Thermal Models for Post-Glacial Evolution of the Gas Hydrate Stability Zone: Storegga Slide, Norwegian Margin*

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Abstract

Focusing on the Storegga slide area, Norwegian margin, we modeled the evolution of the gas hydrate stability zone (GHSZ) by combining the effects of post-glacial bottom-water warming and sea-level rise (SLR), and including the latent heat of GH formation/dissociation. The delayed onset and downward diffusing of bottom water warmth above 800m water depths explains why the Storegga and other slides did not occur until the early to mid-Holocene, if GH dissociation was a factor. Intersections between the base of the GHSZ and the slide base suggest points of slide initiation--below the upper slope or the shelf, where GH could have formed AFTER the Low Glacial Maximum (LGM), but BEFORE the 8.15ka failure. We quantified the sudden thinning pressure-drop induced by SLR and subsequent ''thermal'' rethickening of the GHSZ within the slide scar. Different SLR curve and pore water compositions were tried to test model sensitivities. At water depths below ca. 800m, persistent cold water allowed the SLR to thicken the GHSZ over time, ruling out post-LGM deepwater GH dissociation-induced initiation of failure. The calculated modern GHSZ thickness increases from zero at 400m water depth to 160m at 1200m, and its base agrees well with published bottom simulating reflector (BSR) depths revealed by seismic data. Mass wasting on upper continental slopes may have been triggered in many mid- to high-latitude seas because the effect of post-Glacial ocean warming on GH stability overcame the gas hydrate-stabilizing effects of SLR. In any event, gas hydrate dissociation by Holocene mass wasting cannot have initiated deglaciation.

Introduction

Researchers have suggested that the occurrence of submarine landslides over continental margins must have been triggered by the dissociation of GH associated with the reduced hydrostatic pressure during glacial maxima. One such landslide is the Storegga slide of the Norwegian margin (Figure 1), a compound (three-event) slide that dislodged ca. 5580 km^3 of sediment. Initially described and dated by Bugge et al. (1988), the three separate Storegga slides were first believed to have occurred at different times, the most recent at 8.5 ka. However, more recent work has shown the three main events to have been essentially concurrent, ca. 8.15 ka (Haflidason et al., 2001). Other researchers further hypothesized that the methane liberated by GH dissociation might have found its way into the atmosphere and caused a strong greenhouse effect, initiating ice sheet melt (e.g., Kennett et al., 2003). Thus the early Holocene date of the Storegga slide presents a problem for the hypothesis that GH dissociation, caused by low sea levels of the LGM, contributed to northern hemisphere deglaciation, which by that time (8.15 ka) was substantially complete. At the time of the slide, global sea levels had already risen to within 25 m of modern values, and were still rising (Fairbanks, 1989).

In this paper, we propose that Holocene sliding is not inconsistent with GH dissociation as a trigger for mass wasting--but as a consequence, rather than a cause, of deglaciation. We test the hypothesis that sliding was delayed into the early Holocene by modeling the time it took for late glacial or early post-glacial warming to penetrate to the base of the GH stability zone. Our models take into consideration time variations of both hydrostatic/lithostatic pressure (P) and temperature (T) since the 18-20 ka LGM. We consider the evolution of pressure and temperature as a result of post-Glacial SLR and bottom water warming, and, in the slide scar, the effect of overburden removal and subsequent thermal re-equilibration. Reconstructed and present profiles across the Storegga slide are based on Bugge et al. (1988). We assume heat transport solely by molecular conduction, and apply solutions from Carlslaw and Jaeger (1959). Due to the low horizontal temperature gradients, we assume heat is transferred only in the vertical direction.

