Characterizing the influence of Atlantic water intrusion on water mass formation and primary production in Kongsfjorden, Svalbard

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Abstract

With warming global temperatures and changes to large-scale ocean circulation patterns, warm water intrusion into Arctic fjords is increasingly affecting fragile polar ecosystems. This study investigated how warm Atlantic water intrusion and the tidewater glacial melting it causes impacted water mass formation and primary productivity in Kongsfjorden, Svalbard. Data were collected over a 2-week period during the height of the melt season in August near the Kronebreen/Kongsvegen glacier complex, the most rapidly retreating glacier in Spitsbergen. Since 1998, intruding waters have warmed between 4 and 5.5°C, which has prevented sea ice formation and changed the characteristics of fjord bottom waters. Increased glacial melting in the last decade has changed the characteristics of surface waters in the fjord. Modeled light fields suggest that suspended sediment in this glacial meltwater has reduced the euphotic zone close to the ice face, preventing high primary production in both the consistent and intermittent sediment-laden meltwater plumes. However, measurements collected close to terrestrially terminating glaciers indicate that extremely high primary production can occur in conditions of low turbidity. The results of this study support a three-part model of the effects of warm-water intrusion on water mass formation and primary production, where changes in sea ice coverage and tidewater glacial dynamics affect the optical light field. This model allows for spatial predictions for the most likely impacts of warm water intrusion on primary production in Spitsbergen, and could be extrapolated out to explore potential phytoplankton response in other regions susceptible to warm-water intrusion.
Introduction

*Climate change in Arctic fjord ecosystems*

Anthropogenic climate change is expected to have profound effects on ecosystems globally, with the Arctic predicted to be one of the most impacted regions. Due to polar amplification, or the disproportionate impact of the greenhouse effect at high latitudes (Holland and Bitz 2003), polar ecosystems are already experiencing large air temperature changes (Bekryaev et al. 2010). Based on conservative projections by the International Panel on Climate Change (IPCC), Kattsov et al. (2005) estimate that by 2071-2090, air temperatures will be 5-6°C above the 1981-2000 averages in the Arctic, an increase more than three times greater than the global projections for this same time interval. Ice-albedo feedback largely drives this projected change; as warmer air temperatures increase the melting rates of sea ice in the summertime, the reflectivity of the ocean surface lowers. This causes higher regional heat absorption, which in turn melts more ice (Figure 1, Perovich et al. 2007, Screen and Simmonds 2010). New minima in Arctic sea ice extent and area in September 2007 (Comiso et al. 2008) and again in the summer of 2012 reflect the ways that decreases in ice coverage interplay with and cause increased local air temperatures.

In addition to warming air temperatures, Arctic ecosystems have begun to experience influxes in warmer ocean waters, which have been attributed to changes in atmospheric circulation in the North Atlantic (Holland et al. 2008). These warm water influxes have had devastating consequences for tidewater glaciers in the Arctic, and are also partially responsible for decreases in sea ice production in Arctic fjords (Figure 1). In 1997, following the sudden increase in subsurface ocean water temperature along the west coast of Greenland, the Jakobshavn Isbrae glacier in Disko Bay experienced a doubling in glacial velocity. Holland et al. (2008) attribute this speed increase to a destabilization and rapid melting of the floating ice tongue in front of the glacier, which had previously acted to buffer the glacier against changes in ocean currents. In recent years, further summer speedups in glacial velocity led to the retreat of the terminus of the glacier to a 1300m deep portion of the basin by 2013, exposing an even greater melting face to warming ocean waters (Joughin et al. 2014). At the temperate tidewater LeConte glacier in Alaska, a warm (7.2°C) layer of water extending from 130m depth to the fjord
Figure 1. Diagram of the effects of warm water intrusion and warming air temperatures on glaciers and sea ice, and the albedo effect on local warming, with arrows scaled to represent relative impact. Glacial melting is primarily driven by changes to water temperature, while decreases in sea ice coverage and thickness are ablated by both increases in air and water temperature. As sea ice melts and glaciers retreat, the positive effects of ice-albedo feedback further contribute to local warming trends.

Glaciers were found to have dramatically increased glacial retreat in 2000 (Motyka et al. 2003). Submarine melting due to this warm water is thought to account for 57\% of the total ice loss at the terminus, with late summer melting exceeding 12m d\(^{-1}\) water equivalent (w.e.) over the submerged face of the glacier (Motyka et al. 2003). Another study of submarine melting on three western Greenland glaciers found that, despite finding much lower melt rates than the study in Alaska (0.7 to 3.9m d\(^{-1}\) w.e.), submarine melting accounted for 20-80\% of the summer melting of these glaciers (Rignot et al. 2010).
In addition to impacting marine-terminating glaciers, increases to water temperatures in Arctic fjords and bays cause dramatic changes to local biotic communities. Warm Atlantic water intrusion into Arctic-type fjords allows organisms formerly confined to lower latitudes to persist and even outcompete endemic species as sea surface temperatures rise (Post et al. 2010). Research in the Pacific Arctic by Grebmeier et al. (2010) indicates that, while clams and the bird and marine mammal populations that feed on them are declining rapidly, pelagic fishes once constrained to lower latitudes are thriving. These temperature changes may also prove inhospitable to certain Arctic species, limiting their ranges significantly and even leading to extinctions. In a review of 51 instances of Arctic marine biota response to climate change, range shifts and changes in abundance, growth, behavior, and community were attributed primarily to warming ocean temperatures and associated sea ice loss (Wassmann et al. 2011). As ocean temperatures warm by just a few degrees Celsius in ecosystems that have historically balanced at temperatures close to freezing, seasonality and food availability is greatly impacted, which in turn has permanent effects on species distribution and extent.

Changes to sea ice coverage caused by warm-water influx will also have indirect impacts on organisms in Arctic fjords. Regionally, sea ice loss led to increases in primary production averaging 27.5 Tg C yr\(^{-1}\) between 2003 and 2007; during the 2007 ice extent minimum, primary production was 23% higher than the 1998-2002 mean (Arrigo et al. 2008). Around 70% of this increase in production can be attributed to a longer growing season for phytoplankton, when light availability increases in the early spring due to the lack of overlying ice (Arrigo et al. 2008). Even the thinning of sea ice and the proliferation of melt ponds as a result of increased air temperatures can increase under-ice bloom intensity. Arrigo et al. (2012) found that sufficient light is transmitted through thinning ice to allow for earlier and more extensive blooms in nutrient-rich Arctic continental shelf waters, which may have been underestimated by up to 10-fold in satellite-based estimates of regional primary production. As sea ice extent in fjords decreases with warm water influxes, phytoplankton blooms are predicted to occur earlier in the year, and overall primary production should increase.
Warming ocean temperatures in fjords will also increase glacial melting, increasing the suspended sediment in the water column and greatly impacting nearby primary production. At McBride Glacier in Alaska, rapid retreat (0.25 km yr\(^{-1}\), 1984-86) led to the deposition of 6.6 x 10\(^6\) m\(^3\) of sediment over a two-year period, with sediment accumulation rates as high as 13 m yr\(^{-1}\) over 300m from the glacier (Cowan and Powell 1991). Approximately 2/3 of the total volume of sediment was emitted by meltwater plumes originating at the base of the tidewater glacier, and was subsequently deposited by suspension settling (Cowan and Powell 1991). Despite sediment flocculation close to the glacier, fine particles stay suspended even several kilometers from the meltwater plumes (Cowan and Powell 1991). Because increases in light attenuation are associated with high turbidity in fjords, there is a potential for dramatically limited primary production close to rapidly retreating glaciers.

**Glacial Terminology**

Glaciers can be described and classified in several different manners. Land- or terrestrially-terminating glaciers accumulate snow at high latitudes, and ablate at lower elevations due to higher air temperatures. Meltwater that pools on the surface is transported to the base of the glacier through crevasses or moulins. Once at the base of the glacier, meltwater moves through basal channels to the terminus, where it deposits sediment in moraines and travels in braided rivers to nearby bodies of water. In contrast, tidewater glaciers are marine terminating; they ablate partially due to air temperature but primarily at their terminus, where they contact ocean water. Submarine melting and calving (the process by which glacial ice breaks off and forms icebergs) are the primary causes of this ablation, making tidewater glaciers susceptible to rapid retreat when ocean waters warm. Basal meltwater encounters ocean water at the grounding line, or terminus of the glacier, upwelling to the surface due to its low density. Sediment is transported with this meltwater, and is deposited primarily through suspension settling, greatly affecting nearby marine life. A large areal portion of the glaciers in Greenland, Antarctica, eastern Canada, Southeast Alaska, and Patagonia are tidewater glaciers.

