SUBMERSIBLES

See **MANNED SUBMERSIBLES, DEEP WATER; MANNED SUBMERSIBLES, SHALLOW WATER**

SUB-SEA PERMAFROST

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Introduction

Sub-sea permafrost, alternatively known as submarine permafrost and offshore permafrost, is defined as permafrost occurring beneath the seabed. It exists in continental shelves in the polar regions (**Figure 1**). When sea levels are low, permafrost aggrades in the exposed shelves under cold subaerial conditions. When sea levels are high, permafrost degrades in the submerged shelves under relatively warm and salty boundary conditions. Sub-sea permafrost differs from other permafrost in that it is relic, warm, and generally degrading. Methods used to investigate it include probing, drilling, sampling, drill hole log analyses, temperature and salt measurements, geological and geophysical methods (primarily seismic and electrical), and geological and geophysical models. Field studies are conducted from boats or, when the ocean surface is frozen, from the ice cover. The focus of this article is to review our understanding of sub-sea permafrost, of processes ocurring within it, and of its occurrence, distribution, and characteristics.

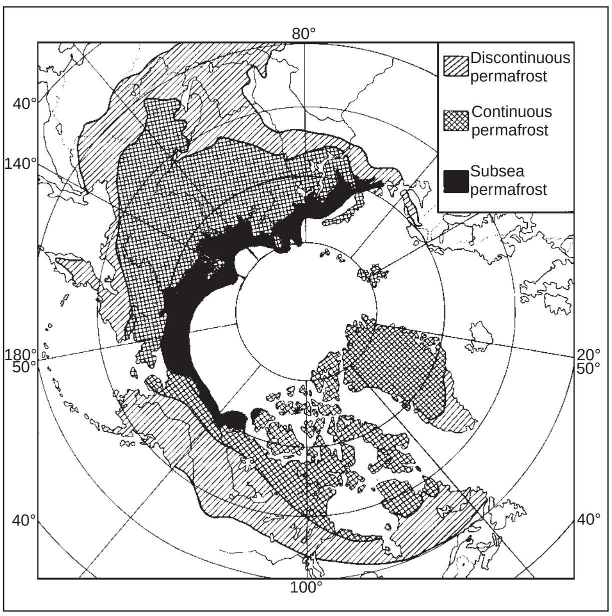

Figure 1 Map showing the approximate distribution of sub-sea permafrost in the continental shelves of the Arctic Ocean. The scarcity of direct data (probing, drilling, sampling, temperature measurements) makes the map highly speculative, with most of the distribution inferred from indirect measurements, primarily water temperature, salinity, and depth (100 m depth contour). Sub-sea permafrost also exists near the eroding coasts of arctic islands, mainlands, and where seabed temperatures remain negative. (Adapted from Pewe TL (1983). *Arctic and Alpine Research* 15(2):145–156 with the permission of the Regents of the University of Coloroado.)

Sub-sea permafrost derives its economic importance from current interests in the development of offshore petroleum and other natural resources in the continental shelves of polar regions. The presence and characteristics of sub-sea permafrost must be considered in the design, construction, and operation of coastal facilities, structures founded on the seabed, artificial islands, sub-sea pipelines, and wells drilled for exploration and production.

Scientific problems related to sub-sea permafrost include the need to understand the factors that control its occurrence and distribution, properties of warm permafrost containing salt, and movement of heat and salt in degrading permafrost. Gas hydrates that can occur within and under the permafrost are a potential abundant source of energy. As the sub-sea permafrost warms and thaws, the hydrates destabilize, producing gases that may be a significant source of global carbon.

Nomenclature

'Permafrost' is ground that remains below 0°C for at least two years. It may or may not contain ice. 'Ice-bearing' describes permafrost or seasonally frozen soil that contains ice. 'Ice-bonded' describes ice-bearing material in which the soil particles are mechanically cemented by ice. Ice-bearing and ice-bonded material may contain unfrozen pore fluid in addition to the ice. 'Frozen' implies ice-bearing or ice-bonded or both, and 'thawed' implies non-ice-bearing. The 'active layer' is the surface layer of sediments subject to annual freezing and thawing in areas underlain by permafrost. Where seabed temperatures are negative, a thawed layer ('talik') exists near the seabed. This talik is permafrost but does not contain ice because soil particle effects, pressure, and the presence of salts in the pore fluid can depress the freezing point 2°C or more. The boundary between a thawed region and ice-bearing permafrost is a phase boundary. 'Ice-rich' permafrost contains ice in excess of the soil pore spaces and is subject to settling on thawing.

Formation and Thawing

Repeated glaciations over the last million years or so have caused sea level changes of 100 m or more (**Figure 2**). When sea levels were low, the shallow continental shelves in polar regions that were not covered by ice sheets were exposed to low mean annual air temperatures (typically −10 to −25°C). Permafrost aggraded in these shelves from the exposed ground surface downwards. A simple conduction model yields the approximate depth (X) to the bottom of ice-bonded permafrost at time t, (eqn [1]).

$$X(t) = \sqrt{\frac{2K(T_e - T_g)t}{h}} \qquad [1]$$

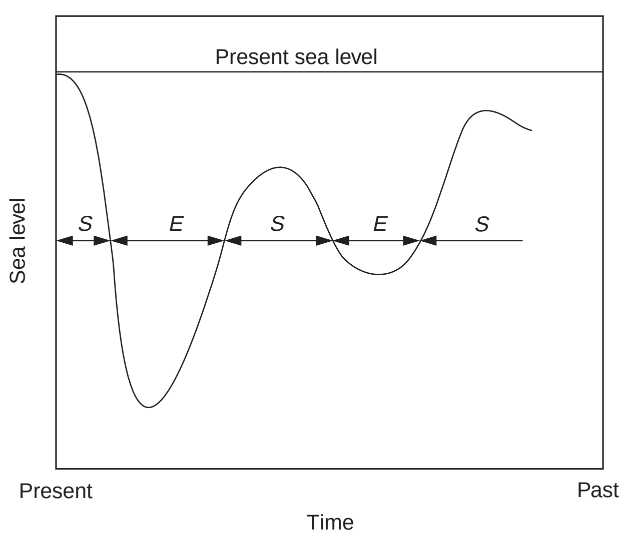

Figure 2 Schematic sea level curve during the last glaciation. The history of emergence (E) and submergence (S) can be combined with paleotemperature data on sub-aerial and sub-sea conditions to construct an approximate thermal boundary condition for sub-sea permafrost at any water depth.

