing the project, gives an outline of all of the chapters in the thesis, and provides the three main objectives of the project. The second chapter is the literature review and gives background information and a review of literature on solar radiation interactions with the atmosphere, snow, ice, melt ponds and the ocean. It also describes heat fluxes in the Arctic sea ice-ocean system and gives a brief summary on primary production in the upper ocean under sea ice.

The third chapter is the methods chapter and provides information on the study area of the project, the International Polar Year (IPY)-Circumpolar Flaw Lead (CFL) system study, the field methods and the datasets acquired during freezing and those during melting, and the data processing. This chapter also gives information on the equations and steps for completing the spatial variability calculations and statistics as well as how the potential heating and primary production results were derived.

The fourth chapter is the results and discussion chapter. It gives some of the following information:

- The physical sea ice environment during freezing and melt
 - Pond or snow depth
 - Ice thickness
 - Albedo
 - Water column PAR, salinity and temperature
- Temporal and spatial variability of PAR transmission
 - Results from profiles completed in the field show the temporal and spatial variability of PAR transmission under the ice during freezing and melting
 - Statistical analysis of profiles at sites with heterogeneous surface cover to determine when surface covers result in significantly different PAR profiles under the ice and to what depth under the ice they differ
 - Effective attenuation profile calculated for a site with a heterogeneous surface cover using profiles under ponds, white or bare ice and a surface melt channel
 - Discuss in what way these results affect how sites for profiles should be selected in the future for the best results

- Potential solar heating during melt using the radiant heating rate calculation
 - Completed under white ice and pond conditions for one site with multiple pond and white ice profiles
- Potential primary production during melt using photosynthesis-irradiance (P-I) calculations
 - Completed under white ice and pond conditions for one site with multiple pond and white ice profiles

Chapter five summarizes the results and discussions and gives a conclusion of the thesis project. This chapter also gives some future direction for research on the topics and results discussed in the thesis. The last and sixth chapter lists the references used in the thesis.

1.3 OBJECTIVES

The overarching objective of my thesis is to understand the role of surface heterogeneity on in the transmission of shortwave radiation through the ocean-sea ice-atmosphere (OSA) interface and to quantify this role in terms of PAR and thermal heating of the ocean near the bottom of the ice. More specifically I address three interrelated objectives:

1. Examine PAR and the ocean mixed layer salinity and temperature distribution from freezing to melt under different sea ice surface conditions;
2. Investigate the role spatial heterogeneity of surface types (e.g., snow depth, ice thickness, melt pond coverage, etc.) plays in determining the vertical distribution of transmitted irradiance in the upper water column;
3. Calculate the potential heating and primary production that occur as a result of radiation transmitted through melt ponds and bare ice.

Taken as an amalgam these objectives will address a current gap in our knowledge of the OSA system, namely the role which surface heterogeneity plays in control of energy to the upper part of the ocean. This is important because of the rapid changes we are currently seeing in the sea ice system, with earlier melt, long melt period conditions and because of a transfer from a multiyear ice dominated to first-year ice dominated Arctic Ocean [*Barber et al.*, 2010].

CHAPTER 2 BACKGROUND

Solar energy transmitted through snow, melt ponds, and sea ice supplies heat to the ocean and energy for photosynthesis. The transmission of energy through the sea ice, snow and melt ponds is highly variable spatially and changes diurnally and seasonally with solar elevation [Perovich *et al.*, 1998]. The understanding of the interactions of solar radiation with the OSA system is important for properly representing the ice-albedo feedback mechanisms and heating in models which can affect prediction of the future sea ice extents in the summer [Holland *et al.*, 2006].

There is limited data studying light, salinity and temperature directly under the ice down to 10 or 20 m. Most studies measure the light above the ice [Perovich *et al.*, 1998; Hanesiak *et al.*, 2001; Ehn, 2006], in the ice or at the ice bottom [Perovich *et al.*, 1998; Ehn *et al.*, 2008; Light *et al.*, 2008]. Measuring the light under the ice in the upper ocean can be difficult, time consuming and expensive. You either need to have a dive program with under ice divers that can carry the instruments under the ice and drop them [Ehn *et al.*, 2011], drill a large hole and use large equipment and cover the hole with some material or you have to auger small holes and manually do profiles. Some also measure further down in the water column since the light under the ice is affected by the large hole or cover used and then model what the light field might look like closer to the ice bottom

[Belzile *et al.*, 2000]. Because of the heterogeneous surface cover in the melting season in the Arctic with melt ponds, many profiles can be needed to get enough data to study the heterogeneity of light field. Two recent studies that have been completed under the ice by *Ehn et al.* (2011) and *Frey et al.* (2011). Both studies found that there is a horizontal spreading of light under the bare ice from melt ponds.

The solar radiation availability for heating and primary production in the ice and in the ocean below the ice depends on a number of different aspects of the OSA system. These factors affect the intensity of the irradiance, distribution and the spectral quality.

The fractions of the different wavelengths of the electromagnetic spectrum that are reflected, transmitted or absorbed by the snow, ice or water is vital to the heating and primary production. The quantities of these different wavelengths of the electromagnetic spectrum will have different effects on the physics, chemistry and biology of the OSA system [Lukas and Soloviev, 2006]. For example, photosynthetically active radiation or PAR is 400-700 nm, which is in the visible part of the spectrum, is what plants need to carry out photosynthesis to live [Melnikov, 1997]. PAR is indicated in Table 1 where the wavelengths and names of the ranges of the electromagnetic spectrum that are most often used are shown.

TABLE 1. Commonly used wavelength ranges and some examples of their interactions [Thomas and Stamnes, 1999].

Subregion	Range	Solar Variability	Comments
X rays	$\lambda < 10 \text{ nm}$	10-100%	Photoionizes all thermosphere species
Extreme UV	$10 < \lambda < 100 \text{ nm}$	50%	Photoionizes O_2 and N_2 Photodissociates O_2
Far UV	$100 < \lambda < 200 \text{ nm}$	7-80%	Dissociates O_2 Discrete electronic excitation of atomic resonance lines.
Middle UV or UV-C	$200 < \lambda < 280 \text{ nm}$	1-2%	Dissociates O_3 in intense Hartley Bands. Potentially lethal to biosphere.
UV-B	$280 < \lambda < 320 \text{ nm}$	<1%	Some radiation reaches surface, depending on O_3 optical depth. Responsible for skin erythema.
UV-A	$320 < \lambda < 400 \text{ nm}$	<1%	Reaches surface. Benign to humans. Scattered by clouds, aerosols, and molecules.
Visible or PAR	$400 < \lambda < 700 \text{ nm}$	$\leq 0.1\%$	Absorbed by ocean, land. Scattered by clouds, aerosols, and molecules. Primary energy source for biosphere and climate system.
Near IR	$0.7 < \lambda < 3.5 \text{ }\mu\text{m}$		Absorbed by O_2 , H_2O , CO_2 in discrete vibrational bands.
Thermal IR	$3.5 < \lambda < 100 \text{ }\mu\text{m}$		Emitted and absorbed by surfaces and IR-active gases.

There are three outcomes that can happen to solar radiation as it interacts with the atmosphere, snow, water and ice, namely: reflection, absorption and transmission (see Figure 1 and Equation 1). The wavelength of the radiation and

the composition of the material it interacts with will have an effect on the relative partitioning into reflection, absorption and transmission.

EQUATION 1

Incident Radiation = reflection + absorption + transmission

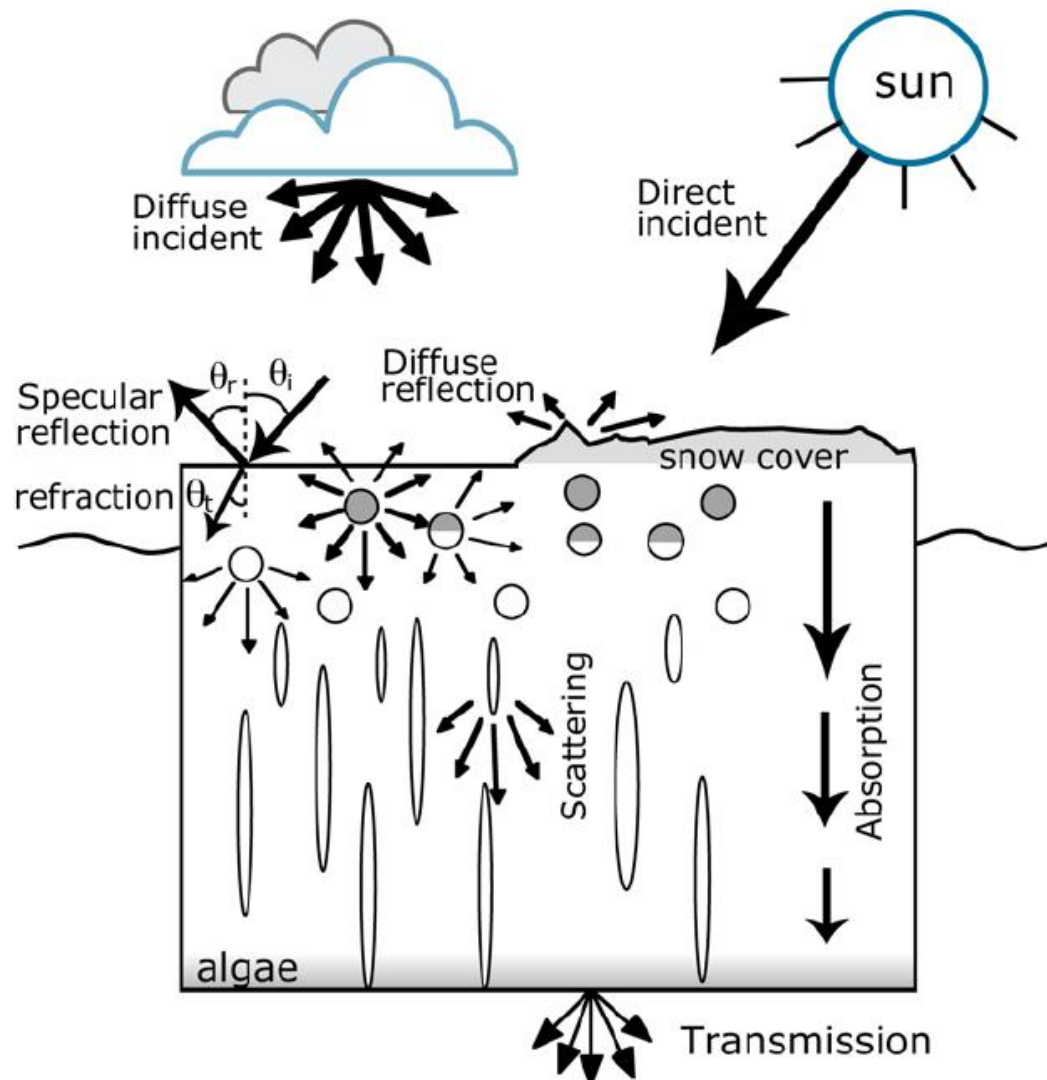


FIGURE 1. Radiative transfer in the atmosphere-sea ice-ocean system in the spring/summer [Perovich, 1998b; Eiken, 2003; Ehn, 2006].

The fraction of the incoming solar radiation reflected off the surface is called the albedo of the surface. When talking about albedo, there is the spectral albedo and then total shortwave albedo. Spectral albedo is the fraction of the incident irradiance that is reflected from the surface using individual wavelength measurements. The total shortwave albedo is the integral of the total (all-wavelength) albedo over the entire range of the solar shortwave spectrum [Eiken, 2003]. In this thesis, there will also be reference to PAR albedo which is albedo calculated using the range of the wavelengths of PAR. The seasonal evolution of surface albedo is one of the most important features when researching solar radiation distribution in the Arctic. The albedo of the surfaces in the Arctic can cover almost the entire range of albedo as seen in Figure 2 [Eiken, 2003].

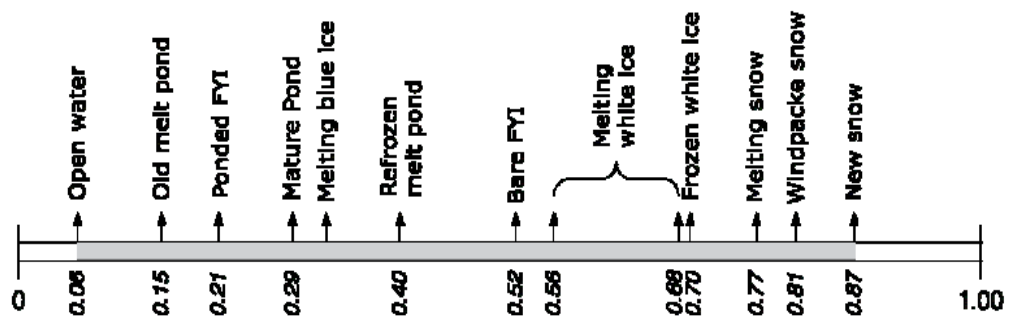


FIGURE 2. Observed total albedo for different ice, snow and melt ponds types in the Arctic [Perovich, 1996; Ehn, 2006].

An interesting thing about ice, albedo and transmission of shortwave radiation is that because of the contents and structure of sea ice, less than half of the incoming radiation enters the ice or reaches the sea water below whereas ice on a freshwater lake can have over 75 % entering the ice and water [*Eiken*, 2003].

The timing of sea ice melt represents a critical period in the sea ice-albedo feedback mechanism. The sea ice-albedo feedback mechanism is most easily described as the period when the surface has a lower albedo such as open water, bare ice or melt ponds then more irradiance will be absorbed and the surface will warm up. Once the surface warms up, the ice will melt, or the melt ponds will increase in size and absorb more irradiance causing more melting and the ocean will warm up which will cause more melting and so on [*Curry et al.*, 1995; *Perovich*, 1998b].

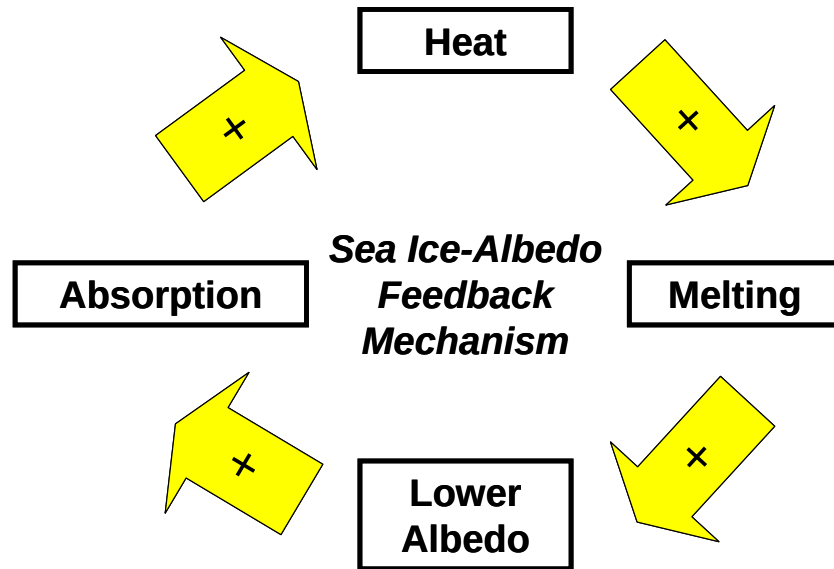


FIGURE 3. The sea ice-albedo feedback mechanism with the arrows showing increases of heat, melting, lowering the albedo and absorption of irradiance.

The incidence angle of solar radiation, θ_i (solar elevation or zenith angle) can have an effect on whether the incident radiation is reflected, absorbed and transmitted. The lower the sun is in the sky then there will be more reflection off the surface of the ice or snow, which means less radiation is absorbed within the snow or ice and less is transmitted through into the water below [Jin *et al.*, 1994]. This is shown in Figure 1 where there is specular reflection and refraction and the larger the angle of incidence the more reflection there will be. The timing of the beginning of the ice melting and the time of year is very important because of the solar zenith angle. The earlier the melt starts, the longer period of time there will be with the sun higher in the sky combined with open water, melt ponds or bare ice because the melt occurs during the period of highest solar

elevation. This also means that there will be less time with a higher reflecting snow cover when the sun is higher in the sky and the incoming solar radiation is more intense. Therefore, there will be higher transmission to the ocean for a longer period of time which will result in increased warming since there is less reflection with a lower zenith angle and the solar radiation has a higher intensity at that time of year. This is another important addition to the sea ice-albedo feedback mechanism which can amplify the reduction of sea ice in the Arctic.

The majority of the discussion in Section 2.1 in this chapter will revolve around the PAR range of solar radiation and its interactions. The range of PAR being 400-700 nm which plants use for photosynthesis is very similar to the visible light range and therefore PAR is sometimes referred to in literature and also in this thesis as light [Kirk, 1994]. Since each of the sections represents such large topics of study on their own, they are only discussed briefly in this thesis.

2.1 SOLAR RADIATION INTERACTIONS IN THE ARCTIC

ATMOSPHERE-ICE-OCEAN SYSTEM

2.1.1 ATMOSPHERE

Solar radiation comes from the sun and is considered direct incident radiation or when it gets to the earth's surface as solar irradiance. The composition of the

atmosphere can affect the composition of wavelengths and intensity of the radiation before it gets to the earth's surface [Siegel *et al.*, 1999; Lillesand and Keefer, 2000; Makshtas and Korsnes, 2001; Lukas and Soloviev, 2006] which is called the spectral distribution. In the atmosphere there are clouds, aerosols, particles, gases, water vapour, ice crystals among others that have an effect on what happens to the incident radiation [Martin, 2004]. The different wavelengths of solar radiation behave differently with each of the atmospheric constituents and can be absorbed, scattered or transmitted. Whether the solar radiation is absorbed, scattered or transmitted depends on the wavelengths of the radiation as well as the size and composition of what it is interacting with. In general, air molecules and dust particles will scatter incoming solar radiation and water vapour, oxygen, ozone and carbon dioxide will absorb it [Kirk, 1994]. Specifically, clouds and aerosols in the atmosphere will scatter radiation in the visible or PAR range (400 - 700 nm) and in the UV-A range (320 – 400 nm) [Thomas and Stamnes, 1999].

Cloud coverage in the atmosphere will influence how much radiation and also which wavelengths reach the earth and also how much reflected radiation from the earth is reflected off the clouds back to the earth. If there is a higher concentration of clouds, the amount of shortwave radiation incoming from the sky will decrease. If the earth's surface has a high albedo there will be multiple reflections off the surface then off the clouds and then back to the earth [Inoue

et al., 2005]. Furthermore, if the solar zenith angle is higher or the incidence angle of the sun is lower and there are clouds then the incident solar irradiance will be even lower [Kirk, 1994, Perovich, 2007].

Once the solar radiation makes it through the atmosphere it interacts with the surface cover on the earth. In the Arctic marine system the first surface to be encountered by this radiation is snow.

2.1.2 SNOW

Snow is a highly reflective surface and a highly scattering medium which light reacts with. The growth history of the snow will determine how light interacts with it.

Snow thickness and distribution is important for determining the distribution of radiation in the OSA system. Snow has an albedo range of 0.77 – 0.87 depending on whether it is new (0.87), wind-packed (0.81) or melting (0.77) [Perovich *et al.*, 1998]. This high albedo means that majority of the incoming shortwave radiation will be reflected back into the atmosphere and the other small amount will be either absorbed in the snow or transmitted into the ice below. This is shown in Figure 4 where there is a significant amount of energy reflected back to the atmosphere by snow and ice compared to the amount reflected by the melt ponds and leads. During the freezing period the snow is dry and has a higher

albedo. Freshly fallen snow has the highest albedo of any Arctic snow, sea ice, melt pond or sea water surface followed by older wind-packed snow, melting snow then different ice types [Perovich, 1998b].

Whether the remaining irradiance reaching the earth, that hasn't been reflected, is absorbed in, or transmitted through the snow, depends on snow depth, density, salinity, and grain size [Warren, 1982]. When studying snow, the snowpack can be split into two or three distinct layers which have different physical properties which evolve over the age of it [Papakyriakou, 1999; Mundy *et al.*, 2005]. The transmission of PAR in the snowpack is affected by snow depth and the extinction coefficient, which incorporates scattering and absorption [Mundy *et al.*, 2005].

Snow scatters and absorbs more of the radiation being transmitted through it than sea ice, sea water or melt ponds. As the snowpack ages before melting, the snow grain sizes increase, the extinction coefficient decreases due to changes in the snow properties and the salinity increases [Warren, 1982; Mundy *et al.*, 2005]. During the melting season, the albedo of the snow will decrease along with the changing properties of the snow. The snow albedo will decrease because of increasing snow grain size, increasing wetness, and decreasing thickness [Haas, 2007]. As time progresses from freezing to melting and the temperature in the snow warms, the snow crystals increase in size [Barber *et al.*,

1995] and the liquid water content changes which in turn affects what happens to the light as it enters or penetrates through the snow. Importantly, the refractive index of water is similar to that of ice. As light is scattered at the crystal boundary, water in the snow acts to effectively increase grain size, hence the prevalence for transmission over scattering [Mundy *et al.*, 2005].

The snow, since it has such a high albedo, can have an effect on the incoming shortwave radiation as well. This can happen if it is cloudy and there is snow with a high albedo because the majority of the incoming radiation is reflected off the snow into the atmosphere then reflected in the atmosphere back to the ice and so on [Inoue *et al.*, 2005]. If there are more areas with a lower albedo, like the open water in the marginal ice zone (MIZ), then there will be fewer occurrences of multiple reflections. There will also be fewer multiple reflections if there is less cloud cover.

One difference in the summer between the seasonal ice zone in the Antarctic and in the Arctic is that during the summer the floes in the Antarctic still remain snow covered whereas in the Arctic the snow will melt and expose the ice surface as well as create melt ponds [Eiken, 2003; Brandt *et al.*, 2005]. This is an important difference since the albedo of snow is higher than that of ice, melt ponds, and open water which means more solar radiation will be reflected back into the atmosphere and not absorbed in the ice or water below the ice.

2.1.3 SEA ICE AND MELT PONDS

Inclusions including their shape and composition such as water, salt, air, algae, other organisms, and detrital matter in the ice will impact the absorbance, scattering or transmittance of light in the ice. Materials such as algae, other organisms, and detrital matter will absorb light passing through the ponds or the ice. Whereas air pockets or bubbles will scatter the light passing through.

As time progresses from freezing to melting and the temperature in the ice warms, the density and percent brine changes which in turn affects what happens to the light as it enters or goes through the ice. The pores in the ice increase in size and as they warm more they join together. This allows more light to go through the ice and less to be scattered in the pores before they are drained [Perovich *et al.*, 1998].

One important thing when dealing with radiation with snow, ice and ponds in the first-year ice that is not landfast is that no location has the same ocean surface for an extended period of time. There can be open water between ice flows and the ice flows are moving. So at one time there is ice and snow or a melt pond covering a location and a short time later there is open water allowing radiation to enter directly and then a short time later there is ice and snow covering that location again restricting the amount of radiation entering the water because of

reflection at the surface, transmission through the snow and ice and absorption in the snow and/or ice, and water.

Determining the aerial albedo or reflectance of a heterogeneous surface cover such as the one described above or during the melting season can be a difficult task due to the multitude of different surface types [Zhou and Li, 2002]. To combat this problem previous researchers have estimated the different ice types in an area along with their albedos. An average albedo is then determined by taking a weighted average of those albedos with the weight determined by the relative area that each ice type covers [Hanesiak *et al.*, 2001; Zhou and Li, 2002; Zhou and Li, 2003; Perovich *et al.*, 2007]. This is important for remote sensing purposes to note that the satellite's footprint across the ice cover in the spring or summer will have many of these different surfaces and can therefore result in a large error if a correction is not used [Zhou and Li, 2002].

