Abstract

Fjords are awesome, and sometimes there are whales in them. This raises some questions: First of all, where do fjord babies come from? (Some Great Ice Cream Scoop in the Sky???) Why should we care about them? (Where to begin?) Where in the world do they happen? (A Goldilocks temperate zone and it’s pertty too!) Why are British Columbia’s fjords the best? (If you’ve ever visited them you wouldn’t be asking such stupid questions.) What do you know about the fjords of the Gitga’at Territory and the Bangarang study area? (Ah, the Kitimat Fjord System! Where to begin?!)
Introduction

Due to the complexity of their ecological space, coastal waters comprise some of the most diverse and productive marine habitats on earth (Levin & Dayton 2009). This high biodiversity is the cornerstone of lucrative fisheries, tourism industries, and innumerable ecosystem services, including nutrient cycling, nursery habitat, food web support, carbon sequestration, and tourism revenue (Turner 2000). And yet, while the health of coastal waters is the most economically valuable and easily monitored of marine systems, they are also among the most endangered (Gray 1997). As the coastal habitats of the world face intensive development, even the historically remote, rugged, and inaccessible coasts of the world are now facing heavy disturbance (Syvitski et al. 2005).

This conflicted condition of our coasts -- superlative ecological versatility met with accelerating anthropogenic exploitation -- finds its hyperbole in those far-flung habitats known as fjords. Though these expansive, glacially carved inlets of temperate and sub-polar latitudes seem pristine, they are developing fast. Intensifying vessel traffic, fish stock depletion, seismic surveys, and oil developments are finding their way into these remote marine habitats, deteriorating their ecological qualities at alarming rates (Orensanz et al. 1998, Ford et al. 2009, McLeod et al. 2010, Boyd et al. 2011).

In fjords, however, these problems can become opportunities. As Syvitski et al. (1987) put it, fjords are “perfect natural oceanographic and geologic laboratories. Source inputs are easily identified and their resulting gradients are well developed”. Isolated from the continental shelf by narrow entrances and shallow sills (Crawford et al. 2007), these isolated environments allow marine scientists to approach systemic questions experimentally in a natural setting. This offers the same conveniences that islands afford terrestrial ecologists, bequeaths similar insights into the ecological rules that govern isolated communities, and brings the need to assess human impacts within reach of limited scientific resources.

In physical, chemical, and biological respects, fjords are unique and bizarre environments. Their characteristics vary widely, both within individual fjords and throughout the fjordlands that occur worldwide. Features also vary according to various factors: latitude, degree of variegation, and bathymetric complexity being a few. As geologically young land features (late Pleistocene), fjords are still in the midst of fundamental upheaval, experiencing interglacial rebound, glacial scouring and massive sediment loading (Miller et al. 1982). Compounding this complexity, strong diurnal, tidal, and seasonal cycles induce punctuated variability in temperatures, terrestrial inputs, and precipitation (Hoskin & Burrell 1972). These inputs are funneled and mixed by the narrow fjord walls in an aggravated two-way “estuarine circulation”, leading to stark gradients in water column density, salinity, oxygenation, and nutrient load (Carstens 1970). In fact, every aspect of fjordic systems can be characterized by inordinate heterogeneity. This is thought to yield patterns in ecology, diversity, and productivity that are unique to fjords (Matthews & Heimdal 1980).

Unfortunately for all their scientific attraction, fjords are difficult places to work. Their rugged topography, remoteness, and unforgiving weather make field science notoriously difficult. For the fjords of the Canadian arctic archipelago or Greenland, where most of the world’s fjords can be found, very little data exist at all (Syvitski et al. 1987). With only a patchy and incomplete picture of this major ecosystem, ecologists consider fjordlands one of the least-understood coastal habitats on earth (Pearson 1989).

No fjordland has been studied equally from all biological, physical, chemical, and oceanographic angles. But when taken together studies of fjords comprise an impressive corpus of knowledge. British Columbia’s fjordland is relatively well-studied in terms of its zooplankton community (e.g. Mackas 1980) and tidal features (e.g. Cummins et al. 2003). Over 125 years of data on physical oceanography have been collected in Norwegian fjords (Farmer & Freeland 1983). New Zealand studies have focused on physical-biological interactions (Beer et al. 2011, Wing & Leichter 2011), land-sea energy exchanges (McLeod & Wing 2009, McLeod et al. 2010), subtidal ecology (Wing & Leichter 2011, Wing et al. 2008), trophic dynamics (Jack & Wing 2012, McLeod & Wing 2007), biogeography (Lawton et al. 2011, Wing 2009), and top predator energetics (Lusseau & Wing 2006).

