Title

Evolution of contourite drifts in regions of slope failures at eastern Fram Strait

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Abstract

Geotechnical characteristics of contouritic deposition often lead to preconditioning slope instabilities and failures along glaciated and formerly glaciated continental margins. However, internal depositional geometry is also an important factor in triggering instabilities. This work highlights the importance of the tectonic and oceanographic evolution of the Northwestern (NW) Svalbard margin in determining the build-up and the internal structure of contourite drifts and the subsequent type of slope instability. The analysis of seismic reflection data reveals that the presence of two contourite drifts on the flank of an active spreading ridge in the Fram Strait - NW Svalbard margin - in an area of extensive slope instability, had a major impact on the evolution of slope failure. The presence of a slope sheeted drift (or plastered drift) led to the development of rotational/translational mass movement at water depth < 2500 ms, whereas at water depth > 2500 ms the presence of sediment waves facilitated the formation of planes of shear that led to internal deformation of the lower slope through a process of slump/creep. The well-documented high seismicity of the area might have provided the necessary energy to trigger the slope instability.

1. Introduction

Contourite drifts are sediment accumulations ranging from 50 to >10^6 km^2 [1] controlled by contour currents (i.e., bottom currents that flow parallel to the slope or the continental rise) [e.g. 2]. Previous work has documented that the initiation of contour currents is strongly dependent on thermohaline circulation, wind-driven circulation systems [1] and ocean circulation changes driven by large-scale processes such as plate-tectonic events [3]. Tectonic induced rifting and subsidence, alteration of the morphology of the slope by erosion and sedimentation can create accommodation space for sediment deposition and may force changes in the flow regime [4,5]. Morphologic obstacles promote variations in flow velocity;

for example, erosion may be promoted in the center of the current, while deposition may take place both downslope and up-dip of the core of the current [2,6-8]. Along-slope contour currents can form a variety of sediment drift morphologies [9-11] depending on grain size, amount of transportable sediments available, current speed and turbulence, slope steepness, or an interplay between these different factors, including downslope processes such as turbidity currents [12-14].

Four main types of contourite drifts exist: sheeted, mounded-elongate, patch and channel-related drifts. Sheeted drifts tend to form in areas characterized by relatively slow deposition rates and they can cover an area of $>10^3$ km^2. A further distinction among sheeted drift, based on their occurrence and yielding, comprises: abyssal sheeted drifts, slope sheeted drifts (also referred to as *plastered*) and channel sheeted drifts [8,7,13].

Many factors may lead to the instability of contourite drifts and trigger small and large-scale submarine landslides [15-20]. When a contourite drift develops along the slope, over-steepening of the slope or undercutting by erosion can have consequences for slope instability. Undercutting or erosion at the toe of a drift has been suggested as a potential controlling factor for slope instabilities in the Mediterranean sea [21,22] as well as in parts of the Fram Slide Complex in the Arctic [23]. Moreover, long-lasting and stable bottom currents tend to result in very well sorted sediment deposits [13,12]. Homogeneity in grain size is one of the characteristic favoring high water content and less friction between individual grains, making well-sorted sediments less resistant to shearing [19].

Slope failures can occur due to the presence of overpressure within the pore space of sediments and subsequent drop in shear strength [24]. On formerly glaciated margins, cyclic sediment deposition and high-fluid content in contourites sealed by thick sequences of impermeable glacigenic debris may furthermore generate overpressure within contouritic layers [15]. Overpressure can also be generated by the accumulation of free gas within permeable contourite layers that are sealed by impermeable layers. Gas from deeper reservoirs or dissociation of shallower gas hydrate during ocean warming may contribute to unstable slope conditions [25].

Gas hydrates are compounds consisting of hydrocarbons entrapped in water cages. They form within a certain range of temperature and pressure conditions, depending on water salinity and the composition of the sourced gas [26]. The base of the zone where gas hydrates are stable on continental margins is often identified by the presence of a bottom simulating reflection (BSR) in seismic profiles, highlighted by a high amplitude, reversed polarity, cross-cutting reflection which mimics the sea bottom [27]. The occurrence of a gas hydrate related BSR is an indicator of free gas beneath overlying impermeable gas hydrate-saturated sediments formed within the gas hydrate stability zone (GHSZ) [28-30]. The high negative impedance contrast indicates a sudden decrease in P-wave velocity at the phase boundary between gas hydrate-saturated sediment above and the accumulation of free gas underneath [27].

