Cold ocean seas and northern hydrology: an exploratory overview

Ming-ko Woo

ABSTRACT

Oceans are expected to bring their maritime influence to coastal zones, and sometimes further inland, to affect the terrestrial energy and moisture fluxes. Heat and water balances jointly govern northern hydrology, both directly through the hydrological cycle and indirectly through cyclogenesis and radiation regimes. Drawing upon reported examples, a survey is presented to examine the roles of ocean currents, onshore winds, coastal storms and sea ice in modifying precipitation (snow, fog, orographic precipitation), evaporation (through radiation and moisture availability) and coastal inundation processes. Coastal currents can alter permafrost distribution on a regional scale and runoff patterns may be modified accordingly. There are also notable feedback mechanisms whereby hydrology affects oceanography. Examples include freshwater discharge that exerts influence on thermohaline circulation in polar seas and on the dynamics of coastal sea ice, and the export of sediments, organic carbon and nutrients to the Arctic Ocean. Given the sensitivity of high latitudes to climate warming, which impacts many aspects of the northern environment, collaborative investigations of ocean–atmosphere–hydrologic linkages are of priority interest.

Key words | northern hydrology, oceanic influence, sea ice, snow, streamflow, water balance

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INTRODUCTION

Hydrologic–oceanic interactions are expected to be important to northern hydrology, at least in the coastal zone. However, there is a scarcity of research on the coupling of terrestrial hydrology with oceanic processes and coastal dynamics. While collaborative studies on atmospheric–hydrologic linkages have led to much improvement of knowledge on the cryosphere (e.g. Woo 2008) no systematic investigation has been carried out on the mutual influences of polar seas and northern hydrology.

Several obvious but important conditions drive or restrict hydrologic–oceanic interactions.

(1) Open water is an unlimited moisture source whereas the land is relatively dry in comparison.
(2) Water has a much larger heat capacity than land and therefore, given the same energy input, oceans heat up and cool down more gradually than the continents.
(3) The atmosphere is a common medium that facilitates heat, moisture and momentum transfers between the land and the sea.
(4) Under a cold climate, the formation of a sea ice cover greatly reduces the heat and moisture contrasts over the land and sea surfaces.

Proximity to the sea moderates variations in temperature. Table 1 provides a comparison of the average minimum and maximum temperatures (1971–2000) of Coral Harbour (64.19°N, 83.6°W, elevation 64 m) and Baker Lake (64.3°N, 96.08°W, elevation 18 m). Baker Lake, about 300 km inland, has slightly warmer summers and colder winters than Coral Harbour on the Hudson Bay coast, though their annual means are similar (−11.8°C for Baker Lake and −11.6°C for Coral Harbour). The effect of coastal cooling in the summer is also noted at a local scale

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Table 1 | Monthly mean minimum and maximum air temperatures (in °C). (Top) for a coastal station (Coral Harbour: 64.19°N, 83.6°W) and an interior station (Baker Lake: 64.3°N, 96.08°W) in Arctic Canada. (Bottom) for a station on the west coast of Norway (Bergen: 60.23°N, 5.2°E) and on the Baltic coast (Turku: 61°N, 22.35°E)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Coral Harbour												
Temp. minimum	− 34.1	− 34.1	− 30.0	− 22.0	− 10.9	− 0.8	4.6	3.1	− 2.0	− 10.7	− 21.4	− 29.9
Temp. maximum	− 25.8	− 25.9	− 20.8	− 11.3	− 2.6	6.3	13.9	11.4	4.1	− 3.9	− 12.8	− 21.6
Baker Lake												
Temp. minimum	− 34.9	− 35.1	− 31.5	− 22.1	− 9.4	0.5	6.0	5.0	− 0.6	− 10.7	− 23.9	− 31.9
Temp. maximum	− 28.7	− 27.9	− 22.9	− 12.6	− 2.2	9.2	16.7	14.0	5.9	− 4.2	− 16.3	− 24.8
Bergen, Norway												
Temp. minimum	− 0.4	− 0.5	0.9	3.0	7.2	10.2	11.5	11.6	9.1	6.6	2.8	0.6
Temp. maximum	3.6	4.0	5.9	9.1	14.0	16.8	17.6	17.4	14.2	11.2	6.9	4.7
Turku. Finland												
Temp. minimum	− 10.0	− 11.0	− 6.7	− 1.8	3.6	8.6	11.2	10.1	5.6	1.4	− 3.2	− 7.9
Temp. maximum	− 3.6	− 3.6	1.1	7.0	15.2	19.6	21.5	19.4	13.2	7.2	1.5	− 1.9

on the Fosheim Peninsula, Ellesmere Island. Edlund *et al.* (1990) reported that Hot Weather Creek, which is 25 km inland of the coastal station at Eureka (80°N, 85.93°W, elevation 10 m) heats up more quickly than the coast and stays warmer during almost all days throughout the summer. For July 1989, Eureka mean temperature at 5.8°C was close to its long-term mean, but Hot Weather Creek (elevation 115 m) averaged 6.3°C.

Northern hydrology is controlled by both heat and water balances. Energy and moisture exchanges between the land and the sea have potential effects on cold region hydrology through modifying land phase processes like precipitation and snow melt, evaporation and condensation, glacier ice accumulation and ablation, flooding and wetland formation, soil freeze–thaw and runoff generation. The magnitude and timing, as well as frequency, of hydrological events are affected. There is also a temporal discord between the warming and cooling of the land vs the sea. While the land gets cold very quickly in the Arctic autumn, the sea remains warm relative to its overlying air. The formation of sea ice always lags behind freezing of the ground. In the spring, the land loses most of its snow cover while ice cover remains on the water. Then the land, deprived of its snow, would heat up quickly while the sea gradually loses its ice. During the year, a heat and moisture gradient often exists between the land and the sea.

Interactions among the ocean, the atmosphere and terrestrial hydrology are intricate but our current knowledge remains limited. This paper is a preliminary survey of the effects of proximity to ocean seas on the hydrology of coastal areas based on information published or inferred. The primary purpose is to stimulate further interest and more intense work on the reciprocated influences of northern waters on the land and in the oceans. This paper is written from a hydrologist's perspective, not due to the author's personal bias but because it reflects his constrained knowledge in ocean sciences.