Over the decades, various studies have brought us closer to a synthesis of fjord ecology. Matthews and Heimdal (1980) and Pearson (1989) offered the such reviews. Freeland et al. (1980) and Farmer & Freeland (1983)
offered impressive compendia of physical oceanography in fjords. These were updated by Tande (2001) in a succinct encyclopedia article of some of the primary ecological processes common to fjords, primarily in Norway. Syvitski et al. (1987) was the first to compile a geography of the world’s fjords, and in many ways it remains the seminal work on fjordic systems.

Defined

Fjords are estuaries according to the definition of Cameron and Pritchard (1963): a “semi-enclosed body of water having a free connection with the open sea and containing a measurable quantity of sea salt”. But the oceanography of fjord estuaries are unique unto themselves, setting them apart from estuaries like Chesapeake Bay and comprising a branch of integrative oceanography that is still undergoing active basic research. A typical temperate fjord is a deep U-shaped basin, a vestige of past glacial scouring with a glacial-mud bottom, a glacier or river at its head, and an underwater ridge (a sill) across the mouth (Pickard 1967, Pickard and Stanton 1980).

Sills are underwater ridges at which receding glaciers once deposited crushed rock and silt in a terminal moraine. In Jervis Inlet the sill stands 300m above the fjord floor. Glacier Bay's entry sill is only ~25m deep, and the depths of its other many sills vary. Some fjord basins have more than one sill, each one marking an interstadial deposit of the glacier responsible for the fjord (Thomson 1981). The slow isostatic rebound of the continent, in response to the melting of the ice sheet's load, is causing these sills to rise still today; some sills, like that of Powell Lake, now protrude above sea-level (Thomson 1981). Deep basins exist between these sills (e.g., Glacier Bay’s basins are up to 458m deep; Hooge & Hooge 2002). The change in depth from one side of the sill to the other can be dramatic (Thomson 1981). As described below, sills are barriers to “communication” between waters within fjords and the continental shelf offshore (Burrell 1986).

The term “head” is used for the inland end, and “mouth” for the seaward end (Pickard and Stanton 1980). Fjords with glaciers at their heads and icebergs adrift within the fjord are called “iceberg” fjords. (Pickard and Stanton 1980). Mainland fjords are often classified according to the amount of freshwater runoff they receive from snowmelt (Pickard 1961).

Worldwide

Silled fjords are a phenomenon of mid- to high-latitude coastlines worldwide, occurring at high latitudes in both the Atlantic and Pacific oceans (Mann & Lazier 1996). Due to their crenellated and complex topography, fjords comprise a major percentage of the world’s coastline (Syvitski et al. 1987). At first-order approximation, 47% of the world’s coastline is to be found in fjordlands2. While the world’s ~1,900 fjords can be found anywhere above a minimum latitudinal belt (above ~42.5 degrees), a specific confluence of climatic and geologic conditions must exist to generate the extensive complexes of densely packed fjordic coasts, or fjordlands. Distributed throughout the two major fjord regions of the world (north of 42 N latitude and south of 43 S), there are 9 “fjordlands”, provinces of dense and persistent fjord networks (Syvitski et al. 1987)3:

1. Northwest North America (up to 390 fjords)
   (North Pacific fjords occur between latitudes 50 and 60, Pickard and Stanton 1980)
   Alaska (220)
   BC (150)
   Washington (20)
2. Northeastern North America (230)
   Maine (30)

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2 If we assume that ~1,900 fjords that are to be found within the 9 major fjordlands (from Syvitski et al. 1987), and that each has the same average coastline length as the average Norwegian fjord (89.36km, arrived at by dividing Norway’s coastline of contiguous fjords, known to be 21,000km, by the number of fjords it has, 235), we find that there are 169,784km of fjord coastline in the world. Divide this by the total coastline of the world (356,000km, CIA 2012 World Factbook), and we find that ~47.6% of the world coastline is fjordland.

