Repeated degradation and progradation of a submarine slope over geological timescales

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ABSTRACT: Submarine slopes prograde via accretion of sediment to clinoform foresets, and degrade in response to channel or canyon incision, or mass-wasting processes. The timescales over which progradation and degradation occur, and the large-scale stratigraphic record of these processes, remain unclear due poor age constraints in subsurface-based studies, and areally limited exposures of exhumed systems. We here integrate 3D seismic reflection and borehole data to study the geometry and origin of ancient slope canyons developed within Late Mesozoic strata of the Måløy Slope, offshore Norway. Slope degradation and canyon incision commenced during the late Kimmeridgian, coincident with the latter stages of rifting. Later periods of canyon formation occurred during the Aptian-to-Albian and Albian-to-Cenomanian, during early post-rift subsidence. The canyons are straight, up to 700 m deep and 10 km wide on the upper slope, and die-out downdip onto the lower slope. The canyons trend broadly perpendicular to and crosscut the majority of the rift-related normal faults, although syn-filling fault growth locally helped to preserve thicker canyon-fill successions. The headwalls of the oldest (late Kimmeridgian) canyons are located at a fault-controlled shelf edge, where younger canyons overstep this fault, which was inactive when they formed, extending across the paleo-shelf. Downslope, Aptian-to-Albian canyons either erode into the older, late Kimmeridigian-to-Barremian canyon-fills, forming a complicated set of unconformities, or in the case of the Albian-to-Cenomanian canyons, die-out into correlative conformities. Boreholes indicate that the canyon bases are defined by sharp, erosional surfaces, across which we observe an abrupt upward shift

from shallow- to deep-marine facies (i.e. late Kimmeridgian canyons), or deep marine to deep marine facies (Aptian-to-Albian and Albian-to-Cenomanian canyons). Missing biostratigraphic zones indicate the canyons record relatively protracted periods (c. 2-17 Myr) of structurally enhanced slope degradation and sediment bypass, separated by >10 Myr periods of deposition and slope accretion. The trigger for slope degradation is unclear, but it likely reflects basinward tilting of this tectonically active margin, enhanced by incision of the slope by erosive sediment gravity-flows. The results of our study have implications for the timescales over which large-scale slope progradation and degradation may occur on other tectonically active slopes, and the complex geophysical and geological record of these processes. We also demonstrate that canyon formation resulted in an abrupt change in syn-rift facies distributions not predicted by existing marine rift-basin tectono-stratigraphic models.

INTRODUCTION

Submarine slope growth is driven by periods of sediment progradation and aggradation (e.g. Rich, 1951; Bates, 1953; Asquith, 1970; Pirmez et al., 1998; Steckler et al. 1999; Adams and Schlager, 2000; Steel and Olsen, 2002; Patruno et al., 2015; Patruno & Helland-Hansen, 2018). Slope progradation and aggradation may alternate with periods of erosion or 'degradation', during which time erosional conduits, such as channel-levee systems, may bypass large volumes of sediment to the lower slope and basinfloor (e.g. Mayall et al., 2006; Neal & Abreu, 2009; Kane et al., 2009; Romans et al., 2009; Sylvester et al., 2012; Figueiredo et al., 2013; Hodgson et al., 2011, 2016; Dalla Valle et al., 2013; Janocko et al., 2013; Hubbard et al., 2014). Constraining the location, timing and duration of these degradational periods is important, as they may allow us to infer the driving mechanisms (e.g. tectonics, eustacy), and predict when, where, and how much sediment is transferred downdip (e.g. Johannessen & Steel, 2005; Di Celma, 2011; Hodgson et al., 2011; Gong et al., 2015). More generally, establishing whether canyons form in the submarine or subaerial realm is important in terms of assessing basin morphology and paleogeography, and the potential timing and magnitude of tectonic events, and/or changes in eustatic sea-level (e.g. Shepherd, 1981; Posamentier & Vail, 1988; Pratson & Coakley, 1996; Fulthorpe et al., 2000; Bertoni & Cartwright, 2005; Zecchin et al., 2011; Maier et al., 2018). For example, do canyons encased in largely marine strata simply represent subaerially formed 'incised valleys' (*sensu stricto*; Van Wagoner et al., 1998) generated during a period of sea-level fall and lowstand? Or can canyons form at any point in the relative sea-level cycle in a fully submarine setting in response to some kind of tectonic or sediment supply forcing?

