

An Improved Weichselian Seismic Stratigraphy,
Subsurface Fluid Migration Patterns and Methane
Hydrate Dynamics, Offshore West Svalbard

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उपाधि की अपेक्षाओं की आंशिक पूर्ति में प्रस्तुत शोध
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2024

Dedicated to

My mother
Mrs. Archana Trivedi



भारतीय विज्ञान विक्षा ँ अनुसंधान संस्थान, पुणे

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Certificate

This is to certify that the work incorporated in the thesis entitled “**An Improved Weichselian Seismic Stratigraphy, Subsurface Fluid Migration Patterns and Methane Hydrate Dynamics, Offshore West Svalbard**” submitted by **Akash Trivedi** has been carried out by the candidate under my supervisions. The work presented here or any part of it has not been included in any other thesis submitted previously for the award of any degree or diploma from any other university or institution.

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Abbreviations used in this thesis

BGHS: Base of hydrate stability
BP: Before the present
BSR: Bottom Simulating Reflection
BTMF: Belsund Trough Mouth Fan
BWT: Bottom water temperature
BWT _{TF} : Bottom water temperature based on Benthic Transfer Function
BWT _C : Bottom water temperature based on benthic foraminifera $\delta_{18}\text{O}$ values
BMHSZ: Base of the methane hydrate stability zone
CF: Chaotic Facies
CMP: Common Midpoint
ES: Erosional surface
GDF: Glacial debris flow
GIA: Glacial-isostatic adjustment
IBCAO: International Bathymetric Chart of the Arctic Ocean
IRD: Ice-rafted debris
ITMF: Isfjorden Trough Mouth Fan
k_o : Intrinsic permeability
k_r : Relative permeability
KTMF: Kongsfjorden Trough Mouth Fan
LGM: Late Glacial Maxima
MCS: Multichannel seismic
MASOX: Monitoring Arctic Seafloor-Ocean Exchange observatory
MeBo: Meeresboden-Bohrgerät
NMO: Normal moveout
ODP: Ocean drilling program
PKF: Prins Karl Forland
RAMMS: Rapid Mass Movement Simulation
RSL: Relative sea-level
SF: Stratified Facies
SSZ: Sulphate Reduction Zone
STMF: Storfjorden Trough Mouth Fan
SU: Seismic unit
SYSIF: SYstème SIsmique Fond
TMF: Trough Mouth Fan
UCS: Upper continental slope
URU: Upper Regional Unconformity
VR: Vestnesa Ridge
WSC: West Spitsbergen Current

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Abstract

The impacts of ocean warming are increasingly noticeable in the Arctic, such as the retreat of ice sheets and the accelerated melting of glaciers. The marine sediments in the Fram Strait gateway, connecting the Arctic and Atlantic Oceans, host methane hydrates (600 to 2000 m water depth). Methane hydrate is an ice-like substance that forms when methane in gaseous form is trapped within the crystalline structure of water molecules. It remains stable under low temperature ($-10\text{ }^{\circ}\text{C}$ – $25\text{ }^{\circ}\text{C}$) and high-pressure (3–38 MPa) conditions. However, the stability of the hydrates is sensitive to ocean warming and relative sea-level (RSL) changes. In 2008, methane bubbles were observed above the present-day upper limit of hydrate stability (~ 400 m water depth) in the western continental margin of Svalbard, suggesting hydrate dissociation, but the exact cause remained undetermined.

This study examines the influence of RSL changes and variations in bottom water temperature (BWT) on methane hydrate dynamics using numerical simulation code TOUGH+Hydrate (v 1.5, 2014). To assess the impact of RSL fluctuations on hydrates over the past 11,000 years, two different reconstructions of local ice history were used to predict RSL changes. The University of Tromsø (UiT) model suggested a fall in RSL, while the ICE-6G_C model predicted a continuous rise. The modelling revealed that hydrates had dissociated before the bottom water cooling phase that began in the late 1970s, during which hydrates could reform. During the period of accelerated bottom water warming from 1978 to 2016, the newly formed hydrates underwent complete dissociation. Therefore, recent ocean warming played a critical role in newly formed hydrate dissociation completely regardless of long-term sea-level history.

The seismic stratigraphic framework for the upper continental slope (top 80 m below seabed) between the Kongsfjorden and Isfjorden Trough Mouth Fans is refined. The study identified four distinct episodes of shelf edge glaciations occurring around 120–110, 90–85, 70–54 and 38–24 ka. On the uppermost continental slope, the Weichselian glacial units are characterized by glacial debris materials transported by a slow-moving ice sheet. Prominent V-shaped indentations carved out by icebergs indicate distinct calving episodes and erosional trough as a result of glaciogenic debris flows. Below the methane seepage area on the upper continental slope, the debris flows exhibit an oblique retrogradational stacking pattern. At water depths > 500 m, the debris flow units transitioned into stratified turbidite deposits, indicating a shift from debris flows to turbidity currents due to progressive dilution with water.

Based on the seismic interpretation and numerical modelling, a comprehensive analysis of fluid accumulation beds and migration patterns in the upper continental slope was conducted. Present-day extent of the base of gas hydrate stability was calculated, which is in a transient state due to the migration of the base of methane hydrate stability over the past 11,000 years due to changes in RSL and BWT. Although the SYSIF seismic data did not reveal a bottom-simulating reflection (BSR) on the uppermost continental slope, numerous gas pockets (~ 10 to 30 m below seabed) represented as bright spots with negative polarity below the present-day base of methane hydrate stability were observed. Beneath the methane seepage area, two distinct sets of reservoirs were identified (first ~8 to 12 m and second ~ 20 to 33 m below seabed), comprising fragmented turbidite beds with gas occurrences. These beds are separated vertically by low-permeability glaciogenic debris (~ 12 to 33 m below seabed), that prevents fluid flow. Based on seismic evidence and modelling, it is shown for the first time that methane hydrate dissociation, induced by ocean warming, results in the release of frozen methane. This gas then migrates through fractures, disrupting the sealing capacity of the debris stacks, leading to fluid focusing toward the seepage locations.

Summarsing, the findings underscore the impact of recent ocean warming on the dissociation of methane hydrates offshore west Svalbard. High-resolution seismic data analysis improved our understanding of glacial stratigraphy, fluid migration patterns and methane hydrate dynamics in the rapidly changing Arctic region. Advanced imaging methods such as 3D seismic and SYSIF enhance our understanding of glacial reservoirs. This knowledge can optimize hydrocarbon exploration by providing insights into complex reservoir architecture and fluid movement within the fractured system, thereby improving exploration methods.

Chapter 1

Introduction

1.1 Gas hydrates

Gas hydrates are solid crystalline compounds with an ice-like structure, comprising guest gas molecule, predominantly methane hosted inside the cage formed by ice (Sloan and Koh, 2007). They are found in marine continental margins and cold permafrost settings as high pressure and cold conditions are necessary for their stability (Sloan, 1998; Collett et al., 2011). Under natural conditions, methane hydrate is stable within a temperature range of $\sim -10\text{ }^{\circ}\text{C}$ – $25\text{ }^{\circ}\text{C}$ and pressures ranging from approximately 3–38 MPa (Ruppel and Waite, 2020). Their presence is detected through scientific drillings (Paull et al., 1996) or the interpretation of seismic profiles, where a crosscutting negative polarity reflection mimicking the seabed, is identified as the Bottom Simulating Reflection (BSR) (Shipley et al., 1979). It marks the base of gas hydrate stability (BGHS) and the transition from hydrate to underlying free gas.

Methane hydrates are sensitive to environmental changes; for instance, increasing temperatures can destabilize them, leading to the emission of methane (Figure 1.1a); conversely, increasing sea level deepens hydrate stability (Figure 1.1b). The methane release through hydrate destabilization is a serious concern (Ruppel, 2011). The exact volume and fraction of methane hydrates that are vulnerable to warming remain uncertain. Buffett and Archer (2004) considered an upper bound estimate of around 3000 gigatons of carbon stored in the hydrate inventory. Rising atmospheric CO_2 levels can induce an increase in intermediate and deep ocean temperatures, dissociating methane hydrates, releasing methane and reinforcing a feedback loop linking atmospheric warming, deep ocean temperature and methane release from hydrate dissociation (Archer and Buffett, 2005; Levitus et al., 2005). This can trigger a self-perpetuating cycle of warming and methane release, with significant implications for accelerating global warming if substantial amounts of hydrate-bound methane are released (Nisbet, 2002).

Methane release from hydrate bearing marine sediments is a complex process influenced by various factors, with microbial community playing a crucial role. Anaerobic methane oxidation and bacterial sulfate reduction, act as filters, preventing methane from reaching the ocean (Barnes and Goldberg, 1976). However, anaerobic oxidation has a limited impact on reducing seabed methane emissions from methane hydrate dissociation in low-permeability marine sediments (Stranne et al., 2022). If methane bypasses this filter and is released into the ocean, it may dissolve or be oxidized to form carbon dioxide.

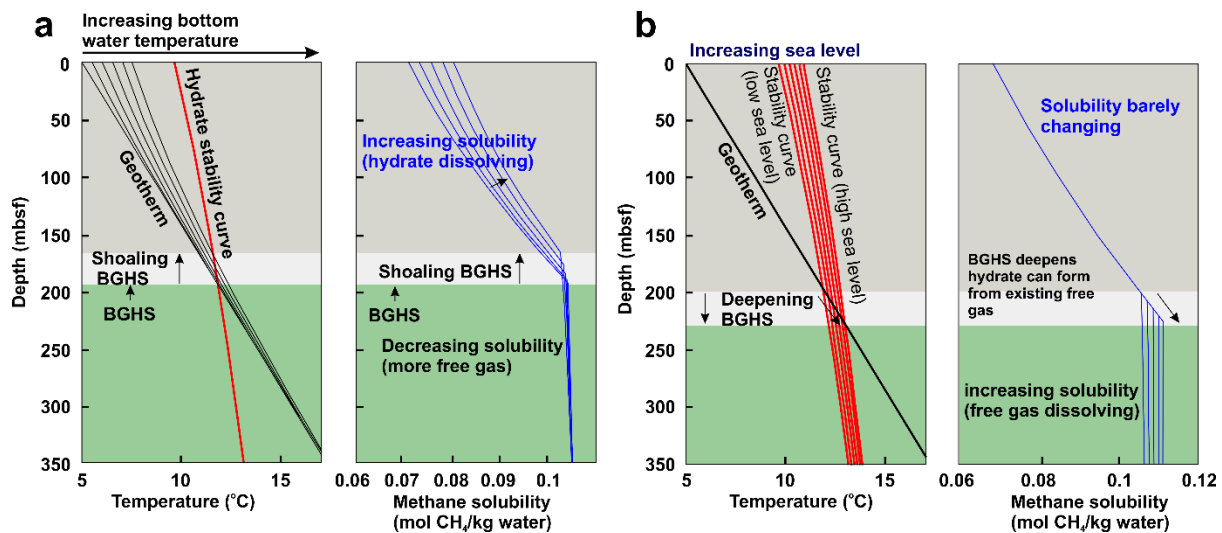


Figure 1.1. Stability of methane hydrates to increases in ocean temperature and sea level, reproduced from Ruppel and Kessler (2017). **(a)** A scenario where seabed temperature rises, the BGHS moves upwards. Below the BGHS, methane solubility decreases, causing dissolved gas to transition into free gas. Increasing solubility within the hydrate stability, results in hydrate dissolution. **(b)** Sea level rise deepens the hydrate stability zone. The sea level is increased by 100 m over thousand years. The hydrate stability curve shifts to a new position in response to the increase in hydrostatic pressure and the BGHS deepens, while the geothermal gradient remains constant. Below the BGHS, methane solubility increases resulting in methane dissolution. Hydrate will form close to the new BGHS if there is excess of methane solubility.

Though infrequent, rapid methane releases from marine hydrates, known as "catastrophic" events, can lead to significant global methane emissions. Submarine landslides, for instance, can trigger such catastrophic methane releases (Maslin et al., 1998). A more likely outcome of climate warming is a gradual, continuous emission of methane because of oceanic hydrate destabilization, termed as a "chronic" event (Archer and Brovkin, 2008). Although less dramatic, chronic methane releases can significantly contribute to long-term climate change when oxidized to carbon dioxide. If significant methane is released from the hydrate reservoir over thousands of years, it impacts the climate where the cumulative CO₂ concentration surpasses the impact of a consistent rise in methane (Archer and Buffett, 2005; Schmidt and Shindell, 2003; Shaw and Mukherjee, 2022).

Methane hydrate dissociation has been associated with past warming episodes (Ruppel and Kessler, 2017). For example, significant carbon injection events from the geosphere to the atmosphere have been linked to episodes of abrupt climate change, exemplified by occurrences like the anomalous thermal maximum at the boundary between Paleocene and

Eocene (Penman and Zachos, 2018). The close correlation observed between the abrupt Late Quaternary glacial terminations, as evidenced by rapid rises in temperature, and the simultaneous increase in methane with a lighter isotope composition, suggests the possible impact of methane hydrate dissociation (Kennett et al., 2003). During interstadials, known as Dansgaard-Oeschger events, the intermediate depth water temperature rises triggered methane hydrate destabilization, releasing CH₄ into the ocean and atmosphere. However, investigations using $\delta\text{D-CH}_4$ question this hypothesis. These studies challenge the conventional belief that fluctuations in atmospheric CH₄ concentrations during specific Dansgaard-Oeschger events are solely attributed to marine gas hydrate dissociation. Instead, the authors proposed that these variations are more likely influenced by wetland methane (Bock et al., 2010).

Global warming results in higher average surface ocean and air temperatures, which can raise sea levels due to enhanced glacier melting (Hugonnet et al., 2021). Marine hydrates at the feather edge of stability are influenced by higher oceanic temperatures, which generally have a more significant impact than the minor stabilizing effect of rise in sea level (Braga et al., 2021). The pressure and temperature effects work on different time scales. A relative sea level fall depressurizes hydrates, allowing a very short time for the pressure effect to transmit through the sediments (Oluwunmin et al., 2022). Conversely, the impact of temperature changes is delayed, takes time to propagate through the sediment, depending on the thickness and thermal properties, such as specific heat capacity and thermal conductivity of sediments. In the polar regions, methane hydrates form substantial methane reservoir in the oceanic sediments and permafrost (Macdonald, 1990). The polar region is unique in that its cold-water mass facilitates the conditions conducive for the stabilization of methane hydrates as shallow as 200 m water depths, significantly shallower than in other regions of the world. The polar oceanic hydrates remain vulnerable to ocean warming, as these regions are highly sensitive to future climate change, including rising temperatures (Houghton et al., 2001). Moreover, due to shallower depths, there is enhanced likelihood of methane reaching the atmosphere (Archer and Buffett, 2005).

The dissociation of methane hydrates is prevalent on Arctic slopes and shelves (Kvenvolden et al., 1993; Stranne et al., 2016). Methane emission in the continental shelf of East Siberia is attributed to thawing of permafrost and methane hydrate dissociation, because of marine transgression on coastal plains since the last deglaciation (~ 15 ka BP) (Shakhova et al., 2010). The western Svalbard margin serves as a globally relevant study site to gain more insights into the role of oceanic warming on methane hydrate dynamics. Westbrook et al.

(2009) attributed the methane emission from the western Svalbard slope at ~ 370-400 m depth to methane hydrate destabilization caused by recent (1978–2008) bottom water warming. Currently, there is no indication of methane entering the atmosphere (Myhre et al., 2016). Repeated geophysical surveys and the establishment have contributed significantly to the knowledge of arctic hydrate dynamics (Westbrook et al., 2009; Berndt et al., 2014, Wallmann et al., 2018).

1.2 Western Svalbard margin

Svalbard is an archipelago situated in the northwestern Barents Sea between 74°N and 82°N. The mid-oceanic ridge and associated transform faults, including the Knipovich and Molloy Ridges, as well as the Molloy and Spitsbergen transform faults are located west of Svalbard (Figure 1.2). The passive continental margin of western Svalbard evolved through a series of specific tectonic events. These are as follows.

Late Cretaceous-Early Eocene: The separation between Greenland and Svalbard was initiated by the dextral strike-slip displacement along the Spitsbergen Shear Zone (SSZ) (Figure 2, Crane et al., 1991). It opened a series of pull-apart basins in an *en-echelon* manner.

Early-Middle Eocene: Following the initial strike-slip, the SSZ entered a regime of transpression known as the West Spitsbergen Orogeny (Figure 3.11, Harland, 1997) that gave rise to the West Spitsbergen Fold and Thrust belt. The transpression was linked with the Hornsund Fault system (Faleide et al., 2008). Transpression persisted from Late Paleocene to the Middle Eocene, after which transtension occurred across the SSZ.

Middle Eocene-present: Oblique extension within the SSZ resulted in the Knipovich Ridge spreading towards north, opening the Fram Strait (Figures 2 and 6, Crane et al., 2001). In the Early Miocene, the Molloy Ridge spreading was initiated (Boebel, 2000). The spreading rates diminish in a northward direction, varying from 0.43–0.49 cm/yr at 75°N to typically < 0.3 cm/yr at 78°N (Ritzmann and Jokat, 2003). The ultraslow asymmetric spreading of the Knipovich Ridge caused spreading on western side being 1.5 times greater than on the eastern segment (Eldholm et al., 1990). The ridge obliquity varies between 35° in the N and 49° in the S (Crane et al., 2001).

Throughout the Pliocene–Pleistocene, the advancement of glaciers on the shelf produced a regional unconformity. Ice streams curved out the cross-shelf troughs building up the Trough Mouth Fans (TMFs) on the slope (Figure 1.2, Vorren et al., 1991; Laberg and Vorren, 1996).

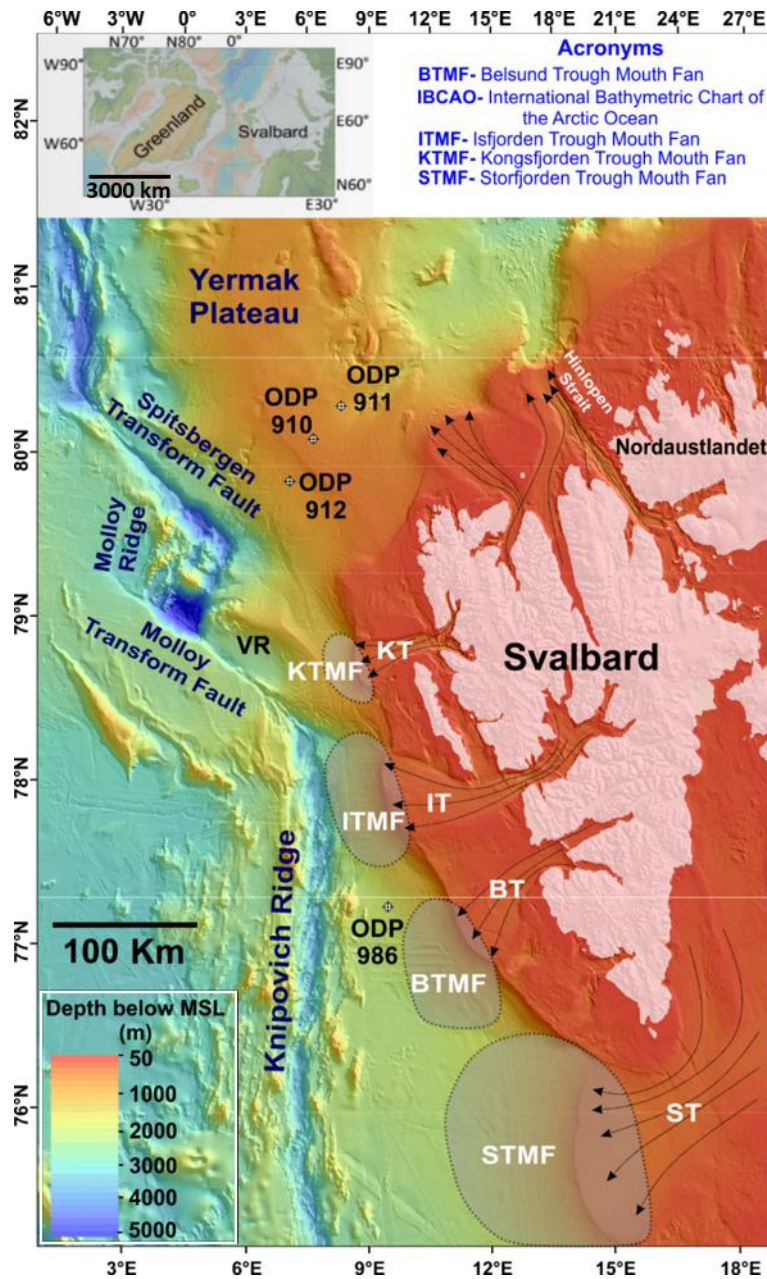


Figure 1.2. IBCAO bathymetry (Jakobsson et al., 2020) reveals glaciated continental shelf of western Svalbard. Several E–W trending troughs dissect the shelf and were carved out by the paleo-ice streams. These troughs are named Storfjorden (ST), Bellsund (BT), Isfjorden (IT), and Kongsfjorden (KT) troughs. Trough mouth fans (TMFs) are the centres of glacial wedge build up by the ice streams at the mouths of the respective troughs. The bathymetry also reveals the ridge and transform fault system, and Ocean Drilling Program (ODP) sites located between BTMF and ITMF (Leg 162, Forsberg et al., 1999) and on the Yermak Plateau (Leg 151, Myhre et al., 1995).

1.3 Methane hydrates off west Svalbard

Gas hydrate BSRs are detected in the Vestnesa Ridge (VR), the Molloy and Spitsbergen Transform Fault (Figure 1.3, Vanneste et al., 2005). The VR, a contourite drift, exhibits a significant sediment thickness variation. It reaches ~ 5 km at the crest, thinning towards the adjoining spreading centre and transform fault system (Eiken and Hinz, 1993; Hustoft et al., 2009). Three-dimensional (3D) seismic data collected on the VR reveal normal faults and fractures facilitating fluid ascent (Figure 1.4, Plaza-Faverola et al., 2015; Plaza-Faverola and Keiding, 2019). Fracture opening due to high tensile stress in glacial forebulges, such as the VR, assisted in methane venting during the periods of substantial ice build-up such as the Mid-Pleistocene Transition and the Last Glacial Maximum (LGM) (Vachon et al., 2022; Cooke et al., 2023). Seafloor pockmarks are connected to gas migration through faults (Plaza-Faverola et al., 2017) on the VR. The depth of the BSR below the seabed was used to determine geothermal gradients that show a gradual rise from $70\text{--}115^\circ\text{C km}^{-1}$ toward the Knipovich Ridge (Vanneste et al. 2005; Sarkar et al., 2012). The organic-rich source rocks in the VR thermally matured due to their proximity to the young oceanic crust. This produced thermogenic methane, which is trapped in hydrate accumulations (Smith et al., 2014; Dumke et al., 2016). Hydrate occurs as slender veins and chunks, measuring tens of cm and embedded within a muddy matrix on pockmarks of the VR (Panieri et al., 2017).

Gas hydrates were recovered from a sediment core in the interfan between the Isfjorden and Kongsfjorden TMFs (Fisher et al., 2011). BSR is identified in the interfan slope (Figure 1.3, Sarkar et al., 2011). However, the BSR could not be identified in the glaciomarine strata at shallower water depths (< 600 m). Below this depth, hydrate is theoretically stable, with the hydrate stability base outcropping at 410 m water depth on the seabed. Analysis of V_S and V_P (S and P wave velocities, respectively) variations from wide-azimuth seismic data off western Svalbard provided further insights into hydrate distribution and saturation (Chabert et al., 2011). In hydrate-bearing sediments, both V_P and V_S increase, while free gas is associated with a reduction in V_P . There is a strong lithological control in influencing methane hydrate distribution. Chabert et al. (2011) provided estimates of hydrate saturation based on observed V_P and V_S variations. The effective medium-based modelling method revealed a patchy distribution of hydrates in the interfan region, influenced by glaciomarine materials near the uppermost slope, with a concentration up to $\sim 5\%$ hydrates and $\sim 3.5\%$ free gas just below the BGHS. In the lower slope, where hydrate is inferred to occur as the load-bearing framework within marine contouritic sediments can have concentrations twice as high as

those found in the glaciomarine sequence. In lower slope, Chabert et al. (2011) estimated 6–12 % hydrate concentration, which matches well with similar estimates for deep waters in northwest Svalbard by Westbrook et al. (2008). Goswami et al. (2016) estimated 20–50 % hydrate, where a BSR is present, while < 12 % in the predicted hydrate stability zone at the upper slope, where the BSR is not found based on resistivity derived from controlled source electromagnetic data.

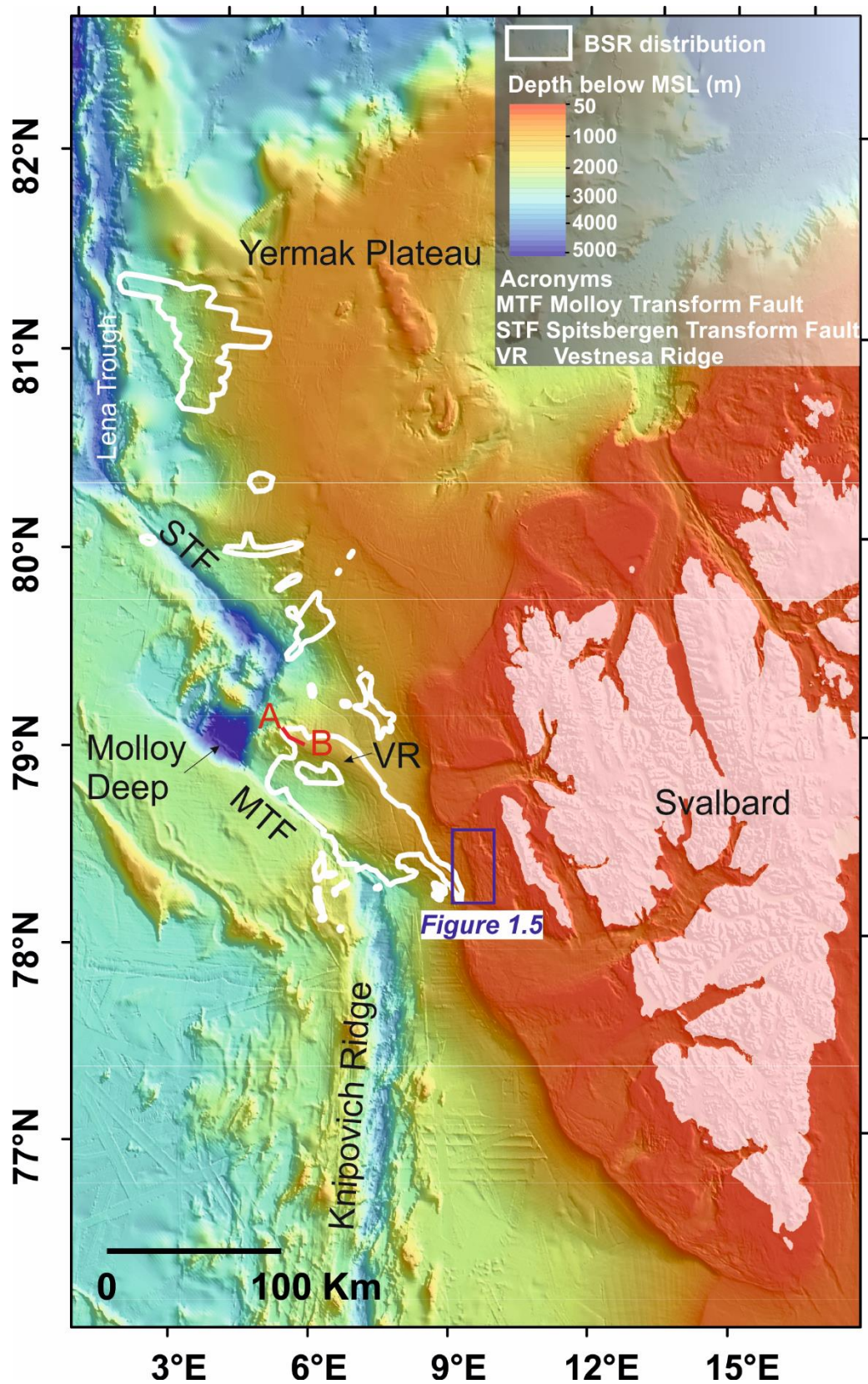


Figure 1.3. Gas hydrate BSR distribution. The BSR map was constructed using multiple 2D seismic lines collected during various marine seismic surveys and compiled by Minshull et al. (2020). Seismic line along A–B from the Vestnesa Ridge is shown in Figure 1.4.

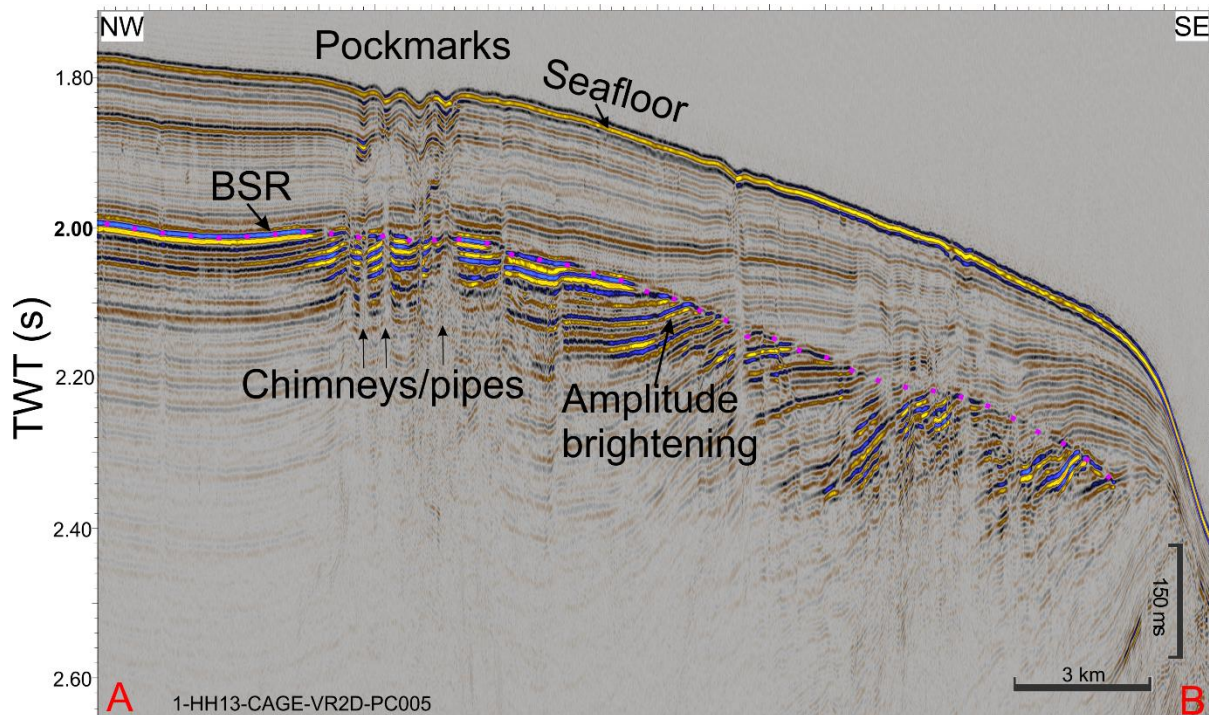


Figure 1.4. Seismic line from the Vestnesa Ridge reveals a crosscutting BSR exhibiting a reversed polarity. Zones of near-vertical/vertical amplitude fading, inferred as seismic chimneys, terminate beneath the present-day pockmarks. Bright amplitudes are due to free gas beneath hydrate.

1.4 Ocean warming and gas hydrate destabilization

Ocean warming can destabilize methane hydrate at the upper continental slope. A geophysical expedition conducted offshore western Svalbard in August–September 2008 revealed active methane bubbles emerging from the seafloor at 400 m and shallower water depths (Figure 1.5, Westbrook et al., 2009). Westbrook et al. (2009) attributed the origin of these seeps to methane hydrate destabilization triggered by the warming of the West Spitsbergen current (Figure 1.6), which has experienced a 1 °C temperature increase from 1975 to 2008. This temperature rise resulted in the deepening of the uppermost theoretical methane hydrate stability limit, shifting from ~ 370 to ~ 400 m (Figure 1.7, Westbrook et al., 2009). As buoyancy-driven methane migrates up-dip to shallower depths, it converts into methane hydrate, within the hydrate stability field particularly up to 400 m water depth. Consequently, a portion of the seeps may be attributed to methane originating from hydrate destabilization. Gas hydrate tends to reduce permeability of sediments and obstruct upward gas migration. It enhances the likelihood of lateral methane migration along the methane hydrate stability base to areas where this zone emerges on the seabed (Figure 1.7).

Thatcher et al. (2013) used 1D computer simulations to examine whether methane emission results from warming-induced hydrate destabilization. They modelled the evolution of methane hydrate in a sedimentary layer at 380 to 410 m water depths by linearly increasing the temperature from 2° C to 3° C over 1975–2008. They considered the porosity, permeability, and thermal properties using expected values specific to glaciogenic setting. Hydrate dissociation began at the hydrate and free gas boundary within five years of warming, and spread throughout the entire hydrate column within 20 years. It took 38 years for the gas to reach the seabed following the complete hydrate dissociation. The duration of seafloor gas emergence after the onset of warming is mainly controlled by thermal properties, such as thermal diffusivity. Other factors that affect gas flow through the sediment column include critical gas saturation, permeability and hydrate saturation. Marin-Moreno et al. (2013) studied the gas hydrate dynamics in response to temperature fluctuations over the last 2000 years. Their timeline considered historical periods, including Little Ice Age (~ 1500-1900 CE), Medieval Climate Anomaly (~ 900-1500 CE), Dark Ages Cold Period (~ 600-900 CE), Roman Warm Period (~ 1-600 CE), and the industrial period (Spielhagen et al., 2011). They showed that hydrate dissociation west of Svalbard is primarily due to recent ocean warming.

1.4.1 Monitoring Arctic Seafloor-Ocean Exchange (MASOX) observatory

The seafloor MASOX observatory (Unmanned) (Figure 1.5, inset) was established at a depth of 394 m to monitor variations in BWT and methane bubble activity. The results were used to build a continuous time series record (Berndt et al., 2014) of BWT and methane flare activity using current meters, temperature probes, and an echo-sounder. The time series record showed strong seasonal variations in BWT, with a temperature difference of ~ 1.5 °C between spring and fall. Seasonal variations can result in repeated hydrate formation and melting episodes.

Carbonate crust and nodules at the MASOX site indicate authigenic carbonate precipitation. Anaerobic methane oxidation caused by sulphate reducing bacteria causes precipitation of bicarbonates and carbonates (Luff et al., 2004). According to Berndt et al. (2014), despite ocean warming over the past century in the western Svalbard, a 100-year timeframe cannot fully explain the observed carbonate thickness. U/Th dating on recovered carbonate samples suggest methane precipitation dating back as far as ~ 8000 years. Ferré et al. (2020) explored the impact of seasonality on gas flare activity. Their findings suggest that the region's

methane flare activity diminishes during cold seasons when BWTs are significantly lower than the summer temperatures.

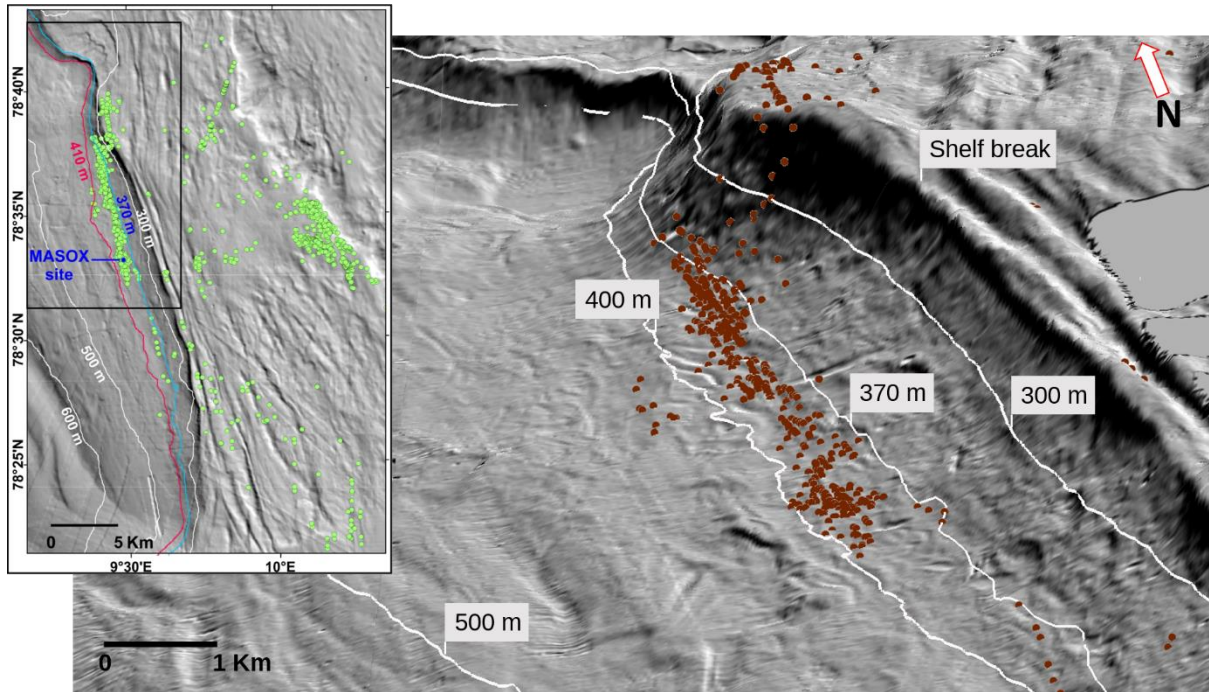


Figure 1.5. Three-dimensional representation of shelf break and upper slope bathymetry off western Svalbard shows seabed gas seepage locations (indicated by red dots). The location of the inset in the upper left corner is shown by a blue rectangle in Figure 1.3 (Inset shows seepage locations as green dots). The seeps are mostly located between the 370 and 400-m contours and are a result of hydrate dissociation due to bottom water temperature (BWT) rise from 2° C to 3° C. The uppermost hydrate stability limit outcrops at 370 m and 400 m water depths at 2° and 3° C BWT respectively. The northern seepage trend, displays a distinctive clustering along the shelf break. These seeps are attributed to lateral gas migration along prograding glacial clinoforms and are not linked to hydrate dissociation (Sarkar et al., 2012).

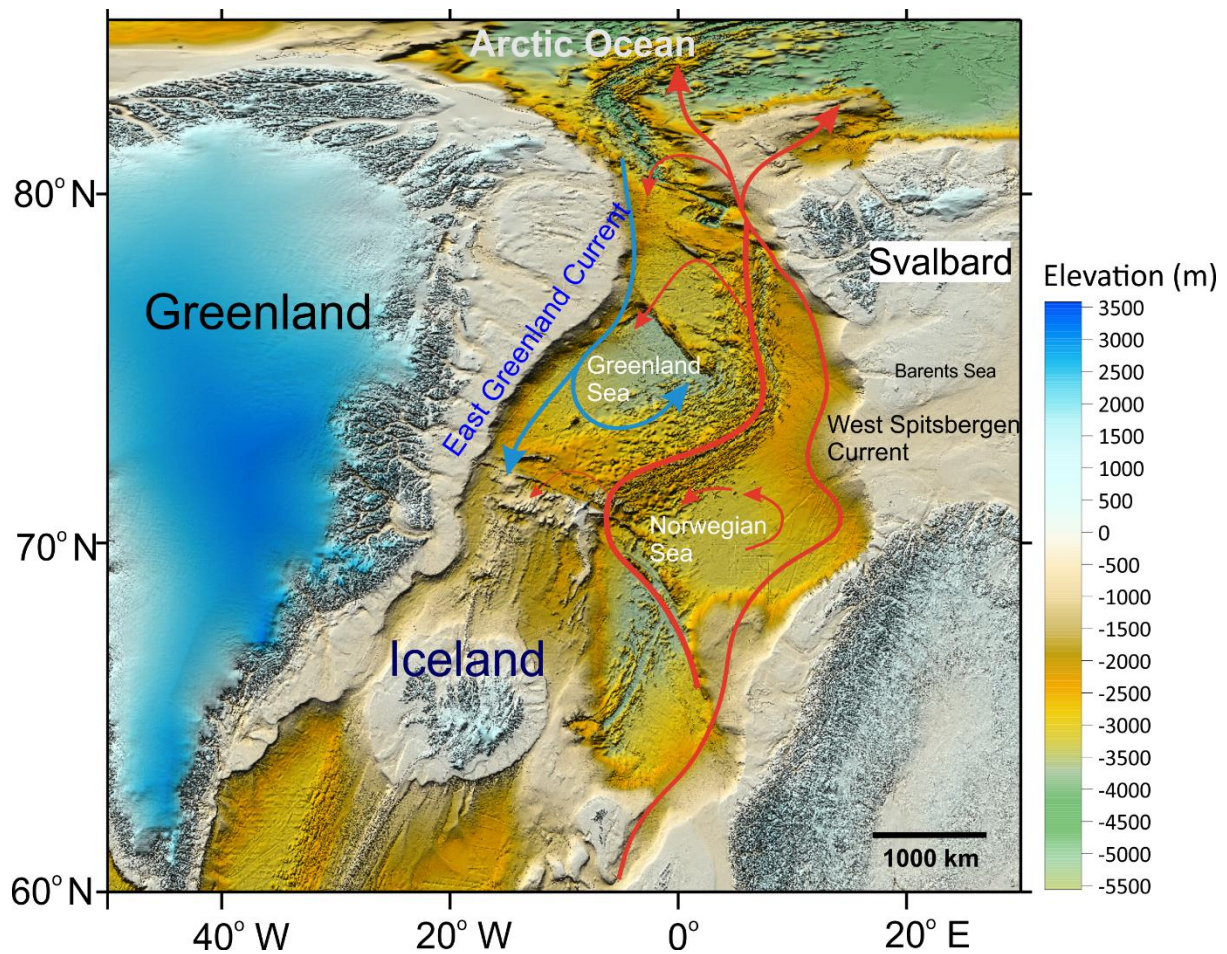


Figure 1.6. Schematic illustration of the Northern Atlantic Ocean circulation system. The northward flowing Norwegian Atlantic Current and its western branch closely follow the mid-oceanic ridges, such as the Mohns and Knipovich Ridges. The West Spitsbergen Current, offshore west Svalbard, transports the warm Atlantic waters. The southward flowing East Greenland Current transports cold Arctic waters along eastern Greenland.

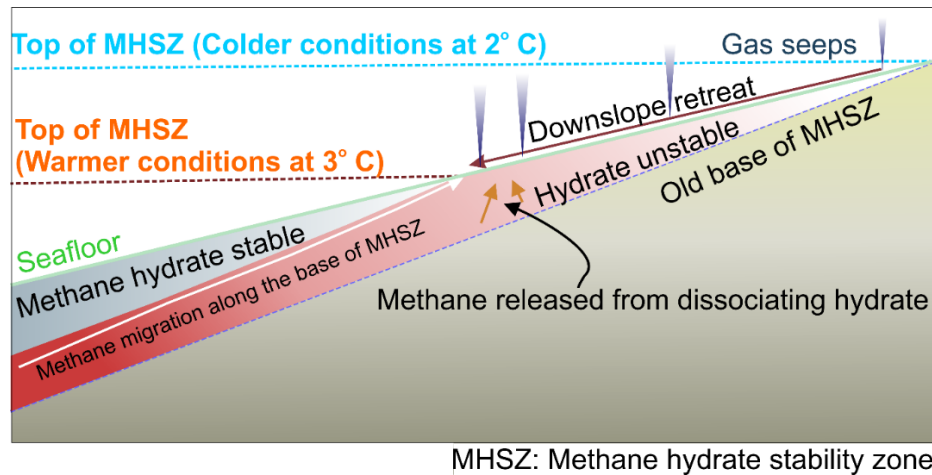


Figure 1.7. Cartoon depicting the retreat of methane hydrate stability zone in response to warming bottom waters. As depicted in the cartoon, a 1 °C increase in BWT causes the uppermost limit of methane hydrate stability to shift downward. At 3 °C, the zone recedes deeper than its position at 2 °C. The dissociation of hydrates due to warming releases trapped methane, results in methane venting within the retreat zone. Additionally, some methane may migrate upwards along the base of remaining hydrate and released at the uppermost limit of its stability, further contributing to methane release.

1.4.2 Hydrate modelling results based on observations from MeBo (Meeresboden-Bohrgerät) drilling experiment

A shallow coring experiment was conducted using the MeBo70 drill-rig (Wallmann et al., 2018) to determine if methane hydrates are present in the marine sediments below the active methane seepage sites on the slope west of Svalbard. The pore fluid chemistry of the samples retrieved by drilling revealed a decline in chloride concentration at depths between 15 and 25 m below the seabed. The reduction in salinity was attributed to freshening of pore waters caused by gas hydrate destabilization. The pore-fluid isotopic composition ($\delta^{18}\text{O}$, $\delta^2\text{H}$) with respect to chloride concentration distinctly indicates the role of hydrate dissociation.

Wallmann et al. (2018) linked hydrate destabilization to a fall in RSL. Around 8 ka before the present, the postglacial isostatic rebound rate surpassed rise in the eustatic sea level, causing a decline in RSL and hydrostatic pressure. The hydrates dissociated completely prior to the recent increase in BWTs (Westbrook et al., 2009; Sarkar et al., 2012; Marin-Moreno et al., 2013; Thatcher et al., 2013).

Wallmann et al. (2018) simulated hydrate behaviour considering the historical records of BWT and sea level. Using BWT reconstructed from benthic foraminiferal $\delta^{18}\text{O}$, the model was initialized at 2.15°C at 13 ka, with the initial hydrate layer occurring at 16–20 m below

seabed. During early Holocene, the seafloor temperature reached a peak of 4.8°C. The simulation indicated that hydrate dissociation was complete around 10.7 ka Before the Present (BP) due to seabed warming. However, the resulting chloride anomaly (salinity) diffused away in approximately thousand years.

In an alternative scenario, Wallmann et al. (2018) examined hydrate behaviour over the past 8,000 years, considering fluctuations in both relative sea level and bottom water temperature. They initiated this model at 8,000 years ago when the BWT was 2.5 °C and a RSL was 12 m higher than present. The BWT was kept constant at 2.5°C until the last 100 years, after which it linearly rose to 3°C (Figure 1.8). The MeBo drilling locations, ~ 50 km west of the Svalbard coast, experienced rebound isostatically after the LGM. Using a glacial isostatic adjustment model, Wallmann et al. (2018) refined the reconstruction of the isostatic rebound at the drilling site (391 m depth). Around 8,000 years ago, a decline in relative sea level (RSL) occurred as isostatic seabed uplift surpassed eustatic sea level rise. Their simulation results revealed complete hydrate dissociation because of the decline in RSL preceding the warming event in the last 100 years. The study highlighted that an initial 60 % hydrate saturation is necessary to generate the observed chloride anomaly in pore fluid during MeBo drilling.

1.4.3 Review on the depressurization-induced hydrate destabilization model

i. Influence of BWT Fluctuations:

In their model, Wallmann et al. (2018) assumed a flat BWT curve for the period of 8,000 to 100 years BP. However, temperature variations were evident during this timeframe (Figure 1.8), including period of seabed cooling, which can influence gas hydrate reformation. Their model did not consider these variations, and shallow hydrate reformation during seabed cooling. Consideration of this process can cause a delay in the complete hydrate dissociation. Consequently, this omission could lead to biased predictions related to methane seepage.

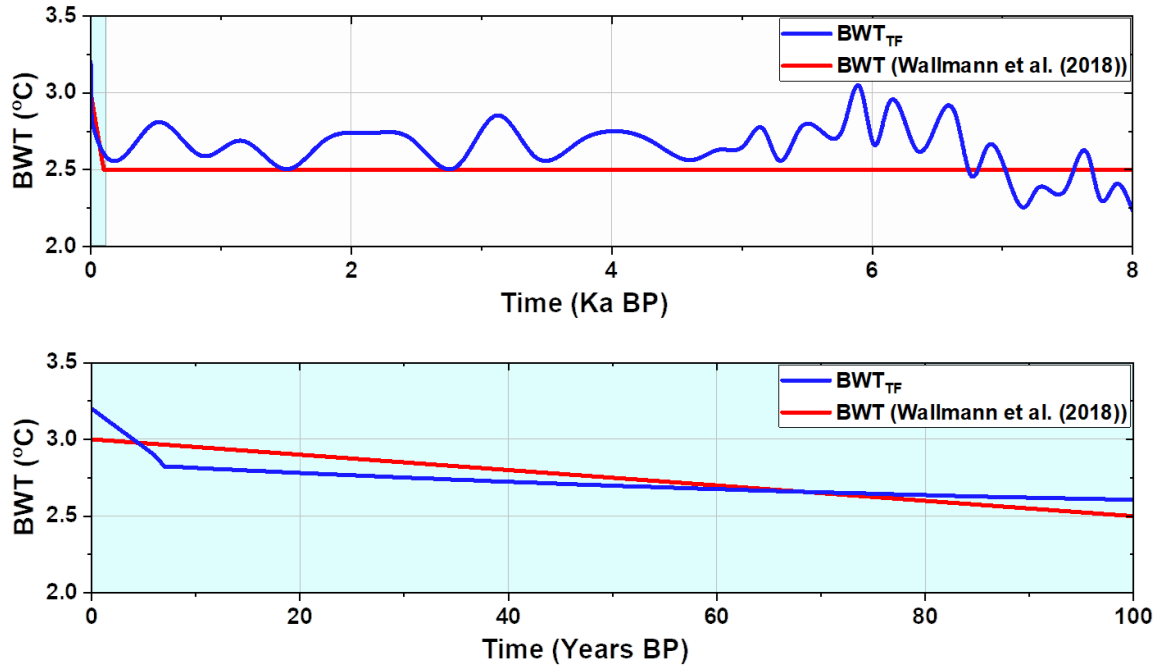


Figure 1.8. BWT_{TF} for the past 8000 years at 327 m water depth on the continental shelf west of Svalbard is based on the data published in Rasmussen et al. (2014). Wallmann et al. (2018) used the red BWT curve for hydrate modelling. They assumed a flat temperature curve with a constant value of 2.5 °C from 8000 to 100 years ago, followed by a linear rise up to 3 °C. Compared to the BWT curve used by Wallmann et al. (2018), the actual BWT_{TF} curve shows oscillations in temperature values. These subtle variations influence formation and dissociation of hydrates in shallower depths and affect the upward migration of methane.

ii. Relative Sea level:

Wallmann et al. (2018) used the University of Tromsø (UiT) ice history and glacial isostatic adjustment model. The corresponding Earth rheology model considered an elastic lithosphere that is 120 km thick and a relatively high viscous (2×10^{21} Pa.s) upper mantle. The model showed a poor match with the observed RSL curve near Prins Karl Forland (Forman et al., 2004). In contrast, the ICE-6G_C model, with a 71 km thick lithosphere and a lower viscosity (0.5×10^{21} Pa.s) upper mantle, showed a better match with the observed RSL variations (Auriac et al., 2016). This discrepancy highlights the need for a reassessment of RSL predictions using alternative models.

1.4.4 Nature of fluid migration patterns in the zone of hydrate destabilization

Despite the observational evidence for the nature of gas seeps at the seabed, very little is known of the fine-scale details of sedimentary structures that carry the gas to the seafloor. Moreover, seismic evidence of a BSR is not identified immediately below the gas seeps,

although a BSR occurs in deeper waters further downslope in that region (Sarkar et al., 2012). The BSR identification over the probable methane hydrate retreat zone is further complicated by the presence of highly heterogeneous glaciomarine sedimentary units, in which there are not enough well-stratified layers, and identifying a cross-cutting BSR becomes difficult. Resolving these uncertainties is crucial for a better understanding of hydrate dissociation.

1.4.5 Hydrate reservoir characterization

Reservoir heterogeneity and fluid migration can strongly influence gas hydrate distribution. Hydrate accumulates in porous and permeable sediments or fractured fine-grained clay beds. For accurate methane hydrate reserve estimates in complex glaciomarine sediments, reservoir characterization is essential. However, seismic images lack the resolution to highlight detailed lithology and their lateral and vertical extents in the shallowest glaciomarine sediments, limiting our understanding of reservoir heterogeneity, including predictions of releasable methane in response to future warming. High-resolution seismic imaging is necessary to enhance subsurface stratigraphy and reservoir characterization.

1.5 Objectives:

The thesis aims to address the aforementioned knowledge gaps in terms of the following objectives–

1. To understand the dynamics of gas hydrate stability in response to the increasing temperature of the Atlantic water and RSL changes since the last deglaciation.
2. To improve the glaciomarine seismic stratigraphy, depositional patterns and the environment of the seepage sites in the interfan region.
3. To ascertain the nature of lithological heterogeneity that either facilitates or impedes subsurface fluid flow beneath the methane seeps.

1.6 Svalbard Glacial History

The glacial history of Svalbard is deciphered from seismic stratigraphic analysis and age models determined from boreholes, for example, Ocean drilling program (ODP) Sites in the Yermak Plateau provide details of northern Svalbard, while ODP Site 986 between Belsund and Isfjorden trough mouth fan (TMF) provide age constraints of glacial development of western Svalbard (Leg 162, Forsberg et al., 1999).

Recent advances in seismic stratigraphic studies of the Svalbard and Barents Sea region have produced a more refined framework, providing a comprehensive understanding of the glacial history and evolution of the regional ice sheet. The seismic stratigraphic framework for this

region has been built incrementally. Laberg and Vorren (1996) established initial seismic stratigraphic framework for the Bear Island TMF, while Hjelstuen et al. (1996) developed the stratigraphic framework for the Storfjorden TMF. Vorren et al. (1991), Faleide et al. (1996), and Ryseth et al. (2003) reconstructed the glacial history of the western and southwestern Barents Sea. Geissler and Jokat (2004), Sarkar et al. (2011) and Mattingdal et al. (2014) contributed to understanding the glacial history of the northern and northwestern Svalbard margin. Building on this foundation, Alexandropoulou et al. (2021) provided a comprehensive seismic stratigraphic framework covering the past 2.7 million years for the Barents Sea and western Svalbard. Their study identified three distinct phases of glacial development. These are summarized below.

Phase 1 (2.7–1.5 Ma): This phase marked the commencement of early glaciation in northern Barents Sea, with two-stage glacial expansion, such as around 2.7–2.5 Ma and later around 1.95–1.78 Ma, as evidenced by glaciomarine sedimentation caused by glacial ice build-up in northernmost Svalbard, while more ice distal conditions prevailed in the southwestern Barents Sea (Figure 1.9a). Around 2.5 Ma, the build-up of the Sjubrebanken TMF marked the development of paleo-ice stream at the northwestern Svalbard (Sarkar et al., 2011; Alexandropoulou et al., 2021). Elevated mountainous terrain, resulting from preglacial uplift in Svalbard and northwestern Barents Sea, facilitated glacial ice build-up. The second stage of glacial expansion resulted in a wider glacial cover that extended beyond the Svalbard coastline encompassing western Barents Sea, but excluding the southern Barents Sea. The decline in river-sourced smectite from the Siberian shelf, potentially due to ice sheet obstruction and the development of northwestern glacial fan, coincided with an increase in illite, the most common glacial clay mineral (Figure 1.9b). This suggests that glacial erosion increased during the early Northern Hemisphere Glaciation (~ 2.7 Ma, Alexandropoulou et al., 2021).

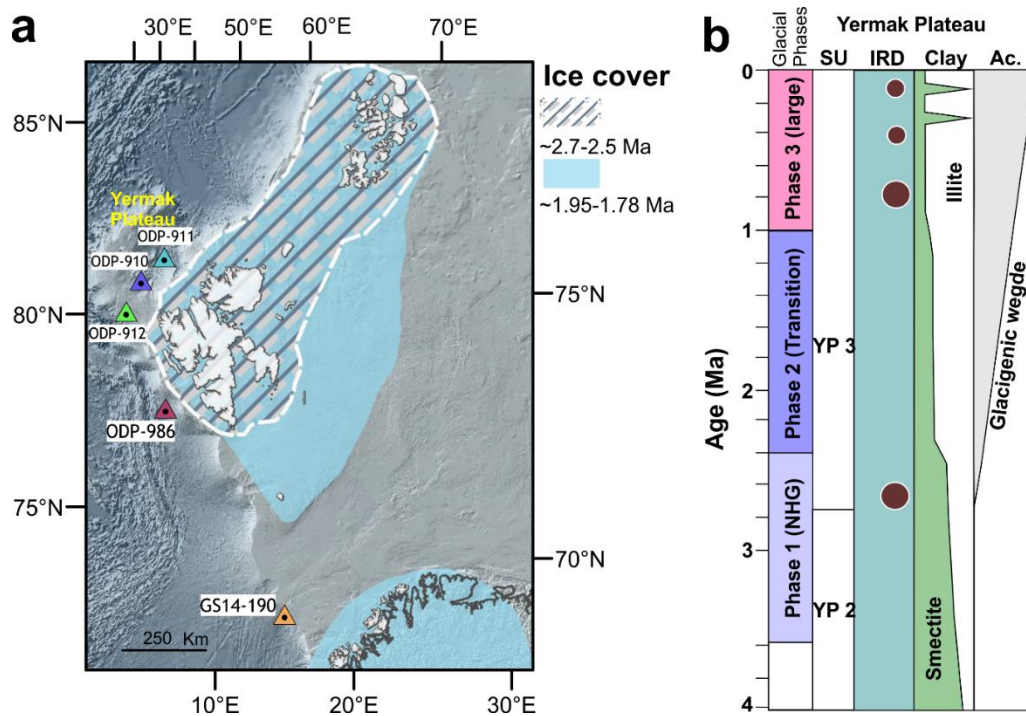


Figure 1.9. Maximum ice sheet limit in the Svalbard and Barents Sea regions during 2.7–1.5 Ma. **(a)** The Svalbard and Barents Sea region experienced a two-step ice growth during 2.7–1.5 Ma. The first phase occurred from 2.7 to 2.5 Ma, and the second from 2.95 to 1.78 Ma. Adapted from a figure by Alexandropoulou et al. (2021). **(b)** Schematic stratigraphy of the Yermak Plateau illustrates ice-rafted debris (IRD) flux indicated by the diameter of circles, glacial wedge development, and clay mineral proportions (smectite and illite) at the Yermak Plateau ODP sites, providing insights into the northwestern region's glacial history. Adapted from Knies et al. (2009), SU: Seismic unit; Ac.: Seismo-acoustic interpretation. The base of seismic unit YP3 marks the onset of northwestern glaciogenic wedge development in the northwestern Svalbard (Eiken and Hinz, 1993).

Phase 2 (1.5 to 0.42 Ma ago): During this transitional phase, there was a wider glacial coverage in the Barents Sea, and repeated glacial advancements to the western Svalbard shelf-edge resulting in increased sedimentation from 1.5 to 1.2 Ma than the previous periods (Figure 1.10a, Sejrup et al., 2005). In southwestern Barents Sea, there was an enhancement in sedimentation rate from ~ 60 cm/kyr to 142 cm/kyr during 1.78–1.5 Ma. This phase witnessed glacial coverage extending beyond the Svalbard margin, supported by the IRD record from ODP site 986. Presence of the IRD after 1.78 Ma indicates shelf-edge glaciations in western Svalbard after that period (Butt et al., 2000). Around 1–0.94 Ma, a large-scale ice sheet expansion occurred in the Barents Sea and Svalbard, coinciding with the Mid-Pleistocene

Transition between ~ 1.25 – 0.7 Ma, during which there was a substantial global increase in ice volume (Mudelsee and Stategger, 1997). This period marked the transition from shorter, more frequent glacial cycles (approximately 41,000 years) to longer, less frequent glacial cycles (approximately 100,000 years). The Kongsfjorden trough formed somewhere between 1.5– 0.99 Ma, because of ice streaming, resulting in the deposition of the KTMF (Sarkar et al., 2011).

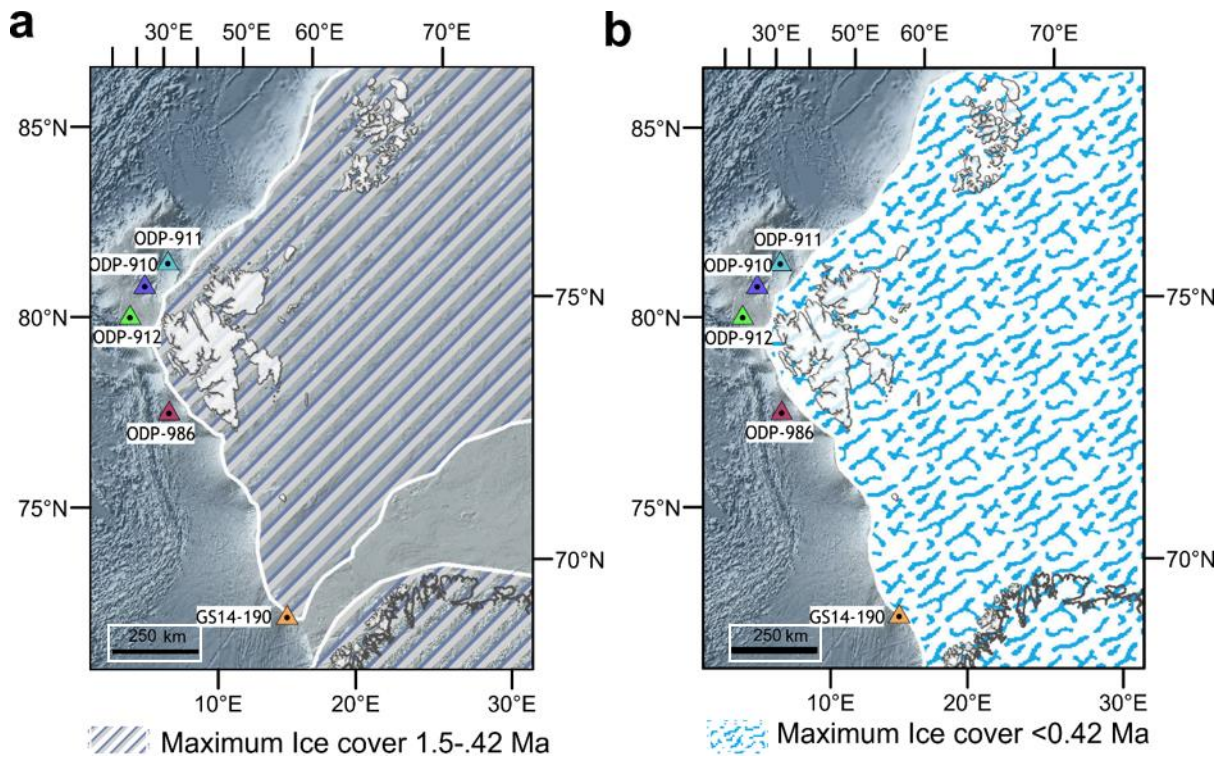


Figure 1.10. Maps showing the maximum ice sheet limit in the Barents Sea and Svalbard regions from ~ 1.5 Ma to the Last Glacial Maximum. **(a)** Maximum ice coverage during 1.5–0.42 Ma and **(b)** maximum ice extent < 0.42 Ma. Adapted from a figure by Alexandropoulou et al. (2021).

Phase 3 (~ 0.42 Ma ago to the present): The development of glacial fans continued during the glacial episodes. The maximum extent of the ice sheet attained during this phase is shown in Figure 1.10b. The glacial sedimentary deposits from the last 0.4 Ma are associated with the Early and Late Saalian, and Weichselian glaciations. Glaciations during Saalian resulted in high sedimentation in NW (281 cm/kyr) and SW Barents Sea (197 cm/kyr) compared to the Weichselian glaciations (30, 56 cm/kyr in NW and SW Barents Sea, respectively) indicating decreasing erosional capacity of the glaciers.

1.7 Past RSL variations

The RSL considers both eustatic sea level changes and vertical seafloor movements at a specific location. Glacial-isostatic adjustment (GIA) is the mechanism through which the crust responds to variations in ice loading. Ice loading causes crustal down-bending (Figure 1.11a), and when the ice sheet melts, the crust beneath it rebounds (Figure 1.11b). The isostatic rebound rate depends on the lithosphere thickness, and mantle viscosity.

The GIA models help to derive isostatic response of various ice history scenarios and crustal deformation in response to ice load and extent, and the timings of glaciations and deglaciations. For example, Auriac et al. (2016) considered six ice-loading scenarios in the Barents Sea and Svalbard, by considering ice sheet configuration throughout the last glacial cycle (**Table 1.1**). The ice loading scenarios were generated following different methods; therefore, they have their own strengths and limitations. Two primary approaches were followed to construct the ice-history scenarios:

Forward modelling: These models use forward modelling approach to simulate ice-sheet evolution by imposing a climate model. Examples of forward modelling scenarios for the Barents Sea include NO5, SO4 and UiT.

Inverse modelling: These models employ data on RSL and the past records of ice margins to determine patterns of ice loading. Examples of inverse modelling scenarios include ANU, ICE-5G and ICE-6G_C. These scenarios are constrained by observed RSL and geological reconstruction of ice history data, but they can be more sensitive to the quality and interpretation of that data. Forward modelling scenarios are less constrained by data, but they can be more sensitive to the assumptions made about the ice sheet model and its parameters. The inherent uncertainties associated with both the approaches necessitate the need to assess multiple ice load scenarios for interpreting results.

The GIA model conceptualizes the Earth as a sphere with an elastic lithosphere and two viscoelastic layers representing the upper and lower mantle. It solves equations governing Earth's deformation in response to changes in loading by ice and ocean, as well as alterations in the gravity field over time. The combined output of Earth deformation and sea surface perturbation provides relative sea-level changes over time. The comparison of GIA predicted RSL with the observed RSL records on the western coast of Svalbard (Figure 1.11c) reveals a close match of the observed values with the ICE-6G_C scenario (Figure 1.11d).

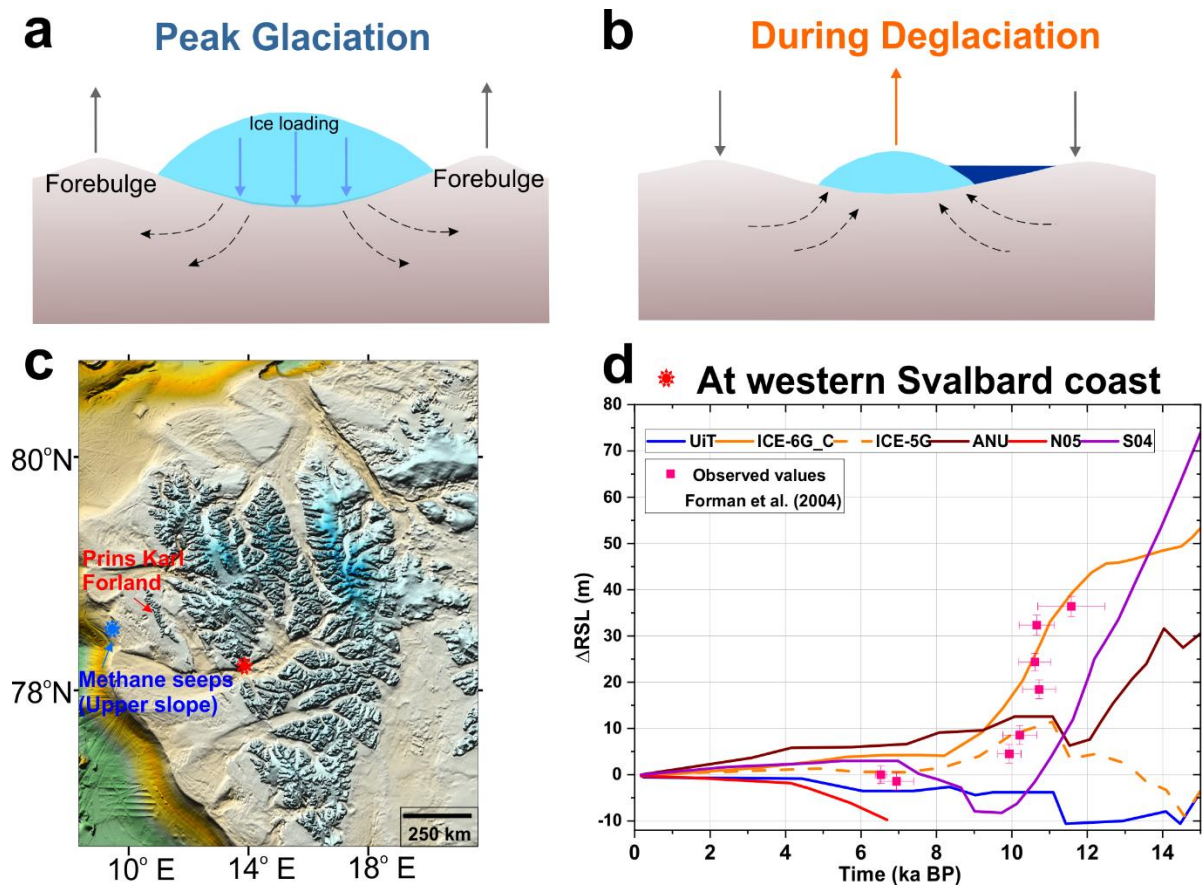


Figure 1.11. Interaction of glacial mass and the crust, and resulting changes in RSL. **(a)** Schematic of glacial isostasy showing crustal downbending beneath ice mass and peripheral forebulge formation due to mantle flow. **(b)** During deglaciation, as the ice sheet melts, its weight diminishes, allowing the previously depressed crust to rebound upwards. Simultaneously, the peripheral forebulges collapse. **(c)** IBCAO bathymetry showing the methane seepage site and elevated beaches found on land at Daudmannsøyra providing information about past RSL variations. **(d)** Observed RSL variations at Daudmannsøyra (Forman et al., 2004) and modeled RSL curves at that site using six different GIA scenarios (Auriac et al., 2016).

Table 1.1. Details of six different GIA models from Auriac et al. (2016)

Ice-history Scenario and GIA model	Spatial coverage	Temporal coverage (ka BP)	Thickness of lithosphere (km)	Viscosity of upper mantle (10^{21}Pa.s)	Viscosity of lower mantle (10^{21}Pa.s)	References
ICE-5G	Global	122-0	96	0.5	1	Peltier (2004)
S04	Local	32-12	46	0.3	10	Siegert and Dowdeswell (2004)
N05	Local	122-0	71	0.5	2	Naslund et al. (2005); Naslund et al. (2006);
ANU	Global	122-0	120	0.5	2	Lambeck et al. (2010)
ICE-6G_C	Global	26-0	71	0.2	2	Argus et al. (2014); Peltier et al. (2015)
UiT	Local	35-7.5	120	2	50	Auriac et al. (2016)

1.8 Overview of the Chapters

Chapter 2:

This chapter provides the details of the two- and three-dimensional multichannel seismic data used in this study along with data processing details. Additionally, it describes the modelling approach utilized for the numerical simulation of methane hydrates in response to the changes in bottom water temperatures (BWTs) and relative sea levels (RSLs).

Chapter 3:

This chapter presents multiple scenarios depicting the evolution of the methane hydrate system over the past 11,000 years off west Svalbard. Given the uncertainty in determining precise RSL and BWT for the western Svalbard slope region, four plausible sets of RSL and BWT curves were considered. The modelling aimed to assess whether, by the end of the simulation, hydrates are completely dissociated, leaving behind a pore water freshening trend that matches the salinity of pore fluids recorded during the MeBo drilling. The results provide clues related to the evolution of methane hydrate over the past 11,000 years. The triggering factors for hydrate dissociation and formation were assessed. Additionally, a separate modelling was carried out to investigate the impact of Atlantic water warming after the late 1970s on methane hydrates and the pattern of methane seeps. This simulation accounted for seasonal fluctuations in BWT. The results were compared with observed modern seepage patterns. The comprehensive scenario testing helps to understand hydrate dynamics, duration of seeps, their patterns and fluxes, which are critical aspects for reconstructing past methane release episodes. The simulations also address the debate whether depressurization caused by RSL fall is the sole reason for hydrate destabilization.

Chapter 4:

This chapter presents the Weichselian seismic stratigraphy of the interfan outer shelf and slope located between the Kongsfjorden and Isfjorden Trough Mouth Fans using 2D high-resolution SYSIF seismic and 2D air gun seismic data. The study provides a detailed description of various seismic units and variations in seismic facies. These results were used to reconstruct the glaciomarine depositional environment, influenced by the slow-moving ice sheet and its waxing and waning behaviour. The glaciogenic sediments were primarily composed of glacial debris flow deposits, and the detailed internal morphology and stacking patterns are presented. An attempt was made to simulate the behaviour of debris flow that best matches seismic observations.

Chapter 5:

This chapter illustrates the stratigraphic controls on fluid flow patterns at the upper continental slope off west Svalbard. Utilizing SYSIF seismic and 3D seismic data from the methane seepage site, the chapter provides evidence of subsurface fluid accumulation and migration. Additionally, it provides a detailed description of reservoirs and seals that govern fluid accumulation and entrapment. Seismic anomalies associated with gas and fluid migration through fractures are discussed based on 3D seismic attributes. The 3D data aids in mapping rudimentary glaciogenic reservoirs that store gas over the gas hydrate retreat zone. Thermal modelling validates the occurrence of gas-charged seismic anomalies and pore-water salinity obtained from drilling. The chapter also explores the possibility of methane migration along the base of methane hydrate stability and its escape at the landward edge of methane hydrate stability. Finally, it presents a conceptual fluid flow model for the interfan region, providing the most plausible explanation for the observed methane seeps.

Chapter 6:

This chapter provides a concise overview and synthesis of the key findings derived from the research, along with an outlook on potential avenues for future research.

Chapter 2

Data and Methodology

2.1 Introduction

This chapter provides a description of two-dimensional (2D) and three-dimensional (3D) seismic data sets and methodologies employed in the present work. The seismic data aided in interpreting the glaciomarine stratigraphy, fluid flow patterns and conducting numerical simulations of behaviour of methane hydrates, in response to variations in ocean temperatures and relative sea level changes. Each seismic data set utilized in this thesis has been obtained and processed during different seismic data collection cruises.

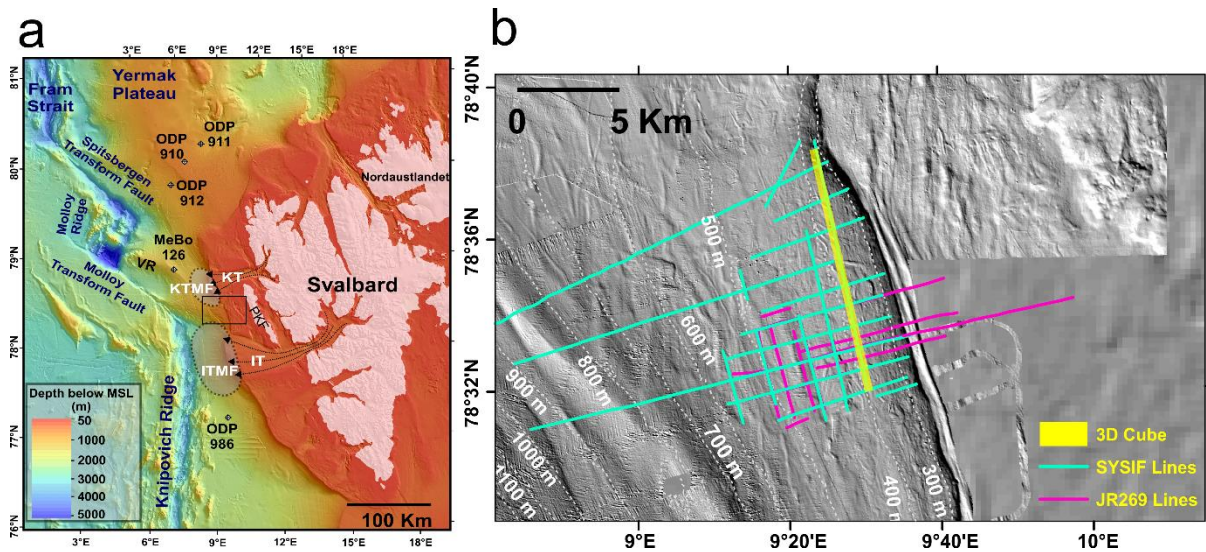


Figure 2.1. Seismic data coverage map. **(a)** IBCAO bathymetry map showing the mid-oceanic ridge and transform system west of Svalbard, Kongsfjorden and Isfjorden cross-shelf troughs (KT and IT). Seismic data were collected in the interfan area (black rectangle) between the Kongsfjorden and Isfjorden Trough Mouth Fans (KTMF and ITMF). **(b)** JR269B 2D MCS were collected on the shelf and slope, while SYSIF seismic lines and 3D P cable seismic data were collected on the upper slope.

2.2 Seismic Data Sets

The study uses three different seismic data sets, viz., 2D multichannel seismic (MCS) data collected using an airgun source, 2D high-resolution SYSIF seismic, 3D seismic data (Figure 2.1). Seismic data collection was aimed to image pathways for gas migration and lithological heterogeneity beneath regions where methane actively escapes into the ocean. The following section provides a discussion of the acquisition and processing details for each dataset.

2.2.1 JR269B 2D MCS data:

A multichannel seismic (MCS) survey was conducted aboard the RRS James Clark Ross during the cruise JR269B in August and September, 2012 when the survey region was free of sea ice. This survey covered the shelf area west of Prins Karl Forland and the interfan region between the Isfjorden and Kongsfjorden Trough Mouth Fans in the upper slope (Figure 2.1). The technical specifications related to seismic data acquisition are provided in Table 2.1. The seismic streamer comprised sixty 1-meter hydrophone groups, spanning an active length of 60 m. GI air gun was used as seismic source with a volume of 1.47 l (0.73 l generator and 0.73 l injector) towed at 1.5 m and operated in harmonic mode at a nominal pressure of 140 bar (2000 psi). During the cruise, the shot interval was maintained at 6 s corresponding to 12 m shot spacing. The seismic data recording duration was 3 seconds, and was sampled at a rate of 0.5 ms. The navigation details were extracted from the vessel's differential GPS system. The airgun source was positioned 2.5 m to the starboard side of the vessel's axis. The length of the source cable from the towing point was 30 m (considering 20 m for horizontal distance only). The streamer was towed from the port side swinging boom, positioned at a distance of 13.5 m from the vessel's axis and located at the stern side. The lead-in cable length from the towing point on the boom was 30 m. The active section of the streamer was 60 m, followed by a 5-m tail section. At the far end of a 20-m rope, a small tail buoy was attached.

Table 2.1. 2D MCS data acquisition parameters

GI-air gun (harmonic mode)	140 bar (2000 psi)
Streamer	60 hydrophone groups (7 elements) at 1 m spacing; immersion: 2 to 5 m
Shooting rate	5 seconds
Seismic sampling interval	0.5 ms
Recording length	3 s
Delay in starting recording	0 s

2D MCS data processing:

(i) The raw shots in SEG-D format were first loaded and 2D marine geometry was set up. Geometry information was incorporated into each trace by defining the locations of both the hydrophones and source, (Table 2.2):

Table 2.2. Location of the air gun and streamer with reference to the vessel axis

Offset direction	Air gun offset (m)	Streamer offset (m)
X	-2.5 (perpendicular to vessel axis)	13.5
Y	75.2 (parallel to vessel axis)	90
Tow depth (m)	1.5	3.5

(ii) Navigation information was integrated into the shots, and the acquisition geometry was configured, considering a 12-m spacing between sources and a 1-meter interval between receivers. Binning in the Common Midpoint (CMP) domain was performed with a 2-m spacing between adjacent CMPs. Once the geometry was set up, it was integrated into the seismic trace headers, and the location of the CMP was computed.