Glaciers can be further subdivided by their thermal structure and their velocity characteristics. Cold-based glaciers are frozen to the underlying bedrock, as they are
below freezing at the ice-ground interface. This prevents basal meltwater from accumulating, slowing glacial velocity and preventing significant erosion. Warm-based glaciers are at or above freezing at the interface, allowing for the accumulation of meltwater at the base, rapid bedrock erosion, and greater potential glacier velocities. Polythermal glaciers either seasonally or partially exhibit both cold- and warm-based
glacial characteristics. Additionally, surge glaciers are found in Svalbard, Canadian Arctic islands, Alaska, and Iceland. These glaciers experience large surging events, where velocities will suddenly increase and there will be significant glacial advance. Although the causes behind the surging events are not yet fully understood, a model developed by Jiskoot et al. (2000) indicates that Svalbard’s surge glaciers tend to have polythermal regimes and are found on fine-grained deformable beds.

**Research Questions**

This study evaluates the impact of warming Atlantic water intrusions and the increased glacial melting that this causes on both the oceanography and the primary production near the Kronebreen/Kongsvegen glacial complex in Kongsfjorden. These changes are assessed through two primary questions: 1) How has the intrusion of warm Atlantic water affected Kongsfjorden oceanography and water mass formation? and 2) How are seasonal increases in glacial melting affecting near-ice face primary production?

Currently, there is little scientific knowledge about production levels close to the ice face of tidewater-terminating glaciers (Hop et al. 2002), and no comprehensive study that links changes in water temperature to sea ice coverage, glacier dynamics, and phytoplankton productivity. Primary production varies across Kongsfjorden; some areas influenced by tidewater-terminating glaciers are extremely light-limited due to high turbidity, whereas other areas proximal to terrestrial-terminating glaciers are not light-limited due to low sediment load in the water. Thus, using different regions of the same fjord as a case study for warm water intrusion, it is possible to predict how primary production will most likely respond as glaciers retreat. By comparing the data collected as a part of this study to satellite data and the broader scientific literature, a model for understanding the impacts of warm water influx on fjord ecosystems is proposed. GIS analysis of the fjords of West Spitsbergen outlines which fjords will most likely be impacted by warm water intrusions in future years, and how that warm water intrusion will affect primary production. The underlying goal of this model is to provide a broad prediction for the ways that warm water intrusions will impact primary production across the Arctic, which will allow for greater understanding for its potential ramifications on upper trophic levels of the food web.
Study Region

The Physical Environment of Kongsfjorden

Kongsfjorden, located in the Norwegian Arctic territory of Svalbard (Figure 3), is considered representative of most other open fjords in the Arctic due to its status as a well-established international scientific station, its lack of a sill (a bathymetric rise in the bedrock at the mouth of a fjord), and the strong Atlantic and glacial influences on its water masses (Hop et al. 2002). Located at 79°N, the 20 km-long fjord runs from southeast to northwest on the western coast of Spitsbergen (Svendsen et al. 2002). With warmer, saltier water masses entering the system from West Spitsbergen Current to the west, glacial inputs of fresh, cold, turbid water from the east, and a gradient of Arctic to boreal conditions in between, the fjord allows for the study of the interaction between Arctic- and Atlantic-type biotic communities (Svendsen et al. 2002, Hop et al. 2002).

Blomstrandbreen, Conwaybreen, Kongsbreen, Kronebreen, and Kongsvegen are surging tidewater glaciers that terminate into the northern and eastern extents of Kongsfjorden (Figure 4; Svendsen et al. 2002). The tidewater glaciers are polythermal, acting largely as cold-based glaciers in the winter and warm-based ones in the summer.
They produce ~1.4 km³ of cold, fresh meltwater each year, primarily during the summer melt season (Svendsen et al. 2002). The interaction between these glaciers and the ocean waters entering the fjord from the southeast is the primary factor in the structuring of the fjord’s physical oceanography. The fjord also has several terrestrially bound valley and cirque glaciers to the south, referred to as the Lovénbreane and the Brøggerbreane glacial complexes. Although these glaciers provide a small input of fresh, cold meltwater and sediment to the fjord, the impact of the tidewater glaciers in the inner fjord is far greater (Svendsen et al. 2002).

**Figure 4.** SvalbardKartet map of Kongsfjorden (Norwegian Polar Institute, 2015). Black text indicates the 5 marine-terminating glaciers in the fjord, while white text indicates the major complexes of terrestrially-terminating glaciers. The three glaciers measured in this study are bolded. Kongsfjorden, Lovénøyane islands, and Colletthøgda, significant because they mark the boundary of many transects used in this study, are marked in blue italics. Orange box demarcates the 2014 study site.
Figure 5. TopoSvalbard 2009 high-resolution image of Kronebreen glacier (Norwegian Polar Institute, 2015), with the consistent sediment plume in 2009 to the NE of the glacier. Inset shows a TopoSvalbard map of Kongsfjorden (Norwegian Polar Institute, 2015), and gold box highlights the region pictured.

The three glaciers measured in this study are Kronebreen/Kongsvegen and Kongsbreen. Kronebreen/Kongsvegen, referred to simply as Kronebreen due to the low modern influence of Kongsvegen on the dynamics of the ice front, has been the most active glacier in Kongsfjorden throughout recorded history. The glacier has a mean annual velocity in the center of the front of -2 m d$^{-1}$ and a peak of -4.5 m d$^{-1}$, causing the glacier to flow an estimated 800 m each year (Lefauconnier et al. 1994). However, starting in the late 2000s, Kronebreen began retreating even more rapidly, making it the fastest-retreating glacier in Svalbard (Brigham-Grette, pers. comm. Summer of 2014). The glacier also has higher sedimentation rates than other nearby glaciers, resulting in high light attenuation. This reduces the euphotic zone to depths as low as 0.3 m near the
glacier (Keck et al. 1999), thereby reducing the growth rate and survivability of nearby phytoplankton (Eilertsen et al. 1989). Basal meltwater, draining from channels under the glacier, upwells at the glacier-ocean interface and brings suspended sediment to the surface in seasonal meltwater plumes.

Directly north of Kronebreen, the southern portion of Kongsbreen has retreated onto bedrock since 2009 (Brigham-Grette, pers. comm.), and basal water upwelling is low to nonexistent. At the northern extent of Kronebreen, an intermittent sediment plume was observed in the summer of 2014 where a strong upwelling plume had predominated from 2009 and 2011 (Figure 5; Rajagopalan 2012, Brigham-Grette, pers. comm. summer of 2014). The southern extent of the Kronebreen ice face featured a highly active sediment plume in the summer of 2014. Although over the last decade there hasn’t been much plume activity to the south of Kronebreen (Brigham-Grette, pers. comm.), during a period of rapid Kongsvegen retreat in the 1980s, the southern portion also exhibited high plume activity. An estimated volume of $4.5 \times 10^6$ m$^3$ of sediment accumulated in the surrounding basin, forming a delta that extended 180 m in a period of 8 years (Lefauconnier et al. 1994).