Outcrop-based studies permit detailed analysis of the sedimentological and stratigraphic expression of only one or a few cycles of slope aggradation and degradation; however, due to limited exposure the longer-term, larger-scale, three-dimensional geometry of large (i.e. kilometre-scale) slope canyons, is poorly constrained (e.g. Wonham et al., 2000; Bertoni & Cartwright, 2005; Giddings et al., 2010; Hodgson et al., 2011, 2016; Di Celma et al., 2013, 2014). In contrast, bathymetric maps of the present seabed and near-seabed geophysical studies permit detailed assessment of the geometry and likely formative mechanisms of degradation-related slope conduits, but not their longer-term (10^3-10^4 Myr) stratigraphic development or the processes that controls their ultimate preservation in the rock record. To better constrain the morphology and long-term stratigraphic evolution of submarine canyons, and thus their importance as 'tape records' of allogenic controls (e.g. tectonics, sea-level variations), we require data that permit detailed mapping of age-constrained canyons over large areas.

We here use 3D seismic reflection and borehole data from the Måløy Slope, offshore western Norway to constrain the geometry, distribution, and stratigraphic evolution of late Mesozoic (Late Jurassic-to-Late Cretaceous) slope canyons through three long-term (i.e. 10^3-10^4 Myr), large-scale (i.e. kilometre-scale) cycles of slope degradation and aggradation (Figs 1 and 2). This is an ideal location to conduct this study, with abundant 3D seismic reflection and borehole data allowing us to map the major structural elements and large-scale stratigraphic patterns, and to thus reconstruct the overall tectono-stratigraphic development of part of this rifted margin. We place our study within a regional, North Sea-wide biostratigraphically constrained, chronostratigraphic framework to investigate the potential regional and local controls on slope canyon formation and evolution. The results of our study have implications for the timescales over which slope aggradation and degradation occur, and the complex geophysical and geological (i.e. stratigraphic) expression of related features in the rock record. Furthermore, our results impact our understanding of rifted margin development, indicating that canyon formation during the syn-rift-post-rift transition can drive major changes in the pattern and style of sediment dispersal, resulting in deep-water facies distributions not captured by existing marine rift-basin tectono-stratigraphic models.

GEOLOGICAL SETTING OF THE MÅLØY SLOPE

Structural framework

The Måløy Slope is up to 40 km wide, and is bound to the east by a series of broadly N-trending, W-dipping normal faults that have >1 km of displacement, and which collectively form the Øygarden Fault Complex (Fig. 1). The Måløy Slope is bound on its western margin by a W-dipping normal fault that defines the eastern margin of the Sogn Graben (Figs 3 and 4). A 10-15 km wide graben, herein called the Gjøa Graben, is developed in the middle of the Måløy Slope. The western margin of the Gjøa Graben is delineated by a relatively large (500 ms TWT or 714-973 m of throw), E-dipping, strongly segmented normal fault, herein called the Gjøa Fault (Figs 1 and 3). The eastern margin of the Gjøa Graben is defined by a series of W-dipping, N-trending, moderately large (15 km long, up to 400 m throw) normal faults that together form part of the Måløy Fault System (Figs 3 and 4). Internally, the Gjøa Graben is dissected by numerous N-S-to-NNW-SSE-striking, E- and W-dipping, relatively small (up to 200 m throw) normal faults (Reeve et al., 2015), whereas several W-dipping, relatively small (up to 360 m throw) normal faults are present into the footwall of the Gjøa Fault (Figs 3 and 4).

Tectono-stratigraphic framework

The Måløy Slope is underlain by Caledonian metamorphic and igneous rocks, with the oldest sedimentary rocks being Early Jurassic (Statfjord Formation and Dunlin Group; Fig. 2) (e.g. Steel & Ryseth, 1990; Reeve et al., 2015). Middle Jurassic (the Aalenian-to-Bajocian Brent Group) rocks overlie the Early Jurassic sequence (e.g. Sørheim et al., 1990), with the complete succession being up to 300 m thick on the Måløy Slope (Figs 2 and 3) (e.g. Helland-Hansen et al., 1989).

During the early part of the Late Jurassic (Callovian and Oxfordian), flooding of the North Sea Basin resulted in deposition of shallow marine sandstone (Krossfjord, Fensfjord and Sognefjord formations), shelf mudstone and siltstone (Heather Formation), and eventually deep-marine mudstone and sandstone (Draupne Formation) (Fig. 2) (e.g. Helland-Hansen et al., 1989; Dreyer et al., 2005; Patruno et al., 2014; 2015; Holgate et al., 2015). Thickening of the Upper Heather and Draupne formations across many of the normal faults on the Måløy Slope indicates extension and normal faulting likely began during the Kimmeridgian (Fig. 4). Late Jurassic deep-water deposition was interrupted by the formation of a major erosional unconformity, which is herein referred to as the Upper Jurassic Unconformity or UJUNC (Figs. 3 and 4). Although dramatic in terms of its seismic expression, and the impact it had on preservation and thus the ultimate distribution of the underlying Heather and Draupne formations (Fig. 4), the exact geometry and processes responsible for the formation of this and younger unconformities remain unclear (Jackson et al., 2008; Sømme & Jackson, 2013; Sømme et al. 2013; Koch et al., 2017).