This study describes a complex geological slope environment covering one contourite drift along the > 3000 m deep slope at the western flank of the Yermak Plateau in the Arctic Fram Strait of the NW Svalbard continental margin. Slope failures and gas hydrates/free gas (based on observations of BSRs) are widespread in this region (Fig. 1) [31,23]. Through the analysis of high-resolution reflection seismic profiles from four downslope transects, including an established seismic stratigraphy, we document the partial extent, the seismic signature and the geometry of the contourite drift. We furthermore reconstruct its growth.

2. Study Location and Oceanic Setting

Our study focuses on contourite deposits along a slope that extends from the western flank of the Yermak Plateau (YP) towards the junction between the Molloy mid-ocean ridge and the Spitsbergen Fracture Zone between 79°31’28’’ N and 80°14’42’’, and 1°32’48’’ and 5°37’33’’ E (Figs. 1 and 2). It covers an area of 5500 km^2 between 850 m and 4200 m water depth and contains 17 slides [23]. Due to the remarkable proximity of the continental shelf break to the mid-ocean ridge, this deep marine setting is not a classic abyssal plain with a continental rise. Instead, the continental slope terminates almost directly over the rift valley where both active tectonic and sedimentological processes occur (Fig. 1).

The eastern Fram Strait is characterized by the continuous northward flow of warm and saline Atlantic-derived water brought by the West Spitsbergen Current (WSC), which is a continuation of the North Atlantic Current (NAC) [32] (Fig. 2). The WSC splits at ~79° into three branches: the western branch joins the southward flowing Arctic-derived water within the Eastern Greenland Current; the eastern branch flows eastward along the northern Svalbard margin, and the Yermak Plateau branch flows along the western flank of the YP entering the Arctic Ocean [33,32]. A mooring deployed from September 2006 to July 2007 near our area (FEVI14, 5.1645°E, 79.6012°N, 2742m depth [34]) revealed an averaged meridional bottom velocity of 2.2 cm/s, reaching up to 23 cm/s in winter and late spring. The deepest water mass flowing within the WSC is the Norwegian Deep Sea Water (NDSW), presenting salinity and temperature values of > 34.91 PSU and < -0.9 C° respectively [35,36].

Sediments are supplied to the deep Fram Strait by downslope transport from the Svalbard shelf [35,37]. Here, dense shelf water is produced in winter due to persistent cold conditions and the consequent formation of polynyas and brines [38-40]. When these water masses reach the shelf edge, their high velocity and turbulence allow for the erosion and the transport of shelf sediments in suspension [41]. Episodically, these dense plumes reach the deep Fram Strait, where the sediments are transported and eventually redeposited by contour currents [35].

3. Seismic Stratigraphy and Geological Setting

The opening of the Fram Strait during mid-late Miocene allowed the onset of oceanic circulation between the North Atlantic and Arctic Ocean which is a prerequisite for sedimentation controlled by oceanic circulation [42]. Rebesco et al. [35] suggested that, in addition to a tectonic pre-conditioning, the onset of strong currents is also connected to the late Cenozoic climate cooling, with the formation of cold and deep water in the Arctic Ocean. They also identified two contourite drifts in front of Isfjorden and Bellsund troughs extending along the deep slope between ~1200 and ~2000 m depth. The drifts are thought to be fed by plumes of dense shelf water generated in the Barents Sea, overflowing the Norwegian Sea Deep Water, which roughly flows at depths where sediment drifts exist [35]. They propose that the onset of the Isfjorden and Bellsund drifts occurred during the Early Pleistocene related to glacial expansion ~1.3 Ma ago.

Three main stratigraphic units have been defined for the region [4]. Correlation to cores from boreholes drilled during Ocean Drilling Program Leg 151 [43,44] provides the age control for these seismic stratigraphic units: YP-1, the oldest unit, is composed of syn- and post-rift sediments deposited directly onto the oceanic crust; the YP-2 sequence represents the onset of contourite facies deposition and is dated between 11 Ma and 14.6 Ma; YP-3 represents the beginning of glacially transported sediments, where contourites, glaciomarine turbidites, and debris flows are the predominant facies.