The concentration of GH in sediments is highly variable and not well known, so we calculated the latent heat effect based on typical hydrate-infested sediment: 55% porosity, and 50% of pore space occupied by GH (Booth et al., 1998). In and below the zones of GH formation or dissociation, latent heat will cause temperature changes to be slower than calculated - depending on the amount of GH actually present and the diffusivity of sediments surrounding the dissociation (or forming) gas hydrates. Vogt and Jung (2002) ignored latent heat effects; thus approximating the ‘‘end member’’ situation of little GH. However, only a very thin, thermally insignificant layer of GH dissociation may suffice to reduce shear strength and trigger sediment failure. The GHSZ does not necessarily contain gas hydrate. Although GH is stable in deeper water, any GH released into the water column would rise buoyantly to the ocean surface, losing methane to solution and oxidation on the way. Within the upper sediments, sulfate in downward circulating seawater would oxidize any methane (Borowski et al., 1999). Based on seismic results near the Storegga slide (Posewang and Mienert, 1999), we take this zone to be 115m thick in the area represented by our models.

Model Constraints

We used the eustatic sealevel curve of Fairbanks (1989) to calculate the time-dependent component of lithostatic pressure. The headwall of the Storegga slide is located near the outer edge of grounded LGM ice sheets, which, although grounded on the shelf, were nearly afloat at their distal edges. Thus, post-glacial rebound, if any, was probably small; therefore, the sea-level history in the slide areas was probably nearly eustatic. The sea level rise following the LGM then expanded the GHSZ, particularly during the periods 12.5-11.5 and 10-9 ka, when global sea levels rose most rapidly (Fairbanks, 1989).

The precise time variation of bottom water temperature in the slide areas is not well known. Based on the paleoceanographic reconstructions of Miller et al. (2001), we place the earliest possible time of emplacement of near-modern water temperatures at 15 ka, and we calculated models (not shown) based on this date. However, we consider the end of the Younger Dryas (11 ka) as the most probable time of significant warming and use this date here. We assume a modern-like (historical) water temperature structure was applied to the seafloor instantaneously at 11 ka. However, the subbottom temperature history several millennia later is not very sensitive to the exact times and rates of water warming. We assumed a constant -1^{o}C for the glacial-age ocean prior to warming.

The equilibrium function (phase boundary in P, T space) for hydrate depends on pore water composition (Sloan, 1998). Admixtures of heavier hydrocarbons tend to increase the range of stability, whereas pore water brackishness has the opposite effect. Our models were calculated for fresh water with no higher hydrocarbons, and with 1% and 2% ethane. We also explored the case of pore water with seawater salinity. Posewang and Mienert (1999) found that seawater and 1% ethane predicted a good fit to the lower BSR (Bottom Simulating Reflector) they detected north of the Storegga Slide, in 1000 m water.

Regional heatflow averages ca. 40-60 mW/m^2 in both slide areas (Sundvor et al., 2000). Typical conductivities in the top few meters of sediment are ca. 1.29 W/m-^{o}K for the Storegga area. There are few thermal data from the continental shelves bounding the slide headwalls. Posewang and Mienert (1999) used a gradient of 50^{o} K/km for an 880-m-deep site just north of the Storegga slide; they based this value on borehole temperature as well as seismoacoustic data on BSR depth. Although a global compilation of heatflow shows typical shelf values of ca.80-120 mW/m^2, very rapid LGM sedimentation (up to 1000m/Ma) along shelf edge depocenters around the Nordic Basin margins would have depressed surface heatflow. Given the above data and the various uncertainties, we computed subbottom temperatures for heat flows of 40, 50, and 60 mW/m^2, based on a conductivity of 1.1 W/m-^{o}K. The thermal diffusivity (3.697 x 10^{-7} m^2/sec) was calculated from this conductivity by the relation of Villinger and Davis (1987).

Results and Conclusions

Our models predict how the GHSZ must have changed during the last 18,000 years in the area of the great Storegga slide. We show this change in two ways: Figure 2 shows conditions at various times along the red line (Figure 1) across the upper slide scar. Figure 3 shows how selected parcels of sediment (numbered in Figure 2) moved in P-T space.