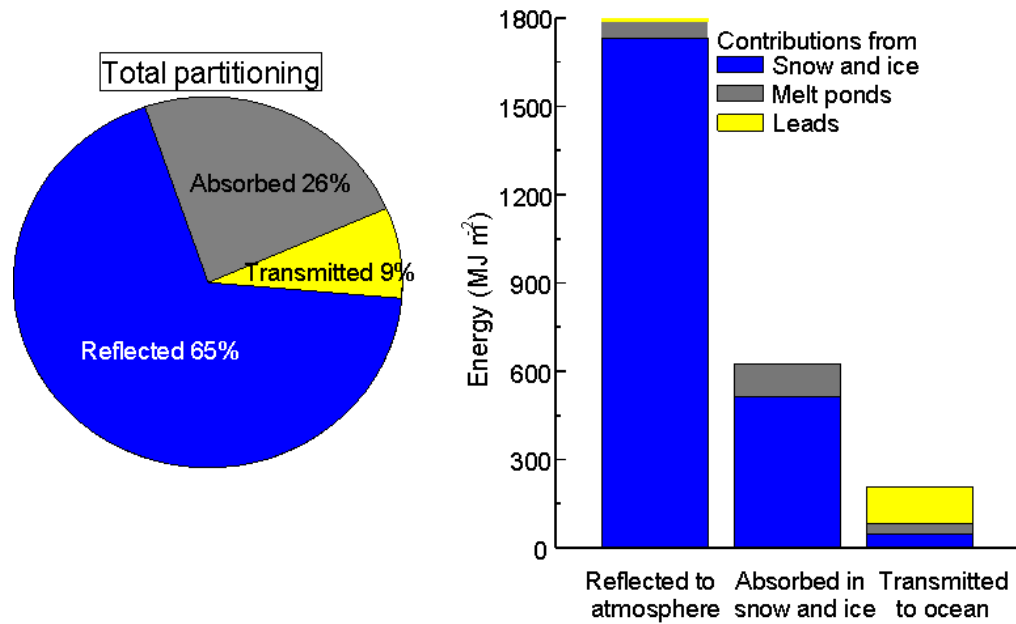


FIGURE 4. The partitioning of radiation into the percentage reflected, absorbed, and transmitted in snow and ice, melt ponds, and leads [Perovich, 2007].

The white surface cover on the ice in the late spring and summer in the Arctic is actually melting sea ice and not snow although it can look a lot like coarse grained snow. This surface is highly scattering and also regenerates itself throughout the melting season so long as it is not covered in a melt pond and it looks quite similar over time. It also has a fairly constant albedo over time at 0.65 ± 0.05 [Perovich, 1998b] or a PAR albedo of 0.70 ± 0.06 [Ehn et al., 2011]. Pond surfaces have recently been found to keep a fairly constant albedo as well even with changing ice thicknesses [Ehn et al., 2011]. The pond covered surfaces allow four to five times more light to pass through the ice than the bare ice surfaces described above with ponds transmitting 38% to 67% and white ice transmitting 5% to 16% [Ehn et al., 2011]. Transmission through melt ponds,

snow covered ice, and bare ice have been found to be significantly different [Maykut and Grenfell, 1975]. There is horizontal spreading of light in the ice which has been shown to increase the light under the bare ice that surrounds the ponded ice and to decrease the amount of light under the edges of the melt pond where surrounded by bare ice [Ehn *et al.*, 2011].

Absorptance is something that is called an inherent optical property (IOP). An IOP is something that depends only on the medium and not the light field. Light will be absorbed in the ice but the amount of sediments and biological particles in the ice will have an effect on the amount of additional absorption [Haas, 2007; Mobley *et al.*, 1998]. If there are more sediments and/or biological particles there will be more absorptance in the ice.

Absorption has a strong spectral dependence meaning that different constituents of what is in the ice, snow and water will absorb more or less radiation at different wavelengths. An example of this is that ice does not absorb a lot of blue light or radiation around 470 nm but it absorbs more light in a shorter distance for radiation at 1000 nm or above [Perovich, 1998b]. Another IOP to be considered for ice is scattering. Brine pockets, air bubbles, and solid salts in the ice will have an effect on the amount of scattering within the ice [Mobley *et al.*, 1998; Haas, 2007].

A property that depends on the medium and the light field are called apparent optical properties (AOP). AOPs are albedo, transmittance, and the irradiance extinction coefficient [Perovich, 2007]. Transmittance is shown in Figure 1 and is the portion of incident irradiance that is transmitted through the medium it is travelling in. Another AOP is the extinction coefficient or K_d which is the amount of light attenuated due to scattering and absorption. The AOPs can be affected by some things like, but not limited to: sky conditions, surface state, sediments and biology [Perovich, 2007].

2.1.4 OCEAN

Water, photosynthetic biota, dissolved yellow pigments (coloured dissolved organic matter (CDOM)), and inorganic particular matter are the four things that absorb light in natural waters [Kirk, 1994]. Though, the inanimate particulate matter also scatters light very strongly. If it is sea water, the red region of the electromagnetic spectrum is attenuated more rapidly than the blue or the green because it is absorbed. The green is attenuated the least and makes it the deepest in the water column and the blue is in between the green and the red [Kirk, 1994].

Depending on where the water, in any particular study area originates, will have a large effect on how much of each absorbing component exists there and how deep the light will reach in the water column before it is absorbed. If the study

location is in a coastal area, the inputs of absorbing and scattering materials into the water column will be high and therefore the light will not reach as far into the water column. Therefore, the available light for primary production will be restricted to the upper water column due to absorption and some scattering. The opposite happens in oceanic areas further from terrestrial inputs. There are fewer absorbing and scattering components other than the sea water and therefore the light reaches deeper into the water column and phytoplankton and the chlorophyll maximum can occur at depths deeper than in coastal locations.

The areas of open water between floes in leads or the MIZ will allow higher values of light into the water column compared to snow and ice covered locations. The albedo of open water is very low and is around 0.08 as shown in Figure 2 where it has the lowest albedo of all the usual surface types in the Arctic [Perovich, 1998b; Ehn, 2006]. This means that it will reflect very little radiation and most of the radiation will be absorbed in the water column. This transmission of radiation in the water column is important for heating but is also important for biological processes such as primary production. This can be especially important during the freezing/melting when there is sufficient incoming radiation but the snow covered sea ice will significantly reduce the light that enters the water column as opposed to the higher irradiance that can enter a lead or the water between floes in the MIZ. Figure 4 shows that the leads or

open water transmit more energy than the melt ponds and snow/ice surfaces put together.

2.2 HEAT FLUXES IN THE ARCTIC SEA ICE-OCEAN SYSTEM

This section briefly investigates the distribution of heat fluxes in the atmosphere-sea ice-ocean system. The energy exchange between the ocean and atmosphere occurs due to horizontal temperature gradients on the surface and the albedo. The temperature gradients create turbulent heat fluxes and albedo causes the heat to either be absorbed or reflected by less or more reflective surfaces respectively. The largest heat flux exchange occurs at the boundary of the ice edge and the open ocean.

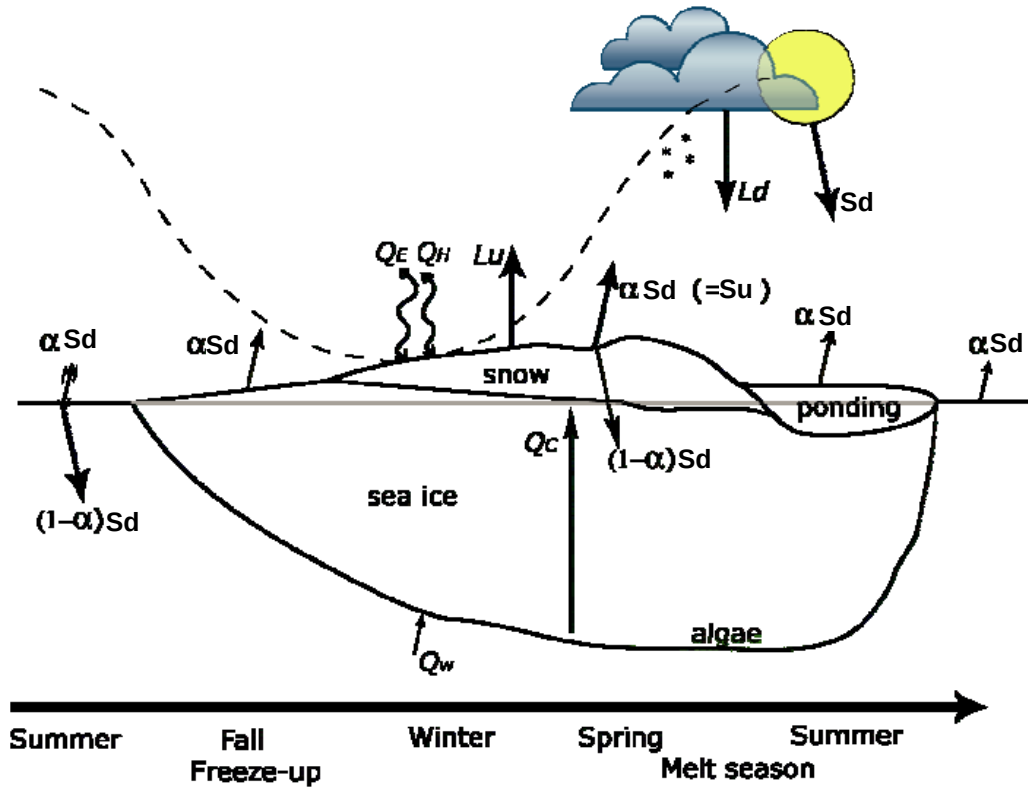


FIGURE 5. Illustration of heat exchanges in the seasonal sea ice system [modified from *Ehn*, 2006].

Figure 5 shows the heat exchanges in the seasonal ice system where S_d is the shortwave radiation downwelling, S_u is the shortwave radiation downwelling, L_d is downwelling longwave radiation, L_u is upwelling longwave radiation, Q_c is conduction, Q_H is the convective exchange of sensible heat, Q_E is the latent heat of evaporation, Q_w is the heat flux from the ocean, α is albedo and Q_M is the phase transition at a surface flux [*Ehn*, 2006]. The net heat exchanges for the system have to be balanced as shown in Equation 2 [*Mundy*, 2007].

EQUATION 2

$$S_d - S_u + L_d - L_u + Q_H + Q_c + Q_E + Q_m = 0$$

The heat fluxes over the ice covered areas are estimated by roughness of the surface and temperature measurements. It becomes difficult to estimate heat fluxes when ice conditions are not uniform, which is a common phenomenon in ice that is not landfast and in leads which have open water between floes and also during the melting season with melt ponds. The surface temperatures in thin, thick and snow-covered ice are highly variable which make it difficult to average out the heat fluxes. If the ice is thin, the surface temperature gets affected by the conductive heat fluxes through the ice. The following two sections will focus on how freezing and melting of sea water, snow and/or ice affects the underlying upper ocean or ocean surface mixed layer.

2.2.1 FREEZING

The freezing or melting of ice is determined by the surface cover, ice water interface salinity, sea surface temperature, water depth and the vertical distribution of salinity. The sea ice has lower salinity and density compared to that of sea water. The freezing point of sea ice of salinity 34 ppm is -1.86°C [Eiken, 2003]. As soon as the sea water temperature reaches the freezing point, ice starts forming and the dissolved solids are left out of the crystallite structure

of the ice. This increases the salinity of the surrounding water and thus lowering the freezing point further. Also, as the water cools it causes convection which means it sinks and is replaced by warmer water which is then cooled and so on until the upper mixed layer is at the freezing point [Eiken, 2003]. As this layer is cooled to the freezing point, ice crystals form called frazil ice. The frazil ice keeps forming and floating to the top until it forms an ice layer.

Once the frazil ice forms a layer of sea ice, there is an additional transitional ice layer which then leads to the vertical ice growth seen in most Arctic ice cores. The cooling of sea ice results in expulsion of salt from it. This causes convection in the surface mixed layer and the depth of mixed layer increases. At the time of melting the fresh water takes the place of salt water and fresh water remains on the surface. The depth and vertical distribution of salinity determines the thickness of the water layer involved in the cooling process [Eiken, 2003].

2.2.2 MELTING

This upper layer of warm water caused by heat supply from the atmosphere is the main cause for the lateral and bottom melting of the ice floes in the Arctic and Antarctic [Ohshima *et al.*, 1998]. The heat supply from the atmosphere is more substantial than the heat supply from deeper in the ocean with the heat from the deeper ocean supplying less than 25% of the total heat [Maykut and McPhee, 1995; Nihashi and Ohshima, 2001]. The lateral and bottom melting

then creates positive feedback since it increases the area of open water which in turn decreases the albedo and increases the absorption of the solar radiation in the open water. The heating of the upper ocean creates a warmer stratified layer at the top. This stratified zone can also be nutrient rich and with the light availability it creates positive conditions for a phytoplankton bloom [Falk-Peterson *et al.*, 2000].

Snow on the sea ice will affect the melting of the ice because snow is so insulating and reflective [Mundy *et al.*, 2005]. Figure 6 shows measured temperature profiles in snow and ice as they warm up. As shown in the profiles, the snow is insulating the ice where the top of the snow cover is warmer during days 150 and 160. This is when the air temperature is warmer but the temperature decreases in the snow towards the ice cover because of the insulating capacity of the snow. Studies have also shown that the thinner the snow is the faster the snow will melt [Iacozza and Barber, 2001].

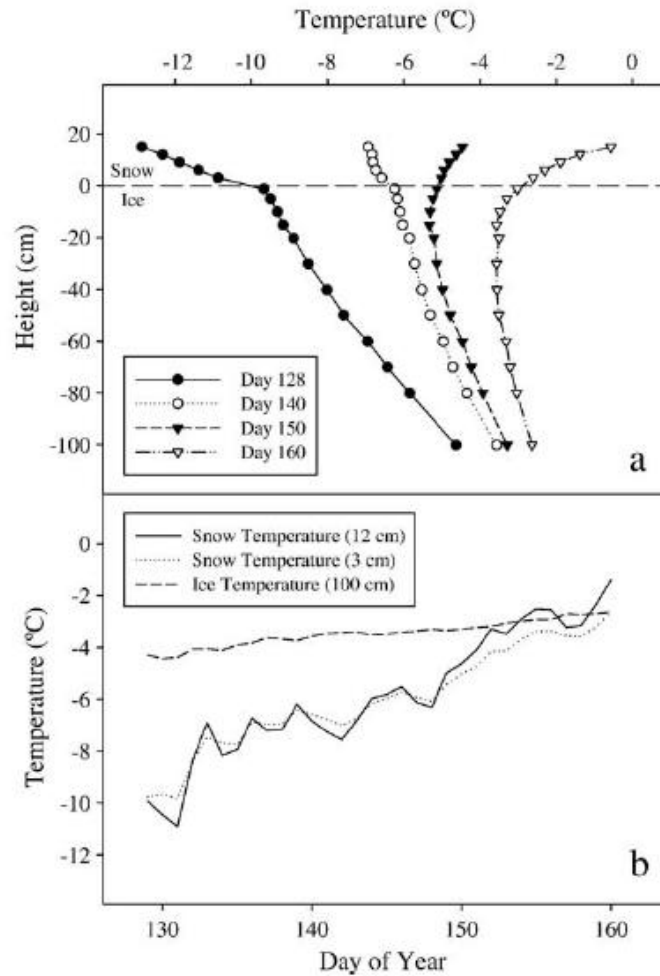


FIGURE 6. Vertical profiles of snow and ice temperatures (a) and time series of daily averaged snow and ice temperatures (b) [Mundy *et al.*, 2005].

Ohshima et al. [1998] carried out a cruise in the Antarctic across highly concentrated ice areas to open ocean during the melting season. They use a term called ice margin which is considered the areas near the open ocean and polynyas which have a lower ice concentration. They found in their study that in the ice margin there is a negative correlation between ice concentration and

water temperature and salinity. The opposite was found in the ice interior areas with a higher ice concentration.

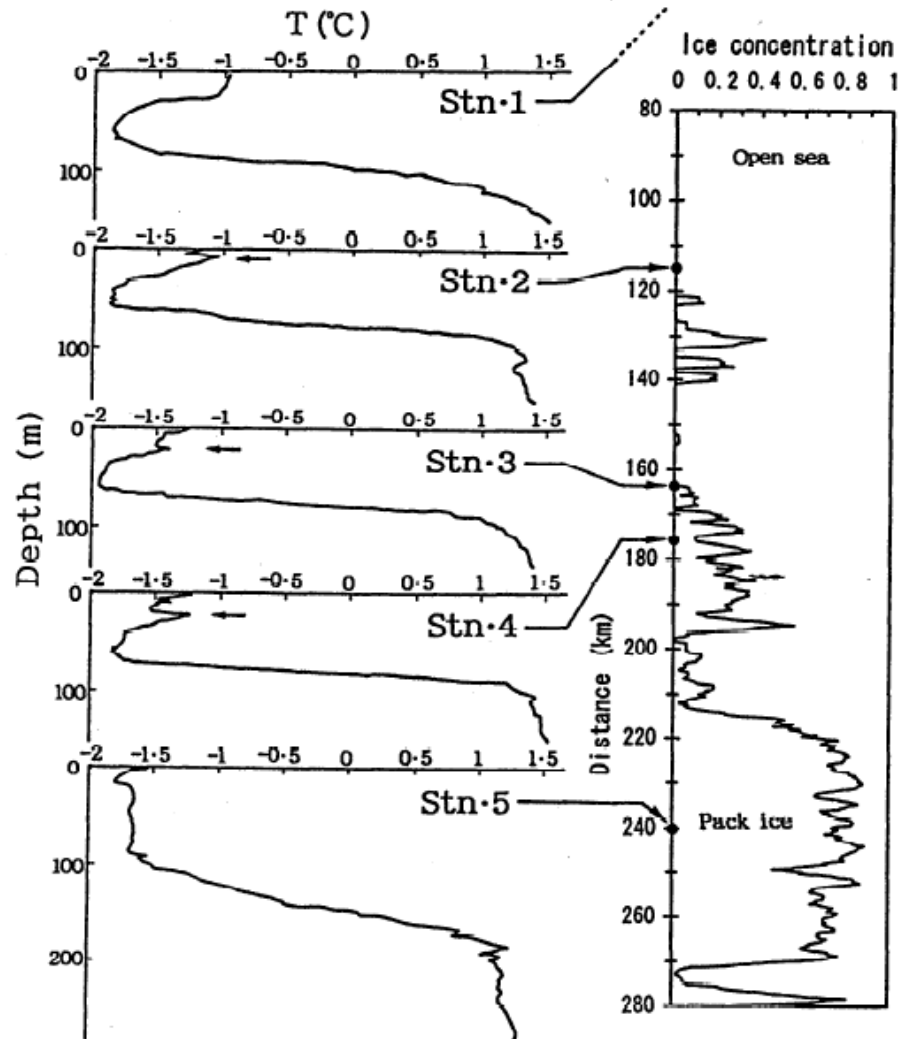


FIGURE 7. Temperature profiles for different ice concentrations from open water to pack ice in the Antarctic Ocean [Ohshima *et al.*, 1998].

What was also found was that the ice floes along the ice edge or areas of low ice concentration were being advected out into the open water which was already warmed by solar radiation. The temperature profiles show that once the ice is

advected into the warmer open water that it melts there and cools the top of the ocean shown in Figure 7 [Ohshima *et al.*, 1998]. The lower temperature is a result of the release of latent heat from the melting ice [Ohshima *et al.*, 1998]. The amount of ice that is advected into the warmer open water is affected by the dynamic activity of the ice [Maykut and McPhee, 1995] i.e. the wind speed and direction.

Another source of heat into the ocean is solar radiation going through melt ponds and the ice beneath and entering the ocean below. This is because the water in the melt pond has a low albedo. Melt ponds are important in the Arctic summer because they can cover between 5 and 80% of the ice area [Barber and Yackel, 1999; Luthje *et al.*, 2006]. The transmission of radiation through these melt ponds can also add to the heat needed to warm the mixed layer beneath the ice and add to the heat already there to assist with bottom ice melting and add to the ice-albedo feedback mechanism. As sea ice warms, the pores in the ice get larger and then start to join. Once the pores join and the ice continues to warm, they get larger and larger increasing the permeability of the ice [Eiken, 2003]. Once the melt ponds get deeper and the ice underneath gets thinner from melting below and above the melt pond bottom, the ice will weaken and the melt pond will drain warmer, less saline water into the ocean below the ice floe [Thomas, 2004]. Melt water will also drain into holes in the ice created by seals and into flaws in the ice [Ehn *et al.*, 2011]. The melt water along with the

melt water from lateral melting forms a stratified warmer, less saline layer of sea water as shown in Figure 8 [Hop *et al.*, 2011]. Once this occurs the ice becomes weaker and what some people call ‘rotten ice’ and it will start to break up into smaller floes. Compared to the 1980s, melt in the 2000s starts on average 13 days sooner in the northern hemisphere [Stroeve *et al.*, 2006].

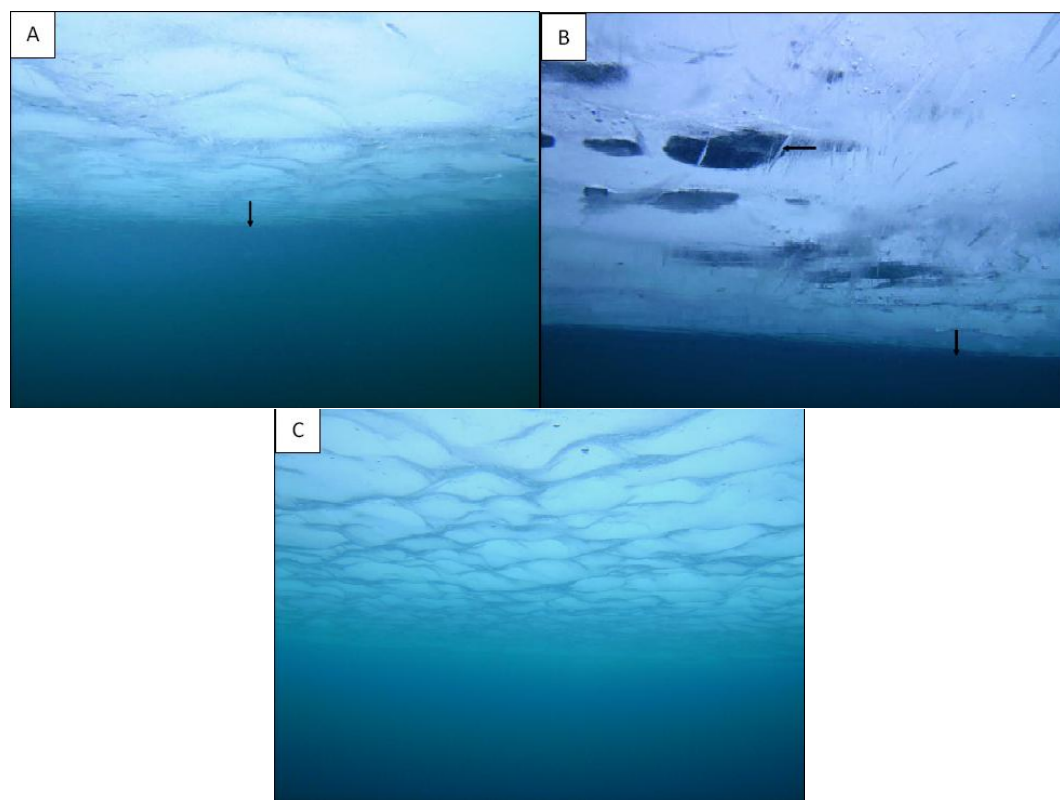


FIGURE 8. Conditions below land fast ice on June 8 and 9, 2008.

A) 50 cm thick melt water layer with small ice platelets (arrow)

B) 50 cm thick melt water layer (arrow) with large ice platelets below soft ice on June 9, and

C) Pock marks below solid ice, created by turbulence [Hop *et al.*, 2011].

The particles on the snow and in the ice can affect the melting of the snow and ice as well as the algae. An example of this is shown in Figure 9 where there were particles on the ice surface in a melt pond in Darnley Bay in June 2008. As shown, the areas where there were particles have melted more than the surrounding areas around them. Algae only use approximately 15 % of the light they absorb with their pigments, ie. their efficiency is only 15 % with the rest of the light is converted to heat [Zeebe *et al.*, 1996]. Therefore, the algae in the ponds and in the ice can affect the ice melt.