The boundary between YP-2 and YP-3 is estimated to be 2.7 Ma [4,45], and has been identified in the region comprising the YP, the Vestnesa Ridge and offshore Prins Karls Forland [46]. The gas hydrate system at the Fram Slide Complex is identified between ca. 50 and 300 meters below the sea-floor within stratigraphic unit YP-3 [31,23], however, seafloor seepage has not been documented.

4. Data and Methods

Six high-resolution 2D seismic lines were acquired in 2013, 2014 and 2015 aboard R/V Helmer Hanssen (Fig. 1). We connected four 25 m long streamers from the P-Cable seismic system (e.g., Petersen et al., 2010) to obtain a 100 m long streamer that recorded data in 32 channels. The source was a mini-GI air gun with a capacity of 15/15 in^3, fired every 5 seconds. Data processing steps included: insertion of navigation data, Common Depth Point-binning every 6.25 m, static corrections, bandpass filtering with a frequency of 10-20-400-500 Hz, amplitude corrections, Normal moveout correction, stacking, 2D Stolt Migration with a 1500 m/s constant velocity. The dominant frequency range of this data is 120-250 Hz allowing for a vertical resolution of 3.2 m (λ/4) at the seafloor assuming a water velocity of 1490 m/s. Seismic signal penetration reaches a maximum of ~1500 ms TWT beneath seafloor. The commercially available seismic interpretation software Petrel was used for seismic interpretation.

Bathymetry data stem from a hull-mounted Kongsberg Maritime EM300 multibeam echo sounder from different research campaigns with R/V Jan Mayen, further renamed R/V Helmer Hanssen, in 2008, 2009, 2010, 2011 and 2013. The EM300 operates with 135 beams, generating a horizontal resolution of ~25 m x 25 m at the depth of the study area. We integrated bathymetric dataset with the bathymetry data from Elger et al. [23].

The seismic stratigraphy was obtained by tracing the seismic horizons identified and dated by Mattingsdal et al. [45] in the Yermak Plateau region, based on data from the Ocean Drilling Program (ODP) Leg 151 [44], Hole 912. Compared to the chronostratigraphy based on the seismic units YP-1, YP-2 and YP-3 [4,44], the horizons dated by Mattingsdal et al. [45] allowed for a more precise age control. One seismic line used in this study crosses the site of the ODP hole 912 location (Fig. 1).

5. Results and Interpretation

5.1 Contourite drift

The seismic profiles (Figs. 3-8, see Fig. 1 for locations) show the sedimentary architecture of the slope over a distance of ~30 km N-S along the eastern Fram Strait (western YP). We have correlated reflections with those presented in (Mattingsdal) and the ODP hole 912 and find that the high resolution nature of the P-Cable data does not allow seismic penetration below the level of the ~7 Ma, within the YP-2 sequence. Consequently, the correlation of the age of the sediment drift with the chronostratigraphy of the IODP site 912 suggests that the sediment column of our study area belongs to the seismic unit YP-2 and YP-3[45,47]. Above ~2500 ms, we observe an extended convex-up mounded body characterizing the sedimentary environment of the slope throughout the study area, whereas below this depth the sedimentary body is highly deformed by faults and shear planes (Fig. 3 and 4). We interpret the mounded body as a sheeted contourite drift. More specifically, the layers of constant thickness over a large area and the slight decrease in thickness towards the shelf break are characteristics of a slope sheeted drift (or plastered drift) [7,3]. The base of the plastered contourite drift is > 5.8 Ma as it is observed below the 5.8 Ma reflection (Figs 3 and 7).