3 Pickard and Stanton (1980) have slightly different numbers for some of these fjordlands: 58 fjords in BC (17 on Vancouver Island and 41 on the mainland); 15 in Alaska; 32 in Chile; 14 in New Zealand. Syvitski’s numbers, being the seminal and most recent, are used here.
Newfoundland (100)
Nova Scotia (35)
Quebec (5)
Labrador (60)

3. Canadian Arctic Archipelago (350)
4. Greenland (335, the least-studied fjords on earth)
5. Patagonia (200)
6. Norway (including Svalbard, 235)
7. New Zealand (30)
8. Iceland (80)
9. Scotland (50)

Other Fjord Provinces:
- Russian arctic archipelago (40)
- Northern Europe
- Antarctic and sub-Antarctic islands

Counting its western, northern and eastern coasts, Canada has twice as many fjords (approx. 700) as the country in second place (Greenland).

**British Columbia**

The British Columbia mainland meets the Pacific in a complex network of interwoven fjords, inlets, passages, and corridors. These inlets consist of long, narrow channels bounded by steep mountainous terrain reminiscent of Norwegian fjords (Thomson 1981). Throughout the scientific literature, such features have been unsystematically referred to as “inlet”, “arm”, “canal”, or “channel” (Pickard 1961). Pickard (1961) wrote that the BC inlets are “more like the fjords of Norway and New Zealand than estuaries like Chesapeake Bay”.

*From Pickard and Stanton (1980):*

![Figure 2. Fjords in Alaska and northern British Columbia.](image1)

![Figure 3. Fjords in southern British Columbia and Vancouver Island.](image2)
The BC coastline length is 850km long as a bird flies (Pickard and Stanton 1980), but the fjordland’s complexity results in a shoreline length of over 27,000 km (Thomson 1981). These fjords are oriented variably with respect to the continental boundary. Some nearly parallel to the general northwest-southeast trend of the BC coastline, while others have a north-northeast trend. These inlets can cut deep into the mainland; for example, the Burke Channel (Herlinveaux 1973), Dean Channel and Fitz Hugh Sound waterway extends 150 km inland from Queen Charlotte Sound, (Crawford et al. 2007). Most BC fjords open to the broad but shallow (20-100m) Queen Charlotte Sound and Hecate Strait before connecting with the ocean (Pickard and Stanton 1980).

Forty two mainland inlets are 10 miles or more in length (Pickard 1961). The mean width of these fjords range from 0.6 to 15 km (Pickard and Stanton 1980). About 40% of BC fjords have rock flour in suspension in the upper layers (Pickard and Stanton 1980). As of 1980, only five of the BC mainland’s fjords had established human communities (Pickard and Stanton 1980).

Few inlets have been studied in any detail, the exceptions being Knight Inlet, the Bella Coola estuary (Fitz Hugh Sound/Burke Channel), Rivers Inlet (Whitney 2002), and to a lesser extent the Kitimat system (Webster 1979) and Rupert-Neroutsos Inlets. Knight Inlet, particularly the hydraulic flow over its, has been studied extensively (Crawford et al. 2007, Cummins et al. 2003, and others).

The Kitimat Fjord System

The Gitga’at Territory, the study area for the North Coast Cetacean Society and the Bangarang Project, and the Confined Channel Assessment Area for the proposed Northern Gateway Project are found within the Kitimat Fjord System: “the system of channels connecting Kitimat with British Columbia coastal waters (MacDonald et al. 1983). This complex extends 140 km inland from the Pacific Coast, and is nested with the Kitimat Ranges, a sub-province of British Columbia’s Coast Mountains (generally 2000-2200m in elevation)(MacDonald et al. 1983).

The Bangarang study area, Gitga’at Territory: 
Gil Basin and adjacent waterways in the Kitimat Fjord System
Fjords of the Kitimat system have a typical morphology, with steep bedrock walls and relatively smooth sediment-floored basins separated by high-relief sills (MacDonald et al. 1983). Douglas Channel, Gardner Canal, and Surf Inlet are the primary fjords that feed into the system (Pickard 1961). Douglas is 45 nm long, 1.9 nm mean width, 330 mean depth, 455m max depth, 210m sill depth (Pickard 1961). Gardner is 70nm long, 1.0nm mean width, 275m mean depth, 500 max depth, with a outer sill depth of 67m (Pickard 1961).

From Fissel et al. (2010):
Table 1. Physical dimensions of the Kitimat fjord system.