K is the thermal conductivity of the ice-bonded permafrost, T_e is the phase boundary temperature at the bottom of the ice-bonded permafrost, T_g is the long-term mean ground surface temperature during emergence, and h is the volumetric latent heat of the sediments, which depends on the ice content. In eqn [1], K, h, and T_e depend on sediment properties. A rough estimate of T_g can be obtained from information on paleoclimate and an approximate value for t can be obtained from the sea level history (**Figure 2**). Eqn [1] overestimates X because it neglects geothermal heat flow except when a layer of ice-bearing permafrost from the previous transgression remains at depth. Timescales for permafrost growth are such that hundreds of meters of permafrost could have aggraded in the shelves while they were emergent (**Figure 3**).

Cold onshore permafrost, upon submergence during a transgression, absorbs heat from the seabed above and from the geothermal heat flux rising from below. It gradually warms (**Figures 4** and **5**), becoming nearly isothermal over timescales up to a few millennia (**Figure 4**, time t_3). Substantially longer times are required when unfrozen pore fluids are present in equilibrium with ice because some ice must thaw throughout the permafrost thickness for it to warm.

A thawed layer develops below the seabed and thawing can proceed from the seabed downward, even in the presence of negative mean seabed temperatures, by the influx of salt and heat associated with the new boundary conditions. Ignoring seabed erosion and sedimentation processes, the thawing rate at the top of ice-bonded permafrost during submergence is given by eqn [2].

$$\dot{X}_{top} = \frac{J_t}{h} - \frac{J_f}{h} \qquad [2]$$

J_t is the heat flux into the phase boundary from above and J_f is the heat flux from the phase boundary into the ice-bonded permafrost below. J_t depends on the difference between the long-term mean

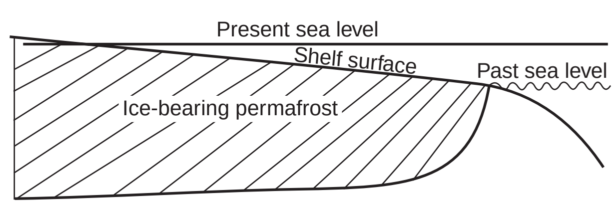

Figure 3 A schematic illustration of ice-bearing sub-sea permafrost in a continental shelf near the time of minimum sea level. Typical thicknesses at the position of the present shoreline would have been about 400–1000 m with shelf widths that are now typically 100–600 km.

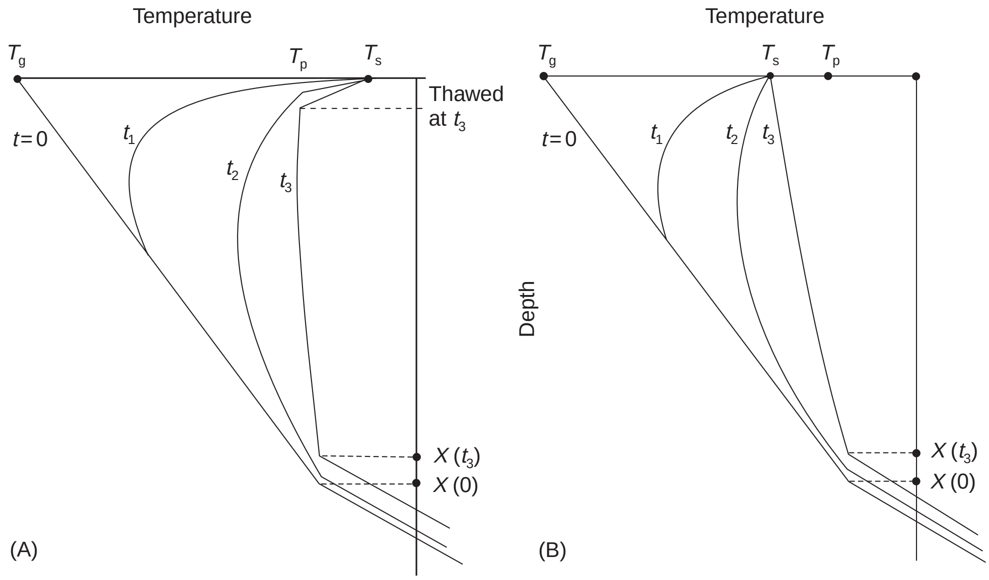

Figure 4 Schematic sub-sea permafrost temperature profiles showing the thermal evolution at successive times (t_1, t_2, t_3) after submergence when thawing occurs at the seabed (A) and when it does not (B). T_g and T_s are the long-term mean surface temperatures of the ground during emergence and of the seabed after submergence. T_p is the phase boundary temperature at the top of the ice-bonded permafrost. $X(0)$ and $X(t_3)$ are the depths to the bottom of ice-bonded permafrost at times $t=0$ and t_3. (Adapted with permission from Lachenbruch AH and Marshall BV (1977) Open File Report 77-395. US Geological Survey, Menlo Park, CA.)

temperature at the seabed, T_s, and phase boundary temperature, T_p, at the top of the ice-bonded permafrost. For $J_t = J_f$, the phase boundary is stable. For $J_t < J_f$, refreezing of the thawed layer can occur from the phase boundary upward. For thawing to occur, T_s must be sufficiently warmer than T_p to make $J_t > J_f$. T_s is determined by oceanographic conditions (currents, ice cover, water salinity, bathymetry, and presence of nearby rivers). T_p is determined by hydrostatic pressure, soil particle effects, and salt concentration at the phase boundary (the combined effect of *in situ* pore fluid salinity, salt transport from the seabed through the thawed layer, and changes in concentration as a result of freezing or thawing).

Nearshore at Prudhoe Bay, $\dot{X}_{top}$ varies typically from centimeters to tens of centimeters per year while farther offshore it appears to be on the order of millimeters per year. The thickness of the thawed layer at the seabed is typically 10 m to 100 m, although values of less than a meter have been observed. At some sites, the thawed layer is thicker in shallow water and thinner in deeper water.