The three-order seismic elements description as proposed in the recent study by Esentia et al. [8] supports our interpretation of the sheeted sediment drift. The first order seismic elements (drift scale) suggest a sheet-like geometry reaching of ~700 ms thickness at the maximum penetration of the seismic signal (Fig. 3) and we do not observe regional discontinuities. The drift consists of medium to low amplitude reflections, indicating slight differences in velocities possibly due to different grain sizes, and therefore sediment sources. The second order seismic elements (depositional seismic units scale) show the characteristic features of a large size drift, as a series of broadly lenticular, convex-up seismic units and gently upslope-downlapping reflections constituting the uniform stacking pattern (Fig. 3) [8]. Indications of downlapping are observed on a surface comprised between the 1.95 and 1.5 Ma reflections (Fig. 3). Below this surface, the reflections present a less enhanced convex-up shape possibly suggesting that the reflections downlap at shallower depths, outside the coverage of our database. We infer that the depocenter deepened during the growth of the plastered sediment drift. The third order seismic elements (facies scale) present a significant change of seismic facies at ~2500 ms depth. Line12 shows the sharpest diagnostic sedimentary features (Fig. 4). The ~2500 ms depth marks the temporal change from continuous, sub-parallel, moderate-low amplitude reflections to a portion of the slope characterized by high amplitude to almost transparent reflections and regular, migrating waves (Fig. 4a). In addition, moat-levee structures appear throughout the sediment column. The wavy features consist of medium-low amplitude to transparent reflections and wavelength varies from 1 km to 2.2 km, showing a decreasing trend in wavelength from older to younger sediments (Fig. 4a). Comparable shapes and wavelength were observed by Lu et al. [48] at the Canterbury Basin. Similar to our interpretation, they interpreted these sediment waves as basinward facies of an elongated drift (A plastered drift being a type of elongated drift according to the previous classification by Faugères et al. [7]). The geometry of the wavy features forms lineaments that pinch out at marked escarpments on the seafloor (Fig. 4a). Similar features were observed by Rodriguez et al. [49] and interpreted as *potential shear planes*, at the Sawqirah contourite drift system in the Arabian Sea. Similar features in our study area, showing pinch outs at the seafloor (~2500 ms) and forming a marked escarpment suggest the occurrence of shear planes (Fig. 4b). This interpretation is supported by the presence of zones of transparent seismic signal (Fig. 4b) that suggest internal deformation, possibly generated by shear movements within the sediment column. We exclude that the moat-levee structures are due to a turbidity current as, in a regime of S-N flowing currents along an eastward-shallowing slope like in the present study, the levee structures develop on the left side of the downstream direction. We therefore interpret that the moat and levee structures relate to the onset of a local and confined paleo-bottom-current [7].

The seismic data available do not allow for the complete mapping of the entire drift extent. However, the data coverage suggests that the drift extends at least 30 km perpendicular to the western flank of the YP and along the slope for 30 km (Fig. 1). As an abrupt along-slope termination of a plastered sediment drift is unlikely and the typical length/width ratio of this type of drifts varies between 2:1 to 10:1 [3], we estimate that it extends further northward and southward for at least an additional 30 km.

5.2 Mass movement in eastern Fram Strait

Evidence of multiple slope instabilities along the eastern flank of the Fram Strait have been recently documented by Elger et al. [23] and Osti et al. [50] and are referred to as the Fram Strait Slide Complex and Spitsbergen Fracture Zone Slide. In this study, we analyzed in detail the geometry of failures constituting the southern part of the slide complex. Our results suggest that the style of mass movement differs with depth and depends on the geometry of the dominant sedimentation. We observe that mass movements present sharp headwalls and glide planes at depths shallower than ~2500 ms, where the contourite drift presents its typical facies characterized by broad lenticular, convex-up seismic reflections. At water depths deeper than ~2500 ms, where the margin consists of a complex mix of wavy structures, moat-related features and planes of shear (Fig. 4), bathymetric data have shown clear indication of mass movement [50,23]. Here, sharp headwalls and glide planes are not evident from seismic data. Rather, sediment deformation and deep-seated faults appear to control the irregular morphology of the seafloor along the deeper slope.

5.2.1 Shallow mass movements

Two distinct submarine slides are recognized in the seismic profiles as chaotic reflections presenting irregular upper boundary, occurring downslope of marked escarpments, which we interpret as headwalls (Figs. 5, 7a and 7b).

The slide in Line017 is part of the Fram Slide Complex and it is referred to as N0, S3 and S5 in Elger et al. [23] (Fig. 1). The slide originates at a depth of ~2500 ms TWT and extends for 4.8 km downslope. Line017 intersects the slide on a marginal part of the headwall (the transition between the headwall and the sidewall is difficult to determine when the detachment niche presents an amphitheater shape), and we therefore assume that the main slide body originates at shallower depths and that it extends for ~7.5 km perpendicular to the slope (Fig. 5b). The slide scar presents up to ~100 ms TWT thick sediment that has not been fully evacuated. The shape of this material, inferred from its appearance in the seismic profile, suggests that it consists of blocks that did not disintegrate during mobilization (Fig. 5b). We observe a transparent seismic unit 2 km downslope the slide scar which may represent the evacuated slide material according to its position and its thickness (Fig. 5b). The main glide plane is ~150 ms deep below the sea floor and it consists of a package of high amplitude reflections deposited between 1.95 and 1.5 Ma. The reflector dated by Mattingsdal et al. [45] at 0.78 Ma is truncated by the headwall. Thus, we infer that the failure event is younger than 0.78 Ma (Fig. 5).