<table>
<thead>
<tr>
<th>Region</th>
<th>Length of Shoreline (km)</th>
<th>Surface Area (km²)</th>
<th>Water Volume (km³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kitimat Arm, Douglas Channel, Lorette Channel</td>
<td>326</td>
<td>380</td>
<td>89</td>
</tr>
<tr>
<td>Kildala Arm, Amos Passage</td>
<td>60</td>
<td>59</td>
<td>9</td>
</tr>
<tr>
<td>Gardner Canal, Devastation Channel, Verney Passage</td>
<td>202</td>
<td>207</td>
<td>37</td>
</tr>
<tr>
<td>Ursula Channel, Princess Royal Channel, McKay Reach</td>
<td>208</td>
<td>185</td>
<td>50</td>
</tr>
<tr>
<td>Wright Sound, Whale Channel, Squally Channel, Campina Sound</td>
<td>250</td>
<td>604</td>
<td>250</td>
</tr>
<tr>
<td>Grenville Channel*</td>
<td>87</td>
<td>43</td>
<td>5</td>
</tr>
<tr>
<td>Total fjord system</td>
<td>1275</td>
<td>1643</td>
<td>470</td>
</tr>
</tbody>
</table>

*Dimensions calculated only as far as the sill.

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Fig. 2a. Longitudinal section of Douglas Channel.

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Figure 3. Channel width B (metres)

Figure 4. Cross-sectional Area A (Metres²)
Physical-Chemical Oceanography

Freshwater Input

The amount of freshwater that flows into an inlet from the river depends on the drainage area, the time of year, and whether the river is fed primarily by rainfall or snow-melt (Thomson 1981). Rain-fed inlets on islands further south (e.g. Vancouver Island) reach their maximum discharge volume in late-winter, at the peak of the winter storms. In contrast large mainland inlets further north, which drain coastal ranges, reach their discharge peak at the height of the snowmelt season, beginning in May. Late spring snow-melt runoff can be an order of magnitude greater than rain-fed volume in winter. In summer, surface waters of inlets are often rendered a milky-green color by the large amount of glacial silt carried downstream (Thomson 1981). Such pulses of freshwater input along the coast can influence shelf currents throughout the year. Year-to-year discharge is highly variable, by as much as 50% (Fissel et al. 2010).

From Fissel et al. (2010):

In the outer channels of the Kitimat Fjord system, the thickness of the freshwater lens tends to be less than 1 m. The freshwater sill deepens to as high as 9.6 in Gardner Canal and 3.6m in Kitimat Arm (Fissel et al. 2010). Often the "jet" of river water that enters an inlet remains partly intact, and moves down inlet as a thin ribbon of silty surface flow. (Thomson 1981). In the northern hemisphere, Coriolis forces deflect tongues of runoff from the mouth of fjords to the right, such that channel surface waters are fresher and more turbid (and sediments have coarser grain sizes) on their right sides (Bornhold 1983).

The BC fjords can be categorized as high, intermediate or low discharge (or runoff) inlets (Crawford et al. 2007). High runoff inlets penetrate far enough inland to tap into glacial watersheds that contribute runoff all summer long, not only in the spring. Their surface layers (up to 10m thick) are typically dilute and turbid (silt- and humic acid-laden), and estuarine flows are the surface can be 10 cm per second. Examples include Burke and Dean channels, Knight and nearby Kingcome inlets. High loads of suspended sediments absorb light as well as heat, creating plumes of warm waters. Secchi discs in these inlets may only be visible down to 30cm or so (Crawford et al. 2007).

Intermediate runoff fjords have less glacial runoff but sufficient snow melt to produce relatively high spring runoff. The surface layer is thinner, only about 2m. Examples include Seymour, Belize and Neroutsos inlets.

Low runoff fjords have high discharge only in the spring and their surface layer is extremely shallow and only partially diluted. Estuarine flow is weak. Rupert-Holberg Inlets are an example.
Pickard (1961) used a different runoff classification system, in which “A” groups are characterized by low surface salinities at the head. He classified Douglas Channel and Gardner Canal as “A-1” inlets, the A-1 subgroup having the lowest salinity (MacDonald et al. 1983). A total of 29 drainages contribute runoff to the Kitimat Fjord System’s outflow, which can be organized into 7 major drainage areas. (Fissel et al. 2010). Gardner Canal has been recorded with the third highest discharge of BC’s fjords (second to Portland Inlet of the Nass River)(Pickard 1961). The Skeena and Nass rivers provide most of the riverine discharge (Fissel et al. 2010). The Bangarang study area comprises two of these seven drainage basins:

1. The southern waterways: Campaia Sound, Caamano Sound, Squally Channel, Otter, Estevan and southern Principe: 1415. Sq km of water area, 253.5 sq km of islands, 953.5 sq km of drainage area.