Sub-sea permafrost also thaws from the bottom by geothermal heat flow once the thermal disturbance of the transgression penetrates there. The approximate thawing rate at the bottom of the ice-bonded permafrost is given by eqn [3].

$$\dot{X}_{bot} \cong \frac{J_g}{h} - \frac{J_f'}{h} \qquad [3]$$

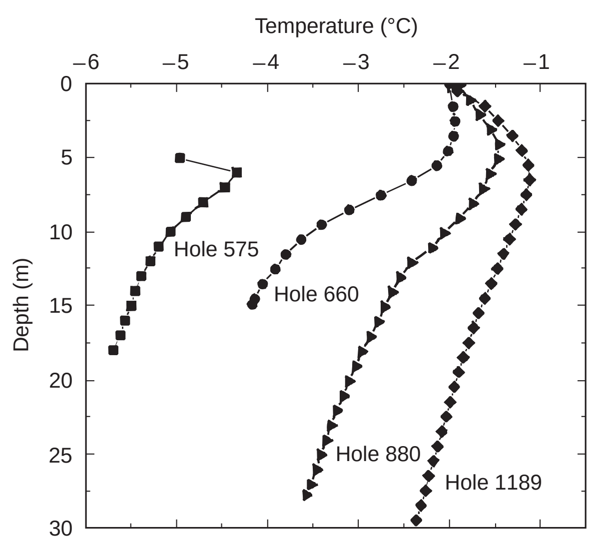

Figure 5 Temperature profiles obtained during the month of May in sub-sea permafrost near Barrow, Alaska, showing the thermal evolution with distance (equivalently time) offshore. Hole designation is the distance (m) offshore and the shoreline erosion rate is about 2.4 m y^{-1}. Sea ice freezes to the seabed within 600 m of shore. (Adapted from Osterkamp TE and Harrison WD (1985) Report UAGR-301. Fairbanks, AK: Geophysical Institute, University of Alaska.)

J_g is the geothermal heat flow entering the phase boundary from below and J_f' is the heat flow from the phase boundary into the ice-bonded permafrost above. J_f' becomes small within a few millennia except when the permafrost contains unfrozen pore fluids. $\dot{X}_{bot}$ is typically on the order of centimeters per year. Timescales for thawing at the permafrost table and base are such that several tens of thousands of years may be required to completely thaw a few hundred meters of sub-sea permafrost.

Modeling results and field data indicate that impermeable sediments near the seabed, low T_s, high ice contents, and low J_g favor the survival of ice-bearing sub-sea permafrost during a transgression. Where conditions are favorable, substantial thicknesses of ice-bearing sub-sea permafrost may have survived previous transgressions.

Characteristics

The chemical composition of sediment pore fluids is similar to that of sea water, although there are detectable differences. Salt concentration profiles in thawed coarse-grained sediments at Prudhoe Bay (**Figure 6**) appear to be controlled by processes occurring during the initial phases of submergence. There is evidence for highly saline layers within ice-bonded permafrost near the base of gravels overlying a fine-grained sequence both onshore and offshore. In the Mackenzie Delta region, fluvial sand units deposited during regressions have low salt concentrations (**Figure 7**) except when thawed or when lying under saline sub-sea mud. Fine-grained mud sequences from transgressions have higher salt concentrations. Salts increase the amount of unfrozen pore fluids and decrease the phase equilibrium temperature, ice content, and ice bonding. Thus, the sediment layering observed in the Mackenzie Delta region can lead to unbonded material (clay) between layers of bonded material (fluvial sand).

Thawed sub-sea permafrost is often separated from ice-bonded permafrost by a transition layer of ice-bearing permafrost. The thickness of the ice-bearing layer can be small, leading to a relatively sharp (centimeters scale) phase boundary, or large, leading to a diffuse boundary (meters scale). In general, it appears that coarse-grained soils and low salinities produce a sharper phase boundary and fine-grained soils and higher salinities produce a more diffuse phase boundary.

Sub-sea permafrost consists of a mixture of sediments, ice, and unfrozen pore fluids. Its physical and mechanical properties are determined by the individual properties and relative proportions. Since ice and unfrozen pore fluid are strongly temperature dependent, so also are most of the physical and mechanical properties.

Ice-rich sub-sea permafrost has been found in the Alaskan and Canadian portions of the Beaufort Sea and in the Russian shelf. Thawing of this permafrost can result in differential settlement of the seafloor that poses serious problems for development.

Processes

Submergence

Onshore permafrost becomes sub-sea permafrost upon submergence, and details of this process play a major role in determining its future evolution. The rate at which the sea transgresses over land is determined by rising sea levels, shelf topography, tectonic setting, and the processes of shoreline erosion, thaw settlement, thaw strain of the permafrost, seabed erosion, and sedimentation. Sea levels on the polar continental shelves have increased more than 100 m in the last 20 000 years or so. With shelf widths of 100–600 km, the average shoreline retreat rates would have been about 5 to 30 m y^{-1}, although maximum rates could have been much larger. These average rates are comparable to areas with very rapid shoreline retreat rates observed today on the Siberian and North American shelves. Typical values are 1–6 m y^{-1}.

It is convenient to think of the transition from sub-aerial to sub-sea conditions as occurring in five regions (**Figure 8**) with each region representing different thermal and chemical surface boundary conditions. These boundary conditions are successively applied to the underlying sub-sea permafrost during a transgression or regression. Region 1 is the onshore permafrost that forms the initial condition for sub-sea permafrost. Permafrost surface temperatures range down to about -15°C under current sub-aerial conditions and may have been 8–10°C colder during glacial times. Ground water is generally fresh, although salty lithological units may exist within the permafrost as noted above.

Region 2 is the beach, where waves, high tides, and resulting vertical and lateral infiltration of sea water produce significant salt concentrations in the active layer and near-surface permafrost. The active layer and temperature regime on the beach differ from those on land. Coastal banks and bluffs are a trap for wind-blown snow that often accumulates in insulating drifts over the beach and adjacent ice cover.

Region 3 is the area where ice freezes to the seabed seasonally, generally where the water depth is less than about 1.5–2 m. This setting creates unique thermal boundary conditions because, when the ice freezes into the seabed, the seabed becomes conductively coupled to the atmosphere and thus very cold. During summer, the seabed is covered with shallow, relatively warm sea water. Salt concentrations at the seabed are high during winter because of salt rejection from the growing sea ice and restricted circulation under the ice, which eventually freezes into the seabed. These conditions create highly saline brines that infiltrate the sediments at the seabed.

