Km-scale polygonal sea-bed depressions in the Hatton Basin, NE Atlantic Ocean - Constraints on the origin of polygonal faulting

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Abstract

Polygonal faulting is a widespread phenomenon in sedimentary basins worldwide. It changes basin-scale fluid flow patterns and alters the physical properties of the sediments making it important for hydrocarbon exploration and geohazard analysis. It is generally accepted that polygonal fault patterns derive from dewatering and compaction of the host sediments, but there is debate regarding the processes that control polygonal faulting. New multibeam-bathymetry data from the Hatton Basin, NE Atlantic, show up to 10 m deep and 200-600 m wide troughs at the sea-bed. They connect to each other forming polygons that are several hundred meters across, i.e. of similar size as buried polygonal fault systems observed in 3D seismic data. The troughs are symmetrical and resemble elongate pockmarks. Previously unpublished high-resolution 2D seismic data from the same area show seismic disturbance zones similar to pipes observed under pockmarks elsewhere as well as faults that have all the characteristics of polygonal fault systems. The observation of the wide disturbance zones is enigmatic, as they appear to follow the polygonal seafloor pattern. The observed extent of the polygonal sediment contraction system is substantial covering almost 37,000 km^2. We calculate that some 2600 km^3 of possibly carbon-bearing fluids have been expelled from this system and we expect that this will affect the benthic ecosystems, although so far there is only limited evidence for chemosynthetic habitats.

Keywords: Polygonal faulting; silicate diagenesis; dewatering; subsurface sediment deformation; seismic data; multibeam bathymetry data

1. Introduction

Polygonal fault systems are networks of small-offset faults. They occur in layers of fine-grained sediments within sedimentary basins. The faults occur in depth intervals (tiers) that seem to be characterized by particularly small grain sizes. However, they can extend for some distances into the over and underlying strata which makes them important for the integrity of reservoirs that have polygonally faulted clays as cap rock. Within the tiers polygonal faults strike in all directions but they tend not to intersect at angles steeper than 10 degree which may be explained by the stress field during propagation (Goulty, 2008). For both reasons, i.e. layer confinement and arbitrary strike direction, they cannot be caused by regional tectonic stresses (Cartwright and Lonergan, 1996). Polygonal faults are up to several hundred meters high and their throws are largest in the middle and decrease both top- and downward (Berndt et al., 2003; Gay et al., 2004; Higgs and McClay, 1993; Stuevold et al., 2003). Typically, the faults dip at angles of 30 to 70° against the vertical and the diameter of the polygons is of the order of 1 to 2 km (Gay and Berndt, 2007) and their throw is roughly increasing with fault plane height (Shin et al., 2010). Although polygonal faults have been documented for more than 50 sedimentary offshore basins from around the world onshore outcrop analogues are scarce (Cartwright et al., 2003).

Individual faults in the Ypern Clays, Belgium have been interpreted as the onshore extension of the polygonal fault systems of the southern North Sea. They show multiple mm-wide ruptures with limited displacement (Verschuren, 1992).

The non-tectonic origin of polygonal faults has been revealed by the use of 3D seismic data in the 1990s (Cartwright, 1994). Apart from being little understood structural phenomena, polygonal faults have some wide reaching implications that merit further investigation. Work on the sedimentary basins off Norway (Berndt et al., 2003) and Angola (Gay et al., 2004; Gay et al., 2003) demonstrated that the polygonal faults are tightly linked to those basins' fluid flow systems. This is evidenced by concentric sediment distortions that rise from the tip of the polygonal faults and up to the sea-bed where they terminate in pockmarks. Although the faults are believed to be linked to pore water expulsion and layer-parallel contraction of sediments, it is not clear whether the fluids focused by the faults originate from sediment dewatering from the deeper parts of the sedimentary basins or from the polygonally faulted interval. The fact that polygonal faults are capable of focusing fluid flow implies that their properties need to be understood for assessment of reservoir leakage. As they only occur in fine-grained sediments they may also serve as a good lithology indicator.

Five hypotheses for the origin of polygonal faults have been discussed in the literature and were thoroughly reviewed in Cartwright et al. (2003), Cartwright (2011), and Goulty (2008). The first hypothesis is that the polygonal faulting is caused by gravitational forces along gently dipping basins floors (Watterson et al., 2000). The problem with this hypothesis is that polygonal faults have been observed in many basins in which they are not bounded by a dipping surface at their base. Also the fact that the faults strike in many different directions and have their greatest throw in the middle of the faulted interval is not easily explained by this hypothesis. The second hypothesis proposes the faulting to be initiated by Rayleigh Taylor instabilities due to lighter under-consolidated sediments at the base of the polygonally faulted interval. Indeed undulations of the expected wavelength are found at the top surface of a polygonal fault tier in the Yper Clays (Henriet et al., 1991) and in the Faeroe Shetland Trough (Davies et al., 1999) that extend to the surface (Long et al., 2004) and the total horizontal shortening seems to be small in some polygonal fault systems (Watterson et al., 2000). However, these are exceptions among the many observed polygonal fault systems, and it is difficult to conceive how these density inversions should actually lead to the observed faulting because it is very different from the structures in response to salt related density inversions (Goulty, 2008). The third hypothesis invokes syneresis of colloidal sediments to initiate the initial fracturing of the rocks (Cartwright and Dewhurst, 1998; Dewhurst et al., 1999). This process has been observed in fine-grain sediments, but this hypothesis was questioned, as polygonal faults occur in a wide range of lithologies and syneresis should be lithology dependent. Laboratory experiments also indicate that this process is occurring very fast (White, 1961) and it is difficult to see how it can lead to long-term deformation as recorded by growth structures along polygonal faults. The fourth hypothesis invokes faulting controlled by the residual shear strength of the faulted sediments (Goulty, 2001; Goulty, 2008; Goulty and Swarbrick, 2005). This hypothesis was questioned (Cartwright et al., 2003) because it requires initial weakness zones spaced at suitable intervals and on its own would not explain the polygonal pattern. Furthermore this hypothesis does not explain well how the faults propagate at larger scales (Cartwright, 2011). Instead Cartwright (2011) proposed that diagenetic processes in general are responsible for a decreased ratio of horizontal to vertical stress which may facilitate initial shear failure. This hypothesis is consistent with the vast extent of polygonal fault systems and their organization in tiers. It is also consistent with laboratory results for fine grained sediments (Shin et al., 2010).