OCEANIC INFLUENCE ON HYDROLOGY

Ocean currents

The temperature and moisture regimes of coastal areas can be much affected by the heat and moisture conveyed by ocean currents that move along their shores. One outstanding example is the Gulf Stream and its northern extension, the North Atlantic Drift, and warm currents bearing other regional labels that enter the cold regions of Scandinavia (Figure 1). The west coast of Norway is much warmer than locations of comparable latitudes on the other side of the Atlantic, or on the shores of the Baltic Sea. Table 1 shows the average minimum and maximum air temperatures of Bergen, Norway (60.23°N, 5.2°E) and

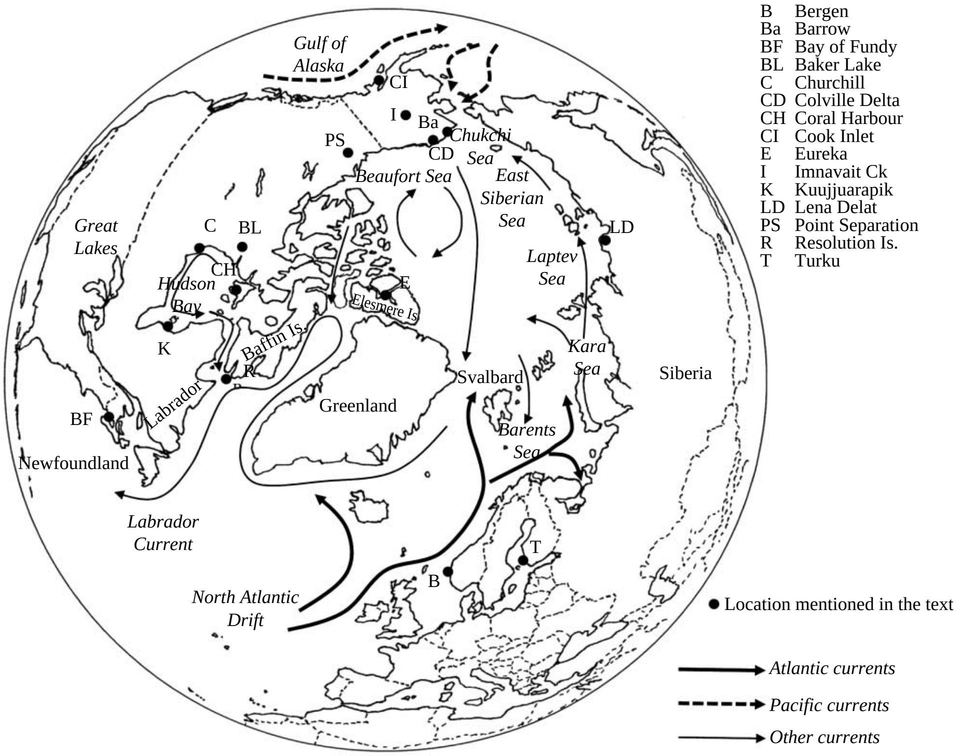

Figure 1 | Major ocean currents of the polar region. Also shown are the locations of most sites mentioned in the text.

Turku, Finland (61°N, 22.35°E). Both are at similar latitudes and both are coastal stations. Bergen is decidedly warmer in the winter. Unlike the Greenland or Baffin Island coasts, or the Baltic Sea with water of lower salinity, sea ice is not formed on the western fiords of mainland Norway. Further north, the extent of ice in late winter is influenced by the oceanic effect due to the warm current and the atmospheric effect of whether the wind is predominantly from the north or from the south (Vinje 2001). In recent years, there is a noticeable increase in sea temperature and ice-free conditions prevailed throughout the winter in the Barents Sea south of 80°N and in most western Svalbard fiords. With an ice-free Barents Sea, airflow over open water entrains moisture and is deposited on land, particularly where it meets the rugged topography of the Svalbard. Humlum *et al.* (2003) noted that, during the winter, strong southerly air flow over open water advects warm air to the Svalbard region, bringing heavy snowfalls and often periods of snowmelt even in mid-winter; but when polar air extends over the region strong westerly airflow brings coldness and heavy precipitation.

The warm ocean current also affects the distribution of permafrost which has significant effects on hydrology. Normally, permafrost is impermeable and this hinders percolation but facilitates rapid surface flow (Woo 1986). On the west coast of Norway, frequent above-freezing air temperatures due to the Gulf Stream influence prevents the permafrost from being formed in the valleys and the shores of the fiordland. Although runoff in the coastal belt responds quickly to snowmelt, glacier melt and rainfall, the rapid and large flow responses are mainly due to the steep topography and shallow soil rather than the frozen ground. Unlike Greenland or the Canadian Arctic Islands, baseflow is not interrupted by soil freezing and the flow continues throughout winter in most of the rivers on the mainland.

Cold currents can cool off the adjacent land, as exemplified by the southward-moving Labrador Current off East Canada (Figure 1). Upwelling of cold water cools the warm, moist air that comes from the south and west (Hertzman 1997) and, as air temperature falls below dew point, fogs are formed along the eastern Nova Scotia coast.

As a warm and moist southeasterly wind blows over the cold ocean current, fog is produced in southeastern Newfoundland and its advection to the land yields fog drips which amounted to 1.8 mm/d (Price 1992*a*). The fog also cuts down net radiation to half its clear-sky values, thus greatly reducing evaporation. Maritime influences through fog drip inputs and low evaporation maintain a high water table despite the steepness of the terrain. These conditions are crucial to the development of blanket bogs in the region (Price 1992*a*,*b*).

Teleconnection

Teleconnection is a large-scale anomalous pattern in atmospheric circulation associated with changes in planetary waves that persists for weeks, months or years, giving rise to climatic anomalies that are related to each other at locations separated by great distances (thousands of kilometres). Through teleconnection, the highly dynamic and mobile nature of the atmosphere is capable of conveying the oceanic effects afar. Thus, low frequency perturbations in oceanic conditions can force a hydrologic response in another part of the world. A much studied phenomenon is the El Niño that arises out of ocean–atmosphere coupling in the Pacific. There, upwelling of cold water off the South American coast is associated with the trade winds, but the resulting tropical sea surface temperature influences atmospheric circulation. El Niño-Southern Oscillation and the El Niño-like climatic variability known as the Pacific Decadal Oscillation (Mantua & Hare 2002) have been related to variations in precipitation (Shabbar *et al.* 1997), snowmelt (Moore & McKendry 1996) and streamflow in northern North America.