The Oceanographic Context of Kongsfjorden

Hop et al. (2002) describe three primary water masses in Kongsfjorden: cool and fresh Surface Water (SW), warm and salty Transformed Atlantic Water (TAW; $T>1^\circ C$, $S>34.7$ psu), and the cold, salty Winter Cooled Water (WCW; $T<-0.5^\circ C$, $S>34.4$ psu). TAW is formed through the mixture of Atlantic Water (AW; $T>3^\circ C$, $S>34.9$ psu) and Arctic Water (ArW; $T=0.5-2^\circ C$, $S=34.7-34.9$ psu; Hopkins 1991), causing its characteristics to vary from year to year, and is water transported by the West Spitsbergen Current (WSC). North Atlantic Oscillation variability appears to account for annual fluctuations in the strength and characteristics of the TAW flow into the fjord (Hurrell 1995, Hop et al. 2002), but TAW intrusion is also partially influenced by sea ice extent in Nordic seas (Vinje 2001). When summer intruding TAW temperatures are high, there is an increase in glacial meltwater. Melting icebergs and basal meltwater draining from channels located under the glacier form the cold, fresh Surface Water (SW) layer, creating a strong halocline in the first few meters of the fjord and preventing deep-water
mixing by surface currents in the summertime (Figure 6; Svendsen et al. 2002). This SW layer is also characterized by high levels of suspended particles associated with the sedimentation outputs of glaciers (Beszczynska-Møller et al. 1997). Water masses proximal to tidewater glaciers experience high levels of mixing, and multiple meter-scale

Summer

![Diagram of water masses in summer](image)

![Diagram of water masses in winter](image)

**Figure 6.** Schematic of water masses and approximate locations in Kongsfjorden in the summer (above) and winter (below), based on the characterization by Hop et al (2002). The influx of warm TAW in the summer causes the Kronebreen glacier to melt, producing cold and fresh SW. In the winter, TAW cools and extrudes salt during sea ice production, forming salty, cold WCW.
turbulent eddies are found proximal to Kongsbreen and Kronebreen, with a strong surface current (>1 m/s) along the glacial front (Svendsen et al. 2002). In the absence of light in the winter, the TAW cools significantly, and sea ice forms. The cold, salty water extruded during this process forms Winter Cooled Water (WCW), the bottom water found at depth, where it is bathymetrically constrained (Figure 6; Svendsen et al. 2002, Rajagopalan 2012).

The tide travels as a Kelvin wave along the west Spitsbergen coast (Gjevik and Straume 1989), and inside Kongsfjorden responds to the tidal elevation of the ocean surface outside the fjord, where the amplitude is about 0.5 m (Svendsen et al. 2002). The dominant component of the tide is the semi-diurnal lunar component, but there is a solar component that constitutes 50% of the lunar component, causing a strong spring-neap periodicity (Svendsen et al. 2002). However, tidal height variation was minimal for the duration of this study, with less than 1 m of change observed for all casts.

Kongsfjorden conditions changed significantly in the early 21st century with an influx in anomalously warm TAW into the fjord at a depth of 15-100m (Willis et al. 2008). Beginning in the early 2000s, TAW began to gradually warm (Svendsen et al. 2002), but in 2006 anomalously warm water in caused a permanent shift to sea ice-free winters and increased primary production in the spring and summer months. These shifts in conditions resulted in changes to phytoplankton species type and distribution throughout the fjord (Willis et al. 2008), and are the focal point of this study.
Methods

Data Collection

From August 4 to August 15, 2014, data on conductivity, temperature, and density (CTD), optical backscattering (OBS), and chlorophyll fluorescence were collected in Kongsfjorden, Spitsbergen. A SAIV A/S CTD/STD model SD204 (manufactured in Tromsø, Norway) with additional SAIV sensors measuring chlorophyll fluorescence and backscatter was used. Data were processed with SD200W software and Microsoft Excel. CTD casts were subsequently analyzed using MATLAB software (MathWorks) as well as TopoSvalbard (Norwegian Polar Institute) to map casts.

Figure 7. TopoSvalbard 2009 (Norwegian Polar Institute, 2015) map of study region, with casts marked in red. The green box indicates low-sediment transect, while the white box marks the intermittent sediment plume area and the brown box marks the high sediment plume region. Inset shows a TopoSvalbard map of Kongsfjorden (Norwegian Polar Institute, 2015), and gold box highlights the region pictured.
Over the 11 days of study, 53 casts were taken (Figure 7) along 5 distinct transects (Table 1). Transect A ran the length of the major plume, starting about 1 km away from the southern ice face and continuing to the Lovénøyane Islands on a day of no plume activity from the intermittent area. Transect B started at the far north terminus of the Kongsbreen glacier in the low-sediment region, running east along the ice face for the first 3 measurements before aiming southwest and running out to the Lovénøyane Islands. It was taken during a time of high sediment plume activity at the intermittent plume, and crosses the plume at about 2 km from the start. Transect C began at the northernmost part of the ice face of Kronebreen glacier and continued west towards the Lovénøyane Islands. This transect was composed of 7 casts taken 250 m apart and of 6 casts taken ~1 km apart, and it was taken during a time of high intermittent plume sedimentation. Transect D curves across the fjord from the southwest to the northeast during a time of no intermittent plume sedimentation activity. Transect E cuts across the fjord from south to north about 7 km from the ice face of Kronebreen glacier, and was taken at a time with no intermittent plume activity.

Table 1. Transect name, date, collection time, and details of cast numbers, names, and distances between the casts. The * symbol indicates a day of low output from the intermittent sediment plume, while the † symbol indicates extremely high output from the intermittent plume.

<table>
<thead>
<tr>
<th>Transect Name</th>
<th>Sediment Load</th>
<th>Date</th>
<th>Time</th>
<th># Casts</th>
<th>Distance between Casts (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transect A</td>
<td>High</td>
<td>Aug. 6</td>
<td>16:31-18:23</td>
<td>7</td>
<td>1.16-1.94</td>
</tr>
<tr>
<td>Transect B</td>
<td>Intermittent</td>
<td>Aug. 7</td>
<td>14:39-16:22</td>
<td>9</td>
<td>0.39-1.84</td>
</tr>
<tr>
<td>Transect C</td>
<td>Intermittent†</td>
<td>Aug 9</td>
<td>12:45-17:08</td>
<td>13</td>
<td>0.22-1.21</td>
</tr>
<tr>
<td>Transect D</td>
<td>Intermittent*</td>
<td>Aug. 11</td>
<td>10:06-12:57</td>
<td>6</td>
<td>0.35-0.91</td>
</tr>
<tr>
<td>Transect E</td>
<td>Low</td>
<td>Aug. 14</td>
<td>09:48-10:54</td>
<td>5</td>
<td>1.18-1.55</td>
</tr>
</tbody>
</table>
Modeling the Euphotic Zone

Solar radiation decays exponentially with depth, and can be represented by the following equation:

$$E_z = E_0 e^{-kz} \tag{Equation 1}$$

Solar radiation attenuated in water due to both absorption and scattering of photons. The absorption coefficient is determined by the optical properties of water itself and by its dissolved and particulate organic matter content, including phytoplankton. Although scattering doesn’t remove light, it redirects the photon path, thereby reducing the vertical penetration of photons. The scattering coefficient is determined by the molecular properties of seawater, and, more significantly, by the characteristics of the suspended organic and inorganic particles. In the waters studied here, suspended sediment and phytoplankton concentration are the major contributors to particle scattering.

The euphotic zone, or the region in which phytoplankton have adequate light to photosynthesize, is generally approximated as extending from the water surface to the depth at which the photosynthetically available radiation (PAR; essentially the visible wavelength range, 400 nm to 700 nm) is attenuated to 1% surface levels. By rearranging Equation 1 and setting $E(z)/E(0)$ as equal to 1%, the following equation approximates the euphotic depth:

$$Z = \frac{\ln(0.01)}{-K} \tag{Equation 2}$$

where $Z$ is the depth of the euphotic zone (units, m), $K$ is the diffuse attenuation coefficient for irradiance (units, m$^{-1}$), and the 0.01 represents the 1 percent light level.

Using a Monte Carlo simulation (in which the path of individual photons are tracked in a theoretical ocean of defined optical properties), Kirk (1994) quantifies the attenuation coefficient as a function of the absorption ($a$) and scattering ($b$) coefficients in the following manner:

$$K = (a^2 + 0.256a \cdot b)^{\frac{1}{2}} \tag{Equation 3}$$

which demonstrates that absorption is typically far more important in determining $K$ than is scattering when they are of comparable magnitudes.

Although photosynthetically available radiation, absorption, and backscattering were not directly measured in this study, the euphotic depth can be modeled via well-
established constituent-specific optical properties. This requires known profiles of optical constituents (in this case, chlorophyll fluorescence and turbidity) as well as published optical specific biogeochemical properties (Babin et al. 2003a).