FIGURE 9. Particles on the ice in Darnley Bay in June 2008 increasing the amount of melt.

2.3 PRIMARY PRODUCTION IN SEA ICE AND THE UPPER OCEAN IN THE ARCTIC

This section includes a very brief description of the primary production in sea ice and the upper ocean. The amount of radiation and the wavelengths of that radiation that penetrate the snow and/or ice flows are important for the growth and survival of under-ice algae [Winther *et al.*, 2004]. When the oceans are covered with snow and ice, the bottom ice algae makes up the bulk of the primary production and is therefore important for the Arctic ecosystem [Gosselin *et al.*, 1997]. In the marginal ice zone this sub-ice algae can grow in addition to the algae growing in the water between the ice floes. Both of these groups can start producing as soon as there is sufficient light available. Bottom-ice algae will also have an effect on the amount of radiation transmitted into the ocean below. The algae will absorb some of the radiation that would have otherwise been transmitted and if there is enough algae or phytoplankton it can cause shading to some of its own population or for populations lower in the water column [Kirk, 1994].

In late winter the primary production in the Arctic is mostly limited to the bottom of the ice whereas in late spring and summer there are additional habitats for algal growth and accumulation, i.e. surface melt ponds, interior ice brine channels, under-ice melt ponds, melt layer under the ice and at the

subsurface chlorophyll maximum lower in the water column [Mundy, 2007; Martin *et al.*, 2010]. In the Canadian Beaufort Sea, primary production at the bottom of first-year sea ice starts at the end of February [Róžańska *et al.*, 2009]. Prior to this time there is not enough light for growth. Measured minimum values of light need for ice algae to photosynthesize have been observed to be as low as 0.3 to 1 $\mu\text{mol m}^{-2} \text{s}^{-1}$ in the Antarctic [Cota and Sullivan, 1990].

The maximum rate of photosynthesis at saturating light is called the photosynthetic capacity. Beyond that amount of light photosynthesis may decrease due to photoinhibition [Kirk, 1994]. Ice algae and phytoplankton can acclimate to higher or lower light levels through physiological adjustments that can be relatively rapid [Palmer *et al.*, 2011]. The physiological changes that can be made are increasing or decreasing the size of the light-harvesting antennae or the number of photosynthetic units [Sukenic *et al.*, 1987]. During a study in the Canadian Arctic in 2008, phytoplankton in the water column were found to acclimate to the changing light levels over a period of approximately 4 to 10 days [Palmer *et al.*, 2011].

Light is not the only limiting factor influencing primary production. It can be affected by nutrient supply, water temperature, sinking and grazing [Mundy, 2007; Martin *et al.*, 2010]. The sea ice algae must be able to adapt to a wide range of light, temperature and salinity variations [Kirst and Wiencke, 1995]. In

the spring and early summer, a melt layer forms under the sea ice which creates a buoyant layer where phytoplankton grow and accumulate [*Mundy et al.*, 2011]. Though there is high light conditions, phytoplankton can become nutrient limited [*Mundy et al.*, 2011]. In addition, phytoplankton can be an important food source for zooplankton in this habitat [*Hop et al.*, 2011].

CHAPTER 3 METHODS

Field data collection was carried out during two field sessions in the Beaufort Sea area in the western Canadian Arctic in 2008 during International Polar Year Circumpolar Flaw Lead (CFL) system study. The first field session was from March 13 to April 24 (CFL Legs 7A & B) during freezing conditions where ice was still continuing to form and snow or frost flowers were present. The second field session was from May 15 to June 26 (CFL Legs 8B & 9A) which started during melting and freezing conditions and progressed into solely melting conditions and ice decay and break-up.

CTD (Conductivity, Temperature, and Depth) and PAR (Photosynthetically Available Radiation) profiles were done opportunistically with a focus on different snow, ice, and melt pond conditions and different times of the day. This was done to interrogate the affect of different snow and ice regimes on PAR in the water column below the ice and to evaluate the seasonal change of salinity and temperature under ice in the upper ocean. Profiles were also done through melt ponds, thin ice and open water between floes from a boat. Transects or moorings from the ice edge and across floes were also done when possible. Sampling was conducted (when possible) on days with consistent sky conditions throughout the time it would take to complete one site. Therefore, it

was preferable to sample on completely overcast days or completely cloud-free days.

3.1 STUDY AREA

The study area was located in the western Canadian Arctic and the sites included in this thesis were located at the entrance to the Prince of Whales Strait, in the Amundsen Gulf, and in Darnley and Franklin bays located along the southern edge of the Amundsen Gulf. Site locations are illustrated in Figure 10. The ice sites sampled in this area consisted exclusively of first year ice floes with a portion of them being landfast.



FIGURE 10. Study area with site locations shown in red.

3.2 CFL PROJECT

Data were collected during the International Polar Year – Circumpolar Flaw Lead System Study in 2008. The study involved over 350 scientists from 27 countries and took place using the Canadian research icebreaker, the CCGS Amundsen. The Amundsen was over-wintered in the Canadian Arctic to allow for sampling in the spring and early winter which would otherwise be difficult to access at those times due to ice conditions. The project was a whole system study including ice physics, marine ecology, microbiology, marine mammals, contaminants, atmospheric science and related system study fields [Barber *et al.*, 2010].

3.3 FIELD METHODS AND DATASETS – FREEZING PERIOD SAMPLING

In the late winter or early spring, March and April, sampling was carried out on an opportunistic basis. Profiles were completed under as many different snow depths and ice conditions for each ice floe as time and ship schedule permitted.

Sampling at a site collected the following maximum amount of data as time allowed:

- Ice
 - Thickness
- Snow
 - Depth, density, temperature, salinity, and albedo
- Water
 - Freeboard, conductivity, temperature, depth and PAR

Once the site was selected, the incident downwelling and upwelling radiation, above the sea ice in was measured to facilitate future calculations of albedo. These data were measured using the LI-COR LI-190 quantum sensor which measures irradiance for wavelengths in the range of 400 and 700 nm, which are the same wavelengths as PAR. The LI-190 was mounted on a rotating arm connected to a tripod as illustrated in Figure 11. The sensor was located at the end of the arm 1 m from the tripod and the arm was leveled at each location 1 m vertically from the ice or snow surface. A measurement of downwelling radiation was first recorded after readings stabilized, the arm was then rotated 180°, levelled, and upwelling was measured in the same fashion. The two measurements were then repeated. These data were not collected at every site due to time and equipment space constraints.



FIGURE 11. Licor PAR sensor set-up for albedo measurements.

Once incident downwelling and upwelling measurements were completed, the site was moved slightly to a spot with the same surface features but undisturbed. At each site the snow was disturbed as little as possible with footprints leading up to the site far apart and the same path on the side opposite the sun direction was taken each time the hole was accessed. At the sampling location, a 5 cm diameter hole was drilled using an ice auger (Kovacs Enterprises, Inc.). This small auger was used to minimize an unnatural addition of solar radiation through the hole. The snow and ice that accumulated around the hole

as a result of augering was gently removed to minimize artificial obstruction of the solar radiation.

Vertical pressure, conductivity, salinity and temperature profiles were completed in the water column with an Idronaut Ocean Seven 304 CTD multiparameter probe. Its operating temperature range is -30 to 100 °C. Its pressure range is 0 to 100 dBar. Its conductivity range is 0 to 6400 μS and 64 mS cm^{-1} . It has a diameter of 43 mm so fit nicely through the 5 cm Kovacs auger hole. The CTD was attached at the top to a 0.64 cm in diameter nylon rope. Nothing was located below the CTD to ensure minimal water column disruption before the readings were made. The CTD was programmed to take six measurements over the span of each second.

PAR was measured with an Alec Ultra-Miniature Light Intensity Recorder the MDS-MkV/L supplied at the time by Alec Electronics Co. Ltd. (now known as JFE ALEC Co Ltd.). The sensor type is silicone photo diode and its measuring range is 0 to 2000 $\mu\text{mol/m}^2\cdot\text{sec}$ and measures the wavelength range of 390 to 690 μm . It has an accuracy of $\pm 4\%\text{FS}$ and a resolution of 1 $\mu\text{mol/m}^2\cdot\text{sec}$ [*JFE Advantech Co. Ltd. Ocean & River Instruments Division*]. It has a white spherical diffuser cap on the top that allows light to enter to the sensor. This sphere is 240° from one side of where it connects to the casing to the other side. This means that it not only allows light to enter from the top it also allows light to enter from the sides and

from some angles below. Its response curve or photosensitivity can be found in the extended specifications of the instrument manual from Rockland Oceanographic Services Inc. It has an internal memory so did not need any cables attached to it while profiling. While the profiles with the CTD were carried out, one of the Alec PAR sensors was attached to the rope above the CTD using plastic electrical clamps to hold the sensor out and away from the rope. The PAR sensor was programmed to measure at one second intervals which is the fastest interval option for the sensor.

The light sensor was attached to the rope above the CTD at the same location each time as shown in Figure 12. They were both programmed to coordinated universal time (UTC) and their time was updated to the GPS satellite time each day to facilitate time matching.

For each cast the CTD was turned on and then a delay of at least 30 seconds was used to allow the CTD program to initialize and for the CTD to obtain the atmospheric pressure values needed during data processing. The CTD was then immersed in water for another 30 seconds, then the cast was started. The CTD was lowered and stopped for a count of 5 seconds at each metre down to 30 metres then was lowered and stopped for 5 seconds every 5 metres down to 60 metres.

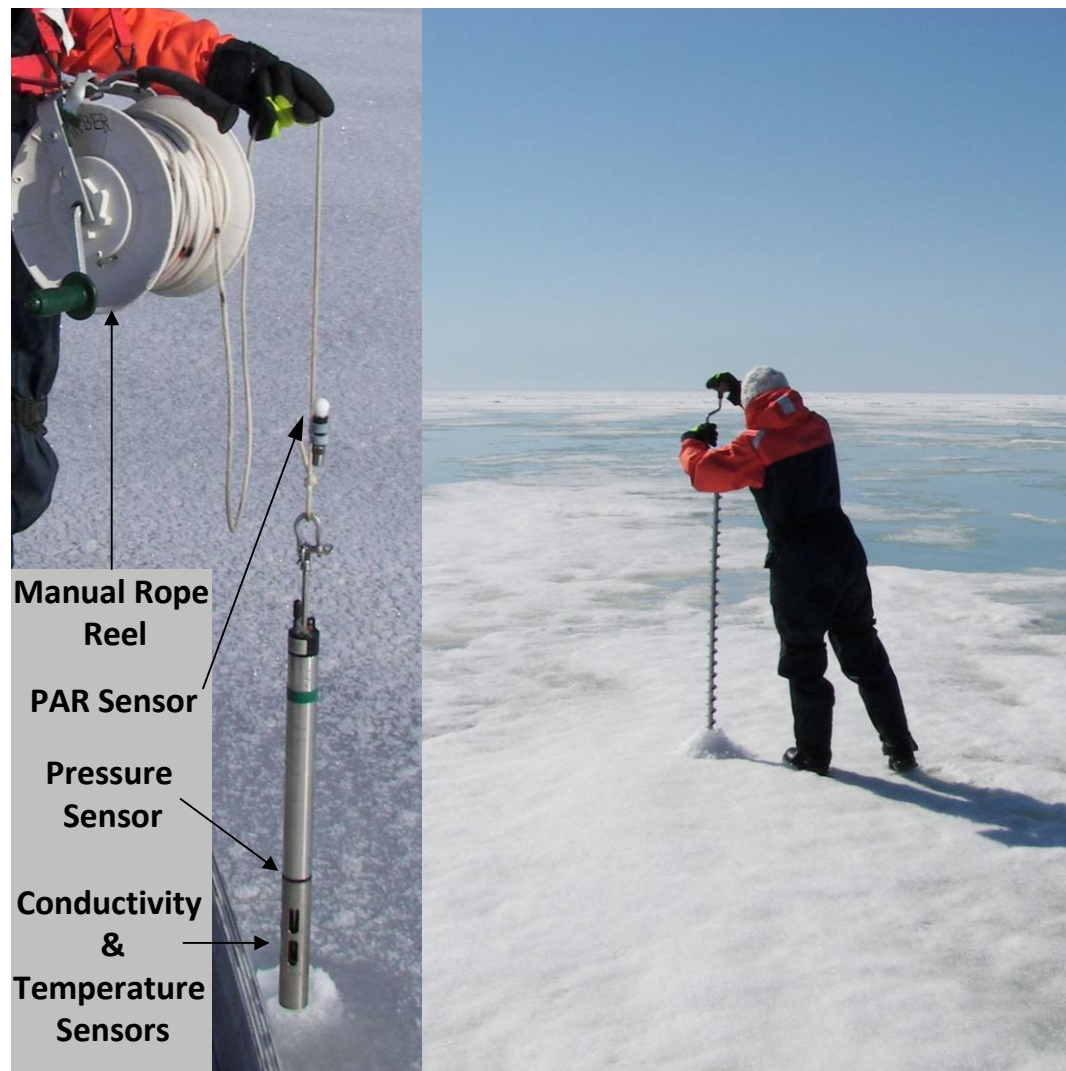


FIGURE 12. PAR and CTD sensors and the profiling instrument configuration illustrated on the left. On the right is a photo showing how the Kovacs auger was used to carefully auger a 5 cm hole through the ice.

Snow depth was measured at every site. Snow pits included a vertical pit photo, and snow grain photos, density and salinity were done for the top, middle and bottom of each snow pack when time permitted.

The location of each sampling site was recorded using a Garmin handheld Global Positioning System (GPS).

The ship was stationed at one site for over 24 hours from April 19 to April 20, 2008. This allowed enough time to install moorings of PAR sensors in the water column below the ice. Four moorings were installed with the first one being through a newly frozen lead, the next being at the edge of the lead and the ice floe, the third one was in the floe, and the last was near a ridge and located furthest from the lead. Sensors were installed at 1.35 m, 3 m, 5 m, 10 m, 20 m, 30 m, 40 m, and 50 m from the ice-air interface.

3.4 FIELD METHODS AND DATASETS – MELTING PERIOD SAMPLING

Light and CTD profiles were done as often as possible during Legs 8B & 9A (May 15 to June 26, 2008). The light sensor was attached to the rope above the CTD at the same height each time as done in Leg 7 (March 13 to April 24, 2008) and as shown in Figure 12. Profiles were done using the same method as described previously.

Sampling at a site collected the following variables as time allowed:

- Ice
 - Thickness
- Snow
 - Depth, density, temperature, salinity, and albedo
- Water
 - Freeboard, conductivity, temperature, depth and PAR

Incident downwelling and upwelling radiation, for wavelengths between 350 and 2200 nm, were sampled with the cosine collector of a FieldSpec FR spectroradiometer (Analytical Spectral Devices, Inc., ASD). Ten measurements of incident downwelling were first recorded then ten measurements of upwelling and this was then repeated an additional time.



FIGURE 13. Setting up the spectroradiometer (ASD) to measure downwelling and upwelling radiation during the melting period.

If there were different surface types at one site then incident downwelling and upwelling irradiance was measured with the ASD at more than one location for each site. For example, if there were ponds and white ice, the measurements were taken at a pond location and a white ice location at that site.

Sampling was done at different times of the day to assess the effect of solar zenith angle. A series of diurnal sampling was completed at two sites 35 m apart. One site was a white patch or bare ice and the other was a melt pond. Vertical

CTD and PAR profiles were done at these two sites in the exact same location consecutively every three hours for approximately thirty hours.

A transect from an ice edge into a floe was completed at Station 405. Profiles were completed at the ice edge and at 3.3, 7.1, 11.3, 15, 18.8, 48, 108 and 168 m from the ice edge across the floe in a straight line.

3.5 DATA PROCESSING

Once the CTD and PAR profiles were completed they were uploaded as text files to the computer. The PAR values were calculated using Equation 3 where A and B are the coefficients for each of the sensors provided by the manufacturer from when they calibrated them and N is the field measured value.

EQUATION 3

$$\text{Light Quantum } (\mu\text{mol m}^{-2}\text{s}^{-1}) = A + B \times N$$

Once each CTD cast was uploaded, the values of each parameter were averaged for every second since the CTD took measurements six times per second. Once this was complete, the PAR dataset and the CTD dataset were joined using the time stamps for each second. The file was then imported into Excel and the depth was calculated from the pressure readings. To correct the pressure value

from the CTD, the average atmospheric pressure value for each particular cast was subtracted from all of the pressure values for that cast. The atmospheric pressure value was the value from when the CTD was held in the air before being placed in the water.

To calculate depth from the pressure readings from the CTD the following formula from the UNESCO Technical Papers in Marine Science No. 44 [Leroy, 1998].

Equation 4

$$x = \sin (\text{latitude}/57.29578)$$

$$x = x \times x$$

$$gr = 9.780318 \times (1.0 + (5.2788 \times 10^{-3} + 2.36 \times 10^{-5} \times x) \times x)$$

$$+ 1.092 \times 10^{-6} \times dp$$

$$mt = \left(((-1.82 \times 10^{-15} \times press + 2.279 \times 10^{-10}) \times press - 2.2512 \times 10^{-5}) \right.$$

$$\left. \times press + 9.72659 \right) \times press$$

$$depth = mt/gr$$

Where:

Press is pressure and is expressed in dbar, dp is depth and is expressed in metres and latitude is expressed in degrees.

Once the CTD and PAR profiles were joined and depth was calculated, the depth of the PAR sensor results were adjusted due to the sensor's location in the instrument set-up and the location of the pressure sensor on the CTD. The distance from the PAR sensor to the pressure sensor on the CTD for each site had been recorded in the field notes when the measurements were taken. That distance was subtracted from the PAR sensor depth values.

3.6 SPATIAL VARIABILITY CALCULATIONS

The Kruskal-Wallis test is a non-parametric test that was used when there were more than two profiles at a site with heterogeneous surface conditions. It was used to determine if at least one of the profiles at each site was significantly different. If at least one of the profiles at the site was significantly different, then the Mann-Whitney U test was used. The Mann-Whitney U test is a non-parametric test that was used to test the variability between two profiles at a time at a site. Non-parametric tests were used because the data for a majority of the PAR profiles, when graphed in frequency histograms, were not normally distributed making a parametric analysis inappropriate [Chapman McGrew Jr. and Monroe, 2000].

One site allowed enough time to complete multiple profiles under a heterogeneous surface cover and an aerial photograph of the area was also

taken. This allowed additional analysis to be completed other than just the Mann-Whitney U test. At FB5 in Franklin Bay on June 16, 2008, five profiles were completed; two in the bare or white ice patches, two in melt ponds and one in a channel connecting ponds. The attenuation coefficient, K_d , was calculated for all profiles at FB5 at each meter. The attenuation coefficient is calculated using the Beer-Lambert Law and is shown in Equation 5.

Equation 5

$$I = I_0 e^{-K_d x}$$

Where:

I_0	-	Intensity at the top of the path ($\mu\text{mol m}^{-2} \text{s}^{-1}$)
I	-	Intensity at the bottom of the path ($\mu\text{mol m}^{-2} \text{s}^{-1}$)
K_d	-	Attenuation coefficient (m^{-1})
x	-	Path length (m)

To be able to calculate one single attenuation coefficient profile for the area, a method was needed to find an average. The goal was to find an effective K_d and the method was taken from *Hanesiak et al.* [2001] and *Perovich et al.*, [2007]. The first step was to select a size for the study area. A study area of 40 m wide by 40 m long was arbitrarily selected which was centred on the most central point between all of the profile locations. The study area is illustrated in Figure 14.



FIGURE 14. An aerial photograph showing the study area of FB5 outlined in red, which was 40 m by 40 m, and the surrounding surface cover.

The study area was able to be measured and selected by using the measurements of the sizes of the melt ponds and bare ice patches measured in the field as indicated in Figure 15. Once the study area was determined, a grid was created over the area. The grid was used to determine the percentage of the area covered by melt ponds, bare ice and channels.

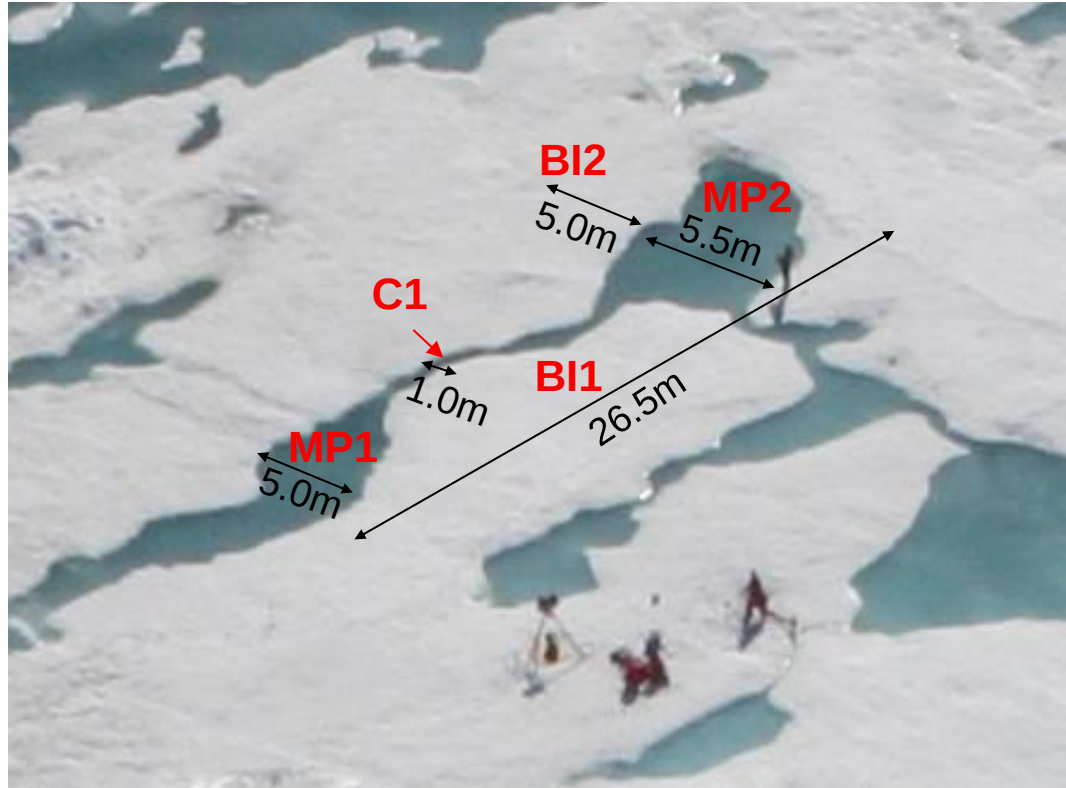


FIGURE 15. The locations of the profiles and the measurements of the melt ponds and bare ice patches at FB5.

Once the percentage of each surface area was determined, the equation that has been used by others to determine the aerial or effective albedo was modified to determine the effective K_d [Hanesiak *et al.*, 2001; Perovich *et al.*, 2007]. The equation used is shown as Equation 6.

EQUATION 6

$$K_{eff} = K_p(z)A_p + K_{bi}(z)A_{bi} + K_c(z)A_c$$

Where:

$K_{eff}(z)$	Effective attenuation (m^{-1})
$K_p(z)$	Average attenuation at depth z under ponds (m^{-1})
$K_{bi}(z)$	Average attenuation at depth z under bare ice (m^{-1})
$K_c(z)$	Average attenuation at depth z under channels (m^{-1})
A_p	Aerial percent cover of ponds
A_{bi}	Aerial percent cover of bare ice
A_c	Aerial percent cover of channels

3.7 HEAT CALCULATIONS

To calculate the heating of the ocean under the ice the radiant heating rate (RHR) equation was used and is shown in Equation 7 [Ohlmann and Siegel, 2000; Ohlmann et al., 2000; Chang and Dickey, 2004].