The Arctic Oscillation (Thompson & Wallace 1998) is an atmospheric pressure pattern in northern latitudes that oscillates between a positive (or warm) and a negative (or cold) phase. Its influence is particularly relevant to the high latitudes. The cold phase brings high pressure over the polar areas and low pressure at mid-latitudes. During the warm phase, there is winter-time warming of Eurasia and North America east of the Rockies but cooling over Labrador and Greenland. Storm tracks shift northward, bringing a strengthened oceanic influence to Alaska and Scandinavia. The relatively warm Atlantic water of high salinity is drawn further into the Arctic and the sea ice becomes thinner than normal. The Arctic Oscillation and the Pacific Decadal Oscillation are found to have teleconnection with river discharges of the Hudson Bay drainage in Canada (Déry & Wood 2004) and with the flow of southeastern Alaskan rivers (Neal *et al.* 2002), though the correlation between the climatic indices and precipitation or streamflow is complicated by local considerations such as location and topography (Moore *et al.* 2003; Woo & Thorne 2008). While evidence of linkage between hydrological variables and climatic indicators remains statistical and is tenuous at times, they do indicate remote connections among the atmosphere, the ocean and surface hydrology on a hemispherical scale.

Onshore winds

On a regional scale, moist-laden onshore wind augments precipitation on the land. This is particularly pronounced in areas of rugged relief where orographic effects are strong. An example is the Gulf of Alaska where annual precipitation reaches 3,000 mm per year. At high elevations, heavy snowfall feeds the many glaciers of the Kenai, Chugach, St Elias and Coast mountains. The hydrology of snow and glaciers, including their accumulation, melt and runoff generation, are of remarkable importance in that region (Meier 1990). Similarly, Norway receives much orgraphic precipitation from the westerly winds. Many glaciers are supported by large snow accumulation at high altitudes, and annual runoff exceeding 1,500 mm per year is generated from most coastal mountains (Killingtveit 1994).

On a local level, onshore and offshore winds give rise to notable changes in temperature and humidity that affect evaporation. During the snow-free period in Churchill, Manitoba, Rouse (1984) observed that the air temperature can drop by as much as 20°C within several hours when the wind shifts from offshore to onshore from Hudson Bay where the sea ice does not break up until mid-June and floating ice remains until mid-July. The onshore cold wind elevates evaporation and sensible heat fluxes while these fluxes decline when warm wind blows offshore. This feature is explained as follows. The surface is heated up by the summer sun, and the onshore wind creates a cold and relatively dry atmosphere immediately above the warm surface. "This imposed very large vertical temperature and humidity gradients that drove large sensible and evaporative heat fluxes" (Rouse 2008). Ground heat flux is also affected at these locations where soil temperature is low. With warm offshore wind, a large soil temperature gradient develops to enhance ground heat input, but the cold onshore wind imposes only a small temperature gradient and hence low heat flux into the ground. Rouse (1984) suggested that the onshore winds permit the permafrost to persist in the coastal zone at latitudes as low as 60°N.

In the autumn and early winter, the land is frozen but there may still be open water areas as sea ice formation lags the cooling of the land. The wind that passes over the water picks up moisture but, as it reaches the cold land, the air is chilled and the moisture-holding capacity is suddenly reduced. This produces the "lake-effect snowfall" which is commonly experienced in the Great Lakes area of the St. Lawrence River drainage at the Canada–US border. It is surmised that such an effect would occur where the prevailing wind deposits snow on the downwind side of large water bodies like the Hudson and James bays. Then, the land on the leeward side of the bays would have more snowfall than the areas upwind of the bays. Table 2 lists the mean monthly snowfall and monthly total precipitation of two stations: Kuujjuarapik, Quebec, on the eastern side of Hudson Bay (55.28°N, 77.75°W, elevation 10 m) and Churchill, Manitoba, on the upwind side (58.73°N, 94.05°W, elevation 29 m). Although it is well known that snowfall suffers underestimation (Goodison 1978), total snowfall for October–December averages 131 cm for Kuujjuarapik, which is higher than the 90 cm for Churchill (to obtain an estimate of the snow water equivalent value, multiply the snowfall by 0.1). It is plausible that the difference in total snowfall is at least partly attributed to lake-effect snow.

Sea ice

The formation of sea ice shuts off the warming effect of the sea so that, once ice is formed, contrasts in winter air temperature between the land and the sea are reduced. On the other hand, lingering sea ice prolongs the cooling season of coastal lands. Maxwell (1981) noted that, because of the sea ice, Resolution Island at the Atlantic entrance of Hudson Strait (61.6°N, 64.63°W, elevation 369 m) is the warmest in winter but its July temperature is the lowest among the islands of the Canadian Arctic Archipelago. The mean annual temperature range at this coastal station is only 22°C. The presence of sea ice also can affect cloud formation, hence the radiation input for snowmelt and evaporation. Sim (1957) reported for Ellesmere Island that "frequently during the summer of 1955 thick banks of low stratus cloud were observed over Fosheim peninsula whereas the skies over the ice-covered area of Eureka Sound, Greely Fiord, and Canyon Fiord were clear. Later, when pools of melt water and open leads appeared, clouds began to form over the channels". High radiation input may compensate for the reduced sensible heat flux due to the presence of ice, and snow sublimation can proceed when air temperature is below freezing. In the Arctic Islands where the many channels and inlets are ice-bound until late summer, the development of leads or the fragmentation of sea ice induces the formation of sea fog (Figure 2). Drifted ashore, these fogs and low clouds cut down solar radiation and reduce evaporation. They also provide "trace" precipitation events that are too small in quantity to be registered

Table 2 | Mean monthly snowfall and monthly total precipitation of two stations: Kuujjuarapik, Quebec, on the eastern side of Hudson Bay (55.28°N, 77.75°W) and Churchill, Manitoba, on the upwind side of the Bay (58.73°N, 94.05°W)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Precipitation (mm)												
Churchill	16.9	15.7	16.1	19.0	31.9	44.3	56.0	68.3	63.4	46.9	33.1	20.0
Kuujjuarapik	27.6	22.2	20.5	23.6	35.1	60.0	79.4	91.5	102.7	80.9	64.3	40.8
Snowfall (cm)												
Churchill	19.8	18.3	18.3	19.8	15.4	3.4	0	0	0.6	28.7	37.0	24.2
Kuujjuarapik	28.7	22.6	19.1	17.8	14.2	4.7	0.1	0	3.4	32.3	56.1	42.4

Figure 2 | Coastal fog formed when sea ice is dissipated in late summer along the shore of Resolute, Cornwallis Island. The fog produces "trace" precipitation and reduces radiation input to limit evaporation on land.

by conventional gauges. Their high frequencies in some years can render these events a major contributor to total summer precipitation in a polar desert environment (Woo & Steer 1979).