In this study, the absorption coefficient \(a, (m^{-1})\) is approximated by the sum of the contributions to absorption by seawater and by phytoplankton, as there is minimal dissolved or particulate organic material in Arctic waters due to sparse land vegetation (Svendsen et al. 2002) and glacial sediments do not contribute substantially to absorption (Babin et al. 2003b; Stramski et al. 2007). Because this study investigates the attenuation of spectrally integrated irradiance in the visible wavelength range, the absorption profile is approximated by:

\[
a(z) = a_{\text{seawater}} + F_{\text{chl}} * 0.018 \\
\]

(Equation 4)

The spectrally integrated absorption by seawater \(a_{\text{seawater}}\) is given by the constant 0.2 \(m^{-1}\) (Smith and Baker 1981) and \(F_{\text{chl}}\), the depth-dependent value of the chlorophyll concentration, was measured with a calibrated profiling fluorometer (units, mg/m^3). The term 0.018 (units m^2/mg chl) is the spectrally averaged value of the chlorophyll-specific absorption coefficient (Bricaud et al. 1995; Roesler and Barnard 2013).

The scattering coefficient \(b, (m^{-1})\) in Equation 2 is approximated by the sum of the contributions to scattering by seawater and glacial sediments, as phytoplankton are relatively inefficient scatterers relative to suspended sediments (Babin et al. 2003a). The scattering profile is approximated by:

\[
b(z) = b_{\text{seawater}} + T_{\text{sed}(z)} * \frac{0.00408}{0.03} \\
\]

(Equation 5)

The spectrally integrated scattering by seawater \(b_{\text{seawater}}\) is given by the constant 0.007m\(^{-1}\) (Morel and Morgan 1972). The second term represents the particle scattering coefficient, where \(T_{\text{sed}(z)}\) is the depth-dependent turbidity obtained by a calibrated turbidity sensor (measured in Formazin Turbidity Units, or FTU). The constant 0.00408 is the formazin-specific backscattering coefficient (m\(^{-1}\)/FTU), which converts calibrated turbidity to the backscattering coefficient (units m\(^{-1}\); WETLabs turbidity/backscattering sensor manual), and 0.03 is the particle backscattering ratio, or the fraction of total particle scattering that is in the backwards direction (Twardowski et al. 2001). The backscattering ratio varies from about 0.5% for phytoplankton, to 1% for detrital organic particles, to 3% for
inorganic sediments. The dominance of the particle field in these waters by glacial sediments is the reason for the selection of the relatively high backscattering ratio was used for euphotic depth calculations in this study.

**Giovanni Satellite Data**

Satellite analyses and visualizations used in this study were produced with the Giovanni Ocean Color Radiometry Online Visualization and Analysis data system, developed and maintained by the NASA Goddard Earth Sciences Data and Information Services Center (GES DISC). The 4-km MODIS-Aqua satellite sensor observations were used to analyze changes to the diffuse attenuation coefficient at 490 nm, sea surface temperature (using the daytime 11 micron channel observations), and chlorophyll a concentrations between 2003 and 2014. The GSM inversion results obtained at 9km resolution were used to analyze changes to particulate backscattering coefficients between 2003 and 2010. The monthly averaged timeseries were spatially averaged over the area from 78° to 80°N and 10° to 13°E (Figure 8). Data was downloaded on April 13, 2015 and analyzed using Microsoft Excel 2011 and MATLAB (MathWorks).

*Figure 8.* Wikimedia Commons bathymetric map of Svalbard, showing the extent of the region analyzed spatially with GIS. The red box represents region analyzed for regional satellite data.
GIS Analysis

The likelihood of warm water intrusion and the predicted effect on primary production in Spitsbergen’s coastal waters was spatially analyzed using ArcMap version 10.2 software. To determine areas most likely to experience warm water intrusion, the International Bathymetric Chart of the Arctic Ocean (Version 3.0, published 2012) was compared to regional sea surface temperature measurements from January to March of 2015. This data was obtained from 4-km MODIS-Aqua satellite sensor observations (using the daytime 11 micron channel observation) recorded on the Giovanni Ocean Color Radiometry Online Visualization and Analysis data system, developed and maintained by NASA’s GES DISC. The time-averaged latitude-longitude map was spatially determined over the area from 75˚ to 82˚N and 7˚ to 30˚E (Figure 8). Areas shallower than 25m were considered unable to experience warm water intrusion based on the depth of peak temperatures observed as a part of this study, and raster analysis allowed these areas to be excluded from consideration. Additionally, coastal areas within 25km of waters warmer than 0˚C during the winter of 2015 were considered most likely to experience warming in the near future. These regions were identified by georeferencing a MODIS-Aqua satellite time-averaged latitude-longitude map (NASA’s GES DISC), digitizing along observed temperature differences, and extracting areas that fell within a 25km buffer from the coastline to include only regions that intersected warmer water. Areas intersecting warmer waters and the corresponding coastal waters were digitized to provide regions most likely to experience warming.

In order to predict how the region would fare with increased warm water intrusion, sea ice coverage was digitized from the same MODIS-Aqua satellite time-averaged latitude-longitude map (NASA’s GES DISC). Areas with sea ice coverage with a high likelihood of experiencing warm water intrusions were digitized. Glaciers were visually inspected and then classified as marine-terminating, terrestrially-terminating, or non-Spitsbergen, and a 5km buffer was created around the terminus of every marine-terminating glacier. Areas with a high likelihood of experiencing warm water intrusion were then digitized, as were all near-coastal waters likely to experience warm-water intrusion.
Results

Water Masses

Three major water masses were observed in Kongsfjorden in August 2014: a cool, fresh surface water (T = 1-5°C, S < 34.7 psu); a warmer, salty water at depth (T = 5.5-7°C, S > 34.7 psu); and a cold, salty bottom water (T < 1.5°C, S > 34.9 psu; Figure 9). Svendsen et al (2002) identified these water masses as surface water (SW), Transformed Atlantic Water (TAW), and Winter Cooled Water (WCW), respectively, based on characteristics first noted by Hopkins (1991).

Figure 9. T-S diagram for all casts, color-coded by depth. Inset shows full range of observed salinities; larger graph shows salinity values from 32 to 35.5 ppt. Black circles indicate water masses measured in 2014.
The Euphotic Zone

Variability in the absorption and backscattering of light through the water column greatly affected the depth of the euphotic zone across Kongsfjorden (Figure 9). Phytoplankton absorption varied from 0.1 to 7 mg/L, with maxima typically occurring in the upper 10m (Figure 10). Turbidity varied from 0 to 135 FTU, with maxima typically occurring at ~5m depth and between 0 and 15m. Both $K_{\text{phytoplankton}}$ and $K_{\text{sediment}}$ impacted total K values, but light attenuation was primarily influenced by turbidity values. 1% PAR values, which indicate the depth of the euphotic zone, occurred between 4 and 21m.

Figure 10. Variation across all casts in A) $K_{\text{phytoplankton}}$, B) $K_{\text{sediments}}$, and C) Total K, and D) % PAR as a function of depth from 0 to 30 m.

Hydrographic Transects

Measurements from this study indicate that, at the height of the summer melt season, the upper several meters of Kongsfjorden experienced a strong pycnocline separating fresh surface water from saltier intermediate waters (Figure 11). There also seemed to be a slight increase in density below 40m due to the cooler waters at depth. In Transect A, taken on August 5\textsuperscript{th} at the persistent sediment plume, temperature and chlorophyll concentrations were low, while turbidity was high proximal to the glacier.
Further from the glacier, turbidity decreased and temperatures increased (Figure 11). A sharp increase in chlorophyll occurred at a depth of about 10-20m at the base of the euphotic zone.