Region 4 includes the areas where restricted under-ice circulation causes higher-than-normal sea water salinities and lower temperatures over the

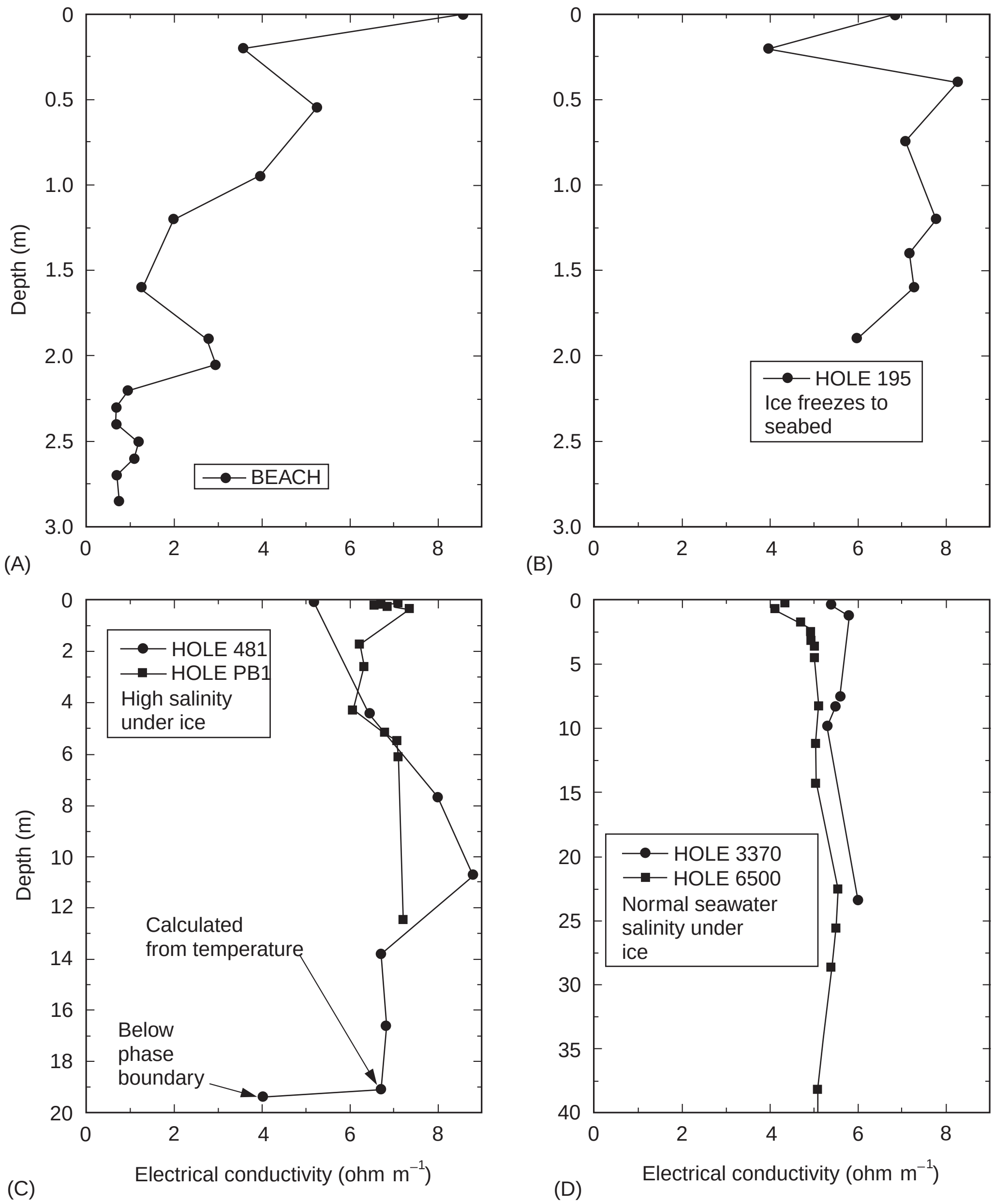

Figure 6 Electrical conductivity profiles in thawed, relatively uniform coarse-grained sediments. The holes lie along a line extending offshore near the West Dock at Prudhoe Bay, Alaska except for PB1, which is in the central portion of Prudhoe Bay. Hole designation is the distance (m) offshore. On the beach (A), concentrations decrease by a factor of 5 at a few meters depth. There are large variations with depth and concentrations may be double that of normal seawater where ice freezes to the seabed (B) and where there is restricted circulation under the ice (C). Farther offshore (D), the profiles tend to be relatively constant with depth with concentrations about the same or slightly greater than the overlying seawater. (Adapted from Iskandar IK *et al.* (1978) *Proceedings, Third International Conference on Permafrost*, Edmonton, Alberta, Canada, pp. 92–98. Ottawa, Ontario: National Research Council.)

sediments. The ice does not freeze to the seabed or only freezes to it sporadically. The existence of this region depends on the ice thickness, on water depth, and on flushing processes under the ice. Strong currents or steep bottom slopes may reduce its extent or eliminate it.

The setting for region 5 consists of normal sea water over the seabed throughout the year. This results in relatively constant chemical and thermal boundary conditions.

There is a seasonal active layer at the seabed that freezes and thaws annually in both regions 3 and 4. The active layer begins to freeze simultaneously with the formation of sea ice in shallow water. Brine drainage from the growing sea ice increases the water salinity and decreases the temperature of the water at the seabed because of the requirement for phase equilibrium. This causes partial freezing of the less saline pore fluids in the sediments. Thus, it is not necessary for the ice to contact the seabed for the seabed to freeze. Seasonal changes in the pore fluid salinity show that the partially frozen active layer redistributes salts during freezing and thawing, is infiltrated by the concentrated brines, and influences the timing of brine drainage to lower depths in the sediments. These brines, derived from the growth of sea ice, provide a portion or all of the salts required for thawing the underlying sub-sea permafrost in the presence of negative sediment temperatures.

Depth to the ice-bonded permafrost increases slowly with distance offshore in region 3 to a few meters where the active layer no longer freezes to it (**Figure 8**) and the ice-bonded permafrost no longer couples conductively to the atmosphere. This allows the permafrost to thaw continuously throughout the year and depth to the ice-bonded permafrost increases rapidly with distance offshore (**Figure 8**).