EQUATION 7

$$RHR_{(z)} = \frac{E_{z1} - E_{z2}}{\rho c_p \Delta z}$$

Where:

E	-	Irradiance at depth ($W m^{-2}$)
$z1$ and $z2$	-	Depths at top and bottom layer respectively (m)
Δz	-	Layer thickness (m)
ρ	-	Density of seawater ($kg m^{-3}$)
c_p	-	Specific heat of seawater ($J g^{-1} ^\circ C^{-1}$)

With the data collected in the field, only the irradiance in the PAR range of 400 to 700 nm was used in this equation because that was all that was available. To account for the total radiant heating rate, the range of 800 to 2500 nm would have to be used which includes visible and near-infrared wavelengths. This equation uses irradiance values in W m^{-2} although the values obtained from the PAR sensors during field sampling are in $\mu\text{mol m}^{-2} \text{s}^{-1}$. To convert the values the PAR spectral data from the HyperOCR profiles completed nearby at the dive site under similar conditions were used [Ehn *et al.*, 2011]. The method of conversion followed that described in [Biggs, 2010]. The radiant heating rate was calculated for each profile at the FB5 site at every metre from 2 to 40 m.

3.8 PRIMARY PRODUCTION CALCULATIONS

To calculate the potential light limited primary production under the ice in the ocean at FB5 Equation 8 was used [Platt *et al.*, 1980].

EQUATION 8

$$P^B = P_s^B (1 - e^{-a}) e^{-b}$$

$$a = \alpha I / P_s^B$$

$$b = \beta I / P_s^B$$

$$P_m = P_s \left(\frac{\alpha}{\alpha + \beta} \right) \left(\frac{\beta}{\alpha + \beta} \right)^{\frac{\beta}{\alpha}}$$

$$\text{Potential Light Limited Production} = P^B [\text{Chla}]$$

Where P^B is the photosynthetic rate at irradiance I , P_s^B is the light-saturated photosynthetic rate in the absence of photoinhibition, α is the initial slope of the primary-irradiance curve, β is the photoinhibition parameter [Arrigo *et al.*, 2010]. The concentration of chlorophyll a was obtained from the Wetlabs fluorometer obtained from the HyperOCR profile at the nearby dive program on the same day under similar ice conditions [Ehn *et al.*, 2011].

The α , β , and P_s^B values were measured at 2 m under the sea ice and at the depth of the maximum chlorophyll a concentration during the same field season in the similar study area [Palmer *et al.*, 2011]. Once the photosynthetic rate was determined it was multiplied by the closest chlorophyll a profile at the same station and the same day to obtain the potential light limited primary production under the white or bare ice patches and the ponds at FB5.

CHAPTER 4 RESULTS AND DISCUSSION

The objective of this thesis is to illustrate using in situ field data what PAR, salinity, and temperature look like in the water column under sea ice from freezing to melting. Furthermore, this thesis will show what PAR looks like diurnally in the water column under sea ice during freezing in April and melting in June.

4.1 THE PHYSICAL ENVIRONMENT DURING FREEZING

The physical sea ice environment encountered during March and April involved many different surface features including open water, leads, newly formed thin ice, ice covered with new frost flowers, moderately thick ice covered in older, decomposed frost flowers, thicker ice covered in thin to thick snow cover and ridged ice cover.

Smooth first-year ice thicknesses (excluding ridges) at sample locations ranged from 0.5 cm to 160 cm. The average ice thickness encountered was 104 cm. The average PAR albedo measured at sites with frost flowers with no snow was 0.76 and at snow covered sites 0.93.

TABLE 2. Site condition information for the sites illustrated in the ice and snow profiles during the freezing period in April.

Station	Date	Ice Thickness (cm)	Snow Depth (cm)	Snow	Sky Conditions
D36_3	2008/04/06	160	7.5	Windpacked small drift	Sunny, clear sky
D38_6	2008/04/11	126	30	New fluffy layer of snow on top, hard icy dense 1 cm layer at bottom	Sunny, Light clouds
D38_7	2008/04/11	110	18	New fluffy layer of snow on top, hard icy dense 1 cm layer at bottom	Sunny, very few light clouds close to horizon
D41_9	2008/04/20	130	10	Smooth, hard, windpacked snow	Mostly cloudy, variable clouds

Profiles were completed during April 2008 through first year pack ice with snow cover in the Canadian Arctic while it was still during the freezing period. The snow cover at the sites included in the profiles ranged from 7.5 cm to 30 cm and the ice thickness ranged from 110 cm to 160 cm.

The PAR values near the ice-water interface, ranged between approximately 6 to 36 $\mu\text{mol m}^{-2} \text{s}^{-1}$ (Figure 16A). D36_3 and D41_9 had the lowest PAR values under the ice between approximately 6 to 14 $\mu\text{mol m}^{-2} \text{s}^{-1}$. 41_9 had also reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at the shallowest depth at approximately 16 m. D38_6 and D38_7 had the highest PAR values at the ice bottom between 28 to 36 $\mu\text{mol m}^{-2} \text{s}^{-1}$. Both profiles also reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at approximately 26 m.

The temperature values at the top of the profiles at the bottom of the ice ranged between approximately -1.78°C at D38_6 to -1.74°C at D38_7 (Figure 16B). Half of the freezing period profiles were cooler at the bottom of the ice compared to the rest of the profile. All of the profiles were around the same temperature at -1.74°C the majority of the rest of the water column down to approximately 46 m. After that depth, D38_7 and D41_9 slightly warm up to -1.72°C. D38_7 warms to -1.69 at 52 m then cools back down to 1.72°C at 53 m and stays at that temperature to the end of the profile at 59 m. The water at D36_3 warmed up after a depth of approximately 49 m and warmed to -1.53°C at the end of the profile at 54 m. D38_6 had a similar shape and warmed to -1.63°C at 60 m. D36_3 had the warmest temperature of all of the freezing period profiles which occurred, as mentioned, at 54 m. These profiles demonstrate the cooling of the water at the ice bottom consisted of a very thin cooler layer where the ice was growing. Then below that is a well mixed layer of constant temperature that

extends approximately 50 m down to the thermocline where the water column then starts to warm up rapidly.

The salinity profiles at the ice and snow freezing sites had slightly more variance than the temperature profiles. The salinity values at the top of the profiles near the ice bottom ranged between approximately 31.72 to 31.80 (Figure 16C). All of the profiles except D41_9 had a salinity of 31.80 at the ice bottom which then slowly got more saline with depth starting at approximately 25 m. The highest salinity reached was 31.96 at 54 m at D36_3. D41_9 increased very little from ice bottom to the end of the profile at 53 m except for four protrusions at 31.5, 36.5, 43.5 and 50.0 m where it dipped briefly to a lower salinity ranging from 31.6 to 31.7. These protrusions may be an artefact of instrumentation and pauses when the CTD cast was completed. The D36_3 salinity profile matches closely to the majority of the temperature profiles from the same location. It had a mixed layer down to approximately 50 m and then the salinity started to rapidly increase at the thermocline. D36_3 had the shallowest thermocline or the thinnest mixed layer out of all of the snow and ice freezing period locations. Some of the other freezing period profiles were problematic because of the instrumentation freezing between casts or getting clogged with ice slush on the way through the hole. This is something to consider when completing profiles at this temperature and to leave enough time in the water lower in the ice hole for the instrument to equilibrate thermally.

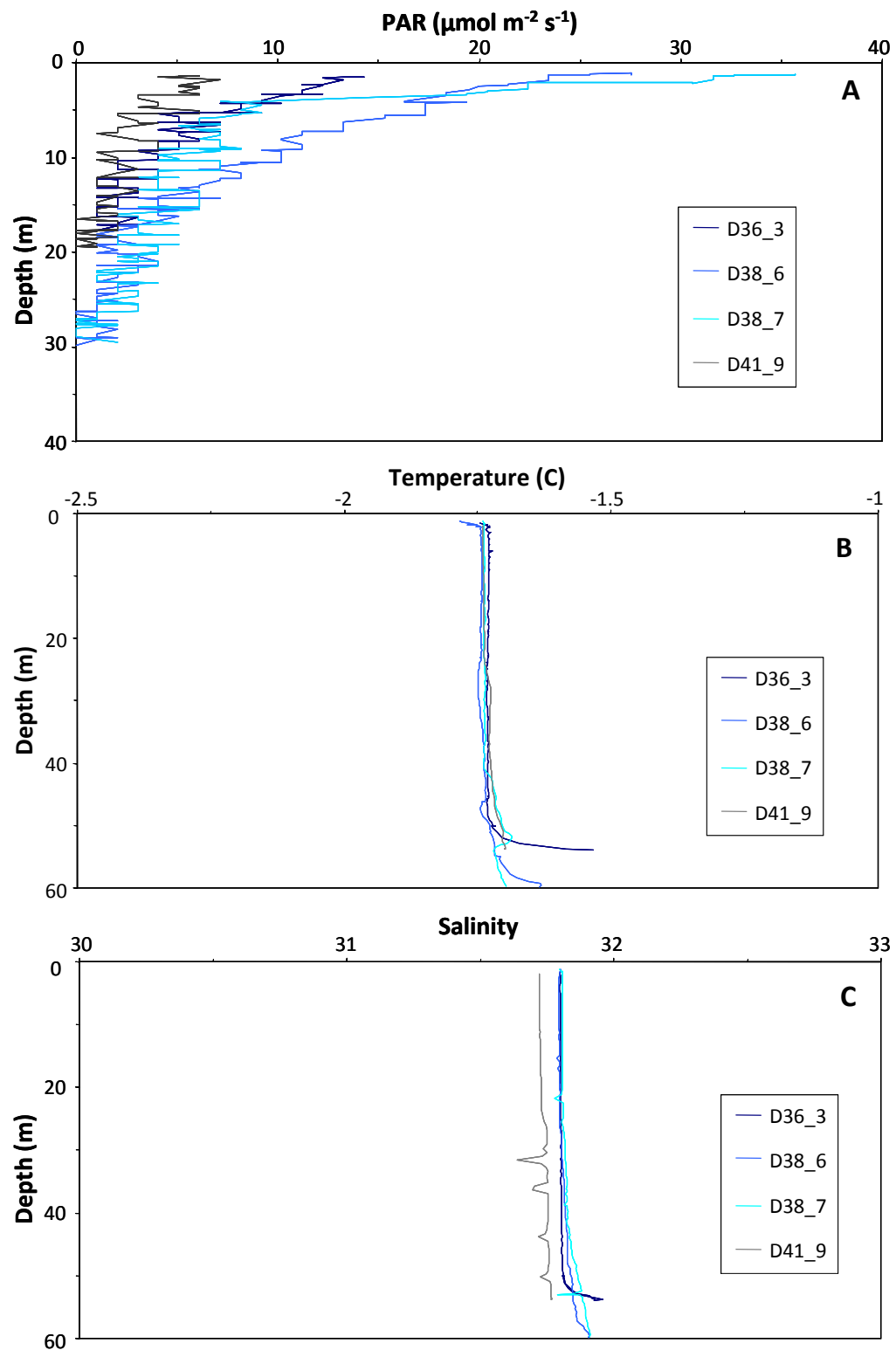


FIGURE 16. PAR (A), temperature (B) and salinity (C) profiles completed during the freezing period through ice and snow with the Alec PAR sensor and CTD.

TABLE 3. Site condition information for the sites illustrated in the leads profiles during the freezing period.

Station	Date	Ice Thickness (cm)	Snow Depth (cm)	Sky Condition	Site Information
D36_4	2008/04/06	lead	-	sunny, clear sky	In centre of lead, 220 m long by 30 m across, no ice
D36_10	2008/04/08	lead	-	sunny, clear sky	Middle of 50 m across lead, no ice
D38_4	2008/04/10	1	frost flowers	sunny, clear sky	Nilas ice covered with frost flowers
D38_9	2008/04/11	1.5	frost flowers	sunny, very light clouds	Nilas ice covered with frost flowers, lead is 1 m across
D41_10	2008/04/20	very thin < 1 cm	-	overcast, light clouds	In newly formed lead, some grease ice, lead is 7-8 m across

Profiles were completed during April 2008 through leads in the Canadian Arctic while it was still during the freezing period. Two of the leads had open water with no ice, one had very thin ice and two had a thin layer of ice covered with frost flowers. The largest of all the leads was 50 m across.

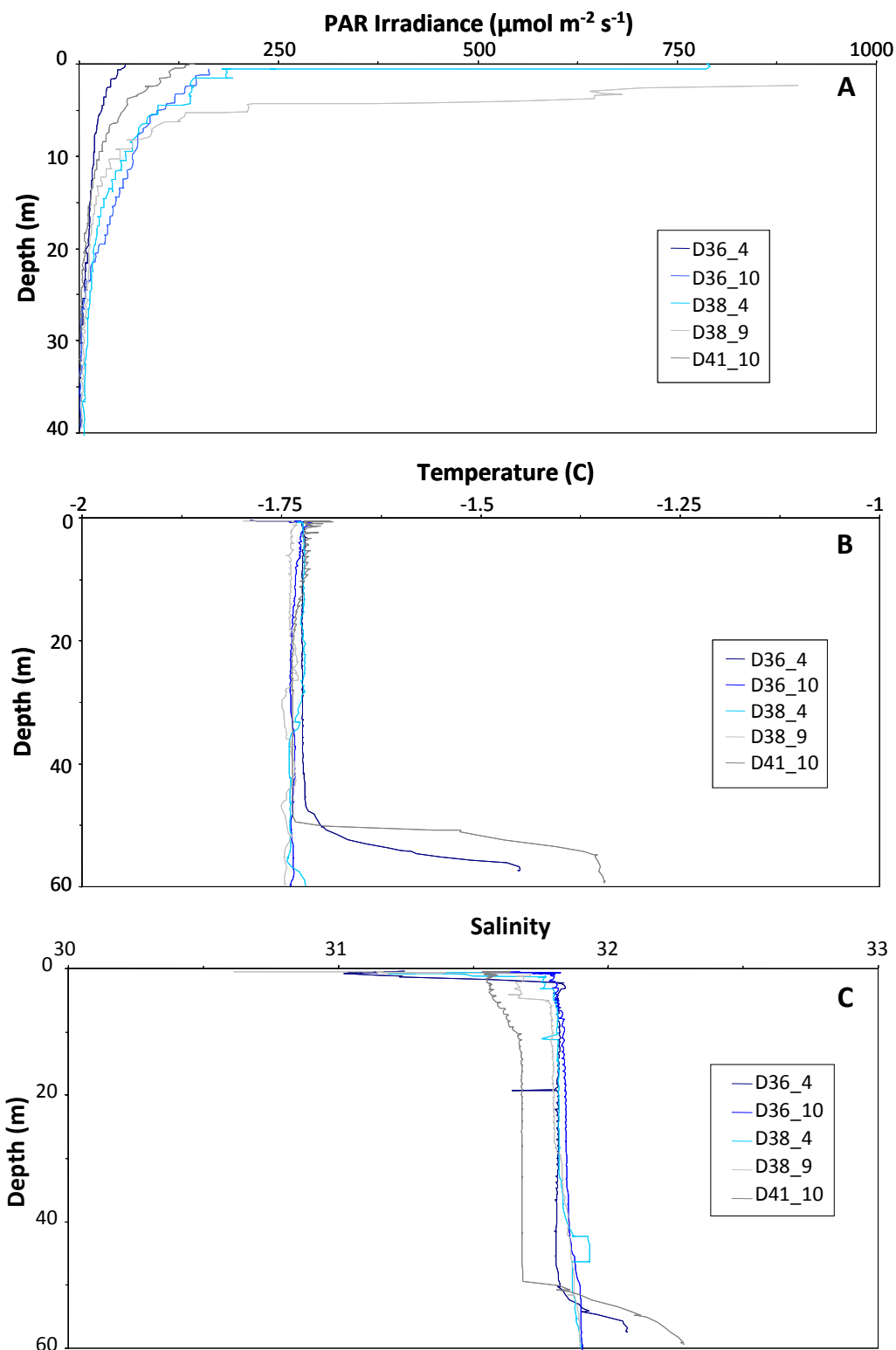


FIGURE 17. PAR (A), temperature (B) and salinity (B) profiles completed during the freezing period through leads with the Alec PAR sensor and CTD.

The PAR values at the top of the profiles ranged between approximately 60 to 900 $\mu\text{mol m}^{-2} \text{s}^{-1}$ (Figure 17A). D36_4, D36_10 and D41_10 had the lowest PAR values at the surface or under the ice between approximately 60 to 160 $\mu\text{mol m}^{-2} \text{s}^{-1}$. These three profiles also had their lowest PAR values at approximately 30 m. D38_4 and D38_9 had PAR values near the surface between 790 to 900 $\mu\text{mol m}^{-2} \text{s}^{-1}$. The minimum values at both locations occurred at approximately 40 m.

The PAR transmission at the lead sites is much higher than at the snow and ice sites as would be expected. The site with the lowest PAR value at the top of the profile in the leads is higher than the site with the highest PAR value at the top of the profile in the snow and ice (Figures 16A and 17A). This is because the snow has such a high albedo that it reflects most of the incoming light and the transmission of light through snow is very low [Perovich, 2005]. The leads have a very low albedo since some of them have no ice and the other ones that are included have only a small amount of ice so majority of the incoming light is transmitted since the albedo would be approximately 0.06 [Perovich, 1996].

Stations D36_10, D38_4 and D38_9 temperature profiles were approximately -1.73°C from the surface down to the end of the profiles at 60 m (Figure 17B). D41_10 is colder at the surface than all of the other profiles with a temperature of -1.80°C then follows the other profiles temperatures down to 49 m. After

49 m it gets to the thermocline and starts to warm up to a maximum of -1.36°C at 55 m. After 55 m it stays around that temperature down to the end of the profiles at 60 m. D36_4 has a similar temperature as all of the other profiles down to 48 m. After this depth it gradually warms up reaching a maximum of -1.45°C at 57 m.

The salinity values at the top of the profiles ranged between approximately 30.6 to 31.7 (Figure 17C). The lowest values at the top of the water column were from D38_9 with the highest being from D36_10. D41_10 and D36_4 have similar salinity profile shapes between 50 and 60 m as their corresponding temperature profiles. D41_10 had lower salinity values than all of the other lead sites at around 31.55 to 31.68 from near the surface to approximately 49 m then sharply increased to 32.28 at the bottom of the profile at 59 m. D36_4 had similar salinity values throughout the water column as the other sites around 31.81 then at approximately 50 m increased to 32.07 at the bottom of the profile at 57 m.

Station D41 during the freezing period had a lower salinity from near the surface to approximately 50 m than any of the other freezing period snow and ice or lead locations. D41 also had the most rapid increase in salinity and temperature between 50 and 60 m showing the end of the surface mixed layer. This is most likely due to being a different water mass since it is located in the centre of the

Amundsen Gulf. Stations D36 and D38 were located to the southwest of Banks Island closer to the Beaufort Sea.

4.2 THE PHYSICAL ENVIRONMENT DURING THE MELTING PERIOD

The physical environment encountered during the melting period was metamorphosed snow, melted and refrozen snow, melt ponds, bare ice and a low salinity layer of melt water underneath the bottom of the ice. All of the sites sampled during this time period had first year ice with most of the sites being landfast first year ice in Franklin or Darnley bays on the southern edge of the Beaufort Sea.

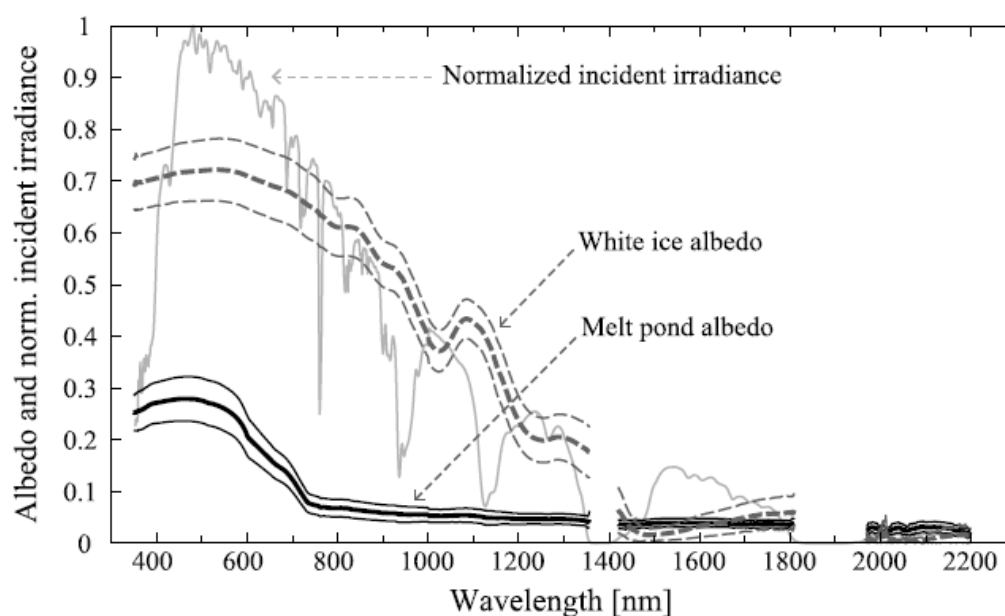


FIGURE 18. Spectral albedos completed with the field methods described earlier at melt ponds and white ice sites during the melting period in similar locations as included in Section 4.2. The thin lines are ± 1 standard deviations [Ehn *et al.*, 2011].

TABLE 4. Site condition information for the sites illustrated in the bare ice profiles during the melting period.

Station	Date	Ice Thickness (cm)	Freeboard (cm)	Snow Depth/ Active Layer (cm)	Skies	Site Information
F6_D1f	2008/06/02	178	12	3	Sunny, very light wispy clouds	White patch between melt ponds, white ice is 10 m across
F6_8	2008/06/03	159	12	2	Sunny, very light wispy clouds	2 surface types each with 50% cover, one is smaller crystals surface white ice layer the other is a harder, bumpy shiny white ice
F7_7	2008/06/09	148	9	3	Sunny, clear sky	A few melted patches in white ice between ponds, white ice is 15 m across
F7_8	2008/06/09	130	15	10	Sunny, clear sky	White patch between melt ponds, white ice is 7 m across
FB5_4	2008/06/16	145	15	10	Sunny, very light wispy clouds near horizon	White patch between melt ponds, white ice is 10 m across
FB5_5	2008/06/16	150	18	10	Sunny, very few light wispy clouds near horizon	White patch between melt ponds, white ice is 15 m across

Profiles were completed during June 2008 through land fast bare ice, or also known as white ice, in the Canadian Arctic in Franklin and Darnley Bays during the melting period. Stations F6 and F7 were at the entrance to Darnley Bay and FB5 was in Franklin Bay. The ice thicknesses at the sites included in the profiles ranged from 130 cm to 178 cm. The freeboard ranged from 9 cm to 18 cm.