The objective of this paper is to contribute to the understanding of the polygonal faulting process by constraining the boundary conditions for the proposed hypotheses. In particular we can provide further detail on the near surface structure of polygonal faults and their overburden, the lithology in which such faults can occur, and the relationship of polygonal faults and fluid expulsion structures. To this end we present newly acquired multibeam bathymetry data and previously unpublished single channel seismic data from the Hatton Basin, Northeast Atlantic **(Figure 1).**

2. Data and Methods

The data used in this study were acquired in the Hatton Basin and consist of multibeam bathymetry data recorded with a SIMRAD EM120 system, which yields 15 m lateral and 1 m vertical resolution at about 1100 m water depth encountered in the study area. We combined these data with the regional multibeam bathymetry survey collected by the Geological Survey of Ireland. The single-channel seismic data were acquired with the British Geological Survey (BGS) mini-airgun array which consists of four 40 cubic inch guns with wave shape kits that operate at a pressure of 130 bar. The data were frequency filtered, deconvolved and post stack time-migrated using water velocity. The vertical resolution is approximately 3 m at the sea-bed and the shot interval is on average 15 m. These data were merged in a KingdomSuite project with multi-channel seismic data from the southern Hatton Basin provided by Irish Petroleum Infrastructure Programme (PIP). The still imagery and video footage was recorded using a SEATRONICS DTS3000 deep water camera system, which incorporates separate still and video cameras and a Valeport CTD. The data shown in this paper were collected in a small area of the northern Hatton Basin **(Figure 1b)**. This area was chosen because it is surveyed with the BGS high-resolution seismic system and ground-truthing by video observations is available.

3. Observations

3.1. Bathymetry

The multibeam bathymetric transects across the northern part of the Hatton Basin show elongate depressions in the sea-bed that define approximately one hundred polygons **(Figure 1b)**. Thousands of polygons exist across the basin as a whole covering an area of approximately 37,000 km^2 (Figure 1a). The depressions are up to 20 m deep, up to 400 m wide, and between 400 and 2000 m long (average 1500 m). These depressions define the polygons which range from 500 to 5000 m in diameter. A typical aspect ratio between the width of the sea-bed depressions and their length is 1:3. The slopes of the depressions are gentle and the overall shape is concave down. The faults strike in all directions with a maximum to 080 **(Figure 1b)**. The angles at which the faults intersect are generally larger than 40 degree. The sea-bed inside the polygons is flat and shows the same trend as the gentle regional topographic variations. The polygons are only found in the central part of the Hatton Basin and they gradually become less connected towards the east, south and west. Although there is minimal multibeam bathymetry data in the north, the seismic data indicate that here also the transition to un-deformed sea-bed at the margins of the basin is gradual. The bathymetric data do not show evidence of sea-bed erosion in the centre of the basin, but moats along the basin margins may be due to non-deposition/erosion.

3.2. Seismic data

The seismic data image the infill of the Hatton Basin. On top of the volcanic basement (Laughton et al., 1972) there is a succession of up to 800 ms TWT-thick Eocene sediments. Due to limited penetration of the BGS high-resolution data this unit is only imaged in the PIP data in the southern part of the basin and at the rising flank of the Rockall Bank in the BGS data in the north. Figure 2 shows a representative section of the seismic line HA04-9005 from the southern part of the Hatton Basin. At depth it shows the Eocene sediments draping onto the Rockall Bank and thicken towards the centre of the basin. On top of this unit there is an up to 700 ms TWT-thick succession of post Eocene to present sediments that is clearly influenced by bottom currents at its southeastern end where it pinches out towards a moat against the Rockall Basin. There are no signs of erosional unconformities within this unit, which is supported by drilling at Site 982 **(Figure 3)**.

There are a large number of vertical disturbances in the Oligocene to Recent succession of the Hatton Basin **(Figure 4)**. These disturbances can be divided into two classes. Class 1 consists of zones of down-bending reflectors underneath the sea-bed depressions. They extend from the sea-bed at 1700 ms TWT or 1200 m beneath sea level down to the bottom of the recorded data at 2600 ms TWT in the north corresponding to approximately 2000 m beneath sea level. Further south, the PIP data show that these disturbance zones extend at least down to the top Eocene reflector. Also the seismic facies of the Eocene succession just underneath the disturbance zones is more chaotic than away from them, but it is not clear if this is real deformation or the result of imperfect seismic imaging. The zones are between 200 and 400 m wide and their spacing is between 200 m and 2000 m. Generally seismic amplitudes in these zones are reduced compared to the surrounding sedimentary reflections. Some of these disturbances are asymmetrical with one side of the disturbance being characterised by a gradual increase in reflector dip towards the centre of the disturbance and a sharp offset of the reflectors on the other side. The number of sharp offsets increases with depth. This is the result of some disturbance zones being more focused at depth changing from gradually increasing dips to discrete faults. Vertical spacing between seismic reflectors is greater in the hanging walls indicating that these sediment disturbances are growth structures. Where the boundaries of the sediment disturbance zones are sharp, i.e. fault like, the throw increases with depth similar to polygonal fault systems elsewhere (Berndt et al., 2003; Higgs and McClay, 1993; Stuevold et al., 2003).