During the greatest extent of glaciation (18ka), sea level, and hence subbottom pressure, was too low for hydrate to be stable on the Storegga shelf. By 11ka rising sea level had expanded the stability zone onto the shelf. Some time after upper waters warmed, the stability zone was forced to retreat from the shelf once more. The time between 18ka until somewhat after 11ka years ago would not have been favorable for dissociation, and therefore slides. However, comparison of profile 11ka-and 8.15ka+ shows how hydrate dissociation could have promoted the Storegga slide when it happened--Hydrate present in the dark blue zone at 11ka- would have dissociated, creating gassy, low-shear strength sediment (stippled at 8.15ka+), which then failed when the right earthquake came along.

Failure most likely began where the dissociation zone (black in Figure 2, 8.15ka+) intersects the known slide base.

While we did not try to model the transient slide event, we know it stripped off most of the hydrate stability zone (8.15ka-), which then rethickened as the warm subbottom exposed by the slide was cooled by the overlying water. Some authors have suggested that reduced post-slide pressure would have caused additional hydrate dissociation and secondary slides, but at Storegga all the hydrate except in deep water was probably stripped off. Even if some had been left, the large (negative) latent heat of dissociation would have cooled the remaining hydrate and delayed dissociation.

The solid red tracks in Figure 3 illustrate how shallower sediment parcels (1, 2, 4, 5) first moved upwards towards hydrate stability as sea levels rose, and then veered back out towards instability due to post-11ka warming. Deep parcels (e.g., 7) were unaffected by warming, while other parcels (e.g., 3, 6) never approached the stability field, except in the slide scar, where they were abruptly jerked towards lower pressure and temperature by the slide event (dashed lines) and then warmed again as thermal equilibrium was restored after 8.15 ka.

Our models suggest the Storegga slide removed and/or re-deposited all sediment within the pre-slide GHSZ, except at water depths exceeding ca. 1350 m. At depths greater than this, GH remained below the slide sole, but was abruptly moved out of its stability field by the slide-generated pressure drop. If GH concentrations were similar to those tabulated by Booth et al. (1998), however, complete dissociation would have taken a number of years, due to strong cooling caused by the (negative) latent heat of GH dissociation. Today the residual thermal anomalies caused by slide-induced latent heat effects are negligible ($<1.5\ ^{o}C$).