**Figure 11.** Interpolated contour plots of Transect A, with color bars representing temperature (A), salinity (B), density (C), turbidity (D), and chlorophyll (E). The white line in subplot E shows the modeled euphotic zone depth. Profiles show changes associated with distance from high sediment plume activity. TopoSvalbard inset (right) indicates location of transect.
Transect B, taken on August 6th on a day of moderate intermittent plume flow preceded by several days of no output, was fairly warm at intermediate depth of 25m but had the lowest temperatures of any profile in the inner fjord at depth (Figure 12). Despite high turbidity values near Kronebreen on a day of moderate output, chlorophyll was found at a depth of ~5m between two distinct turbidity layers but well above the euphotic zone. Chlorophyll was found in much higher concentrations farther away from the glacier, where the euphotic zone deepens.

Figure 12. Interpolated contour plots of Transect B, with color bars representing temperature (A), turbidity (B), and chlorophyll (C). The white line in subplot C shows the modeled euphotic zone depth. The white boxes (above and right) indicate the intermittent plume; the black boxes (above and right) indicate the consistent plume. Profiles show characteristics associated with intermittent sediment plumes on a moderate-output day after 5 inactive days. TopoSvalbard inset (right) indicates location of transect.
Taken on August 9th on a day of high intermittent plume flow, Transect C had the warmest water at intermediate depth (25m) with around 1 m of cold surface water (Figure 13). With turbidity values exceeding 100 FTU through much of the sediment plume, this transect had high suspended sediment in the water column up to 4 km from the glacier face. The euphotic zone shoaled to 8m in the plume, but reached a depth of 21 m at the far extent of the transect, allowing significant chlorophyll concentrations from 0-10 m and at about 20 m at this far extent.
In Transect D, taken on August 11th on a day of low intermittent plume activity, temperatures were fairly high and chlorophyll fairly low throughout, while turbidity was high close to the consistent sediment plume, but low close to the intermittent sediment plume (Figure 14). The euphotic zone shoaled upwards in the consistent sediment plume, but deepened with proximity to the intermittent plume. Despite the low sediment output of the intermittent plume and the deep euphotic zone to the east, there was no corresponding rise in chlorophyll concentration.

**Figure 14.** Interpolated contour plots of Transect D, with color bars representing temperature (A), turbidity (B), and chlorophyll (C). The white line in subplot C shows the modeled euphotic zone depth. Profiles show a cross-section of the fjord on a day of high output from the consistent sediment plume (left) but low output from the intermittent plume (right). TopoSvalbard inset (right) indicates location of transect.
Transect E, taken on August 14th and located 7km from the ice face on a day of high sediment plume activity, had higher surface water temperatures and much lower turbidity values than any of the other transects (Figure 15). Because of the low turbidity, the euphotic depth remained consistently deep throughout the transect, and chlorophyll concentrations were high throughout, with values as high as 7 mg/L.

Figure 15. Interpolated contour plots of Transect E, with color bars representing temperature (A), turbidity (B), and chlorophyll (C). The black line in subplot C shows the modeled euphotic zone depth. Profiles show conditions associated with the retreat of proximal glaciers onto land. TopoSvalbard inset (right) indicates location of transect.
Factors Influencing Phytoplankton Distribution

A direct comparison of chlorophyll and turbidity indicates that high chlorophyll values are only found above 20m and exclusively in low sediment conditions, whereas low chlorophyll is found across a variety of turbidity levels and light (Figure 16). Low turbidity is sufficiently correlated to high chlorophyll, but little light attenuation was necessary for high chlorophyll values to occur.

Figure 16. Chlorophyll vs. turbidity for all casts, color-coded by A) depth, and B) percent of photosynthetically available radiation (PAR). Insets are constrained to chlorophyll concentrations between 0 and 1 mg/L and turbidity values between 0 and 20 FTU.
**Satellite Data**

A comparison of annual SST timeseries from 2003 to 2014 reveals that 2006 had the highest temperatures and earliest maximal temperature of any of the annual time series (Figure 17a). Winter temperatures were between -2 and 1.5°C, while summer temperatures were between 3.5 and 5.5°C, with the exception of the summer of 2014. There was a significant increase in wintertime surface temperatures over this time interval, with February temperatures almost 3.5°C warmer in 2014 than in 2003. However, temperature increases vary significantly throughout the series, and the temperature climatology indicates large standard error throughout the year (Figure 17b).

**Figure 17.** MODIS-Aqua 4km monthly timeseries (left panels) and monthly climatologies (right panels) for years 2003-2014 for (A, B) sea surface temperature (°C), (C, D) chlorophyll a concentration (mg/m³), and (E, F) the particulate backscattering coefficient (1/m). The monthly timeseries was spatially averaged over the area from 78° to 80°N and 10° to 13° E. Error bars indicate standard deviation.
Chlorophyll concentrations in surface waters varied from about 0.2 to 1.0 throughout the timeseries, and were significantly higher in April of 2006 (1.8 mg/m³) than at any other time from 2003 to 2014 (Figure 17c). More striking, however, is the change in the seasonal pattern of chlorophyll from peaks in the early and late season (April-May and July-August) to a single late season peak. Chlorophyll concentrations show indications of decreasing in April through this time interval, with the 3 of the 4 lowest April chlorophyll concentrations occurring from 2012-2014. Trends indicate that this region is shifting from two distinct peaks of chlorophyll in May and July-August (pre-2012) to only one late-season peak of chlorophyll since 2012, occurring in August, a change which is reflected in the variability of the climatology (Figure 17d).

From 2003 to 2010 the backscattering coefficient appears variable, but with a trend towards increased May and June particulate backscattering in recent years (Figure 17e). There is a consistent peak in particulate backscattering in July and August, at the height of the glacial melt season (Figure 17f), although the magnitude of this peak is variable from year to year. There is no data available for 2011 to 2014.

Satellite-derived SST anomalies between 2003 and 2014 indicate that there has been a 1°C temperature increase over the 12 years (Figure 18a). On top of the observed linear trend, a cyclic pattern has been observed, with cool temperature anomalies between 2003 and 2005 and 2009 and 2011 and warm temperature anomalies 2006 to 2008 and 2012 to 2014. The 5th order polynomial approximating the data between 2003 and 2014 indicates that this cyclicity has a period of approximately 4-6 years, suggesting that climate oscillations in this region may mask some longer-term trends. Satellite data additionally indicates that 2014 was a year of anomalously cool summertime sea surface temperatures in the region (Figure 18a).

From the chlorophyll a concentration anomaly, it is clear that 2006 was a year of abnormally high chlorophyll levels and an early-season maximal bloom (Figure 18b). Most of the years after 2006 experienced anomalously low chlorophyll concentrations in April, with gradual peaks towards the end of the summer season.

The shortened record of the particulate backscattering coefficient makes it difficult to gauge how suspended sediment has varied through this time interval and how it compares to measurements from the summer of 2014. There appears to be a slight
Figure 18. MODIS-Aqua 4km satellite anomalies of A) SST, B) chlorophyll a and C) particulate backscattering from 2003 to 2014. The monthly timeseries was spatially averaged over the area from 78° to 80°N and 10° to 13° E. In A, the red line indicates the overall warming trend during this period, while the dashed black line indicates a 5th order polynomial approximating the trend.

cyclicity to this data, with anomalously high particulate backscattering from 2003 to 2006 and anomalously low particulate backscattering from 2007 to 2010 (Figure 18c).

The derived diffuse attenuation coefficient appears to be more closely correlated to chlorophyll a concentration than to the particulate backscattering coefficient (Figure 19). The diffuse attenuation coefficient, which is comparable to Total K in Figure 9, ranges from 0.05 to 0.15, instead of the values up to 1 observed in surface measurements.
From 2003 to 2014, there appears to be significant interannual variation, there does appear to be a slight increase in mid-season light attenuation (Figure 19).

Figure 19. MODIS-Aqua 4km satellite timeseries of monthly variation during the summer months from 2003 to 2014 in the diffuse attenuation coefficient at 490 nm. The monthly averaged timeseries was spatially averaged over the area from 78° to 80°N and 10° to 13°E.