(iii) Trace editing: Trace editing process involves a detailed visual inspection of all input shot records to evaluate seismic trace quality. Traces with low amplitudes or excessive noise were removed. For example, consistently bad traces 13, 15, 45, 52, 53 and 57 were removed from the shot records. To mitigate the delay between triggering and the activation of the airgun, a static correction of -42.0 milliseconds was performed.

(iv) Filtering: Frequency domain filtering was performed to enhance the signal-to-noise ratio. A power spectra was generated to define an appropriate zero-phase Ormsby bandpass filter. A gradual shift between the "pass" and "cut" frequencies was adopted to mitigate the issues associated with the ringy edges. To accomplish this, a trapezoidal shaped filter was designed. The filter parameters were set with a low cut and high cut frequencies of 10 Hz and 1000 Hz respectively, and a "pass" band ranging from 30 Hz to 600 Hz.

(v) True amplitude recovery: This processing step minimizes the reduction in amplitude associated with the propagation of seismic waves originating from a point source. To

counteract this effect, a correction for geometric spreading was implemented, alongside the incorporation of a gain function proportional to the square of time.

(vi) Normal moveout (NMO) correction and stacking: Normal moveout correction was performed considering a constant velocity of 1475 m/s (average sound speed in water) followed by CMP stacking.

(vii) Stolt migration: This technique, also known as Frequency-Wavenumber migration, involves a two-dimensional Fourier transform of the stacked section, represented as a function in the time and distance domains, denoted as $f(t, x)$. It is transformed into the F-k (frequency-wavenumber) domain. Post-stack Stolt migration effectively collapses diffractions while accurately positioning dipping reflectors at their true subsurface locations. The migration primarily occurs in the F-K domain, followed by a 2D inverse Fourier transform to restore the data to its original time and distance representation ($f(t, x)$). A velocity of 1475 m/s was used in migrating this dataset. The vertical resolution of this data is ~ 4 m considering the dominant frequency in the top 50 m is ~ 100 Hz.

2.2.2 Système Sismique Fond (SYSIF) data acquisition and processing

SYSIF source:

The seismic deep-towed acquisition has been developed to overcome major limitations of conventional sea surface seismic acquisition in deep waters, such as the lack of signal penetration and the influence of out-of-plane reflection events caused by a large Fresnel zone (Marsset et al., 2010). The SYSIF deep-towed seismic system (Figure 2.2), developed by Ifremer, enables the acquisition of sub-metric scale seismic data, reaching water depths up to 6000 m (Ker et al., 2010). The Janus-Helmholtz acoustic transducer was used as a SYSIF seismic source. It consists of a structure called the Janus driver placed within a cylindrical casing (Le Gall, 1999). By combining mechanical and fluid resonances, this setup achieves a broad frequency range, generating a consistent acoustic signal that is highly suited for accurately characterizing subsurface lithology. During the cruise JR269A, two different SYSIF seismic sources were used, such as the JH650-6000 and JH250-6000 systems (Table 2.3). JH250-6000 system was used to image the upper continental slope of the interfan region, while on the Vestnesa Ridge both the systems were used. A 3 kVA class D-class power amplifier was used to power the acoustic source. The deep-towed SYSIF vehicle's attitude and heading were logged along with altitude information obtained through a pressure

sensor integrated into the tow fish. The ultrashort baseline (USBL) provided information of the location with respect to the ship. Onboard quality control procedures were conducted for navigation data, encompassing USBL, attitude (roll, pitch, and yaw), and altitude measurements.

SYSIF Receiver:

The streamer comprises a lead-in cable and two hydrophones. It had a length of 15 m. The first channel consisted of a single hydrophone positioned 10 m away from the seismic source. The second channel, located 15 m from the source, was equipped with six parallel arranged hydrophones spaced 30 cm apart. Both channels were equipped with a 26.3 dB pre-amplifier and a bandpass filter between 100 and 3000 Hz. Data sampling was accomplished using an A/D converter at a rate of 10 kHz.

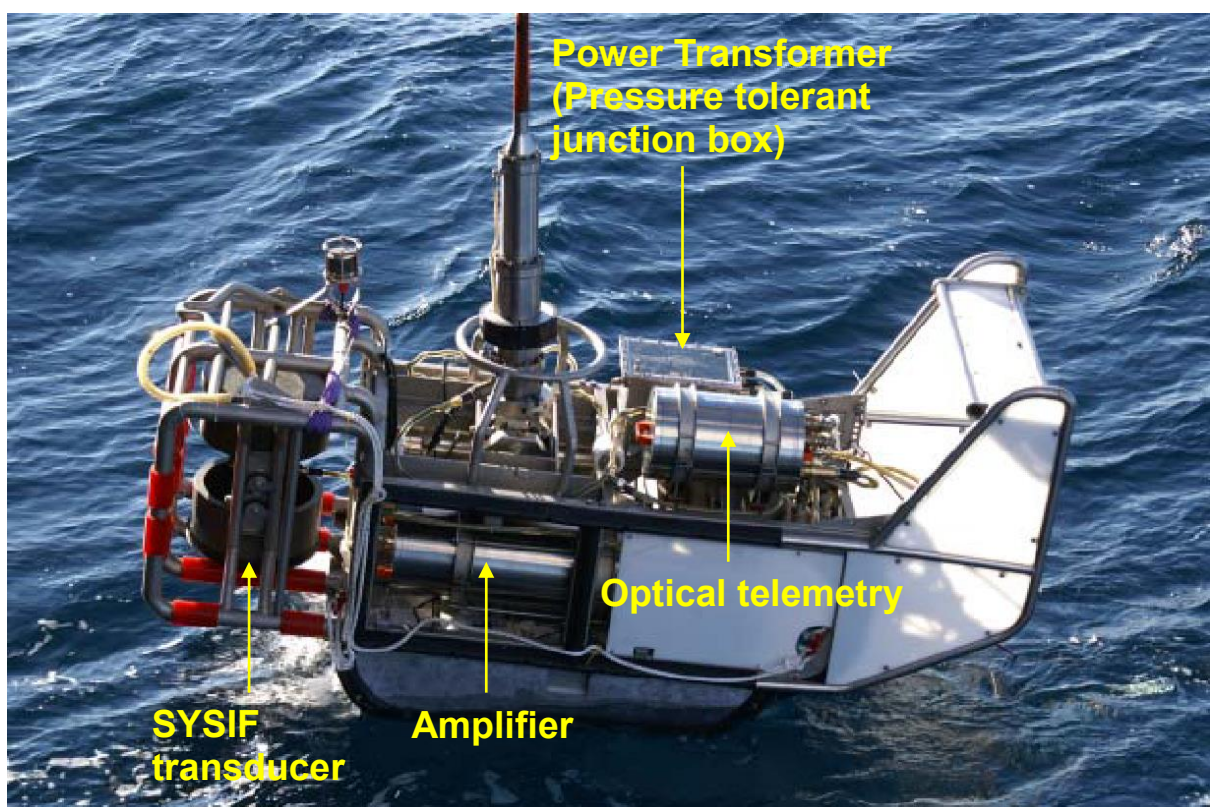


Figure 2.2. SYSIF seismic vehicle carrying the JH650-6000 system as the source. Figure adapted and modified from Leon et al. (2009)

Table 2.3. SYSIF source specifications

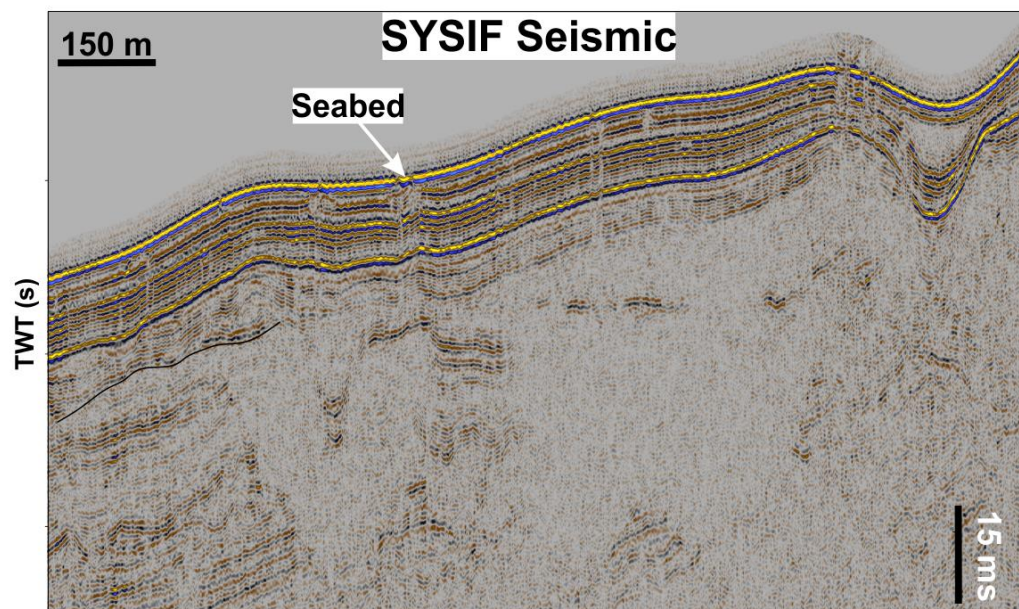
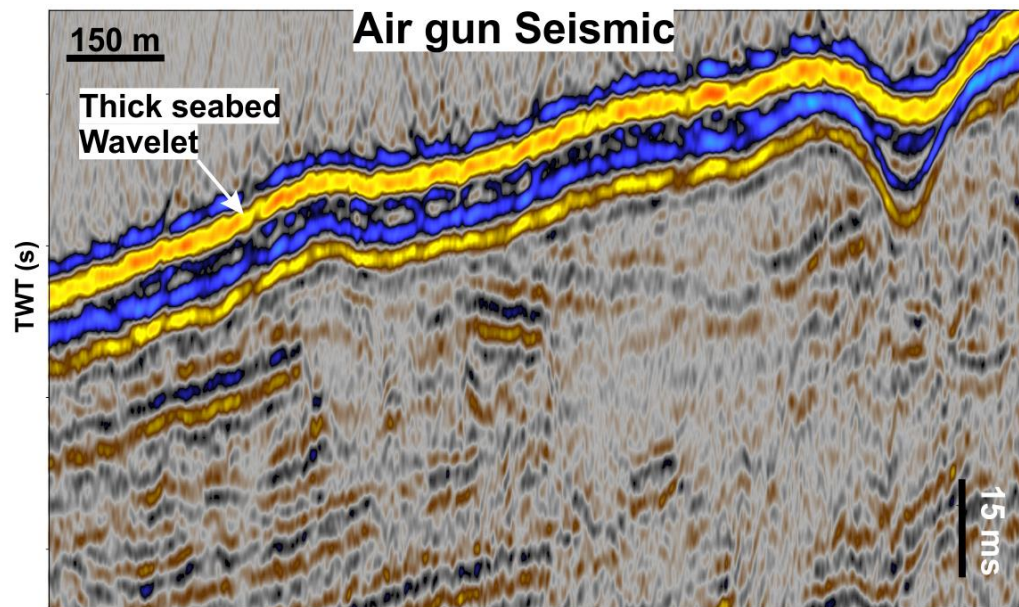
Source Type	JH250-6000	JH650-6000
Dimension (cm)	Height – 112 Diameter – 72	Height – 61 Diameter – 45
Source signal	100 ms frequency-modulated (220-1050 Hz), output 196 dB (referenced to 1 μ Pa per 1 meter)	frequency-modulated signal 580-2200 Hz
Weight (kg)	450	90

SYSIF data processing:

The processing of the SYSIF lines was carried out by Ifremer, and the SYSIF data processing workflow involved four key steps:

- i) Processing of the navigation data and accurate positioning of the source and receiver.
- ii) A source signature deconvolution with the far-field recording at normal incidence to increase the vertical resolution (Ker et al., 2010). In the convolutional model, the seismic trace is expressed as the convolution of the source signature, $s(t)$, and the reflectivity series, $r(t)$. Noise is an additive component to it. Fourier transformation of the seismic data and the source signature helped to carry out the source deconvolution.
- iii) Applying an altitude correction as the altitude of the SYSIF vehicle was 100 m above the seabed.
- iv) Migration with the water velocity (1500 m/s).

The lateral and vertical resolution of the SYSIF lines is 3 m and 0.5 (dominant frequency of 750 Hz)-1 m (dominant frequency of 400 Hz), respectively. A comparison between the 2D air gun and SYSIF seismic data clearly shows superior resolution of the latter (Figure 2.3).



Airgun data

SYSIF data

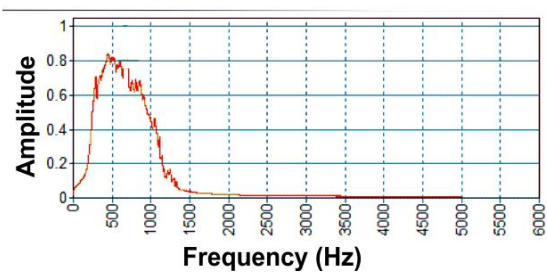
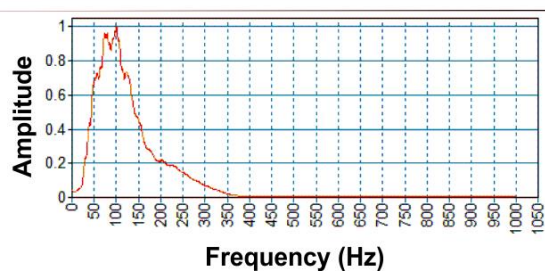


Figure 2.3. Comparison between air gun 2D and SYSIF seismic data. In the case of air gun seismic data, the thick seabed wavelet obscures the underlying geological features. The SYSIF seismic clearly shows the stratigraphic details.

2.2.3 3D seismic data acquisition and processing:

The Centre for Arctic Gas Hydrate, Environment and Climate (CAGE) acquired a 3D P-cable seismic data in the slope where methane seeps were reported (380–395 m water depths) off west Svalbard. The 3D seismic data acquisition system consisted of a cross-cable towed orthogonal to the ships's movement (Figure 2.4). Multiple seismic lines were simultaneously acquired using an array of solid-state streamers, enabling extensive coverage with closely spaced in-line intervals. As the vessel moved forward, the cross-cable stretched out with the help of two paravanes that deflected away sideways. Two trawl winches were used to manoeuvre the paravanes. The cross-cable was 223 m long between the two paravanes and was connected with 14 streamers. The average separation between streamers was 12.5 m, but the curved configuration of the cross-cable lead to streamer shortening (6-12 m) along the crossline direction. Streamers, each measuring 25 m in length and equipped with eight hydrophone groups, had an analog-to-digital data conversion module. The receiver spacing in each streamer was 3.125 m. A signal cable extended from the P-cable winch and linked to the cross-cable at the starboard side. It supported the power supply and transmission of data. Geometrics GeoEel software was used to record the digital data. A mini-GI gun with a total volume of 0.49 l (Generator volume of 0.245 l and injector volume of 0.245 l) was used as a source.

The seismic data had six in-lines with a line spacing of 60 m covering a 5-km² area (12.8 km long and 0.3 km wide). The data was acquired from an area, where methane flares have been consistently discovered for many years. The pre-processing of the data included sorting of navigation data, 3D geometry set up, CMP binning, band pass filtering, true amplitude recovery, gap filling by interpolation, normal moveout correction with a constant 1500 m/s velocity, stacking, and post-stack 3D F-k migration.

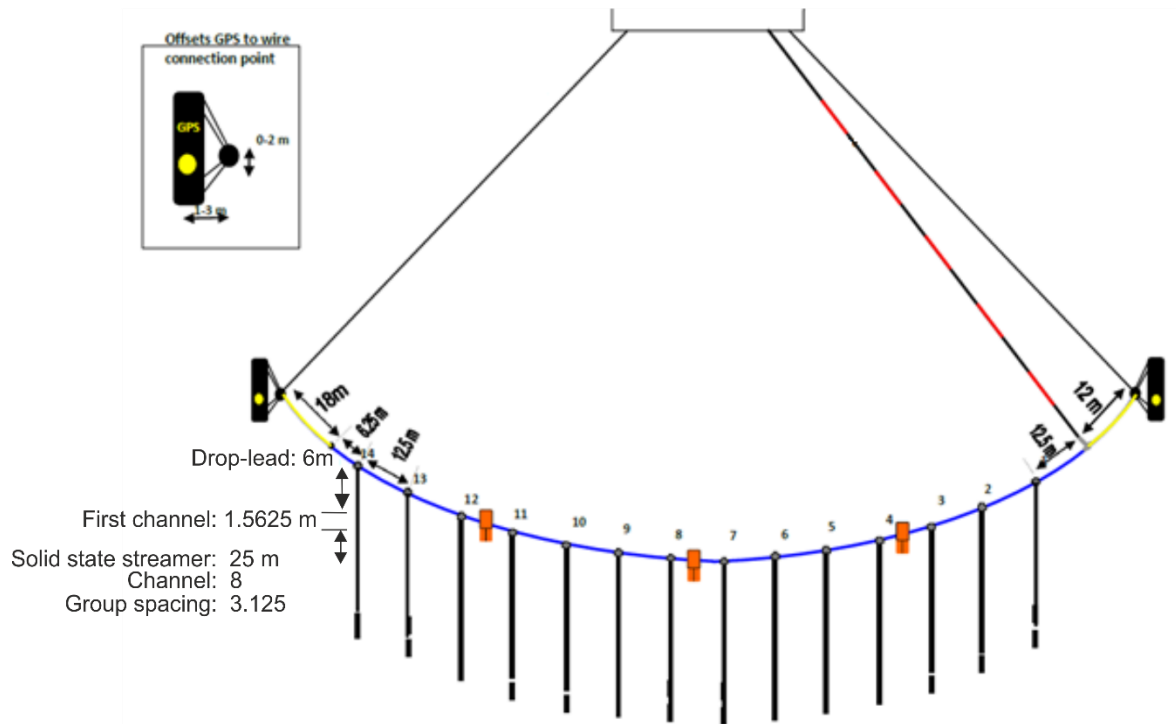


Figure 2.4. Schematic illustration of the 3D P-cable seismic system showing the curved cross-cable and array of seismic streamers (adapted from Andreassen, 2023).

2.3 Seismic attributes

3D seismic data was analyzed to understand subsurface geological features. This included seismic attribute analysis, which helped to decipher key features like lithological variations, fluid distributions, frequency anomalies, and subtle structural discontinuities such as fractures. Specific attributes were chosen and are explained in more details below.

2.3.1 Envelope Attribute

Extracting envelope attribute from raw seismic data requires transformation to an analytical signal, using the Hilbert transform. This separation of amplitude and phase simplifies analysis, allowing for tasks like envelope computation and instantaneous phase determination.

A real signal is converted into a complex signal by adding the real part with its imaginary Hilbert transform component. Mathematically, the Hilbert transform of a signal is given by

$$s(t) = r(t) * h(t) \quad (1)$$

$$\text{Where, } h(t) = \frac{1}{\pi t}$$

In the Fourier domain,

$$H(\omega) = -j \operatorname{sgn}(\omega) \quad (2)$$

$$S(\omega) = -j \operatorname{sgn}(\omega) R(\omega) \quad (3)$$

Where $H(\omega)$, $S(\omega)$, and $R(\omega)$ are Fourier transforms of $h(t)$, $s(t)$, and $r(t)$ respectively.

Mathematically, a complex trace is given by

$$z(t) = r(t) + js(t) \quad (4)$$

Here, $r(t)$ and $s(t)$ are both real-valued, casual functions.

Using equations 2 and 3, the equation 4 can be written in Fourier domain as:

$$Z(\omega) = -j(\operatorname{sgn}(\omega)) R(\omega) \quad (5)$$

On simplification

$$Z(\omega) = 2R(\omega) \quad \omega > 0 \quad (6a)$$

$$= R(\omega) \quad \omega = 0 \quad (6b)$$

$$= R(\omega) \quad \omega < 0 \quad (6c)$$

Thus, a complex trace has the same frequency spectrum as the real stacked seismic signal $x(t)$. The instantaneous amplitude or envelope $a(t)$, instantaneous frequency $\omega(t)$, and instantaneous phase $\theta(t)$ of the seismic trace are derived by the following relations

$$a(t) = \sqrt{r^2(t) + s^2(t)} \quad (7)$$

$$\omega(t) = \frac{d\theta}{dt} \quad (8)$$

$$\theta(t) = \left[\frac{s(t)}{r(t)} \right] \quad (9)$$

2.3.2 Sweetness Attribute

The sweetness attribute highlights areas with high amplitudes and low frequencies, typically corresponding to sandy intervals, which display the highest sweetness values. Conversely, sediments with lower amplitudes and higher frequencies, such as thinly bedded shales, exhibit lower sweetness values. This attribute combination is particularly useful for identifying gas reservoirs. The sweetness attribute is usually computed by taking the amplitude envelope and dividing it by the square root of the instantaneous frequency. In mathematical terms, it is represented as:

$$\text{Sweetness} = \frac{a(t)}{\sqrt{\omega(t)}} \quad (10)$$

Here $a(t)$ is the instantaneous amplitude or envelope, and $\omega(t)$ is the instantaneous frequency.

2.3.3 Similarity Attribute

Mathematically discontinuities in seismic attributes and waveform can be calculated by assuming each seismic trace as a vector, each with N samples.

$$\mathbf{x}_i = \begin{bmatrix} x_{1i} \\ x_{2i} \\ \vdots \\ x_{Ni} \end{bmatrix}$$

This implies that a trace represents a positional vector within an N-dimensional sample space, characterized by N mutually perpendicular axes. The similarity between each seismic trace can then be estimated by constructing its covariance matrix.

First, a sample waveform is calculated from traces within a specified window. For example, an analysis window includes 9 traces. In the first step, a sample vector is created from the nine traces. More such sample vectors are extracted by shifting the sample vector by a value of k, parallel to apparent dips. These sample vectors are arranged in a row forming a data matrix. In

the second step, each matrix column is correlated and cross-correlated to form a covariance matrix.

If $K=1$, then there are 3 rows and 9 columns in a data matrix. The covariance matrix is mathematically calculated by the following expression (Korn & Korn, 1968),

$$c_{ij}(t, p, q) = \sum_{k=-K}^{+K} [u_i(t + k\Delta t - px_i - qy_i) - \langle u(t, p, q) \rangle] [u_j(t + k\Delta t - px_j - qy_j) - \langle u(t, p, q) \rangle] \quad (11)$$

Where $\langle u(t, p, q) \rangle$ is the mean of traces within the window, u_i is the amplitude of the i^{th} trace, x_i and y_i are the distances of the i^{th} trace along two orthogonal direction measured from the centre of the window; apparent dips (p and q) in ms/m. The covariance matrix gives the representation of how the data varies within itself. The first eigenvalue represents the maximum variance in the data, and only a few eigenvalues and eigenvectors represent 95 % dataset variability. The coefficients of the corresponding eigenvector represent the proportion in which each sample value deviates from the sample waveform. Mathematically, eigenvalues and eigenvectors are estimated by the following equation (Korn & Korn, 1968),

$$Cv^{(m)} = \lambda_m v^{(m)} \quad (12)$$

The first eigenvalue, for the coherence or similarity estimate is given by

$$c_e \equiv \frac{\lambda_1}{\sum_{j=1}^J C_{jj}} \quad (13)$$

Physically, λ_1 denotes the energy of traces in the direction of the eigen vector v_1 , and C_{jj} is the variance of the j^{th} trace, representing, the energy attribute. If the trace energy in the direction of the eigenvector is same as the energy of the waveform within the sampled window, then $c_e = 1$, and in other directions it is less than 1.

The above process is repeated for each point in the data. According to the defined window length centred on each point in the volume, the data matrix is extracted to form the covariance matrix. From the covariance matrix the first eigenvalue is used to estimate the ratio of the energy represented by first eigenvalue to that of all the traces to estimate c_e . The 3D output of c_e is named as the similarity volume.

2.3.4 Variance Attribute

The computation of variance attribute requires the gradient structure tensor that considers the waveforms and amplitude variation in three directions of a 3D seismic dataset. It uses a 3×3 covariance matrix. First, seismic amplitude gradient volume G_x , G_y and G_t is generated along x, y, and t direction, by the convolution of the seismic trace with the derivative products of the Gaussian filter $\frac{\delta G}{\delta x}, \frac{\delta G}{\delta y}, \frac{\delta G}{\delta t}$. Here, $G(x_j, y_j, t_j; \sigma) = \exp \left[-\frac{x_j^2 + y_j^2 + t_j^2}{\sigma^2} \right]$ is the Gaussian filter. The seismic full gradient tensor is given by $G = G_x + G_y + G_z$.

The gradient tensor matrix is derived by computing the outer product of the complete gradient tensor with itself.

$$T_{GS} = G \cdot G^T = \begin{bmatrix} G_x^2 & G_x G_y & G_x G_t \\ G_y G_x & G_y^2 & G_y G_t \\ G_t G_x & G_t G_y & G_t^2 \end{bmatrix} \quad (14)$$

The eigenvectors and eigenvalues of the tensor are obtained from

$$T_{GS} \mathbf{v}_m = \lambda_m \mathbf{v}_m \quad (15)$$

Where $\mathbf{v}_m$ are the three eigenvectors corresponding to three eigenvalues λ_m .

The first eigenvalue aligns with the direction of the most pronounced gradient, which runs perpendicular to the primary reflector. The variance is computed using the three eigenvalues using the following equation

$$V = \frac{\lambda_2 + \lambda_3}{\lambda_1 + \lambda_2 + \lambda_3} \quad (16)$$

2.3.5 Curvature Attribute

Curvatures can be useful to highlight structures showing dip variations, such as faults, channels etc. Different curvatures are then computed using the coefficients a, b, c, d and e. Curvature of a surface is estimated mathematically by fitting a quadratic surface $z(x, y)$ to an interpreted horizon, where:

$$z(x, y) = ax^2 + cxy + by^2 + dx + ey + f \quad (17)$$

The dip values along x and y direction can be expressed as $p = \frac{\delta z}{\delta x}$ and $q = \frac{\delta z}{\delta y}$

The first and second derivatives of $z(x, y)$ at $x = 0$ and $y = 0$ gives

$$a = \frac{1}{2} \frac{\delta p}{\delta x}, b = \frac{1}{2} \frac{\delta q}{\delta y}, \quad c = \frac{1}{2} \left(\frac{\delta q}{\delta x} + \frac{\delta p}{\delta y} \right), \quad d = p, \quad \text{and } e = q.$$

2.3.6 Spectral Decomposition

Spectral decomposition separates spectral information from the seismic traces. It is a popularly used seismic attribute, to map thin bed thicknesses and for enhanced geological interpretation of seismic data. In cases, where the lateral or vertical separation of two geological bodies falls below Rayleigh's criteria, which occurs when the peak-to-peak separation of their diffraction pattern is half the wavelet length (Widess, 1973) (Figure 2.5a), these two entities become indistinguishable in a seismic image. Spectral decomposition offers an alternative approach to visualize and quantify thin beds and to identify low-frequency shadow zones beneath gas reservoirs in seismic data. In conventional seismic processing, it is assumed that reflectivity from a geological formation is a random process, yielding a "white spectrum" as it consists of all the frequencies (Figure 2.5b). Thus, the amplitude spectra, which are convolutions of the source wavelet with the reflectivity series, have the same amplitude spectra as the source wavelet. However, Partyka et al. (1999) have shown that for a thin bed, the amplitude spectra estimated over a small window show a characteristic interference pattern, as shown in figure 2.5c.

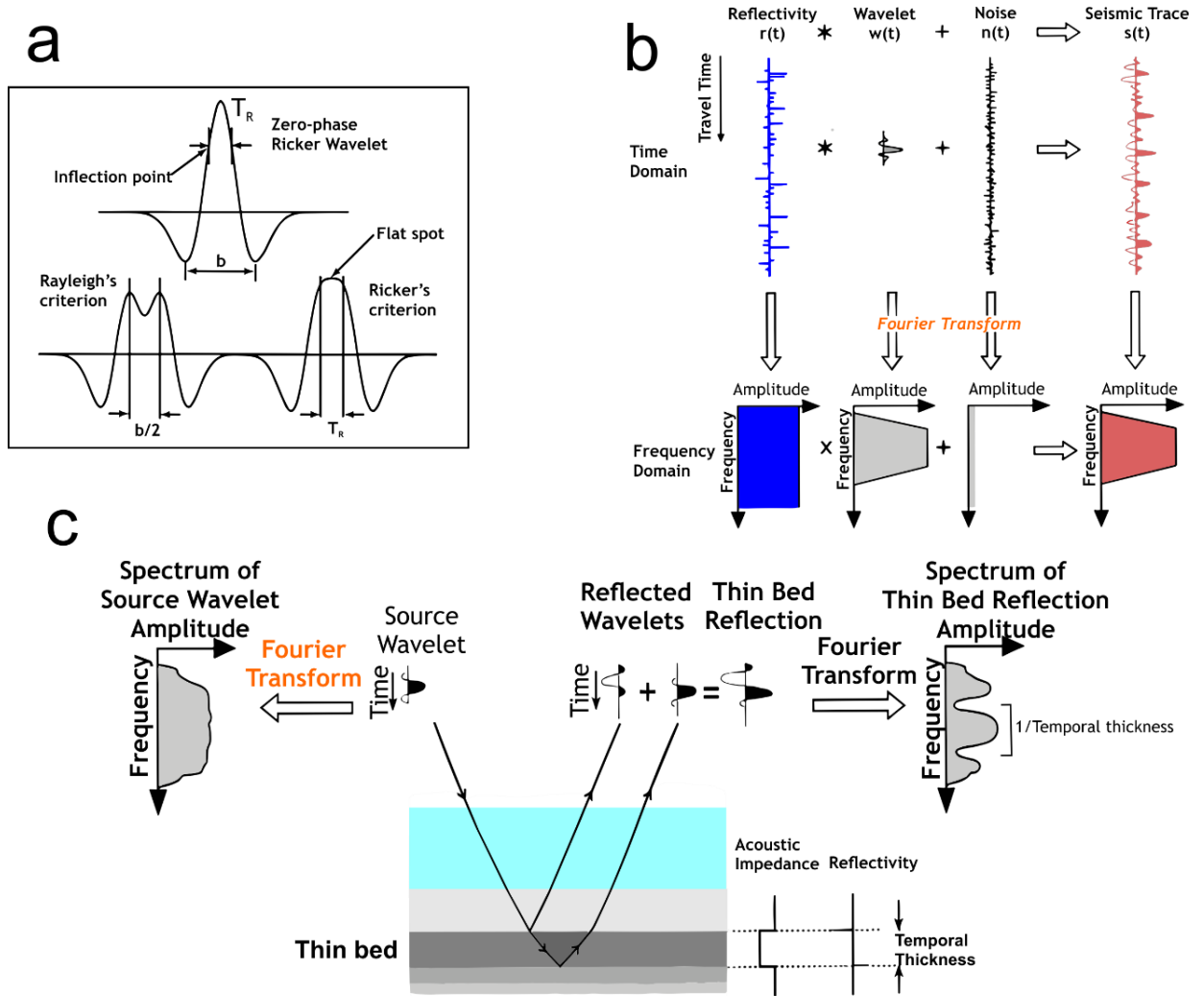


Figure 2.5. (a) Rayleigh resolution criteria. **(b)** Convolution concept. **(c)** Interference pattern for thin bed reflectivity and corresponding frequency spectrum. Figure adapted from Partyka et al. (1999).

In Figures 2.6a, and 2.6b there is a transition in the waveform as the temporal thickness decreases. The wavelet amplitude changes linearly with thickness when the thickness is below $\lambda/4$ (Widess, 1973). Further reduction in the thickness causes the amplitude to decrease due to, attenuation of low frequencies, with no change in the waveform. This transition in the amplitude signifies changes in the bed thickness and impedance of the two layers. Figure 2.6c shows amplitude-frequency spectra for 10 ms and 50 ms thin layers. Note that for a 10 ms blocky bed, the amplitude peaks at 50 Hz. The lateral amplitude variation for a seismic image filtered at 50 Hz will show maximum amplitude for 10 ms thin beds. It means that the reflectivity amplitude for a bed of specific thickness peaks only at specific frequencies. Consequently, filtering the data at those frequencies will highlight the features of that

particular bed thickness, which are otherwise indistinguishable in normal amplitude attributes. One can observe in this case that in contrast to the wide amplitude white spectrum of the seismic image, the thin bed displays a distinctive interference pattern spectrum, which depends upon layer temporal thickness. The wedge model best demonstrates the relation between amplitude and wavelet frequency with varying layer thickness.

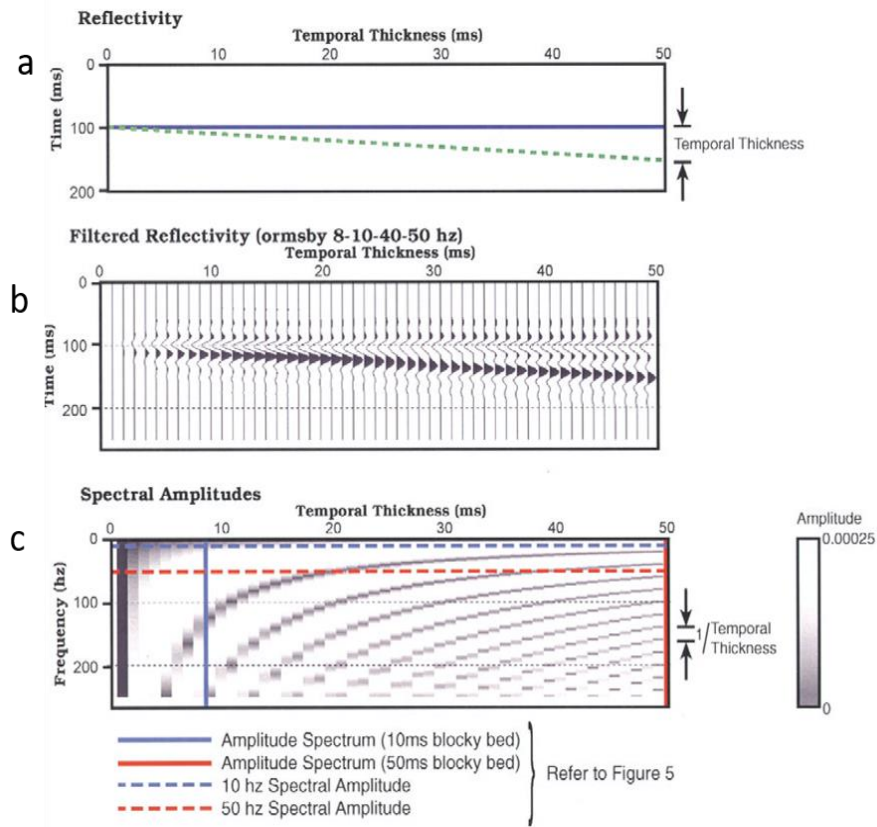


Figure 2.6. The wedge model depicts the change in wavelet and amplitude with bed thickness. Figure adapted from Partyka et al. (1999).

Continuous wavelet transforms and short-term Fourier transform (STFT) are the most commonly employed methods for spectral decomposition analysis. Both methods focus on extracting time variable frequency information from the seismic trace by using the cosine tapered window function. However, CWT integrates a window function of variable size as opposed to STFT, which uses a window function of constant length. The window length is inversely related to frequency resolution. The bandwidth of the frequency spectrum is always the same. However, in CWT, the scale varies with time. In this, good time resolution is evident at high frequencies with a trade-off of reduced frequency resolution. Conversely,

good frequency resolution is achieved at low frequencies, albeit with poor time resolution.

$$w(t) = \begin{cases} \frac{1}{2} [1 + \cos \pi \frac{|k\Delta t| - b + \xi}{\xi}] & \text{if } b - \xi \leq |k\Delta t| \leq b \\ 1 & \text{if } |k\Delta t| \leq b - \xi \\ 0 & \text{if } |k\Delta t| \geq b + \xi \end{cases} \quad (18)$$

STFT window function $w(t)$ with length b , centred at τ such that

$$w(t - \tau) = w(k\Delta t) \quad (19)$$

A continuous wavelet transform was preferred in this study because the STFT showed localized smearing effects resulting in poor vertical resolution that prevented clear identification of the discrete layers. This method convolves the seismic trace with a Morlet wavelet at the frequency f . The zero-phase Morlet wavelet resulted in an improved vertical resolution. The Morlet wavelet is formed by combining a complex exponential with a Gaussian window. CWT of the Gaussian window function, of width or scale σ , where $\sigma = 1/f$ is given by

$$w(t) = \frac{1}{\sqrt{\sigma}} \exp \left[-\frac{(k\Delta t)^2}{2\sigma^2} \right] \quad (20)$$

A commonly used method for visualizing the data from multiple frequency volumes involves using RGB colour blending. This technique combines three spectral components representing low, middle, and high frequencies. In this method, low frequencies are represented by the colour red, intermediate frequencies by green, and high frequencies by blue colour. The resulting RGB displays are interpreted in a qualitative manner. Specifically, red colour indicates stratigraphically thicker features; green suggests moderately thick features, and blue implies thinner features. White suggests an equally strong response at all frequencies, such as isolated bright reflections.

The 2D and 3D seismic data were loaded in the Kingdom Suite software for stratigraphic and fluid flow interpretation workflow. Additionally, PaleoScan was used to interpret the 3D seismic data. The seismic data interpretation workflow is schematically shown in Figure 2.7.

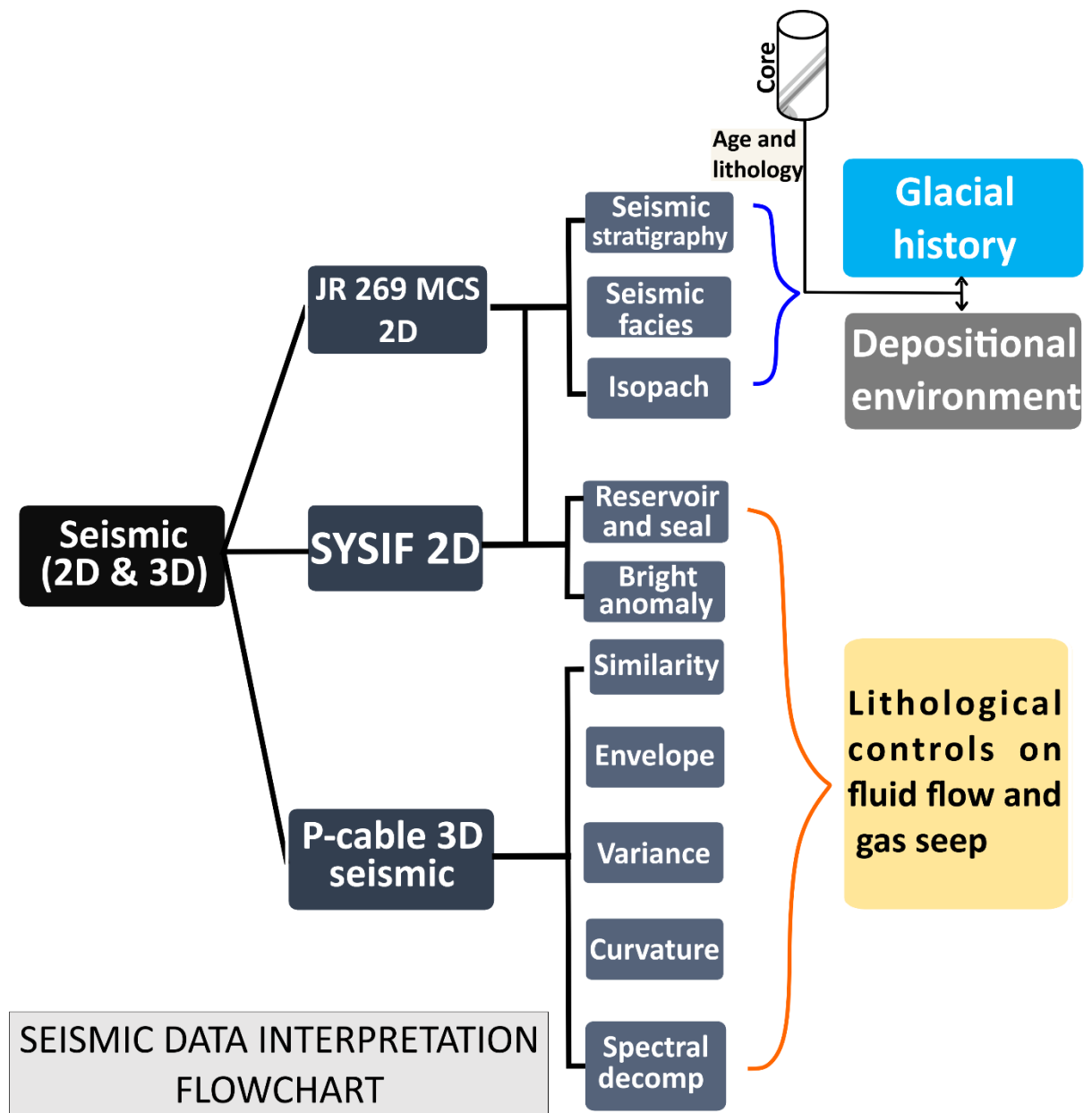


Figure 2.8. Flowchart showing the use of seismic data for stratigraphic, depositional environment, and fluid flow interpretation in this study.

Auxiliary Data

Multibeam Swath-bathymetry data: In this study, we use three distinct swath-bathymetry datasets that collectively provide a comprehensive understanding of seafloor characteristics.

The description of the datasets is provided in Table 2.4.

Bathymetry data source and year of acquisition	Sonar System	Frequency (kHz)	Beams	Depth Range (m)	Cell Size (m)	Geographic Area
JR211 Cruise (RRS James Clark Ross) (2008)	Kongsberg Simrad EM 120	12	191	50-3200	20	Upper Continental Slope western Svalbard
Norwegian Hydrographic Service (2003 and 2004)	Kongsberg Simrad EM1002	95	111	50-350	2	Shelf west of Prins Karl Forland
University of Tromsø (R/V Jan Mayen) (2006 & 2007)	Simrad EM300	30	135	300-2000	50	Slope west of Kongsfjorden, including Vestnesa ridge

2. 4 Modelling the behaviour of methane hydrate using Tough+Hydrate Code

Tough+Hydrate was employed to model the effects of relative sea level changes and fluctuations of bottom water temperature on methane hydrates. Moridis et al. (2014) developed the code. Model parameters governing the behaviour of methane hydrate-bearing sediments were constrained using seismic reflection, geophysical, and geological data. The physical and thermodynamic principles underlying the code, along with its governing equations, are discussed in Section 2.2.1.

2.4.1 Underlying physics and thermodynamics

The foundational physical and thermodynamic principles that underlie the behavior of sediments containing gas hydrates within the framework of the Tough+Hydrate code are discussed below. Moridis et al. (2014) discussed the mathematical principles utilized in the development of this code.

The code is a computational tool designed to solve the equations governing coupled mass and heat balances (as represented by Eq. 21). The code has been utilized to simulate non-isothermal gas discharge, phase change, as well as fluid and heat flow within subsurface environments. This is particularly relevant for hydrates found in porous or fractured marine sediments and in permafrost spanning various scales, such as laboratory-based experiments, and reservoir scale simulations where Darcy's law is applicable.

The code considers four primary components— methane, hydrate, water, and dissolved inhibitors. Inhibitors, like salts, play a role in destabilizing methane hydrates by modifying their thermodynamic equilibrium curve. There are four possible phases—ice, hydrate, gas, and liquid in which the four mass components are distributed. Furthermore, the code can integrate equilibrium and kinetic reactions to simulate the evolution of hydrates. The equilibrium reaction model was chosen for this study. Hydrate formation and dissociation, the transformation of different phases according to the pressure and temperature, and their resulting effect are considered in the code and the governing equations are mentioned below:

$$\frac{d}{dt} \int_{V_n} M^k dV = \int_{\Gamma_n} \vec{F}^k \cdot \hat{n} d\Gamma + \int_{V_n} q^k dV \quad (21)$$

In Eq. 21 V and V_n correspond to the volume of the model domain and subdomain, respectively. The surface area of the domain, and surface area of subdomain are denoted by A

and Γn , respectively. The mass accumulation for the component 'k' is denoted by M^k mentioned in Eq. 22.

Eq. 21 outlines the fundamental mass and energy balance equations for a subsurface system containing various phases and components, including water (w), methane (m), and water-soluble inhibitor (i).

$$M^k = \sum_{\beta=A,G,I} \phi S_{\beta} \rho_{\beta} X_{\beta}^{\kappa} \quad , \quad \kappa \equiv w, m, i \quad (22)$$

Additionally, $\mathbf{F}^k$ stands for the flux associated with the k^{th} component (shown in Eq. 23), while ' $\hat{n}$ ' represents the unit normal vector directed inward. The sink and source contribution terms are denoted by q^k , X_{β}^k denotes the fraction of mass of k^{th} component in the β phase and t denotes time (shown in Eq. 24).

$$\vec{F}^k = \sum_{\kappa=A,G} \vec{F}_{\beta}^{\kappa} \quad , \quad \kappa \equiv w, m, i \quad (23)$$

$$q^k = \sum_{\kappa=A,G} X_{\beta}^k q_{\beta} \quad , \quad \kappa \equiv w, m \quad (24)$$

Similarly, expressions for the heat accumulation is shown in Eq. 25 and mentioned below:

$$M^h = (1 - \phi) \rho_R C_R T + \sum_{\beta=A,G,H,I} \phi S_{\beta} \rho_{\beta} U_{\beta} + Q_{diss} \quad (25)$$

$$Q_{diss} = \begin{cases} \Delta (\phi \rho_H S_H \Delta U_H) & \text{for equilibrium dissociation} \\ 0 & \text{for kinetic dissociation} \end{cases}$$

Eq. 25 considers the density (ρ), heat capacity (C_R), and specific internal energy (U). Q_{diss} is the amount of the heat required to dissociate hydrate. The code incorporates a wide range of heat exchange mechanisms, including heat conduction, advection and convection, radiant heat transfer, reactions involving hydrate, latent heat due to phase transitions, gas dissolution, and dissolution of gas and inhibitors. It also takes into account both advection and molecular diffusion mechanisms for water-soluble gases and inhibitors. Additionally, the code addresses

heat flux originating from conduction and advection within different phases, as well as radiative heat transfer. It takes into consideration the impact of inhibitors (such as dissolved NaCl) on hydrate stability. The code can account for changes in intrinsic permeability associated with the changes in solid-phase saturation, such as hydrate. The methane hydrate phase relationship with the thermodynamic constraints (Pressure and Temperature) is shown in figure 2.8.

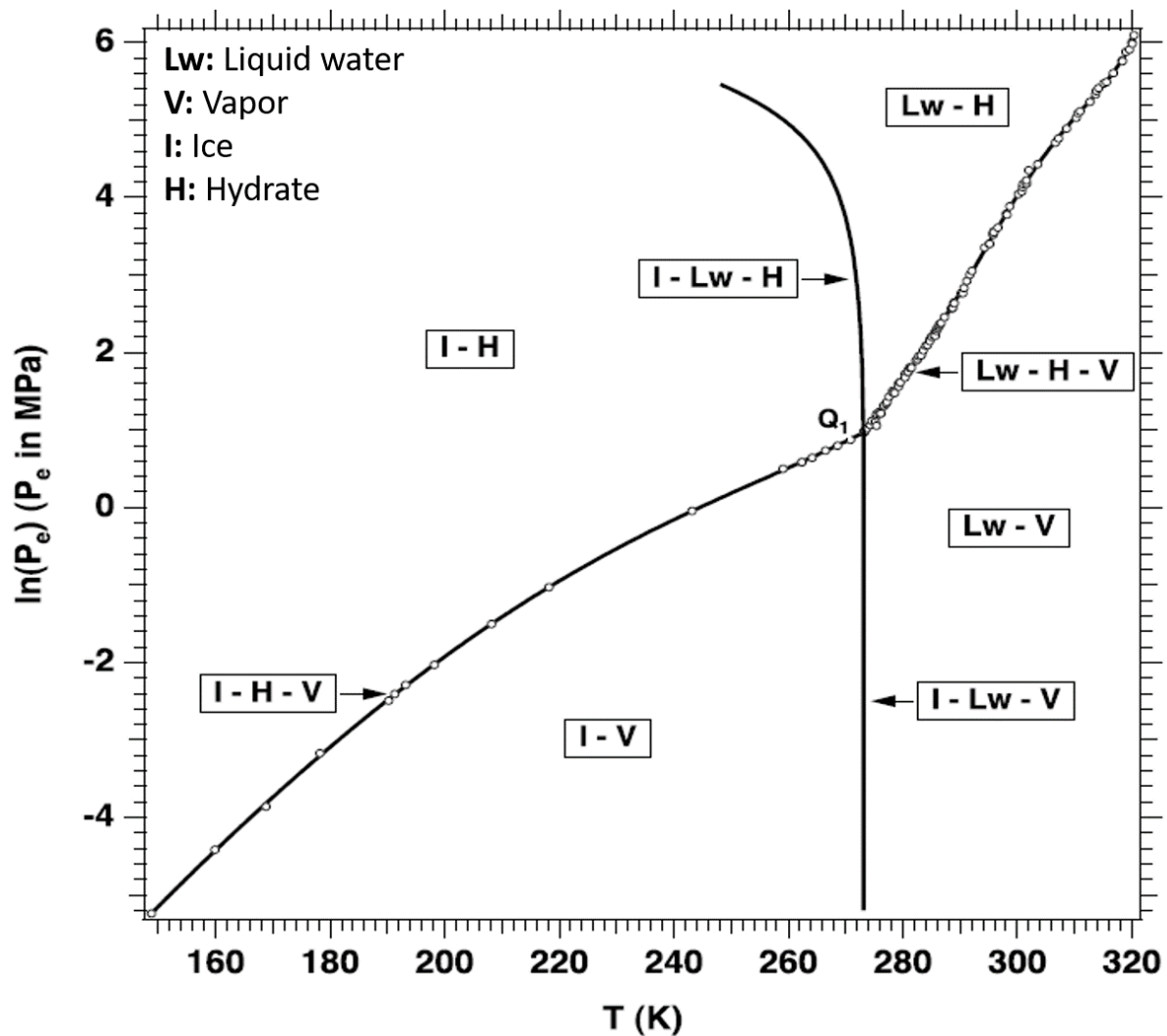


Figure 2.8. Pressure-temperature equilibrium relationship in the phase diagram of the water-CH₄-hydrate system in T+H. Figure adapted from Moridis et al. (2014).

In this study, I have employed the following assumptions and considerations:

1. Within the model domain, Darcy's law holds true under the imposed conditions on the system. The flow behaviour in the pore space is considered to be laminar (Charbeneau, 2006).
2. Dispersion is minimal during the transportation of dissolved gases and inhibitors compared to advection. This consideration makes the simulation computationally optimized.
3. The guest molecule in the hydrate structures is methane. The compositions of gas samples collected in the water column at the seabed from the seepage site suggest 99.9 % methane (James et al., 2011).
4. The medium is stationary while freezing (no soil heaving).
5. The saturation of the inhibitors does not alter the thermo-physical properties of the aqueous phase. In this study, NaCl is the non-volatile inhibitor.
6. Salt precipitation is not considered.
7. The pressure-dependent equations that account for hydrate dynamics are accurate for pressures up to 100 MPa, which is much greater compared to the maximum pressure at which natural hydrates form. Since the maximum water depth in our model is 600 m, the pressure is well within the limit of these models.

The hydrate equation of state governs the distribution of components across potential phases and thermodynamic state. Tables 2.5 and 2.6 shows the primary variables used to define the equation of state. While the primary variables remain fixed throughout the simulation, their values can be adjusted to accommodate the appearance and disappearance of phases and components.

The integral finite difference method was used to discretize the continuum equations (as represented in Eq. 24) in space (Edwards, 1972; Narasimhan & Witherspoon, 1976). Time discretization was achieved using a first-order finite difference method. Fluxes were managed in a "fully implicit" way (Peaceman, 1977). The linearization of the interconnected non-linear equations was achieved by the Newton-Raphson method. Conditioning the matrix was performed by following the Jacobian method. The final solution is obtained by the conjugate gradients method in which a pre-conditioned matrix was considered as an input (obtained from the Jacobian method) (Moridis & Pruess, 1995).

$$R_n^{k,k+1} = M_n^{k,k+1} - M_n^{k,k} - \frac{\Delta t}{V_n} \left(\sum_m A_{nm} F_{nm}^{k,k+1} + V_n q_n^{k,k+1} \right) = 0 \quad (26)$$

Here $R_n^{k,k+1}$ are the residuals. For every grid V_n (volume element), the number of equations are N_k and if a system has a grid numbers N_E , Eq. 26 represents $N_k \times N_E$ couple equations.

Table 2.5. Primary variable and the state identifiers in methane hydrate simulations in equilibrium for pure water (Adapted from Moridis et al. (2014))

Phase	State Identifier	Primary Variable 1	Primary Variable 2	Primary Variable 3
1-Phase: G	Gas	P_{Gas}	Y_{m_G}	T
1-Phase: A	Aqu	P	X_{m_A}	T
2-Phase: A+G	AqG	P_{Gas}	S_{aqu}	T
2-Phase: I+G	IcG	P_{Gas}	S_{ice}	T
2-Phase: H+G	GsH	P_{Gas}	S_{gas}	T
2-Phase: A+H	AqH	P	S_{aqu}	T
2-Phase: A+I	Aqi	P	S_{aqu}	X_{m_A}
2-Phase: I+H	IcH	P	S_{ice}	T
3-Phase: A+H+G	AGH	S_{gas}	S_{aqu}	T
3-Phase: A+I+G	AIG	P_{gas}	S_{aqu}	S_{gas}
3-Phase: A+I+H	AIH	P	S_{aqu}	S_{ice}
3-Phase: I+H+G	IGH	S_{gas}	S_{ice}	T
Quadruple Point: I+H+A+G	QuP	S_{gas}	S_{aqu}	S_{ice}

Abbreviations: **P**- pressure (Pa), **P_{gas}** - gas pressure (Pa), **T**- temperature (°C), **X_{m_A}**- mass fraction of CH₄ dissolved in the aqueous phase, **Y_{m_G}** - mass fraction of CH₄ dissolved in the gas phase, **S_{aqu}** - liquid saturation, **S_{gas}** - gas saturation, **S_{ice}** - ice saturation, **X_{i_A}** - mass fraction of inhibitor dissolved in the aqueous phase.

* For inhibitor: **X_{i_A}** becomes the 3rd primary variable, and the 3rd primary variable (as listed in this table) becomes the 4th primary variable State identifier also known as Equation of state (EOS) determine the thermodynamic state the distribution of the mass components among the different phases. A, Aq - Aqueous, H- Hydrate, G, Gs - Gas, I, Ic - Ice. Example: AGH: Aqueous-Gas-Hydrate, AqH: Aqueous-Hydrate.

Table 2.6. Primary variable and the state identifiers in methane hydrate simulations in non-equilibrium (Kinetic) for pure water (Adapted from Moridis et al. (2014))

Phase	State Identifier	Primary Variable 1	Primary Variable 2	Primary Variable 3	Primary Variable 4
1-Phase: G	Gas	P_{Gas}	Y_{m_G}	S_{hyd}	T
1-Phase: A	Aqu	P	X_{m_A}	S_{hyd}	T
2-Phase: A+G	AqG	P_{Gas}	S_{aqu}	S_{hyd}	T
2-Phase: I+G	IcG	P_{Gas}	S_{ice}	S_{hyd}	T
2-Phase: H+G	GsH	P_{Gas}	S_{gas}	S_{ice}	T
2-Phase: A+H	AqH	P	S_{aqu}	X_{m_A}	T
2-Phase: A+I	AqI	P	S_{aqu}	X_{m_A}	X_{m_A}
2-Phase: I+H	IcH	P	S_{ice}	S_{gas}	T
3-Phase: A+H+G	AGH	S_{gas}	S_{aqu}	S_{gas}	T
3-Phase: A+I+G	AIG	P_{gas}	S_{aqu}	S_{hyd}	S_{gas}
3-Phase: A+I+H	AIH	P	S_{aqu}	S_{ice}	S_{ice}
3-Phase: I+H+G	IGH	S_{gas}	S_{ice}	S_{ice}	T
Quadruple Point: I+H+A+G	QuP	S_{gas}	S_{aqu}	S_{gas}	S_{ice}

Abbreviations: P - pressure (Pa), P_{gas} - gas pressure (Pa), T - temperature ($^{\circ}C$), X_{m_A} - mass fraction of CH_4 dissolved in the aqueous phase, Y_{m_G} - mass fraction of CH_4 dissolved in the gas phase, S_{aqu} - liquid saturation, S_{gas} - gas saturation, S_{ice} - ice saturation, X_{i_A} - mass fraction of inhibitor dissolved in the aqueous phase.* For inhibitor: X_{i_A} becomes the 4th primary variable, and the 4th primary variable (as listed in this table) becomes the 5th primary variable. State identifier also known as Equation of state (EOS) determine the thermodynamic state the distribution of the mass components among the different phases. A, Aq - Aqueous, H- Hydrate, G, Gs - Gas, I, Ic - Ice.
Example: AGH: Aqueous-Gas-Hydrate, AqH: Aqueous-Hydrate.

2.4.2 Changes in Porosity and intrinsic permeability in porous medium

The formation of a new solid phase in a porous medium reduces porosity and permeability. A simple model approximates porous media as capillary tube bundles, yielding a simple power-law relationship between permeability k and porosity ϕ . Eq. 27 shows a mathematical model that conceptualizes the change in permeability due to the formation of a new solid phase, other than the sediment grains, which leads to changes in porosity. k_0 and ϕ_0 refer to the permeability and porosity of a reference medium at reference P , T conditions.

$$\frac{k}{k_0} = F_{\phi S} = \left(\frac{\phi}{\phi_0} \right)^n \quad (27)$$

$F_{\phi S}$ is an adjustment scalar for the permeability and n range from 2 to 3. Porous sediments have pore spaces that show convergent and divergent geometry, made up of a sequence of narrow pore throats and wide pore spaces (Figure 2.9a). This can be visualized as a series of interconnected tubes representing the small-radii pore throats and large-radii pore bodies (Figure 2.9b). In such a model, even a minor amount of solid precipitation can obstruct narrow pore throats, which act as bottlenecks for fluid flow. To mitigate this, the concept of critical porosity was introduced.

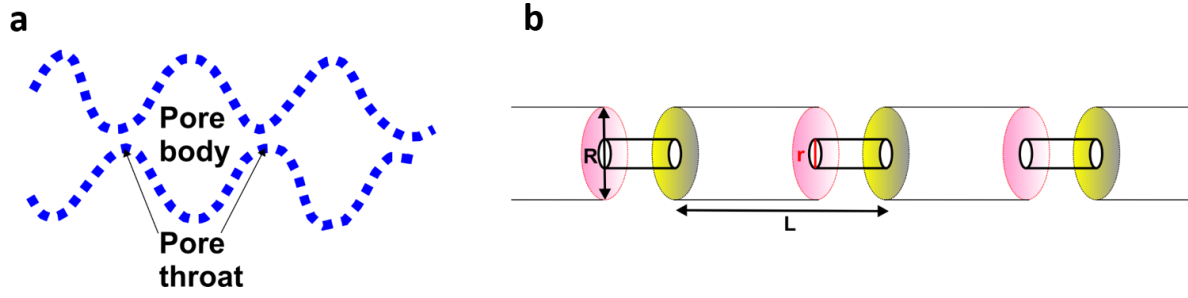


Figure 2.9. (a) The schematic illustration of pore space structure shows constricted pore throats and expanded pore bodies. **(b)** Pore channels are connected as tubes in a series.

Critical porosity (ϕ_c) refers to the threshold porosity at which a porous medium becomes impermeable due to pore throat clogging as new solid phases crystallize, even though the sediment column may still have substantial porosity (ϕ). Below this critical porosity, the permeability of the medium is maintained at a non-zero value, allowing fluid flow.

The mathematical formulation describing the permeability reduction with respect to ϕ and ϕ_c is mentioned in Eq. 28.

$$\frac{k}{k_o} = F_{\phi S} = \left(\frac{\phi - \phi_c}{\phi_o - \phi_c} \right)^n \quad (28)$$

Relative permeability is a critical concept in modelling fluid flow in porous media, especially in multi-phase systems. It quantifies the ability of a fluid phase to flow through a porous medium relative to the total permeability, and it varies with the saturation of the respective phase (Figure 2.10). It is equal to one when there is only one phase in the system. Saturation represents the fraction of pore volume filled by a specific phase. When there is the coexistence of multiple phases, the relative permeability values can be less than one, indicating that not all of the pore space is available for the flow of that phase. In hydrate reservoir systems, the formation of hydrate in the pores represents a new solid phase that occupies space and alters the relative permeability of the remaining fluid phases (gas and water).

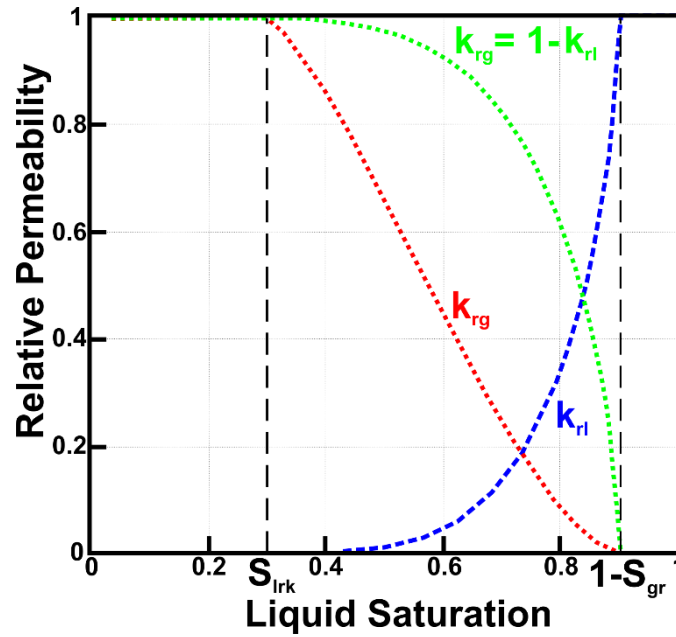


Figure.2.10. Relative permeabilities of liquid and gas phases derived from van Genuchten model (Finsterle,1999).

In the T+H code, two relative permeability models are proposed. The original porous medium (OPM) model calculates the relative permeability of water based solely on water saturation, regardless of the occurrence of gas hydrates. This makes it simple and self-contained, requiring no additional permeability information.

The evolving porous medium (EPM) model treats hydrate as a solid component of the porous medium, allowing intrinsic permeability and porosity to change over time. The EPM model updates relative permeability as new hydrates form, overcoming the limitations of the OPM model. A more sophisticated version of the EPM model considers the critical porosity concept. A fraction of the pore space becomes inaccessible to fluids when a new solid phase (hydrate) forms and blocks the pore throats. This version of the EPM model was used in the present work.

Chapter 3

Reassessment of hydrate destabilization mechanisms offshore west Svalbard confirms link to recent ocean warming

3.1. Introduction

The effects of ocean warming are clearly visible across the Arctic (Hassol, 2004), as evidenced by the disappearance of sea ice (Comiso et al., 2008; Piechura & Walczowski, 2009) and enhanced glacier melting (Osterkamp, 2005; Nagornov et al., 2006; Anthony et al., 2012). Dissociation of methane hydrates has also been linked with past climate warming (Nisbet, 1990; Dickens, 2003; Kennett et al., 2003), and there is concern about the stability of methane hydrate deposits trapped in Arctic marine sediments (Kretschmer et al., 2015). Hydrate at the feather edge of its stability zone, where the base of this zone cropping out on the upper continental slope, are particularly vulnerable to thermal destabilization. The marine sediments of the Fram Strait, the Arctic-Atlantic gateway located in the western Svalbard region, host widespread methane hydrates (Figure 3.1a, Sarkar et al., 2012; Dumke et al., 2016; Minshull et al., 2020). A marine geophysical expedition in 2008 discovered plumes of methane bubbles emanating from the seafloor close to and further up-slope of the present-day upper limit of methane hydrate stability on the eastern margin of the Fram Strait (Westbrook et al., 2009). Here warm and saline Atlantic water brings heat into the Arctic Ocean through the Fram Strait via the West Spitsbergen Current. This Atlantic water is inferred to have warmed the bottom water temperature (BWT) from 2° to 3° C over the past four decades, resulting in the migration of the feather edge of methane hydrate stability from 370 m to 410 m depth (Westbrook et al., 2009). Westbrook et al. (2009) proposed that these methane bubble plumes indicate thermal degradation of the hydrates at the landward limit of hydrate stability.

Several subsequent marine surveys consistently found methane seeps (Sahling et al., 2014; Mau et al., 2017; Veloso-Alarcon et al., 2019) between 370 m and 410 m water depths on the upper slope (Figure 3.1b). The gas flare density and methane release rates are reduced during the winter when the temperatures are substantially lower than in summer (Veloso-Alarcon et al., 2019; Ferré et al., 2020). This suggests that seasonal fluctuations in BWT can lead to repeated episodes of hydrate formation and dissociation (Berndt et al., 2014; Veloso-Alarcon et al., 2019; Ferré et al., 2020). However, the link between methane seepage and anthropogenic ocean warming has been questioned because the presence of seep carbonates in this area suggests that seepage has been active for > 8,000 yr (Berndt et al., 2014). Numerical simulation of the response of gas hydrates to BWT changes was previously carried out considering past and future ocean temperature variations (Reagan et al., 2011; Thatcher et al., 2013; Marín-Moreno et al., 2013; Marín-Moreno et al., 2015). Results from these studies

indicated that for hydrate shallower than a few metres beneath the seabed, bottom water warming since the late 1970s has produced gas from hydrate dissociation. However, these studies did not account for the effect of relative sea level (RSL) change on hydrate stability.

Hydrate dissociation off west Svalbard was first geochemically confirmed by Wallmann et al. (2018). Pore water samples collected from sediments during a “MeBo” seafloor drilling experiment conducted in 2016 at a water depth of 391 m (Figure 3.1b) showed a decline in chloride concentration at 15–22 m below the seabed, indicating freshening caused by hydrate dissociation. Based on the University of Tromsø (UiT) RSL model (Patton et al., 2017), Wallmann et al. (2018) inferred that approximately 8000 years ago, the glacial isostatic rebound rate exceeded the rate of eustatic sea-level rise. The resulting RSL fall and consequent drop in hydrostatic pressure led to hydrate destabilization. In their model, hydrate dissociation was completed 1000 years before the present (BP, where ‘Present’ is considered as 2016). Hydrate could re-form in the model during colder BWT conditions in the late 1970s (Ferré et al., 2012; Wallmann et al., 2018). Wallmann et al. (2018) concluded that a 1 °C linear rise in BWT over 1980–2016 would be insufficient to result in hydrate dissociation. However, in their modelling they did not account for the decadal-scale temperature fluctuations observed in the measured records. Such variations can impact hydrate dynamics and methane flow at the seabed, and therefore, should be included in a realistic assessment of whether recent bottom water warming could have caused the observed hydrate dissociation.

Glacial retreat across Svalbard took place about 13000 years BP (Landvik et al., 2008), causing rapid seafloor uplift and a prominent relative sea-level (RSL) drop over 8–12 ka along the western Svalbard coast (Forman et al., 2004). RSL change at the MeBo drill site is not constrained by data and must be estimated by geodynamic modelling. Comparison of observed and predicted RSL change based on various deglacial scenarios has revealed the inadequacy of some models in reproducing observations of RSL change throughout the Barents Sea region (Auriac et al., 2016). The ICE-6G_C scenario correctly predicts a fall in RSL over 8–12 ka in west Svalbard (Auriac et al., 2016) but, in contrast to the UiT model employed by Wallmann et al. (2018), it predicts RSL rise at the MeBo site during the past 8000 years (see Section 3.3.3). Ultimately it is not possible to confirm which scenario is correct. Therefore, we consider both the UiT and ICE-6G_C RSL scenarios to examine the role of pressure in hydrate dynamics west of Svalbard.

Here we address the limitations of previous studies and place further constraints on hydrate dissociation processes through coupled heat and mass transfer numerical simulations of the impact of 11000 years of ocean temperature and RSL change on hydrate located at the upper continental slope off west Svalbard. We have attempted to reproduce the present-day chloride anomaly observed in the sediment core collected at 391 m water depth (Wallmann et al., 2018) by varying the initial hydrate saturation and considering realistic RSL and BWT reconstructions, including decadal and seasonal temperature fluctuations. Based on this modelling, we conclude that rising BWT plays a dominant role in hydrate destabilization, potentially overprinting the effect of any ongoing RSL rise or fall.

3.2. The West Svalbard Margin

The west Svalbard continental margin has experienced repeated Pliocene–Pleistocene glaciations and deglaciations that affected sea level, sedimentation, and erosion patterns (Landvik et al., 2008). Fast-flowing ice streams carved out the Kongsfjorden, Isfjorden, and Bellsund cross-shelf troughs (KT, IT, and BT, Figure 3.1a). The present-day shelf break represents the maximum extent reached by the ice sheet and grounded ice streams (Solheim et al., 1996; Knies et al., 2009). Our study area (Figure 3.1b) is located in the inter-fan region between the Kongsfjorden and the Isfjorden cross-shelf troughs (Sarkar et al., 2011). Using multi-channel reflection seismic data, Sarkar et al. (2012) identified a gas hydrate bottom-simulating reflector (BSR) (Figures 3.1a and 3.1b) beneath the upper continental slope. The BSR represents the boundary between gas hydrates above and free gas below (Shipley et al., 1979). There is no evidence of a BSR in water depths shallower than ~ 600 m (Sarkar et al., 2012); instead, there is seismic evidence of gas accumulations (400–700 m water depths), as evidenced by low-velocity pockets and negative-polarity bright spots within the top 75 m sediment column below the seabed (Chabert et al., 2011). In addition, regions of anomalously high electrical resistivity suggest the presence of hydrate within its stability field (Goswami et al., 2016).

The Pliocene–Pleistocene sediments on the shelf and upper continental slope are predominantly glaciogenic in origin. The marine hemipelagic sediments are interbedded with glacial debris flow units at the upper slope (Sarkar et al., 2011, 2012). The MeBo drilling experiment drilled a mixture of hemipelagic sediments and poorly-sorted glacial debris consisting of a broad range of grain sizes, from clay to sand with gravel to pebble-sized clasts in variable amounts (Wallmann et al., 2018). During the deglaciations, meltwater discharge on the upper continental slope resulted in turbidity currents cascading down the slope,

depositing sand and silt-rich sediments (Vorren et al., 1998). The lithological heterogeneity strongly influences fluid flow below the methane seepage sites, for example, the flow of gas is impeded by the glaciogenic debris. Lateral fluid migration occurs along the sand and silt-rich permeable sediments, and near-vertical flow occurs through fractures within poorly stratified, low-permeability glaciogenic sediments (Haacke & Westbrook, 2006; Haacke et al., 2009; Sarkar et al., 2012; Thatcher et al., 2013).

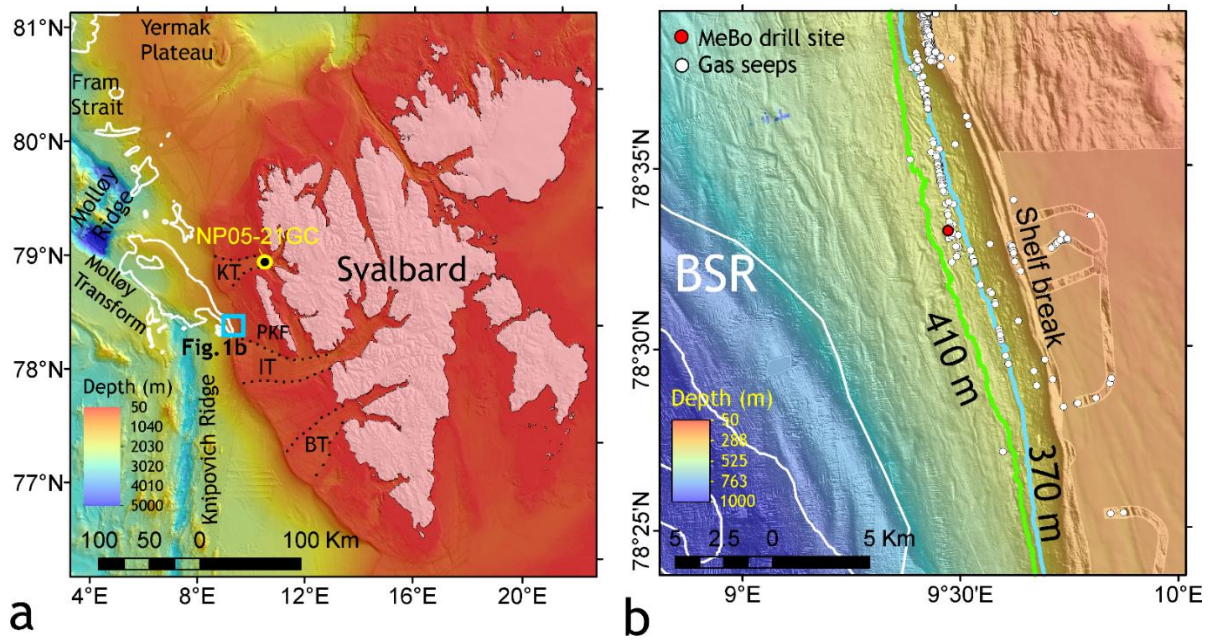


Figure 3.1. West Svalbard continental margin with bathymetry compiled from various sources, such as Norwegian Hydrographic Survey, swath-bathymetry from JR211 cruise (Sarkar et al., 2012), and International Bathymetric Chart of the Arctic Ocean (Jakobsson et al., 2020). **(a)** The Kongsfjorden (KT), Isfjorden (IT), and Bellsund (BT) cross-shelf troughs are the palaeo-ice stream pathways dissecting the shelf. The study area is located between the KT and IT and west of Prins Karl Forland (PKF). The solid white line represents the extent of the seismic bottom simulating reflector (BSR) (Sarkar et al., 2012; Dumke et al., 2016). The hydrate province is bounded by the Molløy Transform Fault that links the Knipovich Ridge and the Molløy Ridge. The sediment core NP05-21GC (yellow dot) is located at 327 m water depth on the Kongsfjorden shelf west of Svalbard (Rasmussen et al., 2014). **(b)** Gas seeps on the upper continental slope are located between 370 and 410 m water depths (Sarkar et al., 2012), which correspond to the theoretical limit of methane hydrate stability at 2 °C and 3 °C bottom water temperatures, respectively. The MeBo drill site is located at 391 m water depth.

3.3. Data and Method

3.3.1. Modelling Approach

We performed 1D numerical simulations of vertical fluid and heat flow coupled with hydrate formation and dissociation using the TOUGH+Hydrate code v1.5 (Moridis et al., 2014). We used an equilibrium hydrate formation and dissociation model with three mass components, i.e., methane, water, and salt (NaCl), partitioned into four possible phases (gas hydrate, ice, gas, and liquid). The model considers conductive and convective heat transport, pressure-driven transport of water and methane gas (Darcy's flow), advective and diffusive transport of dissolved methane and salt in water, and latent heat of hydrate formation and dissociation.

Model initialization at 11200 yr BP was done assuming a hydrostatic pressure gradient, constant heat flow at the base of the model of $10 \times 10^{-2} \text{ W m}^{-2}$ and water saturated sediment thermal conductivity of $2 \text{ W m}^{-1} \text{ K}^{-1}$ (Riedel et al., 2018), generating a geothermal gradient of 50° C/km , 3.5 wt % pore water salinity and appropriate BWT and pressure in the top cell of the model (Figure 3.2a). We used the BWT and pressure values from the reconstructed curves discussed in sections 3.2.3 and 3.3.3. Model initialization is done based on various BWT and RSL scenarios (Supplementary Figure 3.1). The model has a constant cell thickness of 0.5 m from 1 mm to a maximum depth of 750 m and assumes a present-day water depth of 391 m. The topmost cell where the top boundary conditions are applied has a thickness of 1 mm and represents the ocean bottom.

The base of the sulfate reduction zone at the MeBo site is 5–7 m (Thatcher et al., 2013; Marín-Moreno et al., 2013; Marín-Moreno et al., 2015 and Wallmann et al., 2018). To represent this, we assume the top 7 m of the sediment column is hydrate-free in the starting model, but we do not consider that this zone was hydrate-free throughout the simulation. The sulfate reduction zone comprises marine anoxic sediments within which there is a decline in sulfate concentration, and dissolved methane is consumed by anaerobic oxidation (Boetius & Wenzhöfer, 2013; Borowski et al., 1996). We have not modeled this process. The Holocene sedimentation rate is negligible, only 4 cm kyr^{-1} (Panieri et al., 2016); therefore, sedimentation is not included in the modelling.

The initial hydrate layer in our model was placed up to the base of methane hydrate stability zone (MHSZ) (Figure 3.2b). We considered a range of initial hydrate saturations (e.g., 9–60 %) for different scenarios. Since no hydrate was recovered during MeBo drilling experiment, this initial hydrate should completely melt. We assume an initial value for gas saturation below the MHSZ of 2–4 % based on geophysical evidence of gas in the inter-fan region in a

depth range of 480–1285 m on the upper slope (Chabert et al., 2011). Goswami et al. (2016) inferred a much larger gas saturation ($> 10\%$) around the landward edge of the GHSZ based on high resistivity. However, the high resistivity values could also be influenced by the presence of authigenic carbonates. We used 2–4 % gas below the hydrate stability zone based on earlier models (Marín-Moreno et al., 2013 and Marín-Moreno et al., 2015). The gas was pure methane (James et al., 2011).

We assumed an initial hydrate-free uniform intrinsic (absolute) permeability of 10^{-13} m^2 , which is 2–4 orders of magnitude higher than the permeability of hemipelagic glaciomarine sediments (Table S3.1). We adopted the value from Thatcher et al. (2013), who argued that shallow glacial sediments could rarely sustain a low intrinsic permeability during rapid gas release from hydrate dissociation because the pore pressure will surpass the lithostatic load within a few years of the onset of hydrate dissociation (Thatcher et al., 2013), subsequently leading to the development of fractures that enhance permeability. Hence, this high value of intrinsic permeability simulates the effects of fracture permeability. In the scenario they investigated, Thatcher et al. (2013) found that gas appeared at the seabed in response to recent decadal-scale warming, consistent with observations, while if lower permeabilities were used, gas release into the ocean was delayed. In the presence of hydrate in the pore space, there is a reduction in the intrinsic permeability, which is accounted for in the model by using the Evolving Porous Medium model (Moridis et al., 2014; Table S3.1). The irreducible gas saturation and water saturation are 2 and 12 %, respectively, consistent with laboratory measurements (Liu & Flemings, 2007). We note that, to our knowledge, there is no published laboratory data from sediment cores in our study area to better constrain the irreducible gas saturation. We also incorporated multiphase molecular diffusion. Methane flux by molecular diffusion is slow compared to the free gas flow and advection mechanisms, but the simulation time is long; therefore, transport by molecular diffusion can be important.

Changes in BWT and RSL were applied to the model at yearly intervals by changing the temperature and pressure in the top cell. Our model includes the instantaneous response of the sediment column to RSL-induced pressure changes and addresses the process of hydrate dissociation and reformation; for example, the model accounts for the elevated pore-pressure and salinity drop that occur during hydrate dissociation, which can further inhibit melting, as suggested by Liu and Flemings (2009). A summary of the properties used in the models is provided in Table S3.1.