Over a long time span, the persistent presence of sea ice may influence tundra vegetation growth. "The cool and cloudy central polar pack ice climate bulges almost unimpeded into the low-lying islands of the northwest and north-central (Queen Elizabeth Islands). This region has the least vascular plant diversity and is dominated almost entirely by herbaceous species" (Edlund & Alt 1989). The scarcity or absence of vegetation diminishes transpiration water loss.

The overall extent of Arctic sea ice has undergone accelerated decline in 2007 (Figure 3) and 2008. By mid-September 2007, ice covered only $4.13 \times 10^6\,\mathrm{km}^2$. In the following September, the ice-covered area was $4.52 \times 10^6\,\mathrm{km}^2$ but ice was less thick (<1 m at places) as first-year ice gained prominence at the expense of multi-year ice. Ocean surface with open water in the summer enhances the growth of first-year ice in the winter, as was noted by Macdonald *et al.* (2002) during the SHEBA (Surface Heat Budget of the Arctic) investigation. Kwok *et al.* (2009) found that, between 2003 and 2008, the areal coverage of multi-year ice was reduced by $>40\%$ and the ice thinned by about 0.6 m. With over 70% being first-year ice, the total ice volume decreased in 2008 but the freshwater content in the Arctic Ocean increased (McPhee *et al.* 2009). Most of the ice loss was over the Beaufort, Chukchi, East Siberan, Laptev and Kara seas. Commenting on the 2007 ice retreat, Comiso *et al.* (2008) proposed such factors as a warm winter that inhibited ice growth and high temperatures in spring and summer to accelerate melt. An unusually persistent southerly wind from June to August in the Beaufort and Chukchi seas may push the ice to the north (Overland *et al.* 2008) but cyclonic wind centred about 78°N 140°E could forestall further retreat (Comiso *et al.* 2008). Influx of Pacific water through the Bering Sea can be related to the large ice loss. Shimada *et al.* (2006) proposed that delayed ice formation in early winter allows efficient coupling of anticyclonic wind to the upper ocean which increases the flux of warm Pacific water to cause catastrophic ice reduction on the Beaufort Sea.

A decrease of sea ice extent in late autumn has effects on the land. The atmosphere apparently "remembers" the influence of a reduced summer ice cover and extends the effects into the following autumn and winter: warming and destabilization of the lower troposphere, increased cloudiness and weakening of the polar jet stream (Francis *et al.* 2009). The shrunken sea ice extent has effects on the land. A reduced ice cover on the Barents and Kara seas generates a stationary Rossby wave that amplifies the winter high pressure over Siberia and spreads anomalously cold conditions over Eurasia (Honda *et al.* 2009).

The extreme ice loss event of 2007 motivated Lawrence *et al.* (2008) to simulate the impact of similar events on land

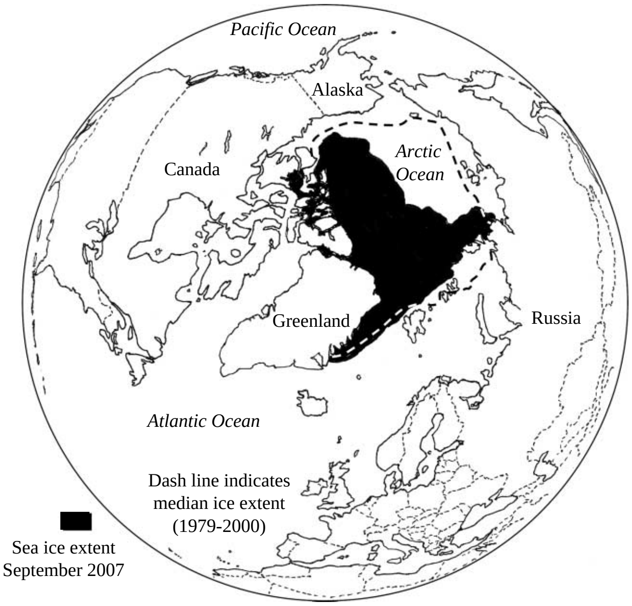

Figure 3 | September minimum extent of Arctic sea ice for 2007 compared with median extent for 1979–2000 (based on images from National Snow and Ice Data Center, Sea Ice Extents file for Google Earth Animations http://nsidc.org/data/virtual_globes).

warming and permafrost degradation. They found that, under climatic change scenarios, the 21st century warming trend can be augmented by a factor of 3.5 during periods of rapid sea ice loss, and that the accelerated warming signal can penetrate up to 1,500 km inland. They speculated that rapid warming, with downwelling of summer heat exceeding winter cooling, will cause dramatic degradation of permafrost. Should their model results be realistic, the hydrologic cycle of many coastal basins would be seriously affected.