During the Early Cretaceous, many of the rift-related normal faults became inactive as the basin underwent a transition from relatively rapid, fault-controlled subsidence to relatively slow, thermal cooling-induced, post-rift subsidence (Gabrielsen et al., 2001; Fraser et al., 2003). In addition, the locus of subsidence migrated westwards into the axis of the Sogn Graben and mainland Norway was uplifted, possibly in response to the initiation of opening of the North Atlantic (Martinsen et al., 1999; Bugge et al., 2001; Gabrielsen et al., 2001). This decline in the rate of normal fault slip and basin subsidence, combined with ongoing deep-water deposition, resulted in healing of underlying rift-related topography. The Måløy Slope thus represented a westward-facing slope during much of the Cretaceous, with the basin floor lying >50 km to the west in the axis of the Sogn Graben (Fig. 1). Although generally considered a period of tectonic quiescence, it is likely the Øygarden Fault Complex was active during the Late Cretaceous (Færseth, 1996; Bell et al., 2014). Furthermore, several authors suggest this major fault controlled the location of the Late Cretaceous shelf-edge (Martinsen et al., 2005; Jackson et al., 2008; Sømme & Jackson, 2013; Sømme et al., 2013). Based on minor offset of Cretaceous seismic reflection events, it is clear that the Gjøa Fault was also reactivated and accumulated a relatively minor amount of displacement during the Late Cretaceous (Fig. 4A).

The Cretaceous succession on the Måløy Slope is up to 800 m thick and dominated by fine-grained pelagic carbonates and hemipelagic mudstone (Fig. 2) (Bugge et al., 2001; Gabrielsen et al., 2001; Kjennerud et al., 2001; Kyrkjebø et al., 2001). However, during both the Early Cretaceous (Albian) and Late Cretaceous (Late Turonian), a series of sand-rich submarine channels and fans were deposited on the Måløy Slope (Martinsen et al., 1999; Jackson, 2007; Jackson et al., 2008). These depositional systems were fed by material derived from the Norwegian mainland and these sediments were delivered to the slope via a series of shelf-edge canyons, which initially formed during the Late Jurassic (Jackson et al., 2008; Sømme & Jackson, 2013; Sømme et al., 2013).

DATASET

We use a 1200 km^2, pre-stack time-migrated, zero-phase processed, 3D seismic reflection dataset to map, in three-dimensions, basin structure and stratigraphy, including the slope canyons and their fill, forming the focus of this study (Figs 1 and 3). A downward increase in acoustic impedance is represented by a peak (black reflection in presented seismic images), and a downward decrease in acoustic impedance is represented by a trough (red reflection in presented seismic images) (i.e. SEG normal polarity; Brown, 2011). Inline and crossline spacing are 12.5 m, and the stratigraphic interval of interest lies at 500-3500 milliseconds two-way time (ms TWT); the frequency content of the data at this depth is 25-30 Hz and the average interval velocity is 2600-3175 m/sec, thereby yielding an approximate vertical resolution of *c*. 22-32 m. Seismic data quality varies from good to moderate, and the key rift-related structures and erosional unconformities are relatively well-imaged. Measurements in ms TWT are converted to metres using velocity data taken from boreholes within the study area. However, marked variations in the depth of burial of the studied succession occur due to the pronounced westward tilt of the basin margin (Fig. 4); we thus use interval velocities of 2600 m/sec and 3175 m/sec to convert values in proximal (i.e. to the east of the Måløy Fault) and distal (i.e. to the west of the Måløy Fault) areas, respectively. A range rather than an absolute value is presented for all measurements to account for ±10% uncertainty in the velocity values used for depth conversion.

We use data from seven exploration boreholes to constrain the age, lithology, thickness, and facies of the studied succession (35/9-1, 35/9-2, 35/9-3, 36/7-1, 36/7-2, 36/7-3, and 36/7-4; Figs 3-5). All of these boreholes, apart from 36/7-4, contain a standard suite of well-log and cuttings data, and one of the boreholes, 36/7-1, has 280 m of core within the interval of interest. 36/7-4 only contains lithostratigraphic top data. In boreholes lacking core data, we use cuttings data to constrain the lithology. Biostratigraphic data, derived principally from micro-palaeontology and palynology, constrain the age of key unconformities identified within the studied succession. These unconformities are related to periods of: (i) *erosion*, being generally defined where stratigraphic units and their associated biostratigraphic events are missing; and (ii) *stratigraphic condensation*, likely caused by marine flooding and/or non-deposition (Fig. 6; see also Table 1).