The BGS high-resolution data lend themselves well for the study of the small-scale nature of the sediment deformation structures. **Figure 4b** shows the detail of a class 1 disturbance with approximately 2 x vertical exaggeration, i.e. assuming 2 km/s P-wave velocity. On a width of approximately 400 m within the depth interval from 1900 to 2500 ms TWT the reflections are interrupted. At the edges these disruptions are frequently sharp and fault-like. Vertically they extend for up to 100 ms TWT. The distance between offsets in the reflector packages is generally less than 80 m and possibly less considering that the seismic line may cut them obliquely. We would like to note, however, that the bathymetry **(Figure 1b)** shows, that seismic line BGS2000-1-44 intersects the shown disturbance structure D at a steep angle and that this effect would therefore be small. The horizontal extent of the disturbance structure coincides with a vertical change of seismic facies that is continuous along the entire line. At its base it coincides with the Top Eocene reflector (Reflector 4 of Laughton et al., 1972) whereas at the top at approximately 1870 ms TWT it changes character where the seismic amplitudes change from being higher above to being lower below. Above 1870 ms the wide disturbance zone is replaced by two normal faults that form a graben above the disturbance zone.

The second class of seismic disturbance zones (class 2) is characterised by narrow 25-75 m wide almost vertical zones of decreased seismic amplitudes. They are up to 500 ms TWT or 400 m high, i.e. 4-5 times higher than the faults observed in class 1 structures, and frequently occur in close vicinity to each other constituting groups of two or three disturbances in the 2D seismic transects. These disturbance zones have vertical displacements, i.e. throws, that increase with depth towards the centre of the faults and decrease further down towards the lower tip of the faults similar to polygonal faults elsewhere (Berndt et al., 2003; Cartwright and Dewhurst, 1998; Gay et al., 2004; Lonergan et al., 1998). The dip of these faults ranges from 30 to 60 degrees against the vertical. However, as we only have 2D seismic available these are apparent dips and may be steeper in instances where the faults are cut obliquely. These disturbance zones are more abundant in the deeper part of the section of the faulted interval. They do not reach the sea-bed anywhere on the seismic profiles crossing these structures.

In summary it is the 200-400 m wide zones of chaotic seismic facies below 1870 ms TWT and the much shorter vertical extent of faults in class 1 disturbances that distinguishes this class from class 2. As the class 1 disturbances are often bounded by sharp faults on either side in their top part, i.e. above 1870 ms TWT, this difference cannot be the result of imaging one of the class 2 structures along strike.

Apart from the seismic disturbances of class 1 and the greater abundance of the narrow seismic disturbances at depth the Oligocene to Pleistocene succession is uniformly stratified. In particular, the seismic data do not show a polygonal fault system underlying the wider (class 1) fluid expulsion structures, which is different from offshore Angola and mid-Norway (Berndt et al., 2003; Gay et al., 2004).

The southeastern end of seismic profiles HA04-9005 **(Figure 2)** and BGS2000-1-1 **(Figure 5)** shows erosional truncation of the uppermost sedimentary reflections against the sea-bed indicating submarine erosion. These top laps are limited to the vicinity of the sea-bed moat that bounds the Hatton Basin to the east against Rockall Bank. Submarine erosion was also reported for the western rim of the Hatton Basin (Laughton et al., 1972), but this cannot be seen in our data. There is no seismic or other evidence for erosion in the central parts of the basin.

3.3. Video transect

During a sea-bed survey in the summer of 2006 we collected a video transect across one of the polygonal sea-bed depressions. There were no signs of fluid expulsion such as vents or crusts of authigenic carbonates along this transect, and there were no indications for abnormal sea-bed fauna such as pogonophera tube worms or cold water coral reefs. The sea-bed shows, however, a large number of light patches which may or may not be bacterial mats. This was corroborated by a recently conducted ROV survey in 2011 (R. James, person. comm.).

4. Discussion

4.1. Sea-bed polygons in the Hatton Basin and polygonal fault systems

The polygonal sediment disturbance structures **(Figure 1)** developed in the post-Eocene sediments of the Hatton Basin (Laughton et al., 1972). The depth of the basin is not well known as basalts covered it during the Paleocene-Eocene and sediments may underlie the volcanic succession. Wide angle seismic data indicate that it is at least 2 and possibly 8 km deep (Morgan et al., 1989; Smith et al., 2005). In the study area the post-volcanic sediments are approximately 1.5 km thick (Hitchen, 2004; Laughton et al., 1972) and fill the trough between the Hatton Bank and the Rockall Bank **(Figure 1)**. The basin formed perhaps during the mid-Cretaceous (Smythe, 1989) as part of the rift history that led to continental break-up between the Rockall Plateau and Greenland in the Early Eocene (Cole and Peachey, 1999).

The sediment deformation does not entirely consist of discrete faults, but shows a continuum from laterally extensive inflexions of the seismic reflectors at shallow depth to discrete faults deeper in the sediment pile. The vertical extent of the faults is quite variable. While some extend from the top-Eocene reflector almost to the surface others appear as part of a network of fractures **(Figure 4b)**. In other respects, i.e. the length of the polygon sides, the variation in strike directions, and variation in throw, they are similar to other polygonal fault systems (**Figure 2**, (Gay and Berndt, 2007; Lonergan et al., 1998). The fact that they almost reach the surface and are overlain by tip folds that ultimately form the sea-bed depressions makes the system in the Hatton Basin similar to the polygonal fault systems on the Gjallar Ridge (Clausen et al., 1999) and offshore Angola (Cartwright and Dewhurst, 1998; Gay et al., 2004).