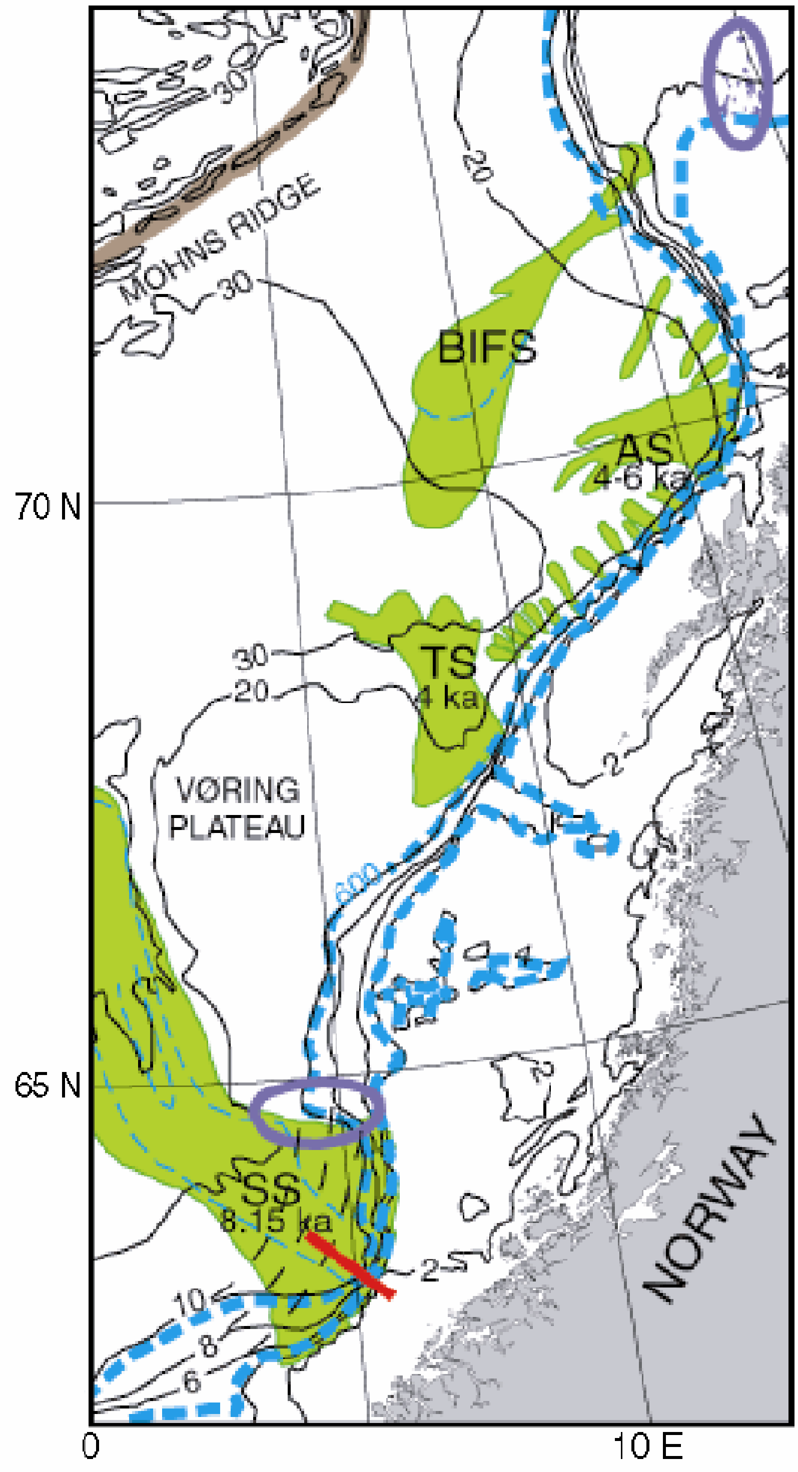

Figure 1. Norwegian continental margin, showing relatively recent giant underwater landslides in green: SS, Storegga slide; TS, Traenadjupet slide; AS, Andoeya slide; and BIFS, Bear Island Fan slide. Gas hydrate stability models shown in Figures 2 and 3 were computed along red line. Black contours show water depths in 100s of meters; thickness of modern methane hydrate stability zone, in meters, shown by dashed blue contours (600m and 0m). Blue ovals and patches show areas where seismic surveying has detected hydrates.