METHODS

To delineate the structure of the study area, and the geometry and distribution of the slope canyons associated with late Mesozoic unconformities, we mapped eight seismic horizons within the 3D seismic dataset (Figs 2 and 4). Isochron maps of the key stratal units constrain syn-depositional variations in accommodation, which in this tectonically active basin are principally related to rift-related normal faulting, and to variable preservation of stratigraphic units below and above the late Mesozoic canyons.

Seismic-stratigraphic relationships, in particular reflection truncation and onlap, were used to define the main canyons in seismic data (Fig. 7). Variations in the lithologies overlying and underlying the canyons mean the seismic expression of their basal erosion surface is highly variable in terms of polarity and amplitude. We therefore employed line-by-line seismic mapping to ensure that the geometry of the canyons, and the stratigraphic relationships between individual unconformities, was accurately captured. Seismically-defined unconformities were tied to boreholes using synthetic seismograms. The quality of the seismic-to-borehole ties was considered to be good-to-excellent, with <30 m mismatch between key reflection events expressed on the synthetics, and those identified and mapped in the seismic data (Fig. 8). Given that the unconformities capture the (preserved) thickness of the sequential canyon-fills, we use time-thickness (isochron) maps generated from key seismically-defined unconformities to show the geometry and distribution of the slope canyons. Isochron maps based on closely-spaced, serial seismic profiles trending broadly normal to the local canyon trend allow us to confidently define the position of the canyon thalwegs, and locally the canyon margins. Several boreholes are located in the footwalls to rift-related faults in locations where the Upper Jurassic to Lower Cretaceous succession is locally thin or even absent (Figs 3, 4b, 5b-c and 7c). In these locations, the unconformities are not expressed as a discrete reflection, and the thickness of stratal units they bound fall below the vertical resolution of seismic data. Where this occurs, we assume the major erosional unconformities mapped in the seismic data correlate to the longest duration, biostratigraphically-constrained unconformities identified in boreholes.

SUBSURFACE EXPRESSION OF SLOPE CANYONS

We identify three slope canyon-defining unconformities in the late Mesozoic succession of the Måløy Slope (UC1-3; Figs 2 and 4-7; see also Table 1). In this section, we combine seismic reflection and borehole data to describe the unconformities and related canyons in stratigraphically ascending order. For each unconformity the descriptions are arranged as follows: (i) a description of the stratigraphic (Figs 5 and 6; see also Table 1) and sedimentological (Figs 9 and 10) expression of the canyon-defining unconformity and flanking strata, based on borehole-derived data; (ii) a description of the three-dimensional geometry and geomorphological features associated with the unconformity, based on seismic reflection data (Figs 7 and 13); and (iii) an interpretation of the tectono-stratigraphic setting for the given time period, with an emphasis on the controls on the origin and evolution of the unconformities.

Unconformity 1 (UC1)

Stratigraphic and sedimentological expression. - Unconformity 1 (UC1) is identified in all six boreholes, although the time gap and stratigraphic expression varies significantly across the Måløy Slope. Towards the eastern basin margin, late Volgian marine mudstones (Draupne Formation) overlie late Oxfordian shallow marine clastics (Sognefjord Formation), suggesting an unconformity spanning *ca*.11 Myr (36/7-2; Figs 5A and 6; see also Table 1; cf. seismic expression of the unconformity described in section 4.1.2). Further west, in the footwall of the Gjøa Fault Zone, UC1 is underlain by Middle Bathonian shallow marine clastics (Fensfjord or Krossfjord formations), and overlain by Early Volgian (35/9-2 and 36/7-3) or early Hauterivian, deep-marine mudstones (35/9-1 and 36/7-1) (Figs. 5 and 6; Table 1). These stratigraphic relationships indicate that the time represented by the unconformity defined by UC1 decreases downslope to c. 4 Myr, and that the unconformity formed in the middle Kimmeridgian-to-early Volgian. Locally, however, on the crests of rift-related structural highs located in the central part of the study area, UC1 merges with UC2 to form a composite unconformity. In this location, the entire pre-Late Jurassic succession is absent, and Caledonian metamorphic and late Aptian marine mudstone subcrop and onlap UC1, respectively (i.e. 35/9-3; Figs 5a and b and 6; see also Table 1), indicating a time gap of >250 Myr.