The time an offshore site remains in regions 3 and 4 determines the number of years the seabed is subjected to freezing and thawing events. It is also the time required to make the transition from sub-aerial to relatively constant sub-sea boundary conditions. This time appears to be about 30 years near Lonely, Alaska (about 135 km southeast of Barrow) and about 500–1000 years near Prudhoe Bay, Alaska. **Figure 9** shows variations in the mean seabed temperatures with distance offshore near Prudhoe Bay where the shoreline retreat rate is about $1\,\mathrm{m\,y^{-1}}$.

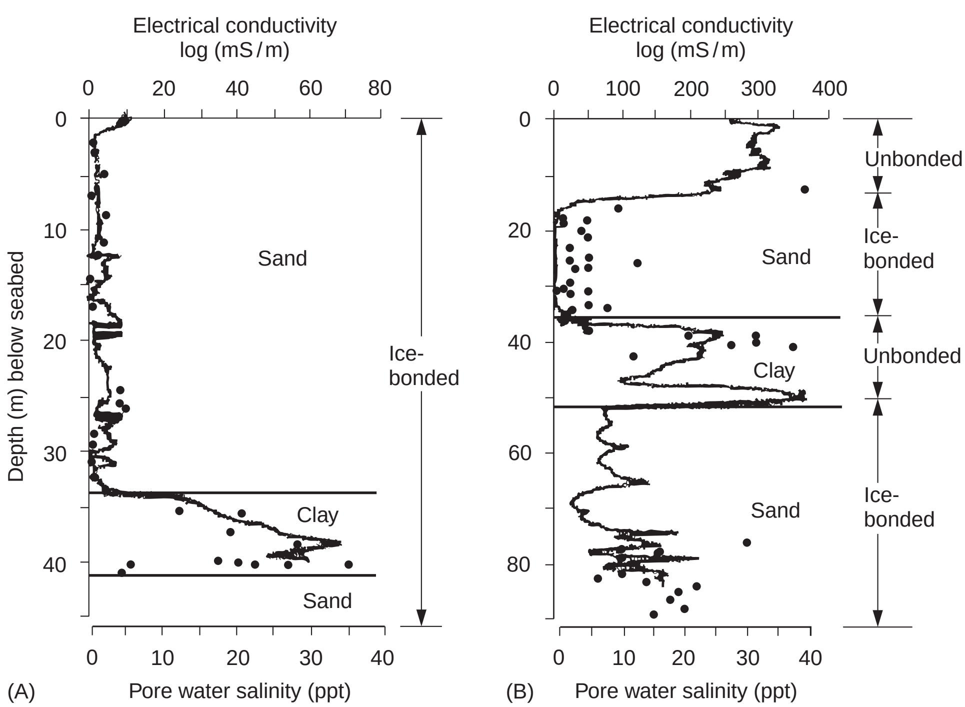

Figure 7 Onshore (A) and offshore (B) (water depth 10 m) electrical conductivity log (lines) and salinity (dots) profiles in the Mackenzie Delta region. The onshore sand–clay–sand sequence can be traced to the offshore site. Sand units appear to have been deposited under sub-aerial conditions during regressions and the clay unit under marine conditions during a transgression. At the offshore site, the upper sand unit is thawed to the 11 m depth. (Adapted with permission from Dallimore SR and Taylor AE (1994). *Proceedings, Sixth International Conference on Permafrost*, Beijing, vol. 1, pp.125–130. Wushan, Guangzhou, China: South China University Press.)

The above discussion of the physical setting does not incorporate the effects of geology, hydrology, tidal range, erosion and sedimentation processes, thaw settlement, and thaw strain. Regions 3 and 4 are extremely important in the evolution of sub-sea permafrost because the major portion of salt infiltration into the sediments occurs in these regions. The salt plays a strong role in determining T_p and, thus, whether or not thawing will occur.

Heat and Salt Transport

The transport of heat in sub-sea permafrost is thought to be primarily conductive because the observed temperature profiles are nearly linear below the depth of seasonal variations. However, even when heat transport is conductive, there is a coupling with salt transport processes because salt concentration controls T_p. Our lack of understanding of salt transport processes hampers the application of thermal models.

Thawing in the presence of negative seabed temperatures requires that T_p be significantly lower than T_s, so that generally salt must be present for thawing to occur. This salt must exist in the permafrost on submergence and/or be transported from the seabed to the phase boundary. The efficiency of salt transport through the thawed layer at the seabed appears to be sensitive to soil type. In clays, the salt transport process is thought to be diffusion, a slow process; and in coarse-grained sands and gravels, pore fluid convection, which (involving motion of fluid) can be rapid.

Diffusive transport of salt has been reported in dense overconsolidated clays north of Reindeer Island offshore from Prudhoe Bay. Evidence for convective transport of salt exists in the thawed coarse-grained sediments near Prudhoe Bay and in the layered sands in the Mackenzie Delta region. This includes rapid vertical mixing as indicated by large seasonal variations in salinity in the upper 2 m of sediments in regions 3 and 4 and by salt concentration profiles that are nearly constant with depth and decrease in value with distance offshore in region 5 (**Figures 5–7**). Measured pore fluid pressure profiles (**Figure 10**) indicate downward fluid motion. Laboratory measurements of downward brine drainage velocities in coarse-grained sediments indicate that these velocities may be on the order of $100\,\mathrm{m\,y^{-1}}$.

The most likely salt transport mechanism in coarse-grained sediments appears to be gravity-