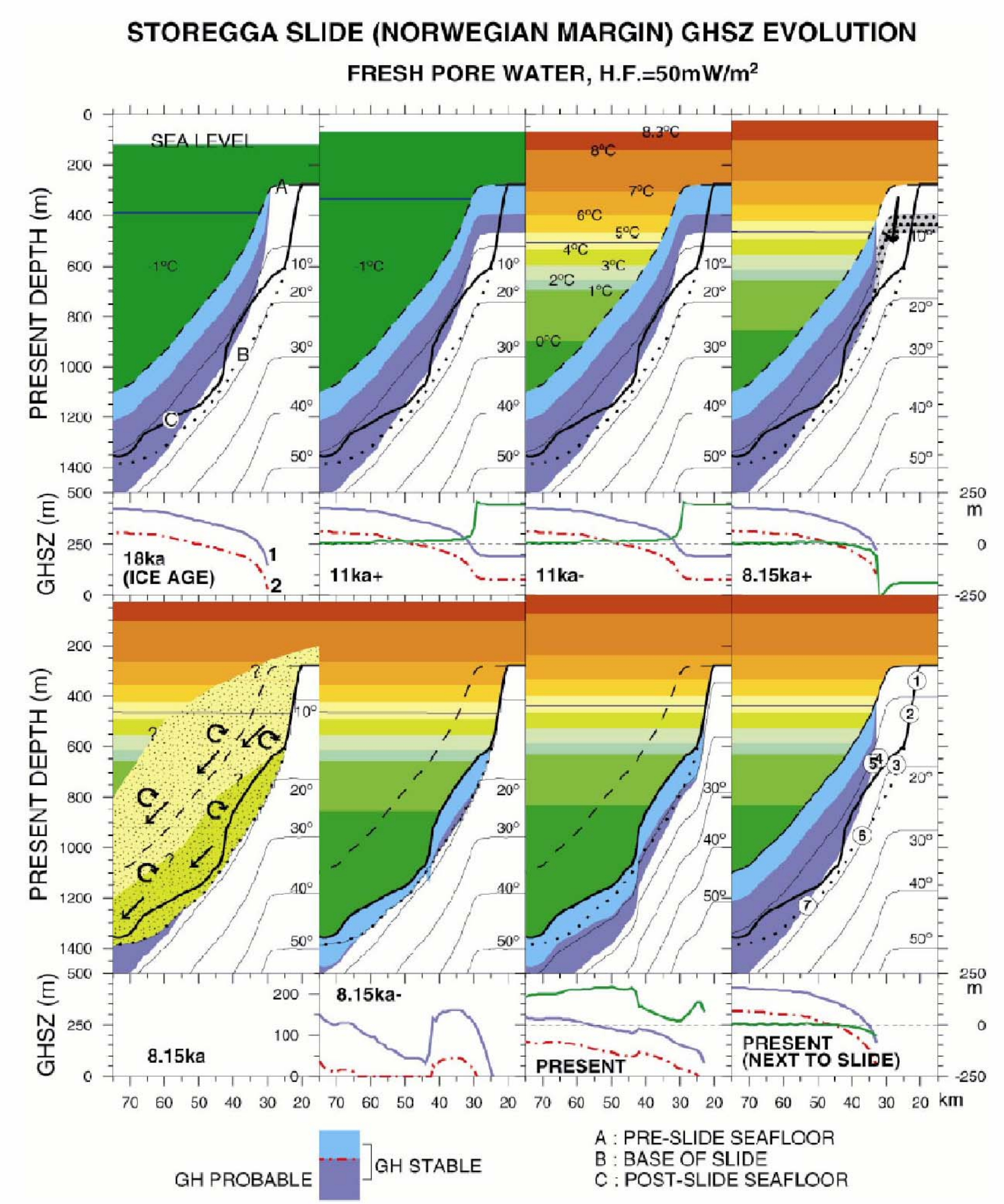

Figure 2. Computed changes in gas hydrate stability zone (GHSZ) over time from last glacial maximum (18,000 years ago, or 18ka) to ocean warming at 11ka, Storegga landslide at 8.15ka, and to present conditions in slide area and on adjacent, not (yet) failed seafloor. 11ka+ means just prior to 11,000 years ago, etc. Stippled zone in 8.15ka+ profile shows area of hydrates dissociated from previous profile (11ka-). Top of each profile is modern sea level. The sub-seafloor GHSZ is shown in blue, with the upper 115m (light blue) probably lacking hydrate. Horizontal blue line is upper limit of GHSZ in ocean. Dashed line (A) shows the restored, pre-slide seafloor, heavy line (B) is the present seafloor, and dotted line (B) is the failure surface (base of slide). Conditions during the actual slide (8.15ka) are schematic only. Small numbers show points whose trajectories in (P, T) space are shown in Figure 3.