Core data from 36/7-1, which was drilled on the northern margin of an UC1-related canyon (Canyon C; Figs 5a, 5c and 7c), constrains the sedimentological expression of UC1 in a relatively downslope position. These data indicate that the Sognefjord Formation, which is composed of shallow marine sandstone (Figs 9 and 10), is sharply and erosionally overlain by a *ca*. 35 m thick interval of deep marine deposits that include: (i) sharp-based, massive, decimetre-scale beds of fine-to-medium grained, turbidite sandstone, which locally are dewatered; (ii) metre-thick beds of very finely-laminated slope mudstone, which contain current-ripple laminated siltstones and very fine-grained sandstone; and (iii) thin beds of very poorly-sorted, and locally conglomeratic, muddy sand debrites (Figs 9 and 11). High gamma-ray values (>80 API) in well-log data and lithological observations from cuttings data indicate the upper part of the canyon-fill, above UC1 but below UC2, is dominated by hemipelagic mudstone (2018-2125 m; Figs 5a, 5c and 9). Well-log and cuttings data from other uncored wells indicate mudstone dominates the canyon-fill (35/9-1 and 35/9-2; Fig. 5); the exception to this is 36/7-3, which is drilled slightly north of the axis of Canyon 1B and that documents several 5-30 m thick, sharp-based, presumably turbidite sandstone-dominated packages (e.g. 2765-2795 m; Fig. 5B) separated by mudstone. Despite being relatively thin, these sandstone-rich packages overlie unconformities that define significant time gaps (up to 8.5 Myr; Fig. 6) that are as long as the major, slope-wide, canyon-defining unconformities (i.e. UC1, 2 and 3; Fig. 6 and Table 1).

Seismic expression and basin-scale morphology. - East of the Måløy Fault, UC1 is represented by a relatively high-amplitude, laterally continuous reflection that is conformable with underlying and overlying reflections (Figs 7a-b). Given that borehole data indicate a time gap of *ca*. 11 Myr along the basin margin, the lack of seismic-scale incision suggests UC1 is, at least in this position, related to a period of stratigraphic condensation and/or non-deposition, perhaps related to downslope sediment bypass (see below). In contrast, downslope, west of the Måløy Fault, UC1 defines a prominent erosion surface, along which four broadly ESE-trending canyons are developed (labelled 1A-D; Figs 7C-E and 13A). The heads of the southernmost canyons are located in the immediate hangingwall of the Måløy Fault (C and D; Fig. 13A). Although the heads of the northernmost canyons are not preserved due to erosion beneath the younger canyons associated with UC2, we infer they were located in the immediate hangingwall of the Måløy Fault (A and B; Fig. 13A). These canyons are up to 700 m deep and 9 km wide, typically widening downslope to the west (canyons C and D; Fig. 13A). The canyons are flat-bottomed, display 'U'-shaped geometry in cross-section, and their margins are smooth and dip up to 5° (Fig. 7C-E). The two northern canyons extend outside of the area of seismic data coverage, thus are at least 35 km long. In the immediate footwall of the Gjøa Fault, UC1-related canyons and their fill are eroded and thus variably preserved beneath younger, UC2- and UC3-related canyons (i.e. Canyon 1C in Fig. 7E; see also Fig 13A).

The four UC1-related canyons trend broadly perpendicular to the majority of rift-related faults on the Måløy Slope (Fig. 13A). However, these canyons are offset by the northern and southern segments of the Gjøa Fault, in addition to a number of smaller faults located in its hangingwall (Fig. 4). We note that the magnitude of base-canyon incision increases into the footwalls of the faults whereas the canyon fill thickness decreases (i.e. footwall of GFN in Fig. 4).

Origin and evolution. - Using seismic reflection and borehole data from the northern Måløy Slope and Slørebotn Sub-basin, Jackson et al. (2008) and Sømme et al. (2013) describe broadly 'Late Jurassic' canyons of comparable geometry and dimension to those associated with UC1. As such, we interpret UC1 to represent the along-strike continuation of the Upper Jurassic Unconformity (UJUNC) as previously defined elsewhere along the margin. We thus infer that UC1-related slope canyons are genetically related to those developed elsewhere offshore western Norway, suggesting slope incision occurred along an at least 300 km strike length of the southern Norwegian margin during the Late Jurassic.