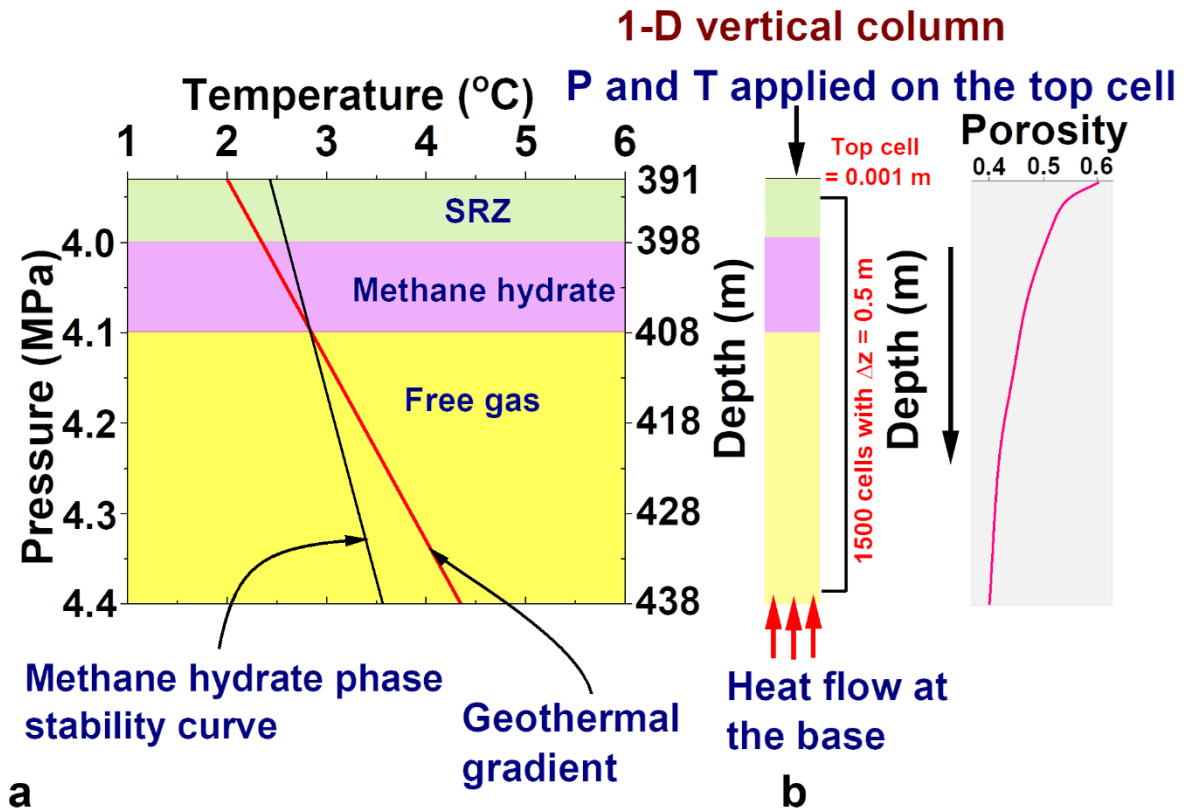


Figure 3.2. Schematic representation of the model domain at a water depth of 391 m. **(a)** Methane hydrate stability shown on the Temperature vs. Pressure (Depth) plot. The thickness of the methane hydrate stability is calculated using the geothermal gradient (red line) and the methane hydrate phase stability curve (black line). SRZ denotes the sulfate reduction zone (green area). Methane hydrate-bearing sediments are represented by the purple, and the free gas zone by the yellow colour. **(b)** Schematic representation of a 1-D vertical sediment column with the boundary conditions. Time-dependent pressure and temperature values are applied as boundary conditions on the top cell while a constant heat source is maintained at the bottom cell. The porosity depth profile is shown by the pink curve.

3.3.2. Sea bed temperature history

Rasmussen et al. (2014) reconstructed the BWT west of Svalbard over the past 11200 years based on two proxies – (1) benthic foraminiferal census counts (BWT_{TF}) and (2) benthic foraminifera $\delta_{18}O$ values (BWT_C) from the sediment core NP05-21GC taken at 327 m water depth on the Kongsfjorden shelf west of Svalbard. NP05-21GC is closest sediment core to the methane seepage site on the upper continental slope (Figure 3.1a). It is currently bathed by warm Atlantic water (Sarkar et al., 2015), and hence we assume that the reconstructed BWT values well represent past temperature variations of the Atlantic water mass.

The BWT_C temperatures, based on $\delta_{18}O$ values in *Cibicides lobatulus*, are influenced by summer conditions since it calcifies during the summer. Temperatures derived using this approach may therefore be biased warm. The BWT_C curve shows incursions of warmer Atlantic water during the early Holocene (11500–6000 years BP) and a subsequent temperature dip during 5000–2000 years BP (Figure 3.3a). BWT_{TF} (also shown in Figure 3.3a) shows a smoother curve compared to BWT_C because its reconstruction relies on annual faunal counts (> 200 specimens) (Rasmussen et al., 2014). Since we do not know which is the most representative temperature curve, we considered both the temperature series.

In addition, Marín-Moreno et al. (2013) compiled measured and estimated temperature records at 400 m water depth on the West Svalbard continental margin for the past two millennia. For the years A.D. 1–1900, they used 50-year-running summer temperature means at 50 m water depth, derived from planktic foraminifera proxy (Spielhagen et al., 2011), and applied appropriate scaling to predict the temperature at 400 m depth based on the ocean/sea-ice model NEMO at $1/12^\circ$ resolution (ORCA12) (Madec, 2012). We use the resulting BWT_M curve of Marín-Moreno et al. (2013) (Figure 3.3b) in a specific modelling scenario (refer to section 3.4.3).

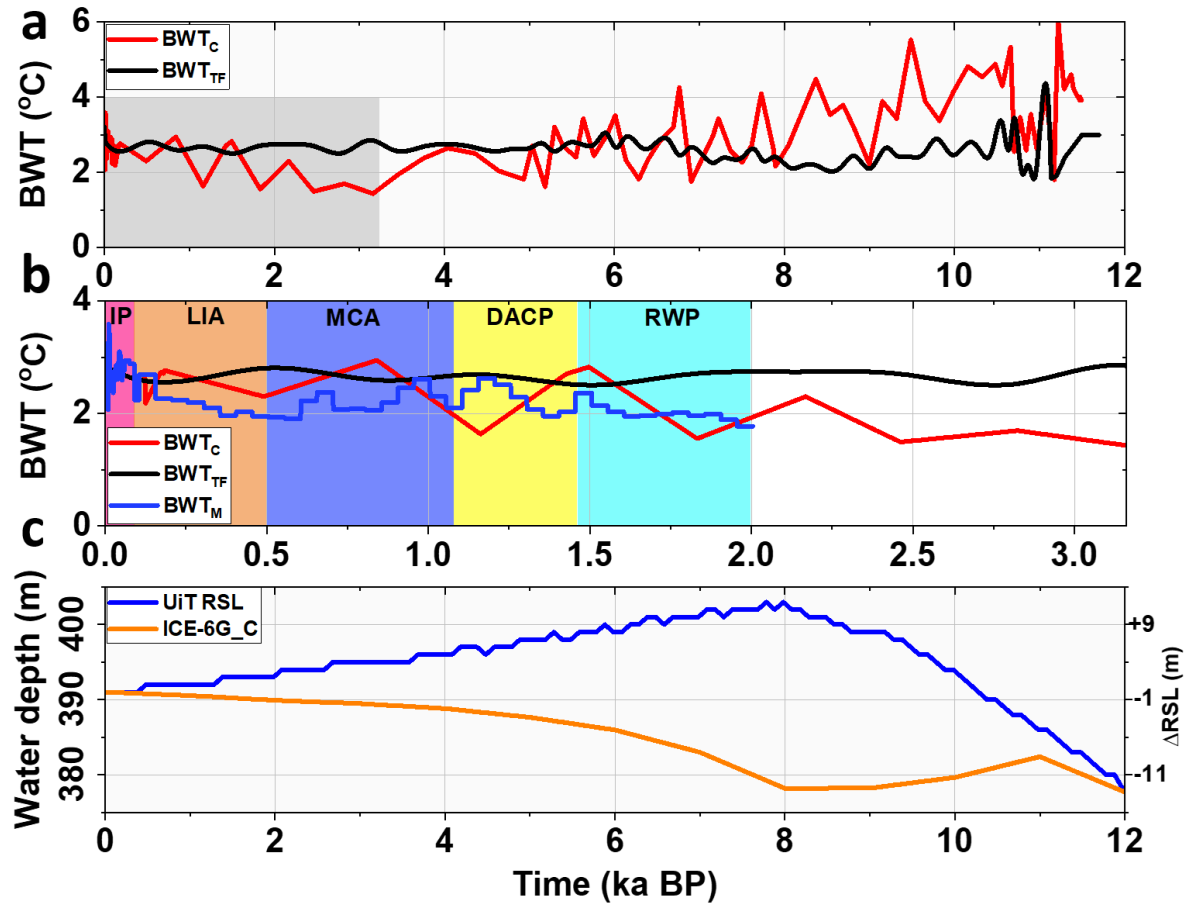


Figure 3.3. BWT and RSL changes used as model boundary conditions. **(a)** Bottom water temperatures derived from shelf core NP05-21GC (327 m depth) based on a benthic foraminifera transfer function (BWT_{TF}) and oxygen isotopes (BWT_C) (Rasmussen et al., 2014) for the period 11.7–0 ka. **(b)** Reconstructed BWT_M values from Marín-Moreno et al. (2013) during the Roman Warm Period (RWP, 2005–1405 BP), the Dark Ages Cold Period (DACP, 1405–1105 BP), the Medieval Climate Anomaly (MCA, 1105–505 BP), the Little Ice Age (LIA, 505–100 years BP), and the Industrial period (IP, 100–present) (Spielhagen et al., 2011). For the IP, BWT values are from Polyakov (2004) for 1900–1950, Holliday et al. (2008) for 1950–1975, and CTD-based near-seabed temperature values from Westbrook et al. (2009) and Thatcher et al. (2013) for 1975–2005. **(c)** Modeled RSL (relative to present) over the past 12000 years at the MeBo site (Fig. 3.1b) based on the University of Tromsø (UiT) ice load history and Earth rheology model (Auriac et al., 2016) (UiT RSL) and the ICE-6G_C scenario (ICE-6G_C RSL).

3.3.3. Relative sea-level change

We considered two relative sea-level (RSL) change scenarios during the past 11200 years (Figure 3.3c) – (1) the RSL values of Wallmann et al. (2018), which are based on the

University of Tromsø (UiT) model of Eurasian ice sheet change (Patton et al., 2017), and (2) the RSL change predicted using the global ICE-6G_C ice history model (Peltier et al., 2015; Auriac et al., 2016). The UiT model of ice sheet change was developed using a thermo-mechanical ice model coupled to a climate model (Patton et al., 2017). The UiT RSL curve used here is an extended version of the curve used by Wallmann et al. (2018), which only extended back to 8 ka BP. It was derived by combining the isostatic response to ice load change, calculated using an elastic lithosphere/relaxed asthenosphere model (Le Meur and Huybrechts, 1996), with a eustatic sea-level curve (Waelbroeck et al. 2002).

The ICE-6G_C model is the updated version of its predecessor ICE-5G (Argus et al., 2014; Peltier et al., 2015). ICE-6G_C was tuned and tested using a suite of relative sea-level data, GPS observations of vertical motion, and time-dependent gravity change. Auriac et al. (2016) found that ICE-6G_C provides the best fit to RSL data across the Barents Sea when combined with an Earth model that has a relatively thin lithosphere (71 km), an upper mantle viscosity of 0.2×10^{21} Pa.s, and a lower mantle viscosity of 2×10^{21} Pa.s. However, when seeking to fit RSL data from Svalbard alone they identified an optimum Earth model with 96 km lithosphere, 0.3×10^{21} Pa.s upper mantle viscosity, and 2×10^{21} Pa.s lower mantle viscosity (Auriac et al., 2016; their Table 4). We combined the ICE-6G_C ice history with two lithosphere thickness values, 71 and 96 km, and a range of combinations of upper and lower mantle viscosities (Supplementary Figure S3.2). A global glacial isostatic adjustment (GIA) model was used to predict gravitationally self-consistent RSL for the ICE-6G_C model at Prins Karl Forland – to compare with existing RSL data from Forman et al. (2004) – and the MeBo site (Supplementary Figure 3.2 and Table S3.2). Differences between model predictions are small (< 5 m) during the period of interest, and we use the model characterized by 71 km lithosphere thickness, 0.2×10^{21} Pa.s upper mantle viscosity, and 5×10^{21} Pa.s lower mantle viscosity.

We use both the UiT and the ICE-6G_C ice history models to provide RSL boundary conditions when modelling the past behaviour of methane hydrates because they predict contrasting behaviour over the past 11000 years (Figure 3.3c). The UiT scenario predicts a rise in RSL (increase in water depth) for the upper slope site over 11200–8000 years BP followed by a fall. The ICE-6G_C model predicts steady RSL prior to 8000 years BP and a steep RSL rise between 8000–4000 years BP, followed by a gentler rise during the last 4000 years.

3.4. Results

We ran the 1-D hydrate model with four different plausible combinations of BWT and RSL changes, viz. BWT_{TF} & UiT RSL, BWT_{TF} & ICE-6G_C RSL, BWT_C & UiT RSL, BWT_C & ICE-6G_C RSL (Figures 3.3a and 3.3c) to understand the past behavior of methane hydrates in response to both temperature and pressure changes. At the start of each model, we assume that any chloride anomaly due to the formation of the hydrate that is already present or due to hydrate dissociation before the start of the model run has diffused away. We present results showing the transient behavior of the hydrate layer that dissociates into methane and freshwater. The MeBo drilling experiment detected a freshwater anomaly, but it did not directly recover hydrate, i.e., we do not have direct constraints on hydrate saturation values. We varied the initial hydrate saturations with the aim to identify the critical hydrate saturation for which hydrates completely dissociated, and the final chloride values mimicked the observed values (Wallmann et al., 2018).

3.4.1. Response to BWT_{TF} variations and UiT RSL change (Scenario 1)

The evolution of the hydrate layer in response to the BWT_{TF} time series and a fall in RSL (UiT), which is considered Scenario 1 (Figures 3.3a and 3.3c), is illustrated in Figure 3.4. The model forcing incorporated in this scenario is similar to Wallmann et al. (2018). Table S3.3 shows the differences between our model and the depressurization model proposed by Wallmann et al. (2018).

We initialized the model at 8.6 ka when the seabed temperature was at 2.05° C, and high sea level favored a thicker initial hydrate column (Figure 3.4a and Supplementary Figure 3.1a) compared to Wallmann et al. (2018). We assumed 12 % initial hydrate saturation and a gas layer beneath it with 2–4 % saturation (Figure 3.4a), 560 mM Cl⁻ concentration (Figure 3.4b) equivalent to pore water salinity of 3.5 wt %, and a steady geothermal gradient of 50° C/km (Figure 3.4c) in the sediment column. For this scenario, the dissociation of 12 % hydrate resulted in the chloride anomaly that best matched the observations (Figures 3.4a–c).

During 8.6–7.6 ka, the seabed temperature increased from 2 to 2.6 °C, and RSL increased by ~ 2 m (Figures 3.3a and 3.3c). During this time, a small amount of hydrate dissociated at the base of the hydrate stability zone (Figure 3.4a). Subsequently, RSL dropped by ~ 3 m, and BWT increased from 2.6 to 3 °C over 7.6–5.9 ka. During this time, hydrate dissociation was governed by warming and depressurization (Figure 3.4a), freshening the pore water and resulting in a chloride anomaly (Figure 3.4b). The methane released due to hydrate dissociation flowed upward through the stability zone and vented into the ocean. When the

BWT reduced to 2.5° C, such as around 5.3 ka, methane was converted into hydrate and chloride concentration increased at shallower depths (Figures 3.4a and 3.4b). In the Wallmann et al. (2018) model, shallow hydrate reformation was not considered, and methane vented rapidly. From 5 ka to 0.4 ka, RSL fall continued to erode the hydrate layer. Between 0.4–0.08 ka, the remaining hydrate layer dissociated due to high BWT conditions ($> 2.5^{\circ}\text{C}$), since the RSL change was negligible, and the geothermal gradient shifted toward the steady geotherm (Figure 3.4c).

According to the UiT model, isostatic rebound resulted in an ~ 8 m fall in RSL between 6 ka and present at our study site (Wallmann et al., 2018). Similar to Wallmann et al. (2018), the consequent drop in hydrostatic pressure was responsible for hydrate dissociation. Hydrate saturation in our model is significantly less than their model; however, the initial hydrate layer is six times thicker than their assumption. In the Wallmann et al. (2018) model, the hydrate layer completely dissociated 1000 years BP producing a large chloride anomaly that progressively diffused away, and methane flow ceased. However, in our case, hydrate dissociation was not completed until 80 years BP, at which point it resulted in a chloride minimum in the system. Subsequently, the chloride anomaly diffused away to match the observed values (Figure 3.4b) and methane flow was maintained at a low rate (Figure 3.4d).

We also tested an initial hydrate saturation of 10 % for this particular pressure-temperature scenario. This model could not fit the observed chloride saturation profile since the hydrate dissociated earlier (Supplementary Figure S3.3).

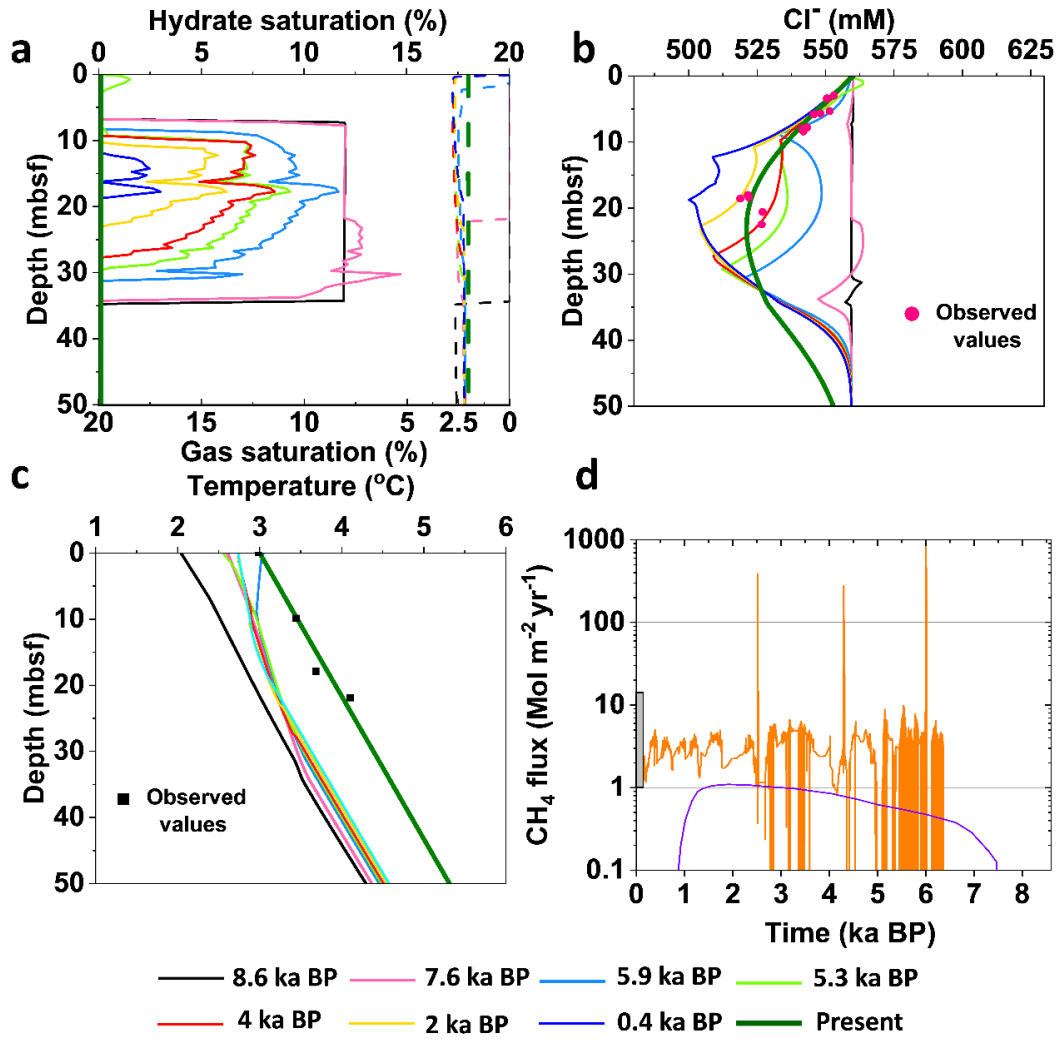


Figure 3.4. Simulation results for pressure-temperature driven (UiT RSL and BWT_{TF} as model forcing) methane hydrate dissociation. The results are plotted at a series of specific times. **(a)** Evolution of the hydrate layer thickness and saturation (solid line) and gas layer beneath it (dash line) over the simulation period. Hydrate completely dissociated 80 years BP. **(b)** Chloride depletion suggests pore water freshening as hydrates dissociate. The chloride concentration started to diffuse away after hydrate dissociation was complete. Pink dots indicate values in cores retrieved at 391 m water depth (Wallmann et al., 2018). **(c)** Temperature profile in the sediment column. Black squares indicate temperatures measured in drill holes at 391 m water depth. **(d)** Methane emission rate from sediments into the overlying water column. Modeled fluxes induced by hydrate dissociation over the past 8610 years (orange curve) are compared to the area-averaged range of present-day methane gas fluxes measured at active seeps (grey vertical bar) in the study area (Sahling et al., 2014; Wallmann et al., 2018; Ferré et al., 2020) and methane flux (purple curve) from the depressurization-induced hydrate dissociation model of Wallmann et al. (2018).

3.4.2. Response to BWT_{TF} variations and ICE-6G_C RSL change (Scenario 2)

In an alternate scenario 2, we tested the impact of BWT_{TF} and ICE-6G_C RSL on the hydrate layer. The ICE-6G_C model predicts RSL rise over the past ~ 8000 years at the upper continental slope off west Svalbard. The rise in BWT during this period could destabilize hydrate while the RSL rise will have a stabilizing effect; the combined effect of these contrasting factors has not previously been considered.

In this case, we initialized the model at 11.138 ka (rather than 8.6 ka). Initialization at 8.6 ka resulted in a very shallow hydrate layer (0.7–1.2 m) due to the low hydrostatic pressure (~ 380 m water depth), and the dissociation of this hydrate over 8.6–0.04 ka resulted in a chloride anomaly at relatively shallow depths, whereas the observed chloride anomaly is deeper (10–20 mbsf).

The colder BWT at 11.138 ka favoured a deeper initial hydrate layer (7–22 mbsf). We used an initial hydrate saturation of 60 % because this resulted in the chloride anomaly that best matched the observations (Figure 3.5a). During the initial 1100 years of the simulation, intermittent bottom water warming led to enhanced hydrate dissociation with substantial methane and freshwater release (Figures 3.5a–c and Supplementary Figure 3.4). At ~ 10 ka, hydrate saturation was reduced to 40 %. Over the period 10–8 ka, there was a slight RSL fall, resulting in a hydrostatic pressure decrease, and this was accompanied by bottom water cooling (2.7–2 °C, Figures 3.3a, and 3.3c). Dissociation slowed due to the bottom water cooling, but the sediment column remained outside the hydrate stability. Since 4 ka, the rise in RSL is predicted to have been small (Figure 3.3c). The remaining hydrate layer completely dissociated at 40 years BP due to warm bottom water conditions (> 2.5 °C). The resulting chloride anomaly started to diffuse away, and the final chloride values are consistent with the observations (Figure 3.5b). The methane flow (Figure 3.5d) is predicted to have been highest during the earliest Holocene (11.2–10 ka), continuing through to the present day at a reduced rate. In this scenario, the BWT rise was primarily responsible for hydrate destabilization despite the RSL rise.

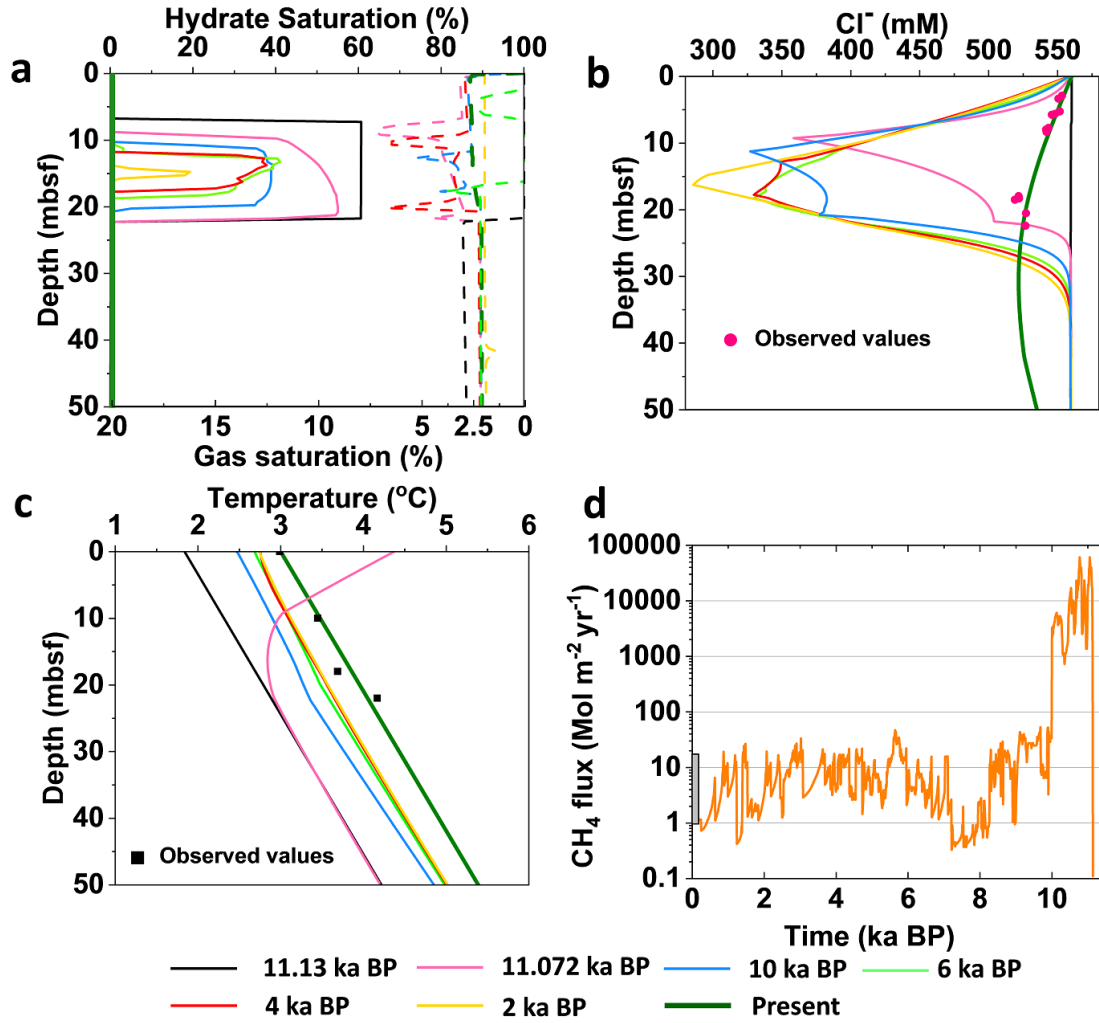


Figure 3.5. Simulation results for pressure-temperature driven hydrate dissociation forced by ICE-6G_C RSL and BWT_{TF} over the past 11000 years. **(a)** Evolution of the hydrate layer thickness and saturation over the simulation period. Solid lines mark hydrate saturation and dashed lines mark gas saturation. Hydrate completely dissociated 40 years BP. **(b)** Dissolved chloride concentration. **(c)** Temperature profile in the sediment column. Black squares indicate temperatures measured in drill holes at 391 m water depth. The steep rise in temperature towards the seafloor at 11.072 ka reflects the sharp increase in BWT (Figure 3.3a) imposed at the top boundary cell. Temperature profiles at selected times are shown for clarity. **(d)** Modelled methane emission rates from sediments into the overlying ocean are compared to the area-averaged range of present-day methane gas fluxes measured at active seeps (grey vertical bar) in the study area (Sahling et al., 2014; Wallmann et al., 2018; Ferré et al., 2020). The methane flow rate was highest during the early Holocene.

3.4.3. Response to BWT_C variations with UiT and ICE-6G_C RSL change (Scenarios 3 and 4)

In two additional scenarios, we applied BWT_C with UiT RSL (Scenario 3) and BWT_C with ICE-6G_C RSL (Scenario 4) to the hydrate layer. To investigate the development of hydrate in each of these cases, we initialized the model when the BWT_C attained a minimum value of 1.4 °C at 3.16 ka (Figure 3.3b) using a hydrate saturation of 5 % and we ran the model to observe changes in the hydrate layer. Simulation results for scenarios 3 and 4 are shown in Figures 3.6a-d and 3.6e-h, respectively.

In scenario 3 (BWT_C+UiT RSL), hydrate dissociation was triggered by depressurization and BWT rise, while in scenario 4 (BWT_C+ICE-6G_C RSL), dissociation was exclusively induced by bottom water warming. Initially, gas flowed through the hydrate stability zone (Figures 3.6a and 3.6e) and was converted into hydrates, increasing hydrate saturation by ~ 10–15 % as well as chloride concentration (Figures 3.6b and 3.6f). The BWT_C record documents two prominent warming pulses (BWT > 2.5 °C) at ~ 1.5 ka and ~ 0.7 ka (Figure 3.3b). During these warming episodes, the shallowest hydrate dissociated in both models, and gas reached the seafloor. Hydrate completely dissociated at 53 and 49 years BP in scenarios 3 and 4, respectively, and the final chloride anomalies matched the observed values at the end of the simulations (Figures 3.6b and 3.6f). The temperature profiles migrated toward the steady geotherm once the hydrate was exhausted (Figures 3.6c and 3.6g). Chloride sensitivity was gauged by varying the initial hydrate saturation. The final chloride values for the model with hydrate saturation more than 5 % did not fit the observed chloride profile.

We also compared the response of the hydrate layer to the BWT reconstruction of Marín-Moreno et al. (2013) (refer to section 3.3.2). In this case, we initialized the model with 5 % hydrate saturation (Supplementary Figures 3.5a and 3.5d) at 2.16 ka and forced it using BWT_M (Figure 3.3b) and both sets of RSL values (UiT and ICE-6G_C) (Figure 3.3c). In these cases, hydrate dissociated completely at 49 and 46 years BP for BWT_M+ UiT RSL and BWT_M+ ICE-6G_C RSL, respectively, and final chloride values matched the present-day observations (Supplementary Figure 3.5). Although there are differences between BWT_C and BWT_M over the past 2 ka, model outputs at the end of the simulations did not show any major differences.

The experiments described above were initialized in the late Holocene. To check whether some hydrates could have survived warming episodes earlier in the Holocene, we initialized the model with a high hydrate saturation of 60 %, as used in Wallmann et al. (2018), at 11.2

ka (Supplementary Figures 3.6 and 3.7). In the BWT_C and ICE-6G_C scenario, hydrate dissociation was completed at 10.6 ka (Supplementary Figure 3.6a) and was driven by seabed warming and minor RSL fall (~ 2 m). In the BWT_C and UiT scenario, the hydrate layer completely dissociated at 9.6 ka (Supplementary Figure 3.7a) in response to early Holocene warming despite ~ 10 m RSL rise. In both cases, the chloride concentration was restored in the next thousand years (Supplementary Figures 3.6b and 3.7b). The equivalent early Holocene warming model by Wallmann et al. (2018) also showed complete hydrate dissociation by 10.7 ka; however, they did not consider any RSL change.

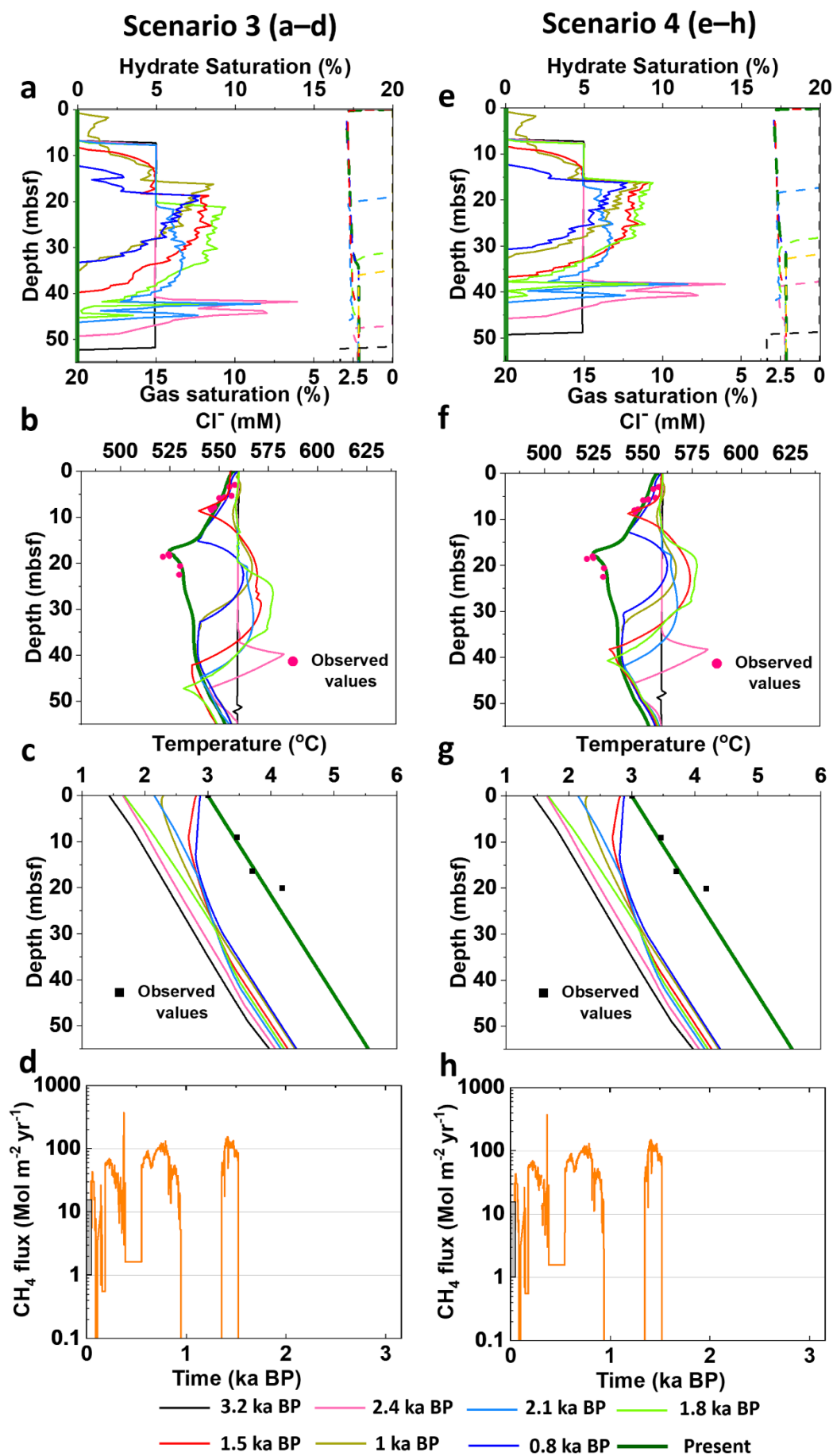


Figure 3.6. Simulation results show methane hydrate dissociation forced by $BWT_C + UiT$ RSL (a–d) and $BWT_C + ICE-6G_C$ RSL (e–h) over the past 3160 years (Figures 3.3b and 3.3c). (a, e) Evolution of the hydrate layer thickness and saturation over the simulation period. Solid lines mark hydrate saturation and dashed lines mark gas saturation. Hydrate dissociated at the base, and gas migrated through the hydrate stability zone. Hydrate completely dissociated at 53 and 49 years BP in scenarios 3 and 4, respectively. (b, f) Change in dissolved chloride concentration. Hydrate dissociation caused pore water freshening. The chloride concentration started to diffuse after the completion of hydrate dissociation, and the final profiles match the present-day observations (dots). (c, g) The temperature profile in the sediment column. Black squares indicate temperatures measured in drill holes at 391 m water depth. (d, h) Methane emission rate from sediments into the overlying ocean (orange curve) and area-averaged range of present-day methane gas fluxes (grey vertical bar) in the study area (Sahling et al., 2014; Wallmann et al., 2018; Ferré et al., 2020).

3.4.4. Response to final 38 years of BWT changes

It is evident from the previous long-time series runs that complete hydrate dissociation at this site was possible well before the accelerated bottom water warming during 1978–2008 (Westbrook et al., 2009). For example, dissociation of 12 % hydrate was complete at 80 years BP in scenario 1 (BWT_{TF} and UiT RSL, Figure 3.4a), while in scenarios 3 and 4, covering the past ~ 3200 years (BWT_C and UiT RSL, Figure 3.6a and BWT_C and $ICE-6G_C$ RSL, Figure 3.6e), 5 % hydrate completely dissociated around 50 years BP.

In such cases, we examine whether hydrate reformation and subsequent dissociation were possible over the last 38 years in our model runs (1978–2016). Several factors provided suitable conditions for hydrate reformation during this period. During the previous episode of hydrate dissociation, pore water freshening would have enhanced hydrate stability. Similarly, gas from previously dissociated hydrate and deeper sources would have been available to be converted into hydrate during cold BWT conditions. Importantly, any new hydrate formation will have erased any earlier chloride depletion anomaly because chloride is excluded from the hydrate during the reformation.

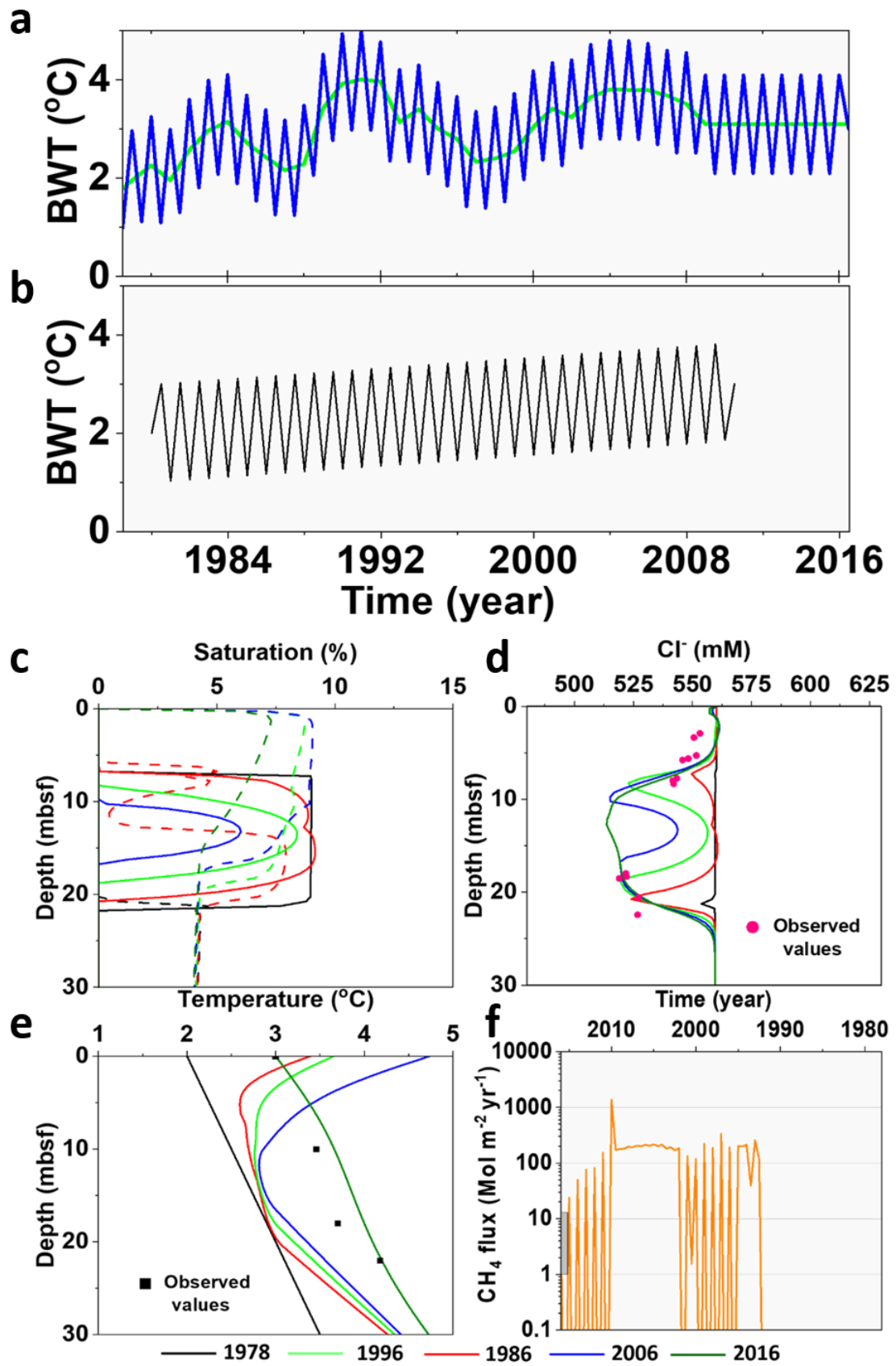


Figure 3.7. BWT variations at 400 m water depth off west Svalbard and simulation results showing the behaviour of hydrate forced by BWT variations over 1978–2016 in a sediment column with permeability 10^{-13} m^2 . **(a)** The instrumental record of BWT during 1978–2016 (Green curve, Ferré et al. 2012) with seasonal fluctuations of $\pm 1 \text{ }^\circ\text{C}$ (Blue) following Riedel et al. (2018). **(b)** The 1980–2010 linear temperature rise from 2–3 $^\circ\text{C}$ was adopted by Wallmann et al. (2018) with seasonal fluctuations of $\pm 1 \text{ }^\circ\text{C}$. **(c)** Methane hydrate saturation over the simulation period. The initial hydrate saturation was 9 % in 1978. Solid lines mark hydrate saturation and dashed lines mark gas saturation. Hydrate saturation decreased with progressive thermal dissociation. Hydrate completely dissociated 2 years before the end of the model run. **(d)** Dissolved chloride concentration. The final chloride profile provides a good fit to the observations. **(e)** The temperature profile in the sediment column. Black squares indicate temperatures measured in drill holes at 391 m water depth. **(f)** Modeled methane fluxes induced by methane hydrate dissociation compared to the area-averaged range of present-day methane gas fluxes measured at active seep sites (grey vertical bar) (Ferré et al., 2020). In the experiment, methane appeared at the seabed 14 years after the onset of warming.

Our 38-year run (1978–2016), driven by the BWT time series shown in Figure 3.7a, finds that an initial hydrate layer with 9 % hydrate saturation completely dissociates and produces a chloride profile that matches the observations (Figures 3.7c and 3.7d). Since the RSL change in past 4 decades is negligibly small, we did not include RSL change in this model. The net 1 $^\circ\text{C}$ BWT (Figures 3.7a and 3.7e) rise led to hydrate dissociation, and the resulting methane flowed through the hydrate stability zone (Figure 3.7c). Methane appeared at the seabed 14 years after the onset of warming when we considered a permeability of 10^{-13} m^2 (Figure 3.7f). We observed periodicity in methane flow (Figure 3.7f) that is attributed to seasonally-induced hydrate formation and dissociation at a shallow depth. The seasonal fluctuations in BWT (Figure 3.7a) strongly affected the top 10 m of the sediment column, where we find repeated episodes of shallow hydrate reformation and dissociation during colder and warmer conditions, respectively.

Previously, Wallmann et al. (2018) ran a 30-year model (1980–2010) driven by a linear temperature increase from 2 to 3 $^\circ\text{C}$ that included seasonal temperature variations $\pm 1 \text{ }^\circ\text{C}$ (Figure 3.7b). The initial hydrate layer in their model was at 10–30 mbsf, and hydrate saturation was varied in the range of 6–8 %. Hydrate did not completely dissociate by the end of their model run, and the chloride anomaly could not be reproduced. Hence, they concluded

that the thermal dissociation of hydrates over the past three decades was unlikely. Significant time and energy are required to dissociate 6–8 % hydrate. Although measured down-core temperature gradients lie in 45–50 °C km⁻¹ (Riedel et al., 2018), their 30-year model assumed a geothermal gradient of 45 °C km⁻¹.

Our model was run for an additional 8 years compared to Wallmann et al. (2018), so there was sufficient time for hydrate dissociation. The heat flux (10×10^{-2} W m⁻²) applied at the base of our model resulted in a geothermal gradient of 50° C/km in the sediment column, which, in turn, resulted in a thinner initial hydrate layer (14 m thick, 7–21 mbsf) than the 20 m thickness considered by Wallmann et al. (2018). Our thinner hydrate layer completely dissociated over 38 years. However, when we considered a lower geothermal gradient (45° C/km) by applying a heat flux of 9×10^{-2} W m⁻², the initial hydrate layer was thicker (22 m thick, 7–29 mbsf). It did not completely dissociate after 38 years.

We also tested lower (10^{-15} m²) and higher permeability (10^{-11} m²) scenarios (Supplementary Figure 3.8) and different irreducible gas saturation (S_{irg}) values ranging between 2–10 % (Ma et al., 2020). We found that hydrate dissociated completely in all cases, and the methane flow rate was very sensitive to the permeability and S_{irg} values (Thatcher et al., 2013). As expected, a reduction in permeability caused a delay in the arrival of the methane at the seabed (Supplementary Figure 3.9a), while an increase in the S_{irg} value caused a delay in the arrival of the methane at the seabed and a decline in the maximum methane flux (Supplementary Figure 3.9b).

The simulations described in this section emphasize the role of recent BWT rise as a potential cause for hydrate dissociation and pore water freshening. They also illustrate the role of interannual temperature changes in controlling changes in seafloor methane flux in response to near seafloor hydrate dissociation-formation stages.

3.5. Discussion

We have examined the impact of RSL change coupled with BWT variations on hydrate west of Svalbard at 391 m water depth, where recent drilling confirmed pore water freshening due to hydrate dissociation. Table 3.1 summarizes model outputs from scenarios where the final chloride anomaly and subsurface temperature profile closely match the present-day observations.

The previous modelling by various authors either emphasized the role of temperature rise or RSL fall in explaining hydrate destabilization. For example, Thatcher et al. (2013), Marín-

Moreno et al. (2013), and Marín-Moreno et al. (2015) considered the effects of BWT on hydrates, while Wallmann et al. (2018) attributed hydrate dissociation to RSL fall at constant BWT. Extending these studies, we have examined the fate of hydrates in response to the combined effects of a plausible set of BWT and RSL values.

We have demonstrated plausible scenarios of hydrate dissociation based on RSL reconstruction and BWT changes that show prominent millennial to seasonal changes over the past 8000 years at different time scales (decadal to seasonal). All of the scenarios considered here predict thinning of the hydrate layer from the base during the Holocene, releasing methane that can get trapped in shallow sediments under cold BWT conditions ($< 2.5^{\circ}\text{C}$). Sharp rises in BWT dissociate the shallow hydrates and enhance gas flow at the seabed. Periods of high methane flow at the seabed could lead to seep-carbonate accumulation at the feather edge of hydrate stability. The precipitated carbonate can influence benthic ecosystems since it provides hardground shelter for them (Wilson et al., 2007). Thicker carbonate accumulations of varying ages could be related to multiple episodes of hydrate dissociation (Berndt et al., 2014). Our modelling shows a range of possible past methane emission scenarios. Future work should focus on mineralogical and isotopic studies of authigenic carbonates (e.g., Liang et al., 2017) to unravel fluid sources, long-term seepage dynamics, and the fate of Arctic methane hydrates.

Our results based on coupled BWT_{TF} and UiT RSL values (Scenario 1) challenge the hypothesis that a fall in RSL was the sole cause of hydrate dissociation over 8–1 ka (Wallmann et al., 2018). The depressurization model proposed by Wallmann et al. (2018) only considered a constant BWT over 8–0.1 ka. By using the UiT RSL in combination with the more realistic BWT_{TF} curve, we find that pulses of warm bottom water temperatures ($> 2.5^{\circ}\text{C}$) also contributed to hydrate dissociation together with depressurization. The initial hydrate saturation was significantly higher (60 %) in the Wallmann et al. (2018) model compared to the initial saturation (12 %) in our BWT_{TF} and UiT RSL scenario. Previously inferred hydrate saturations in the top 50 m of glaciomarine sediments on the upper continental slope of Svalbard, in water depths of 480–866 m, range between 5–22 % (Chabert et al., 2011 and Goswami et al., 2016). While the initial hydrate saturations considered here are within this range in most cases, we infer a higher value (60 %) in the model forced by BWT_{TF} and ICE-6G_C RSL (Scenario 2). We needed such a high initial hydrate saturation to generate the chloride anomaly that mimics present-day observations. In scenario 2, hydrates of such high saturation slowly dissociated over a prolonged period (11138 years). The model

shows a prolonged record of methane emission at the seabed, with methane flux progressively decreasing with time (Figure 3.5d).

Based on their model (UiT RSL), Wallmann et al. (2018) concluded that methane fluxes from hydrate dissociation ceased at 1 ka upon completion of depressurization-induced dissociation, and the present-day gas seeps are not related to warming-induced hydrate dissociation. This inference disagrees with the findings of Veloso-Alarcon et al. (2019) and Ferré et al. (2020), who proposed that the present-day methane flux is likely controlled by temperature-driven seasonal hydrate dynamics. In our simulation using the BWT_{TF} and UiT RSL, complete hydrate dissociation takes longer than in Wallmann et al. (2018) mainly because some time is required between the onset of hydrate dissociation and methane reaching the seabed, and it depends on effective permeability, porosity, irreducible gas saturation, hydrate saturation, distance between the top of the hydrate layer and the seafloor (Thatcher et al., 2013). For this scenario, the modeled methane flux at present (year 2016) lies at the lower end of the range of observed values ($1 \text{ mol m}^{-2} \text{ year}^{-1}$, compared with $1\text{--}13 \text{ mol m}^{-2} \text{ year}^{-1}$, Ferré et al., 2020), suggesting that other possible explanations for the methane seeps should be investigated.

Despite the possibility of several episodes of hydrate formation and dissociation in the past, the latest episode of hydrate formation and subsequent warming (1978–2016) stands out as a strong possibility that can explain the observed pore water freshening and variations in methane flux at the seabed. The observed gas flare activity shows large variations that are strongly influenced by seasonal hydrate dynamics (Berndt et al., 2014; Veloso-Alarcon et al., 2019; Ferré et al., 2020). This pattern is well explained by our 38-year simulation (1978–2016), where the initial 9 % hydrate saturation completely dissociated, and hydrate dynamics are clearly influenced by decadal- and seasonal-scale BWT fluctuations (Figure 3.7f). There is seismic evidence of gas (Chabert et al., 2011; Sarkar et al., 2012) in the top 100 m of the sediment column at the upper continental slope (water depths 350–1200 m). In all the cases we considered (Table 3.1), our modelling predicts that hydrates have completely dissociated prior to 1978. However, the combination of the delayed upwards migration of the resulting methane through vertical fractures and permeable strata (Sarkar et al., 2012) and colder BWTs, such as the remarkable cooling (10-year) during the late 1970s (Westbrook et al., 2009), likely created favourable conditions for hydrate re-formation. Since no hydrates were recovered by the drilling, ocean bottom warming must have completely dissociated these newly-formed hydrates to explain the observed pore-water freshening. Therefore, there is

insufficient evidence to reject the (*recent*) warming-induced hydrate dissociation hypothesis (Westbrook et al., 2009) in explaining the present-day methane seeps off west Svalbard.

Our 38-year model provides valuable insights into the response of methane hydrates to recent warming at the feather edge of hydrate stability in this area. Moreover, it evidences the importance of reducing uncertainty in the predictions of methane emissions, particularly on the continental shelf. Since the shelf regions are at shallower water depths (100–200 m), seafloor methane emissions have a greater chance to reach the atmosphere. Even if methane does not reach the atmosphere in these areas, methane dissolution and its oxidation in the ocean can cause local changes in pH, thereby affecting the local marine biodiversity (Valentine et al., 2001; Riebesell, 2008). Although gas emission activity could substantially subside at the feather edge of hydrate stability during bottom water cooling on the upper continental slope (~ 390 m), deeper gas can still be deflected by the impermeable hydrate layer and migrate upslope towards the shelf through permeable beds (Sarkar et al., 2012; Veloso-Alarcon et al., 2019). The beds outcrop at the shelf, and methane is released directly into the ocean. Veloso-Alarcon et al. (2019) reported enhanced methane flux on the continental shelf (100–200 m water depths) compared to the upper slope during cooler BWT conditions since the shelf is outside the gas hydrate stability. Strong methane emission has a greater chance of reaching the sea surface and atmosphere, contributing to further warming (Westbrook et al., 2009).

Model forcing and duration of simulation (2016 represents 'Present')	Time taken to completely dissociate hydrate (years)	Initial hydrate saturation, initial hydrate layer thickness	Correlation between observed chloride anomaly and final chloride values obtained from the model	Methane flux at the end of the simulation (Observed area-averaged methane flux is 1–13 Mol m ⁻² y ⁻¹)
Scenario 1: BWT _{TF} + UiT RSL (8.6–0 ka)	8520	12 %, 30 m	0.9	1
Scenario 2: BWT _{TF} + ICE-6G_C RSL (11.138–0 ka)	11098	60 %, 15 m	0.8	1
Scenario 3: BWT _C + UiT RSL (3.16–0 ka)	3107	5 %, 45 m	0.9	2.25
Scenario 4: BWT _C + ICE-6G_C RSL (3.16–0 ka)	3111	5 %, 42 m	0.9	2.2
Scenario 5: BWT from (1978–2016) (Ferré et al., 2012)	36	9 %, 14 m	0.94	7

Table 3.1. Summary table showing the modelling scenarios and quantification of how well the final results (chloride values, and methane flux) matched the observations.

3.6. Conclusions

We have re-examined the fate of Arctic marine hydrates in response to BWT and RSL changes at the landward limit of hydrate stability off west Svalbard. We have considered an alternative ice history model (ICE-6G_C) that suggests RSL rise and contradicts the RSL fall predicted by the UiT model over the past 8000 years used in a previous modelling study in the area (Wallmann et al., 2018). We demonstrate that with the available constraints, multiple

episodes of hydrate formation and dissociation are still possible based on a comprehensive scenario testing approach that has never been attempted before for such a long-time scale.

In the simulation that used the UiT scenario, we found that although RSL fall may have triggered hydrate dissociation, decadal-scale BWT fluctuations (BWT_{TF} and BWT_C) played a critical role in controlling hydrate dynamics. Specifically, in these scenarios the eventual fate of methane released from depressurization-induced hydrate dissociation depended on BWT. Warmer seabed temperatures (e.g., $> 2.5\text{ }^{\circ}\text{C}$) prevented the shallow hydrate reformation, allowing gas to flow through the hydrate stability zone and into the ocean. On the other hand, colder bottom water temperatures facilitated shallow hydrate reformation. In the simulations that used the ICE-6G_C scenario, hydrate dissociation was primarily caused by warming of bottom waters, which overrode the stabilising effect of RSL rise.

We have not been able to determine the actual RSL and BWT scenario through which the hydrate system evolved since several scenarios produce outputs that match the present-day observations. However, we demonstrate that earlier hydrates completely dissociated before the late 1970s' bottom water cooling episode, irrespective of the assumed scenario. During this cooling episode, hydrate could have reformed and then dissociated in response to BWT rise over the period 1978–2016. The gas produced by dissociation takes time to migrate to the seabed and can explain the modern methane seeps. Seasonal warming and cooling cycles impact hydrate dynamics and can explain the large observed variations in methane flux. The pore water freshening observed in the marine sediments indicates latest episode of hydrate destabilization caused by the latest episode of seabed warming.

If bottom water warming continues at an accelerated rate, we can expect downslope migration of the landward limit of hydrate stability. The results presented here bring into focus the necessity to closely monitor the effect of warming Atlantic waters on hydrates because continued warming may degrade more Arctic marine hydrates in the impending future at current or enhanced rates.

3.7. Auxiliary materials

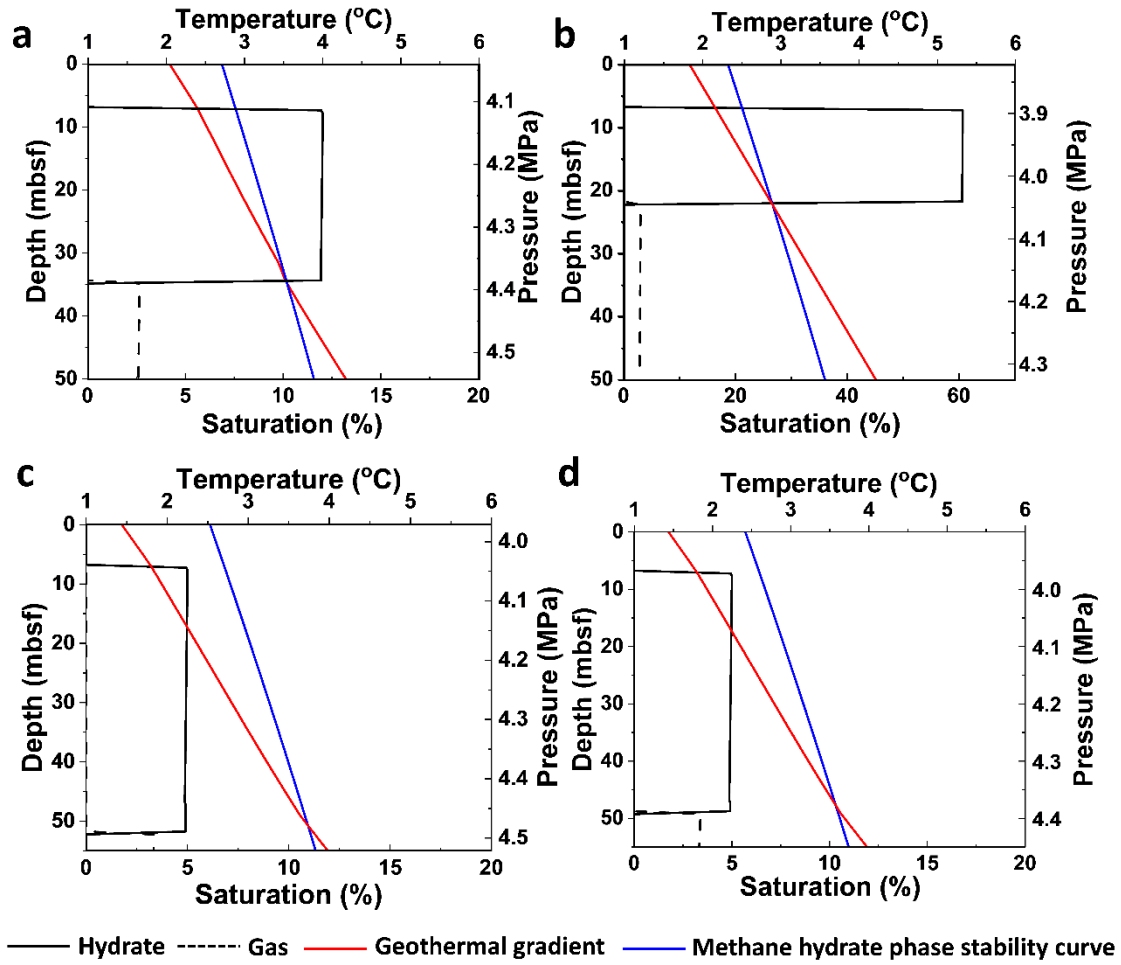


Figure S3.1. Model initialization for different BWT and sea level conditions in the past. Base of the gas hydrate stability zone determined by the intersection of geothermal gradient (50° C/km) and methane hydrate phase stability curve (Moridis et al., 2003) at **(a)** 8.6 ka BP for scenario 1 (BWT_{TF}+ UiT RSL), **(b)** at 11.138 ka BP for scenario 2 (BWT_{TF}+ ICE-6G_C RSL), **(c)** at 3.16 ka BP for scenario 3 (BWT_C+ UiT RSL) and **(d)** at 3.16 ka BP for scenario 4 (BWT_C+ ICE-6G_C RSL).

Table S3.1. Table with physical properties of the methane hydrate system.

Parameter	Value
Initial salinity (wt %) (Moridis, 2003)	3.5
Initial gas saturation below GHSZ (%) (Chabert et al., 2011)	2-4
Initial thickness of SRZ (m) (Thatcher et al., 2013)	7
Gas composition (James et al., 2011)	100 % CH ₄
Water-saturated thermal conductivity (Wm ⁻¹ K ⁻¹) (Riedel et al., 2018)	2
Solid grain density (kg m ⁻³)	2600
Hydrate-free initial intrinsic permeability k_o (m ²) (Thatcher et al., 2013).	10-13
Intrinsic permeability of the hydrate-bearing rock k_i (Moridis et al., 2014)	$k_i = k_o F$ $F = \left(\frac{\varphi_o(1 - S_h) - \varphi_c}{\varphi_o - \varphi_c} \right)^n$ $\varphi_c = 0.05, n = 3$
Initial diffusivity (m ² s ⁻¹) (Marín-Moreno et al., 2013) CH ₄ : aqueous phase, gas phase H ₂ O: aqueous phase NaCl: aqueous phase	 2x10-9, 2x10-5 1x10-9 1.5x10-9
Relative permeability model Modified version of Stone's first three phase relative permeability method (Stone, 1970)	$k_{rA} = \max \left\{ 0, \min \left\{ \left[\frac{S_A - S_{irA}}{1 - S_{irA}} \right]^n, 1 \right\} \right\},$ $k_{rG} = \max \left\{ 0, \min \left\{ \left[\frac{S_G - S_{irG}}{1 - S_{irA}} \right]^{n_G}, 1 \right\} \right\},$ $S_{irA} = 0.12, S_{irG} = 0.02, n = n_G = 4$

Capillary pressure model (Van Genuchten, 1980)	$P_{cap} = -P_0 \left[(S^*)^{-1/\lambda} - 1 \right]^{-\lambda},$ $-P_{max} \leq P_{cap} \leq 0,$ $S^* = \frac{(S_A - S_{irA})}{(S_{mxA} - S_{irA})},$ $\lambda = 0.254, S_{irA} = 0.11, P_0 = 12500 \text{ MPa},$ $P_{max} = 10^6 \text{ MPa}, S_{mxA} = 1$
Heat flow at the model base (Wm^{-2}) (Riedel et al., 2018)	9–10x10 ⁻² (Resulting geothermal gradient of 45–50° C/Km)

Abbreviations: SRZ, sulphate reduction zone; Sh is hydrate saturation; ϕ_0 and ϕ_c are absolute and critical porosity (porosity at which permeability is reduced to zero, due to disconnected pores in the rock); k_{rA} and k_{rG} are relative permeability for aqueous and gas phases, respectively; S_A and S_G are saturations for aqueous and gas phases; S_{irA} and S_{irG} are irreducible aqueous and gas saturations; S_{mxA} is the maximum water saturation; P_{cap} is the capillary pressure; P_{max} is the maximum value of capillary pressure; P_0 is the capillary entry pressure; n , n_G and λ are fitting parameters.

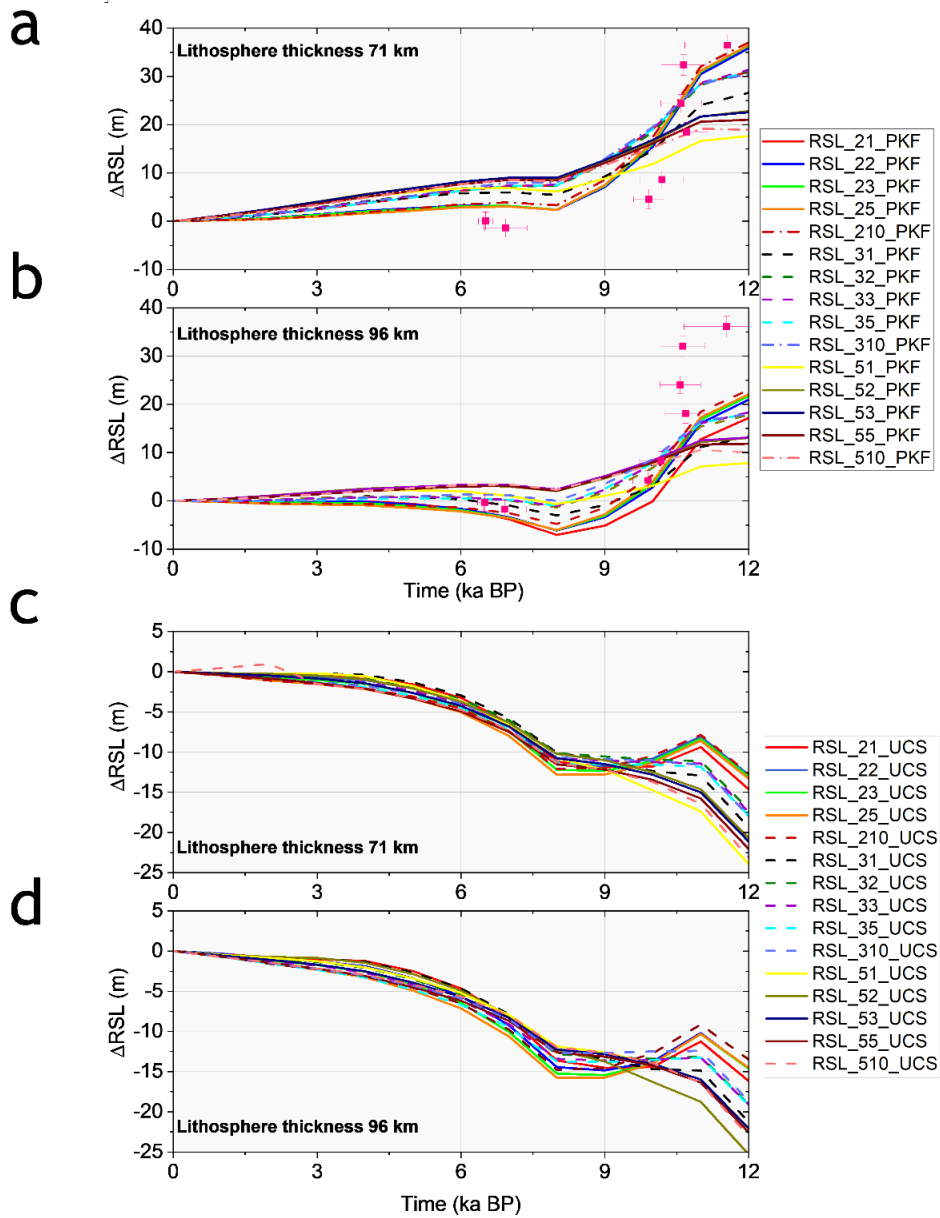


Figure S3.2. RSL change over the past 12000 years predicted by the ICE-6G_C scenario for Prins Karl Forland (PKF) and Upper Continental Slope (UCS, MeBo drill site, present-day water depth 391 m). RSL was computed using a global GIA model. **(a)** RSL change for Prins Karl Forland (PKF) derived using an Earth model with a lithosphere thickness of 71 km and considering variable upper and lower mantle viscosities (Curve details are given in the Supplementary Table S3.2; pink square with error bar denotes RSL observations (Forman, 2004). RSL_22_PKF to RSL_25_PKF curves show good match with the RSL observations. **(b)** RSL change for PKF derived using an Earth model with a lithosphere thickness of 96 km. **(c)** RSL for UCS derived using an Earth model with a lithosphere thickness of 71 km. **(d)** RSL for UCS derived using an Earth model with a lithosphere thickness of 96 km.

Table S3.2: Earth model parameters for two different lithospheric thickness viz. 71 and 96 km.

Location	Lithosphere thickness (km)	Upper mantle viscosity (10 ²¹ Pa.s)	Lower mantle viscosity (10 ²¹ Pa.s)
Prins Karl Forland (PKF)	71	0.2, 0.3 and 0.5	1, 2, 3, 5 and 10
	96		
Upper continental slope (UCS)	71		
	96		

General naming convention of the RSL curves shown in Figure S3.2

RSL_XY_PKF (XY is a two-digit number)-RSL change values with 0.X x 10²¹ Pa.s and Y x 10²¹ Pa.s viscosity values for upper and lower mantle at Prins Karl Forland respectively. UCS represents Upper Continental Slope (MeBo site).

RSL_XYZ_PKF (XYZ is a three-digit number)-RSL change values with 0.X x 10²¹ Pa.s and YZ x 10²¹ Pa.s viscosity values for upper and lower mantle at Prins Karl Forland respectively.

Examples:

RSL_21_PKF stands for RSL change values with 0.2 x 10²¹ Pa.s and 1 x 10²¹ Pa.s viscosity values for upper and lower mantle at Prins Karl Forland respectively.

RSL_22_PKF stands for RSL change values with 0.2 x 10²¹ Pa.s and 2 x 10²¹ Pa.s viscosity values for upper and lower mantle at Prins Karl Forland respectively.

RSL_210_PKF stands for RSL change values with 0.2 x 10²¹ Pa.s and 10 x 10²¹ Pa.s viscosity values for upper and lower mantle at Prins Karl Forland respectively.

RSL_21_UCS stands for RSL change values with 0.2 x 10²¹ Pa.s and 1 x 10²¹ Pa.s viscosity values for upper and lower mantle at Upper Continental Slope (MeBo site) respectively.

RSL_22_ UCS stands for RSL change values with 0.2×10^{21} Pa.s and 2×10^{21} Pa.s viscosity values for upper and lower mantle at Upper Continental Slope (MeBo site) respectively.

RSL_210_ UCS stands for RSL change values with 0.2×10^{21} Pa.s and 10×10^{21} Pa.s viscosity values for upper and lower mantle at Upper Continental Slope (MeBo site) respectively.

Table S3.3. Table highlighting differences between the model proposed by Wallmann et al. (2018) and Scenario 1 (this study).

Model parameters and working	Depressurization model by Wallmann et al. (2018)	Scenario 1 (This study)
Duration	8000 years	8600 years
RSL curve	UiT	UiT
BWT curve	BWT _{TF} was kept constant (2.5° C) between 8 and 0.1 ka, and thereafter, there was a rise to the modern value of 3° C in the past 100 years.	BWT _{TF} curve (Rasmussen et al. 2014) was used (Figure 3a). The curve shows subtle temperature variations that can affect shallow hydrate reformation and methane flow.
Initial hydrate layer thickness	5 m hydrate layer within the potential hydrate stability zone.	30 m
Top of hydrate layer	16 mbsf	7 mbsf
Initial hydrate saturation	60 %	12 %
Geothermal gradient	45–50° C/km	50° C/km

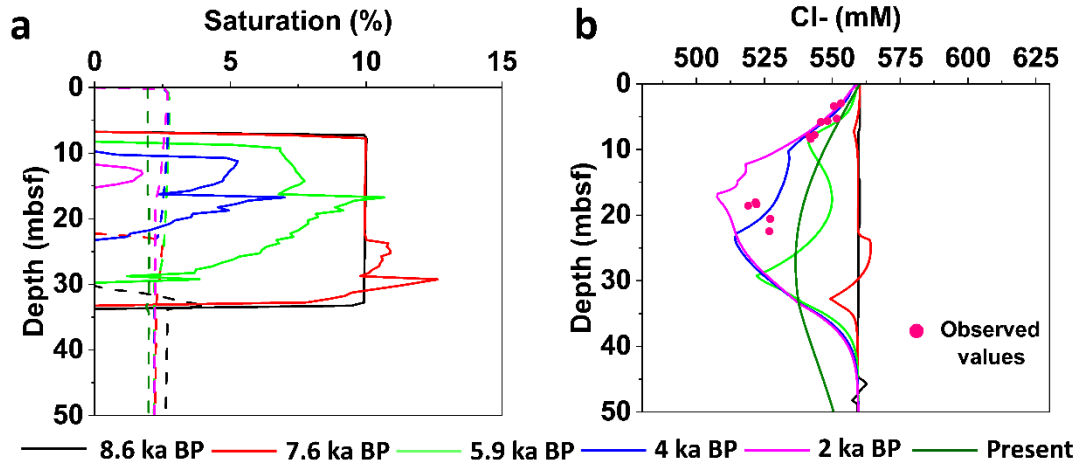


Figure S3.3. Simulation results for pressure-temperature driven methane hydrate dissociation forced by UiT RSL and BWT_{TF} . **(a)** Evolution of the hydrate saturated sediment column. Hydrate saturation is shown by solid line and gas saturation by dash line. The initial hydrate saturation was 10 %. The hydrates dissociated completely at 1.2 ka. **(b)** Dissolved chloride concentration resulting from dissociation of hydrate completely diffused away. At the end of the simulation, the final chloride values did not match the observed chloride values (pink dots).

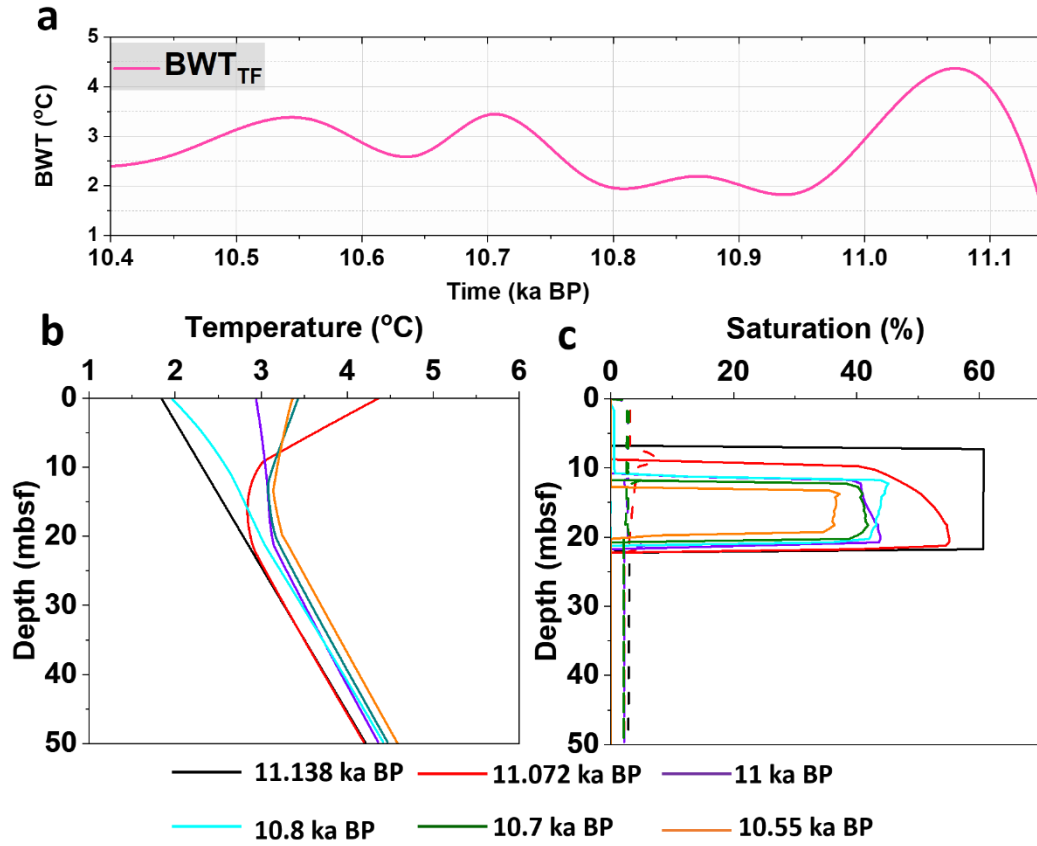


Figure S3.4. Impact of BWT_{TF} and ICE-6G_C RSL variations on the methane hydrate layer during 11.138–10.4 ka. **(a)** BWT variations over 11.138–10.4 ka. **(b)** Temperature profile in the sediment column. **(c)** Evolution of the hydrate saturated sediment column. Solid lines mark hydrate saturation and dashed lines mark gas saturation. The initial hydrate saturation was 60 %. During the initial 66 years (11.138–11.072 ka), the hydrate layer dissociated in response to the rise in BWT (1.83 to 4.36 °C).

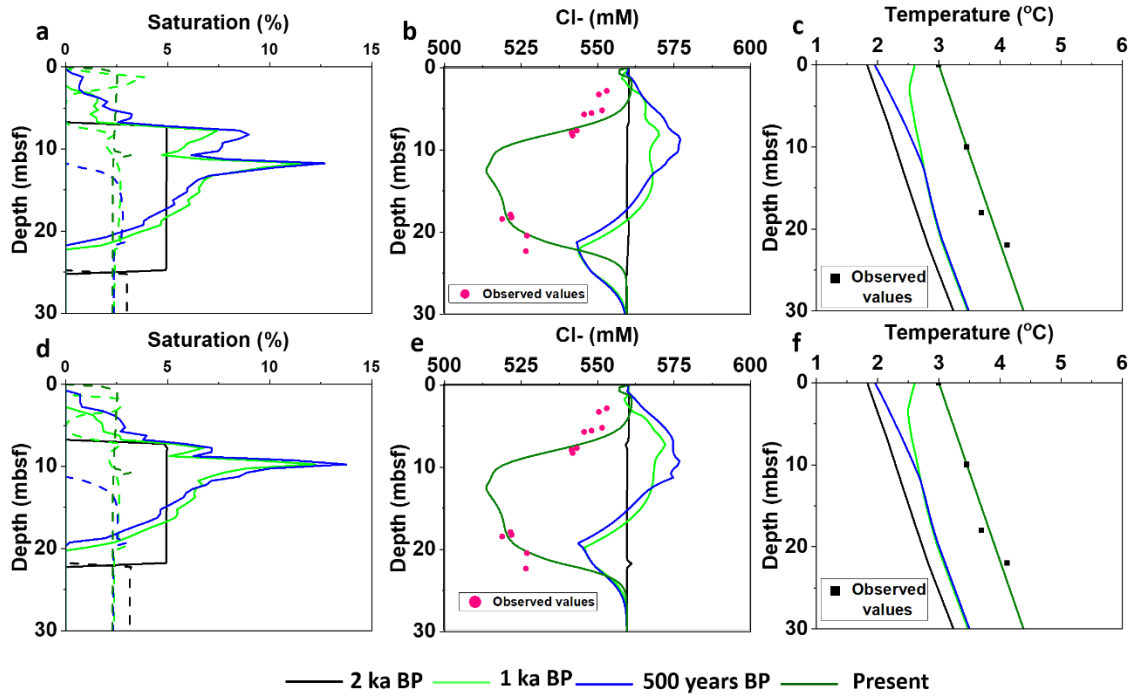


Figure S3.5. Simulation results showing methane hydrate dissociation forced by BWT_M with UiT RSL (a, b, c) and BWT_M with ICE-6G_C RSL (d, e, f) over the past 2016 years (Figures 3b and 3c). **(a)** Evolution of methane hydrate layer with 5 % initial hydrate saturation. Solid lines mark hydrate saturation and dashed lines mark gas saturation. **(b)** Chloride depletion resulting from the dissociation of the hydrate. The chloride values matched with the observations (pink dots). **(c)** Temperature profile in the sediment column. Black squares indicate temperatures measured in drill holes at 391 m water depth. **(d)** Methane hydrate layer with 5 % initial hydrate saturation. **(e)** Chloride values resulting from hydrate dissociation. The chloride values matched with the observations (pink dots). **(f)** Temperature profile in the sediment column.