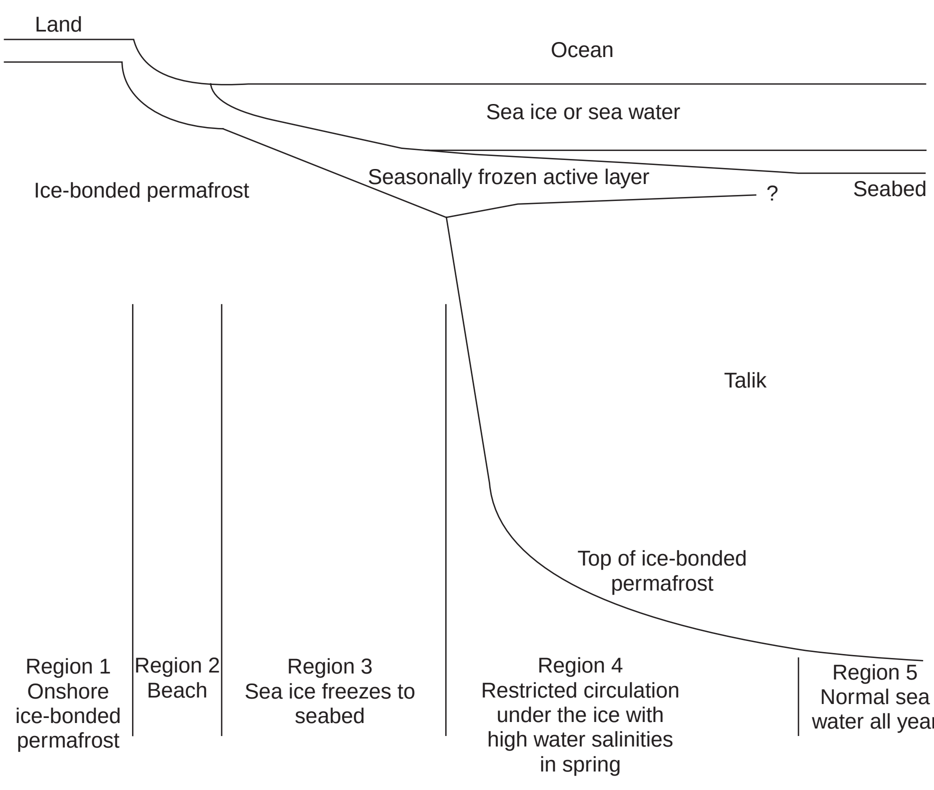

Figure 8 Schematic illustration of the transition of permafrost from sub-aerial to sub-sea conditions. There are five potential regions with differing thermal and chemical seabed boundary conditions. Hole 575 in **Figure 5** is in region 3 and the rest of the holes are in region 4. (Adapted from Osterkamp TE (1975) Report UAGR-234. Fairbanks, AK: Geophysical Institute, University of Alaska.)

driven convection as a result of highly saline and dense brines at the seabed in regions 3 and 4. These brines infiltrate the seabed, even when it is partially frozen, and move rapidly downward. The release of relatively fresh and buoyant water by thawing ice at the phase boundary may also contribute to pore fluid motion.

Occurrence and Distribution

The occurrence, characteristics, and distribution of sub-sea permafrost are strongly influenced by regional and local conditions and processes including the following.

1. Geological (heat flow, shelf topography, sediment or rock types, tectonic setting)
2. Meteorological (sub-aerial ground surface temperatures as determined by air temperatures, snow cover and vegetation)
3. Oceanographic (seabed temperatures and salinities as influenced by currents, ice conditions, water depths, rivers and polynyas; coastal erosion and sedimentation; tidal range)
4. Hydrological (presence of lakes, rivers and salinity of the ground water)
5. Cryological (thickness, temperature, ice content, physical and mechanical properties of the onshore permafrost; presence of sub-sea permafrost that has survived previous transgressions; presence of ice sheets on the shelves)

Lack of information on these conditions and processes over the long timescales required for permafrost to aggrade and degrade, and inadequacies in the theoretical models, make it difficult to formulate reliable predictions regarding sub-sea permafrost. Field studies are required, but field data are sparse and investigations are still producing surprising results indicating that our understanding of sub-sea permafrost is incomplete.

Pechora and Kara Seas

Ice-bonded sub-sea permafrost has been found in boreholes with the top typically up to tens of meters below the seabed. In one case, pure freshwater ice was found 0.3 m below the seabed, extending to at least 25 m. These discoveries have led to difficult design conditions for an undersea pipeline that will cross Baydaratskaya Bay transporting gas from the Yamal Peninsula fields to European markets.

Laptev Sea

Sea water bottom temperatures typically range from $-0.5°C$ to $-1.8°C$, with some values colder than $-2°C$. A 300–850 m thick seismic sequence has been found that does not correlate well with regional tectonic structure and is interpreted to be ice-bonded permafrost. The extent of ice-bonded permafrost appears to be continuous to the 70 m isobath and widespread discontinuous to the 100 m isobath. Depth to the ice-bonded permafrost ranges from 2 to 10 m in water depths from 45 m to the shelf edge. Deep taliks may exist inshore of the 20 m isobath. A shallow sediment core with ice-bonded material at its base was recovered from a water depth of 120 m. Bodies of ice-rich permafrost occur under shallow water at the locations of recently eroded islands and along retreating coastlines.

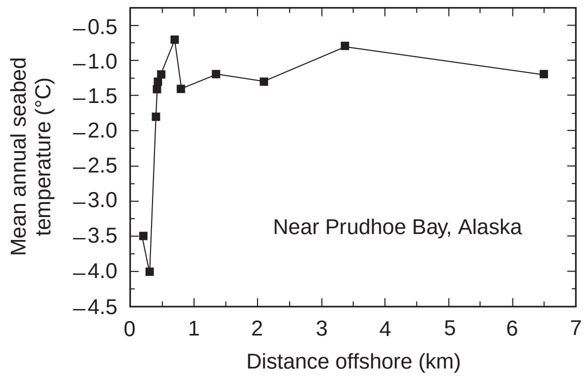

Figure 9 Variation of mean annual seabed temperatures with distance offshore. Along this same transect, about 6 km offshore from Reindeer Island in 17 m of water, the mean seabed temperature was near $-1.7°C$. Data on mean annual seabed salinities in regions 3 and 4 do not appear to exist. (Adapted from Osterkamp TE and Harrison WD (1985) Report UAGR-301. Fairbanks, AK: Geophysical Institute, University of Alaska.)

Bering Sea

Sub-sea permafrost is not present in the northern portion except possibly in near-shore areas or where shoreline retreat is rapid.

Chukchi Sea

Seabed temperatures are generally slightly negative and thermal gradients are negative, indicating ice-bearing permafrost at depth within 1 km of shore near Barrow, Rabbit Creek and Kotzebue.