GIS Analysis of warm water intrusion vulnerability

Vulnerability to warm water intrusion was considered to be dependent on nearby bathymetry and proximity to warm waters. Based upon the depth of warm water in the hydrographic transects, areas shallower than 25m were excluded from analysis (Figure 20a), as were areas more than 25km from waters warmer than 0°C in the winter of 2015 (Figure 20b). Analysis of coastal sea floor depth and proximity to the warm water of the West Spitsbergen Current indicates that the west coast of Spitsbergen is the most likely part of the Svalbard archipelago to experience warm water intrusion (Figure 20c).
Figure 20. GIS analysis of the coastal areas of Svalbard that A) have bathymetry shallower than 25m, B) have winter 2015 sea surface temperatures greater than 0°C, and C) are thus most likely to experience warm water intrusion. Red color indicates areas greater than 0°C in the winter of 2015 within 25km of the coast. Red crosshatching indicates areas within 25km of warm waters that are deeper than 25m and are thus most likely to experience warm water intrusion.
Discussion

Water Temperature and Water Mass Structure in Kongsfjorden

The results of this study confirm that there are three primary water masses found in Kongsfjorden: a cool, fresh surface layer formed by glacial meltwater; a salty, warm intermediate layer; and a cold, salty bottom layer bathymetrically constrained to the inner fjord. However, the characteristics of these water masses differ from those outlined by Hopkins (1991) and Svendsen et al. (2002). As such, Figure 21 illustrates the proposed changes to the water masses in the fjord as a result of warm water intrusion since the early 2000s.

In 1996 and 1997, Transformed Atlantic Water temperatures were no higher than 1.5°C, while from 1998 to 1999, the water mass was consistently warmer than 2°C during their summer field season (Svendsen et al. 2002). More recent studies in Kongsfjorden found significant warming trends: in 2006, anomalously warm water influxed into the fjord, raising the temperatures to 3.5°C (Willis et al. 2009; Cottier et al. 2007). Between 2005 and 2011, TAW temperatures in front of the Kronebreen glacier increased to about 4°C (Rajagopalan, 2012). The additional 1.5-3°C temperature increase observed during the 2014 field season indicates further TAW warming. This warming may indicate that the Atlantic Water in the Western Spitsbergen Current is no longer mixing substantially with Arctic Water (ArW), or that Arctic Water and/or Atlantic Water has warmed significantly over the past few decades. AW has a higher salinity than ArW, but measurements of TAW indicate that its salinity has stayed fairly constant throughout time, implying that there has been little change in mixing patterns between the two water masses. This consistent salinity indicates that, instead of warming occurring through changes to mixing patterns, the substantial warming of one or both external water masses has occurred, causing the warming of TAW in the fjord.

Satellite ocean color imagery estimates sea surface temperature within the first optical depth (or the 37% light level) of infrared radiation, which corresponds to the top few millimeters of water. Because of this, satellite data reflects the ocean skin temperature, and cannot be directly compared to subsurface temperatures found in the summer 2014 profiles. However, the time series observations do place the summer of
2014 in a temporal context, allowing for an assessment of how the 2014 field season compared to previous years.

In Kongsfjorden satellite measurements of temperature most closely reflects changes to SW in the summertime and TAW in the wintertime, preventing an assessment of how temperatures are changing in WCW or at depth in the water column. Satellite SST indicates that regional temperatures have warmed around 1°C through this time interval. Summer surface water temperatures show little change, most likely because of the increases in glacial melting since 2006, leading to the dominance of cool, fresh surface

**Figure 21.** Schematic of proposed changes to water masses and approximate locations in Kongsfjorden in the summer (top) and winter (bottom), before warm-water intrusion (left) and after warm-water intrusion (right). Increased temperatures of TAW has led to a summertime increase in volume of SW and greater surface stratification, causing mixing to occur later in the season. Warmer waters prevent sea ice formation in the winter, changing the characteristics and concentration of WCW.
waters. Instead, most of this 1°C warming has occurred in the winter months, with temperatures from 2006-2014 averaged 2°C warmer than between 2003 and 2005.

Satellite data from 2014 indicates that the summer may have been significantly cooler than previous years. Although this could imply that the conditions observed during the 2014 field season were different from the climatology in the region, it is more likely that it demonstrates that glacial melting is occurring earlier in the season and that surface meltwater represents a larger part of the SST signal in the region than in previous years. Observations of water mass temperatures throughout the water column in the summer of 2014 fit with the trends through recent scientific literature, but comparing the 2014 data set to future summer profiles and satellite data would allow for a better evaluation of whether or not the season was anomalous.

Additionally, the 12-year satellite data appears to show a 4 to 6 year cyclicality in SST, with about 2°C variation over this cycle. Studies conducted prior to 2006 indicated that North Atlantic Oscillation variability accounts for annual fluctuations in the strength and characteristics of the TAW flow into the fjord (Hurrell 1995, Hop et al. 2002), so it is likely that this cyclicality is also due to NAO activity. Linking the North Atlantic Oscillation to this satellite trend and better understanding how warm water intrusion is affected by regional climate oscillations could be an interesting area of further research. Although 12 years of data is not sufficient to resolve a 4 to 6 year cycle, continued satellite observations could be useful in teasing out this trend.

As for the WCW, the bottom water in the fjord, it appears there has been an increase in temperature of about 2°C since 1999, and an overall decrease in the quantity and prevalence of WCW in the inner fjord since 2011 (Rajagopalan, 2012). This temperature increase and reduction in quantity of WCW is most likely associated with a combination of changes to stratification patterns and the complete loss of sea ice in the fjord. Increased summer melting due to warmer TAW influx may have led to a strengthening of the pycnocline between the SW and the TAW. Greater stratification between these layers in the summertime would delay the fall/winter water column mixing and allowing TAW to retain heat longer, which could prevent sea ice formation until later in the winter. This theory is supported by the location of WCW in the fjord. The coldest bottom water in the fjord was found in front of the less-active Kongsbreen glacier, which
has almost entirely retreated up onto land in the last decade. Because terrestrially-
terminating glaciers input less meltwater into the fjord, there is less intense surface
stratification in front of this glacier, allowing for a greater degree of mixing and leading
to cooler WCW formation.

In Kongsfjorden, the combination of warm TAW intrusion and increased water
mass stratification appears to have caused a cessation in sea ice production. Despite the
historic persistence of <1m thick fast ice covering the entirety of Kongsfjorden for 5-7
months each year (Gerland and Renner 2007), there was a complete loss of sea ice in
2006, followed by insignificant coverage in recent years. Satellite data supports the
notion that high winter temperatures, which peaked at 1.5˚C in February of 2014 (as
opposed to -2˚C in February of 2003), are preventing sea ice formation. When surface
waters do not cool sufficiently to produce sea ice in the winter, it seems logical that
WCW characteristics would change, becoming less saline and warmer, and that overall
quantities of WCW would be drastically reduced.

The Euphotic Zone

The hydrographic transects collected as a part of this study indicate that proximity
to a rapidly retreating glacier has substantial impacts on primary productivity. Particulate
backscattering, which is correlated to glacial retreat, played a larger role than absorption
by chlorophyll in determining the depth of the euphotic zone. Calculations of the
euphotic zone were made with the assumption that there was minimal contribution of
detrital organic material to the fjord (Svendsen et al. 2002), and that glacial sediments did
not substantially absorb PAR (Babin et al. 2003b; Stramski et al. 2007). If the second of
these assumptions is incorrect, the euphotic zone may be significantly shallower than the
current model indicates.

Consistent sediment plume

In the transects that cut across the consistent sediment plume to the south of
Kronebreen, turbidity is extremely elevated close to the glacier (with values as high as
140 FTU), but reduces with distance as sediments flocculate out of the water column. A
suspended sediment load of this magnitude causes a shoaling of the euphotic zone such
that 1% PAR values occur at a depth of 5-10m in the center of the plumes. Because light attenuation is high within the consistent sediment plume, chlorophyll values are below 0.3 mg/L and primary production is likely insignificant. However, decreases in suspended sediment around 4-5 kilometers from the input of glacial meltwater allows the euphotic zone to increase in depth, allowing for high chlorophyll values and indicating high primary production.

**No sediment plume**

In Transect E (7 km from the Kronebreen glacier), turbidity values are below 12 FTU, an order of magnitude lower than within the sediment plume. Due to the stable and deep euphotic zone, which ranges from 19-21m, chlorophyll concentrations of up to 7 mg/L were measured, showing that the region is capable of extremely high productivity when turbidity levels are low. Although there are nearby land-terminating glaciers inputting small amounts of fresh water and sediment to the fjord, primary productivity appears uninhibited by the meltwater input of terrestrial glaciers.