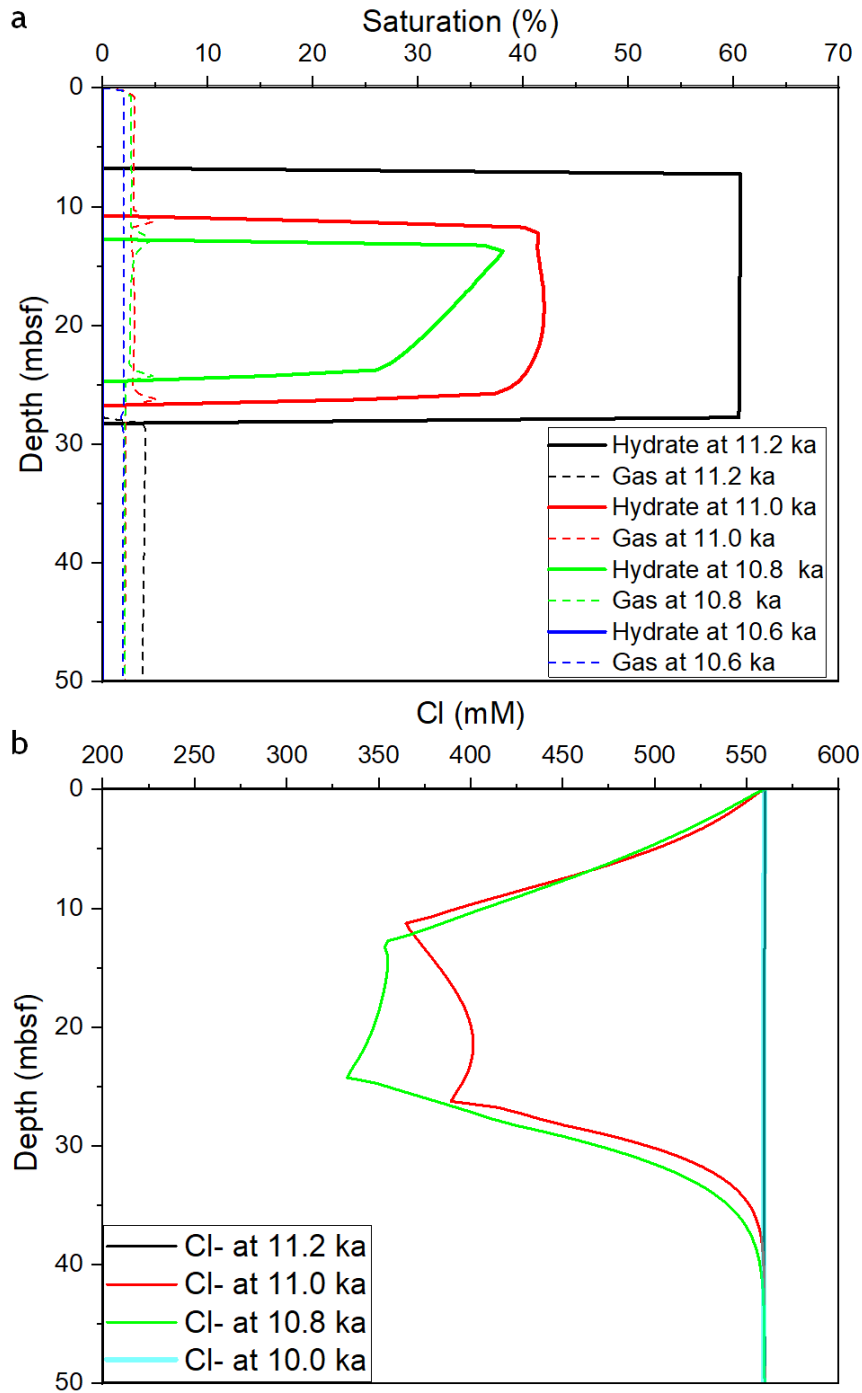


Figure S3.6. Simulation results for pressure-temperature driven methane hydrate dissociation forced by ICE-6G_C RSL and BWT_C over 11.2–10 ka. **(a)** Evolution of the hydrate layer thickness and hydrate and gas saturations over the simulation period. Hydrate completely dissociated at 10.6 ka. **(b)** Change in dissolved chloride concentration in pore water. Hydrate dissociation caused the pore water freshening and chloride depletion. The chloride anomaly diffused away at 10 ka.

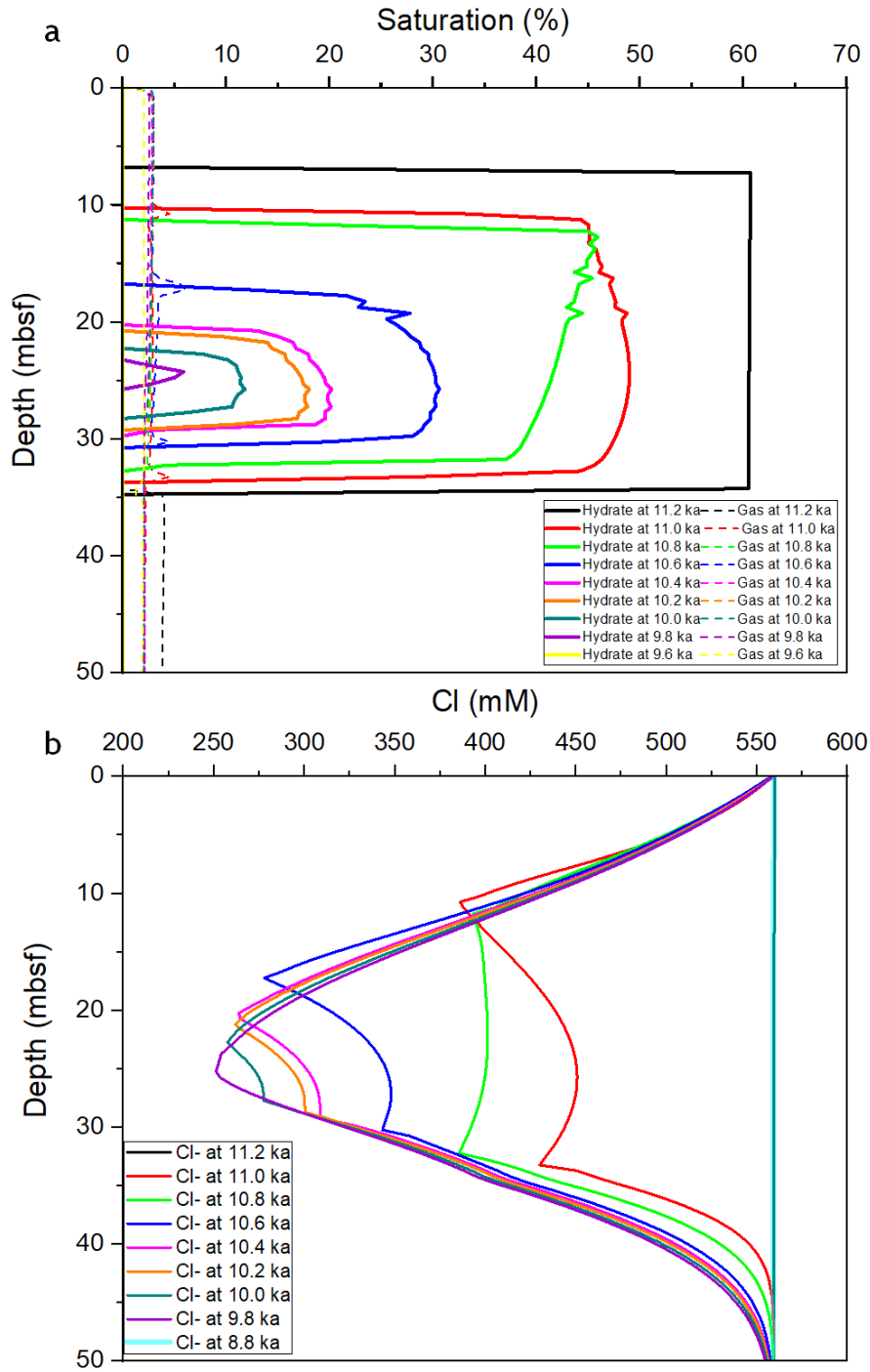


Figure S3.7. Simulation results for pressure-temperature driven methane hydrate dissociation forced by UiT RSL and BWT_C over 11.2–8.8 ka. Initial hydrate saturation was 60 %.

(a) Evolution of the hydrate layer thickness and hydrate and gas saturations over the simulation period. Hydrate completely dissociated at 9.6 ka. **(b)** Change in dissolved chloride concentration in pore water. Hydrate dissociation caused the pore water freshening. The chloride anomaly diffused away at 8.8 ka.

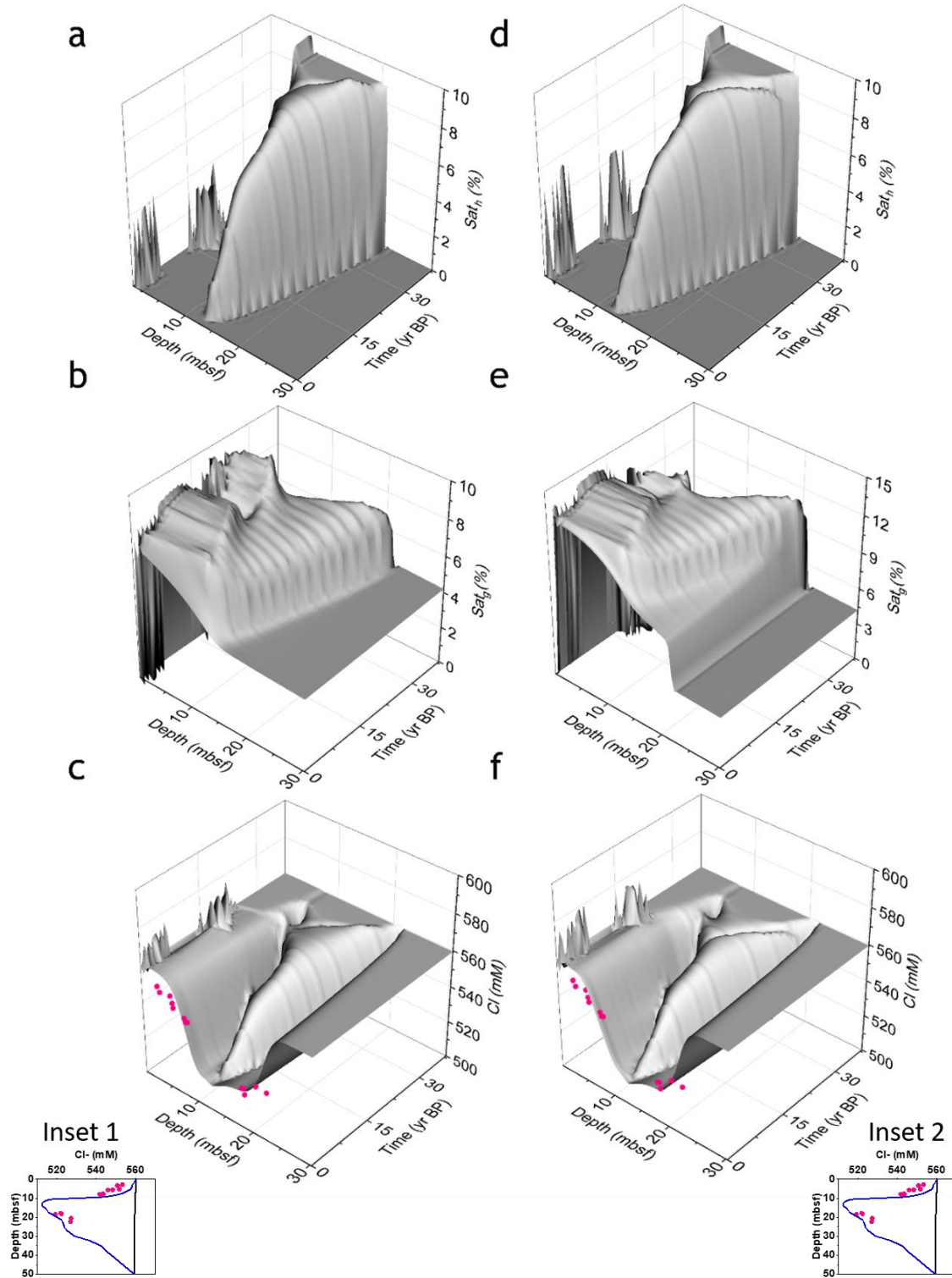


Figure S3.8. Simulation results showing the behavior of hydrate forced by BWT variations over the past 38 years (Figure 3.7b) in a high permeability (10^{-11} m^2 , a, b, c) and low permeability (10^{-15} m^2 , d, e, f) sediment column. The model was run over 1978–2016 (present). **(a)** Methane hydrate saturation (Sat_h) in the high permeability system. The initial hydrate saturation was 9 % (1978). Hydrate completely dissociated 2 years BP (2014). **(b)**

Methane saturation (Sat_g) in the high permeability system. **(c)** Dissolved chloride concentration in the high permeability system. (See Inset 1 for present-day chloride anomaly) **(d)** Methane hydrate saturation (Sat_h) in the low permeability system. Hydrate completely dissociated 1 year BP (2015). **(e)** Methane saturation (Sat_g) in the low permeability system. Gas saturation increased due to hydrate dissociation and is higher compared to the high permeability system. **(f)** Dissolved chloride concentration in the low permeability system. (See Inset 2 for present-day chloride anomaly).

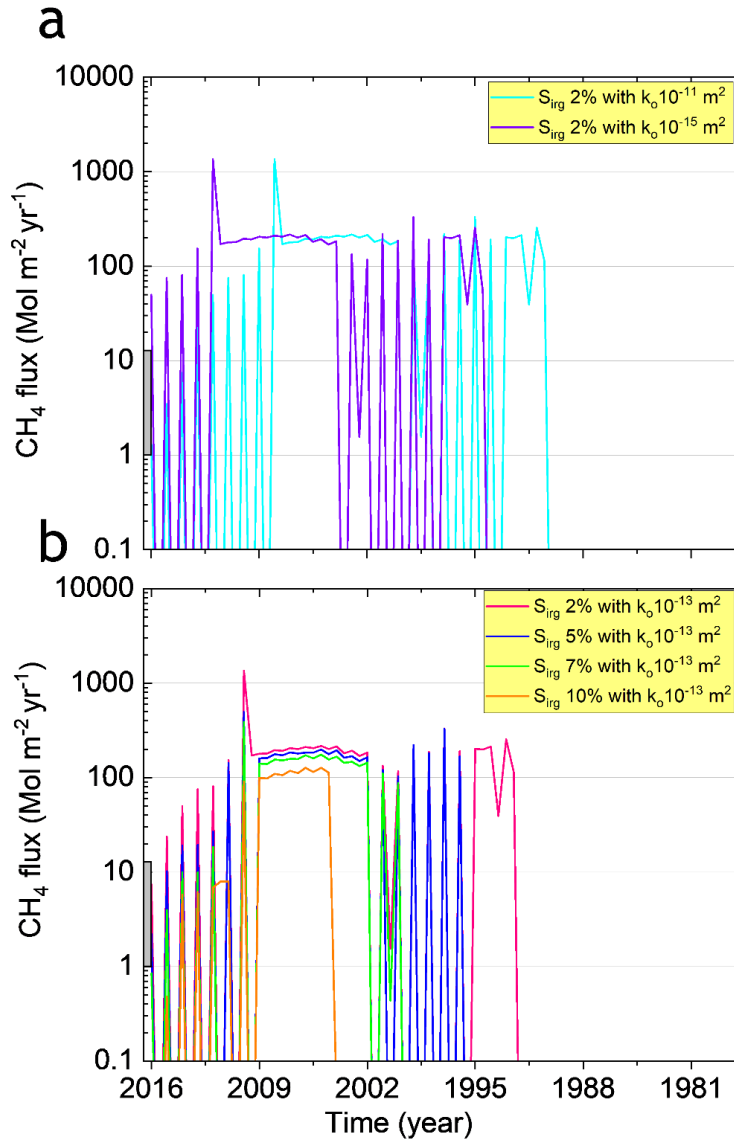


Figure S3.9. Modeled methane fluxes induced by methane hydrate dissociation for different sets of irreducible gas saturation (S_{irg}) and intrinsic permeability (k_o) values. Modeled fluxes were compared with the area-averaged range of methane gas fluxes measured at the active seep sites (grey vertical bar) (Sahling et al., 2014; Wallmann et al., 2018; Ferré et al., 2020). **(a)** Methane flux for S_{irg} values ranging 2–10 % with k_o 10^{-13} m^2 . **(b)** Methane flux in the case of 2 % S_{irg} with high and low k_o values (10^{-11} m^2 and 10^{-15} m^2 respectively).

Chapter 4

New Insights into Depositional Setup of the Interfan Area, off Svalbard

4.1. Introduction and objectives

The continental slope west of Isfjorden and Kongsfjorden cross-shelf troughs host thick, fan-like glaciogenic sediments known as the glacial Trough Mouth Fans (TMFs). These deposits resulted from paleo-ice streams that flowed along the cross-shelf troughs on the continental shelf (Figure 4.1). Slow-moving ice occupied the elevated banks separating the cross-shelf troughs (Ottesen et al., 2005). The glaciogenic sediments in front of the cross-shelf troughs serve as a record of significant sedimentation resulting from various glacial processes, including debris flows, and associated mass transport events capturing the dynamic changes in ice sheets (Ó Cofaigh et al., 2003; Vorren and Laberg, 1997).

Throughout the Pliocene and Pleistocene epochs, multiple episodes of glaciations sculpted the continental shelf of Svalbard, resulting in changes in sedimentation and erosion patterns. (Solheim et al., 1998). The Pliocene-Pleistocene glacial history of Svalbard is characterized by distinct phases of glacier development—an initial growth period (2.7–1.6 Ma) followed by an expansion phase (1.5–0.42 Ma) (Alexandropoulou et al., 2021). The western Svalbard shelf has undergone recurrent cycles of glacial waxing and waning over the past 1.3 Ma (Butt et al., 2000). Sediment core data from boreholes on the northwestern and western Svalbard continental slope (Figure 4.1) helped to establish a comprehensive chronostratigraphic framework for the margin, providing crucial insights into the glacial history spanning 2.7–0.074 Ma (Knies et al., 2009; Alexandropoulou et al., 2021). In the regional stratigraphic framework, Weichselian stratigraphic development is less understood, mainly due to inadequate age controls, creating a gap in our understanding of the Weichselian glacial history.

Although there is a consensus that multiple shelf-edge glacial advancements occurred during the Weichselian, uncertainties exist regarding the exact number and timing of these advancements. According to onshore coastal records, it has been hypothesized that during the Weichselian glacial period, approximately 118,000 to 11,600 years ago, the western Svalbard continental margin experienced at least three separate instances of shelf-edge glaciations (Mangerud & Svendsen 1992; Mangerud et al. 1998). The ice sheet attained its maximum extent during 118–108 ka, 75–50 ka, and 32–20.5 ka (Elverhøi et al. 1995; Mangerud et al. 1998; Ingólfsson and Landvik, 2013). This suggests that the glacial maximum during Marine Isotope Stage 4 (75–50 ka) lasted approximately twice as long as both Marine Isotope Stage 5d (118–108 ka) and the Last Glacial Maximum (LGM), with each spanning approximately 10,000 years. Recent investigations of terrestrial deposits in the

Kongsfjorden area have challenged prior assumptions about the frequency of ice advancements in this region. Eccleshall et al. (2016), proposed that the expansion of ice sheets to the shelf break occurred in Marine Isotope Stage 5b (93–83 ka). However, their model differs from that of Mangerud et al. (1998), as it proposes that glaciers did not extend to the shelf break during Marine Isotope Stage 4 (71–57 ka). Alexanderson et al. (2018) suggested highly fluctuating Svalbard ice sheet configuration during the Weichselian with higher frequency of ice sheet advancements. Peak advancements are estimated to have occurred at approximately 110000, 90000, 70000, 45000, and 25000 years before the present (cal. ka BP) (Alexanderson et al., 2018).

The records of glaciations and deglaciations onshore and on the shelf are affected by glacial erosion, posing challenges to the reconstruction of glacial-deglacial cycles. The well-preserved depositional records in TMFs provide more complete records of the ice sheet activity. High-resolution seismic data, capable of accurately resolving stratigraphic units, combined with improved age controls, is required to enhance the reconstruction of the Weichselian glacial history. Wiberg et al. (2022), improved the understanding of Weichselian glacial history of the Kongsfjorden TMF (KTMF) using high-resolution TOPAS sub-bottom profiler data and ages from a sediment core collected from the KTMF region. However, the knowledge of glacial evolution and depositional processes across the glaciated continental shelf and the upper continental slope in the interfan area between the KTMF and Isfjorden TMF (ITMF), west of Prins Karl Forland (Figure 4.1) is still lacking due to limited high-resolution seismic surveys, preventing correlation between the glacial fan (KTMF) and interfan between KTMF and ITMF. In the present work, I aim to develop the Weichselian seismic stratigraphic framework of the interfan area between KTMF and ITMF by integrating the age model with 2D multichannel seismic data and high-resolution SYSIF seismic data from JR269A cruise, thereby improving our understanding of the glaciomarine environment and depositional processes at the continental slope in the interfan region.

4.2. Study Area: Geological background, brief glacial history and geomorphological setting

The western Svalbard continental margin evolved as a result of a series of tectonic episodes. During the Early Eocene, Svalbard began separating from northeast Greenland as a result of right lateral strike-slip movement along the West Spitsbergen shear zone. The strike-slip movement was followed by a transpressive deformation phase that is referred to as the West

Svalbard Orogeny, resulting in the formation of a foreland fold-and-thrust belt (Harland et al., 1997). The western Spitsbergen thrust belt encompasses the Prins Karl Forland and the western Svalbard coast. The Forlandsundet Graben separates the Svalbard west coast from the Prins Karl Forland (Ritzmann et al., 2004).

During the Middle Eocene to Early Oligocene, a shift from transpression to transtension instigated oblique extension. Seafloor spreading connected the Arctic Gakkel Ridge with the Mohns and Knipovich Ridges, led to the opening of the Fram Strait oceanographic gateway allowing communication between the North Atlantic and the Arctic Ocean (Ritzmann & Jokat, 2003). During the Neogene, westward flowing paleo-ice streams released large masses of glacial sediments, called Trough mouth fans (TMFs), in front of cross-shelf troughs (Figure 4.1). These TMFs left distinct geomorphic features on the slope bathymetry, appearing as bathymetric bulges in front of the cross-shelf troughs (Figure 4.1). The TMFs are mainly built by stacked glacial debris flow lobes (Laberg and Vorren, 1995; Hjelstuen et al., 1996).

The onset of the Sjubrebanken TMF development provides evidence of the first shelf-edge glaciations in northwestern Svalbard around 2.58 Ma (Sarkar et al., 2011; Alexandropoulou et al., 2021) as a result of early Northern Hemispheric Glaciation (Jansen and Sjøholm, 1991; Mudelsee and Raymo, 2005), although glaciers didn't reach the western Svalbard shelf break during that time (Sarkar et al., 2011; Mattingdal et al., 2014). The presence of the ice-rafted debris from ODP site 986 (located between the Bellsund TMF and Isfjorden TMF) after 1.6 Ma indicates shelf-edge glaciations in western Svalbard post 1.6 Ma (Butt et al., 2000). During 1.5-1.2 Ma, glaciers covered a wider Svalbard and Barents Sea region (Alexandropoulou et al., 2021). A prominent and extensive glacial expansion occurred in the Svalbard and Barents Sea regions ~ 1.0 to 0.94 Ma, corresponding to a global increase in ice volume, during the Mid-Pleistocene revolution (Hjelstuen et al., 2007; Mudelsee and Stattegger, 1997). The incision of the Kongsfjorden Trough, initiated sometime between 1.5 and ~ 1 Ma by ice stream activity, lead to the formation of the KTMF (Sarkar et al., 2011).

Shelf geomorphology:

The Kongsfjorden cross-shelf trough extends from east to west. It was formed by the merging of two fjords, Krossfjorden and Kongsfjorden, in the eastern part. The U-shaped configuration of the trough is a characteristic feature resulting from glacial erosion caused by a paleo-ice stream. The floor of the trough is marked by 9–10 km long and ~ 10 m high

glacial lineations, which are a result of soft sediment deformation. The lineations align parallel to the paleo-ice flow directions (Ottesen et al., 2007; Sarkar et al., 2011).

The seabed topography of the shelf surrounding the trough exhibits various large (~ 25 m high and ~ 500 m wide) and small ridges paralleling the shelf break. They are inferred as recessional moraines formed during glacier retreat. Large ridges are associated with more extended still-stands during the glacier recession. Ridges are seen on the Kongsfjorden cross-shelf trough and are known as grounding zone wedges, indicating the halting of the ice front during ice retreat (Howe et al., 2003; Dowdeswell, 2016).

Slope geomorphology:

In the upper continental slope, there are slope-parallel longitudinal ridges, occurring at present-day water depths 750-2000 m (Sarkar et al., 2011). They have a NW-SE orientation and exhibit a moderately sinuous shape. Some of them can be tracked for up to 12 km before fading into the surrounding seafloor topography and are inferred as contourite ridges (Sarkar et al., 2011). On the slope, curvilinear V-shaped grooves, approximately 20 to 30 m in depth, run parallel to the slope. These grooves are roughly 15 km long and are interpreted as ploughmarks caused by deeper keels of icebergs (Vanneste et al., 2007; Zhao et al., 2017).

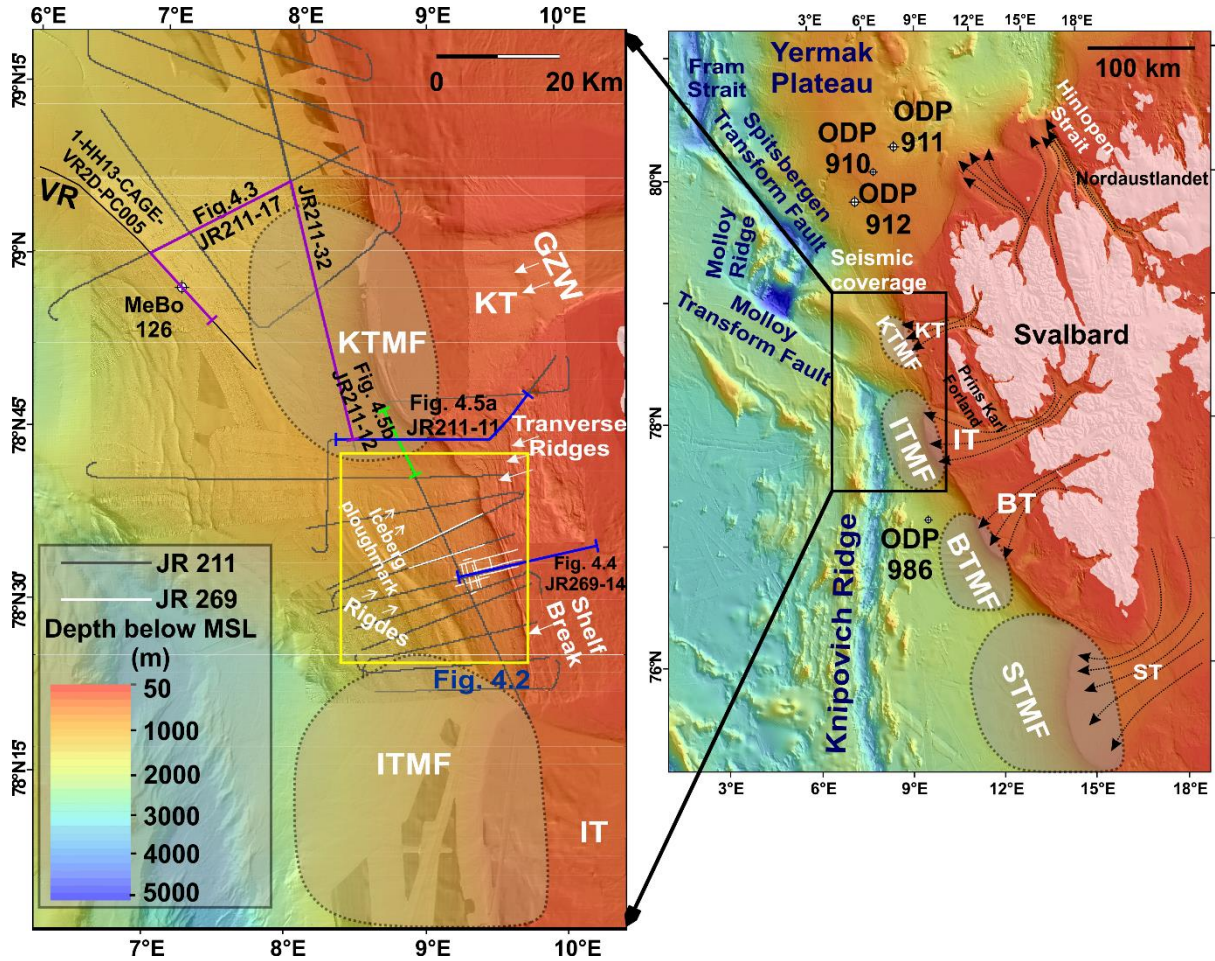


Figure 4.1. Study area map showing the location of 2D air gun seismic lines. The left bathymetry image shows the mid-oceanic ridge and transform fault system west off Svalbard, boreholes on the Yermak Plateau and western Svalbard. Glacial cross-shelf troughs (Kongsfjorden Trough (KT), Isfjorden Trough (IT), Belsund Trough (BT), Storfjorden Trough (ST)) and respective trough mouth Fans (TMF) were formed by paleo-ice streams. The right image displays 2D seismic lines from the JR211 and JR269B cruises. Moraines on the shelf, including a Grounding Zone Wedge (GZW) and N-S trending transverse ridges, indicate glacial retreat post Last Glacial Maximum. V-shaped grooves on the upper slope denote iceberg ploughmarks from the last glaciation, while NNW–SSE wavy ridges are contourites (Sarkar et al., 2011). The bathymetry is a compilation of multiple cruises, such as JR211, and surveys conducted by the Norwegian Hydrographic Survey (Sarkar et al., 2012).

4.3. Data and Methods

4.3.1. High-resolution SYSIF Data

During the JR269A cruise, a total of 125 km of deep-towed SYSIF seismic lines (Figure 4.2) were acquired from the upper continental slope of the interfan region between ITMF and

KTMF in water depths ranging from 250 to 875 m (Ker et al., 2014). A Janus Helmholtz transducer was used as the SYSIF seismic source, which transmitted a chirp signal in the high-frequency bandwidth (220–1050 Hz). The receiver was a single hydrophone with a 10 m offset from the source. The processing workflow of SYSIF data consists of (i) performing a source signature deconvolution with the far-field recording at normal incidence to increase the vertical resolution (Ker et al., 2010), (ii) applying an immersion correction as the altitude of SYSIF is towed at a height of 100 m above the seabed, and (iii) migration with the water velocity. The lateral and vertical resolution of the SYSIF lines are 3 m and 1 m respectively.

4.3.2. JR269B Multichannel seismic reflection data

Multichannel seismic reflection (MCS) data were collected on the shelf and slope of the interfan area (Figure 4.2) during the JR269B cruise. The seismic survey employed a 60-m streamer with 60 one-m hydrophone groups. The seismic source was a 1.46 l GI airgun comprising a 0.73 l-generator and a 0.73 l-injector. The airgun was towed at 1.5 m and operated in harmonic mode at a nominal pressure of 140 bar (2000 psi). The shot interval was maintained at 6 s corresponding to 12 m shot spacing. The duration of the recording of seismic trace was 3 seconds, with a sampling rate of 0.5 milliseconds.

The processing steps of MCS data involved (i) merging marine geometry to the raw shots, (ii) common midpoint (CMP) sorting with the spacing between individual CMP bins at 2 m, (iii) Ormsby bandpass filtering that is characterized by specific corner frequencies set at 10 Hz, 30 Hz, 600 Hz, and 1000 Hz; (iv) normal move-out correction, (v) stacking and (vi) post-stack migration (Stolt f-k).

4.3.3. Ancillary seismic lines

Apart from the seismic lines from the JR269A and JR269B cruises, this study includes 2D MCS lines obtained during the JR211 cruise (Sarkar et al., 2011). These lines were collected using a 96-channel streamer with an active length of 600 m and a GI air gun with a volume of 2.45 l.

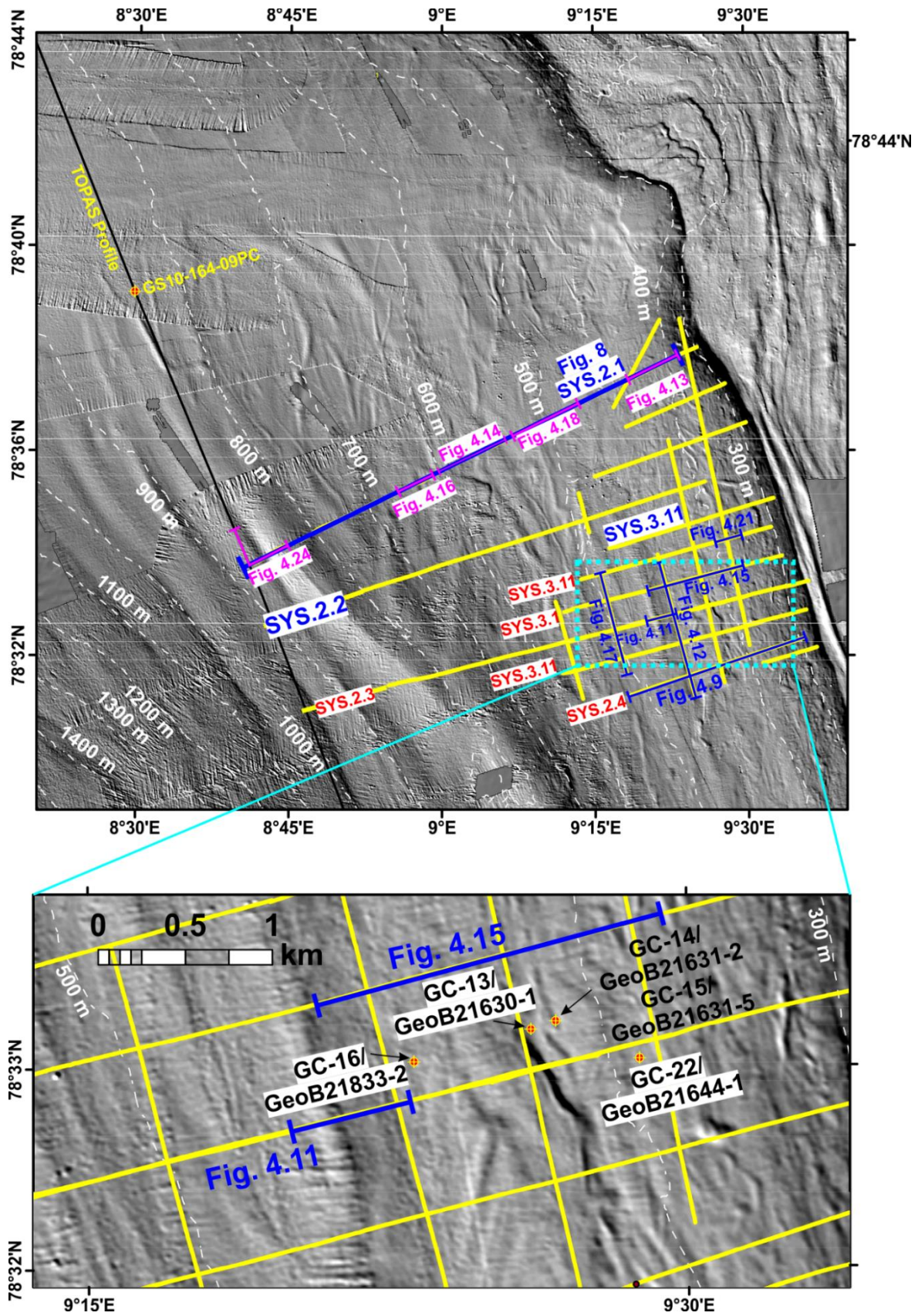


Figure 4.2. SYSIF seismic lines in the interfan region. The enlarged figure below shows the location of various gravity cores during the cruise MSM57 (Bohrmann et al., 2017).

4.3.4. Debris flow modelling with RAMMS

In this study, we employed Rapid Mass Movement Simulation (RAMMS) debris flow numerical modelling to gain insights into the behaviour, and runout characteristics of debris flows. Our aim was to simulate debris flow dynamics, enabling us to correlate model-predicted results with seismic observations of debris flow. To constrain the debris flow model, we took into account the shape, dimensions (length along slope-perpendicular direction and thickness) of debris lobes, and runout distance from the shelf break, considering the release location as the edge of a submarine till-wedge.

The modelling is based on the Voellmy friction law (Salm et al. 1990; Salm, 1993). The Voellmy equation includes a yield stress (cohesion) (Bartelt et al. 2015). RAMMS uses a shallow water equation, based on conservation of mass and momentum (Christen et al., 2010). The frictional resistance S (Pa) considers the frictional coefficient μ representing Coulomb friction and coefficient ξ representing turbulence friction and these parameters control the flow behaviour (Christen et al., 2010). The parameter S is given by equation 1.

$$S = \mu N + \frac{\rho g u^2}{\xi} + (1 - \mu)N_0 + (1 - \mu)N_0 e^{-\frac{N}{N_0}} \quad (1)$$

In this equation, N_0 is the yield stress (Pa), ρ is the density (kg/m^3), g is the gravitational acceleration (m/s^2), u represents x- and y-components of velocity (m/s). N represents the normal stress (Pa). The coefficients μ (unit less) and ξ (m/s^2) are kept constant throughout one simulation. The goal of debris flow modelling is to identify flow patterns for a predefined set of debris flow rheological properties and does not aim for an exhaustive treatment of all debris flow intricacies, such as debris flow transforming into turbidity. The simulations do not account for erosion and sediment entrainment models, focusing on the visco-plastic behaviour of debris flow, considering both cohesion and friction. Modelling parameters used in RAMMS are provided in Table 4.1.

Table 4.1. RAMMS parameters used for simulation

Voellmy-friction (Range for simulation)	
Friction coefficient (μ) (unitless)	0.01-0.04
Turbulence coefficient (ξ) (m/s^2)	150-200
Voellmy-cohesion (Range for simulation)	
Yield strength (Pa)	2000, 100

4.3.4 Previous and updated regional seismic-stratigraphic correlation:

Sarkar et al. (2011) developed the seismic stratigraphic framework of the KTMF, and the interfan region between KTMF and ITMF using six seismic horizons (A1-A6). However, their chronostratigraphic framework lacked comprehensive age controls for these horizons. To address this limitation, an updated seismic stratigraphy with improved age controls for both the KTMF and the interfan region was established. This involved tracing the regional seismic horizons (X1, X2, R1-R4, R4 A, and R5-R7) along a composite seismic line connecting the MeBo 126 drill site to the KTMF shelf regions and upper continental slope and the interfan region between KTMF and ITMF (Figure 4.1). Table 4.2 provides an overview of the regional seismic reflectors and their corresponding age constraints in the western Svalbard slope. The regional horizons R7, R4 and R2 could be tracked from the continental slope to the outer shelf, while the tracking of other reflectors in the shelf region is difficult due to poor reflection continuity.

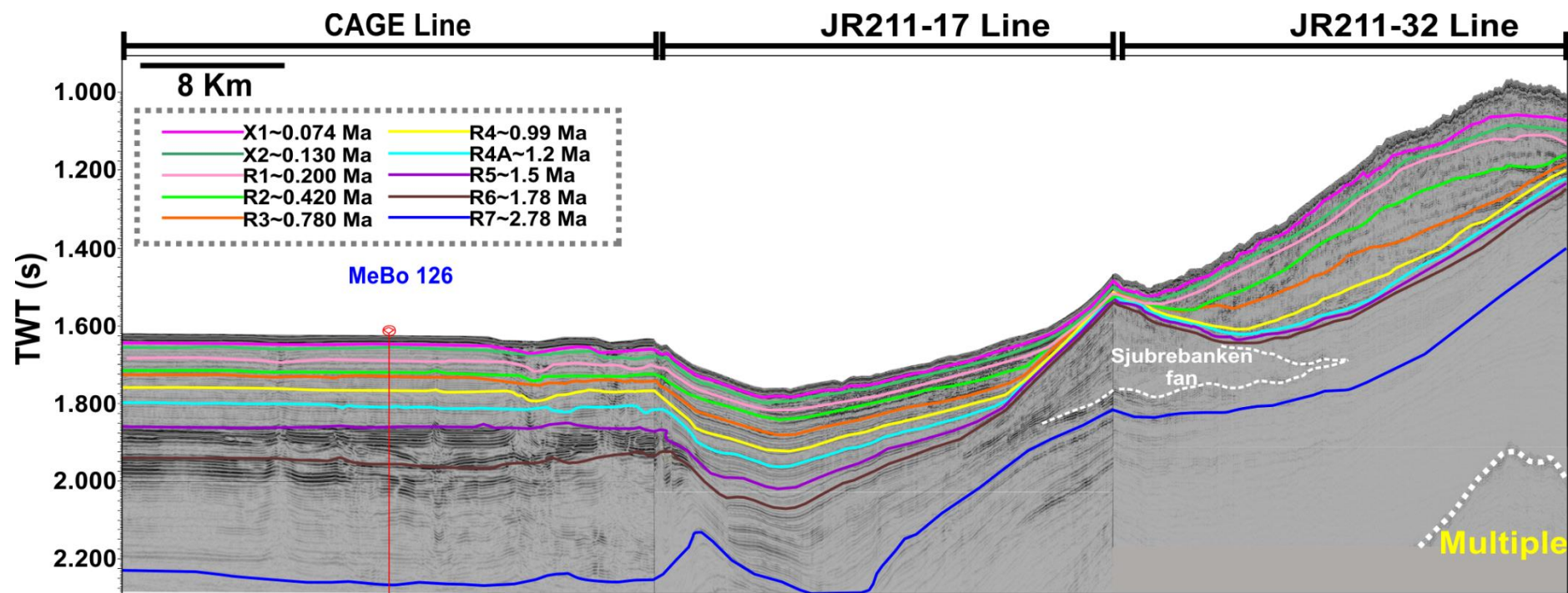


Figure 4.3. Composite seismic profile in the KTMF region showing the updated seismic stratigraphy (line location in Figure 4.1). Reflectors X1, X2, R1–R7 are tracked along the CAGE line (courtesy CAGE) and the JR211 seismic lines (Sarkar et al., 2011).

Table 4.2. Regional seismic reflectors used for stratigraphic correlation off west Svalbard

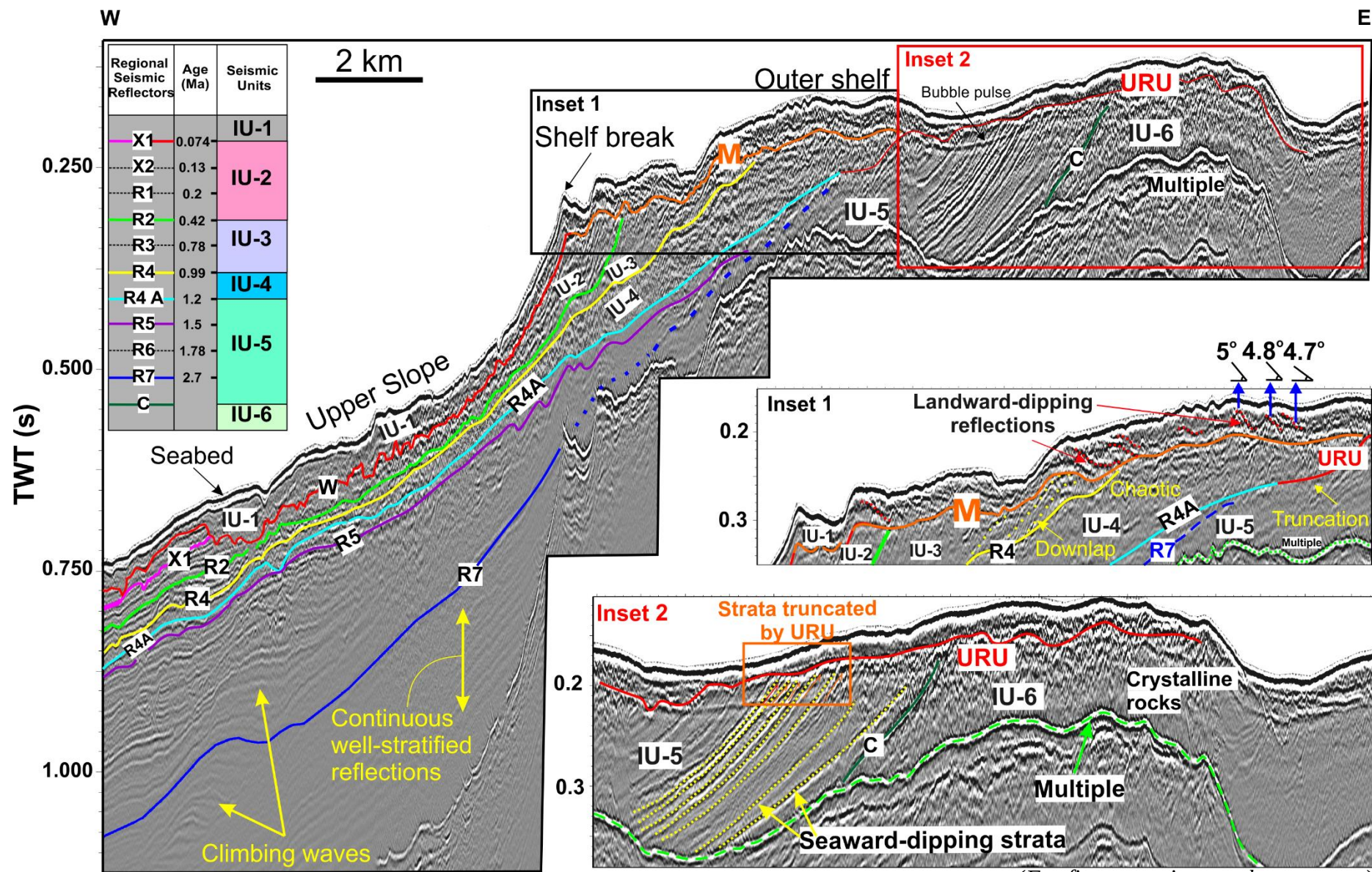
Seismic horizons	Assigned age (Ma)	TWT (ms)	Site	Reference
X1	0.074	1647	MeBo 126	Dessandier et al. (2021)
X2	0.13	1654	MeBo 126	Dessandier et al. (2021)
R1	0.2	1688	ODP 986	Faleide et al. (1996); Fiedler and Faleide (1996)
R2	0.424	1717	MeBo 126	Dessandier et al. (2021)
R3	0.78	1735	ODP 910A, 911A, 912A, 986	Channell et al. (1999); Knies et al. (2007); Myhre et al. (1995)
R4	0.99	1764	ODP 986, 912A	Myhre et al. (1995); Channell et al. (1999)
R4A	1.2	1822	ODP 912A, 910A, 910C	Myhre et al. (1995); Smelror (1999); Knies et al. (2007)
R5	1.5	1867		Faleide et al. (1996)
R6	1.78	1939	ODP 911A, 912A	Myhre et al. (1995)
R7	2.7	2268	ODP 911A, 910A/C, 986	Sato and Kameo (1996); Smelror (1999)

4.4. Results

The seismic stratigraphic description of the interfan region encompasses the continental shelf and continental slope regions. In the present work, the seismic stratigraphy of the continental shelf region is explained by JR269B MCS seismic lines, while the continental slope stratigraphy is established with the help of high-resolution SYSIF lines. Uninterpreted seismic images can be found in the Appendix section (Page no. 227).

4.4.1 Continental shelf stratigraphy of the Interfan area based on 2D MCS lines:

A prominent erosional unconformity identified as the Upper Regional Unconformity (URU) serves as the erosional base for a major glacial advancement in the continental shelf, and it is correlated with the reflector R4A in the interfan region between the Kongsfjorden and Isfjorden Troughs. This correlation suggests that significant erosion, reaching the URU level, happened around 1.2 Ma (corresponding to the age of the R4A reflector, Table 4.2). The URU dips westward at an angle of 0.3-0.9 degrees in the outer shelf. The geological succession below the URU includes seismic units IU-5–6, while the overlying succession is represented by units IU-1–4.



(For figure caption see the next page)

Figure 4.4. Seismic line JR269-14 (location in Figure 4.1) showing the seismic units in the interfan area. Seismic unit IU-6 in the shelf shows a nearly transparent region that likely represents crystalline rocks and IU-5 shows seaward-dipping strata. The Upper Regional Unconformity (URU) truncated the top of IU-5 and merged with the slope reflector R4A (Insets 1 and 2). Seismic units IU-4–IU-1 in the outer shelf region consist of westward-dipping oblique clinoforms. Moderate-amplitude reflections enclose chaotic, transparent, and incoherent reflections, indicating seaward progradation (IU-4–IU-1). Unit IU-1 shows deformed till with landward-dipping reflections on the outer shelf (Inset 1). The prograding clinoforms at the shelf break and uppermost slope indicate net deposition of glaciogenic debris.

4.4.1.1. Pre-URU geological successions (IU-6 and IU-5):

The pre-URU stratigraphic successions, IU-6 and IU-5 are separated by reflector C. Unit IU-6 exhibits a nearly transparent character with weak internal reflections, suggesting the presence of crystalline rocks and the weak reflectivity is presumably due to hard rocks. IU-5 is represented by stratified and westward dipping reflection patterns (Inset 2, Figure 4.4). The top of unit IU-5 is truncated by the erosional surface (URU). The shift from chaotic, transparent facies in IU-6 to a stratified sedimentary sequence in IU-5 is inferred to represent a transition from crystalline basement to marine sedimentary layers.

4.4.1.2. Post-URU geological successions (IU-4, IU-3, IU-2 and IU-1):

The geological succession overlying URU is represented by a stack of seaward advancing wedge-shaped seismic units. The post-URU geological succession is classified into four units: IU-4, delineated by R4A at the base and R4 at the top; IU-3, marked by R4 at the base and R2 at the top; IU-2, defined by R2 at the base and W at the top; and the topmost unit, IU-1, bounded by W at the base and the seabed.

Unit IU-4 predominantly consists of wedge-shaped chaotic facies within the outer shelf region of the interfan area (Figure 4.4). The lateral expansion of this unit is represented by the seaward thickening of individual wedge-shaped chaotic packages. These chaotic reflections are interbedded with oblique westward dipping (3 to 6 degrees) downlapping reflectors, spanning an approximate distance of ~ 2 km in the east-west direction. Reflector M on the outer shelf truncates the upper part of unit IU-4. The chaotic wedge facies is interpreted as a glaciogenic prograding wedge. The transparent and structureless nature suggests the presence of glaciogenic debris, while the high amplitude reflections indicate a transition from debris

flow to the settling of hemipelagic sediments. In contrast to the interfan region, the IU-4 demarcates the onset of glacial TMF development in the KTMF region (Figures 4.5a and 4.5b).

Unit IU-3 reveals seaward prograding clinoforms that extend from the outer shelf to the upper slope, with their top truncated by reflector M (Figure 4.4).

Unit IU-2, in the shelf break and upper slope, consists of prograding clinoforms characterized by high-amplitude reflections separating chaotic regions. They extend basinward, gradually thinning downslope and are interpreted as glaciogenic prograding sequence. Sarkar et al. (2011) observed and compared the extent of prograding wedge facies in the interfan area and the KTMF. They noted that the interfan region underwent moderate shelf progradation (~ 5-7 km), suggesting a depocenter fed by slower-moving ice sheets. In contrast, the KTMF exhibited more extensive prograding packages, extending approximately 30 km in an E-W direction on the upper slope, indicating significant glaciogenic deposition from ice streams. The erosional reflector M in the outer shelf correlates with the slope reflector W in the slope. Reflector M diverges toward the basin from the URU.

Unit IU-1 shows prograding clinoforms in the shelf break and upper slope (Figure 4.4). In the outer shelf region, IU-1 displays high-amplitude, undulating reflections characterized by poor reflection continuity. They represent deposits affected by glacial tectonics, but managed to survive glacial erosion. Additionally, moraine complexes were deposited after the Last Glacial Maximum, marking periods of standstills during the ice sheet retreat.

The URU shows short wavelength undulations, such as upward convex morphology above the crystalline rocks (Figure 4.4, Inset 2). Further westward, the URU is concave, and truncates underlying well-stratified beds. There is a net glaciogenic sediment accumulation, shelf progradation, and aggradation basinward of the concave segment of the URU. The slope reflector R4A correlates with the URU indicating that R4A represents a correlative conformity, marking the distal extension of the URU. The presence of gently seaward dipping sedimentary sequences below the URU and the curvature on the URU likely indicate enhanced erosion of softer marine beds and westward subsidence about a convex hinge zone. Differential subsidence plays an important role in the accumulation of the sedimentary wedge along the margin. Several factors contribute to this subsidence, including compaction of sediments deposited before the URU/R4A boundary, and glacial loading (Løtveit et al., 2019). The seismo-stratigraphic correlation, which is based on regional reflectors (R1-R7)

and interfan reflectors, is detailed in Table 4.3 and provides the age of the interfan seismic units. Since the air gun seismic data could not resolve the younger events that occurred between the present and 75,000 years ago, high-resolution SYSIF data was employed. Table 4.3 serves as a bridge, integrating the air gun and SYSIF seismic data with age information, providing a comprehensive understanding of the stratigraphy.

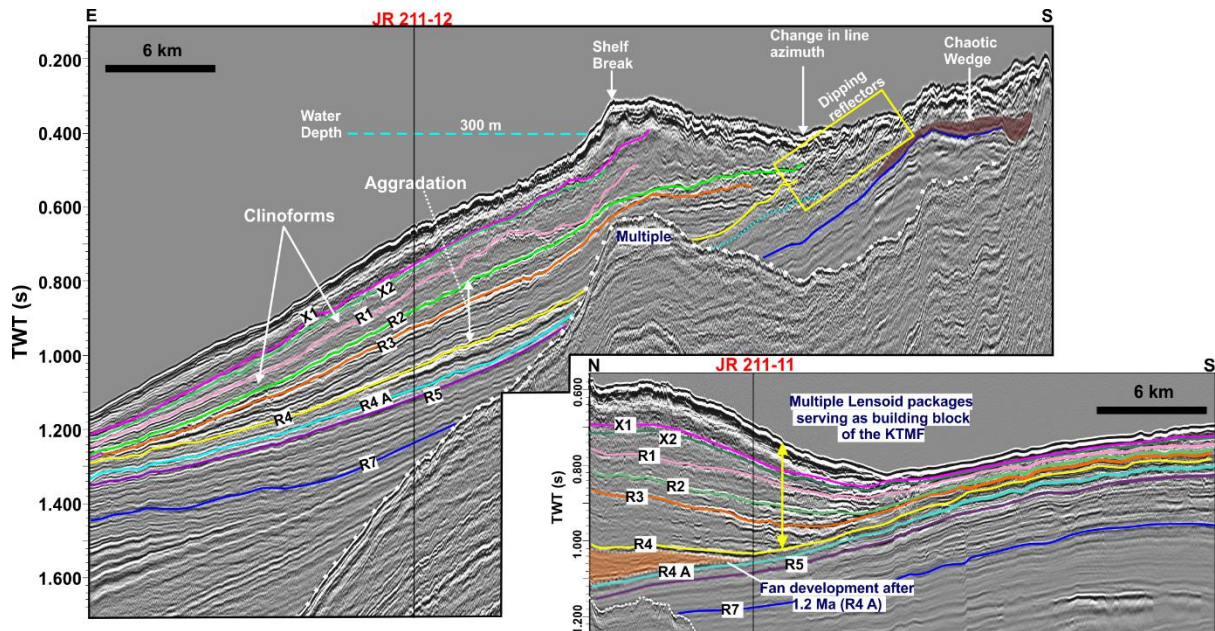
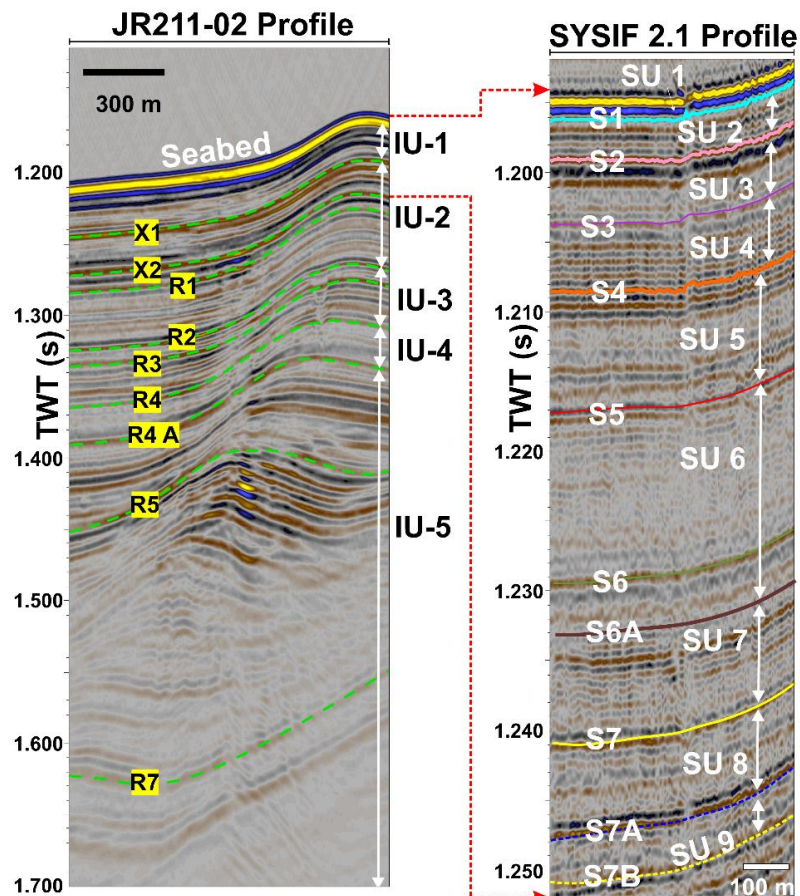


Figure 4.5. Seismic lines JR211-11 and JR211-12 showing the internal structure of the KTMF. **(a)** Line JR211-11 showing large clinoforms formed during aggradation and progradation above R4A (~ 1.2 Ma) demarcating the initiation of Kongsfjorden ice stream activity. **(b)** JR211-12 line reveals the initiation of KTMF. The lensoid package above reflector R4A signifies the commencement of ice stream activity in western Svalbard. Debris lobes pinch into stratified reflectors in the interfan region. The seismic horizons are updated after the initial work by Sarkar et al. (2011). Line locations are shown in Figure 4.1.

Table 4.3. Seismic stratigraphic framework utilizing air gun seismic lines and SYSIF lines. The framework includes seismic units (IU-1–IU-5), regional reflectors (R1–R7) based on air gun seismic data and interfan seismic units (SU 1–SU 9) and reflectors (S1–S7) based on SYSIF data.

Seismic Units (Based on SBIS stratigraphic correlation)	Regional Seismic Reflectors	Age (Ma)	Seismic Units Air gun data (This study)	Present-Day Seabed	SYSIF Seismic Reflectors	Age (ka)	Seismic Units (SYSIF data)
GIII	X1	.074	IU-1	→	S1	10	SU 1
	X2	0.13	IU-2		S2	20	SU 2
	R1	0.2			S3	30	SU 3
GII	R2	0.42			S4	40	SU 4
	R3	0.78	IU-3		S5	50	SU 5
	R4	0.99	IU-4		S6	60	SU 6
	R4 A	1.2			S6A	70	SU 7
	R5	1.5	IU-5		S7	80	SU 8
GI	R6	1.78			S7A	90	SU 9
	R7	2.7			S7B	100	
	C		IU-6			110	
						120	



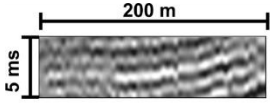
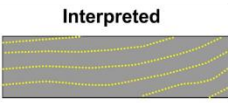
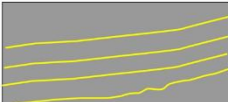
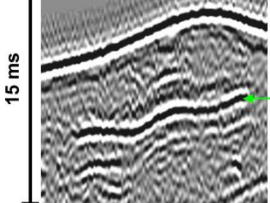
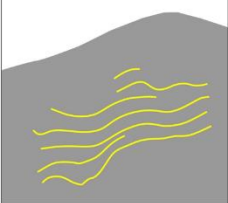
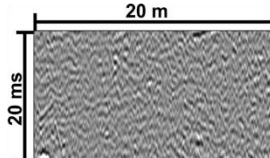
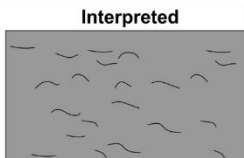
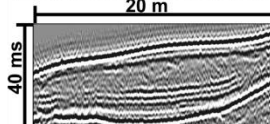
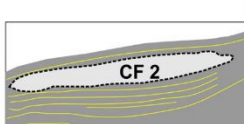
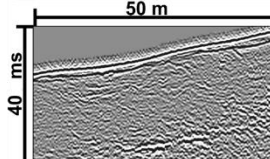
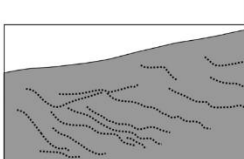
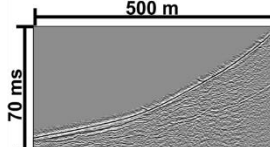
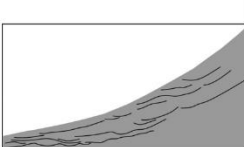
4.4.2. Continental slope stratigraphy of the Interfan area based on high-resolution SYSIF data:

Seismic stratigraphy for the upper continental slope is established based on SYSIF seismic data. The seismic stratigraphy comprises a seismic facies description in Table 4.4, followed by a description of the seismic units, their interpretation, and key bounding reflectors in Table 4.5. The seismic stratigraphic framework is correlated with the established framework by Wiberg et al. (2022).

4.4.2.1. SYSIF Seismic facies

The seismic facies is broadly classified into two categories: Stratified and chaotic facies. Stratified facies represents a set of parallel, continuous reflections, while chaotic facies represents a reflection package dominated by incoherent, discontinuous, semi-transparent to transparent reflections. The stratified facies is subdivided into 3 types, SF 1-3, and the chaotic facies is subdivided into 4 varieties, CF1-4 (Table 4.4).

Table 4.4. Seismic facies template based on SYSIF data for the interfan area.

Stratified Facies (SF)		Description
		SF 1: This facies is characterized by parallel to sub-parallel, low to medium amplitude reflections with a conformable top and bottom. The low amplitude indicates fine-grained deposits like plumite.
		SF 2: This facies is similar to SF 1 but exhibits medium to high amplitude reflections. The high amplitude indicates coarser sediments such as sand.
		SF 3: The facies is similar to SF 2 but exhibits semi-continuous, choppy, and isolated reflections (maximum length ~200m). They show interbedding with chaotic reflections. The medium-high amplitude reflects the deposition of coarser sediments and interpreted as isolated turbidite.
Chaotic Facies (CF)		
		CF 1: This facies exhibits low-amplitude, discontinuous reflections that are nearly transparent and structureless. They fill paleo erosional surfaces. The facies represents glaciogenic debris flows filling pre-existing topographic lows.
		CF 2: Similar to CF1 but have a lenticular/lenoid geometry. CF 2 deposits are structureless with an irregular erosional base, indicating their gravity-driven nature.
		CF 3: Similar to CF 2 but has an overall rhombic geometry, and oblique depositional succession. The boundary between stacked lobes shows a moderate amplitude. This facies represents debris lobes resting on each other, indicating multiple local episodes of debris flow deposition.
		CF 4: This facies exhibits a set of concave low to medium amplitude semi-continuous reflections. These packages represent bottom set of the S-shaped glaciogenic/prograding clinoform.

4.4.2.2. Seismic unit description from the SYSIF data on the slope

Ten seismic reflectors (Seabed, S1-S5, S6A, S7, and S7A–B) divide the SYSIF seismic lines into 9 major seismic units: SU 1 (youngest) to SU 9 (oldest) (Table 4.5 and Inset Figure 4.8). The seismic facies and thickness variations of units SU 1–5 are shown in Figures 4.6 and 4.7, respectively. The deeper seismic units (SU 6–9) are untraceable in the upper slope due to poor imaging and partly because of erosion; hence, facies maps and unit thickness maps for these units are not generated. The detailed characteristics and interpretations of each seismic unit are provided below. Table 4.3 serves as a bridge, integrating the air gun and SYSIF seismic units.

Table 4.5. SYSIF seismic units and their bounding reflectors in the interfan area.

SYSIF Seismic Units	Top reflector	Bottom reflector
SU1	Seabed	S1
SU2	S1	S2
SU3	S1/S2	S3
SU4	S3	S4
SU5	S4	S5/ES-1
SU6	S5	S6A
SU7	S6A/ES-1	S7/ES-2
SU8	S7/ES-2	S7A
SU9	S7A	ES-3/S7B

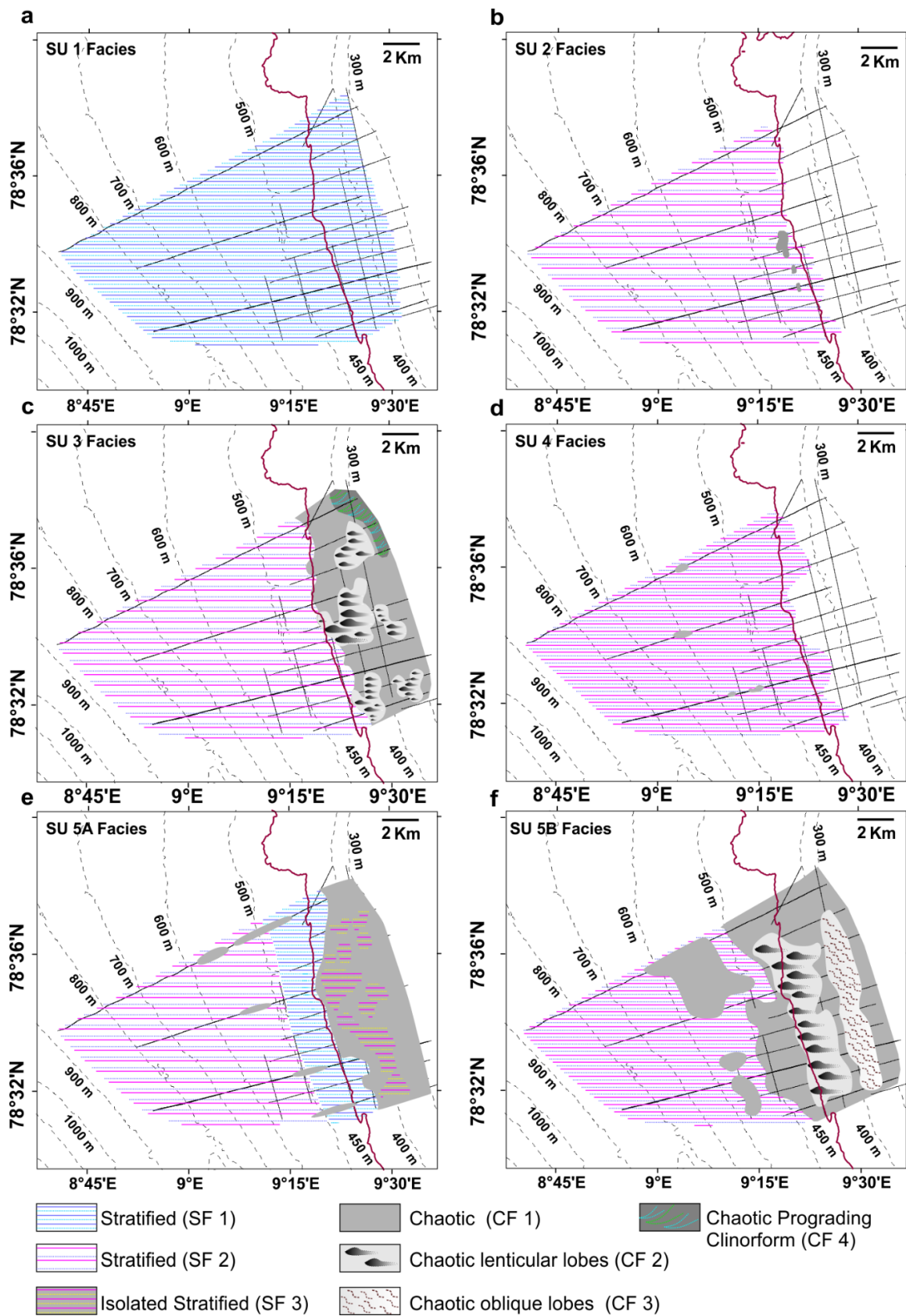


Figure 4.6. Spatial variation of different facies in stratigraphic units SU 1-5. The stratified facies (SF) are predominantly present > 450 m except in SU 1 as it covers shallower depths up to 300 m. The chaotic facies in SU 3, SU 5A, and 5B indicates glaciogenic debris flows.

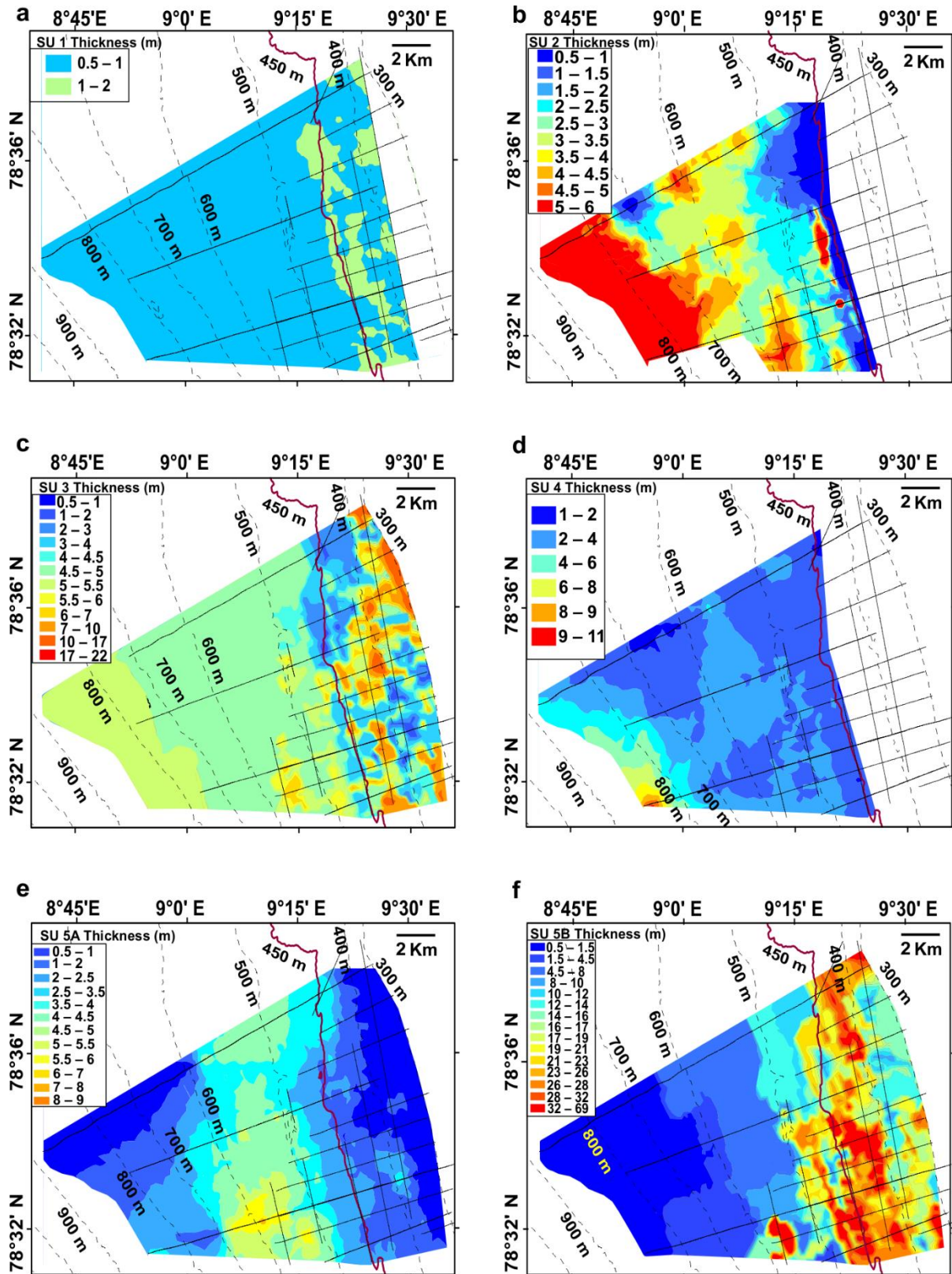
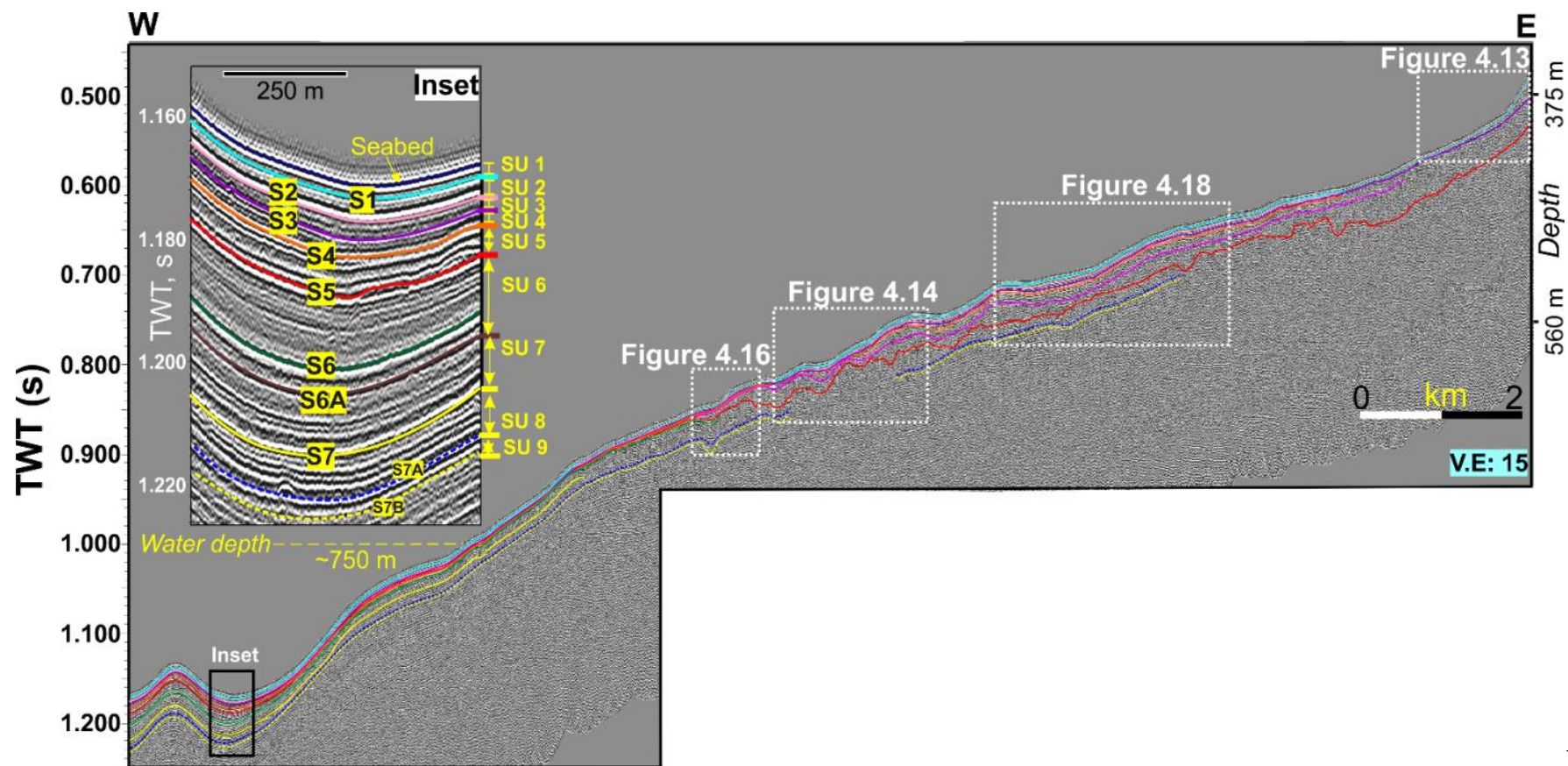


Figure 4.7. Map showing the thickness variations of stratigraphic units SU 1-5. Comparison of Figure 4.6 with this figure shows that the thickness of CF-2 glaciogenic facies in SU 5B was substantially high compared to SU 3



Figure

4.8. Slope perpendicular SYSIF seismic profile 2.1 (location shown in Figure 4.2) showing key reflectors (Seabed, S1-S7, S7A-B) and seismic units in the interfan area. The inset shows the enlarged part with the key reflectors and seismic units. The bottomsets of prograding clinoforms at the uppermost slope is bounded by S1 and S3 (enlarged in Figure 4.13). Reflectors S6 and S7 are truncated by erosional surface (ES-1) in the upper slope (enlarged in Figure 4.16).