The slide observed in Line019 originates at ~1800 ms TWT and extends for ~8 km downslope from the headwall. Within the seismically transparent zone representing the failed material we observe at least five units interpreted as blocks preserving the original structures of the pre-slide sedimentary column (Fig. 7b). Four of five units are located close to the headwall, 100-200 m from each other. The displaced slide material is overlaid by a drape of sediments dated to 0.78 Ma [45] (Fig. 7b), and the uppermost reflection cut by the headwall is likely to be ~1.2 Ma [45] (Fig. 7b). Based on this, we estimate that the slide occurred between 0.78 and 1.2 Ma. Due to their relatively old age, the displaced blocks are not visible on the bathymetric data as they are entirely covered by post-slide sediments. The glide plane is identified as the interpreted reflection above the 2.58 Ma reflection dated by Mattingsdal et al. [45] (Fig. 7b). The depth of void ratio and dry bulk density values, measured at ODP site 910 (leg 151) [51], are adjusted to the thickness of the sedimentary column at the headwall, based on the depths given by the dated reflections in Mattingsdal et al. [45] (Fig. 7b). At the depth of the glide plane, we observe a marked peak in void ratio and a decrease in dry-bulk density.

5.2.2 Deep mass movements

In the deeper portion of the slope (depth < ~2500 ms) the contourite drift (Fig. 4) presents a sediment waves deposition pattern with presence of moats and levee structures (Fig. 9). The wavy pattern forms semi-linear planes along which the succession is condensed, thus the single reflections cannot be followed as their thickness is below the resolution of the seismic data (Fig. 4). The lineaments act like preferential planes of shear, along which the slope undergoes deformation, indicated by the numerous zones of transparent seismic signal in Fig. 4. The deformation along the shear planes affects the entire sedimentary succession, generating escarpment observable on the seafloor [50].

5.3 Faulting and bottom-simulating reflection

Vertical discontinuities in seismic reflections are interpreted as faults. These structures are restricted to a sequence characterized by sub-parallel, continuous depositional layers. Beneath the faulted sequences the seismic character becomes chaotic and/or seismically transparent (i.e., reaching the limit of the seismic penetration) (Figs. 4 and 7c). The vertical discontinuity through a seismically chaotic sequence marked with a dashed-dotted line in Fig. 4b and 7c is interpreted as a fault displacing (seismic) basement blocks. Due to the location of the study area in the vicinity of the Molloy Axial Rift, we interpret this basement as being young crust formed during spreading. In addition to the basement faults identified, we also identify sedimentary faults, which when they occur in relation to basement structure are interpreted as growth faults [52] (Fig. 8). The sedimentary faults along the seismic profiles increases in number and begin to breach the seafloor more frequently with proximity to the spreading ridge (Fig. 8), indicating that deposition of sediment is syn-tectonic. The areas with apparent sedimentary fault activity are distinct, and separated by ~40 km of relatively unfaulted sedimentary strata. We suggest that this is due to the sedimentary faults forming consequently to the movement on the basement faults. This might suggest the potential presence of additional basement faults underneath the upper slope of the West Svalbard Margin that have been accommodating rift spreading in the past. In addition, the breaching at the seafloor by some of the faults indicates that some deformation is ongoing at the present stage.

We observe the presence of a BSR along all the analyzed seismic lines (Figs 1 and 8). The BSR appears patchy along the shallow contourite drift, disturbed by areas of high-amplitude extending at shallower depths. Several high amplitude reflections beneath the BSR (Fig. 8) suggest the presence of a free gas zone underneath gas hydrate-bearing sediments (e.g., [53,54]). The areas presenting anomalies in the BSR trend are spatially coincident with the shallow termination of sedimentary faults suggesting a cause-effect relation between the two features.