Tides, waves and coastal storms

Rise and fall of the sea level exert influences on the hydrology of coastal zones. Extreme cases of coastal flooding are associated with tsunamis. The Alaskan earthquake of 1964, centred 125 km east of Anchorage, for instance, produced a series of tsunamis, one of which reached 20 m high and caused devastation as far as California. More regularly, tides lead to cyclical fluctuations of the base level (the lowest level to which a river can erode its bed) for coastal streams. Streamflows decelerate or accelerate repeatedly in conjunction with tidal rise and fall. Tides give rise to the backwater effect (i.e. tides retard the river flow and push up the river level) at the lowest reaches of rivers on coastal plains. In extreme cases, rivers reverse their flow direction when the tide arrives and recedes, as happens routinely at the mouth of St. John River, New Brunswick, Canada (Figure 4) due to the largest tidal range on Earth in the Bay of Fundy (spring tides are about 12 m). Advantage has been taken of the power from such flow reversal to generate hydroelectricity at the mouth of Annapolis River, Nova Scotia. Tides also can produce a repeated rhythm of flooding and drying of coastal marshes and tidal flats, such as those along the Cook Inlet, Alaska, where the tide reaches 11 m amplitude. Rise of the water table in the marshes usually lags behind the tidal rise while the ebb tide steepens the hydraulic gradient along the shoreline to hasten drainage from the marshes.

Major storms are a source of moisture input to drainage basins. During periods without sea ice cover, heat and moisture fluxes from the water contribute to cyclogenesis (development or strengthening of cyclonic circulation) along the coasts. This often happens at the southern Beaufort Sea which typically becomes open in late June till mid-October. Most of the storms of the northern Mackenzie Basin come from the Arctic, in the direction of the Beaufort Sea (Hudak & Young 2002). However, the 1994 summer sea ice cover over the Arctic Ocean was exceptionally low. It was likely that the more extensive open water condition diverted the storm tracks to lie close to the baroclinic zone along the sea ice margin (Stewart *et al.* 2002, p. 274). This led to an anomalously low frequency of Arctic storms over the Basin. It was suggested that the extremely low and early discharge of the Mackenzie River in the 1994–5 water year may be related to the infrequent occurrence of Arctic storms (Hudak & Young 2002). However, it is also possible that, instead of one phenomenon leading to the other, they were responding together to some large-scale circulation that influenced both the Basin and Beaufort Sea.

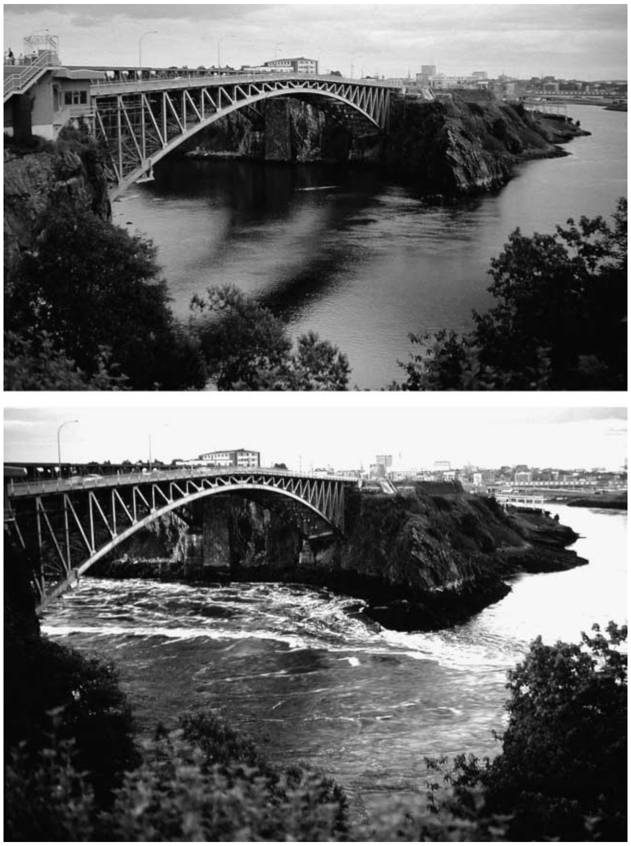

Figure 4 | Reversing Falls at the mouth of St. John River, New Brunswick: (top) tide water from the Bay of Fundy moves up river, from right to left, on 26 July 1998; (bottom) several hours later at 21:00 h, river water tumbles down the Falls from left to right as the tide recedes.

Local storms may be spawned from leads or large cracks in sea ice, and from polynyas which are areas of open water and new ice with thickness of < 0.3 m (see Figure 2 in Mysak *et al.* (1990)). Cyclogenesis along open water margins of polynyas such as the North Water between Greenland and Baffin–Ellesmere islands enhance regional storm development in Eastern Arctic of Canada (Rouse 1993, p. 74). Local or larger-scale cyclones that travel up the Arctic coast in winter can bring unseasonably warm conditions that cause mid-winter snowmelt. Gilbert & McKenna-Neuman (1988) reported an event on 10 January 1985 when strong winds ($36\,\mathrm{km\,h^{-1}}$), low humidity (46%) and air temperatures that reached − 1°C led to snowmelt and flooding of the Weasel River sandur on Baffin Island. Similar previously documented events are listed by Gilbert & McKenna-Neuman (1988). Storms that track northward between the Queen Elizabeth Islands and Greenland encounter high mountains. Orographic uplift and condensation produce heavy precipitation on high elevations, which is important in nourishing the glaciers.

A combination of open water and storms can result in high waves and large storm surges. Low-lying coastal belts are subject to devastation by storm surges which are due to wind stress on the water surface to produce a large displacement of water. Surges, accompanied by waves and tides, can cause extensive flooding of coastal areas. Like many large river deltas, the Mackenzie Delta has myriads of lakes that are periodically linked to the main channel, depending on the river level. Storm surges generate a strong backwater effect that causes the river to overflow the sills and flood the delta lakes that otherwise are not connected to the main channel. A storm on the Beaufort Sea produced a surge on 15–18 September 1985. Marsh & Schmidt (1993) traced the backwater effect up to Point Separation, over 200 km inland (Figure 5). They estimated that, of the 132 lakes studied in Marsh & Hey (1989), about 12% are "no-closure" lakes that are connected to the channel in summer. When this September storm caused the water level to rise by 0.78 m to reach 2.28 m asl, an additional 17% of these lakes would have been flooded by the Mackenzie, thus affecting the sediment, chemical, nutrient and energy inputs to the lakes.