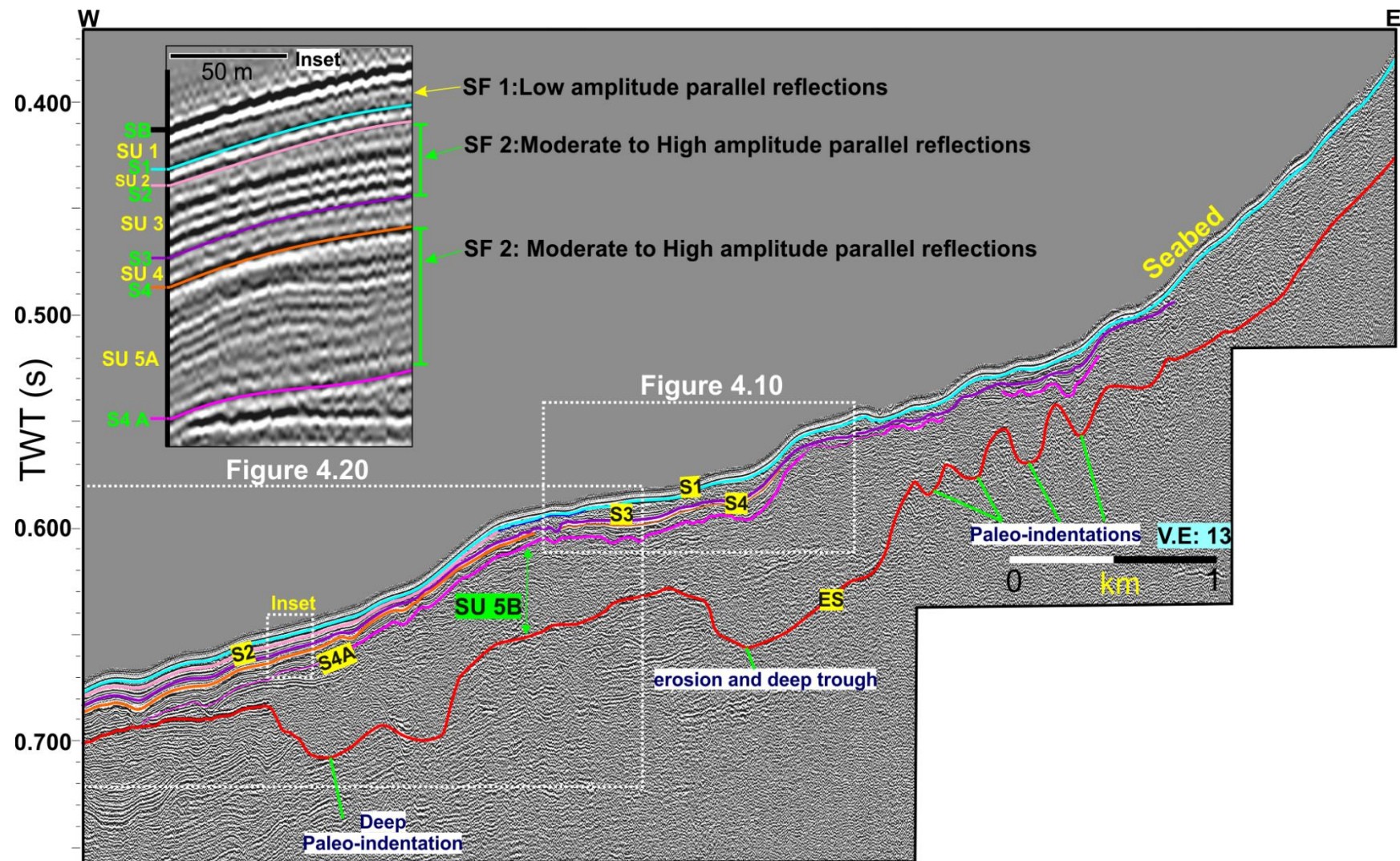


Figure 4.9. Slope perpendicular SYSIF seismic profile 2.4 showing paleo-indentations (ES-1). Seismic unit SU 5B consists of chaotic facies. Seismic line location is shown in Figure 4

Seismic Unit 1 (SU 1):

SU 1 is the topmost unit in the upper slope, bounded by the seabed at the top and reflector S1 at the bottom (Inset, Figure 4.8). It consists of low to moderate amplitude stratified facies (SF1) (Inset, Figure 4.9) that is continuous throughout the region (Figure 6a). The thickness of this unit varies from 0.5 to 2 m (Figure 4.7a). In water depths from 300 to 480 m, thicker sediments (≥ 1 m) are prevalent (Figures 4.7a and Inset Figure 4.10), while in water depths greater than 480 m, the thickness of the unit is < 1 m.

Interpretation: The low-to-moderate amplitude reflections represent hemipelagic sediments and plumites deposited from the meltwater plume. The relatively thicker areas indicate sediment accumulation in pre-existing topographic lows (Figure 4.7a and Inset Figure 4.10).

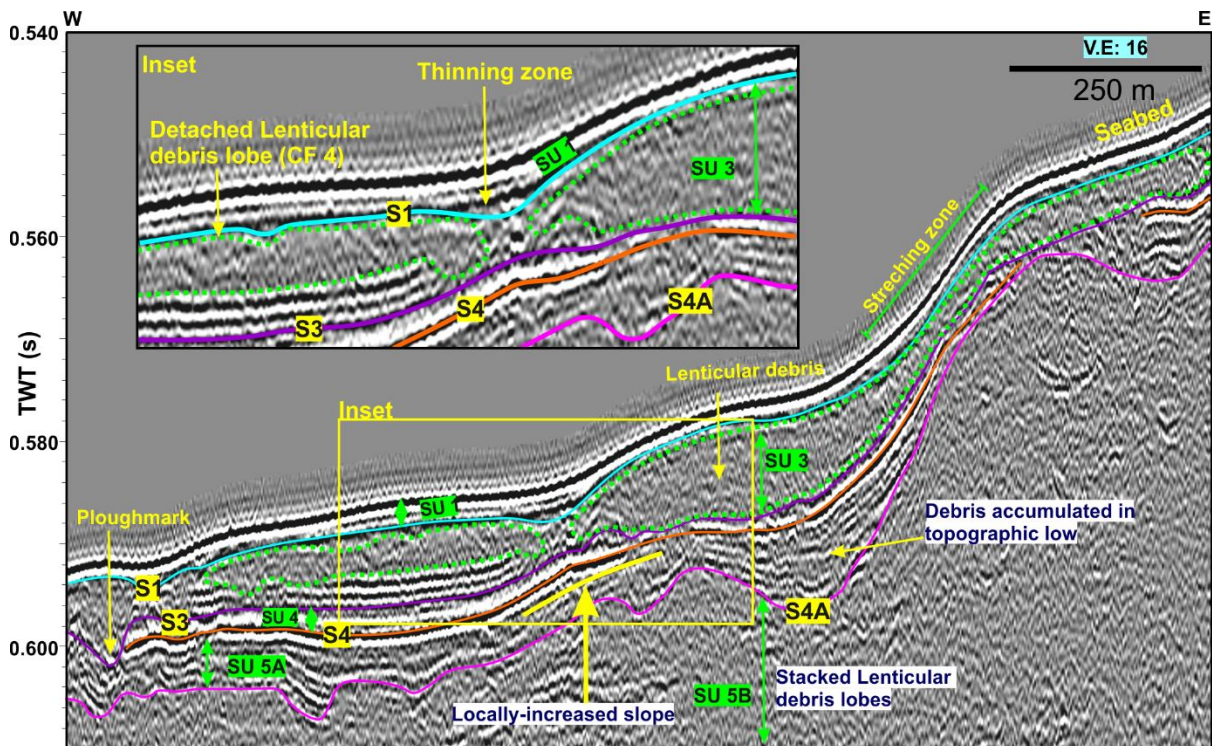


Figure 4.10. Slope perpendicular SYSIF seismic profile 2.4 (location in Figure 4.9) showing the debris lobes of unit SU 3. The debris lobes show intermediate stretching/thinning zones above a locally-increased slope gradient (inset). The depression is subsequently filled by later stage deposits (SU 1).

Seismic Unit 2 (SU 2):

Unit SU 2 is bounded by reflectors S1 at the top and S2 at the bottom (Inset, Figure 4.8 inset). In water depths from 450 to 900 m, SU 2 consists of well-stratified reflections of moderate to

high amplitude (SF2) (Figure 4.8, inset, and Figure 4.6b). The thickness of SU 2 varies from 1 to 6 m (Figure 4.7b). Isolated pockets of chaotic facies (CF1) are observed within SU 2 at a water depth of ~ 475 m (Figures 4.6b and 4.11). SU 2 unit is not preserved above water depths of 450 m (Figure 4.7b).

Interpretation: The presence of well-stratified, moderate-to-high amplitude reflections indicates the deposition of coarser sediments by turbidity currents, particularly at water depths ranging from 450 to 900 m. At a water depth of 475 m, an isolated pod of chaotic facies is interpreted as glacial debris deposits.

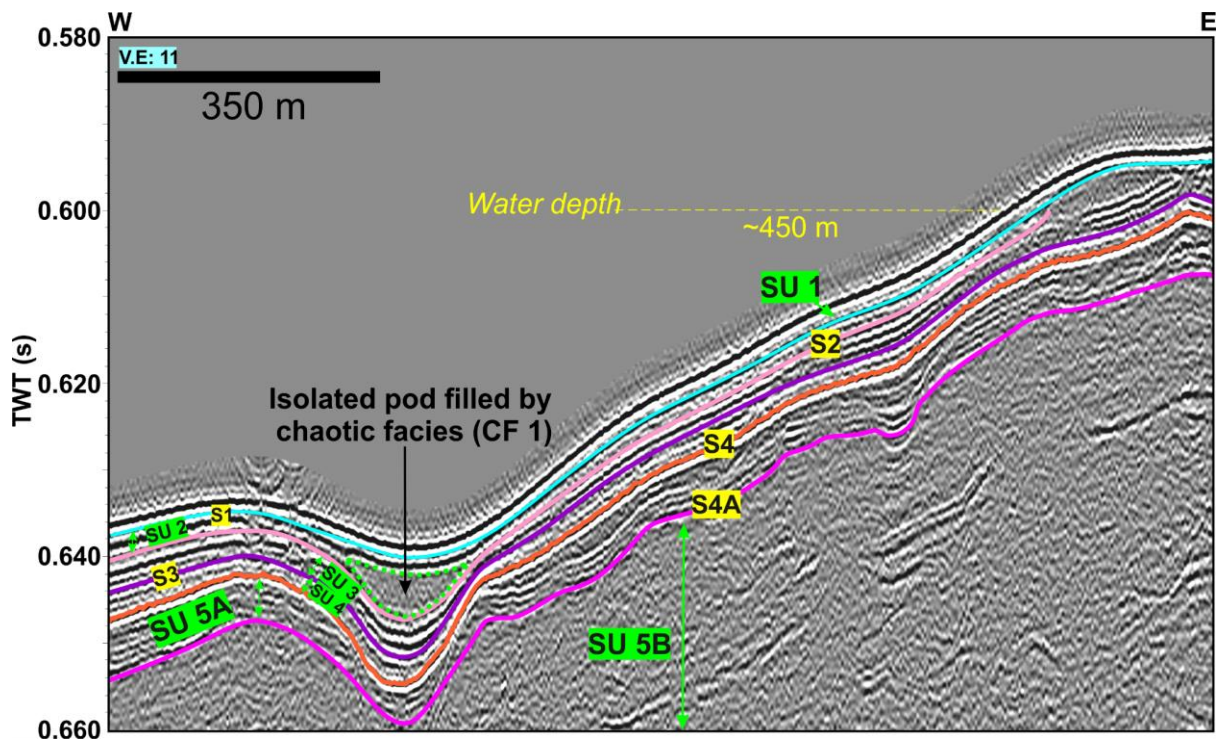


Figure 4.11. Slope perpendicular SYSIF seismic profile 2.3 (location in Figure 4.2) showing the debris filled in isolated pods. Seismic units SU 1-4 show well-stratified reflections. SU 2 pinches out at a water depth of 450 m. The isolated debris lens represented by chaotic lens facies within a depression probably represents a debris lobe detached from a parent debris flow. The concave top of the debris lens could be due to sinking of sediments due to density inversion, considering that the stratified and more compacted package (Units SU 2-SU 5A) got denser than underlying muddy debris (Unit SU 5B).

Seismic Unit 3 (SU 3):

SU 3 is bounded by reflectors S2 at the top and S3 at the base in water depths greater than 450 m (Figure 4.9), and by reflectors S3 at the base and S1 at the top in water depths less than

450 m (Figure 4.10). Reflector S2 is truncated by overlying chaotic lensoid unit, making it untraceable in water depths shallower than 450 m (Figure 4.11).

In the deeper part of the slope (water depths > 450 m), the lower part of SU 3 consists of continuous, parallel, moderate-to-high amplitude reflections (SF 2) that display an upward-aggrading pattern (Inset, Figure 4.9 and Figure 4.6c). The unit thickness is nearly uniform at around 4–5.5 m (Figure 4.7c). In the shallower part of the slope (water depths < 450 m), chaotic facies (CF 1) and chaotic lenticular lobes (CF 2) are observed (Figures 4.6c, 4.10 and 4.12). These lenticular lobes measure 150–600 m in length (N-S) with a 10-m thickness, and extend 500–700 m in the E-W direction (e.g. Figures 4.12, 4.11). The detachment of adjacent lenticular lobes has led to the creation of small depressions on the paleo-seabed, subsequently filled by SU1 (Figure 4.10). The continuity of reflector S3 is affected by indentations, 1 to 5 m deep, filled with chaotic, structureless, and incoherent reflectors (CF 1) (Figure 4.12). In water depths ranging from 345 to 410 m, high-amplitude continuous, parallel reflections are interbedded with chaotic lensoid packages (Figure 4.10). In the uppermost slope (water depths 390 m), the bottomset of prograding clinoforms (CF 4) (Figure 4.13) exhibits undulating wavy reflections. The average height of the wave is ~ 9 m.

Interpretation: In water depths less than 450 m, the chaotic facies in this unit represents glacial debris deposits. The indentations are interpreted as an action of iceberg ploughing, followed by infilling with chaotic debris (CF 1) facies. The indentations resemble the seismic expression of Late Weichselian ploughmarks observed in JR211 seismic lines and TOPAS lines by Sarkar et al. (2012) and Zhao et al. (2017) respectively. It is likely that icebergs were drifted northward by the West Spitsbergen Current, as observed during the Late Weichselian (Zhao et al., 2017). Alternatively, the presence of debris lobes suggests potential debris channelization, which likely contributed to the erosion process. Based on the external geometry and internal reflection, the chaotic lenticular facies (CF 2) pattern is interpreted as debris lobes. The thinned zone between adjacent lenticular lobes is the detachment area for these debris flow lobes. The bottomsets of prograding clinoforms on the uppermost slope display wavy reflections, indicating the action of northward flowing contour currents. This current, coupled with the eastward Coriolis deflection, redistributes the sediments as they cascade down the slope. In water depths greater than 450 m, continuous, parallel reflections with high-medium amplitude (SF 2) indicate coarser sandy and silty sediments deposited by turbidity currents

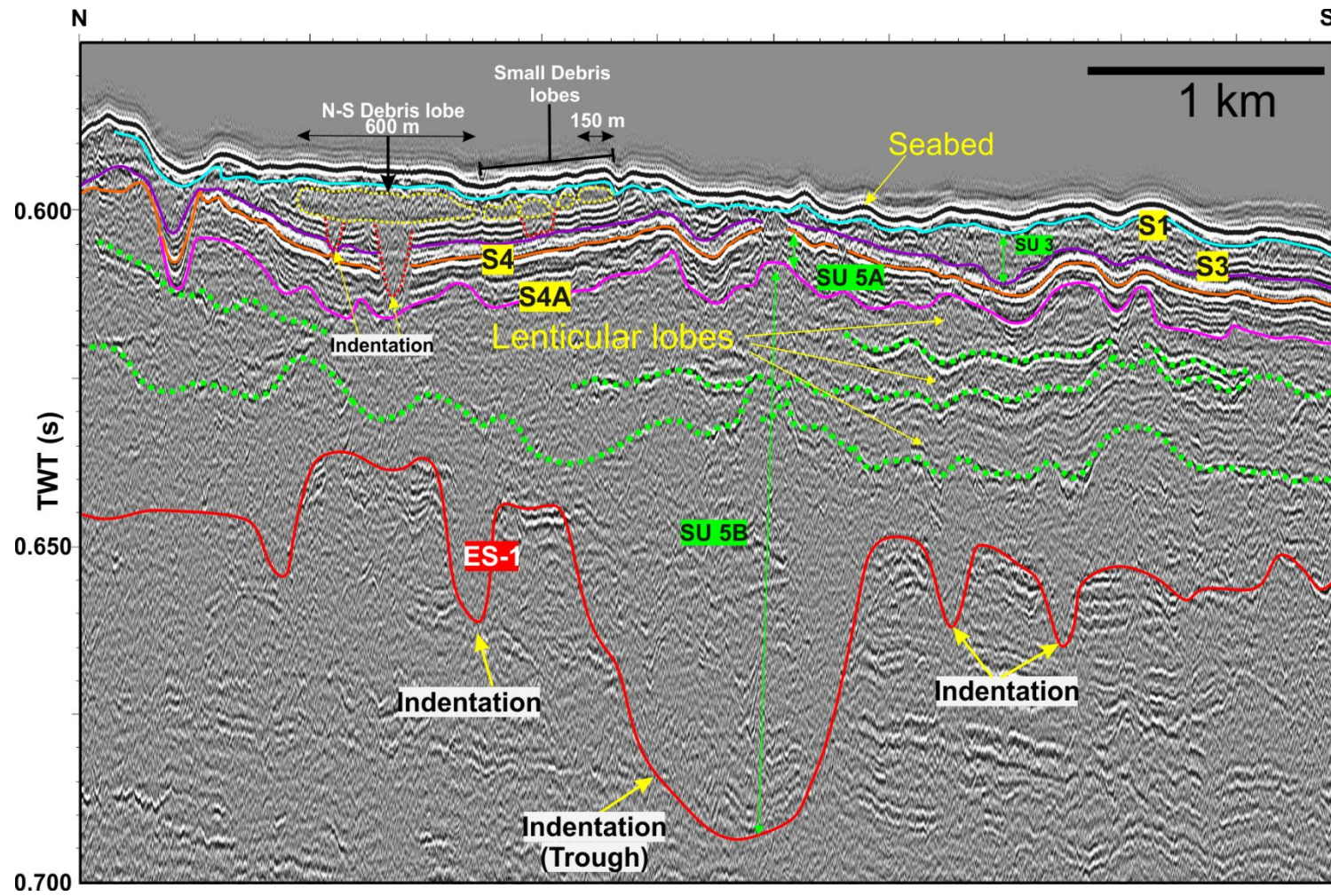


Figure 4.12. Slope parallel SYSIF seismic profile 3.14 (location shown in the Figure 4.2) showing prominent indentations at the base of units SU 3 and SU 5B. The ploughmarks are filled by incoherent chaotic reflection (CF1) and are interpreted as glacial debris flow deposits. Multiple stacks of lenticular debris lobes are present in unit SU 5B, while a single debris lobe set is seen in SU 3.

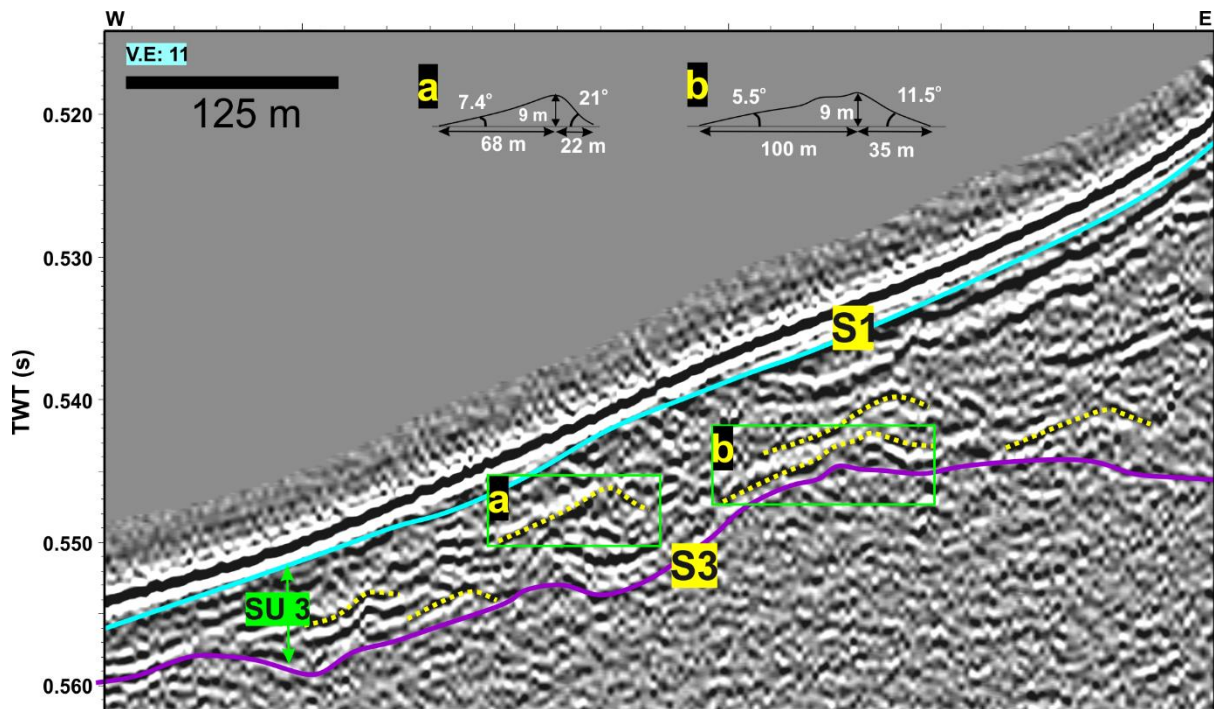


Figure 4.13. Slope perpendicular SYSIF seismic profile 2.1 showing the bottomset of prograding clinoforms. The yellow dotted wavy reflectors (a, b) indicate updip migrating sediment waves. Seismic line location is shown in Figure 4.9.

Seismic Unit 4 (SU 4):

SU 4 is bounded by reflector S3 at the top and S4 at the bottom (Inset, Figure 4.9). This unit consists of moderate-to-high-amplitude stratified reflections (SF 2) with rare occurrence of incoherent semi-transparent to transparent facies (CF 1) in isolated pockets (Figures 4.6d and 4.14). The thickness of this unit is nearly uniform ~ 3–4 m (Figure 4.7d). The unit has undergone erosion, resulting in its absence in water depths < 435 m (Figure 4.15).

Interpretation: Moderate-to-high-amplitude stratified reflections (SF 2) likely represent glaciomarine deposits with finer grained sediments (silt or silty clay) while the presence of isolated incoherent semi-transparent to transparent facies (CF 1) represents detached debris deposits from parent debris flows.

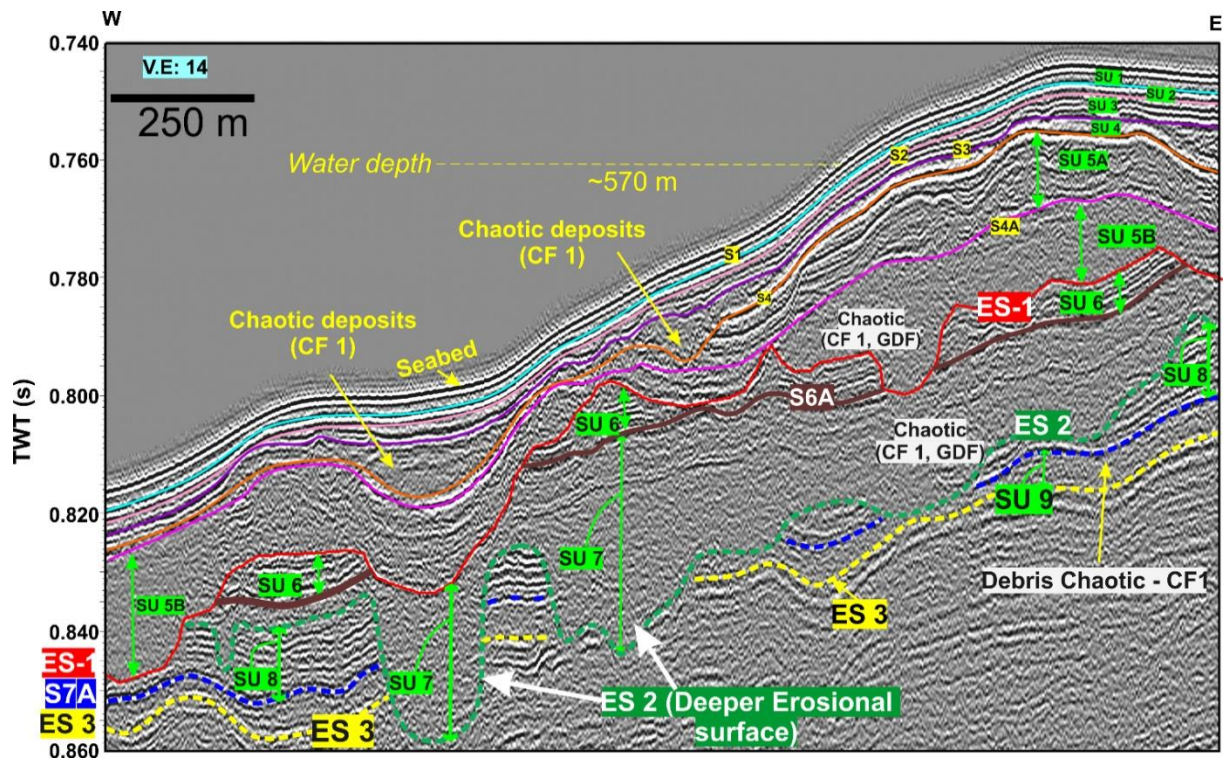


Figure 4.14. Slope perpendicular SYSIF seismic profile 2.1 showing three prominent erosional surfaces (ES 1-3). The isolated chaotic facies (CF 1) is seen in unit SU 4. The erosional surfaces are overlain by chaotic debris. The location of the seismic line is shown in Figure 4.2.

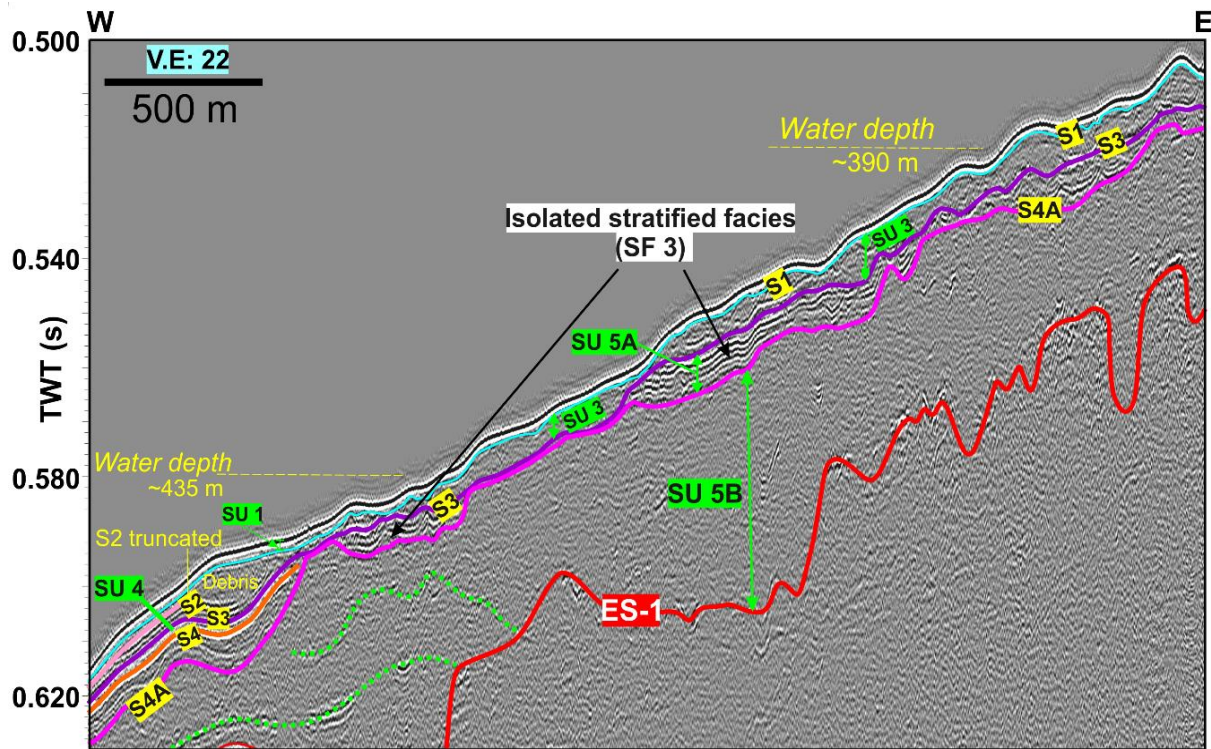


Figure 4.15. Slope perpendicular SYSIF seismic profile 3.1. The isolated stratified facies (SF 2) represents remnants of turbidite beds (SU 5A) that were later eroded by the GDFs (in this case GDFs of SU 3). In addition, the SU 4 deposits are not preserved in water depths < 435 m. The seismic line location is shown in Figure 4.2.

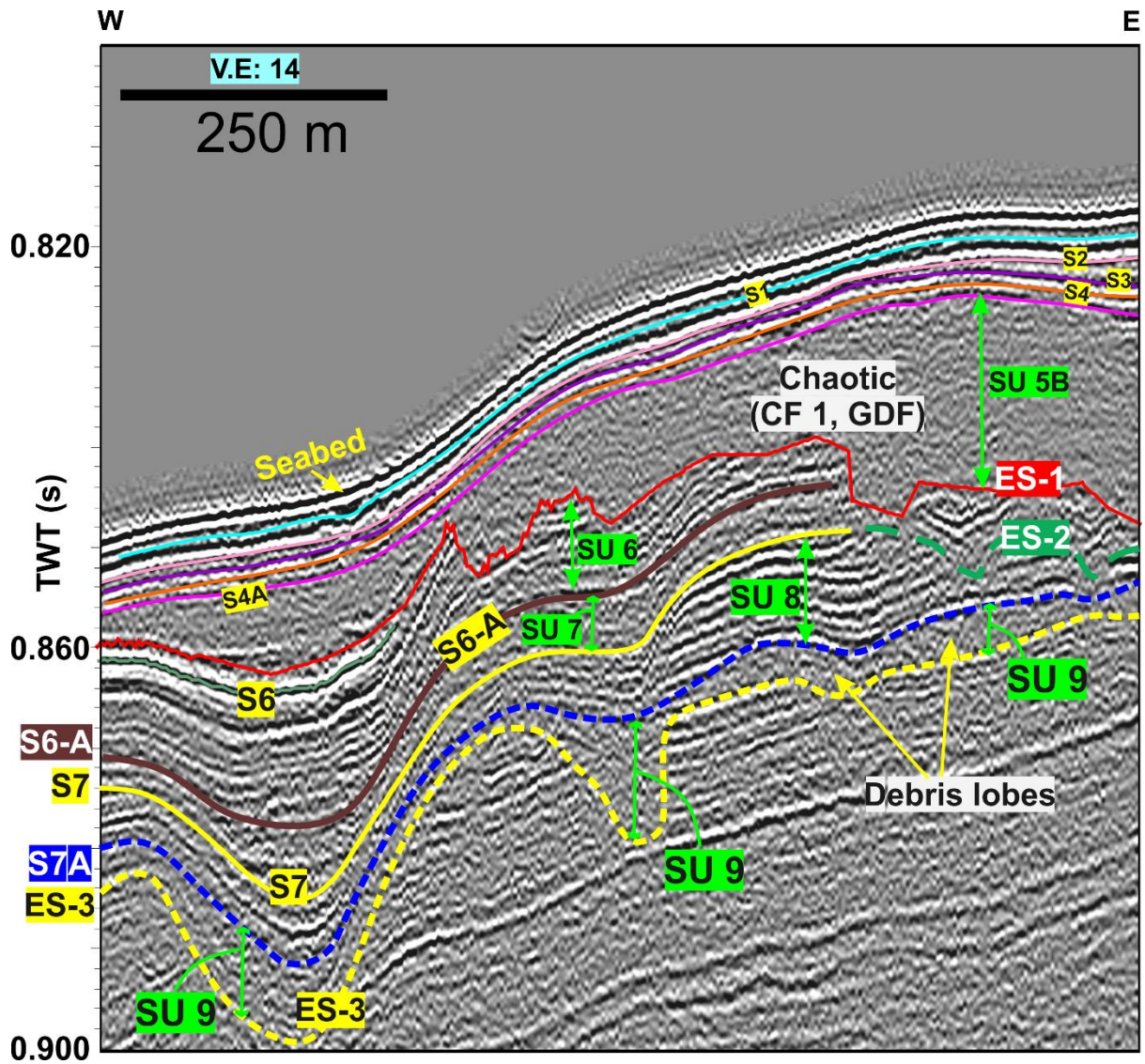


Figure 4.16. Slope perpendicular SYSIF seismic profile 2.1 showing the termination of reflectors S6 and S7 by the erosional surfaces ES-1 and ES-2. ES-3 is the deepest erosional surface showing the presence of grooves. Line location is in Figure 4.2.

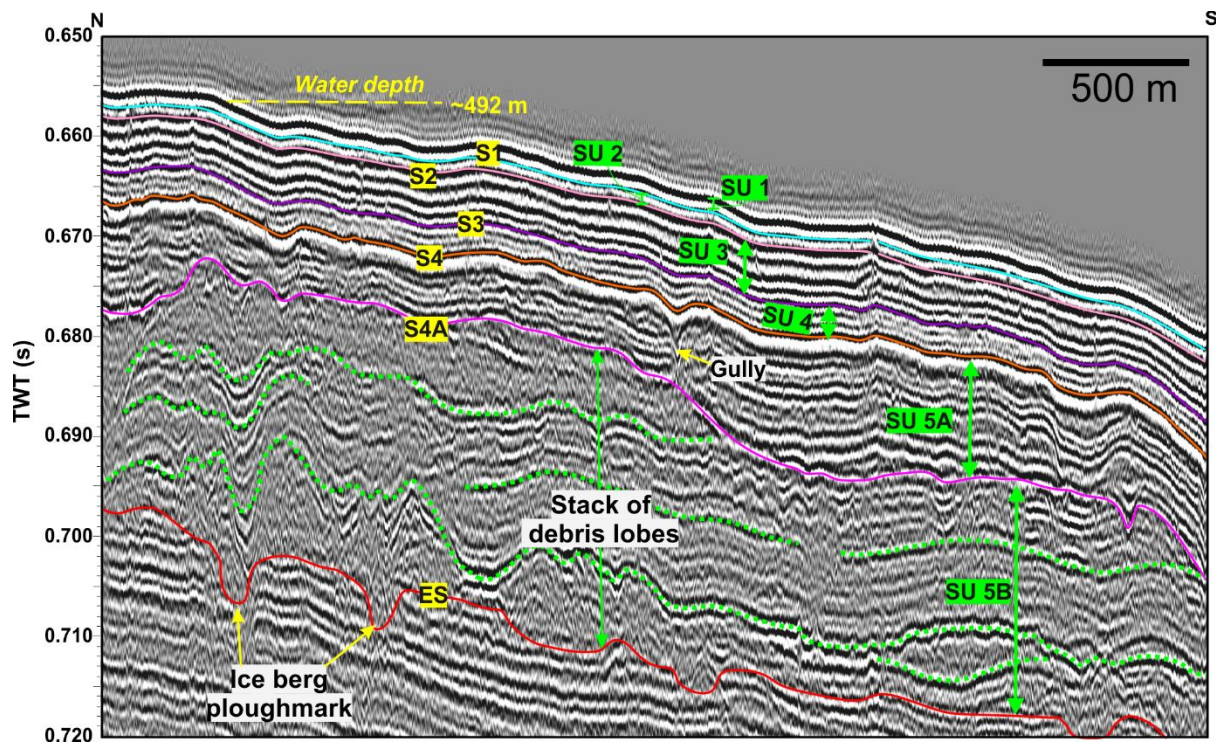


Figure 4.17. Slope parallel SYSIF seismic profile 3.13 showing vertical stack of multiple sets of lensoid packages showing internal wavy and hummocky reflection pattern in SU 5B separated by semi-continuous moderate amplitude reflections. Following the settling of each lensoid package there was a pulse of turbidity currents that resulted in the deposition of turbidites. The rough topography left by the earlier debris flow likely influenced subsequent flow events. This likely caused flow obstruction, flow diversions of debris material. The seismic line location is shown in Figure 4.2.

Seismic Unit 5 (SU 5):

SU 5 is bounded by reflector S4 at the top and S5 at the bottom (Inset, Figure 4.8). The unit is further subdivided into SU 5A and SU 5B by reflector S4A (Figures 4.12, 4.14, 4.16). The slope reflector S5 merges into an erosional surface (ES-1) at water depth ~ 635 m.

Unit **SU 5A** is bounded by the reflector S4 at the top and S4A at the bottom. In water depths greater than 500 m, the unit is predominantly represented by SF 2 facies (Figures 4.6e and Inset, Figure 4.8). In water depths between 600 and 450 m, SU 5A shows intercalation of chaotic facies (CF 1) and well-stratified reflections with medium and high-amplitude (SF 2) (Figures 4.6e, 4.12, 4.14, 4.15, and 4.18).

In water depths < 430 m, isolated well-stratified (SF3) facies and chaotic lensoid facies (CF 2) are present (Figure 4.6e). For example, there are chaotic lenticular facies separated by

parallel reflections in water depths of 400–440 m (Figure 4.15). In these water depths, SU 5A is mostly eroded, with only isolated remnants preserved. This eroded part is overlain by chaotic units of SU 3. The maximum thickness of the well-preserved part of this unit is ~ 9 m between 450–700 m water depths (Figure 4.7e).

Interpretation: Based on the lateral variation in the seismic facies in the upper continental slope, SU 5A shows transition from chaotic debris deposits to stratified turbidites. Isolated stratified reflections in water depths less than 430 m (Figure 4.15) are likely the remnants of turbidites, which were not completely eroded by later glacial debris flows. The thicker parallel-bedded seismic facies that is mostly preserved at water depths of 540–700 m suggest substantial sediment deposition from turbidity currents. The development of turbidity currents is attributed to progressive dilution of debris flow materials as they cascade downslope.

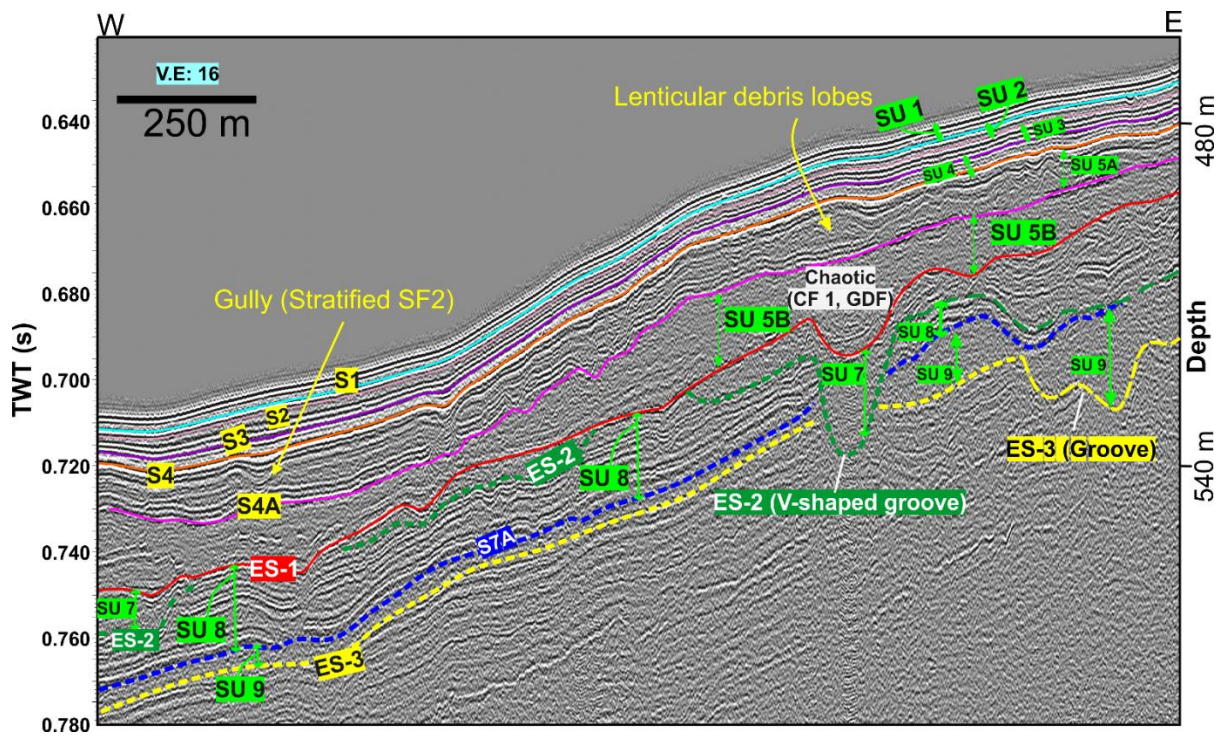


Figure 4.18. Slope perpendicular SYSIF seismic profile 2.1 showing the lenticular debris lobes occasionally encased within the stratified reflectors in seismic unit SU 5A. Below the erosional surface ES-1, two deeper erosional surfaces are present (ES 2 and ES 3). The seismic line location is shown in Figure 4.2.

SU 5B is bounded by S4A as the top bounding reflector and ES-1 as the base reflector (Figure 4.17). The ES-1 reflector shows indentations (Figures 4.12, 4.15 and 4.18). The depth of these indentations in different SYSIF lines is shown in Figure 4.19 and the maximum and minimum depths of indentations for different water depths is presented in Table 4.6 below. The paleo-indentations show V, W, and trough shapes with sidewall slope angles ranging 9-10°, 6-8° (Table 4.6, Figure 4.19). These paleo-indentations are filled with chaotic facies (CF 1).

Table 4.6. Dimensions of paleo-indentations in different SYSIF lines.

Profile name	Water depth range (m)	Max. depth (m)	Min. depth (m)	Shape	Base geometry	Slope angle (degree)	Max. Width (m)	Min. Width (m)	Infill deposit facies
SYSIF 3.11	500-600	No data	No data	N/A	N/A	9-10	N/A	N/A	Chaotic facies (CF 1)
	400-500	30	8	V	Concave up	6-8	400	200	Chaotic facies (CF 1)
	300-400	12	2	V	Concave up	0.5-2	250	150	Chaotic facies (CF 1)
SYSIF 2.3	500-600	35	14	V	Concave up	9-10	200	50	Chaotic facies (CF 1)
	400-500	50	25	W, V	Plane	6-8	500	300	Chaotic facies (CF 1)
	300-400	10	5	V	Concave up	0.5-1	200	50	Chaotic facies (CF 1)

SYSIF 3.3	500- 600	45	5	V	Concave up, Plane	9-9.5	600	40	Chaotic facies (CF 1)
	400- 500	55	27	V, W	Concave up, Plane	6.5-8	500	250	Chaotic facies (CF 1)
	300- 400	2	1	V	Concave	0.5-0.6	200	50	Chaotic facies (CF 1)
SYSIF 2.4	500- 600	No data		N/A	N/A	N/A	N/A	N/A	N/A
	400- 500	30	2	V	Concave up, Plane	6-8	400	200	Chaotic facies (CF 1)
	300- 400	15	5	V	Concave up, Plane	5	300	200	Chaotic facies (CF 1)

This unit consists primarily of chaotic deposits (Figure 4.6f) and is characterized by chaotic structureless transparent facies (CF 1) and chaotic-lenticular facies (CF 2) with isolated medium to high-amplitude stratified reflections (SF 2) (Figures 4.12, 4.17 and 4.20), which separate the lenticular facies. The stack of lenticular facies is predominantly observed in the water depths 420 m to 480 m (Figure 4.6f). Chaotic lenticular facies overlies structureless transparent facies that fills in the indentations. The latter shows a greater run out (up to ~ 500 m water depth, Figure 4.6f), compared to lenticular chaotic sub-units. The maximum length and thickness of the lenticular facies in this unit are ~ 1500 m and ~ 20 m, respectively. In the water depths ranging from 300–420 m on the uppermost slope, there are oblique landward-dipping reflections. They show an apparent dip ranging from 0.8–1.43° and are characterized by chaotic internal reflections and occasional transparent units (CF 3) (Figure 4.21a). The lateral transition from the vertical stack of chaotic lenticular facies (Figure 4.20) at 420–480 m water depths to the landward-dipping debris lobes at 300–420 m water depths (Figure 4.21), represents the shelfward retreating or retrogradational stacking pattern of the debris

lobes. The chaotic facies in Unit 5B thins downslope in the deeper part, transitioning to stratified SF 2 facies (2–8 m thick) in water depths greater than 600 m (Figures 4.6f and 4.7f).

Interpretation: On the upper continental slope below 300–500 m water depths, the Unit 5B is dominated by chaotic facies, which indicates glaciogenic debris deposits. The erosional surface (ES-1) exhibits distinctive irregularity, marked by V/W-shaped indentations and erosional troughs. ES-1 is overlain by incoherent and structureless debris material (CF 1 facies). The contour-parallel alignment of thicker (≥ 23 m) debris-filled paleo-indentations (Figure-4.6f) is likely the result of iceberg drifting influenced by the West Spitsbergen Current, followed by ploughing and subsequent debris infilling. Similar examples of deeper ploughmarks > 40 m were previously reported in the Bear Island Trough (Andreassen et al., 2008) and the East Siberian continental margin (Niessen et al., 2013). The trough shape of the paleo-indentations could result from the erosive action of debris flow channeled along the paleo-indentations. Multiple layers of lenticular debris lobes accumulated on top of one another, forming a multi-storey amalgamation. Some lenticular lobes extend beyond this zone due to their high runout potential. The presence of eastward-dipping reflections on the uppermost continental slope (CF 3 facies) is a consequence of frontal obstruction caused by pre-existing debris mounds in the downslope region (Figure 4.21).

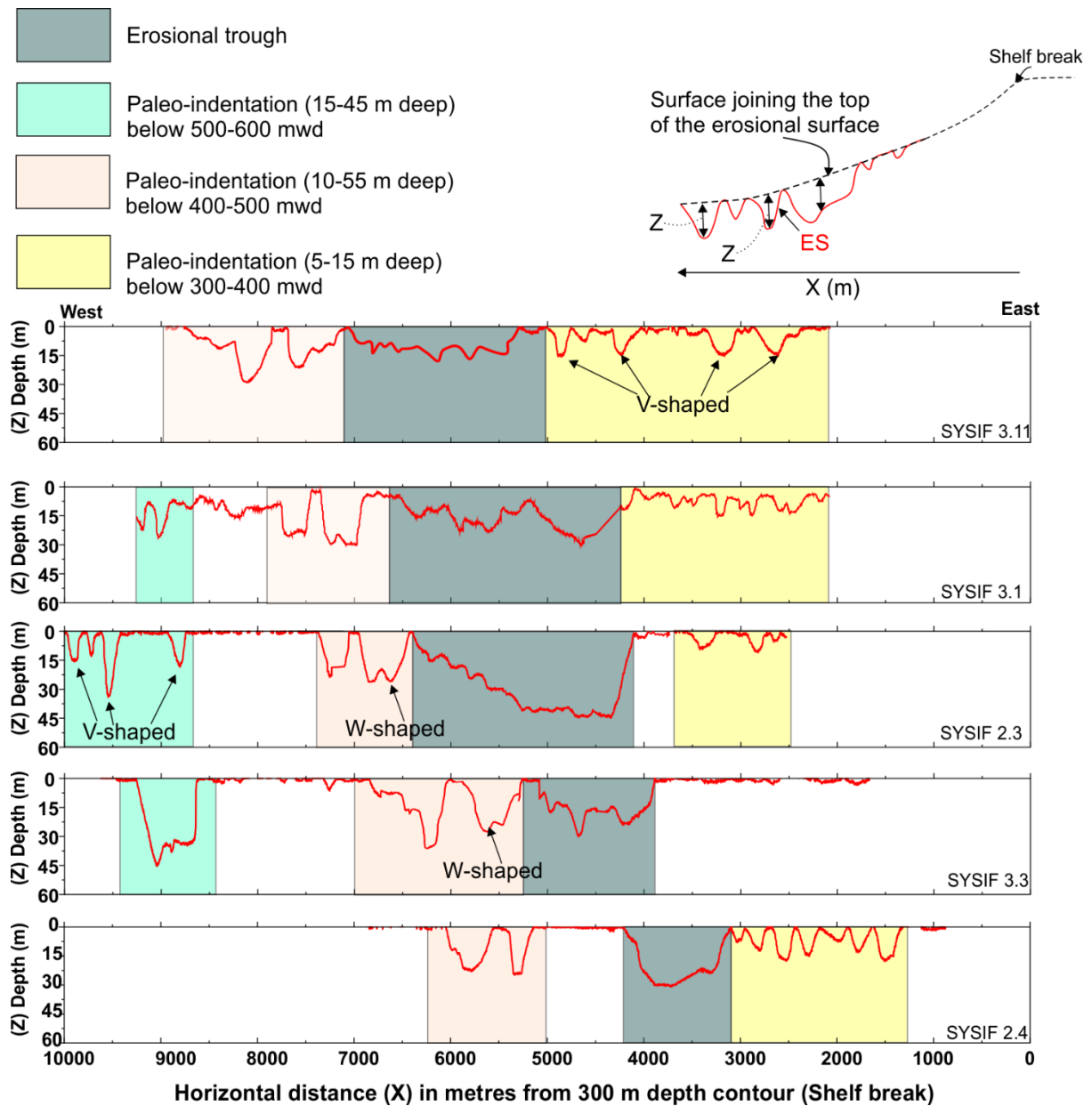


Figure 4.19. Depth variations of the erosional surface (ES-1) extracted from different SYSIF lines from North to South in the interfan area. The indentation depth is minimum in the water depth ranging from 300–400 m, while maximum in ranging from 400–500 m. Some V-shaped indentations are present in the distal slope, water depth ranging 500–600 m. SYSIF line locations are shown in Figure 4.2.

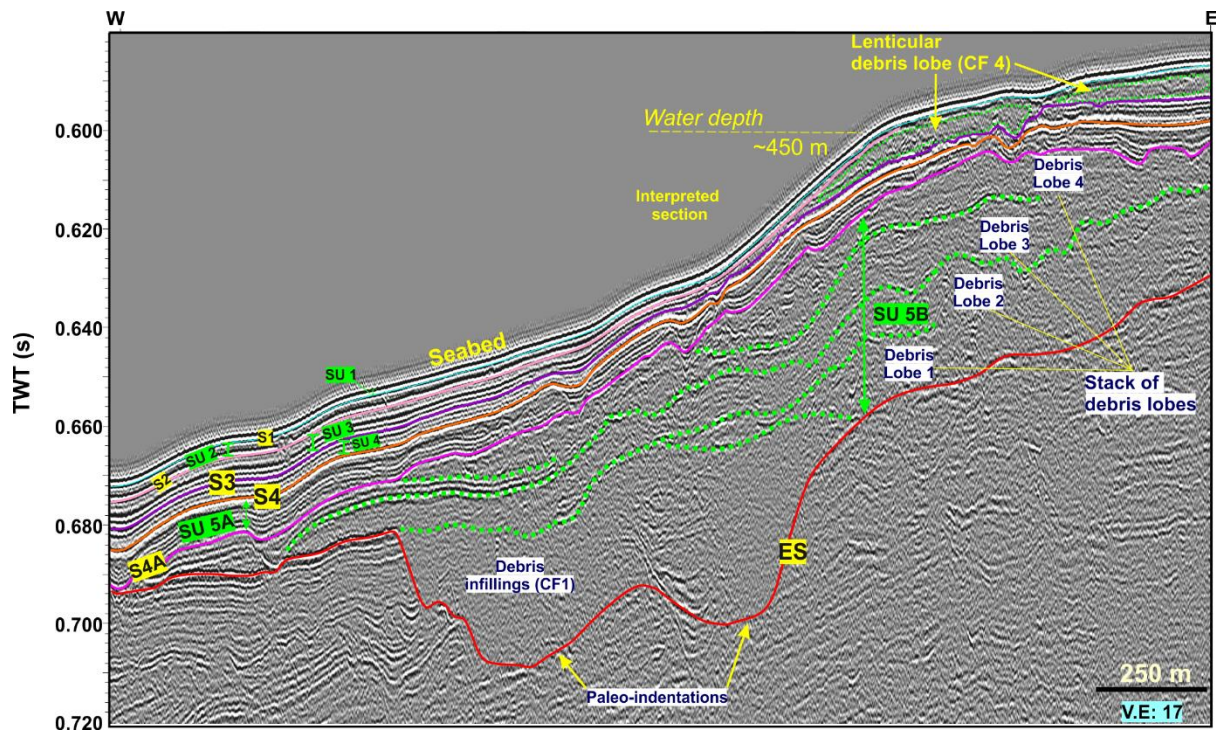


Figure 4.20. SYSIF seismic profile 2.4 showing a prominent erosional surface ES-1 filled by early debris flows and vertical stack of lenticular debris lobes deposited in the seismic unit SU 5B. The stacking of debris lobes on top of each other creates additional obstacles for younger debris lobes to overcome older ones. Profile location is shown in Figure 4.9.

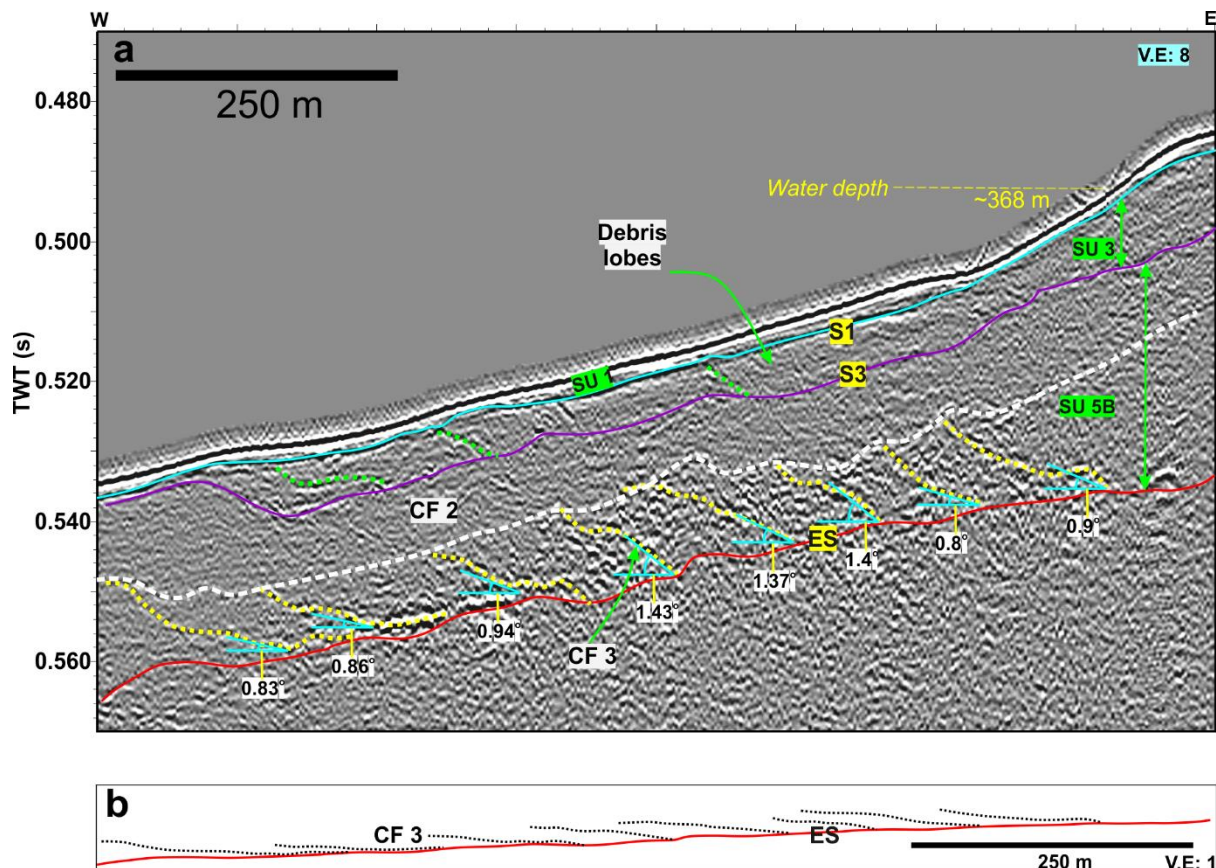


Figure 4.21. Slope perpendicular SYSIF seismic profile 3.11 (location shown in the Figure 4.2) showing the oblique debris stacking (CF 3) facies (yellow dotted lines) below the lenticular debris lobes. (a) The facies exhibits moderate to high reflection amplitudes compared to their surroundings, suggesting deposition of coarse-grained facies. The stacking pattern is indicative of individual debris lobes that were released from a till-wedge at the shelf break. The apparent dip of the individual lobes ranges from 0.8–1.43°. Due to factors such as, limited accommodation space, runout potential, and/or sediment supply constraints, these lobes tend to stack on each other obliquely. (b) Line drawing of the oblique debris stacks at a vertical exaggeration of 1.

Seismic Unit 6 (SU 6):

SU6 is bounded by S5 at the top and S6A at the bottom. This unit is present > 615 m water depths and is eroded by ES-1 (Figure 4.16) in shallower water depths. It consists of stratified SF2 facies.

Interpretation: The facies in this unit is interpreted as being predominantly of glaciomarine origin, in which SF2 facies represents coarse-grained turbidite deposits.

Seismic Unit 7 (SU 7):

In deeper water (depths > 650 m) SU 7 is bound by S7 (base) and S6A (top), represented by moderate amplitude stratified reflectors (SF2). Slope reflector S7 merges with the erosional surface ES-2, while S6A is truncated by reflector ES-1 on the upper slope. Hence, in shallower water (< 615 m water depth), SU 7 is bound by ES-2 (base) and S6A (top) or ES-1 (top), where S6A is eroded and represented by CF1 facies (Figures 4.14, 4.18). Reflector ES-2 shows indentations, e.g. V-shaped grooves.

Interpretation: The V-shaped indentations observed on ES-2 are likely caused by either iceberg ploughing or debris erosion. The seismic unit SU 7 is predominantly made of glacial debris in water depths < 615 m, while in distal part of the slope, the debris grades into coarse-grained silty turbidite deposits indicated by SF 2 facies.

Seismic Unit 8 (SU 8):

The unit is bound by ES-2 (top) and S7A (base) between 480 m and 615 m water depths, while in deeper water (> 615 m) it is bound by S7 (top) and S7A (base) as the erosional surface ES-2 merges with the slope reflector S7 (Figure 4.16). The unit consists of stratified facies of high to moderate amplitude reflections (SF 2). The unit is eroded by ES-2 and cannot be tracked at < 480 m water depth (Figures 4.14, 4.18).

Interpretation: The stratified reflections indicate coarse-grained sediments like sand mixed with silty turbidites.

Seismic Unit 9 (SU 9):

SU 9 is bound by ES3 (base) and S7A (top) and represented by chaotic facies CF1 (Figures 4.14, 4.16, 4.18). In deeper water (> 600 m), ES-3 merges with the slope reflector S7B (Inset, Figure 4.8). ES-3 shows V-shaped indentations.

Interpretation: The V-shaped indentations observed on ES-3 are likely caused by either iceberg ploughing or debris erosion. The unit is represented by glacial debris.

4.4.3. Stratigraphic model based on SYSIF data:

The key reflectors (S1–S7) from the SYSIF data were correlated with those identified from the TOPAS data (Wiberg et al., 2022) in the deeper-water (~ 925 m water depth), where they intersected (Figure 4.2). The wavelet length of the SYSIF and TOPAS data is 2 and 0.2 ms, respectively. The gap between the key reflectors in the TOPAS data was greater than the wavelet of SYSIF data (2 ms). Therefore, the maximum uncertainty of picking and

correlating the reflection events is ± 1 ms. The age of the reflectors from the TOPAS data was derived from the piston core GS10-164-09PC retrieved from the Kongsfjorden TMF region (Wiberg et al., 2022). Reflectors are assigned the following ages: S1 at ~ 15 ka, S2 at ~ 24 ka, S3 at ~ 38 ka, S4 at ~ 54 ka, S6 at ~ 75 ka, and S7 at ~ 90 ka. Additionally, ages for reflectors S5, S7A, and S7B were estimated at 57–61 ka, ~ 110 ka, and ~ 120 ka, respectively, considering the sedimentation rates provided by Wiberg et al. (2022).

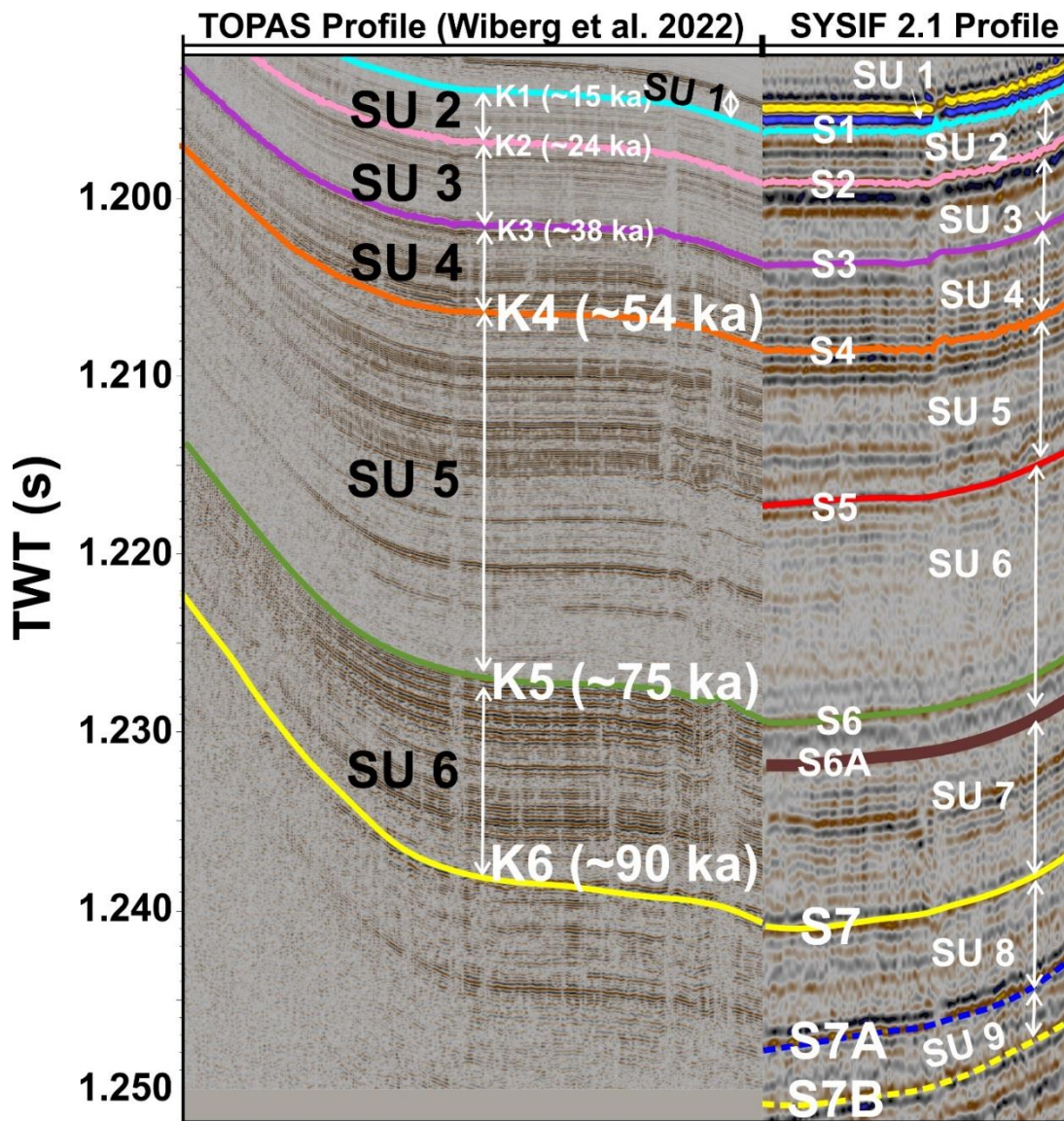


Figure 4.22. A combined TOPAS and SYSIF seismic profile in the interfan region highlights the stratigraphic correlation. Figure 4.2 shows the line locations. The ages were derived from the piston core GS10-164-09PC retrieved at the 945 m water depth in the KTMF region by Wiberg et al. (2022).

4.4.4. Debris flow modelling with RAMMS:

Debris flow modelling using Rapid Mass Movement Simulation (RAMMS) was carried out to understand the travel distance, evolution, and flow regime of lenticular debris flow lobes, including flow velocity, depth, and behaviour considering a Voellmy-cohesion rheological model. The RAMMS model was calibrated to find the best-fitting rheological parameters that would produce the observed runout distance, thickness, and morphology of the debris flow. It was assumed that the flow would stop at a momentum threshold of 6 %, which is determined by comparing the momentum values of every grid cell to the maximum momentum sum. This threshold aligns with the recommended threshold of less than 10 % in RAMMS. In this study, a 50-m-thick block mass was released from a till delta at the shelf break. Since most debris lobes are observed in the steeper upper regions, debris flow modelling attempted to mimic the behaviour of high-strength debris flow considering yield strengths ranging from 100 to 2000 Pa. High-strength debris flows (~ 100 Pa to > 1 kPa) generally occupy steeper gradients and remain closer to their source area, while flows with intermediate strengths (5-100 Pa) can travel down to gentler slopes (Talling, 2013).

The simulation results show various potential run-out distances, thicknesses, and speeds for submarine debris flows in the interfan region. Figures 4.23a–d show a simulation where the debris flow yield strength is assumed to be 2 kPa. The debris flow originates at the shelf break and lasts for approximately thirty minutes before coming to a halt on the upper slope. Five minutes after its release from the shelf break, the debris flow mass starts to elongate and the debris head surges ahead in the form of a lobate geometry with an estimated maximum velocity of 5 m/s. The debris flow typically shows a bulbous head and a tapering tail. There is a gradual reduction of mass in the flow's tail and the progressive accumulation of debris material in the head region. The flow front finally stops at the lower gradient part of the slope, ~ 6 km away from the point of origin. The cessation of the debris flow onto the seafloor is influenced by multiple factors, including a decrease in slope angle from 4° to less than 0.5° and the cohesive properties of the flow, calibrated to correspond to a yield strength of 2 kPa.

In the case of the flow with a yield strength of 100 Pa, the flow ceases at a distance of 8.5 km from its origin reaching the gentler part of the slope (Figures 4.24a–d). Debris flows with lower yield strength tend to break up more easily than those with higher yield strength. The final state shows several localized features with an undulating morphology compared to the

final morphology of the debris flow in the case of yield strength assumption of 2 kPa. The undulating topography of the debris lobe is due to local slope variations of the substrate. The debris mass mostly accumulates on gentler sections while thinning out on steeper sections as the flow velocity is relatively higher in those regions. Cohesive lenses and stretched/thinned debris with internal bulges are observed in the seismic data (Figure 4.10). The modelling aids in comprehending the potential reasons for their creation.

In a separate experiment, the potential of a barrier to halt the debris flow was examined. When the energy needed to overcome the barrier is lacking, the flow front is constrained by the barrier, forming a thicker head leaning against it, and a gradual, landward-dipping tail (Figure 25).

Additionally, the infilling of a ploughmark by debris was checked. The debris flow exhibits a distinct tongue-shaped outline as it advances toward a N-S aligned ploughmark (Figure 26 a-f). The infilling of the ploughmark initiates when the debris flow encounters the groove, progressively extending laterally along the groove as the filling process proceeds.

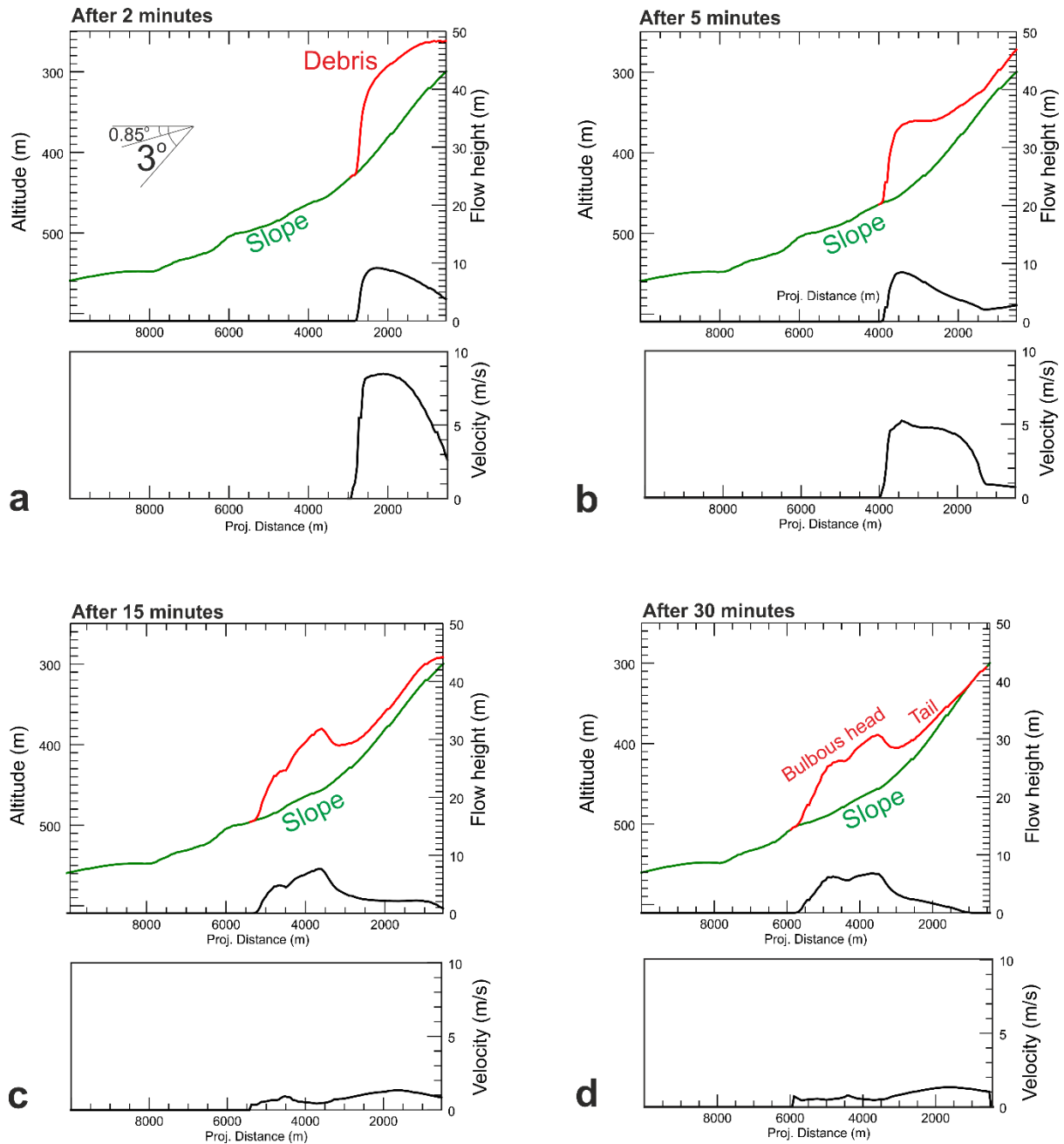


Figure 4.23. Evolution of debris flow after being released as a block at the shelf break, considering a yield strength of 2 kPa. **(a)** The debris flow mass 2 minutes after its release shows a high initial flow velocity of 9 m/s. **(b)** After 5 minutes, the debris mass is shifted towards the flow head, and consequently, the tail thins out. **(c)** After 15 minutes, a distinct bulbous head is formed. **(d)** After half an hour, the maximum thickness of the debris lobe, achieved approximately 7 km away from the initial release site after flow cessation, is 5 m before coming to a halt.

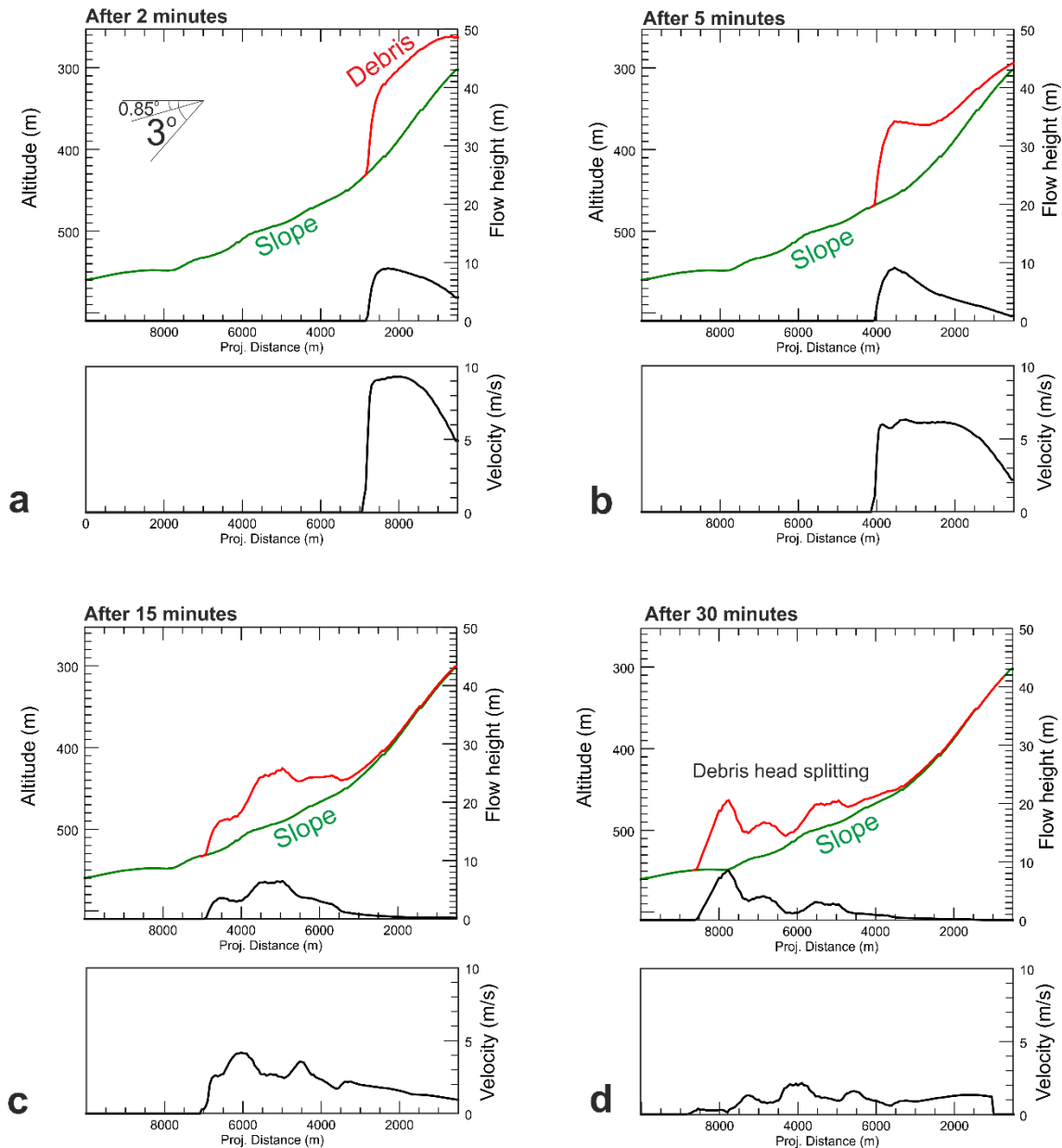


Figure 4.24. Evolutionary changes in debris flow released as a block at the shelf break, considering a yield strength of 100 Pa. **(a)** Two minutes after release, the debris flow mass exhibits a significant initial flow velocity of 10 m/s. **(b)** The debris mass undergoes a transformation in shape after 5 minutes, leading to an overall lensoid shape. **(c)** After 15 minutes, the lensoid debris mass surges ahead at 5-7 m/s. **(d)** Thirty minutes later, the debris flow mass begins to develop distinct internal bulges separated by thinned zones. The debris lobe comes to a halt approximately 8.5 km away from the initial release site.

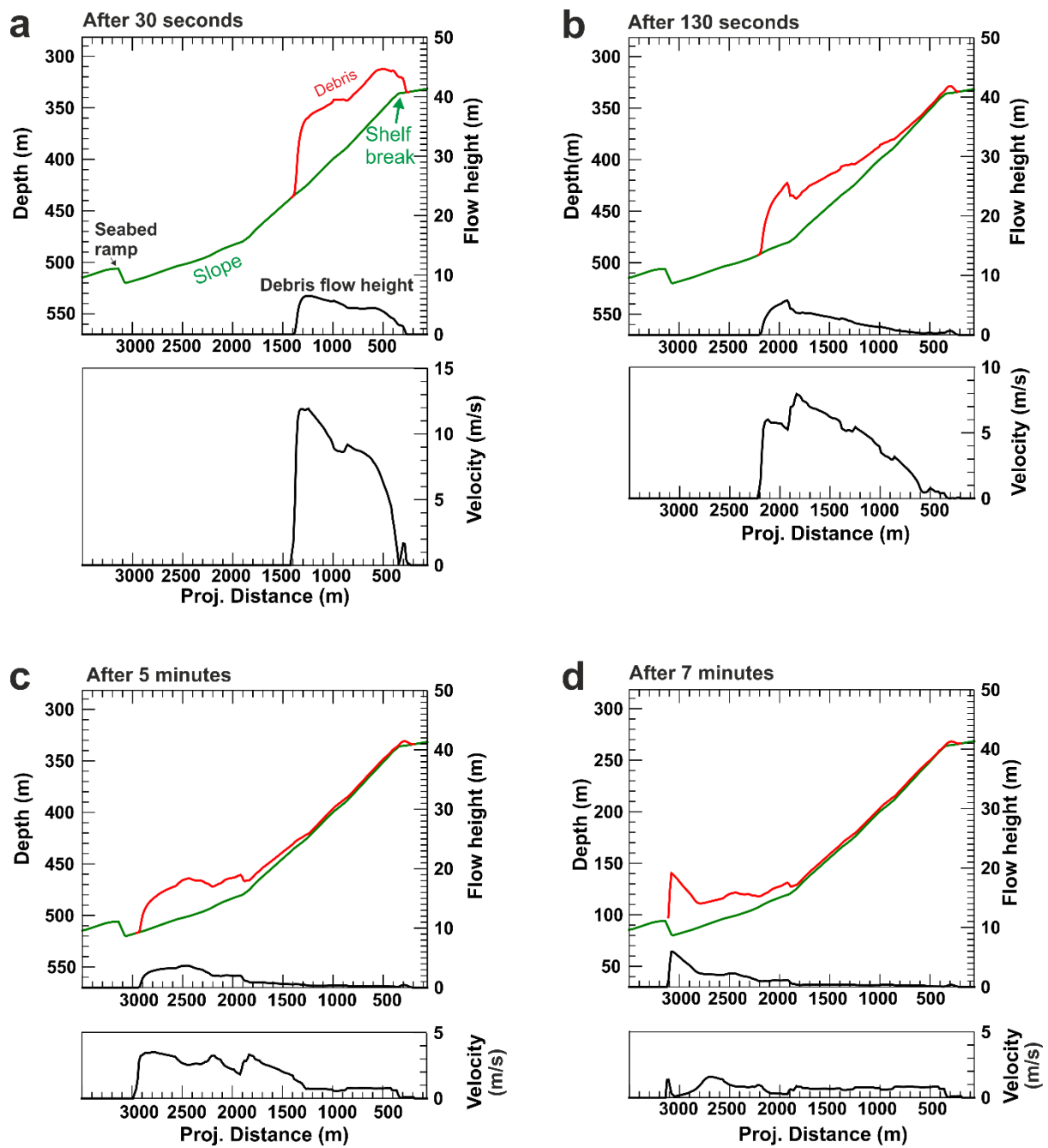


Figure 4.25. Evolutionary changes in debris flow response to an obstacle (ramp). (a) After 30 seconds of release at the shelf break, debris flows downslope with a maximum velocity of 12.5 m/s. (b) After 130 seconds, the shape of the debris mass undergoes transformation, thinning at the tail and bulging at the frontal head. (c) After 5 minutes, the lensoid debris mass surges ahead at speeds ranging from 1 to 4 m/s. (d) Seven minutes later, the debris flow is halted by the ramp due to insufficient energy to overcome the obstacle. The final geometry of the debris lobe reveals a gentle landward sloping tail.