The faults occur in Oligocene to recent sediments that were sampled at DSDP Site 116 and 117 (Laughton et al., 1972) and ODP Site 982 (Shipboard Scientific Party, 1996). At Site 116 and 982 the sediments consist of approximately 700 m of biogeneous oozes with very high calcareous carbonate content (~80%). Only the glacially influenced upper 70 m of sediments have significant amounts of detritus. Beneath 70 m the sediments are increasingly more lithified from watery oozes at the top to limestones at 700 m depth. Also the silica is transforming to chert from approximately 550 m depth. However, the density is only reaching 2.05 g/cm^3 at the bottom of Site 116, i.e. at 854 m depth below the seafloor, and seismic velocities measured with the core logger do not exceed 1.7 km/s even at the base of the borehole, both indicating that dewatering due to silica diagenesis was active but not as pronounced as elsewhere in the North Atlantic (Berndt et al., 2004; Davies et al., 2008) where silica diagenesis leads to the development of bottom simulating reflectors which is not observed in the Hatton Basin. The post Eocene sedimentary succession was deposited without a recognised hiatus and with sedimentation rates of fairly constant 3 cm / 1000 years coinciding with increasing paleo-water depths. The paleontological data indicate neritic sedimentation for the Early Eocene at Site 117 and after a late Eocene hiatus a gradual increase of water depth until the present water depth of approximately 1200 m. Overall, the continuous pelagic sedimentation in the Hatton Basin has caused particularly high water contents which may be the reason why the polygonal fluid escape patterns are so well developed.

The type of available seismic data, i.e. limited bandwidth and short streamer length, does not lend itself to an extensive analysis of pore fill. However, there are some observations that suggest that the class 1 deformations are evidence for past or present fluid migration. First and foremost, it is the disturbance of the primary seismic reflections which is typically observed underneath seep sites (Berndt et al., 2003; Hovland and Judd, 1988). Secondly, there is a general decrease of amplitude within the chaotic zones, which may be the result of pore water expulsion from more water rich layers and a resulting decrease in acoustic impedance contrasts. We interpret the sediment disturbance structures in the Hatton Basin as a polygonal fault system although the occurrence of numerous fluid escape structures of class 1 makes it somewhat unusual. While the polygonal arrangement of seafloor depressions may be explained by the polygonal faults at depth and their accompanying tip folds, it is more difficult to explain a polygonal arrangement of the class 1 deformation structures at depth. In the 2D seismic data they appear as groups of fractures **(Figure 4b).** But it is not clear how they should develop into polygons if they do not propagate as faults due to the stress focusing at their lateral tips (Goulty, 2008). They are not underlain by a mature polygonal fault system (Figure 2), which may lead to a polygonal shape of fluid escape.

An explanation may be found in the observations related to the class 2 anomalies. These are in fact only solitary, i.e. not elongate or joined-up, features such as pipes underlying pockmarks elsewhere. As we do not have 3D seismic control in this area we cannot be sure that they link up in polygons. In this case they may be the result of hydro-fracturing during dewatering of the basin. They may therefore serve as zones of weakness from which polygonal faults nucleate due to their reduced residual strength of the sediments (Goulty, 2008). The fact that the seismic amplitudes decrease lateral consistently at about 1900 ms TWT could be explained by a diagenetic change of silica from opal A to opal CT (Berndt et al., 2004) and it is tempting to attribute the change of style in class 1 disturbance zones to the increased dewatering connected to this diagenetic transformation. However, the changes of silica concentration and type observed at Site 116 do not show abrupt variations (Laughton et al., 1972), and the seismic data do not show a clear crosscutting of this amplitude anomaly across the primary sedimentary reflections, which may of course be explained by the horizontal stratification. Thus, the silica control cannot be corroborated with the available data. We also do not find clear evidence for a transition from class 1 to class 2 which would be expected at the nucleation points, but this may well be due to the limited amount of seismic data. It would take high-resolution 3D seismic data to observe a class 1 structure starting at a class 2 structure.

4.2 Timing – The Hatton Basin a site of present-day polygonal faulting

The polygonal structures of the Hatton Basin reach almost up to the sea-bed and neither the DSDP/ODP drilling results nor the seismic data show evidence for erosion at the present sea-bed. This means that the polygonal pattern develops at shallow burial depth, although proper faulting is not observed until some 30-50 m beneath the sea-bed. In this sense the polygonal sediment disturbances are similar to the structures observed on the Gjallar Ridge on the Norwegian Margin (Clausen et al., 1999) and offshore Angola (Cartwright and Dewhurst, 1998; Gay et al., 2004; Gay et al., 2003). Polygonal deformation affects the sediments above C30 of Hitchen (2004). This means polygonal faulting in the Hatton Basin could be a continuously ongoing process since the Miocene. This is similar to the Norwegian Margin for which the distribution of dewatering pipes that are related to polygonal faulting indicate protracted activity of the polygonal fault system over several million years (Berndt et al., 2003; Gay and Berndt, 2007).

The absence of discrete faults in the upper strata coincides with the change in lithology, i.e. the increase in detritus in the uppermost 70 m caused by the glacial influence. It is not clear if this change in character of the polygonal deformation is a sign for shut-down of the polygonal faulting caused by the change in lithology or whether the focusing of the polygonal deformation would propagate into the present sea-bed sediments with continued burial. The latter seems more likely as the upper termination of the faults is variable and not confined to this depth only.