Alaskan Beaufort Sea

To the east of Point Barrow, bottom waters are typically $-0.5°C$ to $-1.7°C$ away from shore, shoreline erosion rates are rapid (1–10 m y^{-1}) and sediments are thick. Sub-sea permafrost appears to be thicker in the Prudhoe Bay region and thinner west of Harrison Bay to Point Barrow. Ice layers up

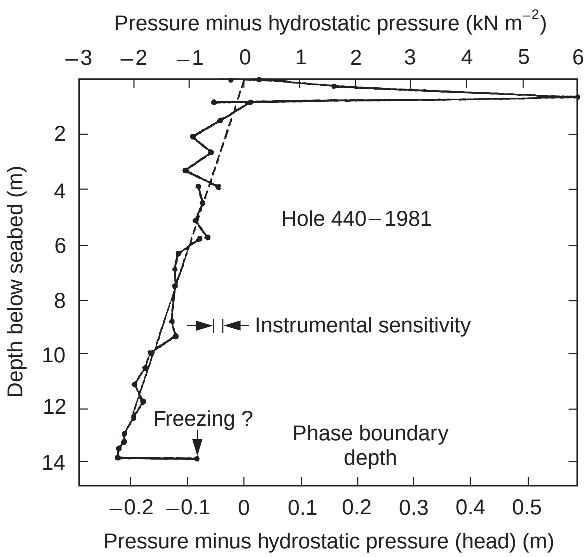

Figure 10 Measured *in situ* pore fluid pressure minus calculated hydrostatic pressure through the thawed layer in coarse-grained sediments at a hole 440 m offshore in May 1981 near Prudhoe Bay, Alaska. High pressure at the 0.54 m depth is probably related to seasonal freezing and the solid line is a least-squares fit to the data below 5.72 m. The negative pressure head gradient (−0.016) indicates a downward component of pore fluid velocity. (Adapted from Swift DW *et al.* (1983) *Proceedings, Fourth International Conference on Permafrost*, Fairbanks, Alaska, vol. 1, pp. 1221–1226. Washington DC: National Academy Press.)

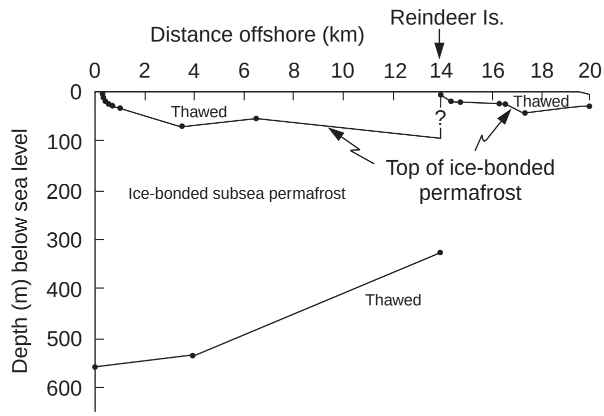

Figure 11 Sub-sea permafrost profile near Prudhoe Bay, Alaska, determined from drilling and well log data. The sediments are coarse-grained with deep thawing inshore of Reindeer Island and fine-grained with shallow thawing farther offshore. Maximum water depths are about 8 m inshore of Reindeer Island and 17 m about 6 km north. (Adapted from Osterkamp TE *et al.* (1985) *Cold Regions Science and Technology* 11: 99–105, 1985, with permission from Elsevier Science.)

to 0.6 m in thickness have been found off the Sagavanirktok River Delta.

Surface geophysical studies (seismic and electrical) have indicated the presence of layered ice-bonded sub-sea permafrost. A profile of sub-sea permafrost near Prudhoe Bay (**Figure 11**) shows substantial differences in depth to ice-bonded permafrost between coarse-grained sediments inshore of Reindeer Island and fine-grained sediments farther offshore. Offshore from Lonely, where surface sediments are fine-grained, ice-bearing permafrost exists within 6–8 m of the seabed out to 8 km offshore (water depth 8 m). Ice-bonded material is deeper ($\sim$15 m). In Elson Lagoon (near Barrow) where the sediments are fine-grained, a thawed layer at the seabed of generally increasing thickness can be traced offshore.

Mackenzie River Delta Region

The layered sediments found in this region are typically fluvial sand and sub-sea mud corresponding to regressive/transgressive cycles (see **Figure 12**). Mean seabed temperatures in the shallow coastal areas are generally positive as a result of warm river water, and negative farther offshore. The thickness of ice-bearing permafrost varies substantially as a result of a complex history of transgressions and regressions, discharge from the Mackenzie River, and possible effects of a late glacial ice cover. Ice-bearing permafrost in the eastern and central Beaufort Shelf exceeds 600 m. It is thin or absent beneath Mackenzie Bay and may be only a few hundred meters thick toward the Alaskan coast. The upper surface of ice-bearing permafrost is typically 5–100 m below the seabed and appears to be under control of seabed temperatures and stratigraphy.

The eastern Arctic, Arctic Archipelago and Hudson Bay regions were largely covered during the last glaciation by ice that would have inhibited permafrost growth. These regions are experiencing isostatic uplift with permafrost aggrading in emerging shorelines.

Antarctica

Negative sediment temperatures and positive temperature gradients to a depth of 56 m below the seabed exist in McMurdo Sound where water depth is 122 m and the mean seabed temperature is −1.9°C. This sub-sea permafrost did not appear to contain any ice.

Models

Modeling the occurrence, distribution and characteristics of sub-sea permafrost is a difficult task. Statistical, geological, analytical, and numerical models are available. Statistical models attempt to combine geological, oceanographic, and other information into algorithms that make predictions about sub-sea permafrost. These statistical models have not been very successful, although new GIS methods could potentially improve them.

Geological models consider how geological processes influence the formation and development of sub-sea permafrost. It is useful to consider these models since some sub-sea permafrost could potentially be a million years old. A geological model for the Mackenzie Delta region (**Figure 12**) has been developed that provides insight into the nature and complex layering of the sediments that comprise the sub-sea permafrost there.