**Intermittent sediment plume**

Unlike the high-sediment conditions in consistent sediment plumes and the low-sediment conditions found proximal to land-terminating glaciers, the effects of intermittent plumes on primary production are far more challenging to characterize due to their unstable nature. The mechanisms driving the on-and-off nature of intermittent meltwater plumes are unknown, but would be an interesting area for future research. The intermittent plume to the north of Kronebreen sometimes acted in exactly the same manner as the consistent plume to the south; sediment concentrations were high enough to reduce the euphotic zone to less than 10m, preventing phytoplankton growth up to 4km from the face of the glacier. However, on other days, there was little to no meltwater emitted from the intermittent plume, and turbidity values were equivalent to those of Transect E. When this occurred, icebergs built up to the north of the fjord, making measurements difficult. However, select profiles suggest that phytoplankton blooms can occur near these intermittent plumes when there is low plume activity. In Transect B (taken on a day of moderate intermittent plume input after 5 inactive days), two distinct
layers of turbid meltwater were observed, above and below a moderately-sized phytoplankton bloom. Although the phytoplankton are located at about 3-5m and are well above the euphotic zone, this particular bloom is still susceptible to mixing below the 1% PAR depth. Despite extremely high turbidity below 2m of depth, several small pockets of high chlorophyll concentration appear at depths of less than 1m, approximately 1km from the face of the glacier. In this case, the turbid meltwater appears to have subducted under the water containing the phytoplankton. In Transect D (taken after several days of consistent high activity from the intermittent plume but on a day of no activity), phytoplankton levels remain at background levels below 0.4 mg/L, despite a euphotic zone of around 20m.

Phytoplankton populations in an intermittent plume are subject to inhospitably variable conditions. While a cessation of plume activity may allow for primary production, increased meltwater input can lead to a sudden shoaling of the euphotic zone, causing stable populations to subduct under the euphotic depth or become confined to the surface waters. Fresh, low-density water upwelling from the base of the glacier enhances mixing proximal to the ice front, making even populations located well above the euphotic depth susceptible to mortality. In addition to the hazards associated with turbidity increases, phytoplankton in Kongsfjorden surface waters and throughout the Arctic are exposed to high levels of UV radiation and the cell destruction it causes. Svendsen et al. (2002) observed 1% incident UV-B light penetration at 13 m in Kongsfjorden, which Hop et al. (2002) noted could have a strong negative impact on marine organisms and would especially act as a stressor for primary producers. Damage from UV radiation prevents phytoplankton from photosynthesizing and producing chlorophyll, which prevents these populations from appearing in surface measurements. In areas of high UV light radiation, such as the polar regions, limited mixing of phytoplankton away from the surface can cause extreme cell damage and death (Neale et al. 1998). In near-glacier areas where the euphotic zone is restricted to the surface 10m and where mixing is limited by the strong stratification of surface waters, phytoplankton populations are unlikely to persist.
Factors Influencing Phytoplankton Distribution

Across all transects, high chlorophyll values were found exclusively in low sediment conditions, but low turbidity only correlated to high chlorophyll when light attenuation was low and % PAR was high. Additionally, observations indicated that 1) the euphotic depth was significantly decreased in the presence of high sediment plume activity produced by Kronebreen glacier’s rapid retreat, 2) large phytoplankton blooms occurred in waters offshore of terrestrially-bound, low sediment-emitting glaciers, and 3) phytoplankton populations could be subducted under high sediment plumes or could exhibit local growth between plumes in intermittent plume conditions. This implied that phytoplankton could co-occur with high sediment as long as they had sufficient light and were close to the surface.

Satellite Evidence of Changes to Primary Productivity and Glacial Melting

Satellite ocean color imagery estimates chlorophyll concentrations, particulate backscattering, and light attenuation within the first optical depth (or the 37% light level) of visible radiation. In the summer of 2014, the upper optical depth corresponded to depths of 1-4m for each cast, indicating that satellite data cannot be directly compared to the deeper portions of 2014 profiles, but is a useful analysis assuming there is a relationship between surface and deep concentrations. The time series observations place the summer of 2014 in a temporal constant, allowing for an assessment of how the 2014 field season compared to previous years.

Regional satellite measurements of chlorophyll indicate that primary production increased dramatically in 2006, with April chlorophyll a concentrations of 1.8 mg/m³ and remaining high throughout the growing season. This corresponded to the sudden loss of sea ice during the winter of 2006. The early timing of this bloom indicates that as SST increased and prevented sea ice formation, phytoplankton blooms were able to occur earlier in the year. Once light became available in the spring, phytoplankton populations were not light-limited by overlying ice and were able to respond to the increased light instantly. However, since 2006, chlorophyll levels in the region have been low. These low chlorophyll levels despite a lack of sea ice imply that increased suspended sediment in the water column has prevented the high primary production observed in 2006.
There have also been significant changes in the month of max bloom between 2003 and 2014. Between 2003 and 2005, there were two blooms, one in May and a second in August. In 2006, April had the highest chlorophyll concentrations, with a second bloom occurring in July. From 2007 until 2011, chlorophyll concentration remained fairly low, and blooms occurred through April and May and in either July or August. Between 2012 and 2014, chlorophyll concentration appeared to increase linearly throughout the season, with steadily increasing primary production from April until August and the only discernible bloom occurring late in the season. Although the lack of early bloom may indicate an early start to glacial melting, this shift towards later blooms does not appear to be associated with clear causal mechanisms.

Because Kongsfjorden’s phytoplankton populations are limited to near-surface areas (almost exclusively between 0 and 7m), this satellite data may provide a fairly accurate assessment of primary production. However, significant sediment input from glacial meltwater generally occurred deeper within the water column, from 5 to 15m, implying that satellites, with a range of 0-4m in Kongsfjorden, may not adequately show changes in particulate backscattering. There were no significant changes to the particulate backscattering coefficient occurred between 2003 and 2010, despite the fact that glacial melting in the region has increased substantially since 2006. Although there is an expectation that turbidity should have increased during this period, particles may be too deep in the water column to be measured adequately with a satellite. Interestingly, despite the finding that light attenuation was more closely linked to particulate backscatter than to chlorophyll absorption, the satellite-derived attenuation coefficient mirrored changes in chlorophyll concentration but appeared less dependent upon the particulate backscattering coefficient. By measuring changes in backscattering, absorption, and phytoplankton throughout the water column over a longer period of time, a more accurate assessment of changes to suspended sediment levels and glacial meltwater input into the fjord could be obtained.
**Model for warm-water intrusions**

Using the findings of this study, a model has been developed to describe the impacts of warm-water intrusions in Arctic fjords on water characteristics and primary production (Figure 22). When intruding waters in a fjord warm substantially, a sudden loss of sea ice in the fjord occurs, as was observed in 2006 in Kongsfjorden. As sea ice production is generally a critical part of the bottom water formation process, this will most likely impact the water mass dynamics of most fjords, potentially changing the quantity and characteristics of bottom waters, as was the case in Kongsfjorden. When sea ice coverage decreases, phytoplankton face less light limitation by overlying sea ice in the spring, allowing blooms to occur as soon as surface radiation increases. Satellite data from the Kongsfjorden region indicates that substantial increases in SST and increased chlorophyll concentrations both occurred in 2006, thus supporting qualitative observations during this period.

![Figure 21. Model for the impacts of warm-water intrusion on water characteristics, the optical light field, and primary production proposed as a part of this study.](image)

In fjords with tidewater-terminating glaciers, loss of sea ice as a result of warming water will be followed by increased glacial melt and retreat. Joughin et al. (2014) observed this timing at the Jakobshavn Isbrae glacier as a result of warm water influx, although their research did not focus on these changes as part of a larger progression resulting from warmer intruding waters. Increased inputs of glacial meltwater in fjords will cause a strong salinity-driven stratification of the surface waters. This may affect water mass formation by preventing mixing until later in the fall or winter. As shown in
both the consistent and intermittent sediment plumes of Kronebreen, primary production is severely limited when glacial melting is high and retreat is rapid. Although some small-scale primary production may occur near intermittent plumes, increased glacial melting generally corresponds to a dramatic reduction in phytoplankton blooms, as was observed in the satellite record of chlorophyll concentrations from 2007 to 2014.