6. Discussion

6.1 Onset of contourite drifts

The stack of sediment composing the shallowest part of the contourite drift, downlaps on a reflection that lays between the 1.95 and the 1.5 Ma reflections (Fig 5b). As the characteristics of the oceanographic circulation in the Fram Strait did not significantly vary since the opening and deepening of the gateway (13.7-10 Ma, Fig. 10) [55], we suggest that the downslope shift of the depocenter of the plastered drift is linked to the climatic variations recorded during the last 2.6 Ma (i.e., since the onset of glaciation in the northern hemisphere) rather than directly to major changes in oceanographic settings. In agreement with Rebesco et al. [35], we infer that the climatic variations had a significant influence on the yield of biogenic and terrigenous sediments as a consequence of increased sediment supply [35] and, thus, on the potential for deeper development of the drift. The significant increase in sedimentation rate marked by the 2.58 Ma reflections that we observe in our seismic profiles supports our hypothesis. As shown in Fig. 7b, the ~60 ms thick interval between the 2.58 Ma reflection and the 5.8 Ma reflection deposited in 3.22 Ma, resulting in a sedimentation rate of ~18.6 ms/Ma. On the other hand, the ~180 ms thick interval between the 2.58 Ma and the 0.78 Ma reflection deposited in 1.8 Ma, resulting in a sedimentation rate of ~100 ms/Ma. Accordingly, the YP-2/YP-3 boundary marks the transition from pure contourite deposition to contourite deposition influenced by glacial sedimentation [45], thus, indicating an increase of sediment yield as a consequence of the intensification of the Northern Hemisphere glaciation [44,45]. Interestingly, the shift in depocenter of the plastered contourite drift (Fig. 10) in this study presents comparable depth and similar age to the onset of the West Spitsbergen drifts (early Pleistocene age) [35].

At the lower slope, an evident change in facies of drift deposition is marked by the transition to a wavy sedimentation pattern (Fig. 3) with the presence of moat-levee structures (Fig. 4, and schematized in Fig. 9). According to our observations, such pattern of stacked features appears to be restricted to the deepest and steepest part of the slope suggesting the strengthening in bottom current regime at greater depths [8,14]. The countourite drift comprises a unit underlying the 5.8 Ma reflection (Fig. 3) indicating a relative age of > 5.8 Ma, probably linked to the opening and deepening of the Fram Strait during middle and late Miocene [42]. In order to initiate the erosive vs. depositional activity of a contourite drift, not only does a bottom current need to be generated, but it also needs to be fast enough to erode bottom sediments. In the case of the Yermak Plateau system, the necessary velocity is reached by constraining the bottom currents [3]. We suggest that the activity of the detected basement normal faults (Figs. 4, 7c and 10) during and after the opening of the Fram Strait contributed to the steepening of the slope along the eastern oceanic gateway. The increased gradient of the eastern Fram Strait slope, combined with the action of the Coriolis effect on a S-N flowing bottom current (bending the currents eastwards), generated the favorable conditions for the confinement of the current and the consequent initiation and growth of the drift [56,3,13]. In addition, the intense slope failure which has been documented to affect the region since > 5 Ma might also have contributed to the steepening of the lower continental slope [50,23]. Interestingly, a long history of sliding events is also recorded at the eastern Faroe-Shetland channel. Similarly to the eastern Fram Strait deep contourite drift, slope failures at the eastern Faroe-Shetland channel are thought to have led to steepening of the lower slope and favored the onset of a contourite drift [11,57].

6.2 Development of slope instability

We propose that the resulting onset of a contourite drift, and specifically the facies characterized by sediment waves at the deepest portion of the slope, had a fundamental impact on the instability of this segment of the slope. The wavy pattern allowed for the formation of planes of shear along surfaces of condensed sedimentary succession. The combination between the steepening of the slope and processes of toe erosion, both controlled by the continuous activity of normal faults, generated the preconditions for slope failure. The area is known to be seismically active and several earthquakes presenting M > 4 have been recorded in the last century (http://www.isc.ac.uk).