2. Southern Douglas, Whale, Verney, Ursula and MacKay: 705 sq km of water area, 536.4 sq km major islands, 853 sq km of land drainage area.
Circulation

In the Kitimat Fjord System, water is circulated by a combination of three processes: estuarine circulation (forced by freshwater discharge), wind stress and tides (MacDonald et al. 1983). The relative importance of each varies according to location and time. The former two are responsible for a vast majority of fjord water “communication” with outer waters; due to their oscillatory nature tides do not function as a strong advective agent (MacDonald et al. 1983). Water movements are also influenced by seasonal patterns and storm events on the adjacent shelf. See the BC Coast Backgrounder for a discussion of those influences.

Regardless of the local prevailing force, currents tend to be aligned with the axes of the channels themselves; cross-channel currents are relatively negligible (Fissel et al. 2010). This is less true in outer, broader channels of fjord systems, like in Caamano Sound where currents take on a much greater range of directions than in the inner waterways of the Kitimat Fjord System (Fissel et al. 2010). There is also a consistent vertical cline in current speeds; surface currents are almost always faster than intermediate and bottom currents (see below figure).

From Fissel et al. (2010):

![Figure H-3 Current Speed versus Depth for Kitimat Arm Site during Winter of 2006](image)

**Estuarine**

Estuarine circulation is a phenomenon endemic to fjords and similar coastal habitats. Bouyant freshwater discharge exits the inlet in a shallow surface layer. As it progresses seaward, it entrains the saline water underlying it. A slow, deep inflow of water must compensate for this upward entrainment of salty water. A two-way flow is therefore established, a seaward surface flow and its landward bottom countercurrent. This is “estuarine circulation”. It is well-established by early summer, at the height of snow-melt (Thomson 1981, Freeland and Farmer 1980).
From Thomson (1981):

Evidence of estuarine circulation in Otter Channel, from Fissel et al. (2010):

Fig. 2.7. Estuarine circulation in a typical British Columbia inlet. Salt water entrained and carried seaward by river outflow is replenished by a net inflow at depth. Sloping isohalines (lines of equal salinity) indicate a down-inlet increase in salinity in surface brackish layer. Turbulent mixing occurs in vicinity of sill.

ESTUARINE CIRCULATION IN AN INLET

To the Open Sea

Fig. B-14 Joint Speed and Direction Distribution in Otter Channel (CM8) at 40-m Depth and at 136-m Depth in Winter

NOTE: In (a), the maximum percentage scale is 50% rather than the usual 40%.

Estuarine flow is strongest during periods of high runoff, but can be also be modulated by tides, winds and oceanographic conditions offshore over short time periods (Pickard and Stanton 1980). This estuarine exchange fluctuates in intensity with the fortnightly spring-neap cycle of tides (Masson and Cummins 2000). Estuarine-driven surface currents in Burke Channel reach speeds of 100-200 cm s⁻¹, with a significant reverse flow below the halocline (Crawford et al. 2007).

Estuarine-type circulation can also occur on a larger scale. Estuarine circulation has been observed in the Juan de Fuca Strait, where the cumulative discharge of the rivers of the Strait of Georgia establish a regular surface outflow that must be replenished by inflowing deep water (Masson and Cummins 2000). It is even thought to be a significant driver of water mass movements in BC shelf waters (Crawford et al. 2007). The combined freshwater discharge of many adjacent fjords, in addition to the upwelling regime usually in place during the
summer, creates an estuarine-like circulation of colder waters up onto the shelf, replenishing fresher surface waters that are moving offshore. This why deep waters along the fjordland shelf are unexpectedly coldest in the summer and warmest in the winter; this cold deep water is a signature of the cross-shelf cycling of cold bottom waters (Thomson 1981). See the BC Coast Backgrounder for more.

As a high-runoff system, the Kitimat Fjord system is expected to have well-developed estuarine circulation (MacDonald et al. 1983). Average net flows are observed to be seaward near the surface and generally landward at depth (Fissel et al. 2010). Typical surface currents are 15-30 cm/s (Fissel et al. 2010). The highest surface currents occur in seaward portions, including Campania and Caamano Sounds: over 1 m/s. (Fissel et al. 2010).