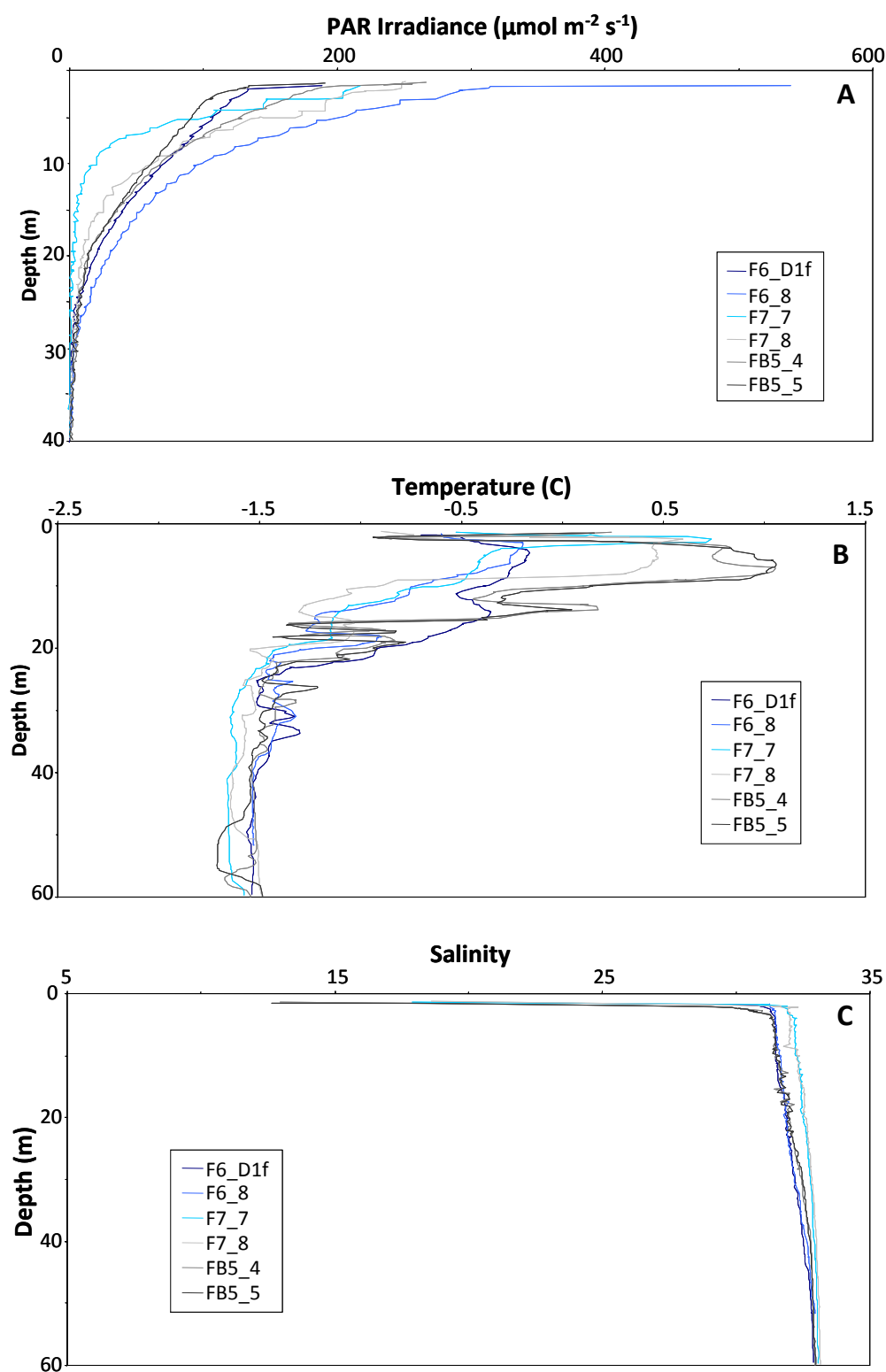


FIGURE 19. PAR (A), temperature (B), and salinity (C) profiles completed through bare ice during the melting period with the Alec PAR sensor and CTD.

The PAR values at the top of the profiles at the ice bottom ranged between approximately 190 to 540 $\mu\text{mol m}^{-2} \text{s}^{-1}$ (Figure 19A). FB5_5 and F6_D1f had the lowest PAR values at the ice bottom between approximately at 190 $\mu\text{mol m}^{-2} \text{s}^{-1}$. F7_7 decreased the steepest and reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at the shallowest depth at 22 m. F6_8 had the highest PAR values at the ice bottom at 540 $\mu\text{mol m}^{-2} \text{s}^{-1}$ and had higher PAR values than all other locations until 27.5 m when all of the profiles joined at 5 $\mu\text{mol m}^{-2} \text{s}^{-1}$ except F7_7 which had already reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at 22 m. The PAR albedo, as shown in Figure 18, was 0.70 ± 0.06 for the bare ice sites [Ehn *et al.*, 2011].

The temperature profiles during melting (Figure 19B) are more complex and variable than those during freezing under the lead (Figure 17B) and snow and ice sites (Figure 16B). Sites in Darnley Bay had the coolest temperatures near the ice bottom which ranged from -0.53°C at F7_7 on June 9 to -0.90°C at F7_8 on June 9. The Sites at F6 in Darnley Bay ranged between the F7 values at the ice bottom. Sites FB5, which were in Franklin Bay on June 16, had the warmest temperatures at the ice bottom with temperatures ranging from 0.16°C to 0.24°C at the ice bottom at FB5_5 and FB5_4 respectively. Lower in the water column the water column there is mixing and it is turbulent and therefore has wavy lines, going from cooler to warmer and back, down to approximately 25 to 30 m at the FB5 sites in Franklin Bay. The maximum temperature out of all of the profiles occurred at FB5_5 and was 1.06°C near 5.5 m from the ice bottom. The

coolest temperature out of all of the profiles also occurred at FB5_5 and was -1.71°C near 53.6 m from the ice bottom. The sites F6_8, F7_7 and F7_8 in Darnley Bay had cooler temperatures than the Franklin Bay throughout most of the water column. This variability will be discussed further in the melt pond temperature discussion later in the thesis.

The salinity values at the top of the profiles near the ice bottom ranged between approximately 12.55 to 18.61 (Figure 19C). The lowest values at the top of the water column were from FB5_5 with the highest being from F7_8. The thickest melt layer under the ice was approximately 0.8 m thick. All of the profiles except F6_8 had a melt layer directly below the ice bottom. The melt layers ranged somewhat and after a metre to two metres the water column salinities remained relatively constant. The profiles from station FB5 had a melt layer that began near 12.55 at the ice bottom and increased steeply to 29.81 approximately 0.8 m below the ice bottom then increased more gradually to 31.27 at 2.0 m below the ice bottom. The sites at station F7 did not have the intermediate salinity increase. They had a melt layer that steeply dropped from around 18.25 at the ice bottom to 31.83 in approximately 0.8 m. The F7 sites had the highest salinities throughout the profiles and the highest salinity measured was 33.14 at the bottom of the profile at 60 m at F7_8.

TABLE 5. Site condition information for the sites illustrated in the melt pond profiles during the melting period.

Station	Date	Ice Thickness (cm)	Pond Depth (cm)	Pond Ice (cm)	Sky Condition	Site Information
F6_2	2008/06/02	150	5-7	1	Sunny, very light wispy clouds	Pond is 10 m across
F6_7	2008/06/03	145	4	0.5	Sunny, very light wispy clouds	Pond is 6 m long by 4 m across
F7_2	2008/06/07	105	10	1	Sunny, very light clouds covering entire sky	Pond is 5 m across
F7_6	2008/06/09	94	28	-	Sunny, clear sky	Huge pond, over 50 m across
FB5_3	2008/06/16	115	12	-	Sunny, clear sky	Pond is 5 m across
FB5_7	2008/06/16	114	20	-	Sunny, clear sky	Pond is 5.5 m across

Profiles were completed during June 2008 through melt ponds in the Canadian Arctic in Franklin and Darnley Bays during the melting period. Stations F6 and F7 were at the entrance to Darnley Bay and FB5 and FB7 are in Franklin Bay. The melt ponds ranged from 4 to 28 cm deep (Table 5). In early June melt ponds had a thin layer of ice on the surface which ranged from 0.5 to 1.0 cm thick. The ponds ranged in size from 4 to over 50 m across with ice thicknesses under the ponds ranging from 94 to 150 cm thick.

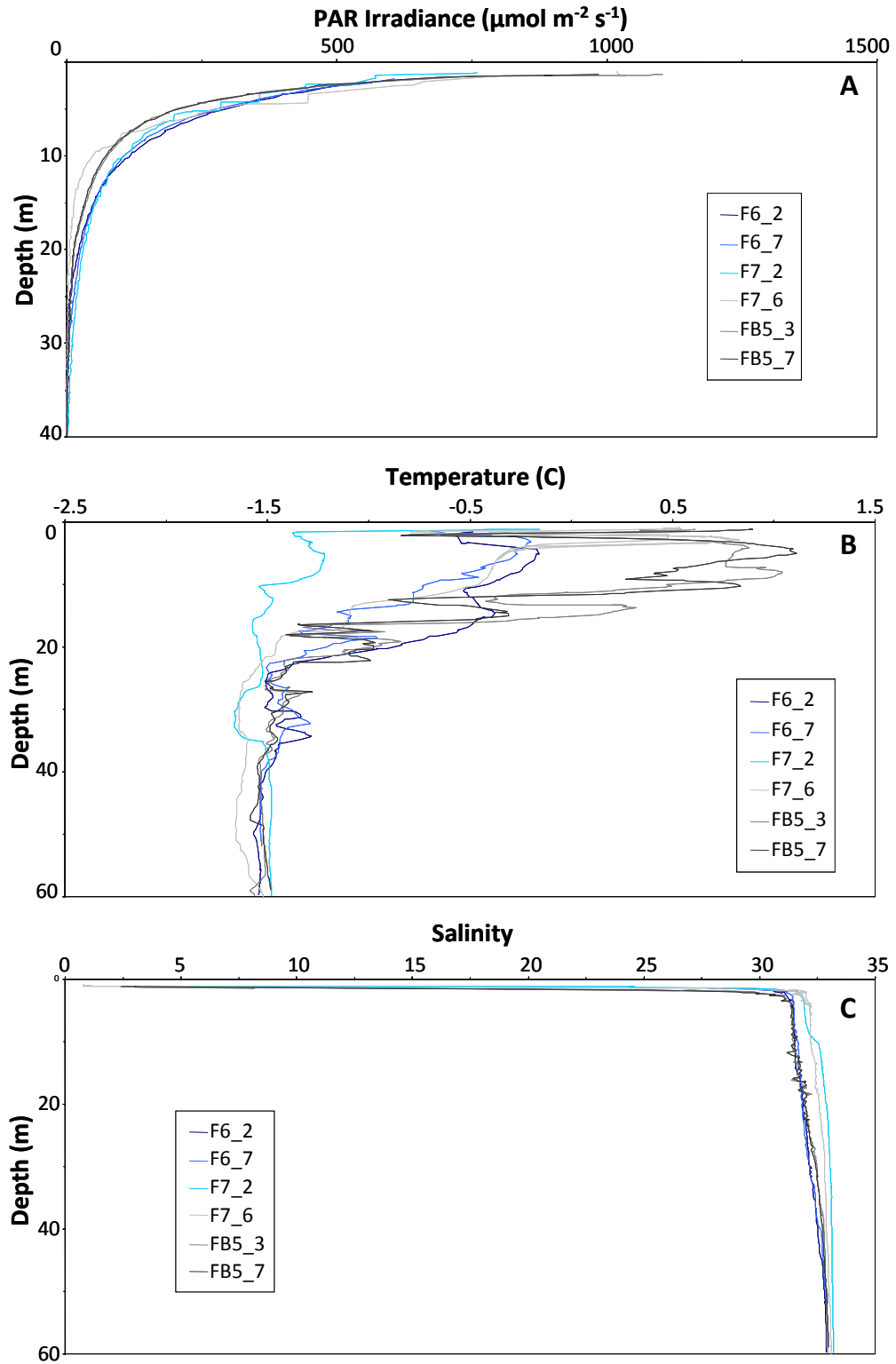


FIGURE 20. PAR (A), temperature (B), and salinity (C) profiles completed through melt ponds during the melting period with the Alec PAR sensor and CTD.

The PAR values at the top of the profiles at the ice bottom ranged between approximately 760 to 1100 $\mu\text{mol m}^{-2} \text{s}^{-1}$ (Figure 19A). F7_2 had the lowest PAR value under the ice and FB5_3 had the highest. F7_6 had the second highest PAR values at the ice bottom at 1019 $\mu\text{mol m}^{-2} \text{s}^{-1}$ but decreases to have a lower PAR value than the other profiles of 87 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at 9 m deep. After 9 m it continues to have the lowest PAR values until it reaches 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at 29.8 m. F7_6 reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at the shallowest depth of all the profiles at 29.8 m and F7_2 reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ at the deepest point of 45.5 m. The PAR albedo, as shown in Figure 18, was 0.22 ± 0.04 for the melt ponds sites [Ehn *et al.*, 2011].

Sites in Darnley Bay had the coolest temperatures near the ice bottom under ponds as well as the bare ice, which ranged under ponds from -0.69°C at F6_7 on June 3 to -0.54°C at F7_6 on June 9 (Figure 19B). Sites FB5, which were in Franklin Bay on June 16, had the warmest temperatures at the ice bottom of ponds as well as the bare ice with temperatures under ponds ranging from 0.61°C to 0.88°C at the ice bottom at FB5_3 and FB5_7, respectively. Lower in the water column the water column is mixing, which is the same as the bare ice sites, and is turbulent and therefore has wavy lines going from cooler to warmer and back, down to approximately 25 to 30 m at the FB5 sites in Franklin Bay. The maximum temperature out of all of the profiles occurred at FB5_7 and was 1.12°C near 4.97 m from the ice bottom. The coolest temperature out of all of the profiles occurred at F7_2 and was -1.66°C near 32.0 m from the ice bottom.

The sites F6_2 (June 2), F6_7 (June 3), F7_2 (June 7), and F7_6 (June 9) in Darnley Bay had cooler temperatures than the Franklin Bay down to approximately 35 m in which they all have around the same temperature except F7_6 which stays colder. F6_8 (June 3), F7_7 (June 9), and F7_8 (June 9), that were shown in the bare ice temperature profiles were also cooler. This variability in Darnley Bay can be explained by an upwelling event that was caused by strong winds blowing at the ice edge. These strong winds caused upwelling of deeper, cooler, more saline, nutrient rich water to be mixed from deeper down in the water column upwards towards the surface [Mundy *et al.*, 2009]. F6_2 and F6_D1f have the warmest of the Darnley Bay profiles which were on June 2 prior to the upwelling event. F6_7 and F6_8 were on June 3 and are starting to show a cooling near the start of the upwelling. The coolest and most saline profile was from F7_2 which occurred at the peak of the upwelling in Darnley Bay. It appears that the upwelling brought the cool water all the way up to being very near the ice bottom. Only a thin layer of low salinity, warmer melt water remained under the ice on this date. This may just be some of the melt water in the under-ice dome that forms under the ponded ice. F7_6, F7_7, and F7_8 were on June 9 after the upwelling and show that the colder water is starting to sink and they are around the same temperatures as the profiles on June 3, except for F7_8 which was warmer in the upper 10 m.

The salinity values from the top of the profiles down into the water column ranged more greatly under the ponds (Figures 20 and 19C) than under the bare ice or white patches (Figure 19C). Under the ice bottom they ranged between approximately 0.78 to 29.71. The lowest values at the top of the water column were from F7_6, FB5_3 and FB5_7 with the highest being from F6_2 and F6_7. The thickest melt layer under the ice was approximately 0.8 m thick. All of the profiles except F6_2 and F6_7 had a melt layer directly below the ice bottom. The melt layers ranged somewhat and after a metre to two metres the water column salinities remained fairly constant. The profiles from station FB5 had a melt layer that began with a salinity of 2.46 at the ice bottom that increased steeply to 28.98 approximately 0.8 m below the ice bottom then increased more gradually to 31.00 at 1.6 m below the ice bottom. The sites at station F7 did not have the intermediate salinity increase. They had a melt layer that steeply dropped from a salinity of around 0.78 and 16.97 at the ice bottom to 31.57 in approximately 0.6 m. Below the melt layer, the F7 sites had the highest salinities throughout the profiles and the highest salinity measured was 33.23 at the bottom of the profile at 60 m at F7_2. The high salinities at F7 can be explained by the upwelling event as discussed previously and in a study in the same time and area [Mundy *et al.*, 2009].

4.3 OBSERVED TEMPORAL VARIABILITY OF PAR TRANSMISSION

4.3.1 FREEZING PERIOD

Moorings, designed to measure temporal variability of PAR in the water column, were installed in a transect across an ice floe from a newly frozen lead to a ridge on April 17 to 22, 2008 at station D41 near the centre of the Amundsen Gulf (Figure 21). The ice thickness ranged from 11 cm in the lead to 200 cm at the ridge. There was also approximately 70 cm of snow on the ridge though the ice thickness and snow depth was hard to measure because of the variability of the ridge. There was a 1 cm thick layer of frost flowers on the lead. The second mooring was located at the edge of the lead adjacent to the ice floe. It had 11 cm of ice and 4 cm of snow that had accumulated because the vertical difference between the lead ice and the floe ice caused snow to accumulate. There was a 4 to 5 cm crack in places where the lead ice met the floe ice. The third mooring was located approximately half way between the lead and the ridge and had 140 cm of ice and 10 cm of snow.

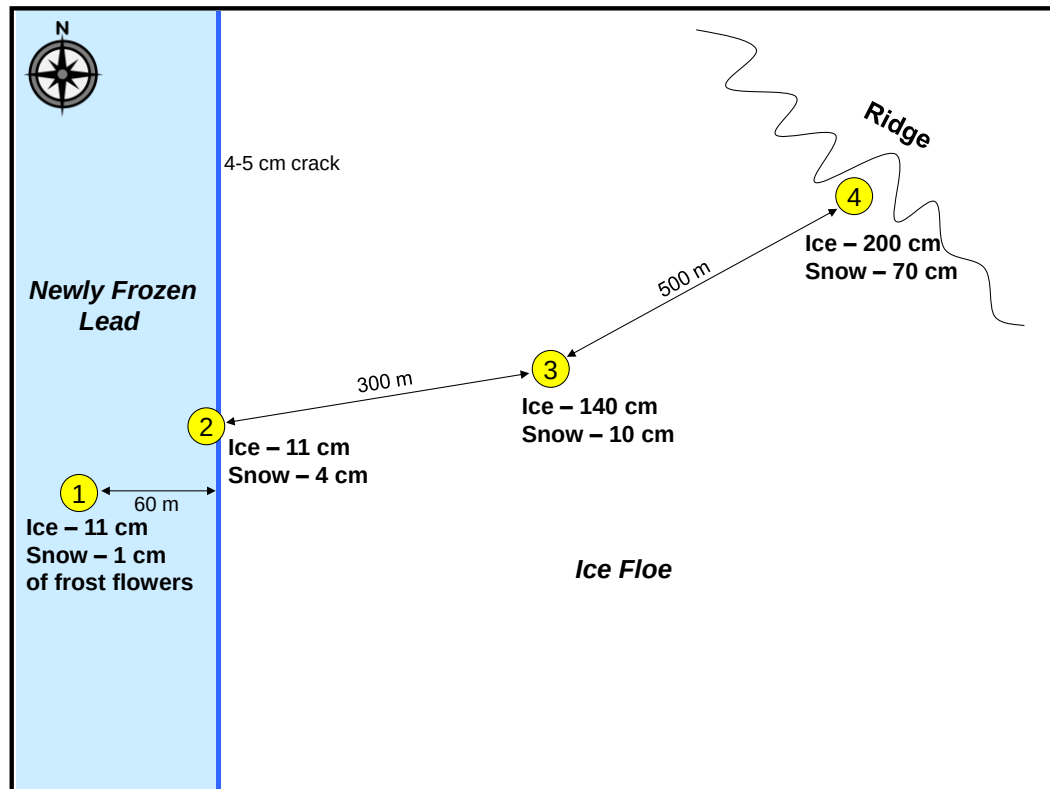


FIGURE 21. Site diagram showing the locations and physical properties for the freezing period moorings in the lead, lead edge, floe centre and near the ridge. Note: Diagram is not to scale.

The data from the 24 hour period starting April 17 at 08:00 UTC to 08:00 UTC on April 18 were used because that time period had completely clear skies. The other days had some cloud cover and snow fall so the light profiles were quite variable over short periods of time making it difficult to use statistics to compare the light at the different locations. The temperature and salinity in the area of these locations at Station D41 is shown in the D41_9 and D41_10 profiles in Figures 16 and 17.

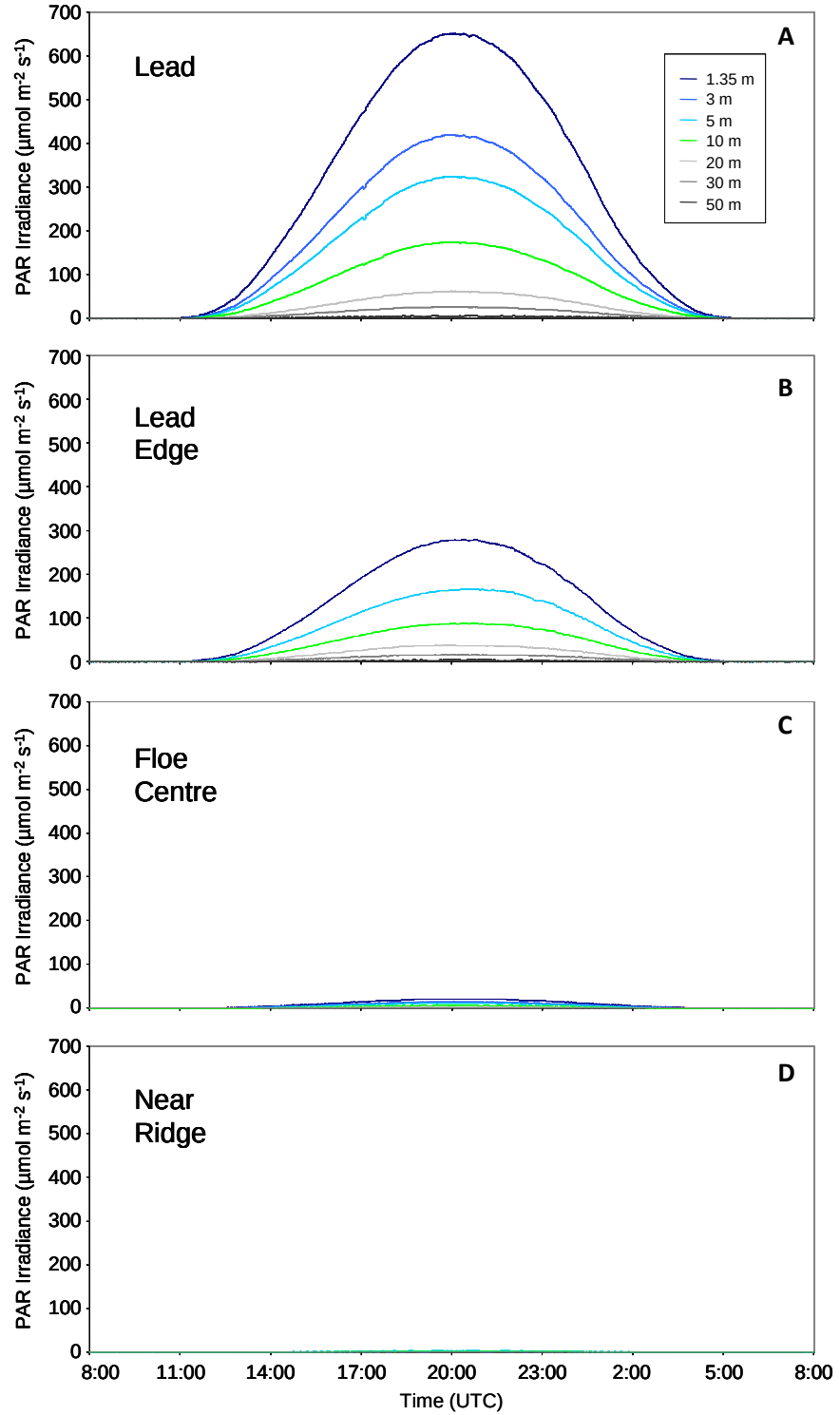


FIGURE 22. Diurnal variations of PAR measured at seven depths under the sea ice. Measurement is taken from the ice-air interface. Moorings were installed at Station D41 from April 17 to 18, 2008.

The site near the ridge had the least amount of light as expected with very little to no light at any depth at any time of day (Figure 22D). The maximum amount of PAR was $2.1 \mu\text{mol m}^{-2} \text{s}^{-1}$ at 19:30 UTC or 12:30 local time at approximately 3 m from the ice surface. This is due to the deep snow (70 cm) as well as the thickness of the ice (200 cm). The snow has such a high albedo that it reflects most of the incoming light and the transmission of light through snow is very low [Perovich, 2005]. PAR decreases to a constant $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ at 3 m from 24:00 UTC (17:00 local time) to 14:30 UTC (09:30 local time). Though, the amount of light the rest of the day between 09:30 and 17:00 local time is very minimal ranging between 0 and $1 \mu\text{mol m}^{-2} \text{s}^{-1}$ (Figure 22D).