4.3 Nucleation – polygonal fault changes with depth

The observations from the Hatton Margin provide further constraints on the formation of polygonal sediment dewatering. The sediment densities encountered at DSDP Site 116 show that there is no inversion at present which rules out Rayleigh-Taylor instabilities in recent times (Davies et al., 1999; Victor and Moretti, 2006), i.e. the first hypothesis discussed by Cartwright et al. (2003). However, if past density inversion was related to undercompaction it may have disappeared during pressure release and fluid expulsion, and it may be difficult to find evidence for it now. Furthermore, the polygonal fault pattern is symmetrical **(Figure 1)**, and the seismic data show that the polygonal sediment deformation occurs in a confined basin withcout a regionally dipping base. This makes gravitational forces (Watterson et al., 2000) an unlikely agent for the development of the polygonal pattern, at least in this area.

Of the four hypotheses proposed by Cartwright et al. (2003) this leaves syneresis and fracturing as a result of low residual shear strength (Goulty, 2001; Goulty and Swarbrick, 2005). Furthermore, diagenetic processes may reduce the ratio of horizontal to vertical effective stress (*ko*) necessary to initiate shear failure (Cartwright, 2011; Shin et al., 2008). The new data show that the dewatering fluids disturb the sediments in a polygonal pattern and it is likely that the disruption caused by pore water movement decreases the shear strength of the sediments. It is therefore of fundamental importance to understand whether polygonal faults develop first (and focussing of fluid flow by the polygonal faults results in the fluid escape structures above), or if fluid expulsion comes first and is already organised in a polygonal geometry when the polygonal faults develop. This may be supported by the observations that (1) the fluid escape seems to be organised in polygons without polygonal faults underneath each of the fluid escape features, (2) the fluid escape features are considerably bigger than the polygonal faults, and (3) most of the polygonal faults do not reach the sea-bed and the sediment deformation is more confined downward, which perhaps indicates that it takes time for the polygonal faults to develop, and that weakness zones are forming as a result of fluid flow focusing.

Dewatering may provide weakness zones that are required by the residual shear strength hypothesis. On the other hand, dewatering will at least partly be related to diagenetic changes. The results of Shin et al., (2010) show that this in itself may generate initial shear failures that develop into polygonal faults. As such the proposed residual shear strength and diagenetic weakening hypotheses are partly complementary as faulting may start at dewatering structures and propagate laterally and upward due to reduced *ko* that is caused by diagenetic processes.

With the limited data at hand it seems most likely that the fluid expulsion structures develop first, followed by polygonal faulting within these weakness zones. The sediment contraction caused by dewatering finally induces further faults within the polygons. These faults only develop at depths at which protracted sediment contraction has generated the necessary reduction of horizontal stress. Overall this process seems to be rather slow and continuous instead of vigorous and episodic, because there are no reflectors in the seismic data that bend upwards toward the fluid pathways which may be expected for fast sediment deforming eruptions. This is supported by the absence of distinct fluid seeps in the video data or pockmarks in the multibeam bathymetry data.

Our observations lend support to an important role of diagenesis in sediment deformation. The observed variations in silica composition at Site 116 show a general decrease in opal A concentration down-hole. Applied to the experimental results of Shin et al., (2010) this would mean that the entire basin is subject to decreased *ko* facilitating initial shear failure. Possibly in some places, i.e. the class 1 deformations, the fluid expulsion from diagenetic processes is so vigorous that focused fluid flow systems form.

4.4 Dewatering of the Hatton Basin and implications for seabed ecology

The new data clearly show that the small offset faults and associated sea-bed depressions are not an analogue to the Feni Drift sediments as proposed previously (Laughton et al., 1972). The sea-bed polygons observed in the multibeam bathymetry data clearly disprove the previous interpretation of the sea-bed depressions seen in 2D seismic data as NE-SW trending sea-bed furrows caused by bottom currents.

The seismic and multibeam data indicate that the sea-bed polygons occur over some 37,000 km^2 in the central part of the Hatton Basin. Using an average thickness of 700 m of sediments that are affected by the polygonal deformation and a porosity loss from 80 to 60 % (based on results from DSDP Hole 116) within this interval (Laughton et al., 1972), we calculate that approximately 2600 km^3 of fluids could have been expelled from this system. If the structures reported by Vanneste et al. (1995) are part of the same sediment body these numbers may still be significantly bigger. So far, it is unknown if this volume is expelled continuously or episodically, but the fact that the deformation zones reach the sea-bed to form polygons shows that the fluid expulsion has been active until the recent geological time, i.e. during deposition of the present surface sediments.

During a sea-bed survey in the summer of 2006 we collected a video transect across one of the polygonal sea-bed depressions. This did not reveal conclusive evidence for active fluid expulsion such as vents or indicative chemosynthetic benthic ecosystems such as tube worms. The video images do show a large number of pale patches at the sea-bed which may be bacterial mats, and decimetre-scale relief which is uncommon in distal, deep-water areas such as the Hatton Basin **(Figure 6)**. This relief may indicate crusts of authigenic carbonates along the video transect. This may indicate episodic expulsion of fluids as continuous dewatering would yield negligible fluxes and would unlikely result in clearly observable carbonate crusts. It is possible that future investigation of this vast area will result in the discovery of benthic ecosystems that have adapted to this special habitat. In addition to the shelter that is provided by the hummocky sea-bed, it is possible that the polygonal dewatering structures sustain chemosynthetic ecosystems such as those recently found in the vicinity of other cold seep sites (Sibuet and Olu-Le Roy, 2003).