Along the shoreline, ice piled up by storms in the cold season protects the beaches and the coastal bluffs from wave erosion. In the thaw season, coastal bluffs rich in ground ice are subject to degradation through: (1) surface wash by runoff from snowmelt and rainfall, (2) debris sliding as the active layer material thaws and slides or flows down the bluff face, (3) slumping when massive ground ice melts and slumps, and (4) thermo-erosion when the bluffs are notched by waves through ice-melt and mechanical action, followed by collapse of the overhang blocks (Carter *et al.* 1987). The resulting retreat rate reaches about 3 m per year, as noted by Grigoriev *et al.* (2004) for the Beaufort, Chuckchi, Kara, Laptev and East Siberian coasts, though shoreline retreat is greatly reduced for the coasts undergoing isostatic rebound. The eroded materials are a source of sediment to the ocean.

On a limited scale, as observed along the Svalbard coasts, autumn and winter storms can build up barrier ridges across the mouth of small basin outlets, if the shores are not protected by sea ice (Akerman 2008). These ridges, 4–6 m higher than the sea level and composed of beach gravels and sand, freeze over and act as impermeable dams in the spring. Impoundment of runoff behind the dam raises the lake level, floods the wetlands and alters the drainage pattern. In recent years, ice has been absent in the fiords and barrier ridges are formed frequently. One hydrologic consequence is that spring peak discharge to the sea is delayed by several weeks, until the frigid ridges are breached.

Small basin water balance

The water balance of two small high latitude basins, one in a continental setting and the other exposed to maritime influence, offers an instructive comparison of the influence of continentality and of polar oceans on cold region hydrology. The Imnavait Creek basin on the North Slope of Alaska (68°28′N, 149°24′W, basin area $2.2\,\mathrm{km}^2$ and a relief of 25–75 m) was studied by Kane *et al.* (2004) and the Windy Creek basin on King George Island (62°13′S, 58°58′W, basin area $1.8\,\mathrm{km}^2$ and a relief of 0–50 m) was investigated by Flügel (1990). Imnavait Basin has a mid-winter temperature that averages − 20°C but can drop to − 40°C; Windy Basin has a winter temperature of − 0.8 to − 6.2°C but can rise to 3.8°C to cause mid-winter snowmelt. The summer temperature averages 6–10°C for

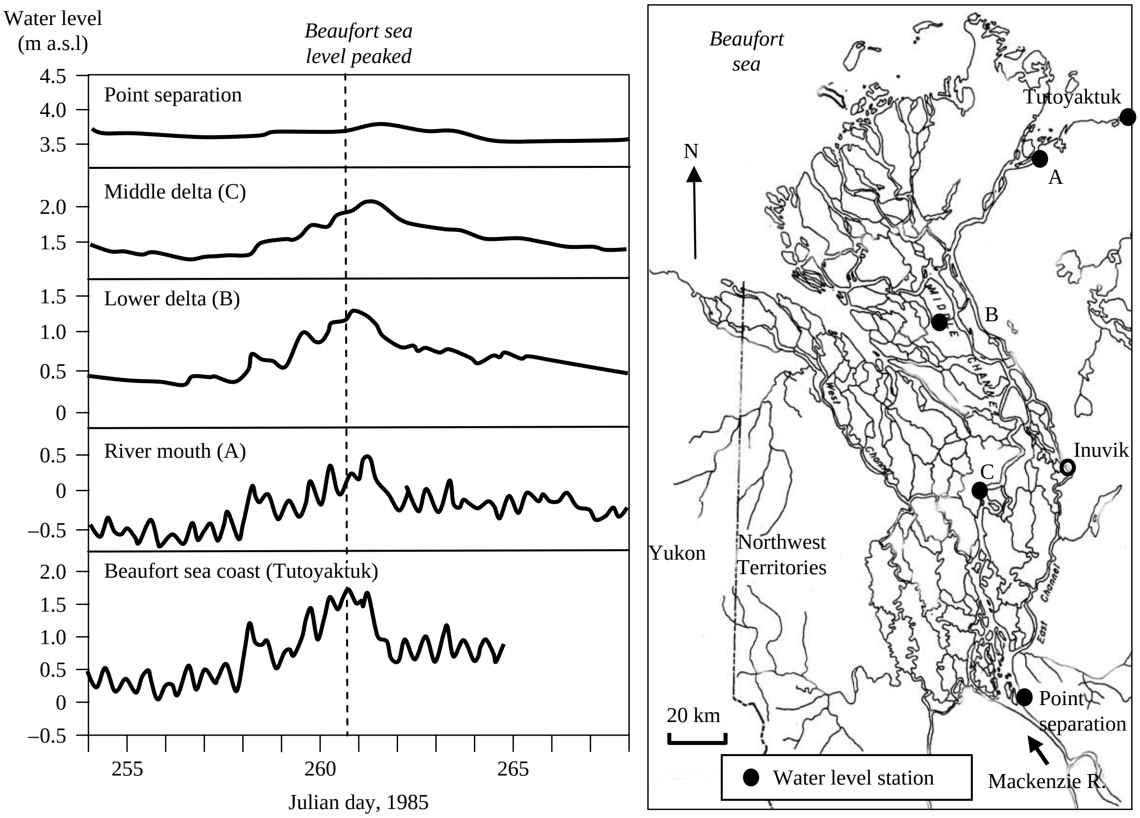

Figure 5 | Storm-surge-induced water level rise along the eastern channel of Mackenzie River from Point Separation at the upper end of Mackenzie Delta to the river mouth (after Marsh & Schmidt 1993).

Imnavait, and is − 3 to 1.9°C in the vicinity of Windy Basin (based on three nearby stations) but can drop to − 3.6°C. Windy Basin has an average wind speed of 6 m s^{-1} and a high relative humidity of 77–92%. Its soil (sand to coarse gravel derived from basalt and volcanic ash, with no visible soil horizons) is saturated throughout the thawed season but the water table in the soil of Imnavait Basin (0.2 m organic layer overlying glacial till) fluctuates more widely during the summer.