The site in the ice floe, halfway between the lead and the ridge, had slightly more light than the ridge site but still did not have a lot of light (Figure 22C). The maximum amount of PAR was $21.1 \mu\text{mol m}^{-2} \text{s}^{-1}$ which was constant between 19:20 to 21:00 UTC or 12:20 to 14:00 local time at the sensor located at the most shallow location which was at the ice bottom. At the same time, at 3, 5, 10, 20, and 30 m, PAR was 14.2, 10.4, 6.2, 1.0 and $0.0 \mu\text{mol m}^{-2} \text{s}^{-1}$ respectively. PAR decreases to $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the shallowest sensor from 03:30 UTC (21:30 local time) to 13:00 UTC (06:00 local time). This location may be affected by ice algae growing in the bottom of the ice. Algae and phytoplankton were not able to be measured due to time and equipment space constraints.

At the lead edge (Figure 22B), there was more light than the ice floe site, but less than the centre of the lead. The maximum amount of PAR was $278.5 \mu\text{mol m}^{-2} \text{s}^{-1}$ which was constant between 20:43 to 20:51 UTC or 13:43 to 13:51 local time at the sensor located at the most shallow location approximately 1.35 m from the ice surface. At the same time, at 3, 5, 10, 20, 30 and 50 m, PAR was 198.3, 166.6, 86.9, 35.5, 15.8 and $4.2 \mu\text{mol m}^{-2} \text{s}^{-1}$, respectively (Figure 22). PAR decreases to $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the shallowest sensor from 05:00 UTC (22:00 local time) to 11:30 UTC (04:30 local time).

The lead site had much more light with the most light at all depths out of all of the locations (Figure 22A). The maximum amount of PAR was $651.7 \mu\text{mol m}^{-2} \text{s}^{-1}$ at 20:00 UTC or 13:00 local time at the sensor located approximately 1.35 m from the ice surface. At this site with the most light during the day, at this time of year, PAR still decreased to $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ from 05:00 UTC (22:00 local time) to 11:00 UTC (04:00 local time). Therefore, all locations and depths will, at a minimum, be dark for this period of the day, at this time of year. The sites with a lower initial amount of light will have no light for a longer period of time than the lead (Figure 22) due to the higher solar zenith angle later in the day and early in the morning. With a higher solar zenith angle, more light will be reflected from the same surface than would be transmitted at a time of day with a lower solar zenith angle [Perovich, 1998a]. This is shown where the site near the ridge with high snow cover is continuously dark for 16.5 hours a day but has very little light

at any time of day. The site in the centre of the floe is dark for 9.5 hours, the site at the lead edge is dark for 6.5 hours and the site in the lead is dark for 6 hours.

There is a temporal affect on the shapes of the profiles at the different depths as well. The shallower locations, especially at the lead site, have more of a peak shape where the maximum irradiance is higher but lasts a short period of time. The deeper locations at the lead site and the edge of the lead as well as all of the depths at the floe centre site have less of a curve and stay at their maximum value for a longer period of time (Figure 22).

The temporal variability shown here with these diurnal profiles will affect the amount of heat supplied to the sea water as well as the biology of the ice and water column below it. An example of this is that the water under the thin ice lead changes in magnitude from 0 to $652 \mu\text{mol m}^{-2} \text{s}^{-1}$ in just 9 hours each day. Therefore, whatever has the ability to grow there will receive a multitude of light levels at one location. The statistical analysis of the spatial variability of these locations will be discussed in 4.4.1.

4.3.2 MELTING PERIOD

PAR profiles were manually completed approximately every 3 hours at two locations approximately 35 m apart at Station F6 near the entrance of Darnley Bay on June 3, 2008 (Figure 23). One location was a white patch or bare ice with an ice thickness of 178 cm and the other was a melt pond site with an ice thickness of 150 cm and a pond depth of 6 cm. The pond had a 1 cm thick layer of ice on the surface. During the daytime the ice on the surface of the pond was melting and was wet on top and was more delicate. This sampling took place during the flooding stage of the melting ice which is the initial melting stage [Ehn *et al.*, 2011].



FIGURE 23. Site photo showing ice and pond conditions at F6. The star is the location of the white ice profiles and the circle is the location of the melt pond profiles.

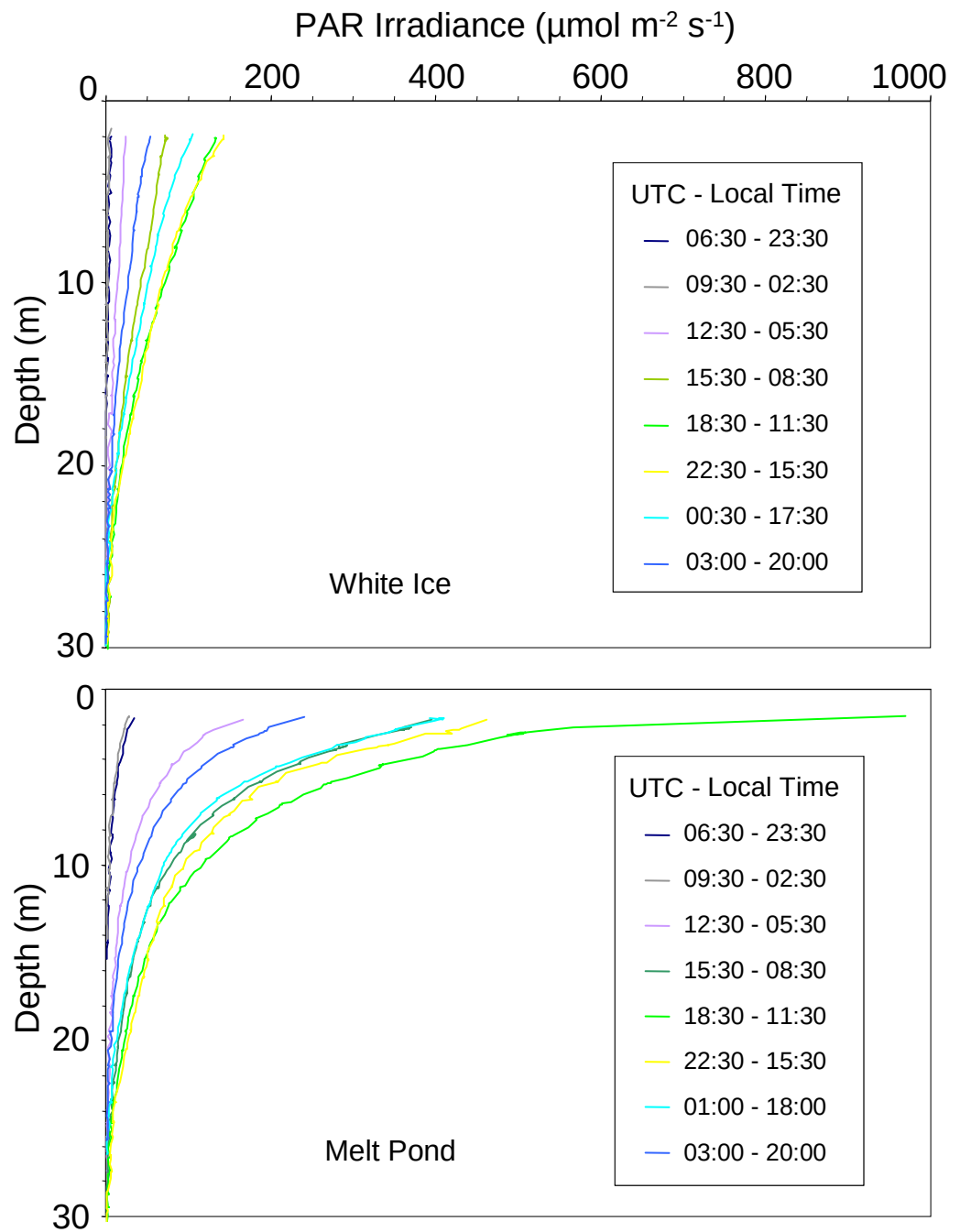


FIGURE 24. Diurnal profiles under white ice (ice-178 cm) and a melt pond (ice-150 cm pond-6 cm) at Station F6 near the entrance of Darnley Bay on June 3, 2008.

The highest sub-ice PAR value under the white ice was $143 \mu\text{mol m}^{-2} \text{s}^{-1}$ which occurred at 15:30 local time and a value of $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached at 30 m deep (Figure 24A). The lowest PAR value at the ice bottom under the white ice was $6 \mu\text{mol m}^{-2} \text{s}^{-1}$ which occurred at 23:30 and 02:30 local time and a value of $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached at approximately 10 m and 15 m deep, respectively. Though, the 02:30 profile hovered around 1 or $2 \mu\text{mol m}^{-2} \text{s}^{-1}$ from 5 m until $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached at 10 m.

The highest sub-ice PAR value under the ice with the melt pond was $970 \mu\text{mol m}^{-2} \text{s}^{-1}$ which occurred at 11:30 local time (Figure 24B). The deepest location a value of $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached was at 30 m deep and occurred at 15:30. The lowest PAR value at the ice bottom under the ponded ice was $28 \mu\text{mol m}^{-2} \text{s}^{-1}$ which occurred at 02:30 local time. A value of $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached at approximately 15 m deep at 11:30 and 02:30. The pond site illustrated much more temporal variability than the bare ice site with a range of $942 \mu\text{mol m}^{-2} \text{s}^{-1}$ happening within a nine hour period between 02:30 and 11:30 local time compared to $137 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the bare ice site. This is also a larger diurnal difference than was observed during the moorings in the freezing period even when compared to a lead with only 10 cm of ice. This is likely because of the higher solar incidence angle and increased incident solar irradiance at this time of year [Perovich, 2005].

It is interesting that the deepest that a value of $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached during any time of day for the pond and the white ice was at a depth of 30 m for both surface covers. Furthermore, the shallowest that a value of $0 \mu\text{mol m}^{-2} \text{s}^{-1}$ was reached during any time of day for the pond and the white ice was at a depth of 10 m for the white ice and 15 m for the ponded ice which was slightly different. This shows that after a certain depth, the light field becomes homogeneous regardless of the surface condition.

4.4 OBSERVED SPATIAL VARIABILITY OF PAR TRANSMISSION

4.4.1 FREEZING PERIOD

Statistical analysis was completed on the freezing period moorings completed at station D41 which was the freezing period mooring site discussed in Section 4.3.1. The Mann-Whitney U test was used to determine if the locations were statistically different from each other at the same depths.

The site near the ridge has the least amount of light as expected with very little to no light at any depth at any time of day (Figure 22D). The maximum amount of PAR was $2.1 \mu\text{mol m}^{-2} \text{s}^{-1}$ at 19:30 UTC or 12:30 local time at the sensor located at the most shallow location at approximately 3 m from the ice surface. This was due to the deep snow and the thickness of the ice. The lead site had

much more light with the most light at all depths out of all of the locations. The maximum amount of PAR is $652 \mu\text{mol m}^{-2} \text{s}^{-1}$ at 20:00 UTC or 13:00 local time at the sensor located approximately 1.35 m from the ice surface.

When comparing all of the sites two at a time at the same depth statistically, all except one are statistically different with p-values of less than 0.01. The only comparison that had a p-value greater than 0.01 was when comparing the site at the centre of the floe and the site near the ridge at 30 m below the ice surface. The p-value from this comparison was 1.00.

These results show that over a 24 hour cycle all of the locations were statistically different at each depth except the centre of the floe and the ridge at 30 m. This is because there is no light at that depth at any time of the day at those two locations. This shows that during the freezing period, different surface covers can cause the PAR in the water column to be statistically different. Generally in the freezing period, the surface is less heterogeneous than the melting period and there are fewer occurrences of the different surface covers compared to a surface covered by melt ponds which would have many differences over a smaller area.

4.4.2 MELTING PERIOD

The spatial variability of PAR transmission during the melting period can be highly variable within small areas due to the formation of melt ponds on the surface of the sea ice. Three different independent study areas are used in this section to show how variable the light can be under an ice floe during melt. The first site is a site in the centre of the Amundsen Gulf and it shows the effect of an ice edge on PAR under the ice. The second site is a pond centre, pond edge and bare ice profile site and it shows the effect of ponds and bare ice under the ice near the pond/bare ice edge. The third site is a larger site with melt ponds, a channel and bare ice sites that show how different that PAR can be under the three surfaces within an area.

At Station 405, in the centre of the Amundsen Gulf, on June 1, 2008, there was the opportunity to complete a transect of profiles from an ice edge across an ice floe. Profiles were completed at the ice edge and in a line perpendicular to the edge at 3.3 m, 7.1 m, 11.3 m, 15 m, 18.8 m, 48.0 m, 108.0 m, and 168.0 m from the ice edge. The ice thickness ranged from open water with no ice at the ice edge and 1.21 m to 1.31 m at all the sites except 2.88 m at the profile location furthest from the ice edge near where a ridge had previously formed. This site was visited during initial stages of melt and did not have melt ponds or flooding. All of the locations on the ice floe had melted and refrozen snow and ice on the

surface forming an uneven hard bumpy surface. The sky was consistently overcast as can be seen in Figure 25 which means that the incoming PAR would be diffuse and homogeneous.

TABLE 6. Physical properties and profile locations for the profiles completed at Station 405 on June 1, 2008 in the centre of the Amundsen Gulf.

Station	Ice thickness (m)	Distance From Edge (m)
STN405_1	0.00	Ice edge
STN405_2	1.29	3.3
STN405_3	1.30	7.05
STN405_4	1.28	11.25
STN405_5	1.31	14.95
STN405_6	1.29	18.75
STN405_7	1.21	48
STN405_8	1.29	108
STN405_9	2.88	168

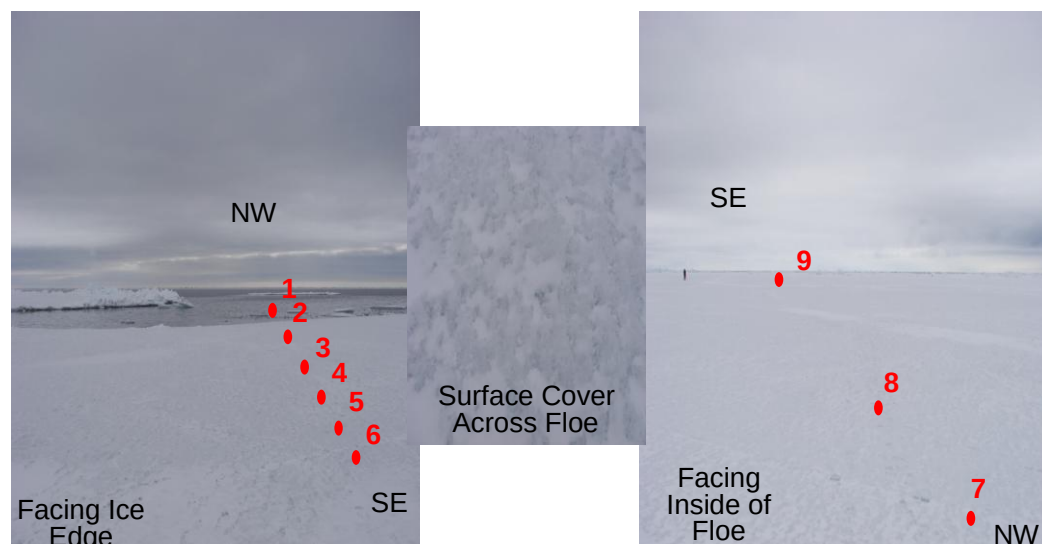


FIGURE 25. Ice and sky conditions and approximate profile locations on an ice floe near the centre of the Amundsen Gulf at Station 405 on June 1, 2008.

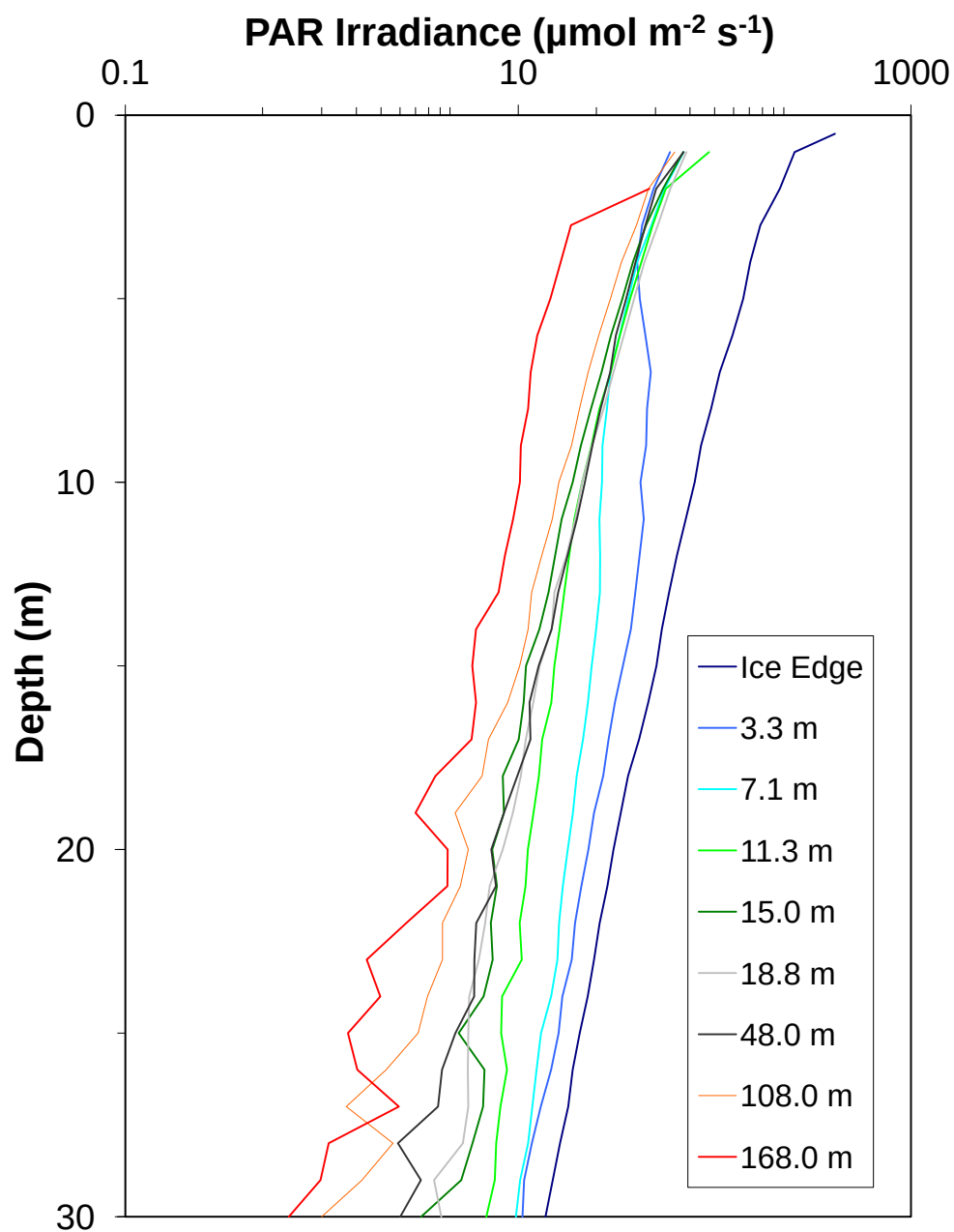


FIGURE 26. PAR profiles completed at Station 405 in a transect from an ice edge across 168 m into the floe. The legend indicates the distance the profile was located from the ice edge.

As would be expected, the profile at the ice edge into the open water had the most light out of all of the profiles in the transect from the surface to the end of the profile (Figure 26). Under the surface of the water, PAR was $456.0 \mu\text{mol m}^{-2} \text{s}^{-1}$. All of other sites, except the furthest one, varied between $67.3 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the ice bottom at 3.3 m from the ice edge to $109.2 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the ice-water interface at 11.3 m from the ice edge. The furthest location, with an ice thickness of 2.88 m and being located 168 m from the ice edge had a PAR value of $19.4 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the ice bottom.

Visually by looking at the profiles, the effect of the ice edge can be seen from the right hand side of the figure to the left hand side of Figure 26. They go in order of increasing distance from the ice edge which also corresponds to decreasing PAR. This pattern seems to stop after the profile 7.1 m from the edge. The profiles 15.0, 18.8, and 48.0 have fairly similar PAR profiles which would visually indicate that the edge only has an effect up to 11.3 m from the ice edge. This result is interesting in terms of the light field under sea ice in a marginal ice zone where floe size will be important in the overall availability of light to under ice production.

To quantify this observation a statistical analysis was completed on the profiles. The Mann-Whitney U test was used to determine if the locations were

statistically different from each other. The profiles from 2 m down from the ice surface were averaged every metre. These averages were compared, using the Mann-Whitney U test, to the other locations down to 30 m.

This analysis showed that the ice edge profile was statistically different than all of the other locations with p-values for every comparison being <0.05 . It also showed that the furthest profile, located 168 m from the ice edge, with an ice thickness of 2.88 m, was statistically different, with a p-value of <0.05 , than all of the locations except for one. The only comparison that resulted in a p-value >0.05 was when comparing the profile location closest to it at 108 m from the ice edge. The comparison of the locations at 3.3 m and 7.1 m from the edge were not statistically different, with a p-value of >0.05 . Though when comparing the locations at 3.3 m and 11.3 m from the edge, they were statistically different with a p-value of <0.05 . The comparison of 7.1 m and 11.3 m resulted in them having a p-value <0.05 as well. Though, the comparison of the profile at 11.3 m into the floe with 15.0 m, 18.8 m, and 48.0 m resulted in p-values >0.05 which means they are all the same. This means that the ice edge has an effect on all profiles up to and including the profile at 7.1 m. After 7.1 m all of the profiles are considered to be the same. These results show this to be true except when the ice thickness increases.

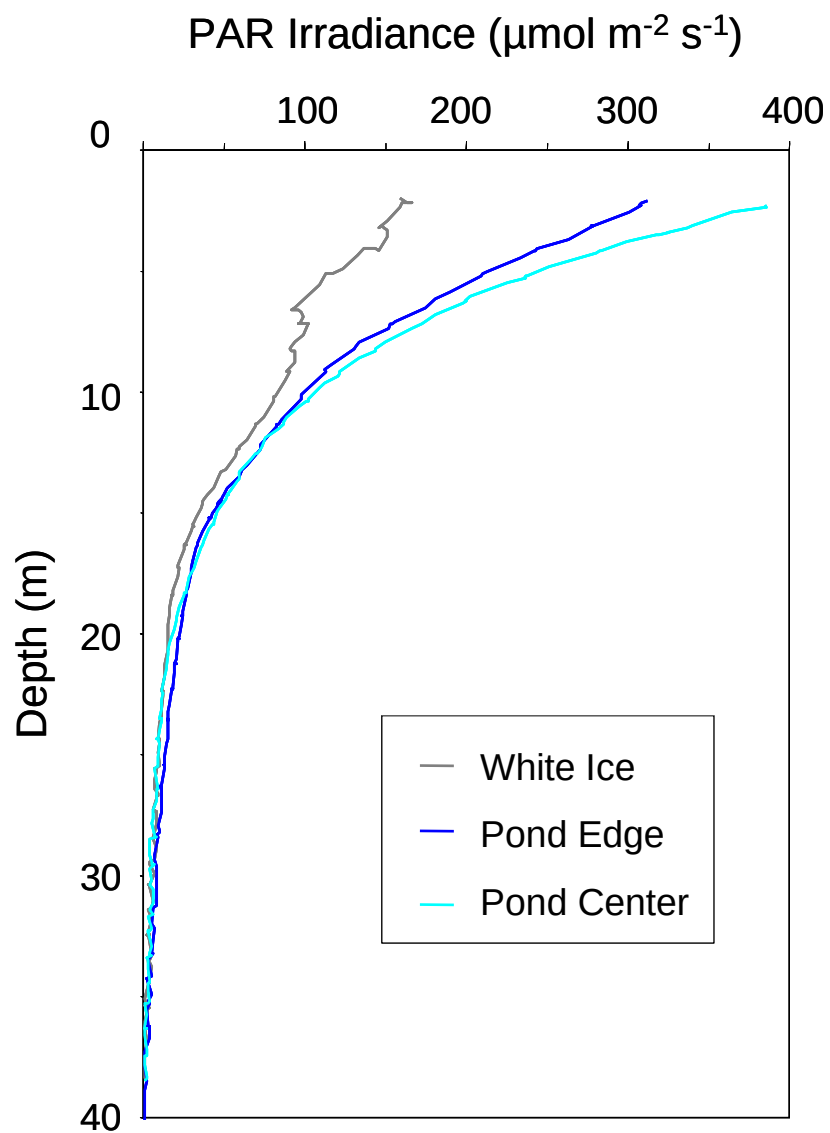


FIGURE 27. PAR profiles at F6 in white ice, 0.3 m from the edge of a pond and the centre of a pond all located in a line from the white ice to the pond centre.