However, the canyons may either have initiated during the late Kimmeridigian, synchronous with the latter stages of rifting, or reflect rejuvenation of older, antecedent systems. We note that the Måløy Slope is located at the northern margin of the Troll Delta, a large, sand-rich, basin margin-attached system sourced from the Norwegian mainland (Helland-Hansen et al., 1989; Husmo et al., 2003; Fraser et al., 2003; Dreyer et al., 2005; Patruno et al., 2014; 2015; Holgate et al., 2015). Shallow marine sandstones belonging to the uppermost, Oxfordian-to-early Kimmeridgian part unit of the Troll Delta (Sognefjord Formation) subcrop UC1 on the Måløy Slope (e.g. 36/7-1; Fig. 5C). Our data do not allow us to determine if these shallow-marine sandstones are confined within erosionally-based, canyon-like conduits; however, relatively recent analysis of the Sognefjord Formation on the Horda Platform, located only c. 20 km to the south, suggests this unit was deposited in an areally expansive, subaqueous delta, which was not structurally or erosionally confined (Patruno et al., 2015). Given this observation, and that the change from shallow- to deep-marine deposition (and erosion) appears relatively rapid (i.e. intra-Kimmeridge; Fig. 2), we infer that major antecedent drainage was not established on the Måløy Slope prior to UC1 incision, and that the UC1 canyons developed during the latter stages of rifting. We cannot rule-out that major incision occurred in the relatively short space of time between Sognefjord Formation deposition and the onset of UC1 erosion, but we see no evidence for this in our seismic (i.e. evidence for pre-UC1, seismic-scale erosion; e.g. Figs 4 and 7) or borehole (e.g. missing biostratigraphic zones; Fig. 6) data.

Given that the UC1 canyons appear to have initiated in the late Kimmeridgian, and based on observations from modern and ancient deep-marine systems, we can now explore the three principal mechanisms typically cited to explain the formation of submarine slope canyons: (i) marine flooding of incised-valleys eroded into a previously subaqueous shelf and cut in a subaerial setting by fluvial processes during a preceding period of relative sea-level fall (e.g. Van Wagoner, 1995); (ii) retrogressive failure of a slope in a fully subaqueous setting (e.g. Twichell & Roberts, 1982; McGregor et al., 1982; Farre et al., 1983); and (iii) incision of a slope by downslope-eroding sediment gravity flows in a fully subaqueous setting (e.g. Spinelli & Field, 2001; Jobe et al., 2011; Lonergan et al., 2013; Prélat et al., 2015; Lai et al., 2016). Given their markedly different modes of formation, and the environments in which they operate, the applicability of these mechanisms to the formation of the UC1 (and younger) canyons can be tested using observations from our geophysical and geological data.

UC1 canyons incised Upper Jurassic shallow marine rocks. Based on this stratigraphic relationship it is possible that UC1 represents a sequence boundary, and that canyons thus initiated as fluvially-cut valleys incised into the shelf in response to a relative fall in sea level. This interpretation implies the deep-marine rocks filling the canyons were deposited during the subsequent period of marine flooding, which eventually established deep-marine conditions across much of the Norwegian margin. However, we do not think that the UC1 canyons formed due to this process for the following five reasons: (i) the Late Jurassic was a time of eustatic sea-level rise, and it seems unlikely that a large magnitude fall in relative sea-level, due to local tectonic uplift, would have occurred during a time of crustal extension and rapid fault-driven subsidence; (ii) the canyons are developed within a fully marine sequence documenting a net increase in water depth with time; although core data are lacking in some boreholes, we have no evidence that UC1 was associated with subaerial exposure of the slope; (iii) canyons of similar dimensions are not observed at the same stratigraphic level elsewhere within the rift and, although some Upper Jurassic fault blocks were locally exposed and eroded, these are related to the formation of relatively narrow (<2 km), strike discontinuous (up to a few tens of km) 'islands' located in the footwalls of large, rift-related faults (e.g. Nøttvedt et al., 2000; Roberts et al., 2019); (iv) the canyons are significantly deeper than (incised) valleys typically formed in response to base-level fall; and (v) the magnitude of erosion increases downslope along UC1-related canyons; this is contrary to that predicted by an incised-valley model. It thus seemly highly likely that UC1 canyons formed in a submarine rather than subaerial setting, in response to mechanism (ii) (i.e. retrogressive failure of a fully submarine slope) or (iii) (i.e. incision of a fully submarine slope by downslope-eroding sediment gravity flows).

When considering these two mechanisms, we note that similar age strata subcrop UC1 across the slope, despite the surface presently displaying a pronounced westward dip (i.e. Late Jurassic; Figs 4A and 6; see also Table 1). This observation suggests only minor tectonic relief was generated at this time; more specifically, this implies that most of the westwards tilting of the Måløy Slope occurred later and that retrogressive slope failure played only a minor role in canyon initiation. This interpretation is consistent with the observation that pre-UC1 deposits are broadly tabular and do not thicken across slope-perpendicular faults, suggesting limited tectonic activity at this time.