We propose that the proximity to the spreading ridge and the continuous supply of seismic energy generated by earthquakes might have been the trigger for the instability of the slope. We suggest that the existence of the planes of shears led to the deformation of the slope through a process of slump/creep [58,59] rather than to failure and disintegration of failed material at the lower slope (Figs. 4 and 10). Our interpretation is based on several observations: a) rotational glide planes are missing in this portion of the slope; b) the almost absence of mobilized material at the toe of the slope [50] suggests a process of slow deformation rather than a failure; and c) the seismically transparent lenses presenting both similar geometry to the adjacent sediments and a weakly preserved internal structure suggest internal deformation rather than complete disintegration after failure and mass transport. A similar pattern of slope deformation has been previously observed along the slope of northern Spitsbergen by Geissler et al. [60]. However, we do not rule out that small size mass movements might have occurred occasionally as observed by Osti et al. [50] and Elger et al. [23].

At shallower depths, where gentle relief and smooth topography allow for a broad non-focused bottom current, the lower gradient and possibly lower velocities favored the development of a facies characterized by broadly lenticular, convex-up sediment units (Fig. 9). Here, conditions of instability have been mainly caused by four factors: a) the lack of support and consequent undercutting at the lower slope caused by the ongoing deformation of the deep slope, b) the homogeneity in grain size, typical of contourite deposits, c) the subparallel geometry of the sediment layers composing the drift and favoring the development of glide planes, and d) the presence of a potential weak layer in the sedimentary sequence [50]. For example, at the depth of the glide plane of the landslide in Line019 we observe a marked peak in void ratio and a decrease in dry-bulk density (Fig. 7). Although relevant changes in lithology have not been observed in the sediments, these values suggest the occurrence of a more porous and, subsequently less dense stack of deposits. These conditions might have favored the accumulation of fluids along this sediment stack and the buildup of overpressure, which, in turn, might have led to the formation of a weak layer. Eventually, the seismic energy generated by the frequent and relatively high magnitude earthquakes acted as the final trigger leading to the development of glide planes. As a consequence, the slope failures above 2500 ms present the characteristic of translational slides, following the classification by Lee et al. [24].

Our observations highlight the importance of the topography of the Fram Strait in relation to slope instability. The vicinity to an active tectonic region as the Molloy Axial Rift and Spitsbergen fracture zone had a crucial implication on shaping the sea-bottom morphology, favoring the onset of the sediment drift (Fig. 10). The occurrence of intense and active normal faulting contributed to the steepening of the continental slope. The generated steep slope created the favorable condition for the confinement of the established bottom currents along the slope. The combination between the geotechnical characteristics, the depositional geometry of the sediment drift and seismicity as the final trigger mechanism may have further contributed to the instability of the slope through internal deformation and/or failure.

The occurrence of a BSR in all the seismic profiles in this study, in addition to the observations by Elger et al. [23] and Geissler et al. [31], suggests the presence of a well-developed gas hydrate system. The BSR in the region is characterized by: a) local interruptions form a patchy BSR pattern and b) the shallowing of high-amplitude areas right at the location of a fault plane (Fig. 8a and 8d). The presence of these fluids might have contributed to overpressure and weakening of the sediment shear strength, and can consequently be a potential additional pre-conditional factor for failure. However, no active venting on the seafloor, nor indication of dissociation of gas hydrate in proximity to headwalls and zones affected by internal deformation have been identified in the stratigraphy. Hence, no evidence of fluid controlled triggers to slope failure are found at the study site.

7. Conclusions

We identified and described one deep-water contourite drift along the eastern Arctic Fram Strait. We analyzed the extent of the sediment drift and its internal geometry to discuss their relationships with the oceanographic settings of the area and their potential implication in the destabilization of the slope. We classified the sediment drift as a plastered, sheeted contourite drift based on the reflection characteristics, its extent and its internal geometry. The drift has an inferred age of > 5.8 Ma. Its onset is likely to be linked to the combined action of the onset of strong bottom currents following the opening of the Fram Strait and consequent steepening of the slope controlled by the activity of normal basement faults. The downslope shift of the drift depocenter during early Pleistocene age, when the deposition of the Isfjorden and Bellsund sediment drift commenced further south, suggest a similar origin, linked to the climatic variations of the last 2.6 Ma, rather than to regional changes in oceanographic settings.

We suggest that the extended slope instability observed within the study area is linked to the active nature of the contourite drift in addition to the tectonic activity in the area. We propose that the instability within the lower slope may be driven by internal deformation, facilitated by its internal geometry. The consequent lack of support to the upper slope may constitute a preconditioning factor for its instability. Evidence of slope failures within the upper slope are the formation of headwalls, sidewalls and glide planes typical of rotational/translational mass movement.