The water balance of Imnavait basin (1985–2003 average) shows winter snow accumulation of 120 mm SWE, summer precipitation of 241 mm, evaporation of 179 mm and runoff of 181 mm. The 1984–85 water balance of Windy Basin shows 727 mm of precipitation (not distinguished between snow and rain), evaporation of 58 mm and measured discharge of 662 mm, which could be 711 mm if the flow response to rainfall in late February is included. Several significant conclusions can be drawn from a comparison of the hydrology of these basins:

(1) In contrast to continentality, maritime influence generates higher precipitation and more frequent variation of the air temperature around the freezing point; thus the maritime basin has more winter melt and summer snowfall events than the interior basin.
(2) Duration of the main snowmelt period in spring is extended for the maritime basin because it has more residual snow and slower melt rates (due to low summer temperature and frequent cloudiness that reduces radiation input).
(3) Persistently high humidity under an oceanic climate suppresses evaporation; high precipitation and low evaporation enables continued soil saturation in the summer.
(4) With low evaporation, Windy Basin has a very high runoff to precipitation ratio.
(5) While both basins exhibit a nival (snowmelt) regime of streamflow, frequent summer precipitation engenders a stronger pluvial (rainfall) response in the Windy Creek discharge.

HYDROLOGIC INFLUENCE ON OCEANS

Hydrologic transfers from the land to the ocean include runoff from rivers and evaporative flux to the air which may then blow over the sea and deposit as precipitation. Terrestrial runoff input is important as it influences sea ice formation and decay, and the heat and water circulations of the polar ocean (Aagaard & Carmack 1989; Manak & Mysak 1989). The supply of freshwater has major impacts on thermohaline circulation in the Arctic Basin.

Groundwater flow to the ocean is limited, particularly in the presence of permafrost. Surface runoff reaches the ocean seas either from the coastal slopes or through river flow. Aagaard & Carmack (1989) estimated that > 3,300 km^3 of fresh water flows into the Arctic Ocean each year, 2,000 km^3 of which comes from four major rivers: the Ob (≈400 km^3/yr), the Yenisei (≈550 km^3/yr), the Lena (≈520 km^3/yr) and the Mackenzie (≈300 km^3/yr), all of which discharge directly into the Arctic Ocean. Other rivers such as the Yukon (≈200 km^3/yr) and rivers that drain into Hudson Bay eventually send part or most of their flows into northern waters. River discharge represents an estimated 38% of mean annual freshwater input to the Arctic Ocean (the other 30% is influx through the Bering Strait and 24% from net precipitation, according to Serreze *et al.* (2006)). Note that all the discharge figures are estimated and can be much refined by more flow measurements, particularly for the vast number of rivers in the immense northern areas that have never been gauged.

Most of the annual freshwater runoff occurs in the spring, as a result of rapid snowmelt. The Lena River, for example, releases 40% of its mean annual flow and 60% of its mean annual sediment to the Laptev Sea at the beginning of June when water and sediments move mostly beneath the shore-fast ice and some overflowing it (Wegner *et al.* 2005). The convection of heat in the high flow period is also notable. Liu *et al.* (2005) estimated that the Lena River has a heat flux of 6.07×10^9 GJ in July compared with 1.92×10^9 GJ in September. This river delivers a total heat flux of 14.28×10^9 GJ from June through September to the Laptev Sea, but the winter flux is small due to low discharge and low water temperature.

Much of the winter inflow from rivers is kept as a large pool of fresh or brackish water under landfast ice off river deltas, and some may be frozen and stored as ice in the nearshore zone (Macdonald *et al.* 1995). For example, the Laptev Sea landfast ice retains about a quarter of the annual total flow of the Lena and Yana rivers (Eicken *et al.* 2005). Sea ice growth rejects brine and, together with shoreward advection of saline water, renders the bottom water increasingly saline in the winter. Off the estuary of Churchill River, Manitoba, winter low flow forms a brackish plume riding on this saline bottom water. With spring snowmelt, river discharge produces a freshwater layer on the saline water or floods the landfast ice (Kuzyk *et al.* 2008). Walker (1973) observed a similar pattern during spring breakup in the Colville Delta, Alaska. He found that, in the pre-breakup period, the water beneath the ice in the subaqueous portion of the delta is as saline or more saline than the Arctic Ocean seawater, suggesting negligible freshwater inflow from Colville River in the winter (due to cessation of flow in this continuous permafrost environment). During breakup, flood water spreads seaward both over and under the sea ice, progressively replacing the saline water below the ice with freshwater. Carmack & Macdonald (2002) reported a similar breakup pattern for the Mackenzie Delta and noted that a rapid influx of freshwater under the fast ice produces geyser-like eruptions through cracks and holes in the ice.

Large river inflow enhances sea ice breakup at the nearshore zone (Searcy *et al.* 1996). While surface net radiative and heat fluxes are key factors controlling ice melt, Bareiss *et al.* (1999) acknowledged that the retreat of fast ice is accelerated by increased surface albedo and solar heating of the floodwater and the dirty ice surface after the water has drained. After breakup, the melting of sea ice and continued discharge of river water maintain a plume of freshwater, stratified above the saline water. It was noted that the presence of the freshwater layer supported by the Mackenzie River discharge helps to maintain heat below the stratification, which then leads to the demise of deep ice keels (Macdonald, pers. comm., 2010). The stratified structure is weakened in late summer and the fall as storms mix the top 10–20 m of water. A generalization of the seasonal rhythm of river inflow, ice cover development and salinity change of the Canadian Shelf of the Beaufort Sea is given by Carmack & Macdonald (2002).

Accompanying the river water are large quantities of sediments, organic carbon and nutrients, much of which is delivered during the spring peak flow season (Walker *et al.* 2008). Hasholt *et al.* (2006) presented a review on the transport of suspended sediments to the polar ocean and seas. They estimated that $324-884 \times 10^6$ t/yr of sediment is discharged to the Arctic Ocean, though this estimate was based on both measurements (56% of the data) and from other sources. A lot of sediment is discharged from large river mouths. Sediment plumes can extend hundreds of kilometres into the ocean and at the nearshore zone, though the plumes often include shelf sediments that are re-suspended from the sea bed (Carmack & Macdonald 2002). Coastal erosion also yields sediments from the land. The deposition of terrestrial materials from river and slope sources enables the seaward expansion of deltas.