Slope incision and canyon development may thus have occurred due to the input of erosive sediment gravity flows, perhaps sourced from basin-margin clastic systems depositionally similar to the stratigraphically older Sognefjord Formation delta (e.g. Dreyer et al., 2005; Patruno et al., 2014; 2015). This interpretation is consistent with data from borehole 36/7-3, which indicate the input of turbidites may be associated with significant time gaps (up to 8 Myr), perhaps related to seabed erosion and sediment bypass (e.g. Stevenson et al., 2015). Borehole and seismic data from 36/7-2 also indicate that the fault-controlled shelf was possibly an area of sediment bypass for up to 11 Myr; these sediments may then have directly entered the canyons at their headwalls, immediately downdip of the Måløy Fault (Fig. 13A).

Reeve et al. (2013) demonstrate that the intra-slope faults were active from the Middle Jurassic until the Early Cretaceous on the Måløy Slope (cf. Fraser et al., 2003; Bell et al., 2014), spanning the late Kimmeridgian period of canyon formation. However, what was the relationship between canyon *incision and infill*, and slip on and relief associated with the intra-slope faults? We envisage two plausible scenarios; (i) sediment accumulation rate was *less* than fault slip rate, meaning the faults generated intra-slope relief at the onset of canyon formation (i.e. intra-slope basins were underfilled); or (ii) sediment accumulation rate was *equal to or more than* fault slip rate, meaning the faults, did not generate appreciable intra-slope relief (i.e. intra-slope basins were balanced or overfilled) despite being active. Considering these two scenarios, it is clear the UC1 canyons maintained a broadly W- to NW-directed course across the intra-slope faults and were not, for example, deflected northward or southward to trend parallel to these broadly N-S-striking structures. This observation suggests the intra-slope faults did not generate intra-slope relief (i.e. scenario (ii)) and that UC1 canyons initially extended along the entire dip-extent of the slope. However, seismic (Fig. 4) and borehole (i.e. between 35/9-2 and 36/7-1; Fig. 5A; 36/7-3 and 35/9-3; Fig. 5B; 36/7-1 and 35/9-1; Fig. 5C) show that Upper Jurassic (upper Kimmeridgian; SU3) and Lower Cretaceous (Ryazanian-Barremian; SU4), canyon-fill strata thicken and that UC1 is itself offset across the intra-slope faults (see also the isochron maps in Fig. 13A and B), strongly suggesting syn-depositional (i.e. syn-filling) slip on at least major structures such as the Gjøa Fault. We thus envisage two plausible scenarios for the relationship between canyon *filling* and intra-slope faulting: (i) the canyons filled with early Volgian-to-early Barremian (SU3) sediment *before* being offset by latest Barremian slip on the Gjøa Fault, just prior to the formation of UC2; in this scenario, thickness changes in SU3 reflect post-depositional, intra-slope faulting, resulting in preservation of a thicker canyon-fill succession in the fault hangingwall below UC2; and (ii) the canyons filled with early Volgian-to-early Barremian (SU3) sediment synchronous with slip on the Gjøa Fault; in this scenario, thickness changes in SU3 reflect syn-depositional, intra-slope faulting. We cannot readily distinguish between these two scenarios with our available dataset, although it is clear that intra-slope tectonics and normal faulting controlled canyon-fill stratigraphy, if not the overall canyon trend.

Unconformity 2 (UC2)