4.4. Implications from other types of patterned ground

Joint-bounded polygonal columns develop in a wide variety of materials ranging from millimetres to hundreds of meters in diameter. Contraction of cooling, solidified magma yields columns that are much taller than broad. This process is called columnar jointing and occurs in almost any kind of solidified lava (DeGraff and Aydin, 1987). Polygonal patterns called desiccation cracks also form when mud (Weinberger, 2001) or starch (Müller, 1998) dry out. In these cases the columns are usually as wide as they are high. Furthermore, polygonally patterned ground develops in permafrost environments, where it is related to complex cycles of freezing, melting and development of secondary ice lenses (Lachenbruch, 1962; Marchant et al., 2002). The new data extend this list of polygonal surface patterns to submarine surface sediment dewatering. The polygons found in the Hatton Basin constitute an end-member in terms of polygon size. The only reported somewhat similar systems are the sediment structures in Lake Superior (Cartwright et al., 2004). However, these structures are not polygonal-shaped but doughnut-shaped and they are not linked to polygonal faults at depth.

The polygonal sediment deformation in the Hatton Basin and polygonal fault system in general are characterised by a higher density of faults at depth than at the surface. This is opposite to polygonal joints that develop in basalts (Saliba and Jagla, 2003) or starch (Goehring and Morris, 2005). Saliba and Jagla (2003) calculate how the stress pattern varies with depth, and that joining of the discontinuity leads to a focusing of displacement with depth and duration of cooling. There are two fundamental differences between columnar jointing and the polygonal sediment deformation in the Hatton Basin. Whereas desiccation cracks and columnar jointing are governed by dispersive laws and starts at the surface and migrates down, polygonal faults nucleate at depth and migrate up and their genesis is probably linked to convective laws of fluid migration. Columnar jointing in basalts also starts at once when lava solidifies, whereas the polygonal sediment deformation in the Hatton Basin develops during ongoing sedimentation and for several millions of years. The fact that the fault density in the Hatton Basin is greater at depth than it is at the surface may therefore imply that the structures at the surface are more mature in the sense that the stress due to contraction and water expulsion has focused. It would require high-resolution 3D seismic data to determine the geometry of the fault terminations at depth and to quantify the stress regime. This geometry information is necessary for finite element modelling of the stress field.

Müller (1998) conducted a quantitative comparison between the column diameter in columnar joints in starch and basalt and concluded that in a first approximation the column diameter depends on the depth gradient of the polygon forming physical property, i.e. the temperature gradient for cooling basalt and the water content for drying starch. Columnar jointing in basalt has a much greater diameter and the temperature gradient is roughly three orders of magnitude lower than the starch gradients agreeing qualitatively with a two orders of magnitude greater diameter for the basalt columns. Following this argument the large diameter of the polygon size of the dewatering structures in the Hatton Basin would suggest even lower gradients in water content. This is intuitively the case in a slowly compacting sedimentary basin in which the water content decreases from 62-68 % volume in the surface sediments to 55 % volume at 700 m depth (Laughton et al., 1972). However, the water content is very variable. Even at 700 m depth there are still sections in which the water content is in excess of 80% indicating the importance of focused fluid migration for these sediments.

5. Conclusions

Polygonal fault development is closely linked to the alignment of fluid escape features in a polygonal pattern. The large-scale pattern seems to be governed by a stress-induced alignment of fluid escape pathways. These in turn may provide the weakness zones required for residual shear strength controlled initial failure. It is crucial that 3D seismic data are collected in the Hatton Basin to corroborate the polygonal layout of the fluid escape pathways, which so far is only deduced from the alignment of the polygonal seafloor patterns with the class 1 disturbance zones in the 2D seismic data. We also suggest that geotechnical experiments be conducted on samples from DSDP Site 116 or the close-by ODP Site 982 to see if their lithology is conducive to syneresis or if there is a correlation between the amount of diagenetically induced horizontal contraction and the depth intervals at which polygonal faulting is best developed.

The hypothesis that the gradient of the property that governs stress build-up, i.e. the reduction in water content, controls the size of the polygons may be valid over very different scales. The polygonal faults in the Hatton Basin extend the scale that was established for millimetre to decimetre-sized polygon patterns to the kilometre size. In this sense, even the development of polygonal faults in a marine environment can be considered as drying of a surface layer. Although of course, the faulting nucleates and propagates at depth and up to the surface. Continuum mechanics have successfully been applied to the modelling of the polygonal patterns within columnar jointed basalts (Saliba and Jagla, 2003). Similar models should be applied to the polygonal fault system in the Hatton Basin in order to predict the length of time that it takes to develop the polygonal patterns, but this would require three-dimensional imaging of the polygonal system at depth.