When warm waters intrude into a fjord and cause increased glacial retreat, many tidewater glaciers ablate into deeper basins, exposing a larger portion of their ice face to the water and leading to further melting. Although some glaciers act as outlets for ice caps (such as those in Greenland and Antarctica) and thus have the potential to retreat for decades to centuries, many glaciers will eventually retreat onto land, becoming terrestrially terminating. When this happens, overall glacial melt rates and the sediment and fresh water input into the fjord will reduce. This would lead to a reduction in the stratification of waters during the summer, which could increase the winter mixing period. Increased mixing would allow bottom water temperatures to gradually decrease once again, as was theorized to have occurred in front of the Kongsbreen glacier. During this stage of warm water intrusion, primary production would increase, with massive phytoplankton blooms occurring earlier in the season than before sea ice coverage was lost. In this study, this stage was observed at transects across from terrestrially-terminating glaciers, where primary production was high enough to act as a significant seasonal sink for carbon dioxide. If these early blooms are large enough, primary production may become macronutrient limited in the late summer, despite having only faced light limitations in the past. Like Kongsfjorden, many Arctic fjords exposed to Atlantic water currently have a gradient of species distributions, with Arctic-type species close to the glaciers and Atlantic-type species near the mouth of the fjord. But because the post-glacial retreat blooms will occur in much warmer waters than previously, Atlantic-type phytoplankton populations will most likely outcompete the Arctic-type species that have previously dominated these fjords. The loss of sea ice and resulting earlier bloom period could potentially cause timing mismatches with local consumers, which may have ramifications for upper-trophic level feeders. Further research into the effects of warm water intrusion on these upper-trophic level species is a critical next step.
Predictions for Svalbard’s Fjords

Spatial analysis of the Svalbard archipelago indicates that the west coast of Spitsbergen is most likely to experience warm water intrusion based on its sharp bathymetry and its proximity to the West Spitsbergen Current. By applying the model to this coast, predictions were made for the future of sea ice coverage and primary production in coastal waters on the west coast (Figure 23).

Although most of the west coast of Spitsbergen had no sea ice by the winter of 2015, several fjords are at risk of losing already limited sea ice coverage, especially Van Keulenfjorden to the south. Large portions of Wijdefjorden, Woodfjorden, and Isfjorden, as well as Van Mijenfjorden in the south, are unlikely to lose sea ice coverage due to their shallow bathymetry. Additionally, sea ice surrounding other islands of the Svalbard archipelago, especially Edgeøya, is likely to continue to form in the wintertime.

In addition to impacting sea ice coverage in the area, warm water intrusion has the potential to cause increased melting of marine-terminating glaciers, thus impacting primary production patterns in the region. GIS analysis of glacier type by area found that marine-terminating glaciers account for 2/3rds of the glaciers on Spitsbergen by area, and with just a few outlets, these marine-terminating glaciers have the potential to drain a large amount of landlocked ice into the ocean. Marine-terminating glacial retreat can dramatically reduce primary production by releasing large amounts of sediment into the water column. In this study, sediment concentrations within 5km of Kronebreen glacier were too high to allow for significant primary production; as such, areas within 5km of marine-terminating glaciers are marked in brown in Figure 23 and are expected to show low primary production in the near future. These areas include the northernmost portions of Spitsbergen, Kongsfjorden, and Hornsund, as well as many glaciers in smaller fjords.

The remaining near-coastal areas are proximal to land-terminating glaciers or have no nearby glaciers, and were thus considered likely to experience high primary production in the near future. However, when warm water intrusion occurs in these areas, new species will most likely become prominent, as most Arctic species are constrained to living in waters with very low temperatures. Because these areas do not have significant winter sea ice coverage, phytoplankton blooms will occur earlier than historically, and there will most likely be two distinct blooms, unlike in areas proximal to marine-
Figure 23. GIS analysis of the coastal areas of Svalbard most likely to experience warm water intrusion, and the predicted impacts of intrusion on primary production and sea ice coverage as outlined by the model in Figure 21. All major fjords are in black, and the four islands that make up the archipelago are bolded.
terminating glaciers. Although nutrient limitation has not historically acted to limit primary production, large enough blooms may deplete nutrients enough to cause nutrient limitation in the future.

Although this prediction serves as a good way to evaluate the potential impacts of warm water intrusion on the Svalbard archipelago, it is very much a first step to understanding how this region will change. The modeled effects of warm water intrusion have not been broadly proven throughout the west coast of Spitsbergen. Additionally, while it is clear that ocean depth and proximity to warm water are critical factors that determine the likelihood of intrusion, other factors not accounted for in this study may also exist. Future studies should attempt to expand the model for warm water intrusion to other parts of West Spitsbergen, consider the impacts on primary production further than 7km from glaciers, more accurately constrain the factors that allow warm water intrusion to occur, and evaluate changes in sea ice coverage, chlorophyll concentration, and particulate backscattering through time to measure the previous impacts of warm water intrusion in the area.

If this model is confirmed to be an effective way to approximate fjords in Svalbard’s response to primary production, several interesting analyses could be conducted. First, the model could be extrapolated out to other regions that are considered at high risk for warm water intrusion, such as Greenland. This would allow a regional prediction for response to warming ocean water to be developed. Secondly, a greater quantity of hydrological measurements of primary production in near-coastal areas would allow for the assessment of how carbon cycling functions in these systems. This would enable researchers to predict changes in carbon sequestration in this area in the near future, which could refine understanding of effects of changes in primary production on atmospheric CO₂ concentrations.
Conclusions

As changing atmospheric and oceanic circulation patterns cause increases in warm water intrusion into Arctic fjords, water masses and primary production will respond predictably to these changes. In Kongsfjorden, a case study location, the intruding water mass, TAW, warmed 4 to 5.5°C in a mere 15 years. This has caused a loss of sea ice, increased glacial retreat, changed water column structure, and changes in chlorophyll in the overall region. Measurements from the summer of 2014 and a 12 year satellite dataset allowed for a broad temporal understanding of changes in this case study location, which made it possible to develop a model for the impacts of warm water intrusion on primary production. According to the model, first sea ice coverage will diminish, as it already has in much of the Arctic, allowing earlier phytoplankton blooms to occur and changing the water mass structure of the fjord. Second, glacial melting and retreat will increase, decreasing light availability and causing primary production to decrease significantly. Finally, as tidewater glaciers retreat onto land, light availability will increase, allowing early and intense phytoplankton blooms. The model proposed as a part of this study explains the changes observed in Kongsfjorden and fits with established mechanisms and data collected in previous research. Testing this model across the Arctic and expanding it to explore other changes caused by warm-water intrusion are critical next steps to this research.
Acknowledgements

This research was conducted as a part of the Svalbard Research Experience for Undergraduates program in the summer of 2014 and thanks to funding from the Grua/O’Connell Faculty Student Research Award. I would like to first and foremost thank my advisor and mentor, Collin Roesler, who spent countless hours helping me to improve and always pushing me to produce my best possible work. I would also like to thank the Svalbard REU team, Ross Powell, Julie Brigham-Grette, Steven (Sven) Ossim, Ryan Pajela, Dominique Seles, Robert Ivey, Jessica Miles, and Peggy MacNeil for their support and assistance during the early days of this project, and for keeping the polar bears away. Thanks also to my advisory committee, Emily Peterman and Michèle LaVigne, for their helpful critiques of my drafts; to Sébastian Barrault and Ingrid Kjerstad of the Kings Bay A/S Marine Lab for their expertise and equipment; and to the Bowdoin Earth and Oceanographic Science department for their support. Finally, I’d like to thank Zachary Burton, Sasha Kramer, Claudia Villar-Leeman, all of my parents (John and Lisa Payne, Murli and Vandana Thirumale, and Wess and Kristina Hoffman) as well as all of my siblings (Stephanie Payne, Antara and Meghana Rao, and Bradley and Gracie Hoffman) for keeping me sane and humble throughout this project.
Works Cited