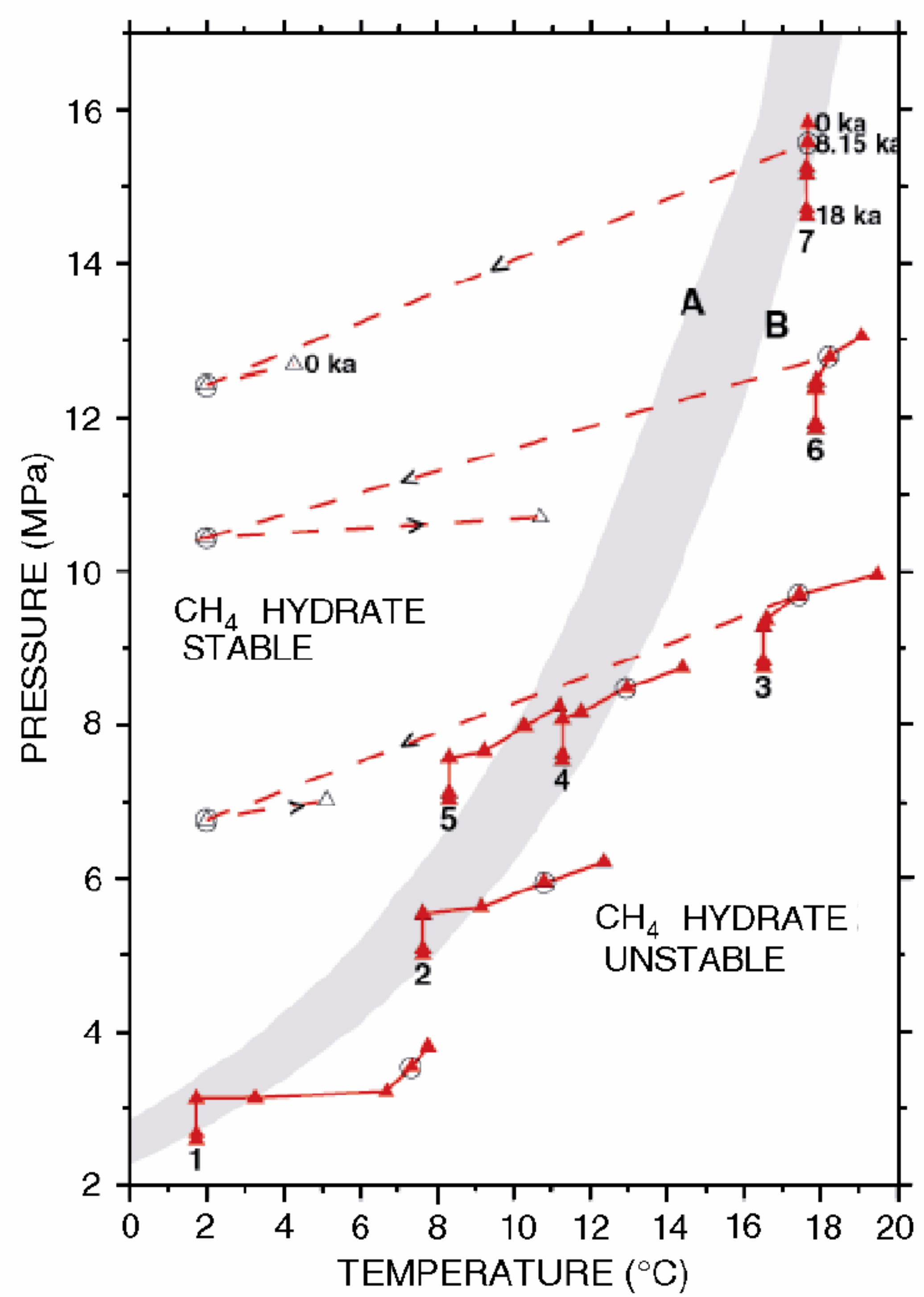

Figure 3. Trajectories (from last Ice Age, at 18ka, to the present) of seven sub-seafloor points (See Figiure 2 for locations) in Temperature and Pressure space, in relation to boundary of methane hydrate stability. Circled points correspond to 8.15ka, time of Storegga slide. Trajectory branches for points within the slide area and covered by slide debris are shown dashed. Because the phase boundary depends on pore water salinity and admixed higher hydrocarbons, it is shown here as a band. The left edge (A) reflects pore water with the salinity of seawater, while the right edge (B) reflects fresh water with 2% ethane.

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