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Figures

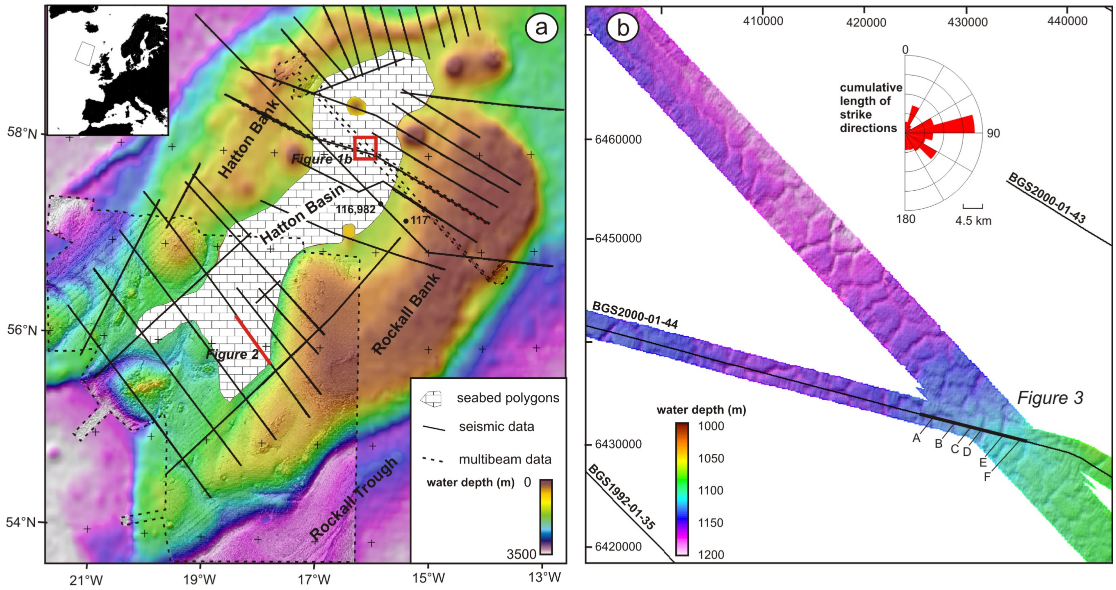

Figure 1: The polygonal sediment deformation structures are observed in the northern part of the Hatton Basin. 1b) Multibeam bathymetry data showing polygonal sea-bed depressions and the strike directions of the sea-bed depressions.

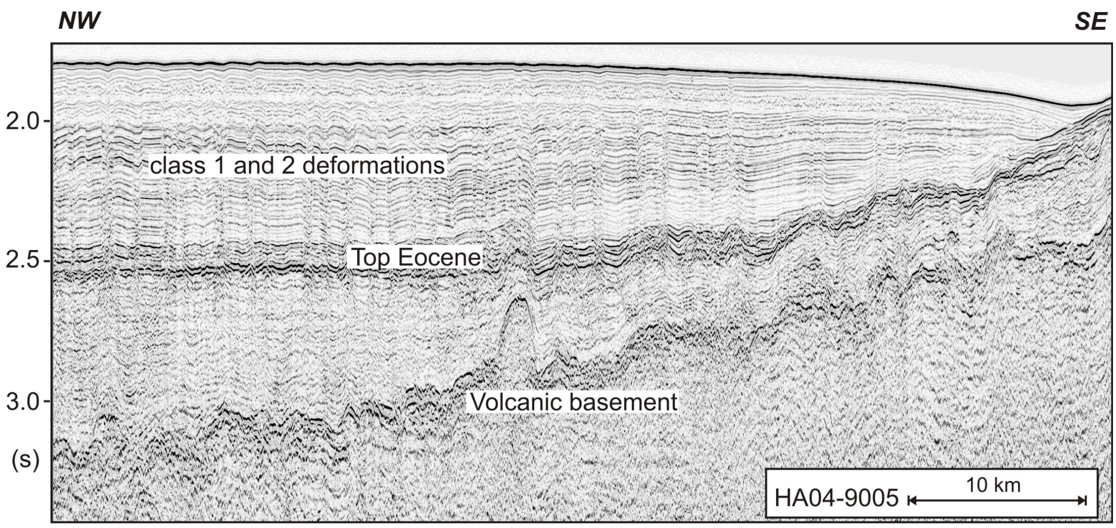

Figure 2: Regional profile (see Fig. 1 for location) showing the depth at which the polygonal deformations terminate at the Top Eocene reflector (reflector 4 of Laughton et al., 1972).

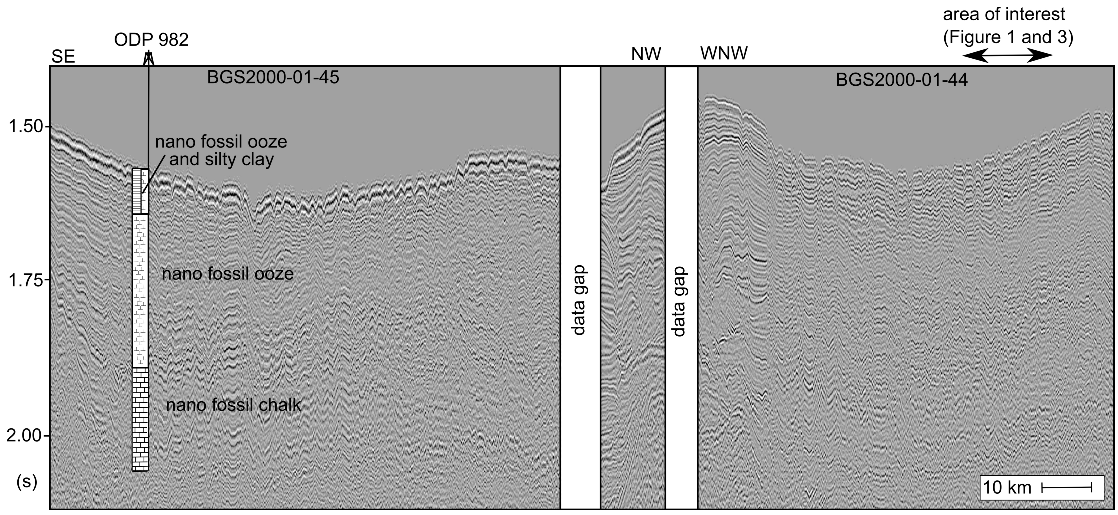

Figure 3: Correlation of the ODP Site 982 lithology to the area with multi-beam bathymetry coverage further north (Figure 1 for location). For the depth conversion of the borehole depth we used seismic velocities of 1600 m/s and 2000 m/s for the top and lower part of the hole.