Another example of observed estuarine flow in the Kitimat Fjord System, from Fissel et al. (2010):

Wind Driven
Due to the height of surrounding hills, winds tend to be directed up or down channels rather than across them (MacDonald et al. 1983). Depending on their orientation up or down channel, winds in fjords can serve to temporarily slow, stop, or exacerbate estuarine circulation. When freshwater discharge is minimal, strong down-inlet winds can maintain a two-way flow similar to estuarine circulation (Dodimead and Herlinveaux 1968; Herlinveaux 1973). Such winds, which are especially common in winter, can also deepen the mixed layer (Thomson 1981, Freeland and Farmer 1980). The katabatic winds of winter can drive surface waters seaward at such rates that both bottom and intermediate waters are pulled into BC inlets (MacDonald et al. 1982).
Tidal Features
Tidal currents and sea level changes in fjords can be dramatic. A 6.2m tide has been observed in the Gitga’at Territory (Fissel et al. 2010). An 8m tide was observed nearby in Principe Channel (Fissel et al. 2010). Tides, though they contribute to mixing, are oscillatory and so are not a strong advective agent (MacDonald et al. 1983). Baroclinicity in tides have been observed in Douglas Channel (Buck Inrigham 1980, Webster 1980, Webster 1982).

In BC and Alaska, the times of high and low water and the range of tide are largely the same at the head and mouth of each fjord, since their natural periods of oscillation are much less than the period of the semi-diurnal tidal component (Pickard and Stanton 1980). However, in the Kitimat Fjord System, tidal behavior in the inner channels is quite different from that of the waterways surrounding Gil Island (Narayana 1980). Tides in the interior channels can be thought of as standing waves, but around Gil Island there is a strong propagating wave component (Narayana 1980). There tidal currents wrap around islands and seafloor features, creating a frontal zone on their far side where the bifurcated flows re-converge.

Vigorous tidally-induced mixing occurs in the bottlenecks within coastal inlets, where waterways are constricted either vertically by sills or laterally with narrowing channel walls — or both. The sills at the mouth of fjords are critically important in local flow dynamics in fjords. As waters are pulled over them and into the inland basin by estuarine circulation (which in turn displaces deeper water upwards), extensive mixing and upwelling can occur. Upwelling is apparent at the shallow sills of Verney Passage and Ursula Channel during periods of fresh water input. Here, estuarine circulation is coupled with vigorous tidal mixing (MacDonald et al. 1983). According to MacDonald et al. (1983), “McKay Reach and Wright Sound may read the benefit of this upwelling with sustained primary production.”

Frontal zones often establish over sills and can be characterized by high primary productivity (Parsons et al. 1983). The position and magnitude of the frontal zone, and the vigor of its associated mixing, has been observed to change as water velocities oscillate with the tidal cycle (Cannon and Laird 1978). 180 degrees out of phase with maxima in this periodic tidal exchange, there is a 14-day cycle in chlorophyll maxima (observed in Saanich Inlet; Hobson 1981, Parsons et al. 1983). As a result, primary productivity appears to peak during the velocity minima of neap tides. While other mid- to high-latitude regions have shown the opposite pattern — i.e., a synchronization of lunar and chlorophyll cycles (e.g. Gulf of Maine; Balch 1981) — in fjords it seems that primary productivity maximizes during periods of greatest stability (i.e. neap tides) and minimizes during periods of maximum water exchange (Parsons et al. 1983). If so, this may be important in the timing of biological events, such as zooplankton life cycles (Parsons et al. 1983).

Tidally accelerated flows of stratified estuarine waters over sills and other channel constrictions and seafloor features can produce internal waves. Internal waves of both short period (minutes) and longer period (hours) are common features of Pacific fjords (Pickard and Stanton 1980).

Phenomenal motionless internal wave trains can develop over some sills, such as those in Knight Inlet (Cummins et al. 2003). As currents increase over a sill throughout a tidal cycle, the flow upstream of a sill bifurcates into a nearly motionless surface layer and an actively flowing bottom layer, separated by an intermediate layer of mixing. Above the sill crest an unstable interface between surface and deep layers occurs, where slow mixing gradually expands the intermediate layer. The deep layer accelerates with further tidal forcing, plunging over the sill then forced hydraulically into oscillations as it adjusts to downstream conditions. Meanwhile, waves are held motionless just upstream of the sill crest. Then, as the tidal current slackens, these standing waves are released and begin to propagate upstream, slowly at first, but more rapidly as the tidal flow continues to wane. The original bore upstream of the sill evolves into a train of upstream-propagating internal waves. As they propagate into deep waters beyond the sill, tidal current diminishes further and the waves escape into the far field.