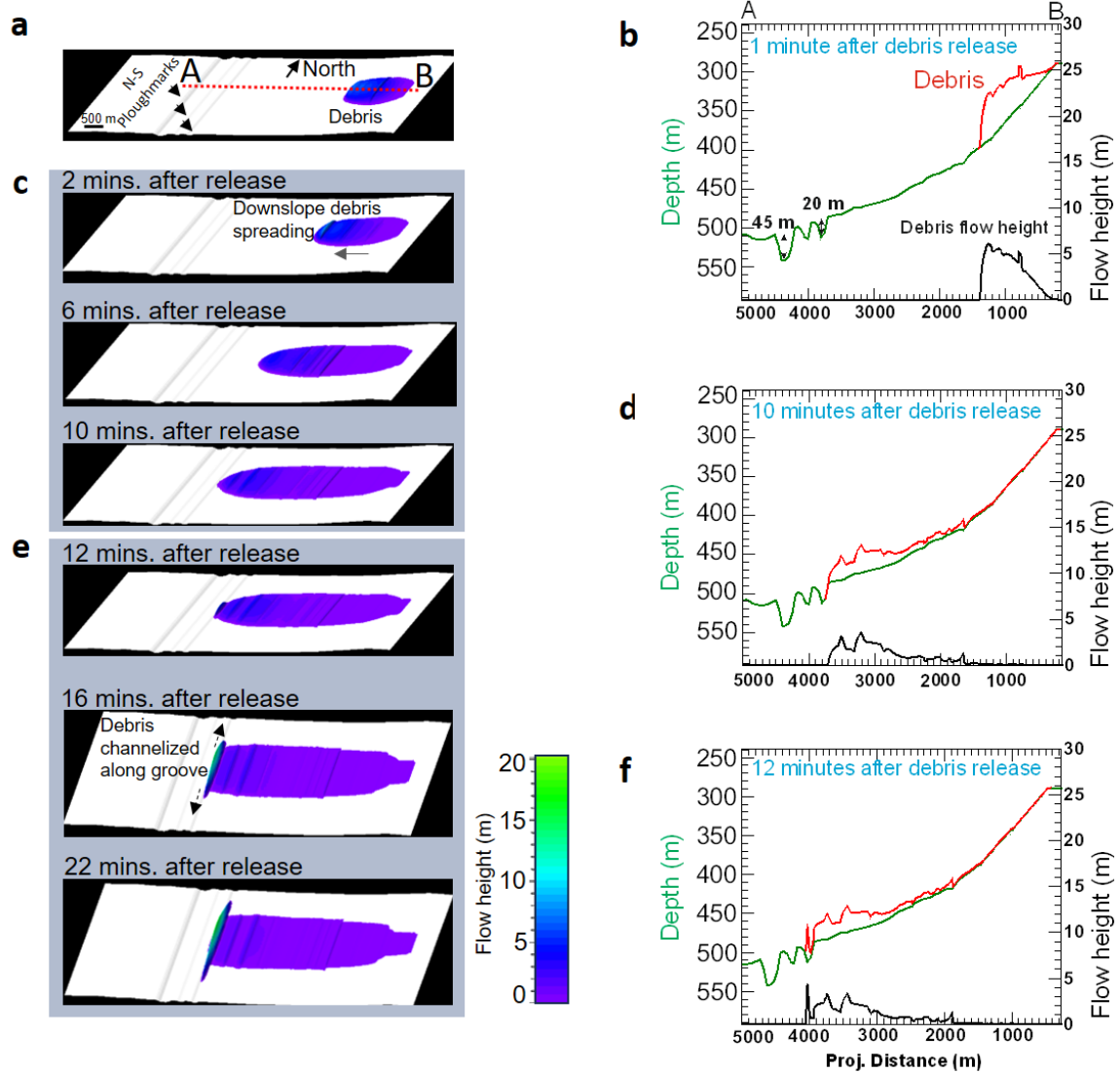


Figure 4.26. RAMMS simulation showing ploughmark infilling. **(a)** Three-dimensional depiction of a westerly dipping slope showcasing the release of debris mass from the shelf break, with three N-S aligned ploughmarks approximately 4 km away from the shelf break. **(b)** Section AB presents the profile of the slope and ploughmark grooves. **(c)** Visualization of debris flow spreading and changes in debris flow height at 2, 6, and 10 minutes post-release. **(d)** Profile section illustrating the debris snout reaching the brink of a ploughmark. **(e)** Illustration of debris filling the ploughmark on the most landward side of the slope. **(f)** Profile section displaying debris infilling the ploughmark.

4.5. Discussion

4.5.1. Reconstruction of glacial history and paleo-depositional environment:

The phases in the development of glaciers are identified by analyzing variations in seismic facies in various seismic units. The following sections provide an overview of the evolution of glacial ice sheet in the interfan area between KTMF and ITMF over the last 2.7 million years and discuss the changes in paleo-depositional environments and glacial dynamics.

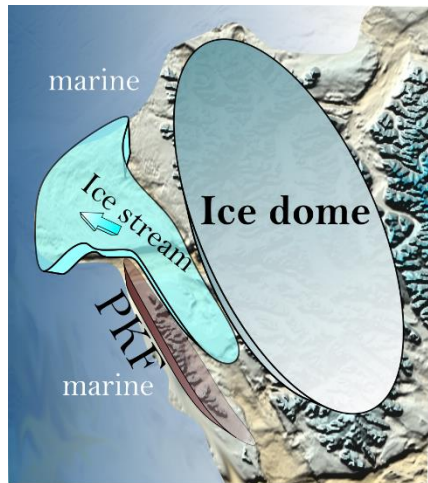
4.5.1.1 Period > 2.7 Ma:

The parallel continuous to semi-continuous low to medium amplitude well-stratified facies in IU-5 in the outer shelf suggests sedimentary strata while the incoherent facies in IU-6 represents the crystalline rocks (Figure 4.4). In the upper slope, the northward flow of the West Spitsbergen Current (WSC) led to upslope climbing contourite deposition with a distinct wavy pattern (Figure 4.4).

4.5.1.2 Period between 2.7-1.2 Ma:

The sedimentary strata deposited between 2.7 million years (R7) to 1.2 million years (R4A) shows climbing up wavy sedimentary succession in the upper slope that thins out towards the shelf break and outer shelf (Figure 4.4). The initiation of North Hemisphere glaciations, exemplified by the shelf edge glaciation in northwestern Svalbard, led to the formation of the Sjubrebanken fan at ~ 2.5 Ma-1.2 Ma (Figure 4.27). However, during this period, there were no indications of fan development in the Kongsfjorden area (Sarkar et al., 2011).

**Sjubrebanken
ice stream
(~ 2.5-1.2 Ma)**



**Kongsfjorden
ice stream
(initiating at 1.2 Ma)**

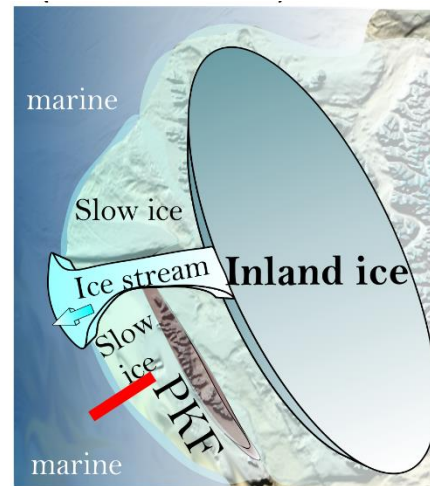


Figure 4.27. Cartoon showing ice stream switching in western Svalbard. At 2.58 Ma, a northwesterly flowing ice stream initiated the development of the Sjubrebanken Trough Mouth Fan (Sarkar et al., 2011; Alexandropoulou et al., 2021), with Prins Karl Forland (PKF) acting as a topographic barrier. Around 1.2 Ma, as the inland ice thickness increased, the ice overcame the PKF barrier and flowed westward along the Kongsfjorden Trough. The red line in the interfan region shows the location for Figures 4.28 and 4.29.

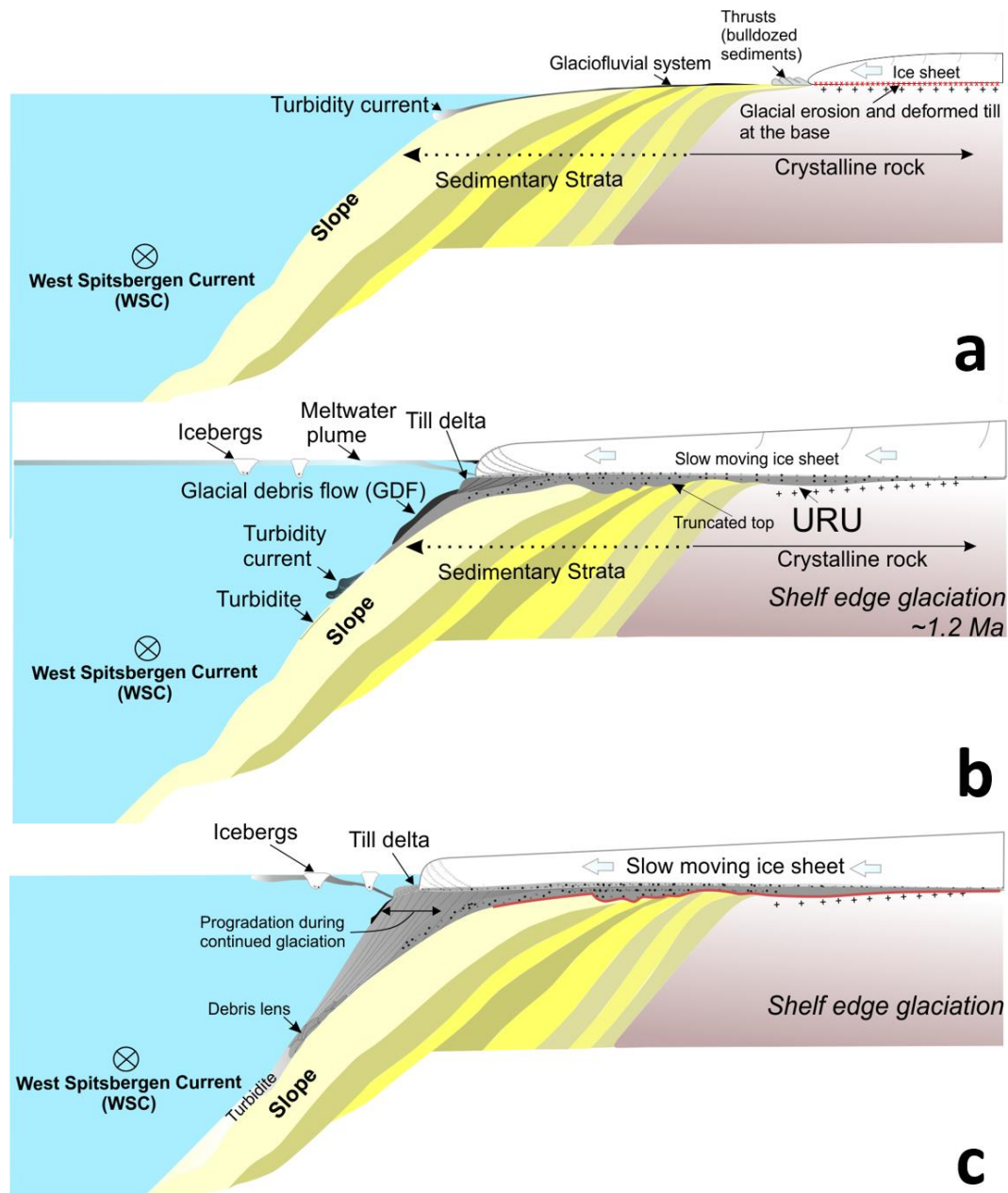


Figure 4.28. Cartoon showing ice sheet advancement toward shelf break, forming a prograding glacial wedge. **(a)** The ice sheet moves over the crystalline basement. **(b)** Erosion of the underlying marine beds creates a distinct unconformity (URU) during a glacial advancement at 1.2 Ma. A wedge-shaped till delta forms at the shelf break as glacial sediments are brought in. Glacial debris are released from the unstable till wedge, cascading downslope as cohesive debris flows. **(c)** Ice sheet advancing and building up a prograding glacial wedge consisting of oblique basinward downlaps on the upper slope. Icebergs form due to calving at the termini of grounded ice. On the slope, the debris flow mixes with ambient water, getting diluted into turbidity currents and depositing turbidites.

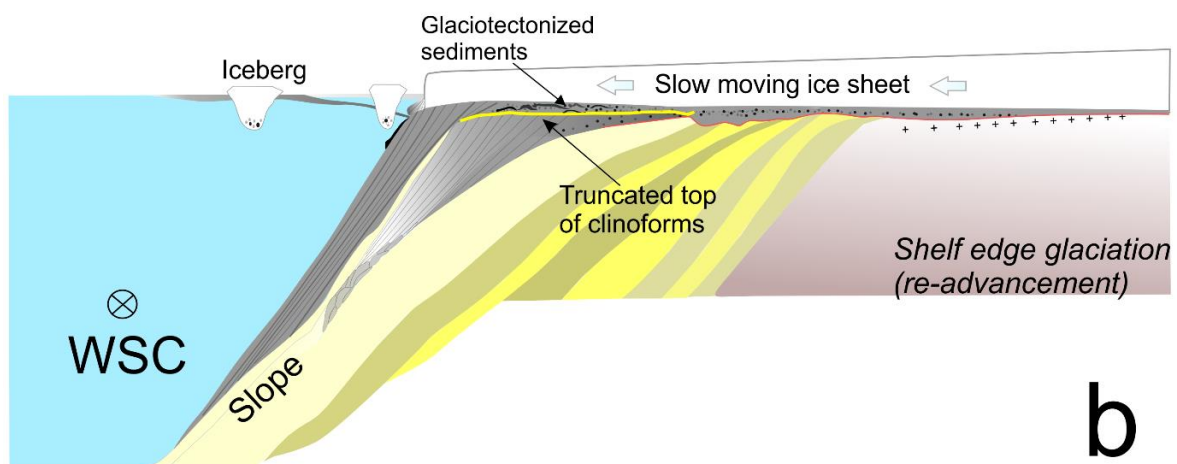
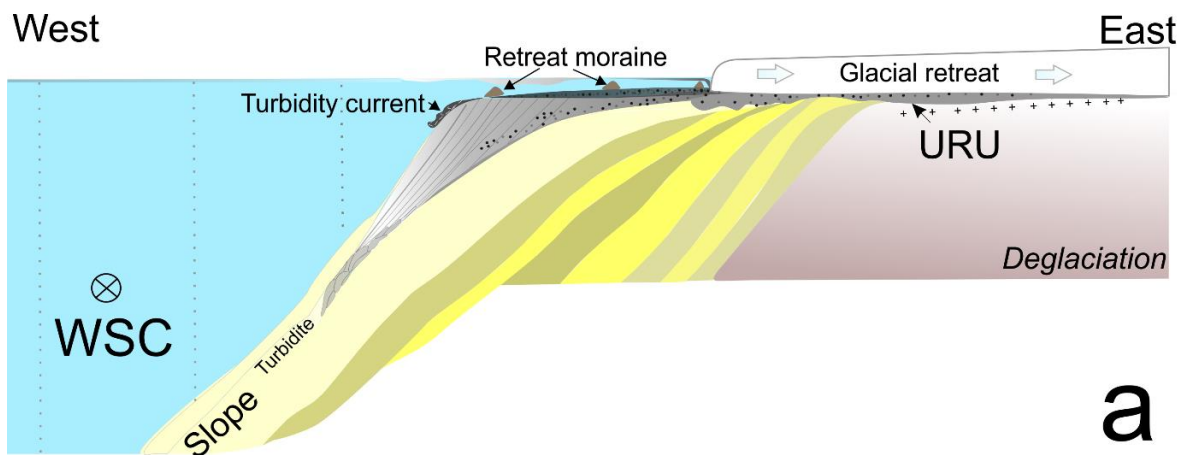
4.5.1.3 Period from 1.2–0.074 Ma (IU-4, IU-3, and IU-2) and < 0.074 Ma (IU-1):

The chaotic wedges above the R4A reflector (1.2 Ma) indicate the presence of glaciogenic deposits formed by the advancing ice during the basinward progression of the ice sheet in the interfan area. Shelf progradation is evidenced by the prevalence of lensoid packages alternating with moderate to high-amplitude basinward advancing clinoforms downlapping reflector R4A (Figure 4.4). These lensoid packages denote glaciogenic debris, while the moderate and high amplitude reflections result from the draping of hemipelagic sediments, deposited through suspension fallout from meltwater plumes, and sandy turbidites settled from turbidity currents. The URU on the shelf marks a substantial erosion of underlying strata and shelf progradation above the URU. Figures 4.28a, 4.28b, and 4.28c schematically depict the initial ice sheet advancement, resulting in the erosion of underlying marine units, creating the URU and transporting glacial debris to the outer shelf and shelf break. Shelf edge glaciations at 1.2 Ma and in each subsequent cycle resulted in the formation of a prograding glacial wedges. The presence of prograding sequences on the outer shelf implies availability of accommodation space, while the inner shelf region was susceptible to either erosion or sediment bypass. In the KTMF region, R4A (~ 1.2 Ma) serves as the base of the lensoid debris flow packages extending to the upper slope, indicating the initiation of ice-streaming along the Kongsfjorden Trough (Figure 4.27). However, the interfan region shows moderate progradation compared to the fan development in the KTMF. Between 1.20 Ma (R4A)–0.074 Ma (X1), the shelf witnessed multiple glacial advancements, evident in distinct sets of oblique westward-dipping lensoid debris packages. Due to the lack of age constraints in the shelf and the difficulty in tracking regional reflectors on the upper slope and outer shelf, dating multiple episodes of glacial advancements and retreats remain challenging. However, by utilizing the available age reflectors, we correlate the prograding clinoforms downlapping the reflectors R4A (1.2 Ma) and R2 (0.42 Ma) to the glaciations of Mid-Pleistocene and Early Saalian, respectively.

Figure 4.29a illustrates a hypothetical scenario where the ice sheet retreats from the shelf break area. During this period, there might possibly be shorter pauses in ice retreat, resulting in the deposition of retreat moraines during glacial stillstands on the outer shelf. On the distal slope, turbidity currents became more prevalent due to the diminished supply of glacial debris and abundant meltwater causing dilution of debris. In this phase, turbidites are deposited over the glacial debris flows. Figure 4.29b depicts the re-advancement of the ice sheet towards the shelf break. As the ice advances, it erodes the existing deglacial sediments. The top of the

prograding glaciogenic sequence in units IU-4 –2 are truncated by reflector M (Figure 4.4), indicating such glacial erosion during the readvancement of the ice sheet.

On the shelf break and upper slope, the shallowest seismic unit IU-1 exhibits prograding clinoforms, suggesting ice sheet advancement to the shelf break. Reflector W marks the base of this unit, truncating X1 (~ 75 ka) on the upper slope, indicating Unit IU-1 is < 75 ka. The distinct episodes of glacial advancements and retreats in this unit cannot be resolved with the air gun seismic data since it lacks the resolution required to decipher the intricate stratigraphic patterns in the upper slope. To address this limitation, SYSIF seismic lines from the upper continental slope were interpreted to reconstruct the glacial (Weichselian) history during the last 120,000 years and glaciomarine sedimentation patterns.



(For figure caption see the next page)

Figure 4.29. Cartoon showing the sedimentation style during a cycle of deglaciation and ice sheet readvancement towards the shelf break. **(a)** During deglaciation, the ice sheet retreats, forming retreat moraines during glacial stillstands. While the supply of glacial debris diminishes at the shelf break, abundant meltwater gives rise to energetic turbidity currents. **(b)** The readvancement of the ice sheet results in a distinct erosional unconformity that truncates the top of the underlying glacial clinoforms from previous cycles, initiating the development of a new prograding wedges.

Weichselian glacial history and depositional environments based on SYSIF lines:

In the seismic lines on the slope, the SYSIF seismic units reveal an alternation of chaotic debris and stratified deposits separated by prominent erosional surfaces ES 1–3. These features help in reconstructing glacial advances and retreats from the shelf break.

The farthest extent of the grounded ice sheet marks the maximum basinward advancement during a specific glaciation. If the ice continues its advance, the part of the ice floating in the water becomes prone to calving events. Calving events are likely to occur during or even after ice sheet grounding, particularly when the termini of the ice sheet has stretched to their maximum dimensions and has become unstable. The deep-keeled icebergs gouge the uppermost slope, leaving behind ploughmarks.

A submerged "till delta" forms at the shelf edge by the grounded ice sheet during the period of shelf-edge glaciation. The glacial debris cascades downslope from the unstable sediment wedge established by the ice sheet. However, when the ice sheet retreats from the shelf break, the supply of glacial debris diminishes and small volumes of debris are supplied to the upper slope. With continued retreat of the ice sheet towards the inner shelf, a significant release of meltwater dilutes the remaining debris, transporting it downslope as turbidity currents.

Turbidites from these flows are deposited over the earlier debris units. Based on this interpretation, the glacial history of the margin was inferred and corroborated with the knowledge of repeated shelf edge glaciations during the Weichselian and shown in Figure 4.30.

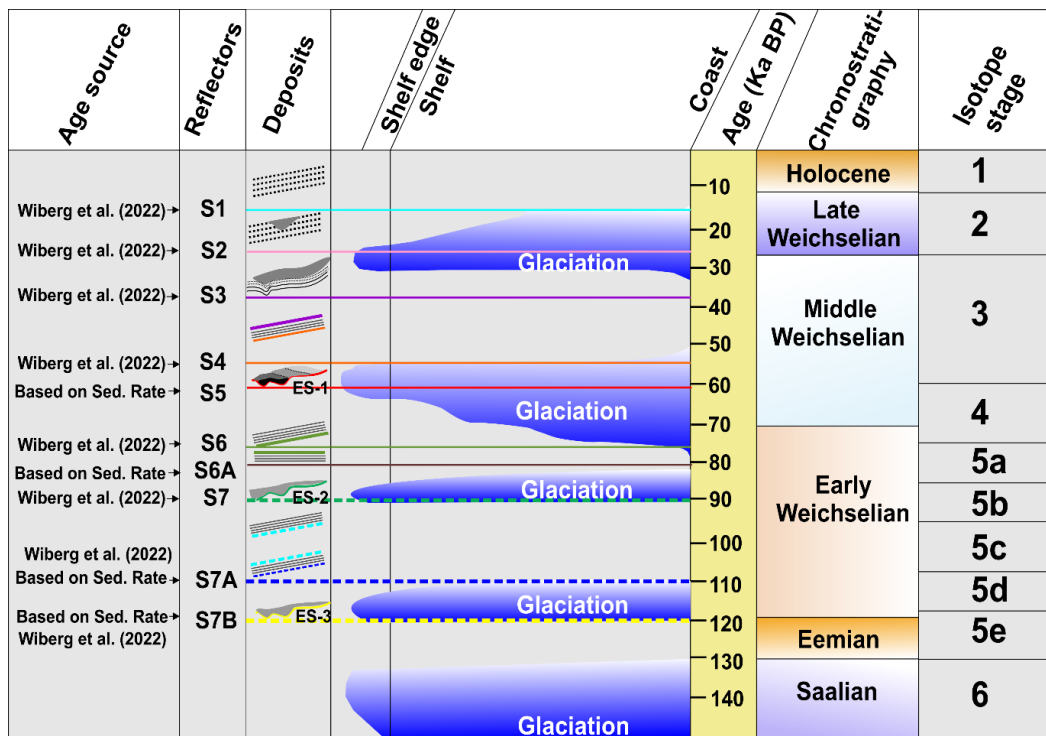


Figure 4.30. Correlation of SYSIF seismic horizons with different Weichselian shelf-edge glaciations. Information for coastal glaciation is obtained from Mangerud et al. (1998), Eccleshall et al. (2016), and Wiberg et al. (2022). Information on Saalian glaciation is from Eccleshall et al. (2016).

Early Weichselian 120–90 ka (SU-9, SU-8, SU-7, and SU-6):

The base of unit SU-9, ES-3 shows V-shaped indentations and is covered by debris materials (SU-9) and stratified deposits (SU-8) in the vertical succession. The indented surface is likely a product of ploughing by deep-keeled icebergs. These icebergs are presumed to have originated from ice-streams situated farther south and subsequently carried by the northward flowing West Spitsbergen Current. Debris materials, which descended downslope from a grounded ice sheet at the shelf break of the interfan region, filled in these indentations. Alternatively, the indentations may have resulted from erosion associated with debris flow. Owing to constraints in tracking ES-3 across the study area, it is challenging to make inferences regarding the lateral continuity or extent of the indentations. Specifically, to determine whether they are slope-perpendicular gullies or nearly slope-parallel iceberg ploughmarks. ES-3 laterally correlates with the reflector S7B (~ 120 ka). Unit SU-9, which is bounded by ES-3 and S7A (~ 110 ka) is corroborated with a period of shelf edge glaciation during ~ 120-110 ka. SU-8 (~ 100-90 ka) shows stratified turbidites, deposited by meltwater-driven turbidity currents that likely formed during the glacial retreat from the shelf break to

the inner shelf, when meltwater supply was abundant. This interpretation is in accord with the ice sheet reconstructions by Mangerud et al. (1998) and Ingolfsson & Landvik (2013) that suggest that the shelf edge glaciation outside Isfjorden ceased around 108 ka.

The base of seismic unit SU-7 shows a prominent erosion (ES-2) and correlates with the slope reflector S7 (~ 90 ka). ES-2 shows V-shaped indentations, likely representing plough marks created by icebergs or erosion caused by debris flow. The presence of chaotic debris flow in SU-7 on the uppermost slope provides evidence of debris supply by a grounded ice sheet ~ 90 ka. The reflector S6 within the overlying turbidite unit SU-6 is dated to 75 ka, while the top bounding reflector of SU-7 (S6A reflector) is dated to approximately 80 ka based on the average sedimentation rate. The stratified turbidites of unit SU-6 suggest a subsequent shift in the depositional environment. Alexanderson et al. (2018) identified interstadial deposits at the Kongfjordhallet in the outer Kongsfjorden, dating them to 85 ka during a period of relative ice-free shelf conditions. Based on this data and the nature of seismic facies, the well-stratified reflectors of Unit SU 6 (bounded between S6A and S5) likely represent turbidite sediments of deglacial origin (85–75 kyr).

Early Mid-Weichselian 75–54 ka (SU 5B and SU 5A):

The base of unit SU 5B is marked by an erosional surface ES-1. This surface cuts the underlying strata, including reflector S6 (~ 75 ka), but merges with reflector S5 (~ 61 ka). Additionally, ES-1 shows iceberg ploughmarks. In the slope of the interfan area, the ploughmarks are deep (maximum depth ~ 55 m), and mostly occur in the water depth interval of 410–520 m, while a few are found around 600–700 m depths. The infilling glaciogenic debris deposits of Unit SU 5B suggests the sediments were released from a grounded ice sheet on the upper slope. The overlying unit SU 5A, bound by S4A at its base and S4 (~54 ka) at its top, shows a transition from chaotic facies to stratified beds with intermittent interbedding of debris in water depths 450–600 m. This transition implies the dilution of glaciogenic debris as meltwater content increases. Units SU 5B and 5A represent deposition under the shelf edge glaciation spanning 61–54 ka, equivalent to MIS 4. Wiberg et al. (2022) proposed shelf-edge glaciation between 70–54 ka, since glacial debris flows are observed after ~ 75 ka; however, based on SYSIF data analysis, a peak in iceberg calving occurred around 57–61 ka, suggesting that ice grounding likely preceded the peak period of ice sheet calving and an enhancement in ice flux during ~ 57–61 ka. This period is considered as the most extensive Weichselian glaciation in the Barents and the Kara Seas prior to the Last Glacial Maximum (LGM) (Svendsen et al., 2004).

Mid-Weichselian (54–38 ka) unit SU-4:

The unit SU 4 bounded by reflectors S3 (~ 38 ka) and S4 (~ 54 ka), exhibits moderate-to-high-amplitude stratified reflections (SF 2) in water depths < 450 m. The glacial debris lobes are conspicuously absent in these depths, except for a few isolated debris pockets. The moderate-amplitude stratified facies of this unit represents deposition from turbidity currents and meltwater plumes generated by an ice sheet situated further landward from the shelf break, resulting in a significantly reduced glacial debris supply. Although the SYSIF lines could not resolve this unit at water depths < 420 m, the sediments retrieved from shallow core at the uppermost slope suggests a transition from sand and pebbly clay to overlying silty clay (Bohrmann et al., 2017, Auxillary Figures 4.1 and 4.2) This indicates a shift in the depositional environment, from debris flow to turbidity currents at the uppermost slope as the ice sheet started to retreat from the shelf break and debris supply dwindled.

Middle to Late Weichselian (38–24 ka):

The seismic unit SU 3 in the shallower depths (< 450 m water depth) consists of chaotic facies (CF1) and lenticular facies (CF2) interpreted as glacial debris lobes indicating the reestablishment of grounded ice sheet at the shelf break supplying glacial debris on the upper continental slope. This observation is corroborated with the MSM57 gravity cores (Bohrmann et al., 2017, Auxillary Figures 4.1 and 4.2), which show the predominance of pebbly and gravelly clay related to poorly sorted debris flow materials. Based on seismic facies and information from gravity cores, the SU 3 unit represents glacial sediments deposited at the maximum extent of the grounded ice sheet during the Last Glacial Maximum (LGM).

Late Weichselian (24–15 ka):

The moderate-to-high-amplitude SF2 SF1 facies in SU2 (24–15 ka) indicates coarse-grained sediments, such as sandy/silty turbidites in water depths > 450 m deposited by turbidity currents cascading downslope. The debris influence started to diminish on the uppermost slope as the ice sheet started to retreat from the shelf break. The turbidity currents are likely caused by a substantial influx of meltwater.

Late Weichselian and Holocene (< 15 ka):

The youngest unit SU1, which consists of low-amplitude stratified SF1 facies, indicates predominance of hemipelagic sediments. The topmost retrieved sediments in the gravity cores (Figure 4.2) on the uppermost slope consisted mainly of silty clay to clay, with occasional occurrences of pebbly clay (Bohrmann et al., 2017, Auxillary Figures 4.1 and 4.2).

In some cores, the clay layer is absent possibly due to winnowing effect of the West Spitsbergen Current. Slubowska-Woldengen et al. (2007) suggested that an inflow of warm Atlantic Water began during the Bølling–Allerød interstadial, commencing around 15 ka, with a more pronounced impact from ~ 14.3 ka. According to these assessments, it becomes evident that the West Spitsbergen Current started to rework the sediments on the seafloor shortly after the onset of the retreat of the Last Glacial Maximum (LGM) ice sheet.

4.5.2. Depositional processes in the shelf and slope

4.5.2.1 Depositional processes in the shelf:

The shelf was shaped by repeated cycles of glacial expansion and the accumulation of glacial materials from the Svalbard ice sheet. As the glacier advances, the immense pressure from the ice causes the deformation of the underlying sediments and the creation of an erosional unconformity as easily erodible sediments are removed. Thrust sheets on the shelf result from the bulldozing effect of the advancing glacier (Inset 2, Figure 4.4).

During maximum glacial advancement, the glacier is grounded at the shelf break and forms a submerged till delta at its terminus (Vorren and Laberg, 1997). The deposits within the till delta are inherently unstable and prone to mass failures. Various factors can trigger these failures, including excess pore water pressure from rapid sedimentation, slope oversteepening beyond the angle of repose, and fluid pressure from seismic activity like earthquakes. These mechanisms contribute to the failure of the deposits, resulting in the sliding of glaciogenic sediments. Following slope failure, the remolding of porous and saturated materials leads to slumping. The slumped materials cascade down the prograding glide planes and evolve into debris flow lobes with increased internal deformation and turbidites.

4.5.2.2 Debris flow processes in the slope

Seismic units SU 5A, SU 5B, and SU 3 are dominated by debris flow deposits on the uppermost slope and shelf break. These debris flows originated at the shelf break near the till delta in water depths of 290–300 m and cascaded down the slope to depths of 450–500 m. The ability of debris flows to travel long distances depends on the presence of a thin, lubricating water layer at the base of the flow (Sobiesiak et al., 2018). This water layer reduces friction between the debris flow and the seafloor, preventing the transfer of shear stress from the flow to the substrate. As the front of the debris flow moves through the surrounding water, a combination of front pressure and lift forces allows a layer of water to

intrude beneath the front of the flow. This process is known as hydroplaning (Mohrig et al., 1999). Hydroplaning can cause the front of the flow to accelerate, leaving behind a stretching zone. This stretching can detach the hydroplaning head, forming an outrunner lobe.

Unit SU 5B contains chaotic debris flows that lie above an irregular and heavily grooved erosional surface (ES-1). V-shaped indentations (15–55 m deep) at the base of unit SU 5B below water depths of 400–550 m suggest that icebergs ploughed the seafloor (Stage-I in Figure 4.31). When debris flows cascade down the slope from the shelf break, these indentations trap the debris (Stage-II in Figure 4.31). The debris cannot advance beyond the indentations until the deep indentations are filled in. The trough-shape of ES-1 is indicative of erosion. Debris flow modelling indicates that the ploughmarks can trap debris cascading downslope, causing the flow to be channelled along the contour-parallel groove (Figure 4.26). The debris flow can also erode the substrate. The depth of substrate erosion depends on factors such as the strength of the flow and the shear stress at the base. Erosion occurs when the shear stress exceeds a certain threshold value. The loss of basal lubricating water layer can lead to enhanced coupling between the debris flow and the substrate leading to free-slip condition. Without a lubricating basal water layer, the shear stress at the base of the debris flow is transferred directly to the seafloor, causing erosion and incorporating substrate material into the flow. Hiscott and Aksu (1994) showed that the snout of a large debris lobe can be very erosive. The interaction of debris flow with the substrate and the transfer of mass from the substrate to the debris flow caused by entrainment is critical for the change in momentum of the flow. Incorporation of substantial amounts of substrate material, such as clay may increase the cohesive behaviour, enhancing viscosity and reducing mobility of the debris flow. For example, Talling (2013), showed 1-m thick debris flow over 0.1-degree gradient slope will increase in bulk flow velocity as the mud content increased from 1 to 6 %, however, flow velocity slackened with 10 % clay volume as viscosity becomes a dominant factor.

Unit SU 5B, below 400–500 m depth, reveals a pattern of waning debris flow energy. In Stage-III (Figure 4. 31), numerous debris flow lobes developed and accumulated in a vertically aggrading manner while the ice sheet was grounded at the shelf break. This filling pattern is a plausible explanation for the observed debris-stacking pattern in unit 5B, illustrated in Figure 4.20. The stacked debris lenses act as obstacles, influencing antecedent flow behaviour. Subsequent low volume cohesive debris flows (due to dwindling debris supply) get trapped behind the frontal obstruction. This results in a unique "retrogradational

stacking pattern" as the younger debris stack up upslope progressing from deeper depths to shallower depths (Stage-IV in Figure 4. 31).

As debris flow cascades downslope, it can become diluted and transform into high- and low-density turbidity currents. The stratified layers of unit 5A in water depths 500 – 700 m are juxtaposed against the chaotic facies in 300-450 m water depths. The presence of stratified facies in unit 5A, in the distal part of the debris, marks the transition to well-sorted turbidite sediments. Occasional channel facies within unit 5A indicate the channelization occurred due to an increase in meltwater. The transition from debris flows to turbidites in a sedimentary environment can occur due to a decrease in the ratio of sediments and water. This can happen, if the sediment supply decreases as a result of change in sediment source or glacial erosion potential. Debris flows are primarily controlled by gravity and may have limited mobility due to high cohesive strength. More water content the flow becomes more fluid. The high clay content of the debris flow gives it greater shear strength compared to turbidity currents. However, if the hydrodynamic conditions change, such as an increase in meltwater content, it can trigger the transformation of debris flows into turbidity currents. The internal processes within debris flow can also lead to flow transformation. For example, the hydroplaning debris flow snout is sometimes thinned and stretched. The stretching zone is also prone to interaction with the basal lubricating water layer, leading to cracks at the base of the flow due to increased pore pressure (Ilstad et al., 2004). These cracks allow water to penetrate the debris flow, diluting its base and the head of debris flow, reducing its shear strength, and forming a turbidity current (Hampton, 1972). These currents exhibit Newtonian fluid behaviour and have zero yield strength. The higher energy and greater fluid mobility associated with turbidity currents enable the transport of finer-grained sediments over longer distances.

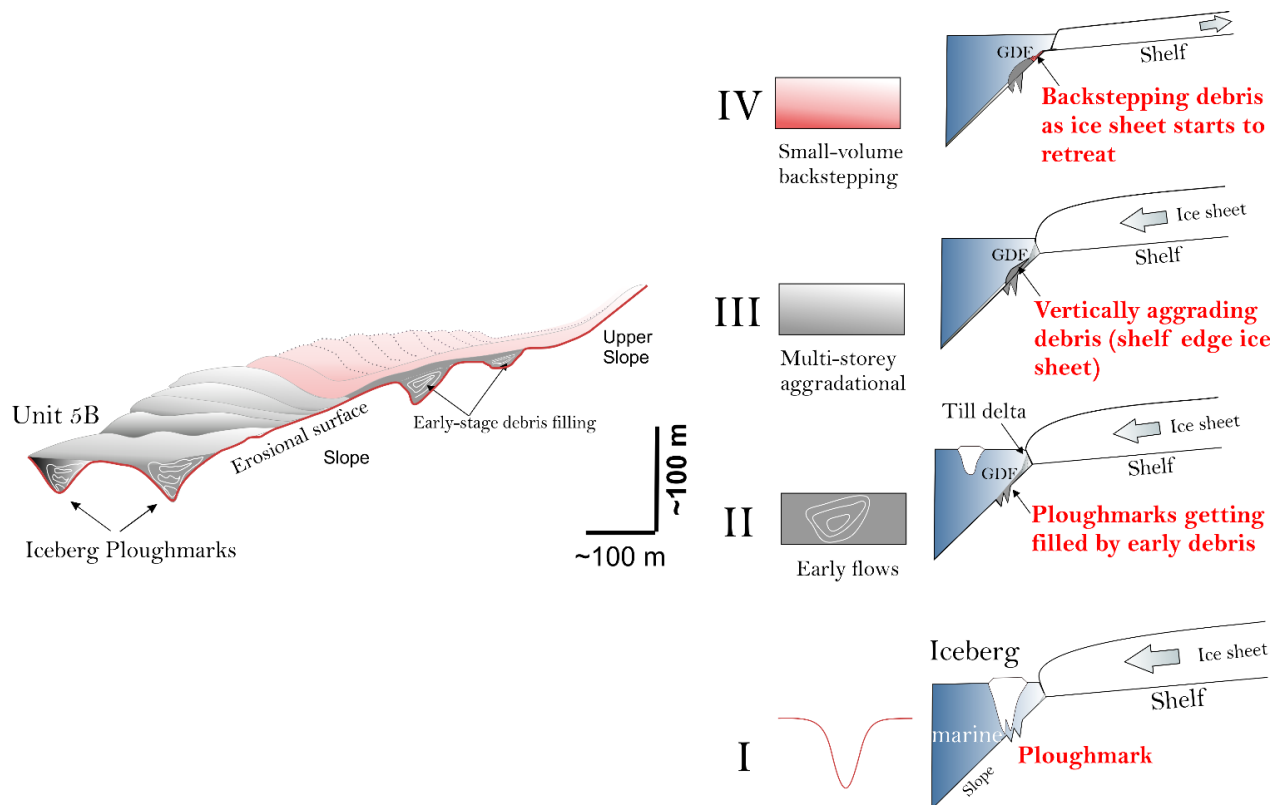


Figure 4.31. Schematic cartoon illustrating the patterns of glacial debris filling indentations caused by iceberg keels on the upper slope in relation to the position of the ice sheet at the shelf break. Stage-I depicts ploughmarks formed by deep-keeled icebergs during a major calving episode. At Stage-II, early glacial debris flows (GDFs) originated from the shelf edge, cascading downslope and filling the grooves. During Stage-III, multiple debris flow lobes emerged and stacked up in a vertically aggrading manner when the ice sheet remained grounded at the shelf break. This filling pattern could be applied to explain the debris-stacking pattern observed in unit 5B, as shown in Figure 4.20. The retreat of the ice sheet from the shelf break during Stage-IV resulted in a decrease in debris supply, leading to the deposition of smaller volume of glacial debris lobes on the upper slope.

4.6. Conclusions

The interpretation of the JR269B air gun data has yielded valuable stratigraphic insights, revealing the extent of the upper regional unconformity (URU) from the outer shelf to the shelf break region in the interfan area for the first time. The post-URU geological succession indicates accretion of glacial materials that extended the shelf break to its present-day position. Sarkar et al. (2011) suggested that a topography-constrained ice stream switched from northwest Svalbard to the Kongsfjorden between 1.5 and 0.99 million years ago. The age constraints from the MeBo 126 drill hole and reinterpretation of existing seismic lines

indicate that the ice stream switching event occurred approximately 1.2 Ma ago, coinciding with the Mid-Pleistocene Transition. At that time, the increase in ice sheet thickness exceeded the topographic barrier created by Prins Karl Forland, enabling the westward flow of ice along the shortest and steepest path, thus forming the Kongsfjorden cross-shelf trough. In the interfan region between Kongsfjorden and Isfjorden cross-shelf troughs, the ice sheet expanded to the shelf break approximately 1.2 million years ago, resulting in the erosion of pre-existing strata and the progradation of the margin, delivering glaciogenic debris at the shelf break. Following this, multiple cycles of deglaciation and shelf edge glaciations occurred, leading to erosion and the development of basinward advancing clinoforms. Examples include shelf edge glacial advancements during the Early Saalian period (~ 0.42 million years ago) and Weichselian period. The exact number of these cycles could not be determined in this study due to lack of age controls in the shelf region except for the Weichselian cycles.

From the high-resolution SYSIF seismic dataset, four advances of the ice sheet leading to shelf-edge glaciations during the Weichselian period were inferred: earliest advancement: ~ 120–110 ka, second advancement: ~ 90 ka, third advancement: ~ 61–54 ka, and the fourth advancement: ~ 24 ka. The third phase is relatively shorter compared to the known timing of 70–54 ka due to a prominent erosion and calving phase at ~ 61 ka, which erased evidence of the earliest part of glacial sedimentation. This phase also witnessed the maximum runout of glacial debris flows, reaching water depths of up to ~ 600 m. In contrast, the penultimate Last Glacial Maximum (LGM) shelf edge glaciation saw glacial debris flows extending onto the upper slope but limited to ~ 410 m water depths. The ages of the first two glaciations require confirmation, as their current estimates rely on sedimentation rates.

The findings from the marine domain help to resolve existing discrepancies associated with the Weichselian glacial history. Eccleshall et al. (2016) suggested that the expansion of ice sheets to the shelf break occurred during Marine Isotope Stage 5b (93–83 ka), while previous models did not consider this expansion. The new results support this interpretation, as there is evidence of plough marks and glacial debris filling at ~ 90 ka, indicating that the ice sheet expanded to the shelf break at this time. This interpretation also agrees with the observations of Wiberg et al. (2022), who suggested the presence of glacial debris as a result of ~ 90 ka shelf-edge glaciation. Eccleshall et al. (2016) also suggested that the ice sheet did not reach the shelf break during Marine Isotope Stage 4 (71–57 ka). The results from this work,

however, contradict this, suggest a peak in ice flux during a shelf-edge glaciation between 70-54 ka and at ~ 61-57 ka, during which deep-keeled icebergs were calved.

The lateral and vertical changes in seismic facies were used to infer the spatial and temporal changes in glaciomarine depositional environment that was linked to multiple cycles of shelf edge glaciations in the margin. During maximum glacial expansion to the shelf break, iceberg calving caused ploughing of the seafloor. A till delta formed by the grounded ice sheet at the shelf break during its maximum extent, supplied debris that cascaded downslope and deposited debris materials on the upper slope. Retrogradational stacking of smaller volume debris indicates the ice sheet started to retreat leaving the shelf break ice free. Influx of meltwater during glacier retreat, produced turbidity currents and turbidite deposition on the slope.

SYSIF seismic images and simulations based on RAMMS provide insights into glacial debris flow travel distance and evolution. High-strength flows occupied steeper slopes with shorter runout distances, while lower-strength flows travelled further and split into smaller lensoid lobes due to local slope variations. In terms of debris flow transformation, the seismic data provided evidence of the transition into turbidity currents. Stratified facies in the distal part of the debris marked the shift to well-sorted turbidite sediments. Potential factors for this transformation are a decrease in sediment supply, an increase in meltwater content, and internal processes within the debris flow, such as hydroplaning and stretching introducing cracks that allow water infiltration into the debris lobes.

4.7. Auxiliary materials

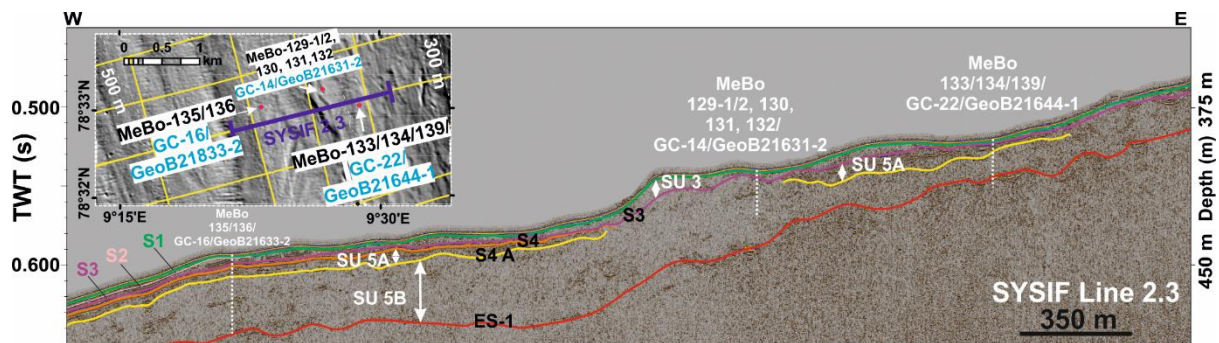


Figure S4.1. SYSIF line showing coring locations during MSM57 cruise at 445, 405, and 391 m depths (Inset map) (Bohrmann et al., 2017). Time-to-depth conversion was performed using a velocity of 1550 m/s.

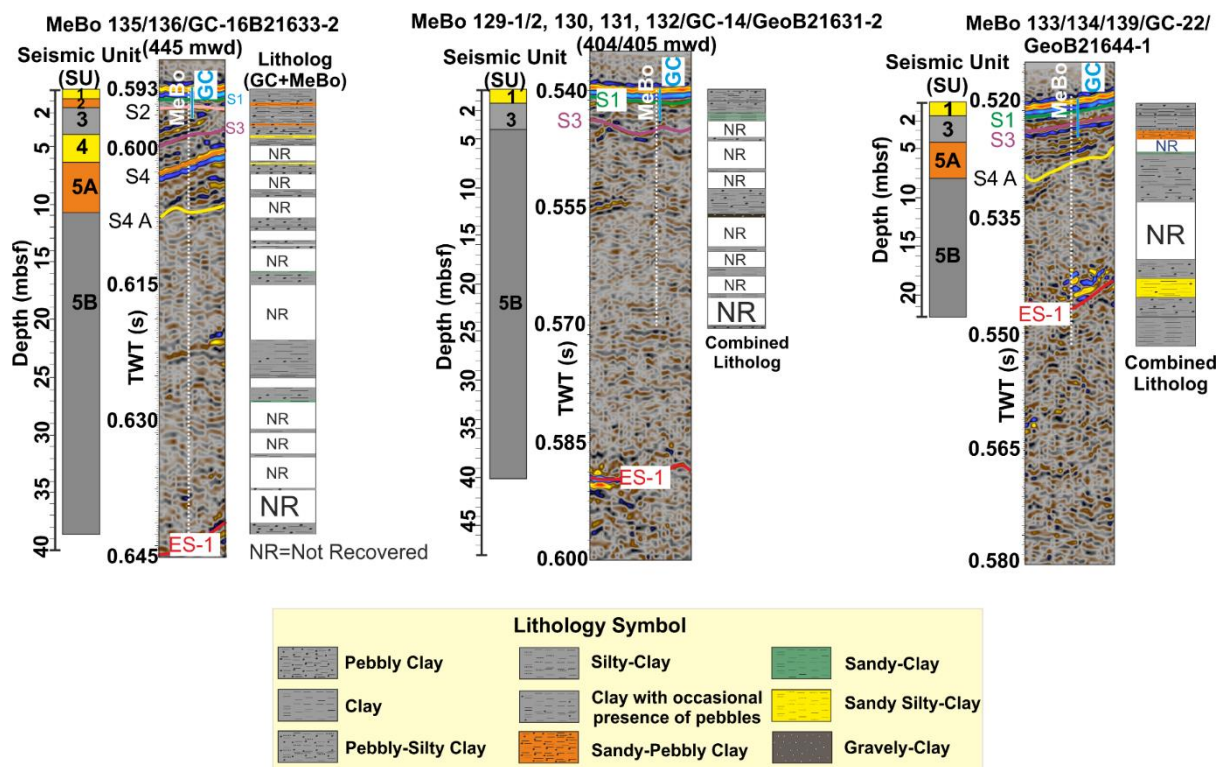


Figure S4.2. Lithological information derived from MeBo, gravity coring (GC) (locations in Figure S4.1) shown alongside the interpreted seismic section.

Chapter 5

Role of Methane hydrate dynamics and lithological heterogeneity in controlling fluid accumulation and migration below the seepage sites

5.1. Introduction

5.1.1 Motivation and Objectives

Extensive methane plumes have been observed along the upper western Svalbard continental margin, ranging from 80 to 415 m in water depth (Figure 5.1) (Sahling et al., 2014; Westbrook et al., 2009). The highest concentration is noted around the 400-m isobath. Initial interpretations by Westbrook et al. (2009) suggested methane emission from gas hydrate dissociation due to the recent warming of bottom waters. The dating of authigenic carbonates indicates that gas seepage at these locations has been active for several thousand years (Berndt et al., 2014).

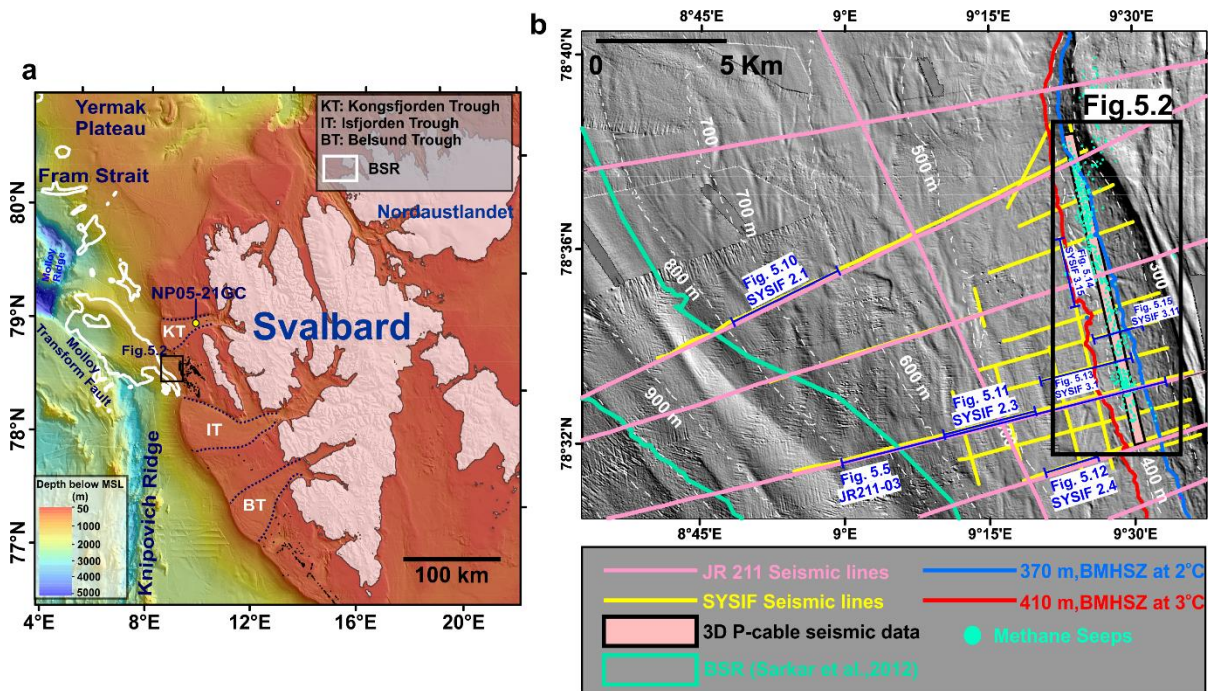


Figure 5.1. IBCAO bathymetry (Jakobsson et al., 2020) reveals glaciated continental shelf of western Svalbard. Several E–W trending troughs dissect the shelf and were carved out by the paleo-ice streams. These troughs are named Bellsund (BT), Isfjorden (IT), and Kongsfjorden (KT) troughs. The bathymetry also reveals the ridge and transform fault system. The BSR map was constructed using multiple 2D seismic lines collected during various marine seismic surveys and compiled by Minshall et al. (2020). The sediment core NP05-21GC (yellow dot) is located at 327 m water depth on the Kongsfjorden shelf west of Svalbard (Rasmussen et al., 2014). Gas seeps on the upper continental slope and shelf region were mapped by Sahling et al. (2014), Mau et al. (2017) and Ferre et al. (2020). **(b)** Location of JR211, SYSIF 2D seismic lines, 3D P-cable seismic cube and methane seepage points in the interfan area. The extent of BSR is also shown.

On the upper continental slope (UCS) in the interfan area between the Kongsfjorden Trough Mouth Fan (KTMF) and Isfjorden Trough Mouth Fan (ITMF), methane seeps occur within 300-400 m water depths, forming a narrow band (Figure 5.2a). This occurrence could stem from the thermal dissociation of methane hydrates as the intersection point of hydrate stability and seabed migrates downslope, influenced by bottom water warming (2 °C to 3 °C) over the past four decades (Figure 5.2b). However, the nearly linear trend might also result from structurally controlled gas migration or focused fluid flow conduits, like seismic chimneys. The scattering of gas seeps within the narrow band suggests subsurface heterogeneity as a key factor in controlling fluid flow. Sarkar et al. (2012), emphasized strong influence of stratal heterogeneity, such as glacial incisions and channels, on fluid flow. Bathymetric images reveal glacial ploughmarks and pits, locally influencing fluid migration (Sarkar et al., 2012). However, the lack of 3D seismic survey hinders the detection of these features in deeper parts, and the 2D seismic lines with gaps between them further limits understanding of fluid migration patterns, reservoir and seal distribution, and interconnections between reservoir elements.

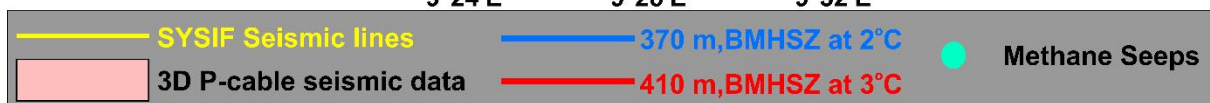
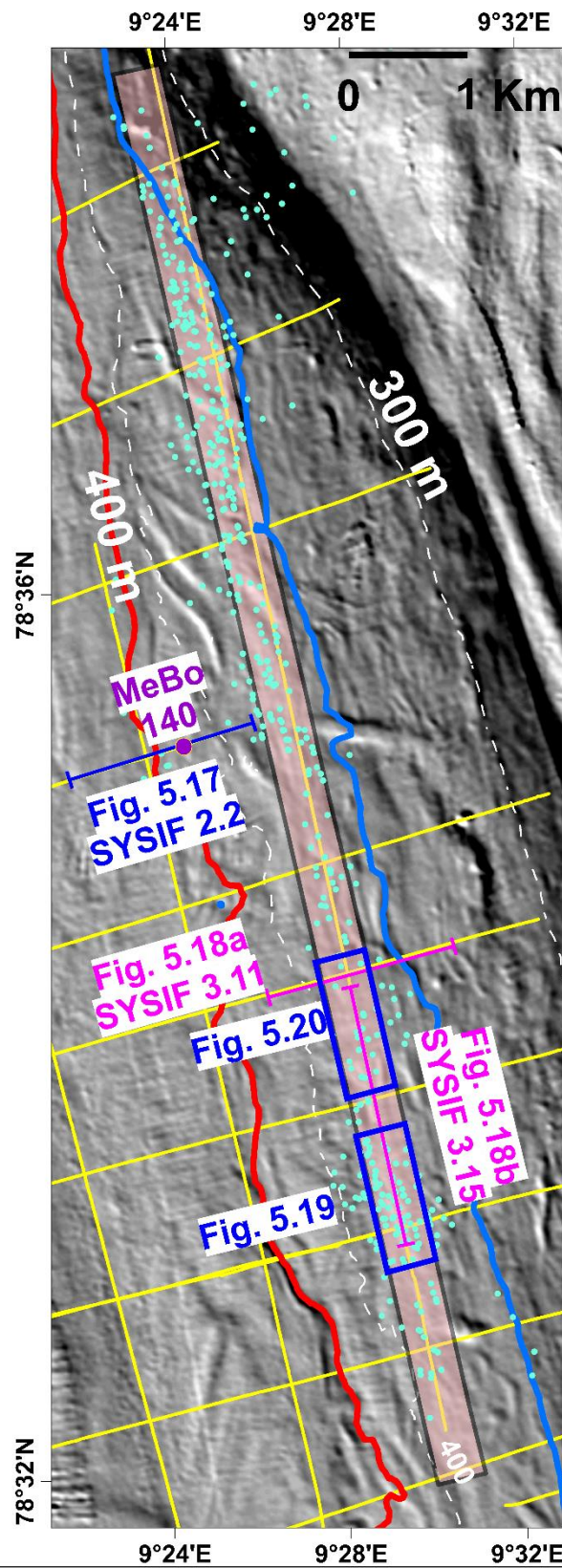


Figure 5.2. Shaded relief bathymetry map of the seepage zone in the interfan area. The methane seeps predominantly occur between the 370 and 400-m contours and result from hydrate dissociation due to an increase in bottom water temperature (BWT) from 2 °C to 3 °C (Westbrook et al., 2009). The upper limit of methane hydrate stability is at 370 m and 400 m water depths under BWT conditions of 2 °C and 3 °C, respectively. The northernmost seeps display a distinctive clustering along the shelf break. These seeps are attributed to lateral gas migration along prograding glacial clinoforms and are not linked to hydrate dissociation (Sarkar et al., 2012).

Chabert et al. (2011) proposed that lithology strongly influences gas hydrate occurrence, with glacially influenced UCS sediments hosting hydrates as inclusions up to 5 % due to limited availability of porous and permeable reservoirs. In contrast, deeper contourite sediments may host up to twice the hydrate amount, due to increased porosity and permeability of these sediments. Thus, understanding of lithological variability and in turn reservoir quality is crucial for quantifying methane hydrates in highly heterogeneous glacial sediments. Since methane is climate-sensitive, the concentration of methane hydrates, influenced by sedimentological properties within the hydrate instability zone, plays a crucial role in understanding the potential release of methane due to environmental fluctuations.

Gas can migrate along the base of methane hydrate stability and gets concentrated on the upper continental slope. 2D seismic lines reveal a bottom simulating reflection (BSR) up to 600 m and deeper; however, lateral tracking of the BSR in the upper slope is complicated by heterogeneity of glacial strata. The seabed wavelet, 5-10 m thick, and interference from the seabed and shallow glaciomarine strata further complicates identification of BSR. Higher resolution seismic data can aid in tracing the BSR, allowing accurate delineation at the feather zone of methane hydrate stability and providing insights into fluid flow behaviour influenced by hydrate dynamics. Methane flare density and release rates decrease in winter when temperatures are significantly lower compared to summer (Ferré et al., 2020). This implies that the recurring changes in bottom water temperature (BWT) seasonally contribute to the formation and dissociation of hydrates in the top 10 m below the seabed (Berndt et al., 2014; Veloso-Alarcon et al., 2019; Ferré et al., 2020). High-resolution seismic data can provide details of the shallow subsurface beds that facilitate hydrate and gas accumulation.

Here, I aim to improve our understanding of subsurface lithological heterogeneity that controls methane transport and accumulation of methane hydrate below the active methane

venting sites and re-assess the role of the methane hydrates in guiding fluid focusing. Using high-resolution SYSIF seismic, I investigate the subsurface lithological heterogeneity influencing fluid focusing, and hindrances in the interfan area. With the 3D P-cable data that provide continuous coverage beneath the methane seeps, I analyze the subsurface presence of fluid-storing glaciogenic reservoirs, their distribution, and orientation. Given the challenge of tracking the hydrate-related BSR in shallower depths ($< 600\text{m}$), I conducted numerical simulations employing coupled heat flow and mass transfer to determine the base of methane hydrate stability. These simulations span 11,000 years and incorporate ocean temperature and relative sea level (RSL) changes as model forcing. The primary goal of these simulations is to estimate the depth of the Bottom of the Methane Hydrate Stability Zone (BMHSZ) at the upper continental slope off west Svalbard, within water depths ranging from 370 to 600 m.

5.1.2 Study area

The western Svalbard margin experienced significant ice sheet movements in the Pliocene and Pleistocene epochs, impacting sea levels and sediment dynamics (Landvik et al., 2008). Ice streams shaped the Kongsfjorden, Isfjorden, and Bellsund cross-shelf troughs (KT, IT, and BT) (Figure 5.1), with the current shelf break marking the ice sheet's maximum extent (Knies et al., 2009; Solheim et al., 1996). This study focuses on the inter-fan region between the Kongsfjorden and Isfjorden cross-shelf troughs (Figure 5.2a).

Sarkar et al. (2012) used multi-channel reflection seismic data to identify the gas hydrate BSR beneath the upper continental slope (UCS). The BSR acts as a boundary between the presence of gas hydrates above it and free gas below (Shipley et al., 1979). Notably, there is no evidence of a BSR in water depths shallower than 600 m, with seismic data suggesting the existence of gas accumulations in the depth range of 400 to 700 m (Sarkar et al., 2012). This finding is corroborated by identifying low-velocity zones and bright spots with negative polarity within the upper 75 m of sediment beneath the seafloor by Chabert et al. (2011). Additionally, regions with notably high electrical resistivity likely indicate the presence of hydrates within their stable range (Goswami et al., 2016).

The sediments in the shelf and upper continental slope (UCS) from Pliocene–Pleistocene epochs primarily originate from glacial processes. Marine hemipelagic sediments are interbedded with layers of glacial debris flows on the upper slope (Sarkar et al., 2011, 2012). The MeBo drilling experiment revealed a mixture of hemipelagic sediments and poorly-sorted glacial debris with a diverse range of particle sizes from clay to sand and gravel to pebble-sized clasts in varying proportions (Wallmann et al., 2018). During deglaciations, the

outflow of meltwater triggered deposition of plumites. On the steeper slopes, these plumites could get destabilized and transported by turbidity currents descending the slope, depositing sediment rich in sand and silt (Vorren et al., 1998).

The contrasting lithology plays a crucial role in shaping fluid dynamics beneath methane seepage regions. Gas flow encounters resistance in glaciogenic debris, while lateral fluid migration is predominant through sandy and silt-rich permeable sediments. Vertical flow occurs through fractures within poorly layered and low-permeability glaciogenic sediments, as discussed by Haacke & Westbrook (2006), Haacke et al. (2009), Sarkar et al. (2012), and Thatcher et al. (2013).

5.2. Data and methodology

5.2.1 Seismic Data

The survey JR269A in 2011 used a deep-towed high-frequency SYSIF (SYstème Sismique Fond) system that used a Chirp signal in the high-frequency bandwidth of 220–1050 Hz and collected a dense grid of seismic lines over the gas seepage area west of Svalbard (Location shown in Figure 5.2a). The source was towed 100–150 m above the seabed in the inter-fan region. The SYSIF data can help to resolve the shallow structures up to 250 ms below the seabed. The vertical and lateral resolutions of this data are 1 and 3 m, respectively, providing unprecedented subsurface details. Processing the SYSIF data comprised correcting the source-receiver geometry for each shot, source deconvolution, and migration (Ker et al., 2014).

The Centre for Arctic Gas Hydrate, Environment, and Climate (CAGE) conducted a 3D P-cable seismic survey in the slope region with reported methane seeps (380–395 m water depths) off west Svalbard (Location shown in Figure 5.2b). Multiple seismic lines were acquired using an array of solid-state streamers, providing extensive coverage with closely spaced intervals. The cross-cable was 223 m long and connected 14 streamers with an average separation of 12.5 m. The streamers, each 25 m in length, featured eight hydrophone groups and had analog-to-digital data conversion modules. The receiver spacing in each streamer was 3.125 m. The seismic source was a mini-GI gun with a total volume of 0.49 l.

The seismic dataset comprised six in-lines with a line spacing of 60 m, covering a 5-km² area (12.8 km long and 0.3 km wide). The acquisition targeted an area known for consistent methane flares over many years. Data pre-processing steps included navigation data sorting, 3D geometry setup, CMP binning, band-pass filtering, true amplitude recovery, gap filling by

interpolation, normal moveout correction at a constant 1500 m/s velocity, stacking, and post-stack 3D F-k migration.

5.2.2 Seismic attributes

The 3D seismic data analysis involved a comprehensive examination of subsurface geological features. Seismic attribute analysis was employed to unravel crucial aspects, including lithological variation, fluid distribution, frequency anomalies, and subtle structural discontinuities like fractures. Table 5.1 shows the seismic attributes used in this study, and examples are illustrated in Figure 5.3.

Table 5.1. List of seismic attributes and their applications in the study.

Seismic attribute name	Applied for
Similarity	Distinguishing the homogenous and heterogeneous lithology (e.g., chaotic diamicton)
Sweetness	Determining and highlighting the reservoir beds containing hydrocarbon
Variance	Determining the occurrence of fractures
Curvature	Determining the undulations in the strata

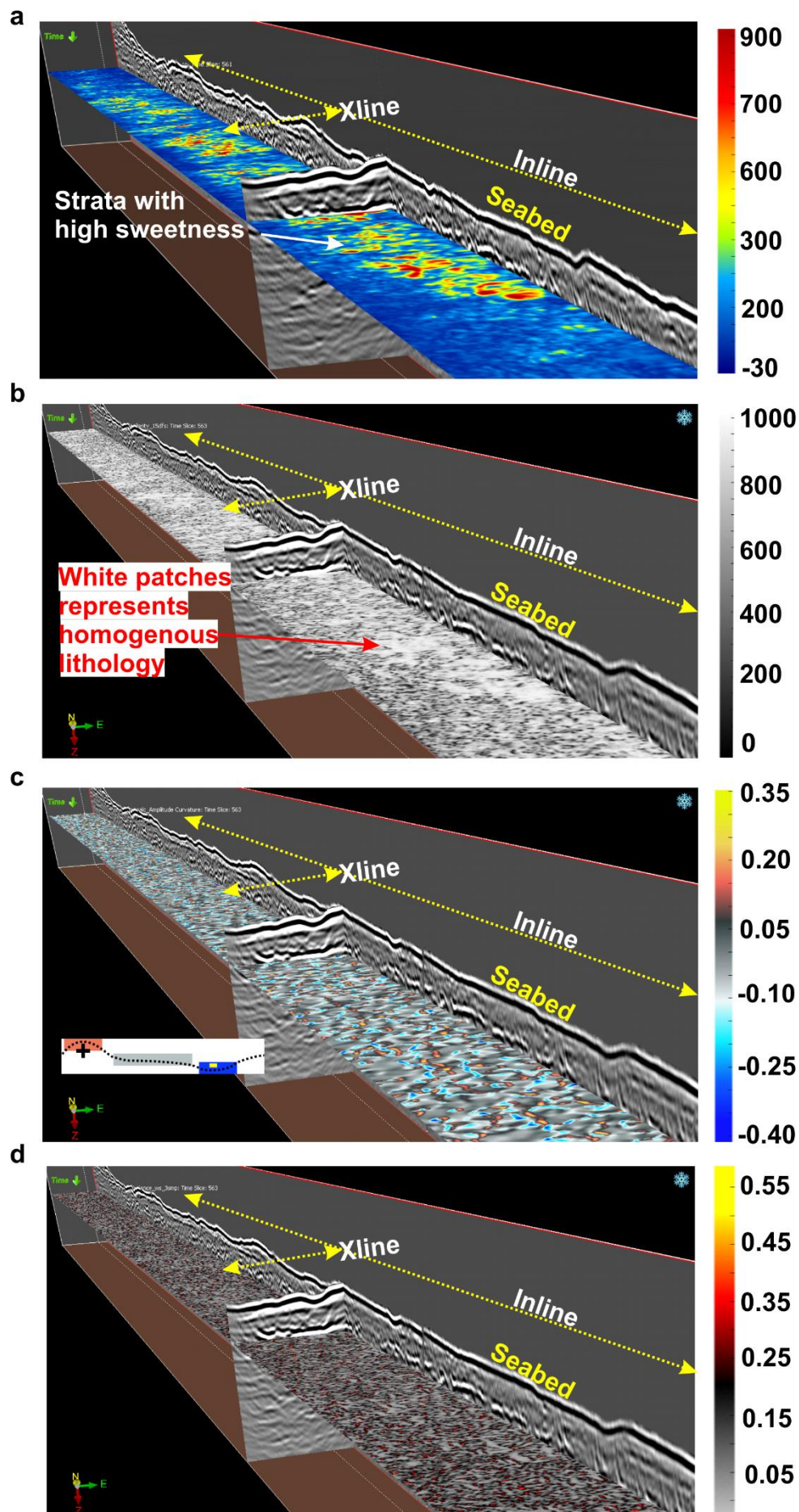


Figure 5.3. A time slice of the seismic attributes is shown along with the chair cut of the 3D seismic cube below the seepage site. **(a)** The sweetness attribute shows the zone of bright spots. **(b)** Similarity attributes distinguish the homogenous (bright white) and heterogenous (grey) lithology. **(c)** Curvature attribute shows the undulation in the subsurface beds **(d)** Variance is used to delineate zones of fractures and discontinuities.

In addition to seismic attributes, spectral decomposition was utilized to verify the presence of gas. The RGB blend of 25 Hz, 40 Hz, and 85 Hz band frequencies derived from continuous wavelet transform (CWT) using a Ricker wavelet revealed distinct colour signatures in the glaciomarine deposit (Table 5.2).

Table 5.2. The RGB colour combinations along with the interpretation.

RGB Colour	Interpretation
White (Red + Green + Blue)	Gas, Carbonates
Cyan (Blue + Green)	Well-sorted preserved turbidite beds
Red	Gas shadow
Magenta (Blue + Red)	Fractures and lithological discontinuity, e.g., base of stratified beds, separation between two debris lobes, fractures

5.2.3 Modelling approach for Methane hydrate dynamics

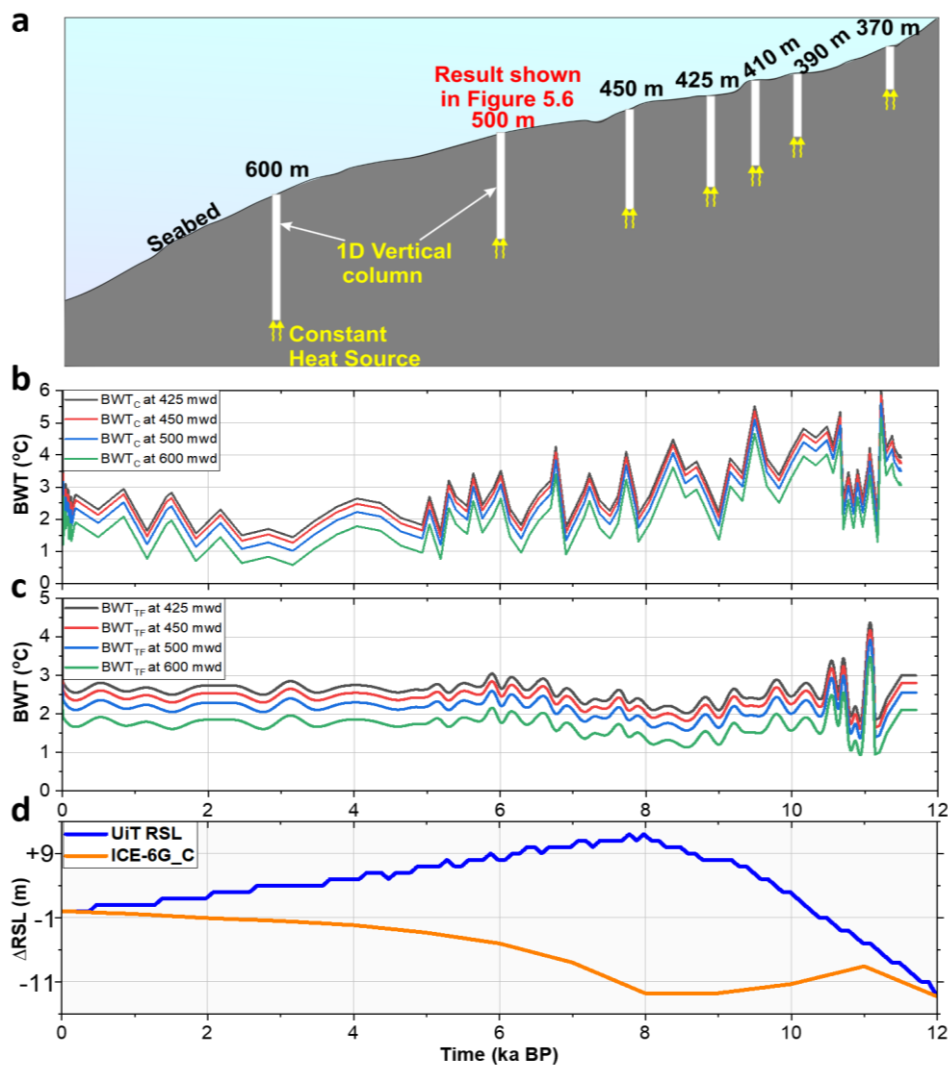
I conducted 1D numerical simulations for water depth ranging from 370–600 m (Figure 5.4a) using coupled heat and mass transfer to ascertain the base of the methane hydrate stability zone (BMHSZ), using the Tough+Hydrate code. These simulations were run for 11,000 years, incorporating bottom water temperature (BWT) and relative sea level (RSL) changes as model forcing.

The BWT values were derived from Rasmussen et al. (2014), who reconstructed the BWT west of Svalbard over the past 11200 years. This reconstruction is based on two proxies: (1) benthic foraminifera $\delta_{18}\text{O}$ values (BWT_{C}) (Figure 5.4b) and (2) benthic foraminiferal census counts (BWT_{TF}) (Figure 5.4c) from the sediment core NP05-21GC, obtained at 327 m water depth on the Kongsfjorden shelf west of Svalbard (Figure 5.1). For deeper depths (425–

600 m), the BWT curve was scaled using the paleo-temperature curve reconstructed for the southeastern Svalbard margin by Rasmussen et al. (2007).

The values for RSL fluctuation were derived from two sources: (1) the University of Tromsø (UiT) model of Eurasian ice sheet change by Patton et al. (2017), and (2) the RSL change predicted using the global ICE-6G_C ice history model by Peltier et al. (2015) and Auriac et al. (2016) (Figure 5.4d).

Yearly changes in BWT and RSL were implemented in the model by adjusting the temperature and pressure in the top cell. The model considers the immediate response of the sediment column to RSL-induced pressure changes and incorporates the dynamics of hydrate dissociation and reformation.



(For figure caption see the next page)

Figure 5.4. Schematic representation of the 1D hydrate models at 370, 390, 410, 425, 450, 500 and 600 m water depths. **(a)** Time-varying pressure (RSL) and BWT values are applied as model forcings on the top while maintaining a constant heat source at the bottom. **(b)** Past variations in BWT derived from core NP05-21GC based on oxygen isotopes (BWT_C). **(c)** BWT determined using a benthic foraminifera transfer function (BWT_{TF}) (Rasmussen et al., 2014) for the period 11.7–0 ka and scaled according to paleothermal profile from Rasmussen et al. (2007). **(d)** RSL variation calculated from the University of Tromsø (UiT) ice load history and Earth rheology model (Auriac et al., 2016) (UiT RSL) and the ICE-6G_C scenario (ICE-6G_C RSL).

5.3. Results

5.3.1 Modelling the base of methane hydrate stability:

The presence of a bottom-simulating reflection (BSR) in the interfan region at a water depth of ~ 600 m (Figure 5.5) suggests the existence of methane hydrate in this area. However, there is no indication of a BSR in the 2D air gun seismic data at shallower depths (< 600 m water depths).

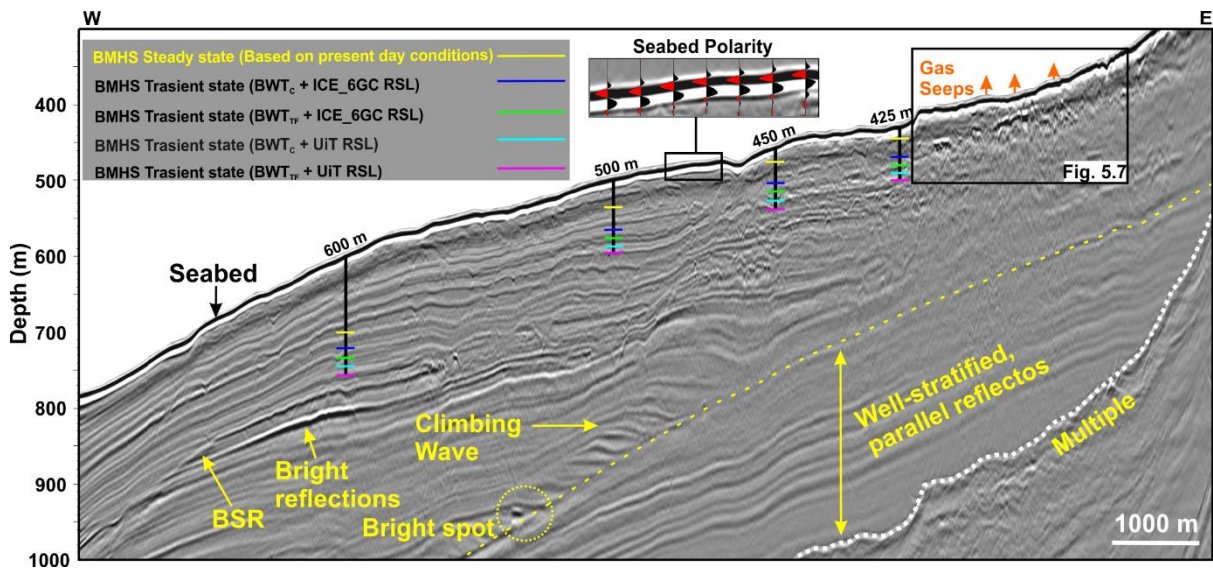


Figure 5.5. Seismic line JR211-03 collected using an air gun source shows the presence of BSR in the interfan region and modeled BMHSZ at water depths 600, 500, 450, and 425 m for the 4 model forcing scenarios. A BSR is seen in the mid-slope, where it cross-cuts the well-stratified, dipping reflectors but cannot be tracked to the uppermost slope (shallower than 600 m). Bright reflections below the BSR are due to gas. The line location is shown in Figure 5.1b.

Given the absence of a methane hydrate-related bottom-simulating reflection (BSR), numerical modelling was employed to determine the depth of the base of the methane hydrate stability zone (BMHSZ). One-dimensional (1-D) hydrate models were simulated for water depths of 450, 500, and 600 m under four plausible combinations of bottom water temperature (BWT) and relative sea level (RSL) changes: BWT_{TF} and UiT RSL, BWT_{TF} and ICE-6G_C RSL, BWT_C and UiT RSL, and BWT_C and ICE-6G_C RSL (refer to Figure 3.3, Chapter 3). Geothermal gradient values ranging from 50 to 60 °C/km (derived from the depth of BSR below the seafloor, Sarkar et al. 2012) were used to investigate the past behaviour of methane hydrates in response to both BWT and RSL fluctuations.