Rivers bring organic carbon and nutrients to the sea. Stein & Macdonald (2004) presented a comprehensive treatment of the sources, fluxes and distribution of organic carbon in the Arctic Ocean. Nutrient fluxes influence the biological productivity of northern oceans. Kuzyk *et al.* (2008) concluded that, where the composition of river and sea water complements each other, their mixture offers greater availability of nutrients than either mass alone; but if the river water is deficient in particular nutrients, the addition of terrestrial inflow would dilute their concentration in the sea water. The availability of nutrients is not the sole consideration in phytoplankton growth, which is often limited by low light level due to a lingering ice cover and turbidity of river water in the spring season. Furthermore, the nutrient contribution from rivers has to be considered in relation to the non-fluvial sources. For example, currents from the Pacific Ocean that enter the Arctic Basin through the Bering Strait bring in much greater quantities of nutrients than the more localized inputs such as from river water of the Mackenzie.

Streamflow regulation in boreal areas may have limited impact on sea ice development offshore. LeBlond *et al.* (1996) analysed the possible effects of regulating river flow into the Hudson and James Bays, from Manitoba, Ontario and Quebec. They concluded that, with an increase in winter flow and an elimination of the spring freshet, the oceanographic effects may be felt only on a local scale, especially at the river mouths; but have little bearing on the North Atlantic thermohaline circulation or the climate of North America and Europe. In an exceptional case, an artificial release of a large pulse of freshewater is detectable in the ocean. Woo & Thorne (2003) examined the release of a huge volume of water from the Bennett Dam, British Columbia, in 1996, which cascaded down the Mackenzie River system to contribute to a large flood pulse in 1997. This river discharge was detected as a freshwater plume in the Beaufort Sea by the oceanographers of the SHEBA project (Macdonald *et al.* 1999).

DISCUSSIONS

Oceans exert influences on northern coastal zones either directly or indirectly through atmospheric circulation and amelioration of ground freeze. Interactions can occur at a range of scales, from teleconnection of hemispherical dimensions to drainage basins large and small, to individual estuaries and river channels. All aspects of hydrology, including precipitation, evaporation, runoff and long and short term changes in storage (snow, glacier, wetlands and lakes), can be affected. On the other hand, terrestrial hydrology produces feedbacks on polar seas and their ice cover, principally through the discharge of freshwater together with the sediments, organic carbon and nutrients it carries, and a mismatch between the warming of the land and the seas. While the overall scheme of ocean–hydrologic interactions can be generalized (Figure 6), the dynamics are not precisely known.

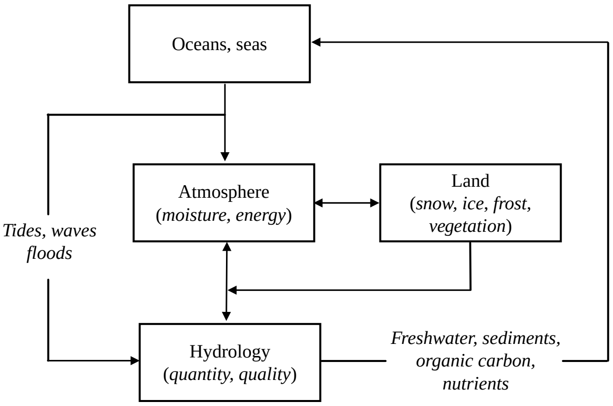

Figure 6 | Schematic of ocean–hydrology interactions showing feedback loops among ocean seas, atmosphere and terrestrial hydrosphere.

High latitudes are very susceptible to climate change, partly because of the long temporal overlap between the snow-covered period and the high-sun season so that any perturbation that reduces the duration of snow and ice cover will raise the surface radiation receipt by an order of magnitude (Woo & Ohmura 1997). Potential impacts on the Arctic cryospheric system have been described in ACIA (2005). Interactions between hydrological elements and the oceans may intensify under a changing climate. Ice conditions and Arctic weather are rendered less predictable, as is indicated by observations made by natives (Carmack & Macdonald 2008). Less sea ice may modify the existent energy and moisture fluxes, hence alter the coastal storm pattern and the basin water balance. Accelerated melting of glaciers which adds more water to the oceans, in concert with steric sea level change (i.e. variation of sea level due to changes in water temperature and salinity), will contribute to the global rise in sea level. A higher sea level will directly impact the coastal zone through flooding of existing lands, increased shoreline erosion and permafrost thaw, and re-arrangement of the coastal drainage configuration. Furthermore, Abeysirigunawardena & Walker (2008) warned that coastal hazards can be more severe if sea level fluctuations forced by climate variability are superimposed on the long term trend of sea level rise.

Change in river flow regime affects the magnitude and timing of freshwater input. Recent changes in annual flows are spatially varied, with reported increases for the large Eurasian rivers (Peterson *et al.* 2002; Yang *et al.* 2002) but decreasing trends for rivers in eastern Canada (Déry & Wood 2005). There is closer agreement regarding spring freshets: breakup is earlier in many major Arctic-flowing rivers but the peak magnitudes are lessened. These appear to be the response to a warming climate. More importantly, human intervention through increased water consumption and flow regulation along rivers usually generate stronger signals of hydrologic change than does the climate (Ye *et al.* 2003; Woo *et al.* 2008). As resource development accelerates in the north, human-induced changes raise the level of uncertainty regarding future hydrologic responses.

There are lots of unknowns concerning how various forcing and feedback mechanisms perform under the present climate and even greater uncertainties regarding how oceanic–atmospheric–hydrologic linkages will operate in a warming climate and its associated changed environments. The zone of direct contact between land and ocean is areally limited and the transfer of oceanic influence on hydrology is mainly through the atmosphere. Future studies have to consider such tripartite linkages. In addition to the physical processes involved, special attention should be given to the discord in the response and residence times, the capacity to store and the rate of transport of energy and moisture among the three media of land, water and air. The atmosphere is the most dynamic; it has the least capacity to store but the fastest rate of conveyance. The ocean, with its immense capacity to retain heat and water, is able to modulate the high frequency perturbations imposed by the atmosphere and inputs from the land. The variegated topography, geology and vegetation of the land have diverse storage capacity for water and energy and impose internal boundaries that complicate the rates of their transfer. These features create seasonal contrasts and gradients at the interfaces between the sea, land and atmosphere. Given our currently incomplete knowledge of the interactions and the pressing need to project future impacts, many fruitful research questions await investigation. Trans-disciplinary collaboration is essential but natural. After all, both oceanographers and hydrologists deal with the same substance, viz. water.

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