Considering the proximity of the study area to the mid-oceanic spreading ridge, we suggest that the trigger mechanism for the overall instability have been the seismic energy generated by frequent and high-magnitude earthquakes.

Figure caption

Fig. 1 Merge of our data set with regional bathymetry data from the area [61]. a) Extent of the Fram Slide Complex and the location of our seismic dataset. The bathymetric map is modified from Jakobsson et al. [61]. The Fram Slide Complex was investigated by Elger et al. [23],Elger et al. [62],Osti et al. [50]. SFZ: Spitsbergen Fracture Zone. b) Close up on the bathymetry of the Fram Slide Complex where several escarpments are observable at the sea bottom. The dashed black lines mark the escarpments interpreted as the surface indication of planes of shears within the sediment column

Fig. 2 IBCAO Bathymetric map of the mid-ocean ridge west of Spitsbergen, modified from Jakobsson et al. [61] showing the main ocean currents (in red). WSC: West Spitsbergen Current. NSC: North Spitsbergen Current. YSC: Yermak Spitsbergen Current. MR: Molloy Ridge. The yellow square indicates our study area

Fig. 3 Seismic profile of Line12. The seismic reflections show the occurrence of a plastered contourite drift. At > ~2500 ms depth the facies of deposition is characterized by sediment waves and the slope presents deformation by faults and shear planes. The dashed red line indicates the 2500 ms depth which marks the change in facies. Indications of a patchy BSR suggest that gas hydrates are potentially present in the sediments, but the absence of a zone of high amplitudes beneath the BSR suggests that no significant free gas accumulation is occurring.

Fig. 4 Close-up of seismic profile of Line12 (deep slope > ~2500 ms). a) The deep part of the contourite drift is characterized by moats and levees and presents an internal deformation indicated by the occurrence of shear planes and zones of transparent seismic signal. b) Interpretation

Fig. 5 Seismic profile of Line017. a) A sharp headwall and a zone of chaotic seismic reflections indicate a submarine slide. b) Sediment blocks that remained intact during the failure event are present within the failed material. The slide headwall cuts the reflection dated at 0.78 Ma, indicating that the failure occurred more recently than 0.78 Ma. The seismic stratigraphy is interpreted from Mattingsdal et al. [45]

Fig. 6 Seismic profile of Line018. An uncommonly continuous BSR can be observed throughout the sediments of the plastered contourite drift

Fig. 7 Seismic profile of Line019. a) Seismic data reveals the occurrence of a submarine slide at ~1800 ms of depth. b) The headwall truncates the 1.2 Ma reflection whereas a stack of layers in which we can trace the 0.78 Ma reflection drapes the failed material. This indicates that the slide occurred between 1.2 Ma and 0.78 Ma. The seismic stratigraphy is interpreted from Mattingsdal et al. [45]. Several sediment blocks that remained intact during the failure are identified within the failed material. c) Seismic data show the presence of a detachment fault interpreted to play a major role in displacing basement blocks, steepening of the lower slope and generating seismic energy for triggering the instability within the contour current deposits

Fig. 8 Overview of seismic profiles of a) Line017, b) Line018, c) Line019 and d) Line12. The shallow termination of sedimentary faults in Line017 and Line12 is spatially coincident with anomalies in the BSR trend, suggesting a cause-effect relation between these two features

Fig. 9 Schematic representation of the onset of the contourite drift on the eastern flank of the Fram Strait. The activity of the basement fault steepened the lower slope confining the contour current and favoring the onset of the sediment drift. Locally, erosion takes place along the moat and deposition occurs downslope and along the direction of the current. The internal depositional geometry of the drift favors the development of planes of shears (dashed red lines)

Fig. 10 Schematic representation of the evolution of the slope from the first opening and deepening phases of the Fram Strait to present day. Intense faulting during and following the opening of the Fram Strait contributed to the steepening of the continental slope. The steep slope created the favorable condition for the confinement of bottom currents along the slope. The interplay between the geotechnical characteristics, the depositional geometry of the contourite drift and frequent high M earthquakes led to the instability of the slope through internal deformation and/or failure

Conflict of interest

On behalf of all authors, the corresponding author states that there is no conflict of interest.

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