The simulation results at 500 m water depth under the BWT_{TF} and UiT RSL scenario are presented in Figure 5.6. The model was initialized at 11.2 ka before the present (BP) with 5 % and 4 % hydrate and gas saturation, respectively (Figure 5.6a), a steady geothermal gradient of 50° C/km (Figure 5.6c), and a pore-water salinity of 3.5 weight percent (wt. %) (Figure 5.6b) within the sediment column. Until 6 ka BP, the BWT rise and RSL fall caused the hydrate layer to dissociate from the base, producing methane (Figure 5.6a) and freshwater. The freshening trend in the sediment column led to a drop in salinity (Figure 5.6c). From 6 ka BP to the present, there was a very subtle change in the BWT, but the RSL fall continued. The upward-flowing methane and freshwater reformed hydrates. The present-day base of the methane hydrate stability zone (BMHSZ) at the end of simulation is ~ 100 m below the sea floor. This depth significantly exceeds the steady-state BMHSZ.

The steady state of hydrate stability is observed when the conditions beneath the seafloor remain relatively stable over an extended period, due to consistent BWT, RSL, and constant heat flux. This means that the BMHSZ has remained constant over an extended period. On the other hand, the transient state of hydrate stability is observed when there are fluctuations in temperature and pressure affecting gas hydrate stability. The theoretical steady-state BMHSZ at 500 m water depth, assuming present-day seabed temperature, is approximately 40 m below the seafloor (mbsf). However, the hydrate stability evolved over time due to changes in BWT_{TF} and RSL (UiT) leading to transient stability conditions. Although hydrate dissociates at the base, the reformation process further delays the upward migration of the hydrate stability base, resulting in a deeper transient state BMHSZ (100 mbsf in this case) compared to the steady state. The difference between the transient and steady-state BMHSZ for depths ranging from 425 to 600 m water depths for each model scenario is shown in Figure 5.6d and Table 5.3. This difference increases from 600 to 450 m water depth but

decreases at 425 m. Figure 5.5 displays both the present-day steady and transient state BMHSZ, considering a $50\text{ }^{\circ}\text{C km}^{-1}$ geothermal gradient, for water depths ranging from 600 to 425 m. The observed BSR at 600 m water depth is much close to the transient state BMHSZ predicted considering the UiT RSL and BWT_{TF} scenarios.

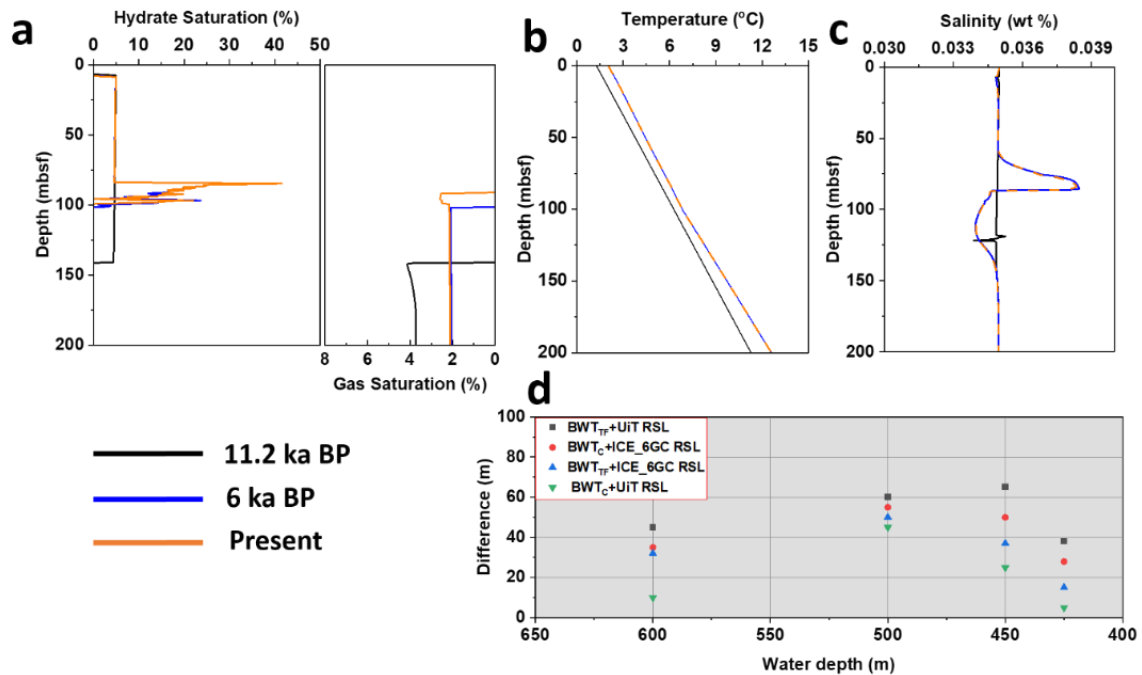


Figure 5.6. Simulation results showing the evolution of hydrate layer in response to BWT_{TF} and UiT RSL at 500 m water depth and comparison with the steady state BHMSZ. **(a)** Evolution of the hydrate layer thickness, saturation, and gas layer beneath it over the simulation period. Hydrate dissociated from the base (140 mbsf) and shrank to 100 m at 6 ka BP. The methane and freshwater released from the hydrate dissociation move upward in the hydrate stability zone, reforming a high hydrate saturation up to $\sim 40\%$ and locally changing the thermal conductivity, and delays the melting of the hydrate. **(b)** The temperature profile in the sediment column exhibits a kink at a depth of 90–100 mbsf, coinciding with elevated hydrate saturation observed at 75–95 mbsf due to reformation processes. **(c)** Salinity profile shows a negative anomaly below the present-day hydrate base, suggesting the release of meltwater. The depth ~ 75 –95 mbsf shows a positive anomaly, indicating the hydrate reformation process. **(d)** The difference between the steady and transient BMHSZ for different water depths 600, 500, 450, and 425 m and various BWT and RSL scenarios.

Trivedi et al. (2022) conducted comprehensive models of the long-term (past 11,000 years) evolution of methane hydrate stability based on plausible BWT and RSL scenarios. The results indicated a complete dissociation of hydrates approximately 50 years before the

present (BP). However, a hydrate reforming phase occurred in the late 1970s due to cooling of bottom waters and the presence of residual methane from the prior dissociation process. The evolution of methane hydrate stability field from 1978 to 2016 at water depths of 370 to 410 m is depicted along the seismic line JR211-03. The current hydrate stability, considering a BWT of 3 °C, outcrops at 400 m. Seasonal fluctuations in BWT strongly affect the top 10 m of the sediment column, leading to repeated episodes of shallow hydrate reformation and dissociation during colder and warmer conditions, respectively.

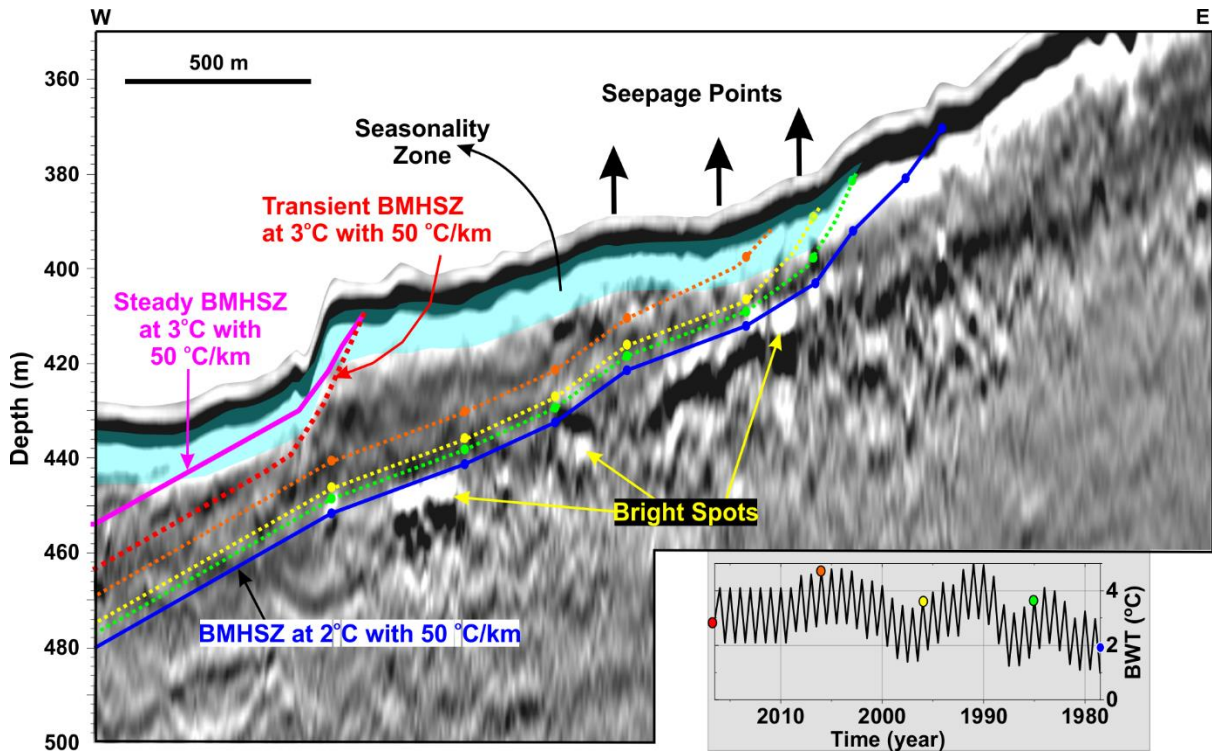


Figure 5.7. Seismic line JR211-03 illustrates the evolution of the BMHSZ in the interfan region at water depths ranging from 350 m to 430 m, considering BWT variations from 1978 to 2016 with seasonal fluctuations of ± 1 °C (inset). The base of hydrate stability outcrops at 370 m depth in the year 1978, while it outcrops at 400 m in the year 2016. The top 10 m below the seabed, highlighted in cyan, is susceptible to seasonal hydrate reformation.

Table 5.3. The depth of transient and steady state BMHSZ for 600-425 mwd.

Water Depth (m)	Model Forcing	Geothermal Gradient (°C/km)	Initialised BMHSZ Below seabed (m)	Transient BMHSZ Below seabed (m)	Steady BMHSZ Below seabed (m)
600	BWT _{TF} + UiT RSL	50	197	148	136
	BWT _{TF} + ICE-6G_C RSL		186	153	136
	BWT _C + UiT RSL		197	148	148
	BWT _C + ICE-6G_C RSL		186	140	137
500	BWT _{TF} + UiT RSL	50	141	99	22
	BWT _{TF} + ICE-6G_C RSL		166	87	12
	BWT _C + UiT RSL		140	85	22
	BWT _C + ICE-6G_C RSL		129	82	12
450	BWT _{TF} + UiT RSL	50	101	53	26
	BWT _{TF} + ICE-6G_C RSL		101	59	20
	BWT _C + UiT RSL		101	46	20
	BWT _C + ICE-6G_C RSL		92.5	53	22
425	BWT _{TF} + UiT RSL	50	74	53	18
	BWT _{TF} + ICE-6G_C RSL		74	59	10
	BWT _C + UiT RSL		74	46	10
	BWT _C + ICE-6G_C RSL		63.5	53	12

5.3.2 Glaciomarine stratigraphy:

The UCS in the interfan area has undergone multiple glaciations and deglaciations, resulting in varied facies alterations from well-stratified to chaotic. The glaciomarine stratigraphy of the interfan region is divided into five seismic units (refer to Table 5.4). The facies in these seismic units are broadly categorized into two types: stratified and chaotic. Stratified facies exhibits parallel, continuous reflections, while chaotic facies are incoherent, discontinuous, and semi-transparent to transparent reflections (see auxiliary table S5.1 for details). Figures 5.8 and 5.9 illustrate the seismic facies and thickness of each unit. The subsequent sections describe the glaciomarine stratigraphy for three zones divided based on the water depth.

Table 5.4. Seismic units and their bounding reflectors in the interfan area

Seismic unit (SU)	Top reflector	Bottom reflector	Age (ka BP)(See Chapter 4 for age)
SU 1	Seabed	S1	Present-15
SU 2	S1	S2	15 – 24
SU 3	S2	S3	24 – 38
SU 4	S3	S4	38 – 54
SU 5A	S4	S4 A	54 – 75
SU 5B	S4 A	S5 (Up to 635 mwd) ES (< 635 mwd)	

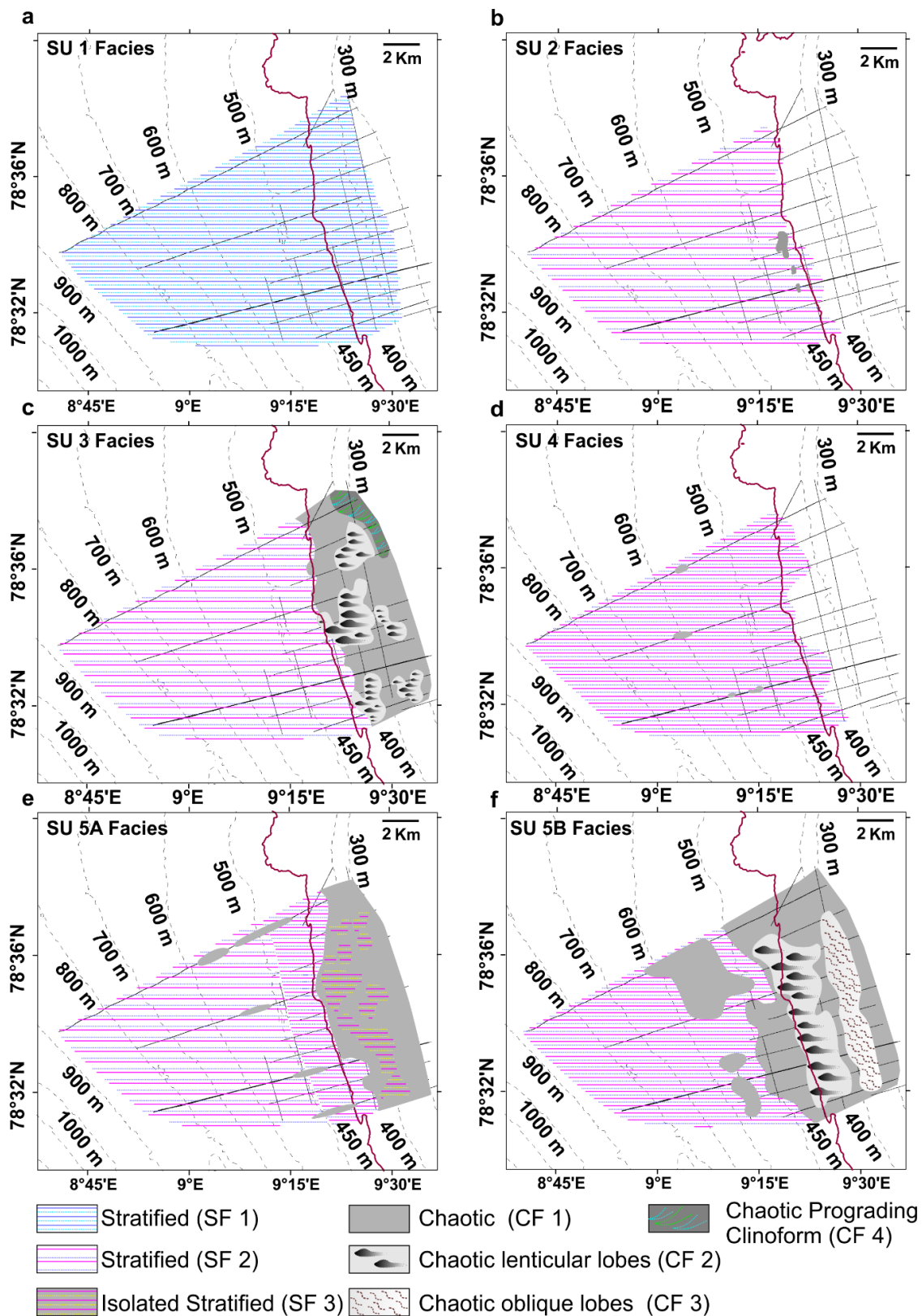


Figure 5.8. Spatial variation of different facies in stratigraphic units SU 1-5A-B. The stratified facies (SF) are predominantly present > 450 m except in SU 1 as it covers shallower depths up to 300 m. The chaotic facies in SU 3, SU 5A, and 5B indicates glaciogenic debris flows.

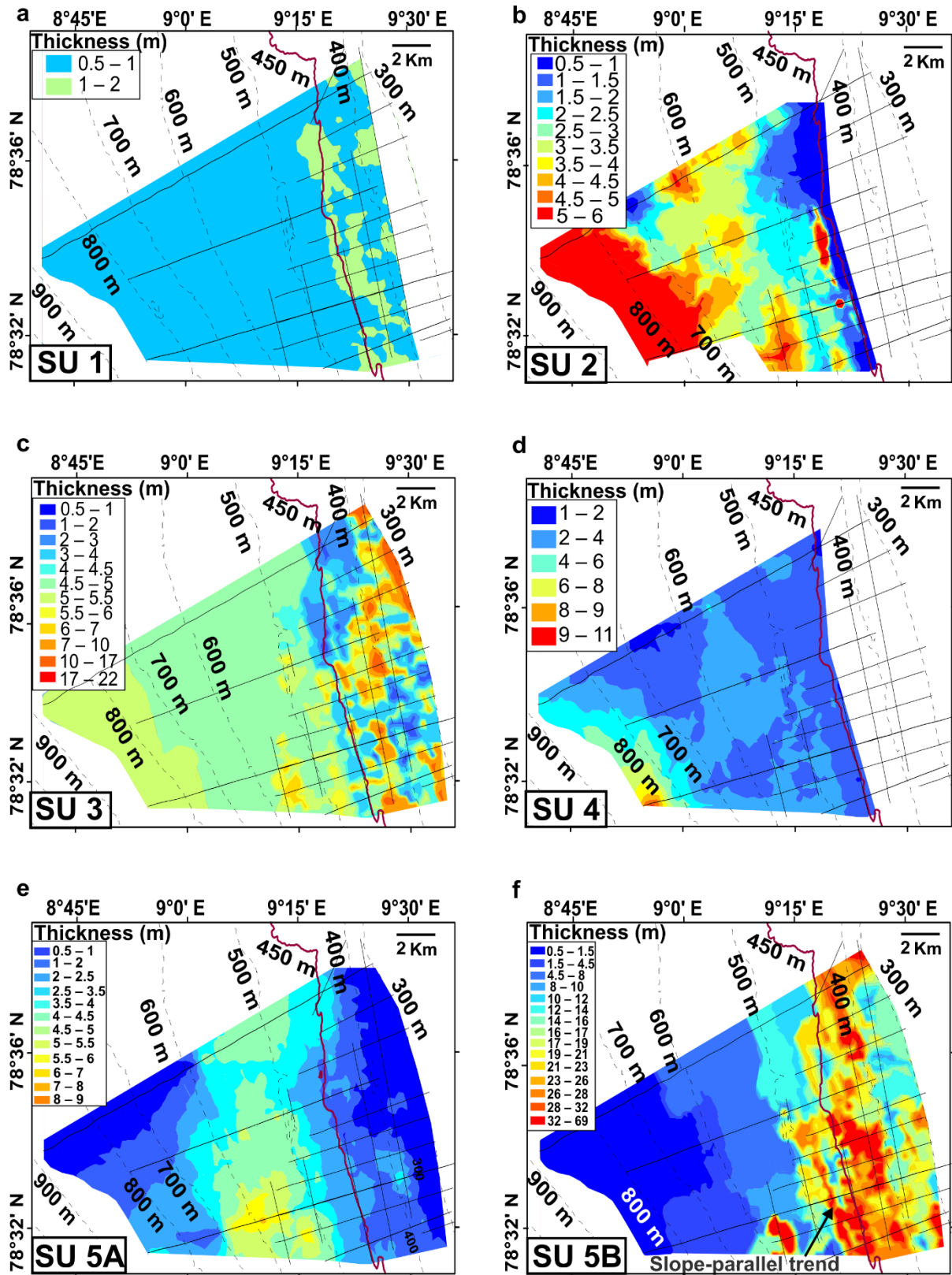


Figure 5.9. Map showing the thickness variations of stratigraphic units SU 1-5. Comparison of Figure 5.8 with this figure shows that the thickness of CF-2 glaciogenic facies in SU 5B was substantially high compared to SU 3.

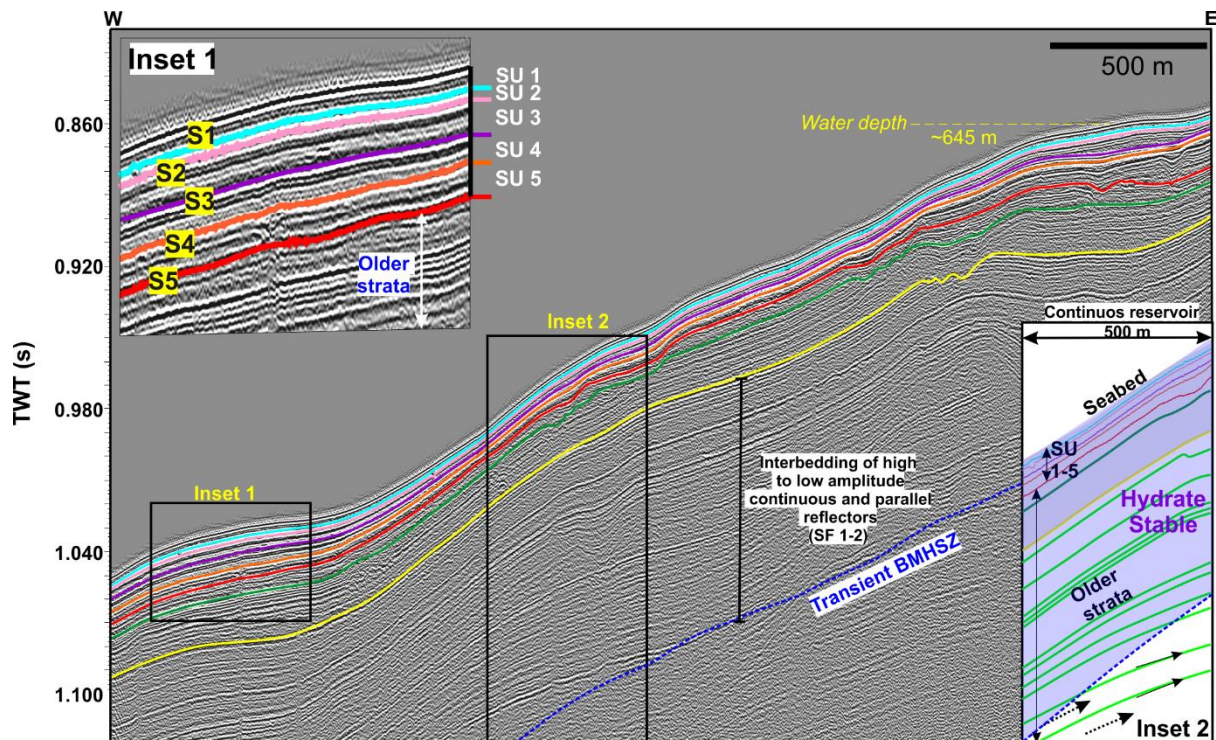


Figure 5.10. SYSIF line 2.1 shows the well-stratified beds in water depths > 645 m. All the units and the older strata show excellent lateral continuity (Inset 1 and 2). The transient BMHSZ (blue dashed line) is also shown. A BSR is not seen in this line. Gas will be trapped as hydrate in the zone above the BMHSZ, however, gas can laterally flow along the layers below the BMHSZ.

5.3.2.1 Seismic facies between 825–600 m water depths (Zone 1):

Within this specific depth interval, spanning all seismic units (SU 1-5), a consistent presence of well-stratified facies (SF) is observed, as shown in Figures 5.8 and 5.10. The well-stratified facies exhibits reflections with varying amplitudes, ranging from low to high and wavy characteristics. Figure 5.9 illustrates the lateral continuity of SF facies in this depth interval for each unit.

Further characterization based on reflection amplitude reveals two distinct subcategories of SF facies: SF 1 and SF 2. SF 1 is characterized by well-stratified low to medium reflection amplitudes, interpreted as fine-grain clay deposits resembling materials like plumite and suspension fallout in Unit 1. SF 2 exhibits medium to high reflection amplitudes and is identified as coarse-grained deposits resembling contourites.

5.3.2.2 Seismic facies between 600–450 m water depths (Zone 2):

In this depth interval, seismic units SU 2-4 consist of moderate to high amplitude stratified facies (SF 2), and SU 1 shows low amplitude stratified facies (SF 1) and have a nearly uniform thickness (Figures 5.11 and 5.12a). In contrast, unit SU 5A consists of stratified

facies with some encased chaotic facies (Figures 5.8 and 5.12a) and have high thickness (Figure 5.9e). Unlike the upper units, SU 5B consists of chaotic facies (CF 1 and CF 2) (Figure 5.8f), and the thickness is high because of the stack of debris lobes (Figures 5.9f and 5.12a). Older strata below SU 5B consist of well-stratified facies (SF) with low to high-amplitude reflections (Figures 5.11d, 5.11e, and 5.12a) interrupted by incisions and erosional troughs.

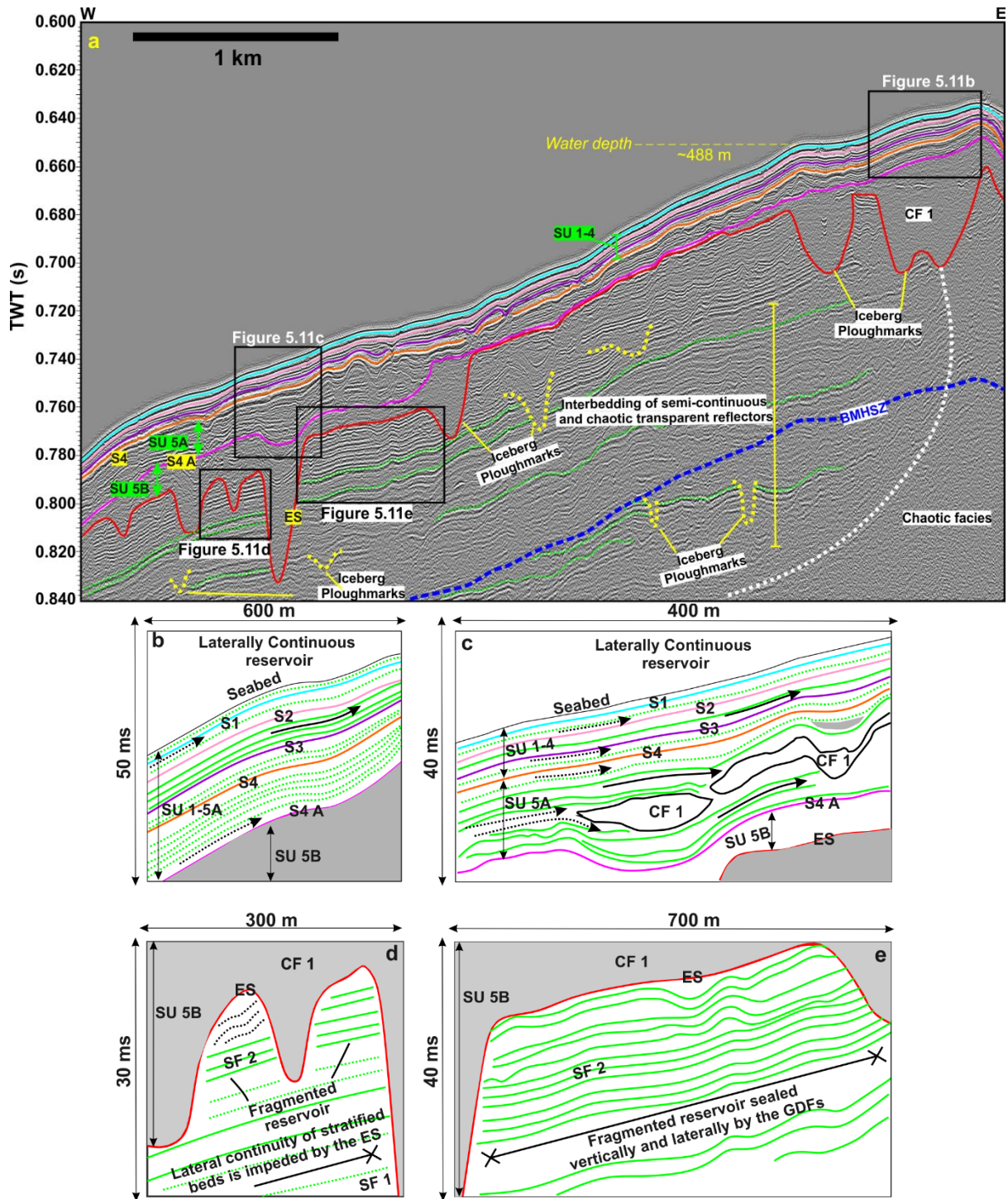
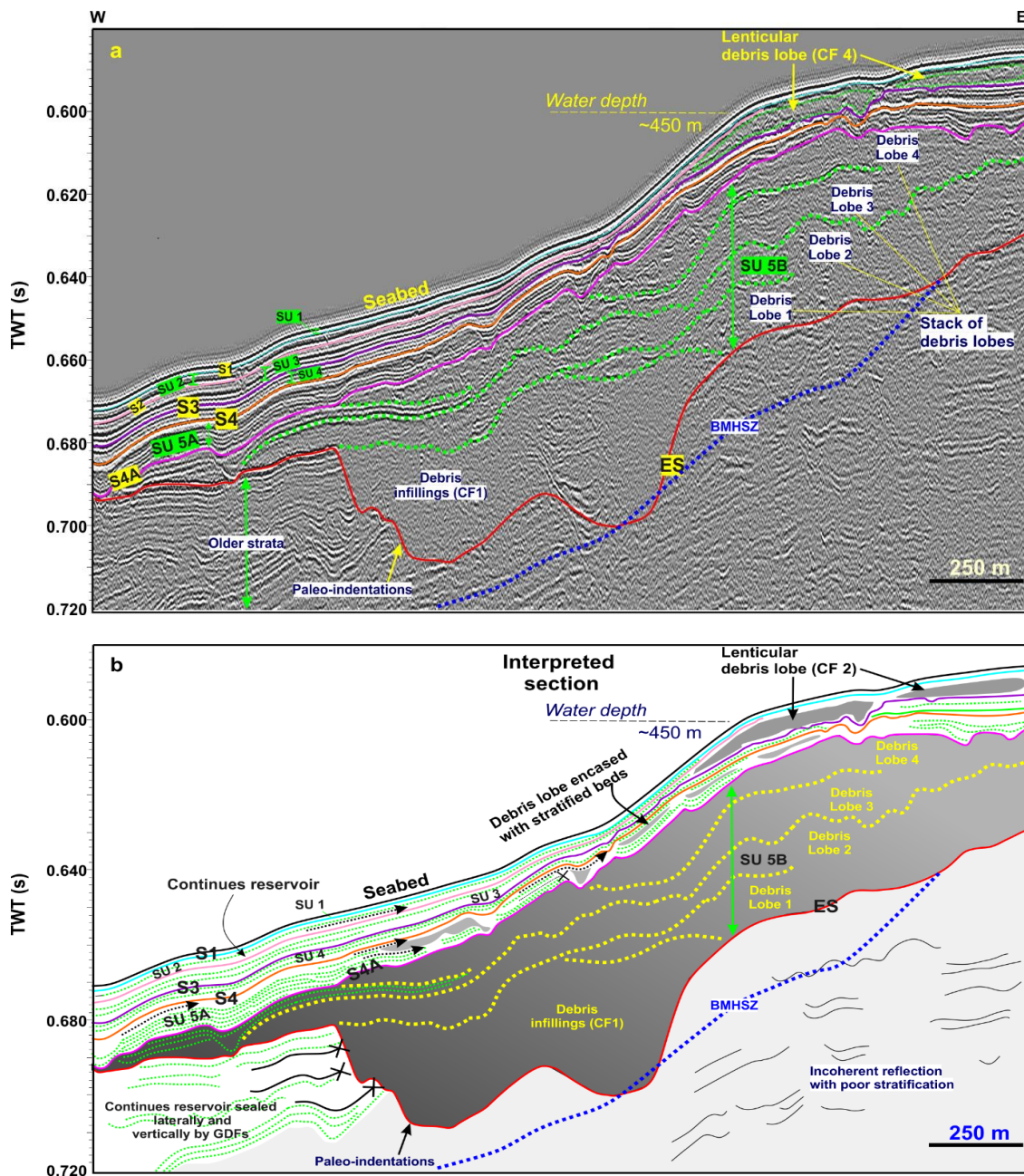


Figure 5.11. (a) SYSIF line 2.3 shows multiple iceberg ploughmarks filled in with GDF deposits. **(b)** Seismic units SU 1-5A show a good lateral continuity, while there is no stratified bed in SU 5B at this depth range. **(c)** Seismic units SU 1-4 are continuous. There is some bifurcation in the lateral continuity in SU 5A due to debris lobes. **(d)** Strata incised by the iceberg plough marks and the debris filled incisions interrupt the lateral continuity of the beds. **(e)** Parallel-bedded strata bounded by the iceberg ploughmarks on either sides. Fluid in such strata will be laterally and vertically trapped by the GDF deposits. Inset locations are shown in Figure 5.11a.



(For figure caption see the next page)

Figure 5.12. (a) SYSIF line 2.4 shows multiple stacks of lenticular debris lobes in the SU 5B. The iceberg ploughmarks trap the GDFs that cascade down the slope. Stratified reflectors of seismic units SU 1-5A cover the lenticular facies of SU 5B. **(b)** In the interpreted section, well-stratified facies in seismic units SU 1-5A facilitate the lateral migration of the fluid. However, the occasional presence of debris lobes on the uppermost slope at ~ 450 m water depth, can defocus the fluid pathways in unit SU 5A. The glaciogenic deposits interrupt the lateral continuity of the older beds below SU 5B.

5.3.2.3 Seismic facies between 450–350 m water depths (Zone 3):

In this depth interval, seismic units SU 1, 3, 5A-B have been preserved, each exhibiting distinct facies characteristics, while units SU 2 and 4 pinch out at 435 m water depth (Figure 5.13a). SU 1 is primarily comprised of stratified facies (SF 1), as shown in Figures 5.8a and 5.13a.

SU 3 exhibits chaotic facies (CF 1 and CF 2), with variations in spatial extent and thickness (Figures 5.8c and 5.9c). In the upper slope region, the presence of lenticular facies (CF 2) contributes to a moderate increase in thickness, as shown in Figures 5.13a and 5.14a.

SU 5A is characterized by isolated stratified facies (SF 3), as seen in Figures 5.8e, 5.13a, and 5.14a. The higher thickness observed in certain zones corresponds to the areas representing remnants of turbidity-related deposits (Figure 5.9e). Some of these beds consist of vertical-sub-vertical fractures (Figure 5.14b). Since these beds are 10 mbsf, the fractures might be related to shallow overpressure caused by repeated seasonal hydrate formation and dissociation.

SU 5B displays chaotic lenticular facies (CF 1, CF 2, and CF 3 in Figure 5.8f). The observed decrement in thickness from 400-m to 300-m water depth results due to facies transitions from the vertical lenticular stacks to oblique stacks of smaller debris lobes (CF 2 to CF 3) (Figure 5.9f). Oblique reflections (CF3) represent distinct debris lobes that were deposited in a retrogradational fashion as the ice sheet started to retreat from the shelf break and debris supply progressively diminished (Figure 5.15). This stacking pattern is either due to a lack of accommodation space, runout potential, or diminishing sediment supply. In water depths ranging from 355-405 m, some isolated vertical/near-vertical fractures are seen within unit SU 5B (Figure 5.15). These fractures were likely formed due to high pore pressure induced by hydrate dissociation.

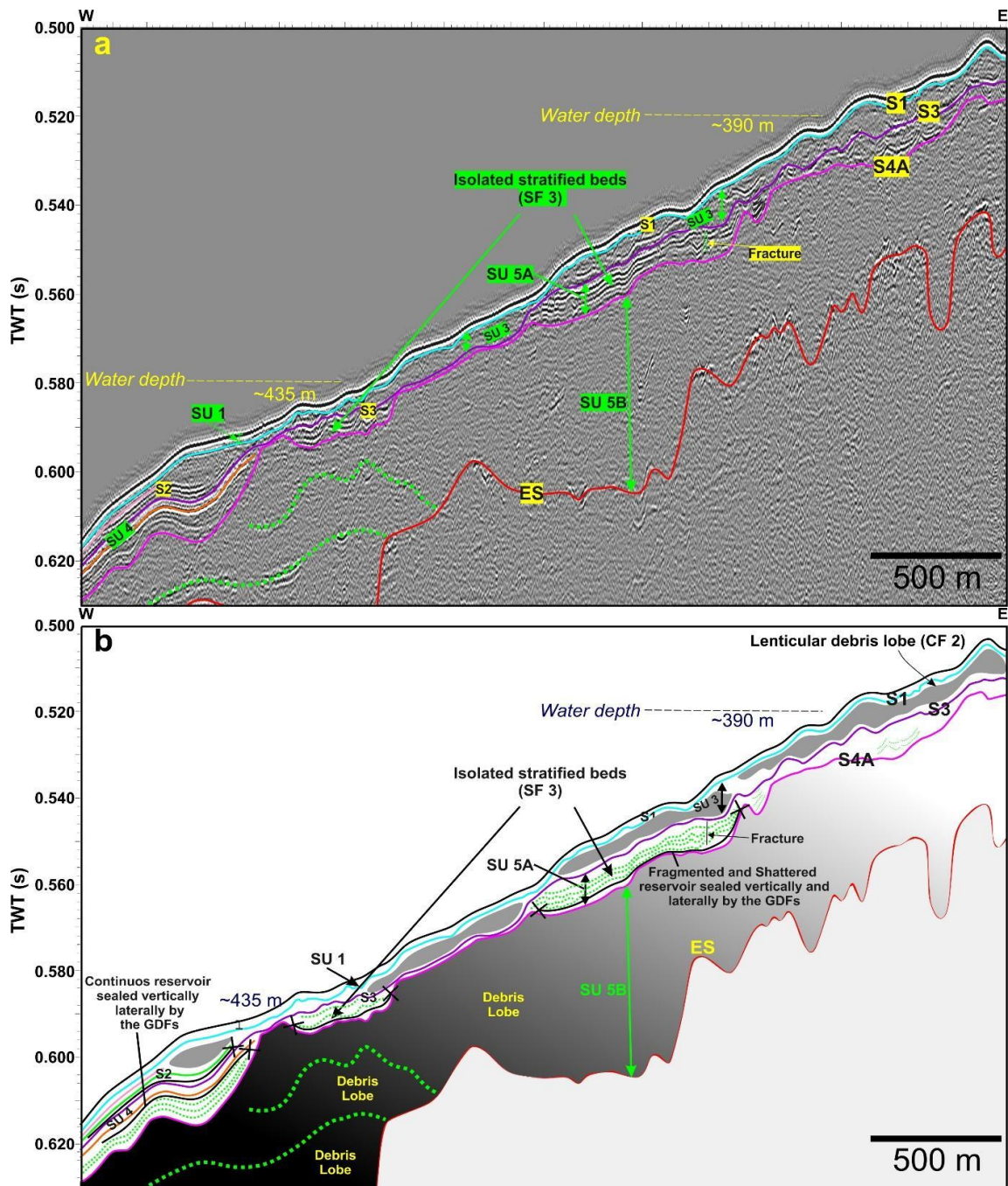


Figure 5.13. (a) SYSIF line 3.1 shows an erosional surface (ES) filled by glaciogenic debris. Seismic unit SU 4 pinches out at a water depth of ~ 435 m due to erosion by the glaciogenic debris of SU 3. The isolated stratified seismic unit SU 5A package shows moderate to high-amplitude reflections that are laterally juxtaposed against lenticular debris lobes. **(b)** Shows the presence of vertical-near vertical fractures in the stratified package of seismic unit SU 5A. (b) Line drawing of seismic section shown in (a). Lateral migration of fluid in Unit 5A is only limited within the well-stratified segments, as the lenticular debris lobes impede lateral and vertical fluid migration.

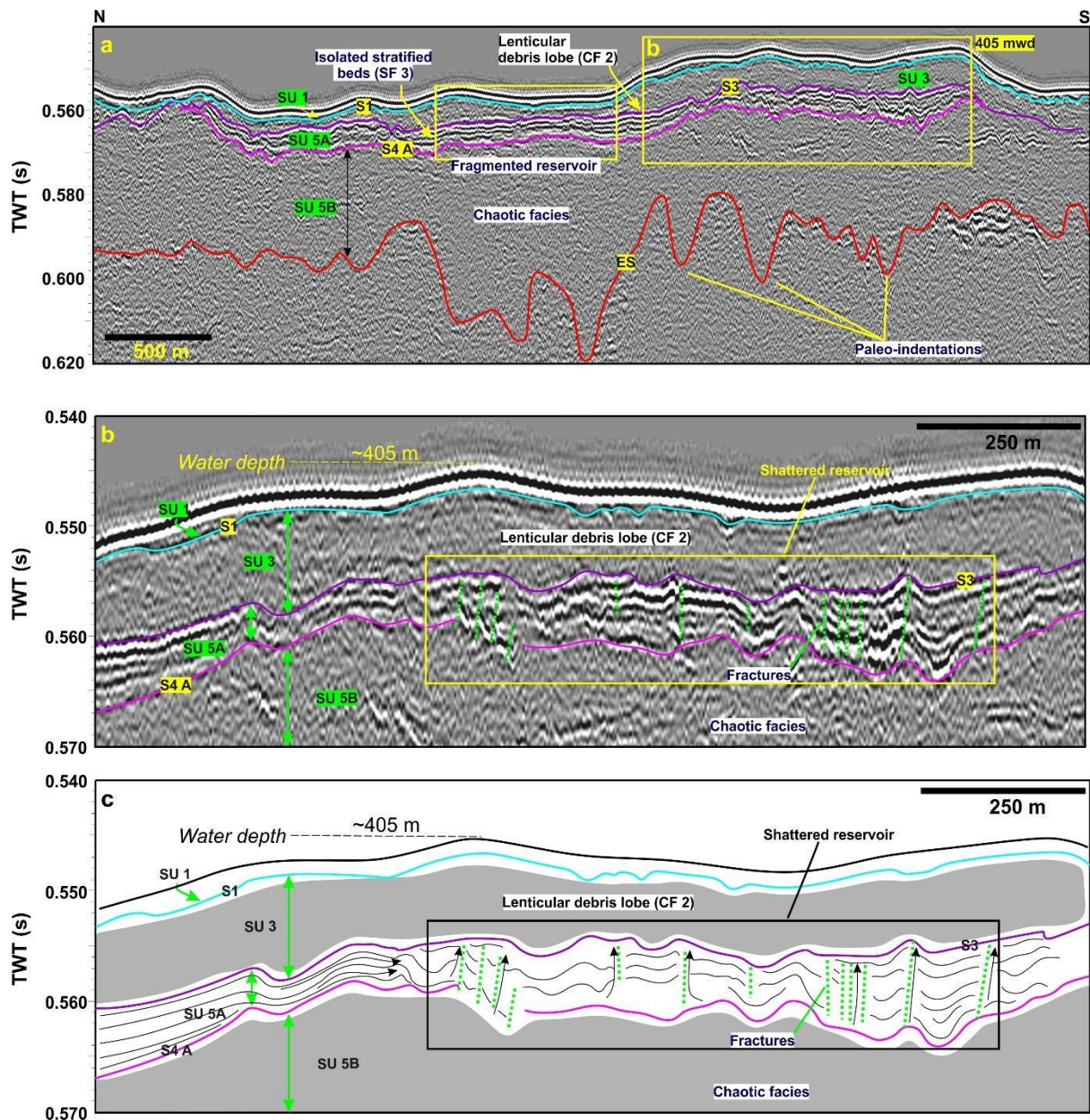


Figure 5.14. (a) N-S oriented SYSIF line 3.15 shows multiple iceberg ploughmarks (ES) filled by glaciogenic debris followed by isolated well-stratified facies in vertical succession. GDFs characterize seismic unit SU 3. **(b)** Vertical/near vertical fractures in the stratified package of seismic unit SU 5A are marked. **(c)** Line drawing of the seismic section shown in (b). Fluid can migrate laterally along SU 5A and vertically through the fractures. However, the lenticular debris lobes of seismic unit SU 3 likely impede vertical migration of the fluid. The Black arrows indicate the possible direction of fluid flow.

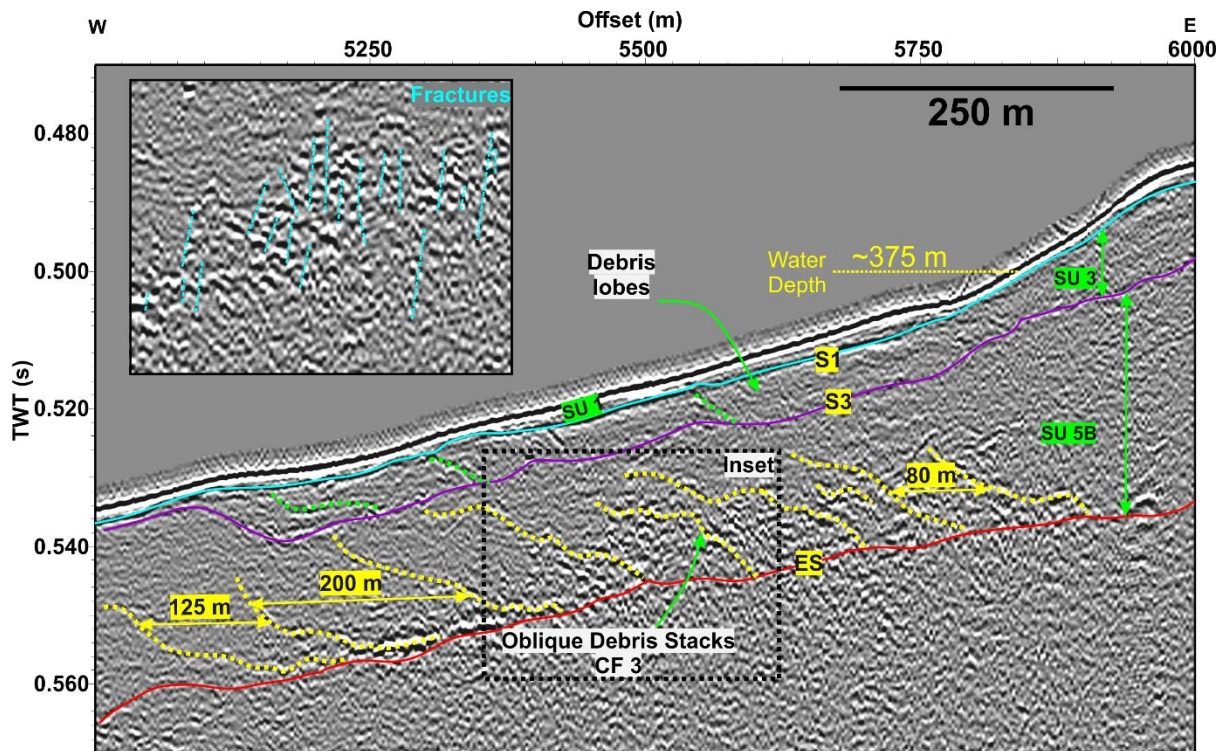


Figure 5.15. Slope perpendicular SYSIF line 3.11 shows the subsurface in the UCS. Obliquely stacked debris lobes accumulated in a retrogradational pattern. Some vertical to near-vertical fractures are seen within them.

5.3.3 Fluid-related anomalies:

5.3.3.1 Seismic Chimneys:

In the SYSIF line 2.1 (northernmost line), only a limited set of buried seismic chimneys (C1-C4, Figure 5.16a) are identified as vertical or near vertical zones. In the envelope attribute, these chimneys show weak energy (Figure 5.16b). Table 5.5 shows chimney height, which is derived from TWT and sediment velocity of 1800 m/s. The base of seismic chimney C1 coincides with the transient BMHSZ, while others fall within the methane hydrate stability zone (MHSZ). The summits of seismic chimneys C2-C4 show prominent up-bending, possibly related to fluid release and associated deformation at an ancient time (Inset, Figure 5.16a).

Table 5.5. The height of seismic chimneys.

Seismic chimney	Height (m)
C1	140
C2	115
C3	85
C4	15

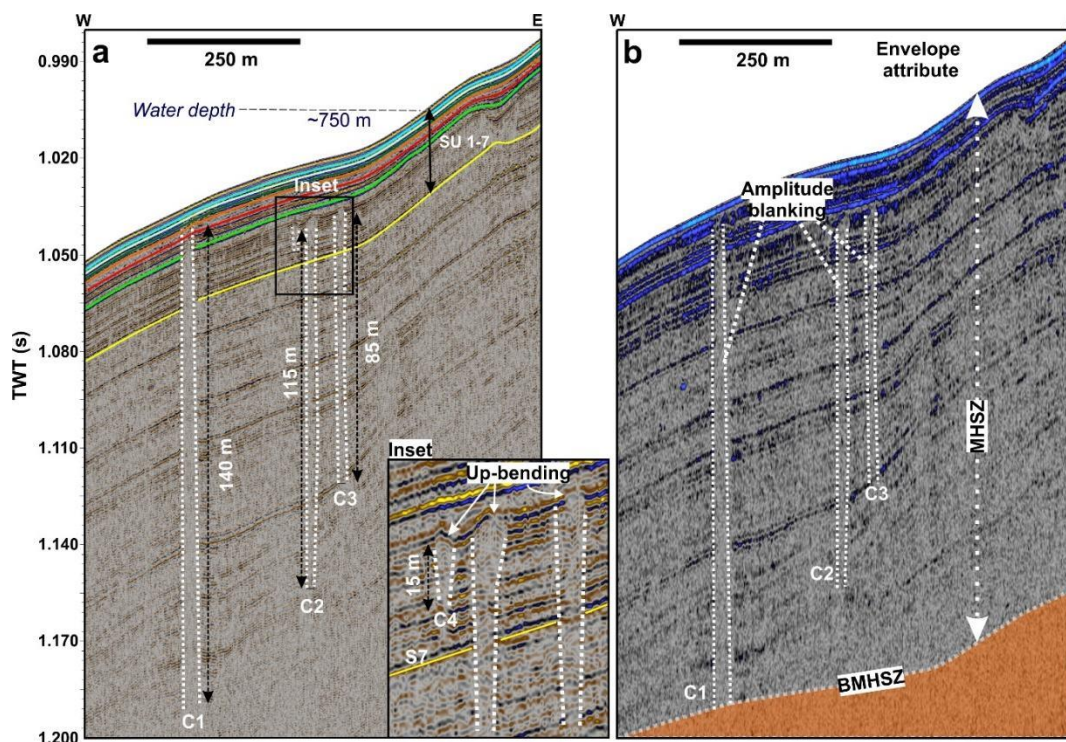


Figure 5.16: SYSIF line 2.1 shows seismic chimneys. **(a)** Four buried seismic chimneys (C1-C4) of different heights show amplitude fading and reflector up bending at their tip (Inset). **(b)** The envelope attribute of the seismic section shows vertical zones of amplitude blanking. Chimney C1 originates at the present-day BMHSZ (modelled), while the root of other two chimneys occur within the hydrate stability zone. The modelled BMHSZ is also shown.

5.3.3.2 Bright spots:

Several bright spots were identified within the glaciomarine sedimentary succession in the interfan area. These bright features could be of stratigraphic origin, or related to fluid flow. Some of these acoustic anomalies exhibit a bright, negative amplitude reflection compared to the lower amplitude background signature (Figure 5.17). These anomalies suggest the presence of free gas or gas-charged pore fluids. Additionally, a few isolated positive polarity

bright spots are observed, although they are less common and likely represent thinner gas layers (Westbrook et al., 2008). Mebo 140 drill core (location in Figure 5.2b) confirms the presence of gas in the sandy-clay sediments at the base of a GDF deposit (Figure 5.17). Based on the SYSIF data, two types of bright spots were identified, which are mentioned below:

Type 1: Isolated bright spots in SYSIF data:

These bright spots are present at the base of SU 5B (Figure 5.17b) and in isolated stratified beds of seismic unit SU 5A (Figure 5.17c). At the base of SU 5B, these bright spots are present in isolated pods that are embedded within a chaotic matrix. The adjacent moderate amplitude reflections show poor continuity, rugose topography and minor grooves. The isolated brights could represent lithological heterogeneity, such as pebbly/sandy clay (harder event) above a muddy matrix (soft event) that gives rise to a negative polarity and not necessarily a gas anomaly. The presence of biogenic methane may result in localized gas pockets. MeBo drilling confirmed the existence of trapped gas in sandy clay at this site (Figure 5.17a and S5.2). Isolated bright spots along the strata, could also signify instances of gas accumulation in pockets within very poor quality glacial diamictic strata made up of sandy clay lenses with limited lateral continuity.

In Unit SU 5A, the beds containing the bright spots represent isolated well-stratified reflections with moderate to high amplitude (Figure 5.17c), indicating the coarser sediments like sand/silt are the remnants of the turbidites beds that partly escaped the erosive action of later GDFs. The bright spots likely indicate gas in the sandy/silty layer.

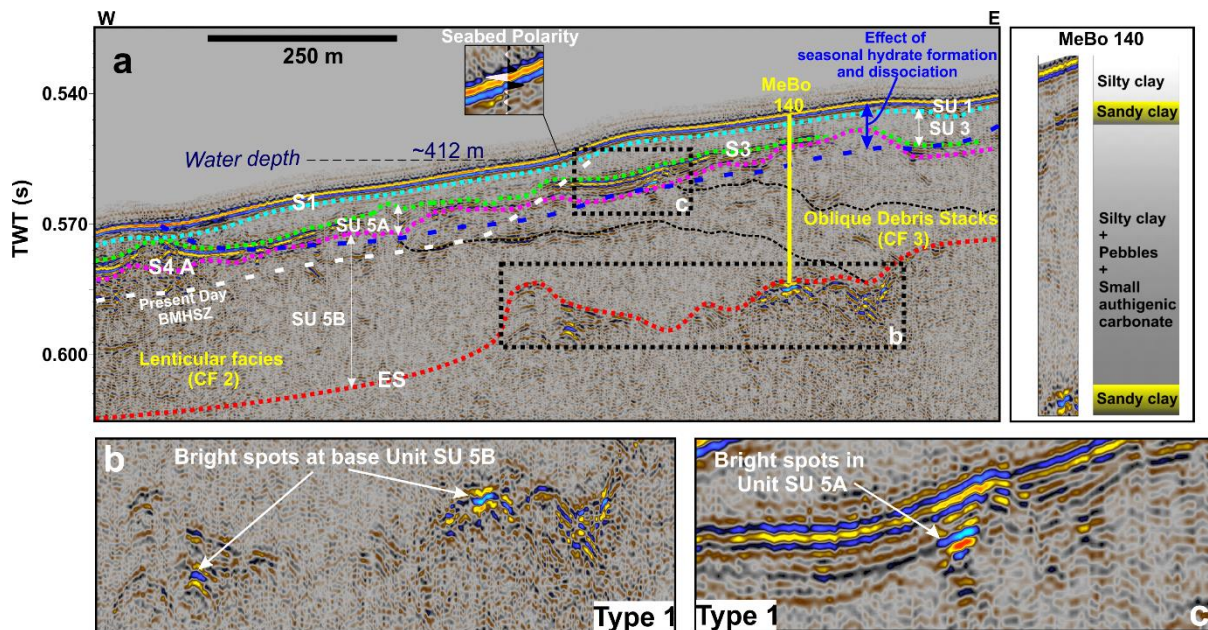


Figure 5.17. The presence of fluid-related anomalies is shown on the slope perpendicular line SYSIF 2.2. **(a)** Bright spots with reversed polarity are present at two stratigraphic levels: in the SU 5A and at the base SU 5B. MeBo 140 drill core, present at a water depth of 405 m, shows two distinct sandy clay zones trapped within glaciogenic debris that can hold fluid in it. Present-day BMHSZ pinches at 410 mwd (white dashed line), while the top 10 mbsf is affected by seasonal hydrate dynamics (blue dashed line) **(b)** Type 1 bright spots are present beneath the SU5B covered by the lenticular facies. **(c)** Type 1 bright spots are in isolated, well-stratified beds of SU 5A.

Type 2: Scattered bright spots:

These bright spots are aligned obliquely at the base of seismic unit SU 5B (Figure 5.18).

These lobes exhibit chaotic, incoherent transparent reflections separated by oblique moderate amplitude reflector dipping eastward in the E-W profile (Figure 5.18a) and northward in the N-S profile (Figure 5.8b). These facies represent stacks of debris lobes deposited due to glacial debris flows, and their particular orientation indicates (i) the lack of accommodation space, (ii) dwindling glacial debris supply, and (iii) limited runout potential. The bright spots show high positive and negative amplitude reflections primarily due to coarser grains such as pebbles and gravel. These brights mostly represent stratigraphic heterogeneities and are not likely related to fluid, even though they may trap traces of gas. The poor permeability within multiple debris lobes slices likely baffle fluid flow.

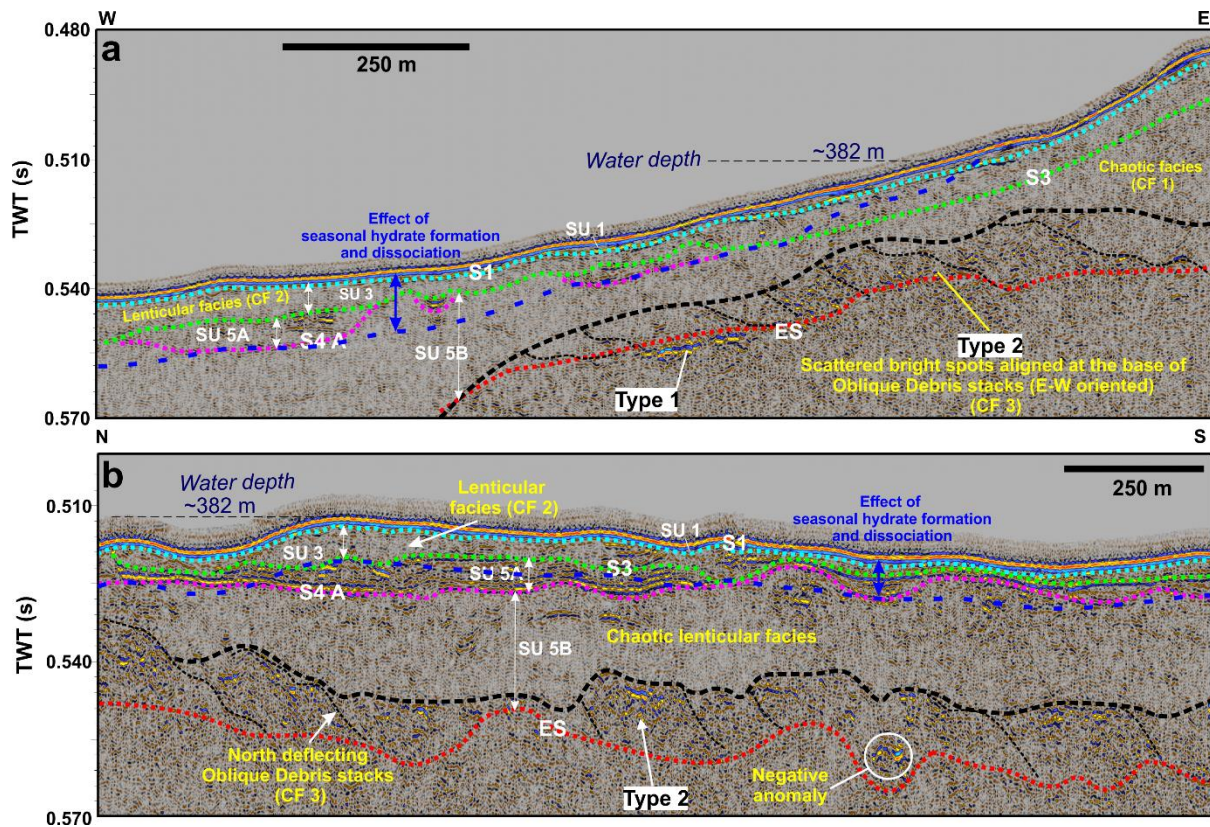


Figure 5.18. The presence of fluid-related anomalies is shown on the slope perpendicular SYSIF 3.11 and slope parallel line SYSIF 3.15. **(a)** Type 1 Bright spots with reversed polarity are present at the base SU 5B, while Type 2 scatter brights are present within oblique debris lobes (E-W oriented) of SU 5B. **(b)** Type 2 scattered bright with occasional negative polarity are present within debris stacks (N-S oriented) of SU 5B.

The 3D airgun data provides continuous spatial coverage between 318–405 mwd in the interfan area (Figure 5.2b) and shows bright spots at 20–25 mbsf, below the methane seepage sites. Two types of bright spots are observed in the 3D data set, as mentioned below:

A: Isolated bright patches:

The majority of bright spots observed in the 3D data across the study area are isolated (Figures 5.19a, and 5.21), have negative polarity, and high amplitude (Figure 5.19b) with respect to the low amplitude of the background.

B: Linear/curvilinear bright patches:

In the sweetness attribute, some of the bright spots show a better lateral continuity and a linear or curvilinear shape on a horizon tracking the bright spots (Figure 5.21). These bright spots are interpreted as stratified and moderately sorted turbidite beds within NNW-SSE

aligned ploughmarks. These patches are seen at the base of debris stacks. Turbidites were likely deposited in pre-existing topographic lows and are much more homogeneous than the glacial debris in which they are encased (Figure 5.19c). The topographic lows were formed due to the erosive action of debris or due to icebergs (mostly NNW-SSE aligned) ploughing the seabed (Figures 5.19e, 5.20a-c).

Some bright anomalies are offset by vertical to near-vertical linear fractures (Figures 5.19a, 5.19d, and 5.20a) and are classified as fractured brights.

A RGB blend of 25 Hz (R), 40 Hz (G), and 85 Hz (B) frequencies reveals isolated (Figure 5.19f) or curvilinear white patches (Figure 5.20d), signifying the presence of gas in turbidite beds charged with gas. Red halos are seen below such brights indicating gas shadows formed due to high-frequency attenuation caused by gas. Cyan-coloured areas represent turbidite beds, where gas is not likely present as there is no bright anomaly in the 'Red' channel signifying any detectable low-frequency gas. These narrow threads of cyan colour bands are separated by sharper and darker linear features, mostly showing NNW-SSE orientation. These are interpreted as fractures and correlated with linear features showing high variance (Figures 5.19d and f). A conceptual model for the spatial distribution of the gas pockets is shown in Figure 5.22.

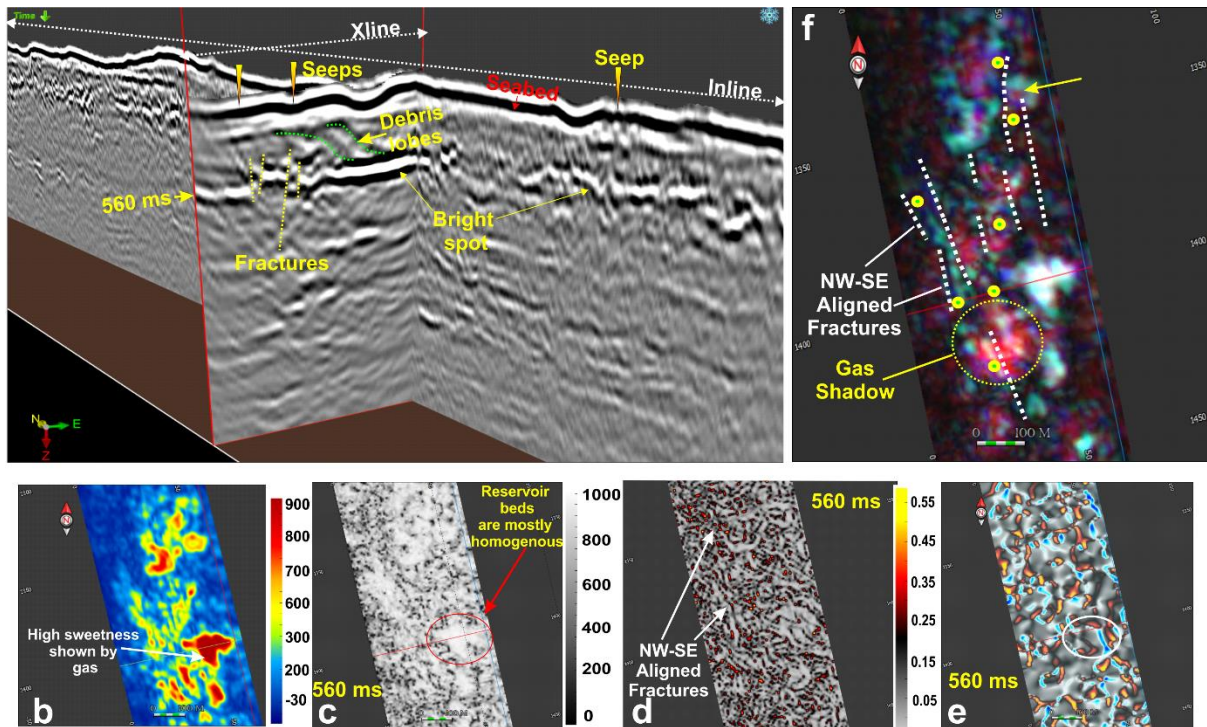


Figure 5.19. The 3D chair-cut view, along with different seismic attributes and RGB blend of spectral decomposition, shows the complex stratigraphy associated with glacial system and associated fluid anomalies. **(a)** The in-line and cross-line show the distribution of bright spots, oblique landward-dipping debris lobes and fractures. **(b)** The sweetness attribute extracted at a time slice 560 ms highlights the fractured bright anomalies in orange and red colour and inferred to be gas. **(c)** The similarity attribute shows the turbidite beds (white patches separated by chaotic surroundings), which are characterized by more homogenous lithology compared to the chaotic glacial debris. **(d)** Variance attributes show NW-SE fractures. **(e)** Curvature attribute shows positive and negatively curved nature of the horizon tracked along the bright spots indicating that pit and grooves were filled by plumites and turbidity-related deposits and subsequently overlain by debris lobes. **(f)** RGB composite derived from spectral decomposition products (25 Hz (R), 40 Hz (G), and 85 Hz (B)) shows two types of bright anomalies: isolated bright white anomalies that appear fractured/segmented with a red halo signifying gas shadow effect caused by an overlying bright spot. Yellow dots indicating the seepage points, are present near the fractures.

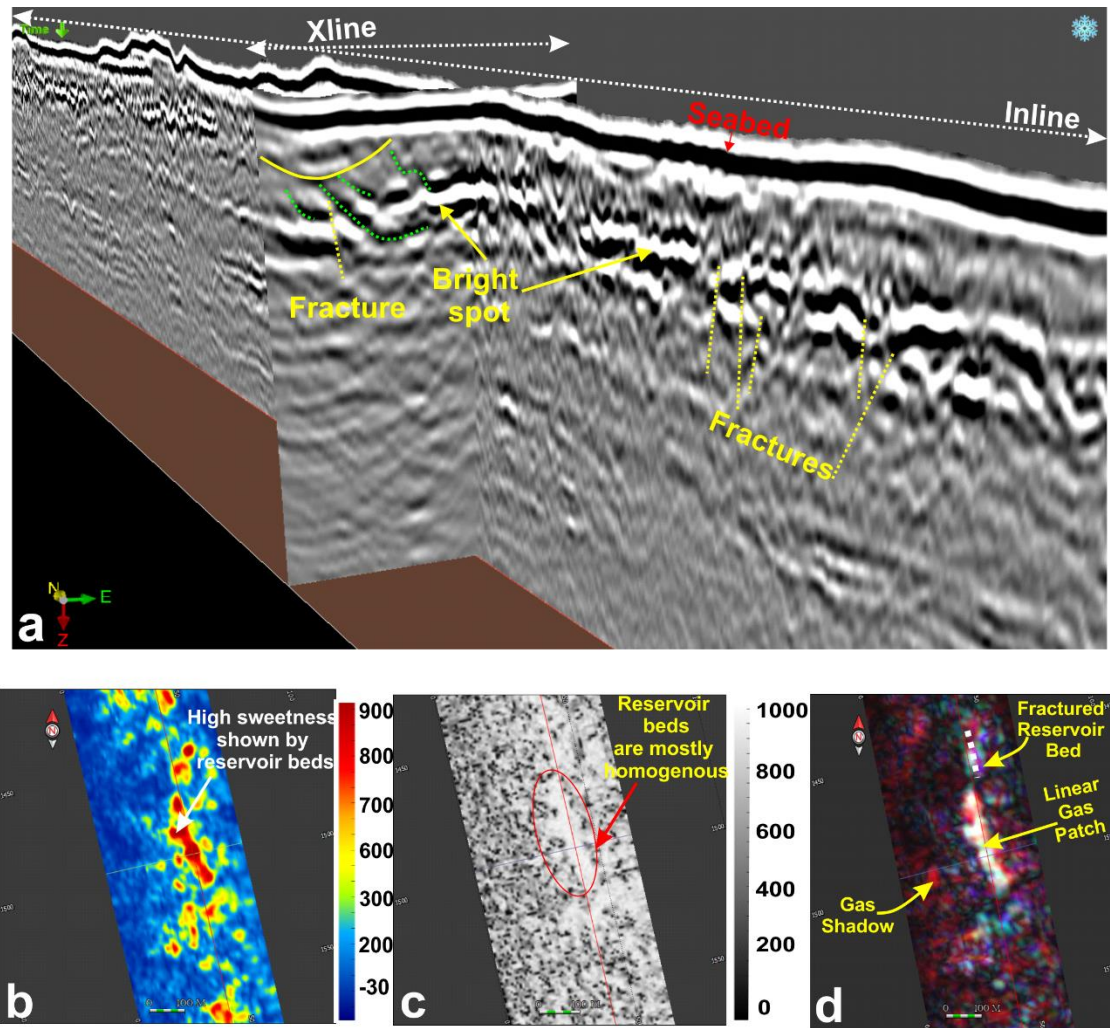


Figure 5.20. The 3D chair-cut view and RGB blend and seismic attributes (sweetness and similarity) showing complexities of glaciogenic sediments and bright anomalies. **(a)** Scattered bright anomalies visible in the in-line and cross-line. **(b)** The sweetness attribute shows a curvilinear segment in high sweetness that likely indicates presence of gas. **(c)** The similarity attribute shows nearly homogenous lithology indicating better sorting within this patch (e.g. turbidite beds). **(d)** The RGB composite of a horizon encompassing the bright spots shows a white curvilinear patch.

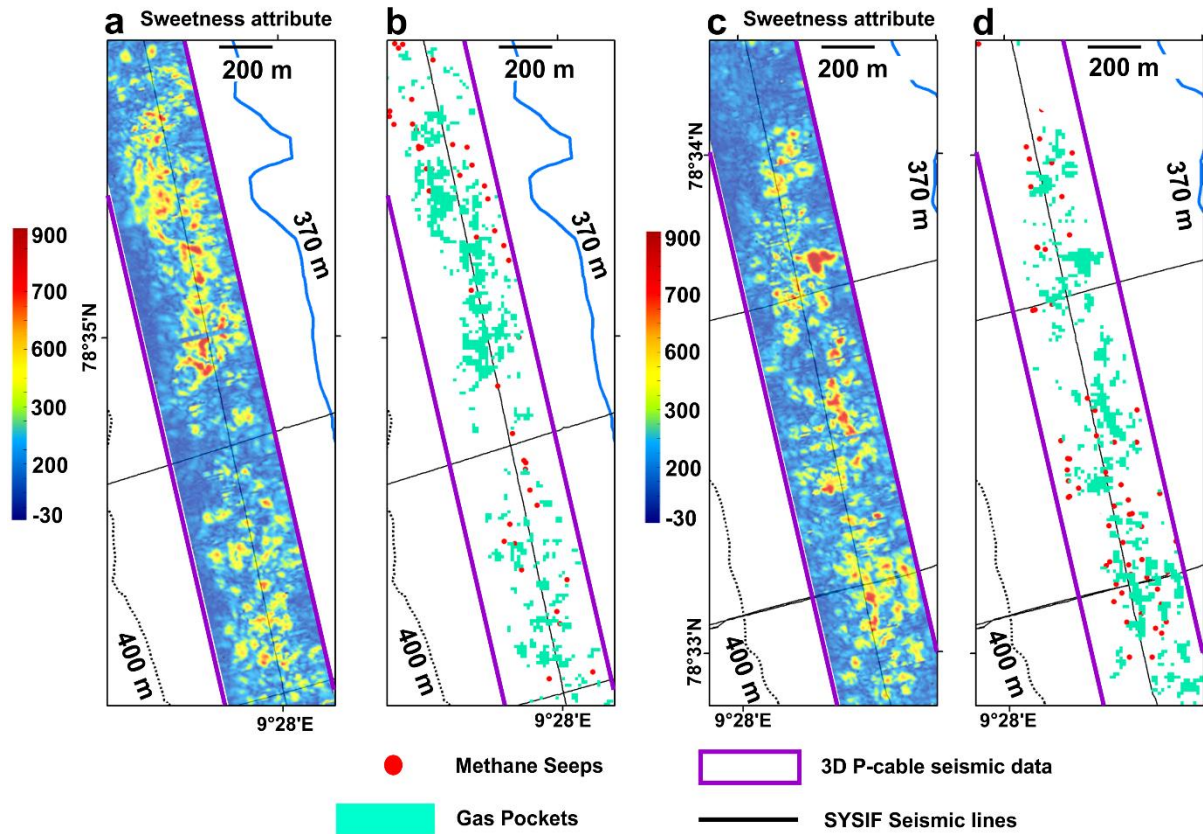


Figure 5.21. Spatial distribution of subsurface occurrence of bright spots and seabed methane seepage points. (a), (c) Sweetness attribute employed as a gas indicator. It highlights isolated gas pockets, which form linear to curvilinear patterns. This arrangement is likely due to the deposition of coarser sediments on the uneven surface of pre-existing debris lobes. (b), (d) The presence of gas pockets beneath the seepage points suggests that gas is transmitted from these pockets through vertical cracks. The seemingly random distribution of these seepage points results from the subsurface distribution of gas-charged layers.

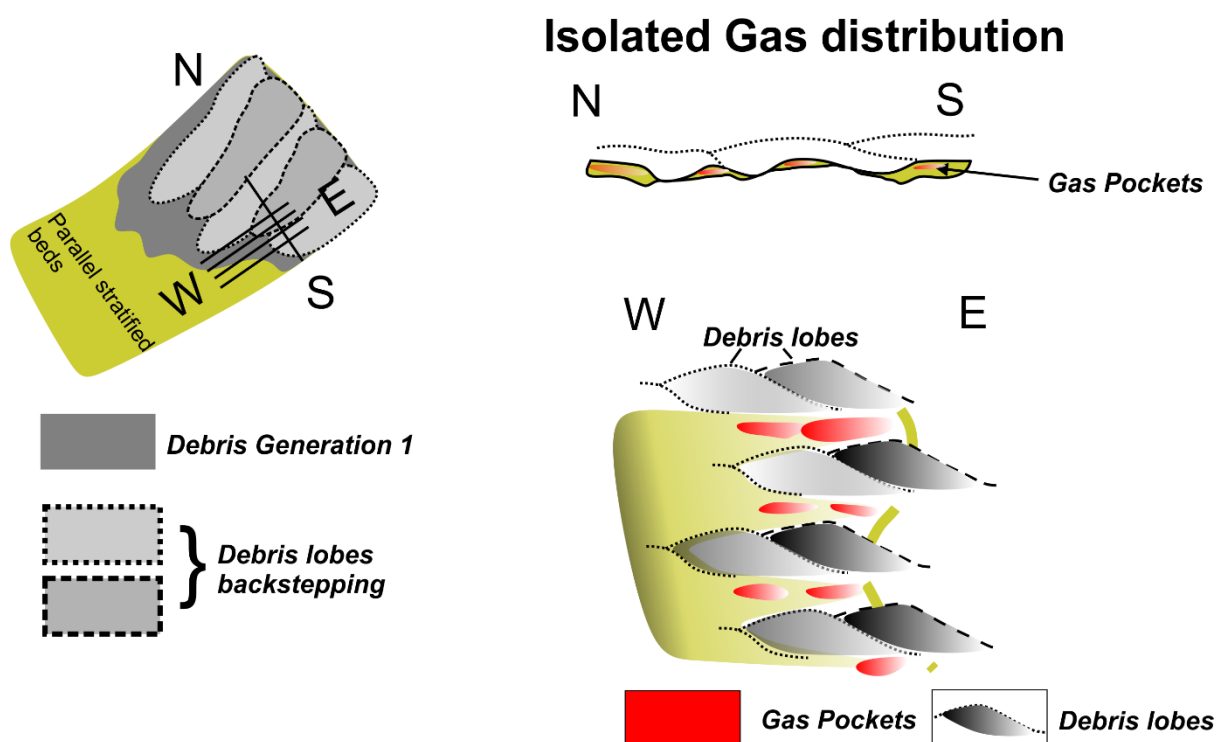


Figure 5.22. Schematic representation of accumulation and spatial distribution of isolated and curvilinear gas pockets sealed by impermeable debris stacks.

5.4. Discussion

The methane seepage area exhibits a distinct linear trend that aligns with bathymetric contours, confined within depths ranging from 370 to 400 m in the uppermost slope. This depth range corresponds to the zone where methane hydrate was previously accumulated but is now depleted due to bottom-water warming by 1 °C (shifting from a depth of 370 m at 2 °C to a depth of 400 m at 3 °C). Therefore, hydrate dissociation is considered a contributing factor to the methane seeps (Westbrook et al., 2009). On closer analysis of the seepage sites (Figure 5.2b), a random distribution of methane seepage is observed, indicating the likely role of subsurface lithological heterogeneity in controlling fluid seepage at this depth. In this context, I analysed the subsurface occurrences of gas and explicit role of lithology and hydrate dynamics in governing the lateral and upward migration of fluid flow in the interfan methane seepage area of the uppermost slope.

5.4.1 Indication of deep and shallow gas occurrence

Bright spots with reversed polarity often indicate free gas accumulations (Judd and Hovland, 1992). Gas accumulations are observed at the base of debris lobes in SU 5B (Figure 5.17b) and within isolated well-stratified reflectors (SF 3 facies) of SU 5A (Type 1) (Figure 5.17c). At the base of SU 5B, isolated, poorly stratified beds indicative of coarser grains, represent

sand lenses embedded within clay-dominated diamicton. These sediments locally exhibit higher porosity and can retain fluids that remain trapped since the overlying clay acts as a seal. Moreover, these beds can generate methane from microbial origin as reported by Sahling et al. (2014). In the shallow glaciomarine strata of Unit SU 5A, isolated stratified facies exhibit better sorting, facilitating fluid accumulation. Glaciogenic debris deposits hinder the lateral continuity of these beds on the uppermost slope, resulting in gas entrapment (Figure 5.17c). Bright spots within the oblique debris stacks (Type 2) of seismic unit SU 5B is limited to specific locations (Figure 5.18b). Within oblique debris stacks, gas might exist in coarser sand rich sediments but the debris sediments act more as seal layers rather than a good gas reservoir.

The analysis of 3D seismic data revealed that the zone of bright spots (Figures 5.19 and 5.20) represents a more homogeneous lithology than the surrounding area, featuring isolated and disconnected gas pockets. Fractures in these beds likely compartmentalized them, leading to the formation of isolated gas pockets, while interconnected gas pockets indicate more continuous beds.