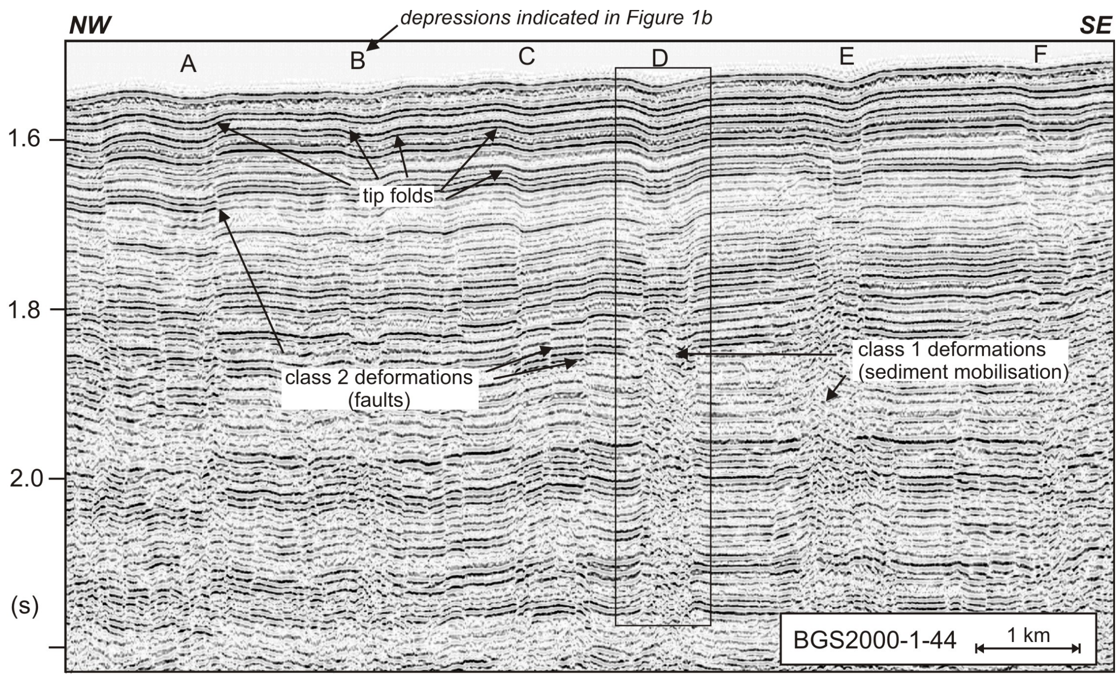

Figure 4: Single-channel seismic line intersecting the multibeam bathymetry transect. The arrows B and C at the top indicate the location of sea-bed depression annotated in Figure 1b. Note, different types of sediment deformation and vertical variation in deformation style. 3b) Seismic example with approximately 2 x vertical exaggeration showing the nature of the class 1 deformations and the typical 30-50 degree dip of the polygonal faults.

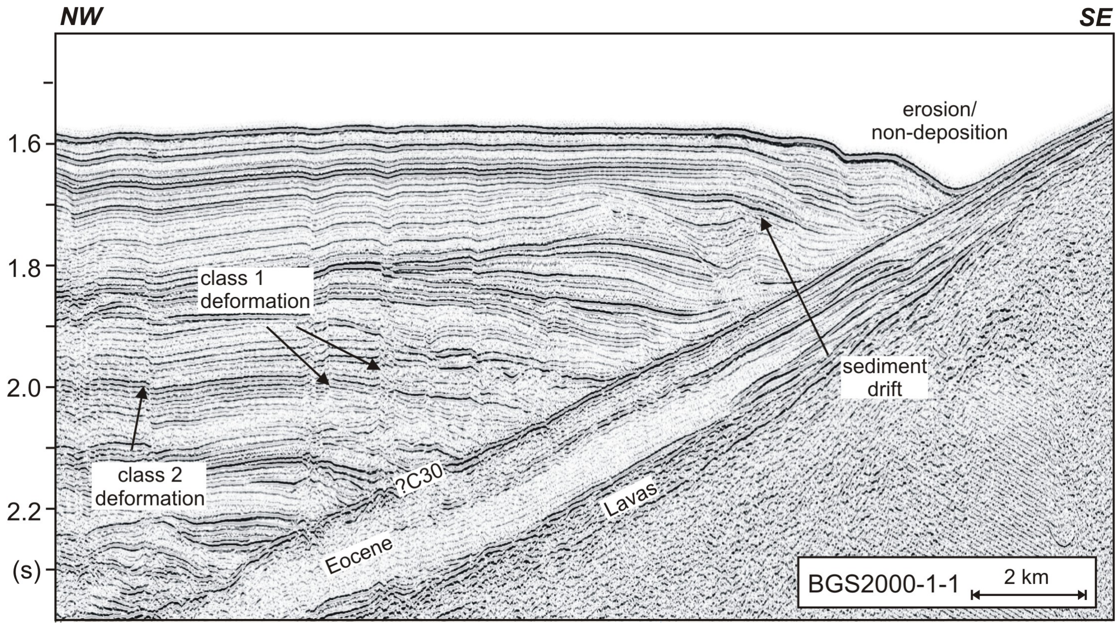

Figure 5: Single-channel seismic line from the northeastern parts of the Hatton Basin. The Late Eocene and younger sediments overlie the volcanic successions of the Rockall High. The sea-bed is scoured by bottom currents leading to erosion or non-deposition at the flank of the Rockall High. C30 as defined by Hitchen (2004).

Figure 6: Video still showing the small-scale topography and pale patches within one of the polygonal sea-bed depressions. These may result as bacterial mats from fluid escape. For scale: the fish is approximately 20 cm long.