Such waves are most conspicuous in the oscillating position of the halocline (Pickard and Stanton 1980), but they sometimes develop surface signatures too. Surface manifestations of these internal oscillations can also develop (Crawford et al. 2007, Cummins et al. 2003). All of these tidally induced features mix waters, accumulate nutrients, aggregate plankton, and attract a cadre of predators (Gilmartin 1964, Crawford et al. 2007).
Property Distribution

Vertical Structure
In Gulf of Alaska fjords, typical vertical structure consists of an upper mixed layer, an intermediate layer between the base of the seasonal thermocline and the halocline, and a sub-halocline layer, which is slowly advected shoreward during much of the production season (March-October) (Coyle et al. 2005). MacDonald et al. (1983) split the water column in the Fjord Kitimat System into four horizontal “regions” (MacDonald et al. 1982):

1. **Surface to nitrate minimum**, silicate maximum at the surface. Just under the pycnocline at 10m there is a silicate minimum accompanied by a local oxygen maximum.
2. **Intermediate remnant of winter cooling: temperature** minimum at 20-30m. Relatively enriched in silicate. Found only at the head of inlets.
3. **Deep water** with a very narrow range of temperature and salinity, probably results from uplift of bottom water during renewal of the outer two basins. Elevated nutrient and reduced oxygen concentrations. Relatively enriched in silicate, nitrate and phosphate.
4. **Bottom water**: salinity exceeding 32.7 parts per thous. Temperature continues to decrease with depth.

The depth of the thermo-, halo- and pycnoclines will be dependent upon local and upstream freshwater discharge conditions as well as local temperatures. Therefore, fjords of distinct discharge “types” have diagnostic water column property profiles.

*From Pickard and Stanton (1980):*

![Diagram of water column properties](image)

**Horizontal Structure**
Estuarine circulation, together with heat forcing in surface waters, largely dictates the vertical structure of fjords throughout the year (Coyle et al. 2005). Horizontal trends in salinity and temperature are dramatic in the surface waters of fjords, but negligible in intermediate and deep waters (Pickard and Stanton 1980). Estuarine circulation establish a hydraulic gradient (lateral clines in salinity – the predominant determinant of density in the shallow
waters of fjords (Pickard 1961)—properties along a channel's surface⁵—and leads to a net annual conservation of salt in the fjord basin: Surface salinity increases from river mouth to fjord mouth, and deeper waters show the opposite gradient, freshening towards the river mouth. Sills can establish strong localized salinity gradients. Across a shallow sill south of Victoria the salinity at 100m might change from 33.5 per mil on the seaward side to 31.5 per mil on the landward side (Thomson 1981).

Seasonality

Salinity and temperature in fjords exhibits an annual cycle: In February, after a winter of extensive and deep mixing by storms, winds and convection, surface waters are found to be highly saline and very cold while bottom waters are at their freshest and warmest (MacDonald et al. 1982). Everything is better mixed, and all properties are more homogenous throughout the water column. In spring, as the water column becomes more stratified and estuarine circulation is reestablished, surface salinities decrease and deepwater salinities increase. Summer peaks in estuarine circulation, corresponding to peaks in freshwater discharge (typically in September), bring about annual minima in surface salinities and maxima in deepwater salinities. Surface salinities will be low during periods of comparatively large runoff and high during periods of relatively low runoff. (Thomson 1981). In late autumn, there is strong near-surface stratification: a shallow, brackish (26 psu) upper layer and a deep, cold, more saline (32 psu) bottom layer. Iso-, thermo-, and halo-clines are sharp (e.g., in Knight Inlet, the thermocline is typically 2-5m thick; Trevorrow et al. 2005), though not always overlapping (Coyle et al. 2005). As winter digs in and mixing intensifies, the pattern reverses back to February conditions.

Other Properties

PH is generally high in the fjords, except in the presence of acidic runoff (Fissel et al. 2010). Freshwater discharge has lower pH values because of the acidity of dissolved carbonate in it. Streams with strong discoloration from humic acid tend to be more acidic. The low salinity of surface waters also leads to lower pH values than open ocean waters (Crawford et al. 2007).