PAR profiles were completed at station F6 in white ice, 0.3 m from the edge of a pond and the centre of a pond all located in a line from the white ice site to the pond centre (Figure 27). The white ice and the location 0.3 m from the ice under the pond had a thickness of 165 cm, a freeboard of 12 cm and an active layer of approximately 4 cm. The ice under the pond was 172 cm thick and the pond was 5 cm deep. There was a thin 0.5 cm layer of ice on the top of the pond and ice along the bottom of the pond had bubbles in it. It was a sunny day with a few clouds.

As can be seen in Figure 27, the melt pond profile has the highest PAR values out of the three locations as can be expected, with the pond edge having the second highest and the white ice having the lowest. The white ice profile had some cloud interference as can be seen by the rough profile line. At the bottom of the ice the PAR values for the white ice, pond edge, and pond were 159.2, 312.2, and 384.6 $\mu\text{mol m}^{-2} \text{s}^{-1}$ respectively. They all reached 0 $\mu\text{mol m}^{-2} \text{s}^{-1}$ around approximately 39 m from the ice surface.

The Mann-Whitney U test was used to determine if the locations were statistically different from each other. The profiles from 2 m down from the ice surface were averaged every metre. These averages were compared, using the Mann-Whitney U test, to the other locations from 2 m down to 30, 20, 15, 14, 13, 12, 11, and 10 m. The pond and the bare ice were statistically different with a

p-value of <0.05 when the profiles from 2 m down to 13 m were compared. The bare ice and the pond edge were statistically different with a p-value of <0.05 when the profiles from 2 m down to 12 m were compared. The pond and the pond edge comparison had p-values of >0.05 at all depths compared and therefore were not statistically different. This shows that the light spreading from the pond has a greater effect on the ice 0.3 m from the edge of the pond under the white ice than the shading from the white ice. This is comparable to recent results [Ehn *et al.*, 2011].

A similar statistical analysis was completed on the profiles at station FB5 (Table 7). The Mann-Whitney U test was used to determine if the locations were statistically different from each other. At this site there were two melt ponds, two bare ice sites and one channel within a 40 m by 40 m area. The physical properties of the locations are shown in Table 7, close-up photos of the pond and bare ice are shown in Figure 28, and an oblique aerial photograph of the site is shown in Figure 29.

TABLE 7. Physical Properties of station FB5 in Franklin Bay.

Site	Ice Thickness (cm)	Freeboard (cm)	Active Layer (cm)	Pond Depth (cm)
MP1	115	-	-	12
MP2	114	-	-	20
C1	115	-	-	14
BI1	145	15	10	-
BI2	150	18	10	-

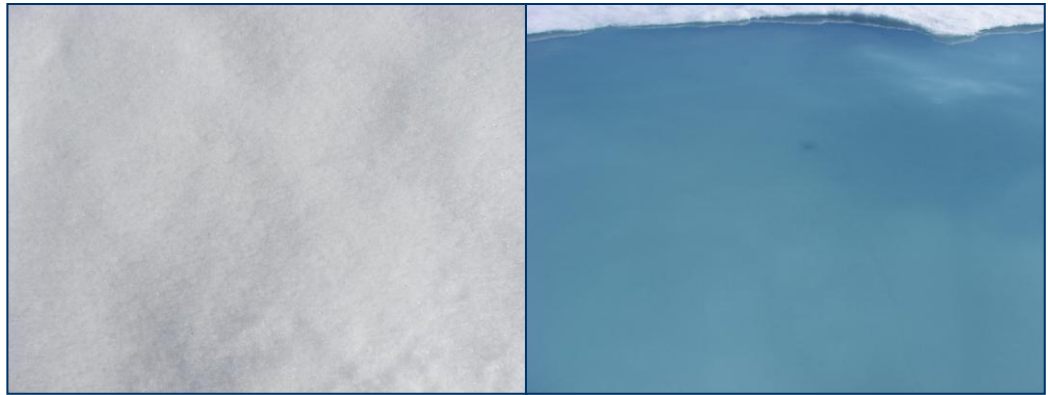


FIGURE 28. The image on the left is the bare ice surface at the FB5 study site and the image on the right is the melt pond surface.

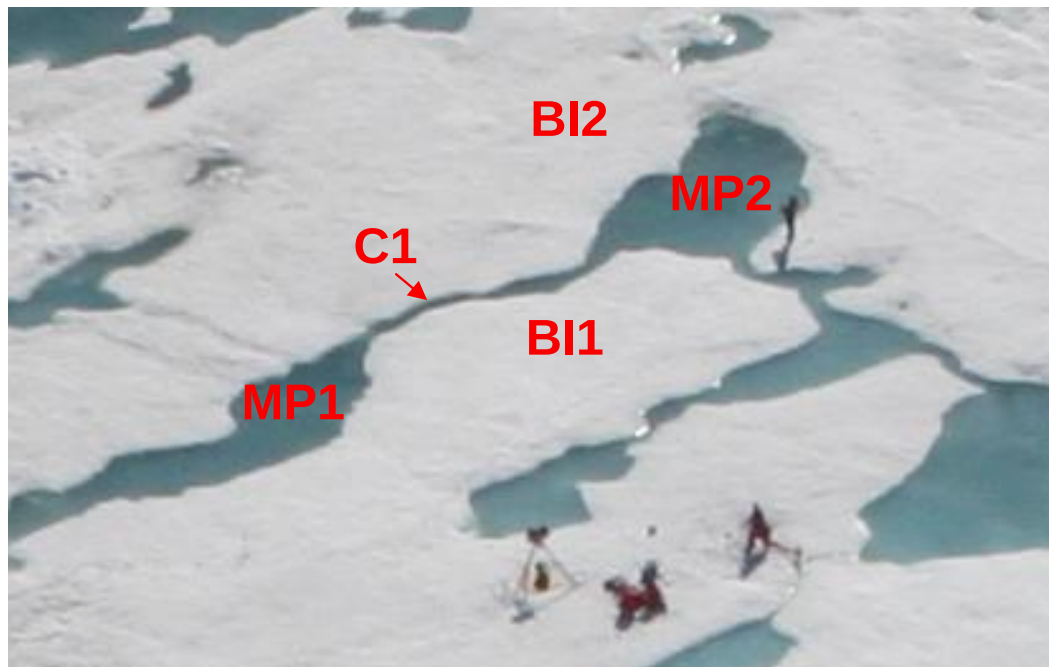


FIGURE 29. Profile locations at FB5 site in melt ponds, bare ice, and a channel on June 16, 2008. Note: The photograph is oblique therefore features that are closer to the top of the photo may look larger than they are.

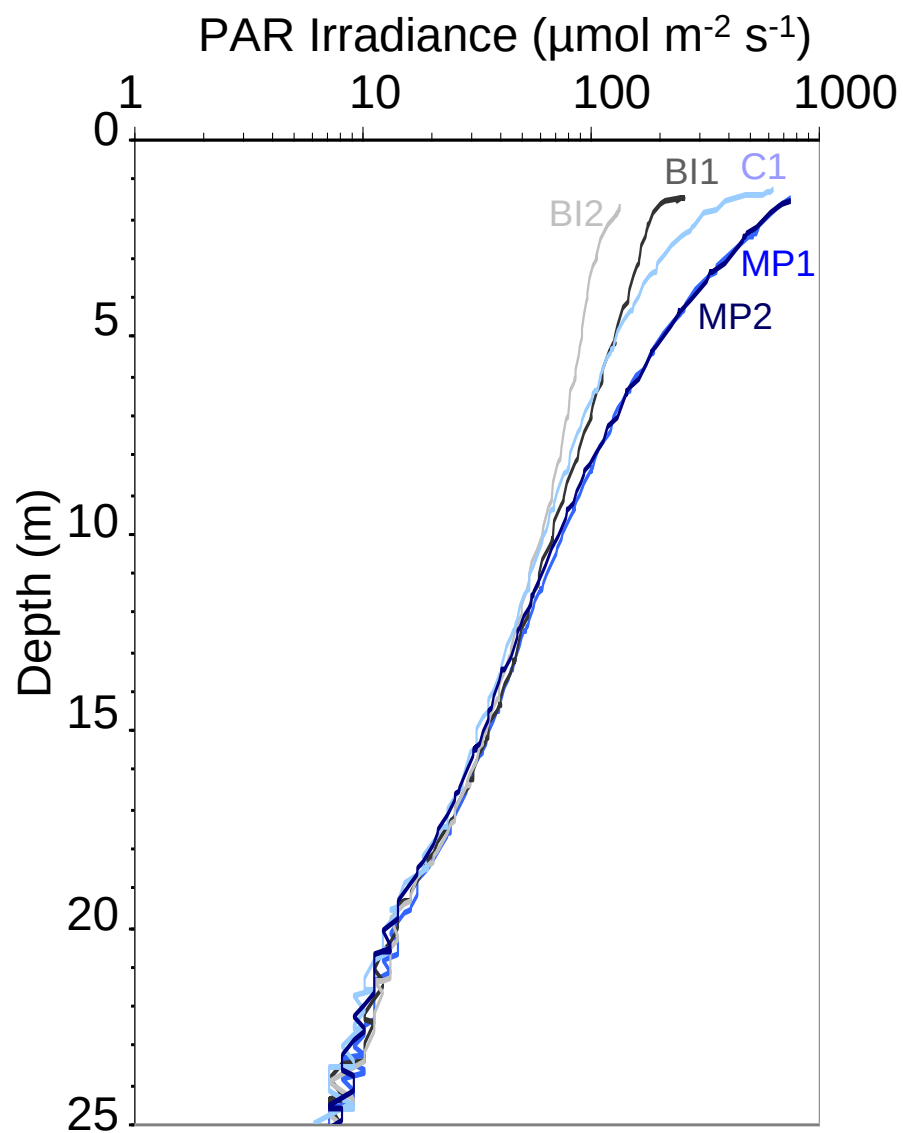


FIGURE 30. PAR under melt ponds, bare ice and a channel at FB5 in Franklin Bay on June 16, 2008 with all of them completed consecutively near solar noon.

The profiles from 2 m down to 10, 11, 12, 15, 20 and 30 m were compared for two sites at a time (i.e., pairwise comparisons). All of the locations had p-values greater than 0.05 except for two comparisons. This means that the profile under the channel is not statistically different than any of the other locations including the bare ice patches and the melt ponds. This also means that the snow patches were not statistically different from each other and that the ponds also were not statistically different from each other.

The two comparisons that were statistically different from each other were both involving the melt ponds and BI2, which was the larger of the two bare ice patches. The Mann-Whitney U, z-values and p-values of these two comparisons at each of the depths are shown in Table 8 and Table 9. MP1 and BI2 are significantly different (p-value <0.05) when compared from 2 m down to 11 m. MP2 and BI2 are significantly different (p-value <0.05) when compared from 2 m down to 10 m. This can also be seen in Figure 31 where MP1 and BI2 reach a p-value at a slightly deeper depth than MP2 and BI2. Also, as can be seen in Figure 31 and in the tables is that MP1 and BI2 have consistently lower p-values at every depth when comparing them all the way down to 30 m. This means that they are more different all the way down to 30 m than MP2 and BI2 and all of the other profile combinations at this site.

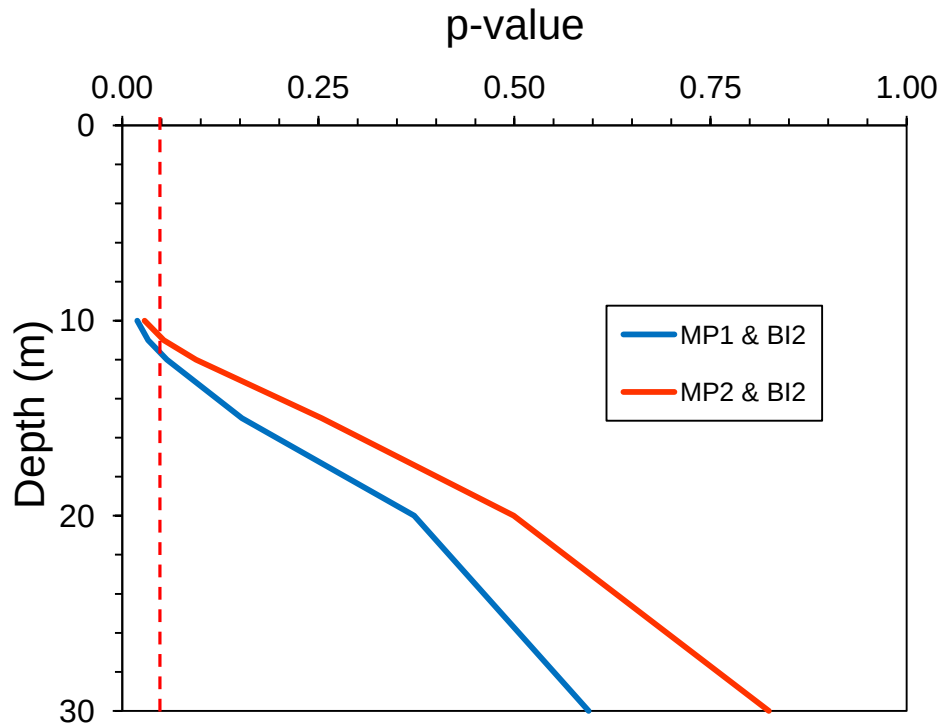


FIGURE 31. P-values from the Mann-Whitney U tests when comparing profiles from MP1 and BI2 and also MP2 and BI2 at FB5.

TABLE 8. Statistical results from the comparison of MP1 and BI2 from 2 m down to depths between 10 and 30 m showing significance of <0.05 from 2 m to 11 m.

Sites and Depths	Mann-Whitney U	Z	p-value
MP1 and BI2 from 2 m down to 30 m	414	-0.53	0.59
MP1 and BI2 from 2 m down to 20 m	167	-0.89	0.37
MP1 and BI2 from 2 m down to 15 m	78	-1.43	0.15
MP1 and BI2 from 2 m down to 12 m	39	-1.91	0.06
MP1 and BI2 from 2 m down to 11 m	28	-2.13	0.03
MP1 and BI2 from 2 m down to 10 m	19	-2.34	0.02

TABLE 9. Statistical results from the comparison of MP2 and BI2 from 2 m down to depths between 10 and 30 m showing significance of <0.05 from 2 m to 10 m.

Sites and Depths	Mann-Whitney U	Z	p-value
MP2 and BI2 from 2 m down to 30 m	435	-0.22	0.82
MP2 and BI2 from 2 m down to 20 m	175	-0.68	0.50
MP2 and BI2 from 2 m down to 15 m	85	-1.14	0.25
MP2 and BI2 from 2 m down to 12 m	43	-1.67	0.09
MP2 and BI2 from 2 m down to 11 m	31	-1.94	0.05
MP2 and BI2 from 2 m down to 10 m	21	-2.19	0.03

The percentage of each surface cover within the study area was calculated and it was determined that 82% of the surface cover was covered with melt ponds, 15% was covered with bare ice and 3% was covered with channels that connected the melt ponds. The light attenuation coefficient was calculated at each metre below the ice for each PAR profile. The calculated light attenuation coefficients from the ice bottom down to 20 m are shown in Figure 32.

In addition to the calculated light attenuation coefficients for each profile, an effective or aggregate attenuation coefficient was calculated using the percentage of each surface cover to produce a weighted average similar to what has been done by others for albedo [*Hanesiak et al.*, 2001; *Perovich et al.*, 2007]. Furthermore, the average attenuation coefficient was calculated the same as a regular average is calculated. By looking at these averages in comparison to the profile attenuation coefficients in Figure 32A, one can visibly see that there is a

difference. Figure 32B shows the difference between the profile attenuation coefficients and the effective attenuation coefficient at each depth. The vertical red dashed line indicates no change. This analysis and Figure 31 show that if you just did one profile through one of these surfaces and used it for whatever analysis you were doing you would most likely be under or over estimating the average amount of light for the area. The bare ice profiles that are located to the left of the dashed line would underestimate the average light or the attenuation coefficient at depth and the melt pond and channel profiles to the right hand side of the line would overestimate the light or attenuation. A regular average was included to show the difference that would result in just averaging the profiles versus using an aggregate or effective average. This result conforms with a recent paper by [Ehn *et al.*, 2011] which also showed that it is inaccurate to use one profile or point measurements to estimate the light under heterogeneous surfaces such as the one in this study.

The PAR sensor used during this study is a 240° sensor which may cause some concern for the K_d or radiant heating rate calculations. According to a study completed in sediments, after a very small layer <1 mm of sand, the photon irradiance and photon scalar irradiance became identical [Kühl *et al.*, 1994]. This means that the downwelling attenuation coefficient, K_d , and the scalar attenuation coefficient, K_o , are identical after a very thin layer of sand which would scatter more than the water column under the ice. In addition, profiles

from FB5 under similar ponds completed with the 240° Alec PAR sensor and the 180° Satlantic HyperOCR profiler at similar times were compared. The coefficient of determination, R^2 , value for the comparison was 0.99. Therefore, the upwelling irradiance is considered negligible in this study and the light field is assumed to be asymptotic leading to K_d and K_o being identical or near identical. This results in the Alec PAR sensor data being suitable for K_d and heating calculations in this study.

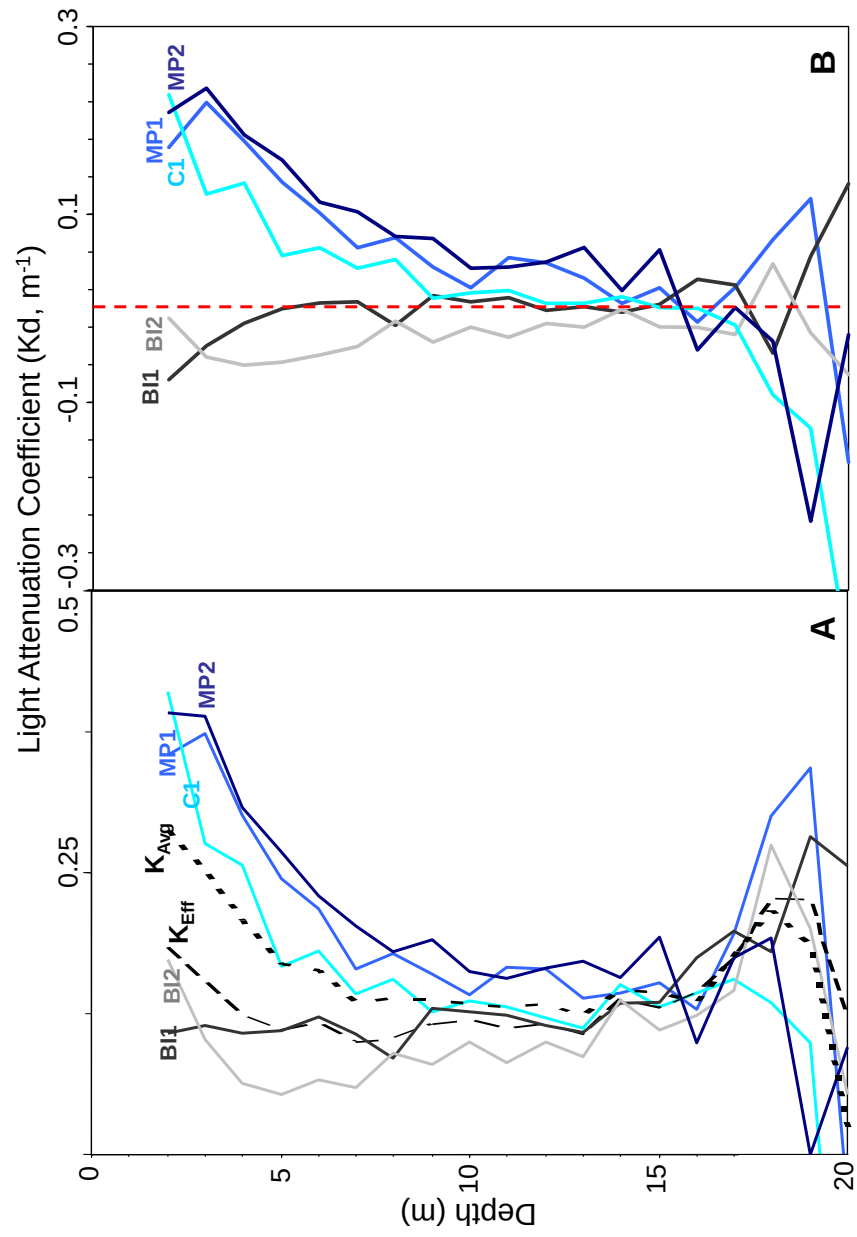


FIGURE 32. A. Attenuation coefficient profiles for melt ponds, bare ice patches and a channel at FB5, the calculated effective attenuation, and the average attenuation based on those profiles. **B.** The difference between the attenuation profiles and the calculated effective attenuation. The red dashed line indicates 0 change.

4.5 CASE STUDY OF SOLAR HEATING AND POTENTIAL PRIMARY

PRODUCTION DURING ADVANCED MELT

4.5.1 SOLAR HEATING USING RADIANT HEATING

When melt begins it will impact the amount of solar energy absorbed during the melt season more than when freeze-up occurs [Perovich *et al.*, 2007]. Earlier melt onset will result in earlier ponds which means more solar energy through ponds because of higher incidence for a longer period of time.

Near the ice bottom, at 2 m from the air-ice interface, the radiant heating rate was the highest out of all the depths and ranged from 0.004 to $0.033^{\circ}\text{C h}^{-1}$ or $0.08^{\circ}\text{C day}^{-1}$ to $0.80^{\circ}\text{C day}^{-1}$. The lowest value occurred under BI1 and the highest value occurred under MP1 with the range being $0.029^{\circ}\text{C h}^{-1}$. After 3 m BI2 was the lowest and stayed fairly consistent at $0.001^{\circ}\text{C day}^{-1}$.

This apparent K_d is due to shading and not due to IOPs. It is because of the shading mechanism and the window effect of the melt ponds with horizontal spreading of light. There is a sharp decrease in temperature between the salt and melt water layer. One reason for this could be because of algae absorption. Algae only use approximately 15% of the light they absorb and the rest of the light is converted to heat [Zeebe *et al.*, 1996].