5.4.2 Glaciomarine reservoir and Fluid migration in the interfan region

Based on the glaciomarine stratigraphy of the interfan area, I have categorized the reservoir and fluid migration potential for the different zones (Zones 1-3), as mentioned below:

Zone 1 (800 – 600 m water depth)

In the examined depth interval, the strata exhibiting SF 2 facies are identified as very good-quality reservoirs due to coarser-grained sediments, contributing to favourable porosity and strong lateral connectivity, as highlighted in Inset 2, Figure 5.10. These characteristics make SF 2 beds particularly promising for fluid storage and transmission.

On the other hand, beds featuring SF 1 facies are categorized as seals. This is likely due to the prevalence of fine-grained sediments, such as clay, which have high porosity and low permeability within SF 1. Vertical zones with reduced amplitudes are interpreted as seismic chimneys and serves as conduits that facilitate the flow of fluids from deeper levels to shallower depths (Hegglund, 1997). The base of chimney C1 coincides with the BMHSZ and confirms that it probably originated because of overpressure arising from hydrate dissociation. Other chimneys (C2-C4) falling inside the hydrate stability zone indicate gas and hydrate coexistence. This can happen if there is insufficient water in the sediment column and/or increased pore water salinity due to salt expulsion during hydrate formation (Liu and

Flemings, 2007). There are alternative explanations for chimneys. They could be high-velocity hydrate plugs, which are often associated with amplitude wipe-outs and pull-ups beneath them. These features are frequently observed as faded amplitudes in short streamer data. Due to the short streamer of the SYSIF, this possibility could be considered but cannot be confirmed.

Zone 2 (600 – 450 m water depth)

Similar to zone 1, in this depth interval, the strata (SU 1-4) showing SF 2 facies are identified as very good-quality reservoirs (Figures 5.11b, 5.11c, and 5.12b). Unit SU 5A has stratified facies that are parallel and continuous. However, the occasional presence of chaotic debris can split the fluid movement towards shallower areas, creating a more tortuous pathway (Figures 5.11c and 5.12b).

The strata giving rise to chaotic facies (interpreted as glacial debris flow deposits) are interpreted to poorly sorted glacial diamicton, implying poor fluid storage capacity in this strata. Also, the permeability of the glacial debris is poor because of finer clays fill in the pore spaces between larger clasts, such as gravels, pebbles, and sand. The isolated fabric of moderate amplitude reflections dividing the two lenticular facies (Figure 5.12b) might act as fluid carriers. However, their lateral continuity is severely impeded by the debris deposits. Debris filled ploughmarks disturb the lateral continuity of stratified beds (Figures 5.11d and 5.12b) and impede lateral updip fluid flow (Figure 5.11e). At water depth ~ 500 m, erosion caused by deeper ploughmarks in SU 5B limit the updip continuity of many stratified beds. This factor will hinder lateral fluid migration and deflection toward the shelf break because of the termination of stratal continuity as envisioned by Sarkar et al. (2012).

Zone 3 (450 – 350 m water depth)

Seismic Unit 1 (SU 1) is characterized by stratified facies 1 (SF 1), indicating the prevalence of fine-grained clay deposits. In this context, SU 1 is designated as a seal, as illustrated in figures 5.13, 5.14, and 5.15.

In Seismic Unit 5A (SU 5A), isolated stratified facies represent well-sorted turbidites; however, the lateral continuity is disrupted by the presence of chaotic debris packages. These beds are thus compartmentalized (Figures 5.13b and 5.14c). Moreover, some isolated stratified beds are fractured (Figures 5.14b and 5.14c). While maintaining some lateral continuity similar to the fragmented reservoir, the fractures in the beds introduce inter-

connected pathways that can facilitate the upward migration of fluids leading to the formation of fractured reservoirs at the methane hydrate retreat zone.

The strata of SU 3, consisting of chaotic, and lenticular facies (interpreted as GDFs deposits), are represented by poor sorting, and therefore, poor permeability, (Figures 5.13b, 5.14c, and 5.15) making them a seal.

SU 5B consists of chaotic lenticular debris deposits and oblique debris stack (Figure 5.13a and 5.15), which act as seals. These layers can alternate between coarse-grained diamicton and matrix-supported clayey diamicton, forming zones with varying porosity that create intricate barriers to fluid communication. In addition, the primary depositional fabric within the oblique debris lobes can impact fluid transmission, for example, if well sorted silt or sand is trapped between two adjacent stacked debris lobes, there could be improved porosity and permeability in that layer. Accordingly, the oblique landward-dipping debris deposits may exhibit significant directional dependency in permeability (for example, landward dipping reflectors shown in Figure 5.15). Fluids flow more easily along the preferred direction of clast imbrication than in perpendicular directions. This anisotropy impacts flow patterns and transmission efficiency in specific directions.

Also, at the base of SU 5B, below the methane hydrate retreat zone, there are some isolated, discontinuous and irregular beds with relatively better sorting than the surrounding glacial diamicton (Figure 5.15). The limited lateral continuity of stratified beds from the deeper slope to this area hinders lateral fluid migration. However, vertical fluid flow through fractures can supply gas to these beds, along with local shallow biogenic gas. In the 3D data, vertical/near-vertical fractures break the continuity of the stratified layer containing gas (Figure 5.19). On a 3D seismic time-slice, these fractures are aligned NNW-SSE. Other than hydrate dynamics, fracture reactivation could also be triggered by glacial unloading and subsequent drop in confining pressure (Vachon et al., 2022).

5.4.3 Effect of Methane Hydrate dynamics on Fluid flow:

The BSR in the inter-fan region is present at water depths > 600 m, but it is not found at shallower depths and below the gas seepage sites. In the absence of a BSR as the direct seismic evidence for the base of methane hydrate stability (BMHS) beneath the uppermost continental slope, modelling was used to predict the BMHSZ.

Based on the long-term transient hydrate stability model for zones 1 and 2, for a water depth of ~ 712 m, the present-day BMHSZ is positioned at a depth of 150 mbsf at ~ 712 m water depth and 120 mbsf at ~ 490 m water depth. This implies that in zone 1, gas within the

BMHSZ is likely to be trapped as hydrates (Inset 2, Figure 5.10). Gas can potentially migrate upslope along the BMHSZ. On the other hand, in zone 2, gas can also potentially migrate upslope along the BMHSZ. However, chaotic facies (debris with poor permeability) will impede updip fluid migration (Figure 5.12b).

In the water depth ranging from 370–410 m, hydrate was stable when colder bottom water conditions prevailed during late 1970s favouring hydrate formation, and methane that was originally stratigraphically trapped within the isolated turbidite beds got locked in the hydrate cage. During 1978–2016, BWT rise caused complete hydrate dissociation. A simulation was run to understand the evolution of a methane hydrate layer that was originally located 10–30 mbsf at 405 m water depth (MeBo drill site 140) using the Tough+Hydrate code. The simulation was conducted by forcing the BWT curve at the seabed. The simulation results are shown in supplementary Figure S5.1. At the end of simulation, methane hydrate completely dissociated producing fresh water and gas in the system. The final salinity matched the results observed during drilling at the MeBo site 140 (Supplementary Figure S5.1). The production of methane in sediments of poor permeability can generate overpressure, if the gas produced by hydrate dissociation cannot vent out. These overpressured zone can trigger fracturing and methane release. Fractures in the stratified beds of SU 5A might be indicative of fracturing induced by overpressure.

5.5. Conclusions

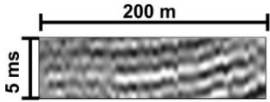
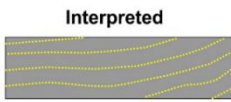
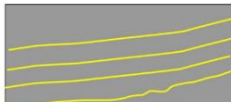
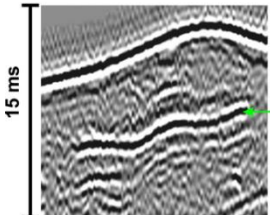
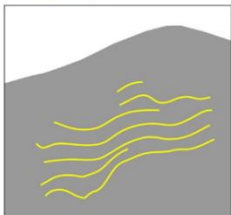
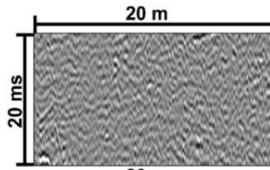
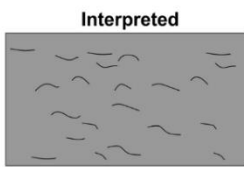
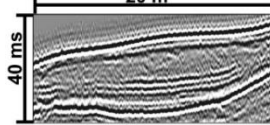
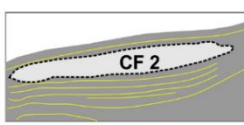
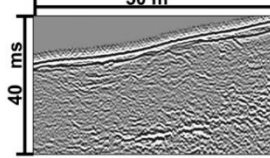
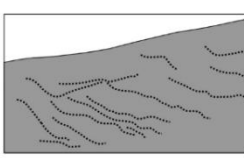
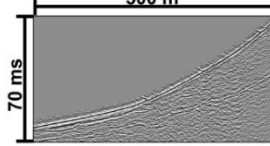
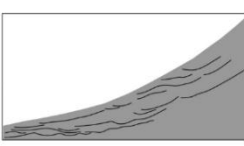
SYSIF seismic data analysis in the interfan area helps to determine reservoir characteristics and fluid migration patterns in the uppermost continental slope. As the water depth shoals, the degradation of the reservoir potential is attributed to the influence of glaciogenic deposits. The interconnected fractures in the shattered reservoirs at the base of unit SU 5B and in unit SU 5A are likely responsible for fluid focusing, and the observed methane seeps at water depths of 350–390 to 410 m. The high-resolution SYSIF data facilitated the identification of potential gas occurrences at two distinct stratigraphic levels characterized by limited stratification and separated by intervening glacial debris.

The SYSIF data did not reveal a BSR, but through numerical modelling, the BMHSZ was predicted. Analysis of seismic facies and the location of the modelled BMHSZ suggested that lateral gas migration along the base of hydrate stability was constrained to approximately 490 m water depth. In shallower depths, strong lithological heterogeneity strongly influences fluid flow, impeding updip migration. Below the active methane seepage sites, the top 10 m

below sea floor experienced seasonal hydrate formation and dissociation episodes, potentially leading to increased fracturing that can enhance seepage during warmer bottom water conditions.

5.6. Auxillary material

Table S5.1. The seismic facies template is based on SYSIF data for the interfan area.

Stratified Facies (SF)		Description
		SF 1: This facies is characterized by parallel to sub-parallel, low to medium amplitude reflections with a conformable top and bottom. The low amplitude indicates fine-grained deposits like plumite.
		SF 2: This facies is similar to SF 1 but exhibits medium to high amplitude reflections. The high amplitude indicates coarser sediments such as sand.
		SF 3: The facies is similar to SF 2 but exhibits semi-continuous, choppy, and isolated reflections (maximum length ~200m). They show interbedding with chaotic reflections. The medium-high amplitude reflects the deposition of coarser sediments and interpreted as isolated turbidite.
Chaotic Facies (CF)		
		CF 1: This facies exhibits low-amplitude, discontinuous reflections that are nearly transparent and structureless. They fill paleo erosional surfaces. The facies represents glaciogenic debris flows filling pre-existing topographic lows.
		CF 2: Similar to CF1 but have a lenticular/lenoid geometry. CF 2 deposits are structureless with an irregular erosional base, indicating their gravity-driven nature.
		CF 3: Similar to CF 2 but has an overall rhombic geometry, and oblique depositional succession. The boundary between stacked lobes shows a moderate amplitude. This facies represents debris lobes resting on each other, indicating multiple local episodes of debris flow deposition.
		CF 4: This facies exhibits a set of concave low to medium amplitude semi-continuous reflections. These packages represent bottom set of the S-shaped glaciogenic/prograding clinoform.

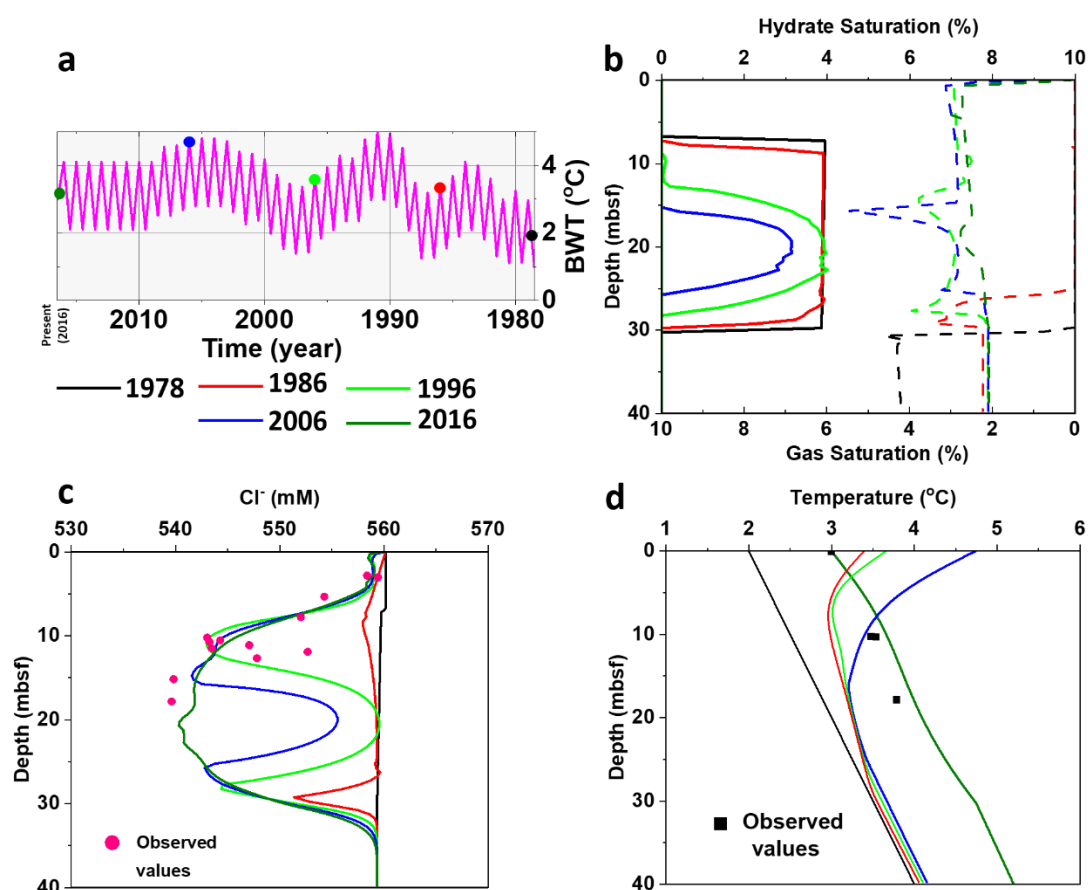


Figure S5.1. Modelled evolution of hydrate layer in response to variations in BWT at a water depth of 405 m off west Svalbard and simulation results showing the impact of BWT fluctuations on hydrate dynamics from 1978 to 2016. **(a)** The BWT record during 1978–2016, incorporating seasonal fluctuations of ± 1 °C. **(b)** Methane hydrate saturation throughout the simulation period, starting at 4 % in 1978. Solid lines indicate hydrate saturation, while dashed lines represent gas saturation. There is a decline in hydrate saturation due to progressive thermal dissociation. **(c)** Dissolved chloride concentration shows a freshening trend as hydrates dissociate. The final chloride anomaly matches with salinity observed during MeBo drilling. **(d)** The temperature profile in the sediment column, with black squares denoting measured temperatures in the MeBo drill hole at 405 m water depth.

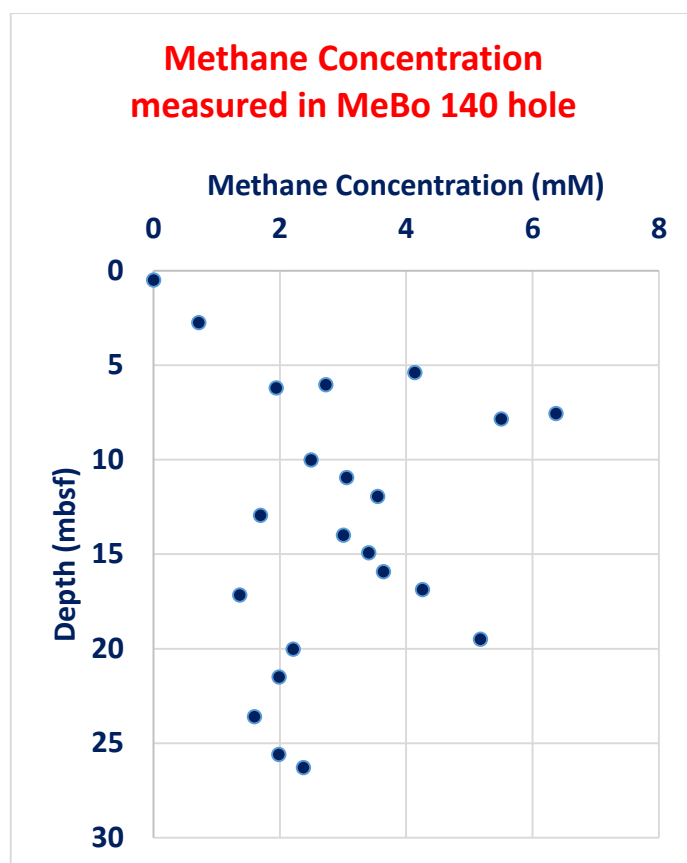


Figure S5.2. Subsurface Methane concentration profiles of dissolved methane from MeBo 140 drill hole at 402 m water depth, UCS.

Chapter 6

Conclusions and outlook

- (1) Accurate assessments of methane hydrate occurrence, distribution, and dissociation in Arctic continental shelves and slopes are crucial for comprehending their response to long-term oceanographic changes. To achieve this, effective ground-truthing strategies are required, employing high-resolution seismic surveys to refine hydrate distribution and characterization, ocean drilling to directly sample hydrates and assess dissociation processes, and seafloor observatories to monitor methane seeps and track environmental fluctuations. This research provides insights into the behaviour of methane hydrates, the Weichselian glaciomarine stratigraphic framework, and fluid migration patterns in shallow glaciogenic sediments of the interfan region between the Kongsfjorden and Isfjorden Trough Mouth Fans, utilizing high-resolution seismic data. It marks an advancement over preceding studies conducted by Sarkar et al. (2011) and Rajan et al. (2012).
- (2) The cause of methane hydrate dissociation, leading to methane seeps on the continental slope off west Svalbard, remains uncertain. Computer simulation results are helpful in visualizing how methane hydrates respond to oceanic changes under various scenarios. This work has contributed to refining our understanding of methane hydrate dynamics in response to historical oceanographic changes over the last 11000 years. It combined the impacts of ocean warming and relative sea level (RSL) changes in governing marine hydrate dynamics. Previous research by Wallmann et al. (2018), utilizing the UiT RSL change model, predicted a RSL fall over the past 8000 years, with the resulting depressurization as the cause of hydrate dissociation (Figure 6.1). However, this work introduces an alternative ICE-6G_C model, suggesting a RSL rise during that period, contradicting the earlier findings.
- (3) Comprehensive scenario testing reveals multiple episodes of hydrate formation and dissociation over the past 11,000 years. In the simulation considering the UiT RSL curve, along with realistic bottom water temperature (BWT) fluctuations, it is evident that BWT also influences hydrate dynamics and methane seepage patterns. Warmer BWTs prevent shallow hydrate reformation, allowing gas venting, while colder BWTs facilitate reformation (Figure 6.2). The simulation considering the ICE-6G_C RSL curve shows hydrate dissociation is exclusively caused by bottom water warming. Although the exact scenario remains undetermined, hydrate dissociation was complete before the late 1970's bottom water-cooling, when new hydrates could

reform. Subsequent episodes of BWT rise caused more recent hydrate dissociation, explaining the present-day methane seeps. Seasonal BWT variations strongly influence hydrate dynamics and methane flux. Accelerated ocean warming could lead to downslope migration of the hydrate stability limit, emphasizing the importance of monitoring warming trends in Atlantic waters. Continued warming may accelerate hydrate destabilization process in the near future and underscores the importance of sustained research efforts.

- (4) Methane-derived carbonates provide insights into environmental conditions during their formation, aiding in the assessment of past methane hydrate stability. Geochemical studies on seep carbonates can help to reveal gas hydrate dynamics offshore west Svalbard. Oxygen stable isotopes can be used to infer composition of seepage fluid, source and past precipitation temperatures (Feng et al., 2014 and the references therein). U/Th and ^{14}C dating methods can help to determine the timing of carbonate precipitation (Aharon et al., 1997). While previous endeavours by Berndt et al. (2014) concentrated on carbonates at specific seepage sites, a more systematic transect across the hydrate retreat zone could be useful to reveal lateral variations in seepage dynamics, validating numerical simulation-based predictions and contributing to our understanding of seepage patterns along the featheredge of gas hydrate stability. Seafloor backscatter images and high-resolution 3D seismic data covering a broader area on the upper slope can aid in the identification of the seafloor carbonate crusts.

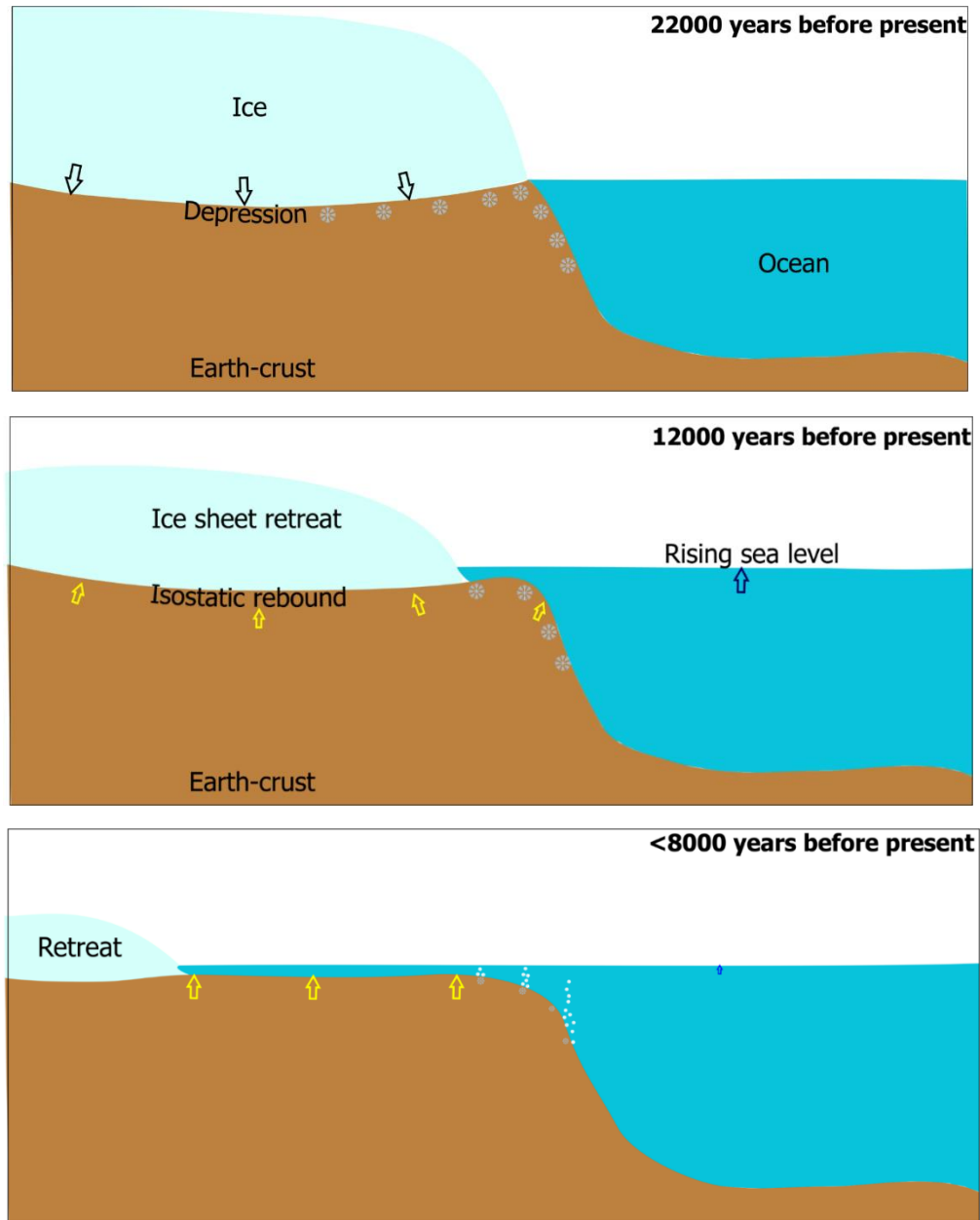


Figure 6.1. Schematic cartoon illustrating the oceanographic changes proposed by Wallmann et al. (2018) that lead to hydrate dissociation. **(a)** During the Last Glacial Maximum (LGM), the shelf was glaciated, and hydrates were stable on the upper slope. **(b)** Around 12 ka, as the ice retreats and melts, there is isostatic rebound and eustatic sea level rise, but the rising sea level rate exceeds the uplift rate of the seabed. **(c)** Around 8000 years before the present (as of 2016), the isostatic seabed uplift rate surpasses the eustatic sea level rise, causing a shallowing of water depth at the uppermost slope. The depressurization resulting from this shallowing triggers methane hydrate dissociation.

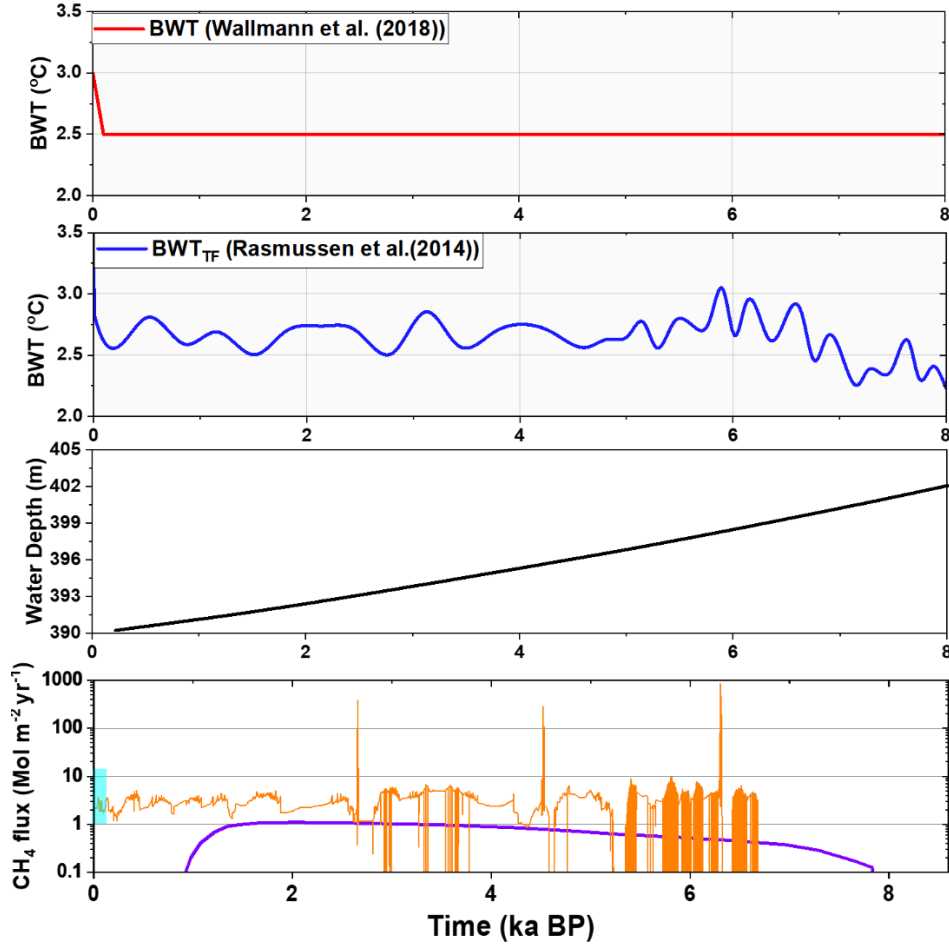


Figure 6.2. Comparison of computer simulation results between Wallmann et al. (2018) and the present study. The simulations utilized past bottom water temperature (BWT) reconstructions and relative sea level (RSL) predictions from the UiT GIA model to understand methane hydrate behaviour. **(a)** The simulation by Wallmann et al. (2018) considered a flat BWT curve at 2.5 °C over 8000-100 years before the present, followed by a linear rise to 3 °C in the last 100 years. **(b)** The actual BWT_{TF} curve showed temperature fluctuations that were not considered by Wallmann et al. (2018) and was used in this study to simulate the behaviour of methane hydrates. **(c)** Water depth changes predicted by the UiT GIA model reveal a shoaling trend over the past 8000 years. **(d)** The magenta curve represents methane flux at the seabed, predicted by Wallmann et al. (2018), resulting from depressurization-induced hydrate dissociation. The orange flux curve, derived from a simulation incorporating BWT_{TF} curve, shows diminished methane seeps during episodes of cooler BWTs when methane was trapped as shallow hydrates. The methane flux persisted into the present, matching with observed fluxes (cyan bar). This study highlights the role of BWT temperature, in addition to RSL changes, in hydrate dynamics—a factor that was not considered by Wallmann et al. (2018).

- (5) Understanding the dynamic behaviour of past ice sheets in response to climate-controlled fluctuations, relative sea level (RSL) changes, and the interaction of the ice sheet with the underlying topography is crucial for predicting future climate change impacts on ice sheets. Despite the availability of a substantial amount of stratigraphic information for offshore western Svalbard region, challenges persist in regional correlation. These issues arise from differing models based on land sections and the absence of comprehensive correlation between terrestrial and marine records. The former is often fragmentary and prone to contradictory interpretations, while the latter provides insight into changes in ocean circulation and glaciomarine records that have better preservation potential compared to terrestrial counterparts.
- Reconstruction of marine record is based on seismic data. Improved vertical and horizontal resolution of seismic data along with the age of key horizons are necessary to refine stratigraphic models. Although 2D seismic lines collected with air gun source were effectively used to infer the glacial history spanning the past 2.7 Ma, this data shows limited vertical resolution in resolving the Weichselian section because of the broad wavelet that masks stratigraphic details. To this end, TOPAS and SYSIF data are more effective. The study showed the potential to use the SYSIF seismic lines along the Svalbard and Barents Sea regions for more refined stratigraphic characterization and improvement in understanding of glaciomarine sedimentation patterns providing invaluable clues related to glacier build-up and decay. In this thesis, a conceptual model for inferring shelf-edge glaciations and retreat by analyzing glacial debris stacking patterns is presented (Figure 6.3). This model can be applied to similar studies elsewhere in glaciated margins. The across-slope seismic lines effectively show the lateral facies changes from the shelf break and upper slope. Conducting similar studies that encompass the inter-fan and Trough Mouth Fans (TMFs) will enhance Weichselian stratigraphic correlation, provide estimates on glacial erosion and sediment volume, and shed light on subglacial processes influencing glacial dynamics. Moreover, the stratigraphic results could provide crucial inputs to coupled models integrating ice-sheet dynamics, erosion, and sediment transport (Patton et al., 2022). These models may play a key role in comprehending a more refined glacial history of Svalbard.

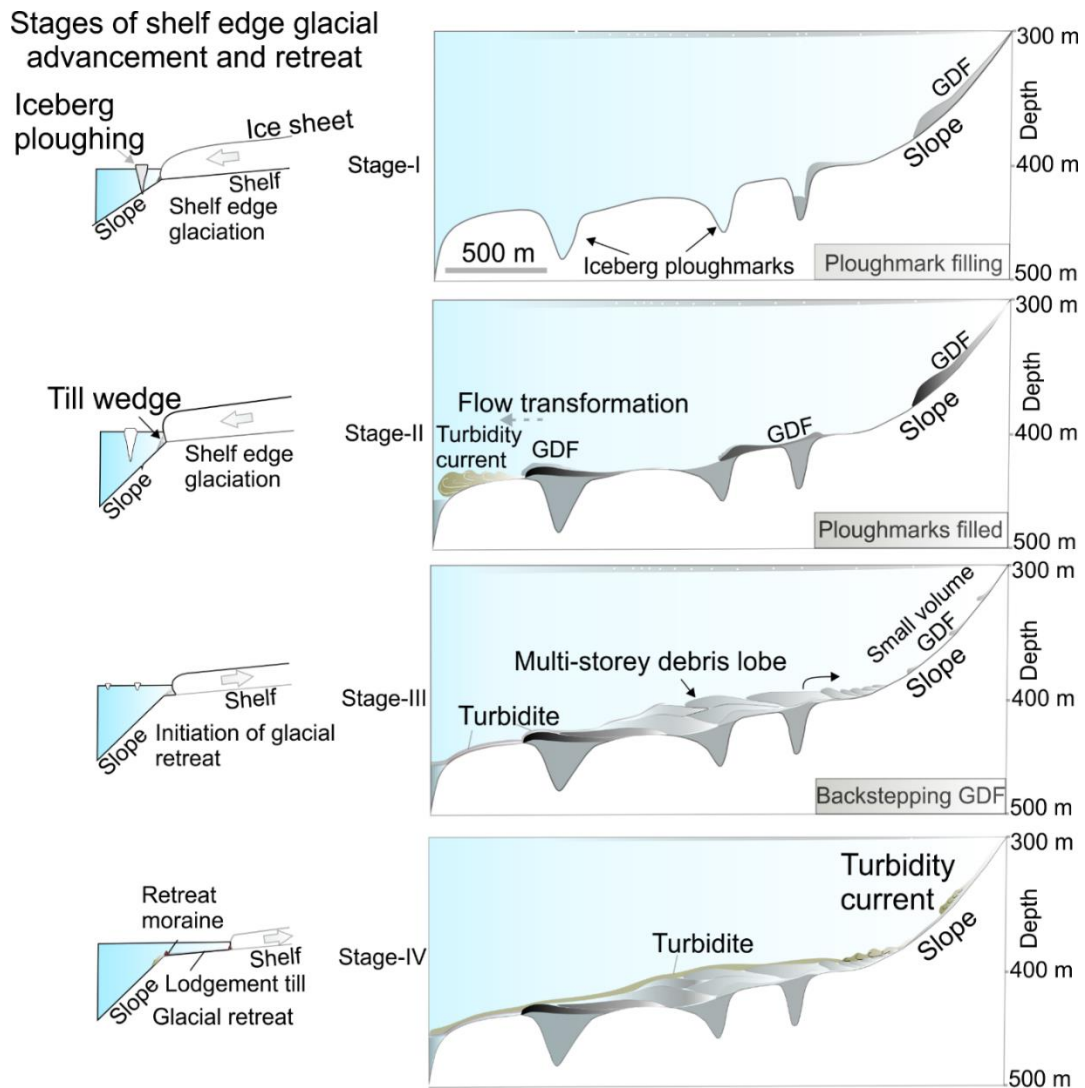


Figure 6.3. Schematic cartoon illustrating the patterns of glacial debris filling an erosional surface on the upper slope in relation to the position of the ice sheet at the shelf break. Stage-I depicts ploughmarks formed by deep-keeled icebergs during a major calving episode. At Stage-II, early glacial debris flows (GDFs) originated from the shelf edge, cascading downslope and filling the grooves. During Stage-III, multiple debris flow lobes emerged and stacked up in a vertically aggrading manner when the ice sheet remained grounded at the shelf break. This filling pattern could be applied to explain the debris-stacking pattern observed in unit 5B, as shown in Figure 6b and Figure 7a. As the debris supply begins to diminish, a retrogradational backstepping of small-volume debris is observed. The retreat of the ice sheet from the shelf break during Stage-IV resulted in a decrease in debris supply, leading to the deposition of smaller volume of glacial debris lobes on the upper slope. Enhanced melt water supply generates turbidity currents that deposit turbidites.

- (6) The stability of the seabed in Polar continental shelf and slope regions is of paramount importance for the development of marine infrastructure. A comprehensive geotechnical assessment is required prior to establishing marine infrastructure and laying submarine telecommunication cables in these vulnerable areas. Common challenges include shallow overpressure, caused by gas hydrate destabilization or under compacted clay, leading to submarine slides that may progress into slumps, debris flows, and turbidity currents.

Understanding the likely behaviour of glaciogenic diamicton in the context of sliding and remobilization is crucial. In this context, the reconstruction of past glacial debris flow behaviour and the measurement of material strengths become essential. Although this study utilized a simple modelling approach to understand the plausible pattern of past glaciogenic debris flow evolution, the geotechnical and sedimentological properties of these flows remain undetermined.

Few shallow cores from the uppermost slope in the interfan provided insights into the glacial debris flow materials, revealing the matrix-supported nature of diamicton ranging from gravely/pebbly-clay to silty clay. Seismic data showed the potential to study the internal structures and stacking pattern of the glacial debris lobes.

A thorough understanding of the rheology of glacial debris flow materials needs to be developed both along and across the slope to enhance our understanding of past glacial debris flow dynamics in the interfan and glacial fan regions, especially those with longer runouts. Comparisons with the glacial debris flows at the Trough Mouth Fans (TMFs), which are either strongly influenced by meltwater or devoid of substantial meltwaters (Bellwald et al., 2020), with those of the interfan regions will further contribute to broadening our understanding of the commonalities and differences in glacial debris flow hazard potential (e.g. erosion and run out potential) between the TMFs and the interfan regions.

Understanding the behavior of glaciogenic diamicton is essential for several marine industry applications. It aids in hazard assessment by predicting submarine landslides, protecting submarine telecommunication cables, and ensuring the stability of oil and gas platforms. Additionally, it supports the placement and design of

foundations for offshore wind turbines and informs environmental impact studies to better manage and conserve marine ecosystems. These insights help mitigate risks and enhance the safety and effectiveness of various marine infrastructures and activities.

- (7) The study utilized 3D seismic data and high-resolution SYSIF images to reveal the characteristics of glaciomarine strata capable of storing methane in the hydrate retreat zone. Investigating the storage and release mechanisms of methane at this feather edge remains a key focus of research. Limited seismic coverage, until now, hindered the examination of this zone. The study brought attention to fragmented strata, characterized by patchy sandy or silty turbidite remnants, which may serve as localized reservoirs. These reservoirs are encased within muddy diamicton, primarily acting as an effective seal. Despite this sealing, methane emerges at the seabed, suggesting sustained fluid flow through interconnected pathways. Complex oblique debris stacking and erosional unconformities baffle the vertical fluid flow to the seafloor. Insights from seismic data indicate that these flows are primarily sustained by open fractures capable of transmitting fluids. These fractures could be generated during the overpressure resulting from hydrate dissociation in muddy diamicton, where gas flow is usually less efficient. In this work, a narrow 3D seismic dataset above existing methane seepage sites was studied to examine subsurface reservoirs, seals, and fluid flow patterns. Expanding this area laterally can provide insights into the 3D architecture of potential reservoirs, particularly those vulnerable to ocean warming in the near future.

The study's findings have significant implications for hydrocarbon exploration and production. Revealing glaciomarine strata capable of storing methane enhances the identification of new potential hydrocarbon reserves. Understanding the characteristics and distribution of fragmented glacial reservoirs improves the accuracy of reservoir models, which is crucial for efficient resource extraction. Additionally, insights into seal integrity and fluid pathways help assess the risks of methane leakage, ensuring safer drilling operations. The use of 3D seismic data and high-resolution imaging enhances the detection and mapping of subsurface reservoirs, providing a more detailed understanding of the subsurface. Moreover, knowledge of how ocean warming affects these reservoirs informs strategies to manage and mitigate risks associated with hydrate dissociation and methane release, which is increasingly important in the context of climate change.

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Appendix: Uninterpreted seismic images

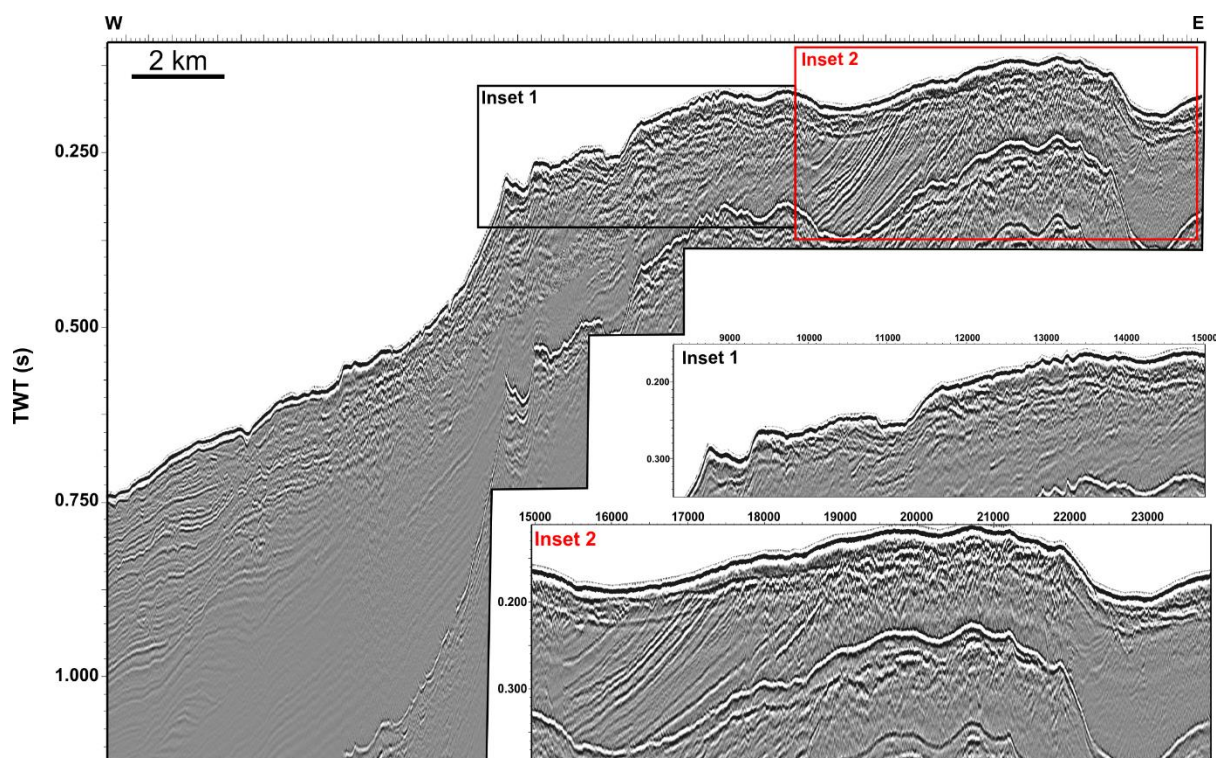


Figure A1: Uninterpreted slope perpendicular seismic line JR269-14, shown in Figure 4.4

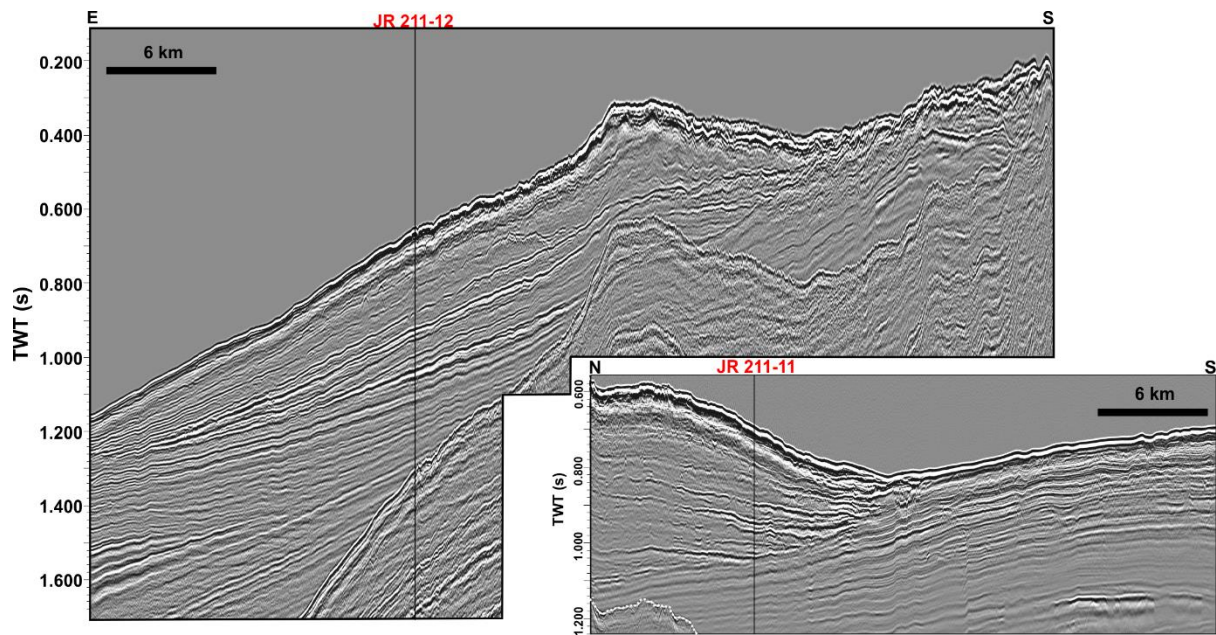


Figure A2: Uninterpreted slope perpendicular seismic line JR211-11 and slope parallel seismic line JR211-12, shown in Figure 4.5



Figure A3: Uninterpreted slope perpendicular SYSIF seismic line 2.1, shown in Figure 4.8.



Figure A4: Uninterpreted slope perpendicular SYSIF seismic line 2.4, shown in Figure 4.9.

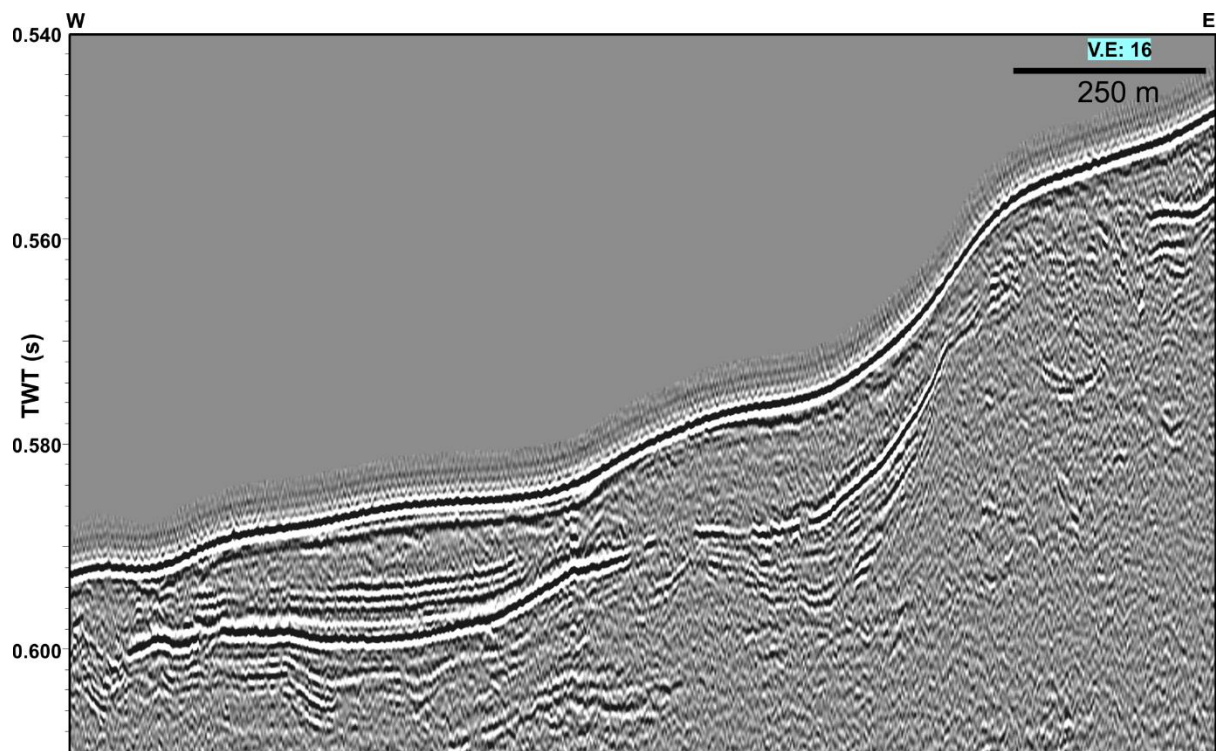


Figure A5: Uninterpreted slope perpendicular SYSIF seismic line 2.4, shown in Figure 4.10.

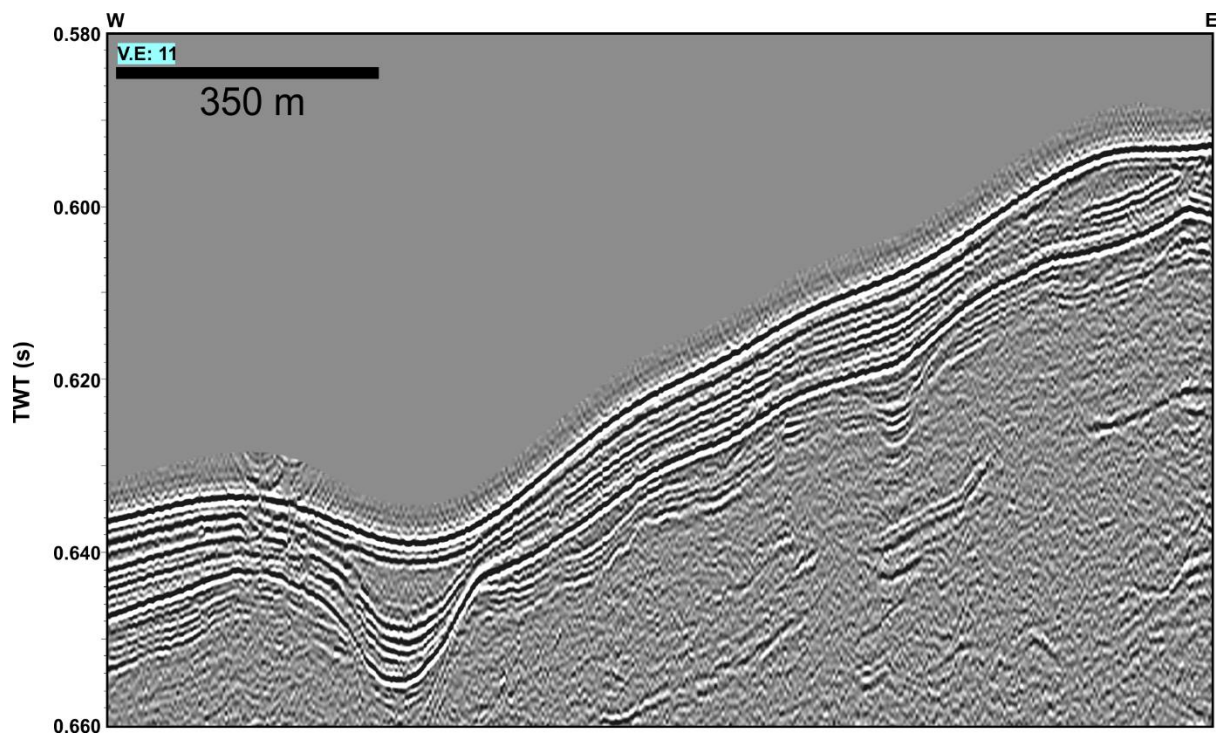


Figure A6: Uninterpreted slope perpendicular SYSIF seismic line 2.3, shown in Figure 4.11.

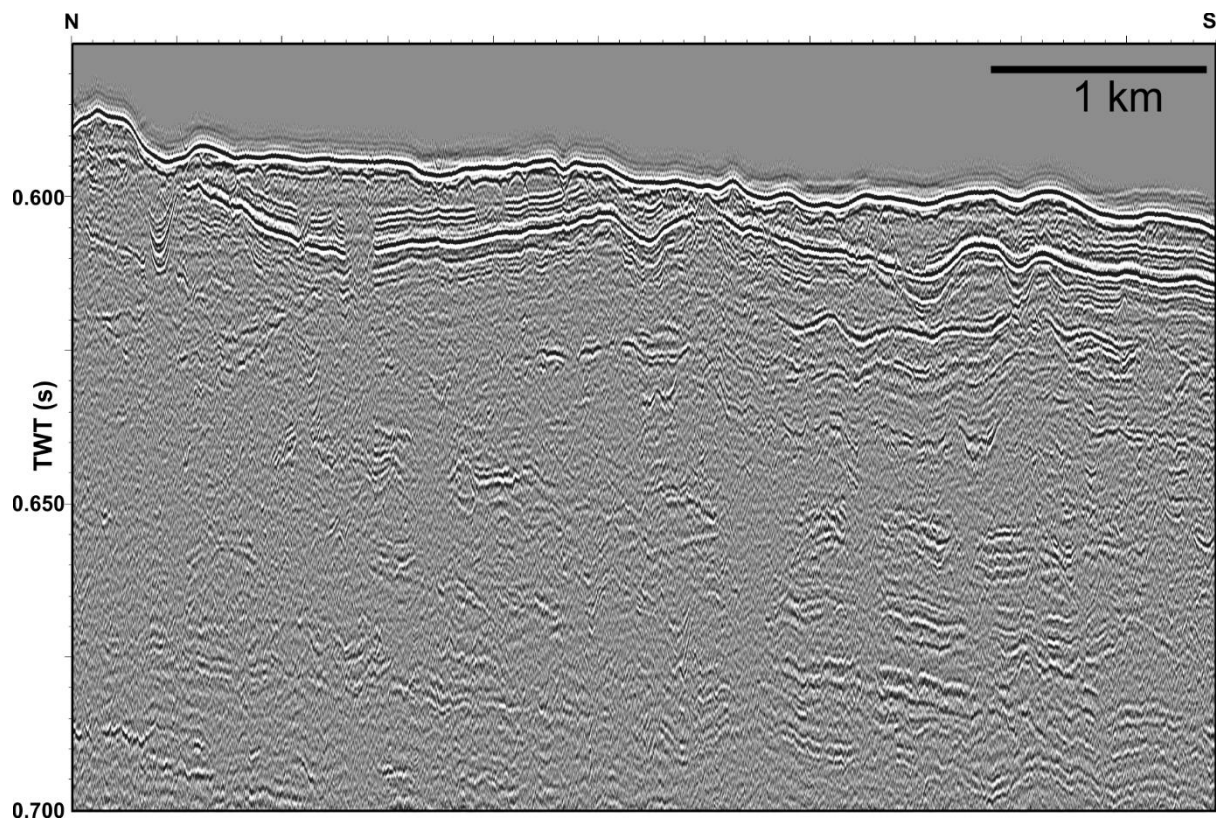


Figure A7: Uninterpreted slope parallel SYSIF seismic line 3.11, shown in Figure 4.12.

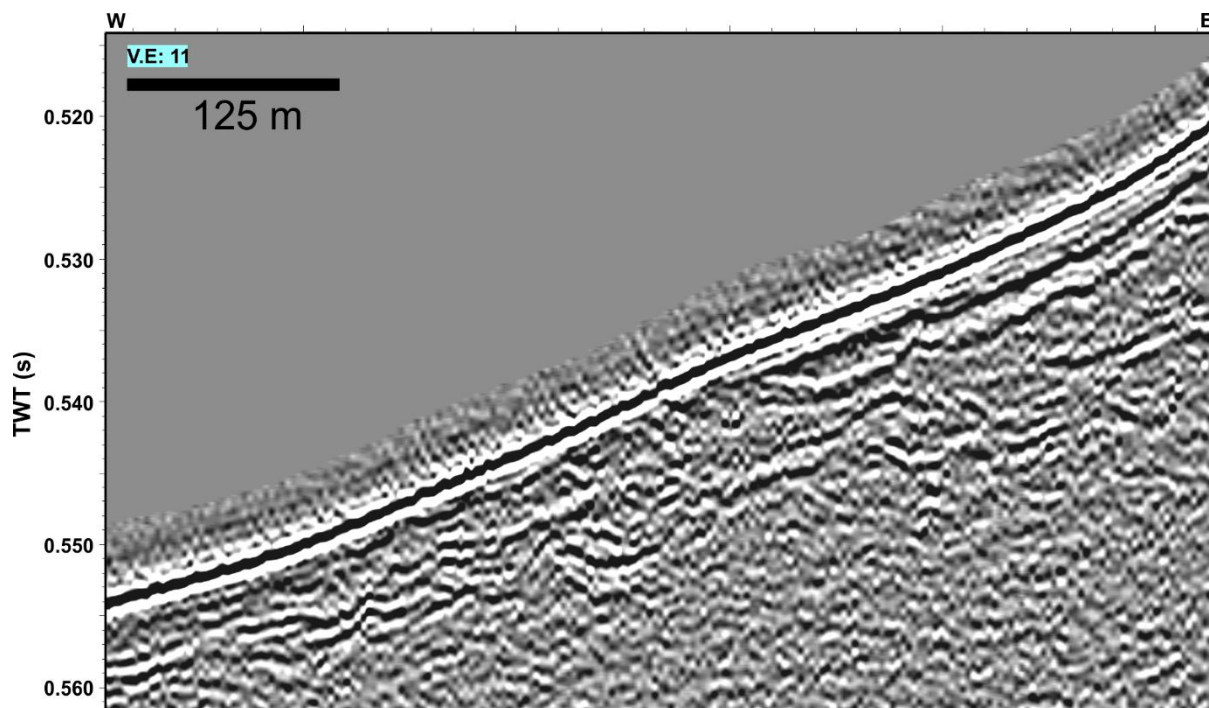


Figure A8: Uninterpreted slope parallel SYSIF seismic line 2.1, shown in Figure 4.13.

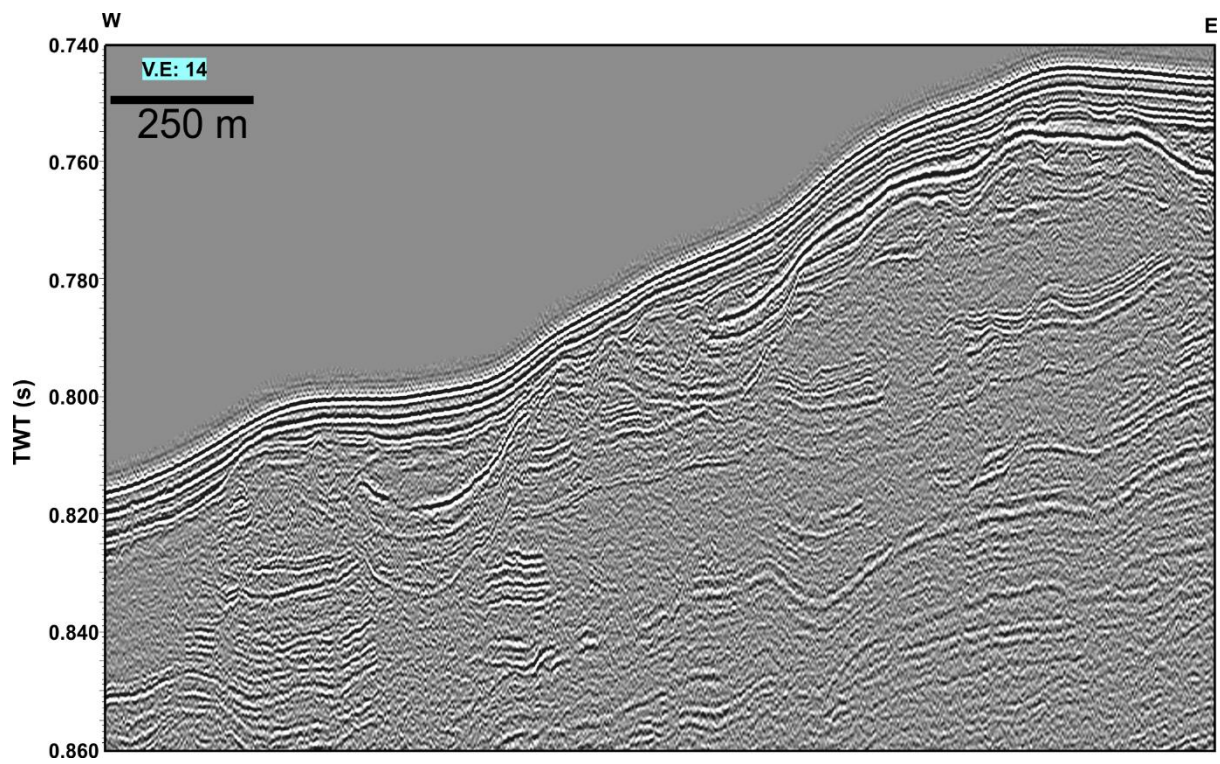


Figure A9: Uninterpreted slope perpendicular SYSIF seismic line 2.1, shown in Figure 4.14.

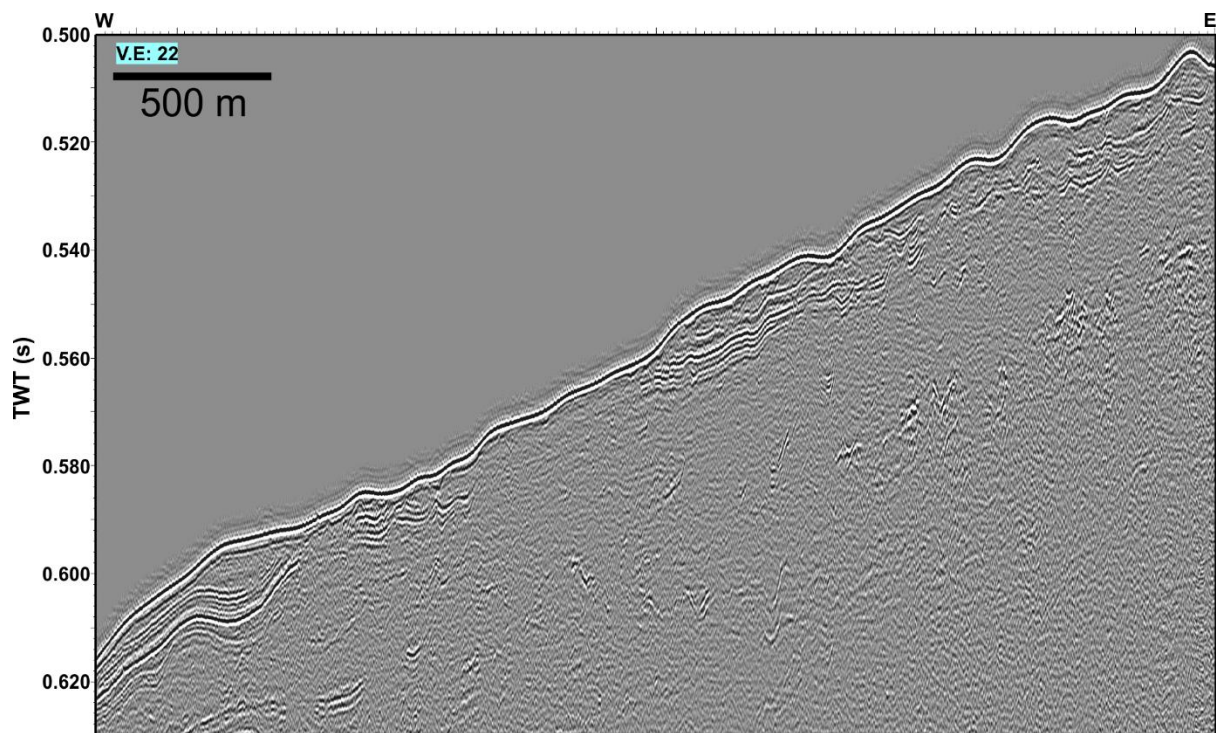


Figure A10: Uninterpreted slope perpendicular SYSIF seismic line 3.1, shown in Figure 4.15.

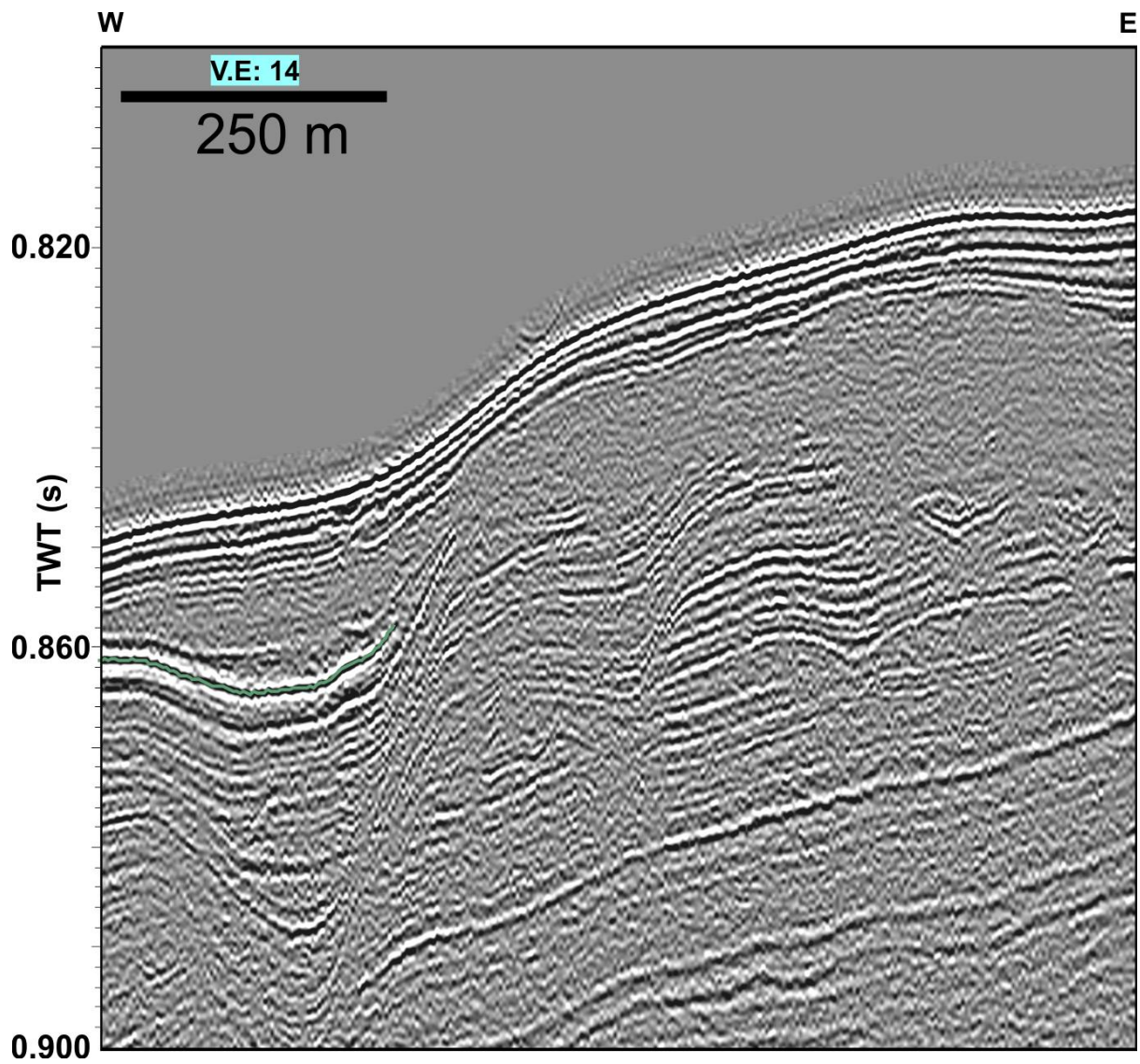


Figure A11: Uninterpreted slope perpendicular SYSIF seismic line 2.1, shown in Figure 4.16.

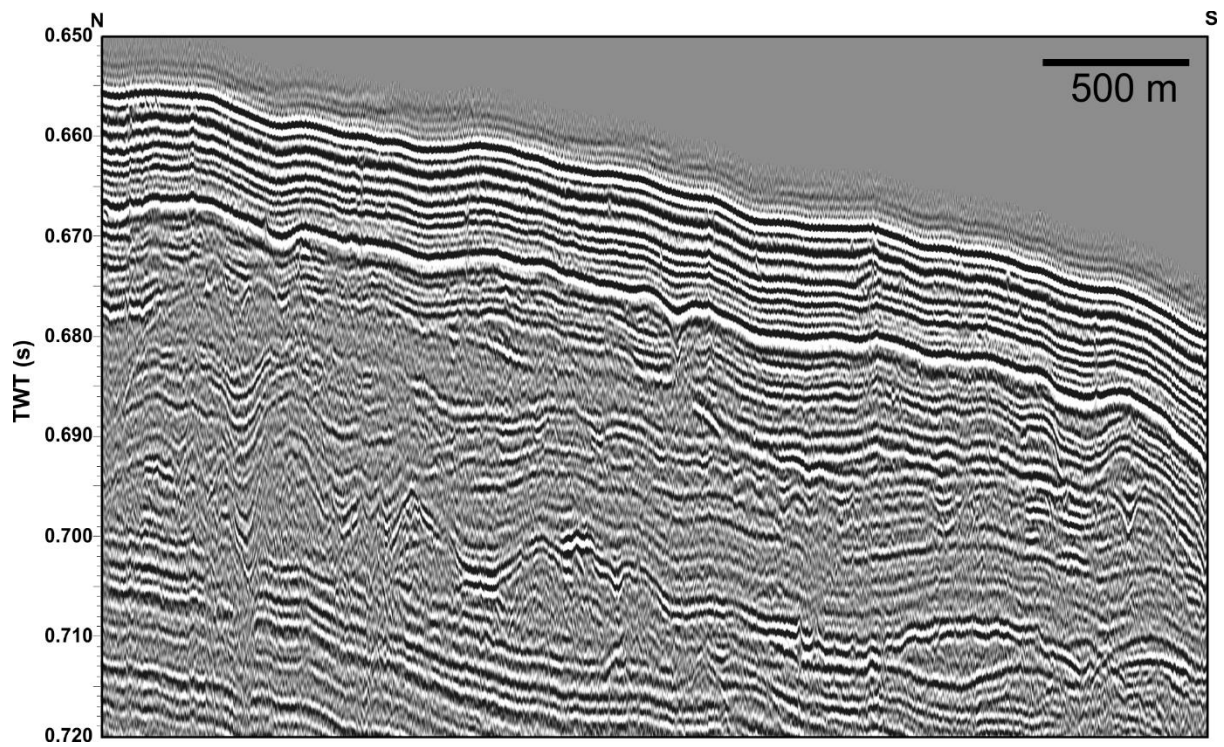


Figure A12: Uninterpreted slope parallel SYSIF seismic line 3.13, shown in Figure 4.17.

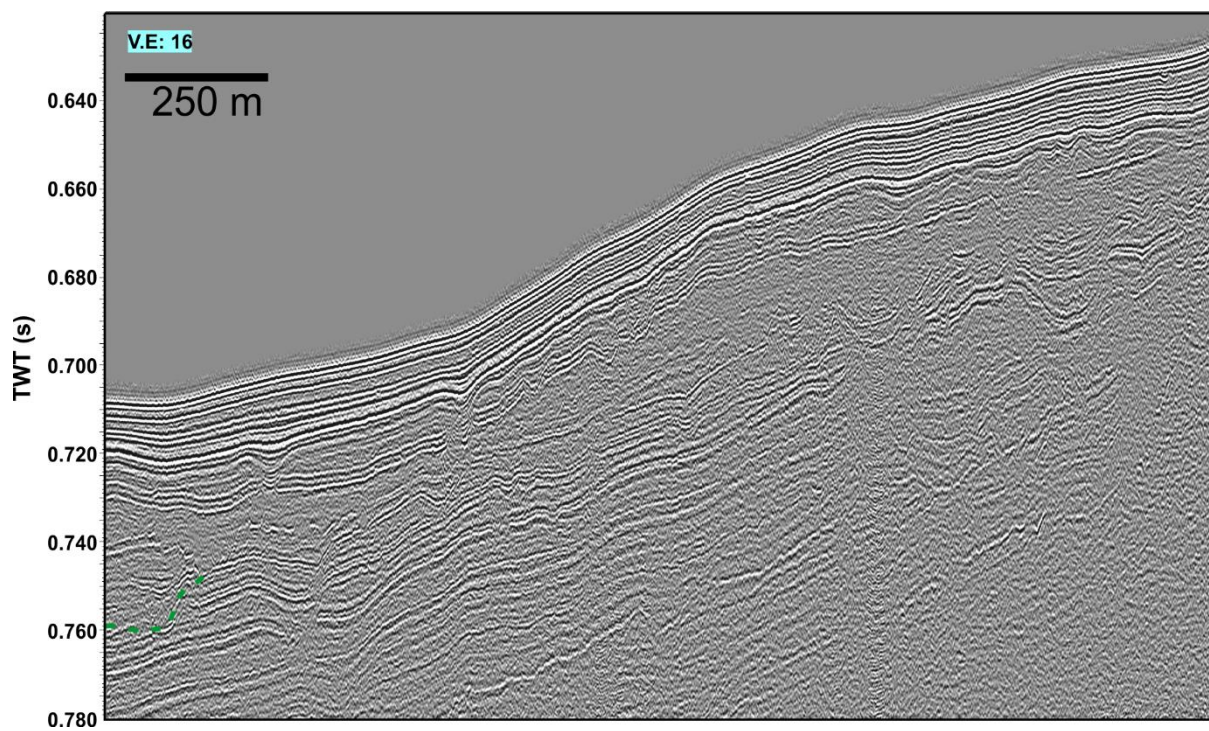


Figure A13: Uninterpreted slope perpendicular SYSIF seismic line 2.1, shown in Figure 4.18.

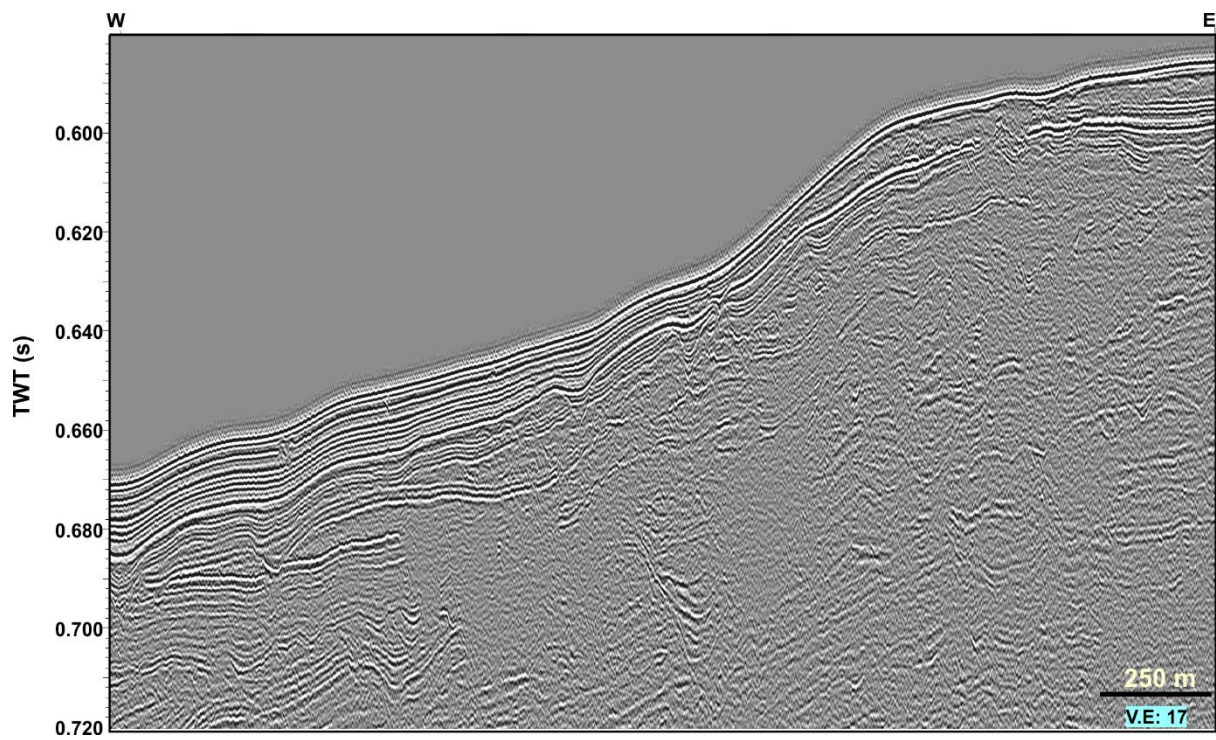


Figure A14: Uninterpreted slope perpendicular SYSIF seismic line 2.4, shown in Figure 4.20.

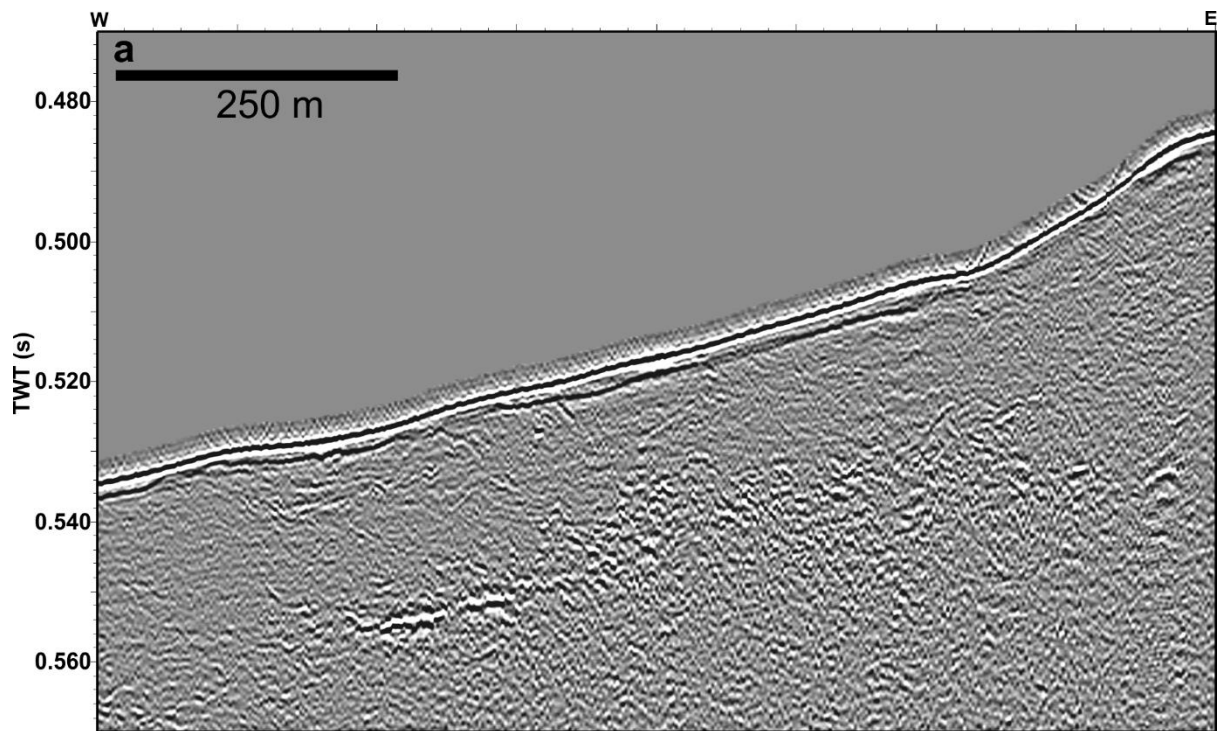


Figure A15: Uninterpreted slope perpendicular SYSIF seismic line 3.11, shown in Figure 4.21.