In BC fjords, surface DO values at or above saturation (especially during spring plankton blooms) are usual (Crawford et al. 2007). Oxygen levels tend to decrease with depth (Crawford et al. 2007) but the regularly renewed bottom waters are rarely, if ever, anoxic (Pickard and Stanton 1980). Vertical maxima and minima in dissolved oxygen are common but do not show any strong pattern (Pickard 1961). Marine oxygen levels are declining worldwide due to warming, and this depletion may be more pronounced in coastal inlets. The ecosystem impacts of the expansion of oxygen minimum zones in BC are unknown (Crawford et al. 2007).

Nutrient levels on the BC continental shelf and in its inlets are generally high. (Fissel et al. 2010). Fjords act as nutrient traps where organic and inorganic materials in freshwater discharge accumulate and are concentrated, thus laying the foundation for a highly productive and intricate ecosystem (Thomson 1981).

Deep and Bottom Waters

Sills are barriers to the exchange of nutrients, salinity and oxygen in the deep waters of fjords. Conversely, sills also contribute to the retention and concentration of river-borne nutrients within fjords, which is a primary cause of their remarkable productivity (Thomson 1981). Sills can be the cause of sharp gradients in salinity at fjord mouths (2 psu differences are not uncommon at 100m depth; Thomson 1981).

In some fjords of the world, accumulating freshwater discharge can deepen the fresh surface layer until the pycnocline lowers to the depth of the sill. This “caps off” the denser waters at the bottom of the inner basins, barricading these bottom waters from renewal by outside flows; in these conditions, exchange can only occur once stratification weakens (Hooge & Hooge 2002). However, BC fjords differ from Norwegian fjords in this one sense: very few, if any, of the BC mainland coast inlets are sufficiently stagnant to exhibit anaerobic conditions (Pickard 1961). The pycnocline rarely meets sill depth and deepwater renewal tends to be regular. One notable and very well studied exception is Saanich Inlet. (Pickard 1961)

When deepwater exchange does become blockaded in fjords elsewhere, waters trapped behind it are typically hypoxic and even anoxic near-bottom, where nutrient concentrations and microbial activity are high and large

concentrations of hydrogen-sulfide can be found (Thomson 1981, Fujiwara and Yamada 2002, Arneborg et al. 2004). In such fjords, deep water is typically only renewed during winter when water column structure is homogenized by colder surface temperatures and vigorous mixing (Hooge and Hooge 2002). Again, anoxia is not common in BC fjords (Pickard and Stanton 1980).

In the Kitimat Fjord System, BC, bottom waters in its three basins undergo regular renewals (perhaps two replacement per year) and there are no records of anoxic conditions being observed (MacDonald et al. 1982). Both estuarine and wind-driven circulation can lead to the replacement of bottom water (MacDonald et al. 1982). In Gil Basin, deepwater renewal appears to completely flush its waterways between June and October, while estuarine circulation and fair-weather northwest winds are well established (MacDonald et al. 1982).

Mid-water oxygen minimum layers (OMLs), which can occur in marine offshore regions and freshwater lakes, are also commonly present in fjords, though their variability is highly seasonal (Fujiwara and Yamada 2002; Arneborg et al. 2004). OMLs and stagnant bottom waters can govern demersal, planktonic and nektonic community composition and distribution (e.g. Bell and Eggleston 2005, Horpilla et al. 2005; see Plankton Processes Backgrounder).

Bottom waters in BC fjords have been warming in step with global temperature trends (Stucchi 2003).

**Sediments**

Waters at the heads of fjords are thick with suspended sediments. The largest turbidity values (corresponding to the smallest Secchi disc readings) are found at the head in the low-salinity surface layer (Pickard 1961). As a result of sedimentation from the rivers that feed them, fjord channels are often flat-bottom (Pickard 1961). However, vigorous bottom currents can limit or prohibit sedimentation and even erode glacial deposits in fjords. In the Kitimat Fjord System, bottom currents are preventing sedimentation in most areas except in the deepest corners of Gil Basin (Bornhold 1983).

Although many fjords including the Kitimat system are generally protected from oceanic tsunamis, the steep walls of their channels are vulnerable to submarine land-slides that can induce tsunamis within a fjord. Two such events, dated to the early Holocene, have been observed in the stratigraphy near the mouth of Douglas Channel in the study area (Thomson et al 2012). These massive landslides (approx. 65 million cubic meters) could have induced a tsunami wave that reached an amplitude of 20m or more on the opposite shore; but the complex bathymetry of the channel would have attenuated the wave to less than a meter in the 15 minutes it would have taken to travel the 45km to Kitimat’s shores.
Literature Cited


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