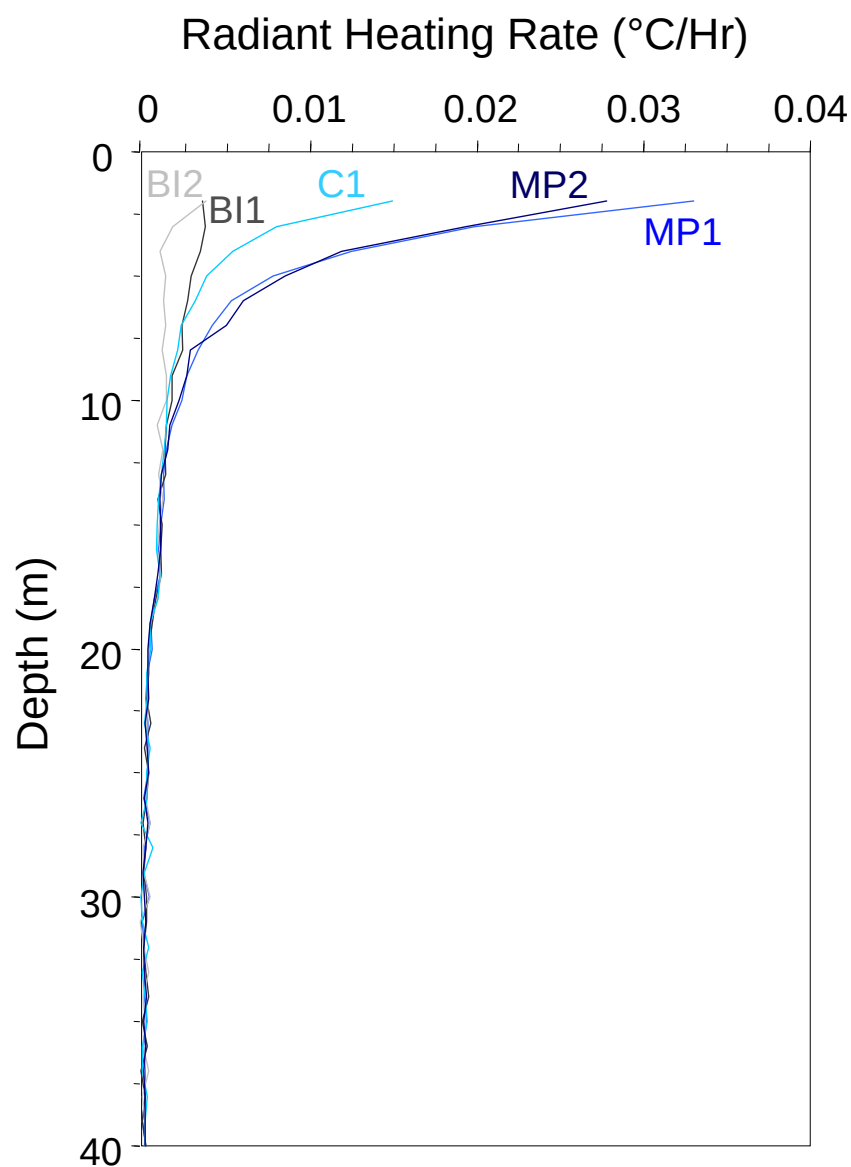


FIGURE 33. Radiant heating rate profiles under melt ponds, bare ice patches and a channel at FB5 in Franklin Bay on June 16, 2008 with all of them completed consecutively near solar noon.

4.5.2 POTENTIAL LIGHT LIMITED PRIMARY PRODUCTION

Chlorophyll *a* was measured during a cast with the Satlantic, Inc., HyperOCR spectroradiometer which also had a Wetlabs in vivo fluorometer on it. Under the ice was accessed by a diver through a hole in the ice created by a seal. Potential light limited primary production values were calculated for the FB5 site in Franklin Bay using the measured chlorophyll values as well as the photosynthesis-irradiance curve values calculated and provided in a recent paper [Palmer *et al.*, 2011]. These values were obtained through analysis and data collection during similar time periods and locations in the CFL project.

Near the ice bottom, at 2 m from the air-ice interface, the potential light limited primary production rates ranged from 0.82 to 0.94 mg C m⁻³ h⁻¹. The lowest value occurred under BI1 and the highest value occurred under both of the melt ponds. After the production peak at 2 m it decreases down to around 0.44 mg C m⁻³ h⁻¹ at 3 m. At 2 m the potential primary production is the highest out of any depth in the water column. Nutrients were not considered though likely limited in this melt layer [Mundy *et al.*, 2011].

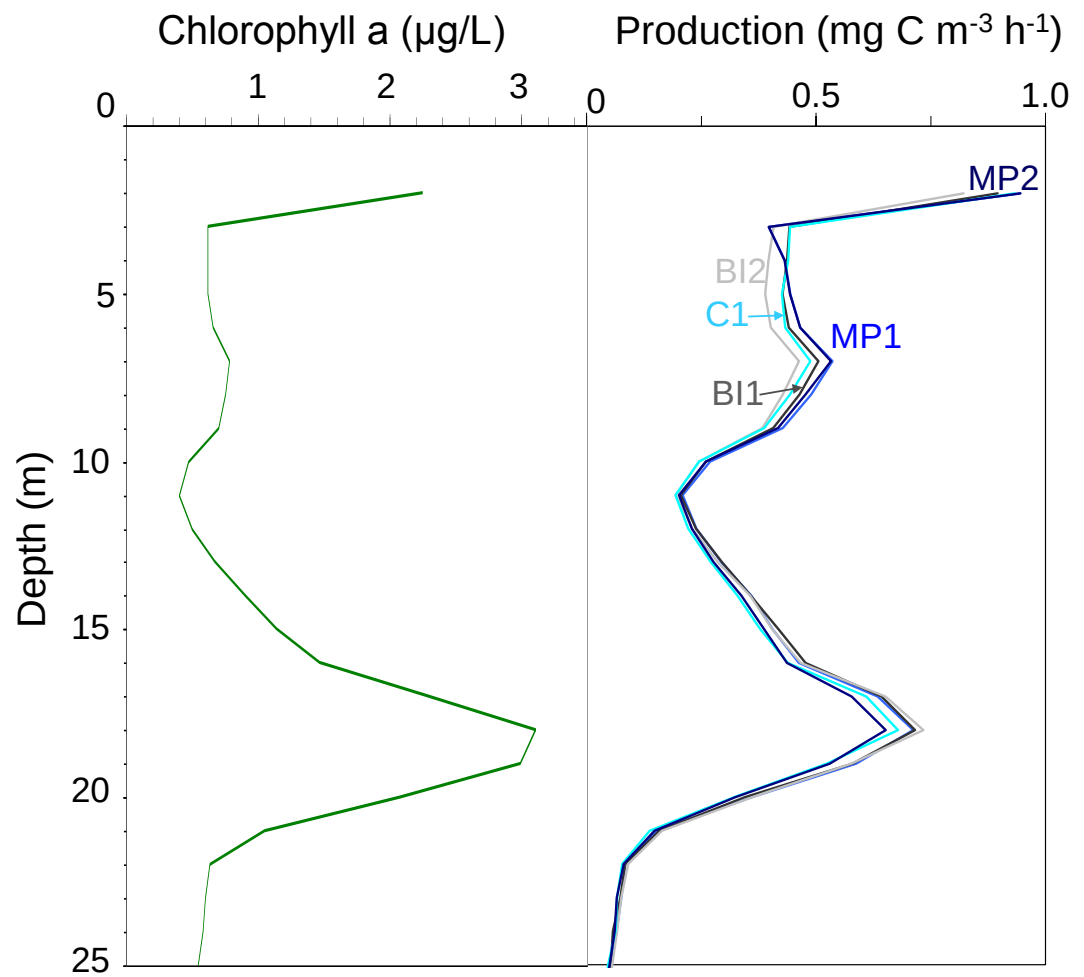


FIGURE 34. Chlorophyll *a* measurements and calculated potential primary production under melt ponds, bare ice patches and a channel at FB5.

The highest variability in potential primary production occurred at 7 and 18 m down from the surface. At 7 m the lowest potential primary production was $0.46 \text{ mg C m}^{-3} \text{ h}^{-1}$ which occurred at BI1 and the highest was $0.54 \text{ mg C m}^{-3} \text{ h}^{-1}$ which occurred at MP1. At 18 m the lowest potential primary production was $0.65 \text{ mg C m}^{-3} \text{ h}^{-1}$ which occurred at MP2 and the highest was $0.73 \text{ mg C m}^{-3} \text{ h}^{-1}$ which occurred at BI1. Since the light field becomes homogeneous at approximately 11 m down at this site, the variability in potential primary production that occurs at this depth is not explained by light. It is explained by the high amount of chlorophyll *a* so that even if there is only a small difference in the amount of light at that depth between profiles, the larger amount of chlorophyll *a* will amplify the variability. This location is where the chlorophyll maximum occurred.

The integrated production in the water column from the ice bottom down to 25 m ranged from 8.19 and $8.20 \text{ mg C m}^{-2} \text{ h}^{-1}$ at BI2 and C1 respectively to 8.32 and $8.66 \text{ mg C m}^{-2} \text{ h}^{-1}$ at MP2 and MP1 respectively which is a range of $0.47 \text{ mg C m}^{-2} \text{ h}^{-1}$. These results show that the amount of production under the ice is variable and one factor could be the heterogeneous surface cover.

CHAPTER 5 CONCLUSIONS & FURTHER DIRECTIONS

The understanding of the interactions of solar radiation with the atmosphere-sea ice-ocean system is important for properly representing the ice-albedo feedback mechanisms and heating in models which can affect prediction of the future sea ice extents in the summer [*Holland et al.*, 2006].

There is limited data studying light, salinity and temperature directly under the ice down to 10 or 20 m. Majority of studies measure the light above the ice [*Perovich et al.*, 1998; *Hanesiak et al.*, 2001; *Ehn*, 2006], in the ice or at the ice bottom [*Perovich et al.*, 1998; *Ehn et al.*, 2008; *Light et al.*, 2008]. Therefore, the data in this thesis provide an invaluable addition to the data and studies of PAR, salinity and temperature in the upper ocean.

During the freezing period, the temperature and salinity variability under the ice is much less than during the melting season with less impact from the surface cover. The sea water directly under the ice is slightly cooler than the rest of the water column and is well mixed down to the thermocline where the water warms up and the salinity increases. For some locations this happened around approximately 50 m and for other locations it wasn't present within the top 60 m of the water column where my profiles ended.

During the freezing period there can be large temporal variability in PAR under the ice and in leads. Though, even in areas with the largest variability measured, the greatest difference in one study area between the maximum and minimum values during the day during the freezing period is still less than the greatest difference in the melting season. In the melting season, the pond site had a range of $942 \mu\text{mol m}^{-2} \text{s}^{-1}$ happening within a nine hour period between 02:30 and 11:30 local time compared to $137 \mu\text{mol m}^{-2} \text{s}^{-1}$ at the bare ice site. Whereas, the lead site in the freezing period under thin ice had a change in magnitude of $652 \mu\text{mol m}^{-2} \text{s}^{-1}$ within a 9 hour period between 04:00 and 13:00 local time each day.

During the freezing period, there can also be large spatial variability in PAR under the ice and in leads. The PAR values at the ice bottom at the snow and ice sites, ranged between approximately 6 to $36 \mu\text{mol m}^{-2} \text{s}^{-1}$ and at the lead sites at the water surface or ice bottom ranged from approximately 60 to $900 \mu\text{mol m}^{-2} \text{s}^{-1}$. Under the ridge in the diurnal moorings, there was very little to no light whereas in the lead during the same time and same ice floe there was $652 \mu\text{mol m}^{-2} \text{s}^{-1}$.

The melting season profiles showed that there is more temperature and salinity variability under the ice than in the freezing period and more variability throughout the water column in the upper ocean. The temperature profiles were extremely variable during this time in the upper 40 m and showed turbulence

due to mixing. The salinity under the ice showed a melt layer of low salinity with the largest melt layer being approximately 0.8 m thick. Furthermore, the melt ponds have a dome that forms under the ice and the salinity directly under the pond in the domes was lower than under the bare ice. Therefore, if you only took a water sample under the bare ice versus under the dome to estimate the entire under ice area it would be inaccurate. Also, if you assumed that the water column in the 10 m below the sea ice is homogeneous, especially during the melting season, this would be extremely inaccurate as can be seen in the variability in the PAR and temperature profiles during this time period. An effective or aggregate average of the water column properties under the ice could be calculated using sample values from an area. This would be more accurate than calculating a regular average from the area. There is spatial PAR variability both in the freezing and melting periods. One difference between them is that there is more spatial variability in a smaller area during the melt due to the melt ponds.

Many further studies could be completed with the data collected during this thesis project. Some of the data yet to be used and projects to be completed are listed below:

- A more in depth comparison of sites during the freezing period using the snow grain, albedo, and PAR data.

- Investigate the role of new sea ice growth on PAR over a period of time using the mooring located in the lead at Station D41.
- A comparison and/or calculation of melting and/or heating using the temperature and salinity profiles from the 30 hour melting season time series under the pond and the bare ice at Station F6.
- A look at temperature and salinity from an ice floe transect in Darnley Bay during June starting at the ice edge and doing a profile every 500 m going into the floe about 2.5 km then going across to another location and completely profiles at the same distances from the ice edge up to the edge.
- Further investigate the role of the sizes and distribution of the melt ponds and bare ice patches on the PAR distribution under the ice at Station FB5.

Some other future work to be done that would involve additional data collection, would be to do a more intensive PAR profile site with PAR profiles in the ponds, at the edge of the ponds, across the bare ice, etc. If possible, it would be useful to have PAR moorings over a few days in those locations down to 20 m to see the full spatial and temporal effect of the heterogeneous surface cover. This study could also include ice algae data and phytoplankton data.

As shown through the chapters in this thesis one water column profile is not representative of the PAR, salinity or temperature under the ice when the surface cover is spatially heterogeneous.

CHAPTER 6 REFERENCES

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APPENDIX A - DATA TABLE

Freezing Period Sampling (CFL Leg 7, March & April)

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	SNOW	ICE THICKNESS	CTD
D29	22/03/08	Yes	No	Dep	Yes	Yes
D33_1	25/03/08	Yes	No	Dep	Yes	Yes
D33_2	28/03/08	Yes	Yes	Dep	Yes	Yes
D33_1a	29/03/08	No	Yes	Dep	Yes	Yes
D33_1b	29/03/08	No	Yes	Dep	Yes	Yes
D33_1c	29/03/08	No	Yes	Dep	Yes	Yes
D33_d	29/03/08	No	Yes	Dep	Yes	Yes
D33_e	29/03/08	No	Yes	Dep	Yes	Yes
D33_3	31/03/08	Yes	Yes	Dep	Yes	Yes
D33_4	3/04/08	Yes	Yes	Dep	Yes	Yes
D36_1	6/04/08	Yes	Yes	Dep	Yes	Yes
D36_2	6/04/08	Yes	Yes	Dep	Yes	Yes
D36_3	6/04/08	No	Yes	Dep	Yes	Yes
D36_4	6/04/08	No	Yes	n/a	Lead/Open Water	Yes
D36_5	6/04/08	No	Yes	n/a	Lead/Open Water	Yes
D36_6	6/04/08	No	Yes	n/a	Lead/Open Water	Yes
D36_7	6/04/08	No	Yes	No snow	Yes	Yes
D36_8	7/04/08	No	Yes	No snow	Yes	Yes
D36_8b	7/04/08	No	Yes	No snow	Yes	Yes
D36_9	8/04/08	Yes	Yes	SGA, D, Den	Yes	Yes
D36_10	8/04/08	No	Yes	No snow	Yes	Yes
D36_11	9/04/08	Yes	No	Dep	Yes	Yes
D36_12	9/04/08	No	No	Dep, SGA,D,S	Yes	Yes
D36_13	9/04/08	No	No	Dep, SGA,D,S	Yes	Yes
D36_14	9/04/08	No	No	Dep	Yes	Yes

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	SNOW	ICE THICKNESS	CTD
D38_1	10/04/08	No	Yes	No	Lead/Thin Ice/Open Water	Yes
D38_2	10/04/08	No	Yes	Dep, SGA,D,S	Yes	Yes
D38_3	10/04/08	No	Yes	Dep, SGA,D,S	Yes	Yes
D38_4	10/04/08	No	Yes	Dep, SGA,D,S	Yes	Yes
D38_5	10/04/08	No	Yes	n/a	Lead/Thin Ice/ Open Water	Yes
D38_6	11/04/08	No	Yes	Dep, SGA,D,S	Yes	Yes
D38_7	11/04/08	Yes	Yes	Dep, SGA,D,S	Yes	Yes
D38_8	11/04/08	No	Yes	Dep	Yes	Yes
D38_9	11/04/08	No	Yes	Frost Flowers	Lead/Thin Ice	Yes
D38_10	12/04/08 to 13/04/08	No	Yes	Dep	Lead/Thin Ice	Yes
D38_11	12/04/08 to 13/04/08	No	Yes	Frost Flowers	Lead/Thin Ice	No
D38_12	12/04/08 to 13/04/08	No	Yes	Frost Flowers	Lead/Thin Ice	No
D38_11a	13/04/08	No	Yes	Frost Flowers	Lead/Thin Ice	Yes
D38_10a	13/04/08	No	Yes	Dep	Lead/Thin Ice	Yes
D41_1	16/04/08 to 22/04/08	No	Yes	Frost Flowers, Dep	Lead/Thin Ice	No
D41_2	16/04/08 to 22/04/08	No	Yes	Dep, SGA,D,S, Cap	Yes	No
D41_3	16/04/08 to 22/04/08	No	Yes	Dep, SGA,D,S, Cap	Yes	No
D41_5	16/04/08 to 22/04/08	No	Yes	Frost Flowers, Dep	Yes	No

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	SNOW	ICE THICKNESS	CTD
D41_5b	19/04/08	No	Yes	Frost Flowers, Dep	Yes	Yes
D41_6	17/04/08	No	Yes	Frost Flowers, Dep	Yes	Yes
D41_7	17/04/08	No	Yes	Frost Flowers, Dep	Yes	Yes
D41_8	19/04/08	No	Yes	Frost Flowers, Dep	Yes	Yes
D41_9	20/04/08	No	Yes	Dep, SGA,D,S, Cap	Yes	Yes
D41_10	20/04/08	No	Yes	n/a	Lead/Open Water	Yes

**Snow Pits: T = temperature profile, Cap = capacitance plate, SGA = snow grain analysis,
D = density Dep = Depth**

Freezing/Melting Period Sampling (CFL Legs 8b & 9a, May & June)

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	ASD DW & UW	SNOW	ICE THICKNESS	CTD
F2_1	16/05/08	Yes	Yes	No	Dep, T, S	Yes	Yes
F2_2	17/05/08	Yes	Yes	No	No Snow	Yes	Yes
F2_3	17/05/08	Yes	Yes	No	Dep, T, S	Yes	Yes
F2_4	17/05/08	Yes	Yes	No	Dep, T, S, D	Yes	Yes
F3_1a	20/05/08	Yes	Yes	No	Dep, T, Cap,S,D	Yes	Yes
F3_1b	20/05/08	No	Yes	No	Snow from F3_1a cleared	Yes	Yes
D45_1	21/05/08	No	Yes	No	Dep, T, S	Yes	No, PAR Mooring
D45_2	21/05/08	No	Yes	No	Dep, T, S	Yes	No, PAR Mooring
D45_3	21/05/08	No	Yes	No	Ice Edge	Yes	No, PAR Mooring
D45_4	21/05/08	No	Yes	No	Open Water	Open Water	Yes
F4	24/05/08	No	Yes	Yes	Dep, T, S, D,SGA	Yes	Yes
MY1	25/05/08	No	Yes	Yes	Dep, T, S, SGA	Yes	Yes
MY2	26/05/08	No	Yes	Yes	Dep, T, S	Yes	Yes
F5_1	28/05/08	No	Yes	No	Dep	Yes	No, PAR mooring
F5_2	28/05/08	Yes	Yes	No	No	Ice edge	Yes
F5_3	28/05/08	No	Yes	Yes	Dep, T, S	Yes	Yes
D46_1	30/05/08	No	Yes	Yes	Dep, T	Yes	Yes
D46_2	30/05/08	No	Yes	Yes	n/a	Lead/ Open Water	Yes
D46_3	30/05/08	No	Yes	No	n/a	Lead/Open Water	Yes
D46_4	30/05/08	No	Yes	No	n/a	Ice Edge	Yes
D46_5	30/05/08	No	Yes	Yes	Dep, T	Yes	Yes
D46_6	30/05/08	No	Yes	Yes	Melt	Yes	Yes

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	ASD DW & UW	SNOW	ICE THICKNESS	CTD
					Pond		
D47_1	31/05/08	No	Yes	Yes	Dep	Yes	Yes
D47_2	31/05/08	No	Yes	Yes	Melt Pond	Yes	Yes
D47_3	31/05/08	No	Yes	No	Ice Edge	No	Yes
D47_4	31/05/08	No	Yes	No	Dep	Yes	Yes
STN405_1	1/06/08	No	Yes	No	n/a	Ice Edge	Yes
STN405_2	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_3	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_4	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_5	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_6	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_7	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_8	1/06/08	No	Yes	No	Dep	Yes	Yes
STN405_9	1/06/08	No	Yes	No	Dep	Yes	Yes
F6_1	2/06/08	No	Yes	Yes	Dep, T	Yes	Yes
F6_2	2/06/08	No	Yes	No	Melt Pond Edge	Yes	Yes
F6_3	2/06/08	No	Yes	Yes	Melt Pond	Yes	Yes
F6_D1	2/06/08 to 3/06/08	No	Yes	No	Dep	Yes	Yes
F6_D2	2/06/08 to 3/06/08	No	Yes	No	Melt Pond	Yes	Yes
F6_4	3/06/08	No	Yes	No	Dep	Yes	Yes
F6_5	3/06/08	No	Yes	No	Dep	Yes	Yes
F6_6	3/06/08	No	Yes	No	Dep	Yes	Yes
F6_7	3/06/08	No	Yes	No	Dep	Yes	Yes
F6_8	3/06/08	No	Yes	No	Melt Pond	Yes	Yes
F7_1	07/06/08	No	Yes	Yes	Dep, T	Yes	Yes
F7_2	07/06/08	No	Yes	Yes	Melt Pond, Dep	Yes	Yes
F7_3	07/06/08	No	Yes	Yes	Melt Pond	Yes	Yes

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	ASD DW & UW	SNOW	ICE THICKNESS	CTD
F7_3b	07/06/08	No	No	Yes	Melt Pond, Dep, T	No	No
F7_4	07/06/08	No	Yes	Yes	Dep, T	Yes	Yes
F7_5	09/06/08	No	Yes	Yes	Dep, T	Yes	Yes
F7_6	09/06/08	No	Yes	Yes	Melt Pond, Dep	Yes	Yes
F7_7	09/06/08	No	Yes	Yes	Dep, SGA	Yes	Yes
F7_8	09/06/08	No	Yes	Yes	Dep, T	Yes	Yes
F7_9	09/06/08	No	Yes	Yes	Melt Pond, Dep	Yes	Yes
F7_10	11/06/08	No	Yes	Yes	Dep, T	Yes	Yes
F7_11	11/05/08	No	Yes	Yes	Melt Pond, Dep, T	Yes	Yes
F7_T1	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T2	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T3	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T4	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T5	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T6	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T7	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T8	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T9	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T10	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T11	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_T12	12/06/08	No	Yes	No	Dep	Yes	Yes
F7_13	13/06/08	No	No	Yes	No	No	No
FB6_1	15/06/08	No	Yes	No	Dep	Yes	Yes
FB5_1	16/06/08	No	Yes	Yes	Melt Pond, Dep, T	Yes	Yes
FB5_2	16/06/08	No	Yes	Yes	Dep, T	Yes	Yes
FB5_3	16/06/08	No	Yes	No	Melt Pond, Dep	Yes	Yes
FB5_4	16/06/08	No	Yes	No	Dep	Yes	Yes

STATION	DATE dd/mm/yy	LICOR PAR	ALEC PAR	ASD DW & UW	SNOW	ICE THICKNESS	CTD
FB5_5	16/06/08	No	Yes	No	Dep	Yes	Yes
FB5_6	16/06/08	No	Yes	No	Melt Pond, Dep	Yes	Yes
FB5_7	16/06/08	No	Yes	No	Melt Pond, Dep	Yes	Yes
HR1	17/06/08	No	Yes	No	Dep, T	Yes	Yes
HR2	17/06/08	No	Yes	No	Open Water, Between Floes	No	Yes
FB5_8	18/06/08	No	No	Yes	Melt Pond, Dep, T	No	No
FB5_9	18/06/08	No	No	Yes	Melt Pond, Dep, T	No	No
FB5_10	18/06/08	No	No	Yes	Melt Pond, Dep, T	No	No
FB7_1	20/06/08	No	Yes	No	Dep	Yes	Yes
FB7_2	20/06/08	No	Yes	No	Melt Pond, Dep	Yes	Yes
FB7_3	21/06/08	No	Yes	Yes	Melt Pond, Dep, T	Yes	Yes
FB7_4	21/06/08	No	Yes	No	Dep	Yes	Yes
STN1216	23/06/08	No	Yes	No	Open Water	No	Yes

Snow Pits: T = temperature profile, Cap = capacitance plate, SGA = snow grain analysis, D = density Dep = Depth

ASD: DW = Downwelling, UW = Upwelling