Analytical models for investigating the thermal regime of sub-sea permafrost include both one- and two-dimensional models. All of the available analytical models have simplifying assumptions that limit their usefulness. These include assumptions of one-dimensional heat flow, stable shorelines or

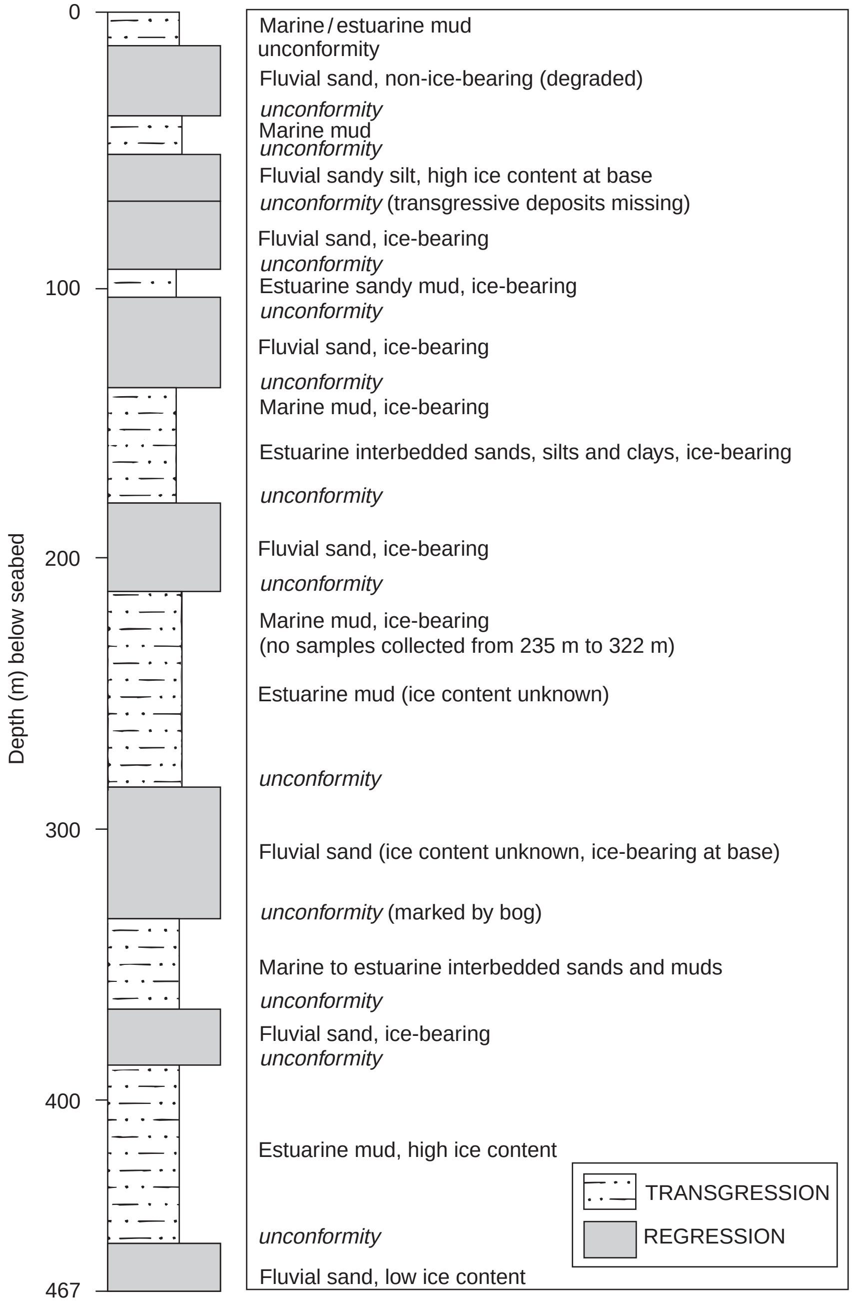

Figure 12 Canadian Beaufort Shelf stratigraphy in 32 m of water near the Mackenzie River Delta. Eight regressive/transgressive fluvial sand/marine mud cycles are shown. It is thought that, except for thawing near the seabed, the ice-bearing sequence has been preserved through time to the present. (Adapted with permission from Blasco S (1995) GSC Open File Report 3058. Ottawa, Canada: Geological Survey of Canada.)

shorelines that undergo sudden and permanent shifts in position, constant air and seabed temperatures that neglect spatial and temporal variations over geological timescales, and constant thermal properties in layered sub-sea permafrost that is likely to contain unfrozen pore fluids. Neglect of topographical differences between the land and seabed, geothermal heat flow, phase change at the top and bottom of the sub-sea permafrost, and salt effects also limits their application. An analytical model exists that addresses the coupling between heat and salt transport but only for the case of diffusive transport with simplifying assumptions. Nevertheless, these analytical models appear to be applicable in certain special situations and have shaped much of the current thinking about sub-sea permafrost.

Two-dimensional numerical thermal models have addressed most of the concerns related to the assumptions in analytical models except for salt transport. Models have been developed that address salt transport via the buoyancy of fresh water generated by thawing ice at the phase boundary. Models for the infiltration of dense sea water brines derived from the growth of sea ice into the sediments do not appear to exist.

Successful application of all models is limited because of the lack of information over geological timescales on initial conditions, boundary conditions, material properties, salt transport and the coupling of heat and salt transport processes. There is also a lack of areas with sufficient information and measurements to fully test model predictions.

Symbols used

h	Volumetric latent heat of the sediments (1 to $2\times10^{8}\,\mathrm{J\,m^{-3}}$)
J_g	Geothermal heat flow entering the bottom phase boundary from below
J_f	Heat flow from the bottom phase boundary into the ice-bonded permafrost above
J_t	Heat flux into the top phase boundary from above
J_f	Heat flux from the top phase boundary into the ice-bonded permafrost below
K	Thermal conductivity of the ice-bonded permafrost (1 to $5\,\mathrm{W\,m^{-1}\,K^{-1}}$)
t	Time
T_s	Long-term mean temperature at the seabed
T_p	Phase boundary temperature at the top of the ice-bonded permafrost (0 to -2°C)
T_e	Phase boundary temperature at the bottom of the ice-bonded permafrost (0 to -2°C)
T_g	Long-term mean ground surface temperature during emergence
$X(t)$	Depth to the bottom of ice-bonded permafrost at time, t
$\dot{X}_{top}$	Thawing rate at top of ice-bonded permafrost during submergence
$\dot{X}_{bot}$	Thawing rate at the bottom of ice-bonded permafrost during submergence

See also

Arctic Basin Circulation. Coastal Circulation Models. Glacial Crustal Rebound, Sea Levels and Shorelines. Heat Transport and Climate. Holocene Climate Variability. Methane Hydrates and Climatic Effects. Mid-ocean Ridge Tectonics, Volcanism and Geomorphology. Millenial Scale Climate Variability. Penetrating Shortwave Radiation. Polynyas. River Inputs. Sea Level Change. Sea Level Variations Over Geologic Time. Sea Ice: Overview; Variations in Extent and Thickness. Sub Ice-shelf Circulation and Processes. Under-ice Boundary Layer. Upper Ocean Heat and Freshwater Budgets.

Further Reading

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