Stratigraphic and sedimentological expression. - Like UC1, the time gap represented by UC2 varies markedly across the Måløy Slope, with the unconformity locally forming a composite surface with older (UC1) and younger (UC3) unconformities. Upslope, towards the eastern basin margin, in the immediate hangingwall of the Øygarden Fault Zone, early Aptian deep-marine strata directly overlie early Barremian deep-marine strata across UC2, which represents an unconformity with a duration of *ca*. 6.5 Myr (36/7-2; Figs 5A and 6; see also Table 1). Downslope, in the immediate hangingwall of the Gjøa Fault Zone and near the axis of a large canyon, late Aptian deep-marine strata directly overlie late Barremian deep-marine strata across UC2, thereby indicating an unconformity of *ca*. 18 Myr (36/7-3, on the northern flank of Canyon 1C; Figs 5A, 5B and 6; see also Table 1). We observe a slightly longer duration unconformity of *ca*. 21.5 Myr on the southern margin of the same canyon, defined by the juxtaposition of Early Hauterivian deep-marine strata above late Oxfordian-Kimmeridgian deep-marine strata (36/7-1; Fig. 6; see also Table 1). Slightly further downslope to the west, in the immediate footwall of the northern segment of the Gjøa Fault Zone, UC2 cuts down to merge with UC1, forming part of a composite unconformity documenting a time gap of at least 300 Myr; here, late Aptian deep-marine strata directly overlie metamorphic rocks (i.e. 35/9-3; Figs 5A and 6; see also Table 1). UC2 also forms a composite unconformity with UC3 further downslope to the south, in the immediate footwall of southern segment of the Gjøa Fault. Here, lower Turonian and late Aptian deep-marine strata are juxtaposed, recording an unconformity of *ca*. 6 Myr (i.e. 35/9-2; Figs 5A and 6; see also Table 1). In summary, we find the magnitude of erosion associated with UC2 increases downslope. However, it must be noted that UC2 typically incises down to broadly the same stratigraphic level within the early Barremian to earliest Aptian, with spatial variations in the associated unconformity reflecting onlap of progressively younger strata upslope (i.e. eastward); for example, late Albian and early Turonian strata overlie UC2 in distal areas, whereas early Aptian and early Albian overlie UC2 in proximal areas (Fig. 6; see also Table 1). Given the minimum unconformity documented, we interpret UC2 formed over a *ca*. 2 Myr period in the early Aptian (i.e. youngest rocks of Early Aptian age below UC2 in 35/9-2; oldest rocks of Early Aptian age above UC2 in 36/7-2).

Due to a lack of core data in the Lower and Upper Cretaceous and in basement rocks, we constrain the stratigraphic expression of the UC2 using only well-log and cuttings data. These data indicate that, where UC2 forms a discrete stratigraphic surface separate from UC1 and UC3, its stratigraphic expression is subtle, with deep-marine mudstone (Åsgard Formation) directly overlain by deep-marine mudstone (Sola Formation). As a result of this stratigraphic juxtaposition, UC2 has no distinct expression in well-log data (Fig. 5). However, in the north of the study area, where a large canyon developed along UC2, thick (up to 100 m) packages of turbidite sandstone occur in the Rødby Formation (i.e. Agat Formation in Canyon 2G in 36/7-3; Fig. 5A and 5B) (Martinsen et al., 2005). Biostratigraphic data do not resolve erosion-related unconformities at the bases of these sandstone-rich packages (Fig. 6).

Seismic expression and basin-scale morphology. - On the proximal, eastern part of the Måløy Slope, east of the Måløy Fault, UC2 is expressed as a relatively low-amplitude, laterally continuous reflection that is conformable with underlying and overlying reflections (Fig. 7A), or that truncates underlying reflections basinward at a relatively low angle (Fig. 4A). Given that borehole data indicate an unconformity of *ca*. 6.5 Myr along the basin margin, the lack of seismic-scale incision suggests UC2 is, at least in this position, related to a period of stratigraphic condensation and/or non-deposition, perhaps related to downslope sediment bypass (see below). In contrast, further downslope to the west, UC2 forms a prominent erosion surface, along which four, very broad, canyon-like features are developed (Figs 7B-E and 13B). Constraining the position of the canyon heads is problematic due to deep incision below younger, UC3-related canyons; however, we infer the heads of UC2-related canyons were located either in the immediate hangingwall or the immediate footwall of the Måløy Fault (Fig. 13B). UC2 canyons are straight, and trend SE or SSE, thus are slightly oblique to those developed along UC1 (cf. Figs 13A and 13B). The canyons are ‘V’- or ‘U’-shaped in cross-section (margins dips of up to 5°), and are <2 km wide and up to 400 m deep, thus are generally narrower and shallower than those along UC1. UC2 canyons abruptly widen and display less relief downslope to the NW, passing into a low-relief erosion surface lacking canyons (Figs 7A-D). However, further downslope, in the footwall of the Gjøa Fault, the magnitude of erosion along UC2 increases dramatically and a very wide (up to 10 km wide), deep (c. 550 m) canyon-like feature is developed (Canyon 2G; penetrated by 35/9-3; Fig. 7E and D; see also Figs 5A-B and 13B). A second canyon-like feature, which is at least 10 km wide, c. 320 m deep and superimposed on an underlying, UC1-related canyon, is developed in the south of the study area (Canyon 2F, superimposed on Canyon 1D; Figs 7E and 13B).

The four canyon-like features developed along UC2 trend broadly perpendicular to the majority of rift-related faults (Fig. 13B). Furthermore, the majority of normal faults on the Måløy Slope tip-out beneath UC2 (Fig. 4), although the depth of incision along its base increases markedly across the Gjøa Fault, such that erosional relief and canyon-like features are present in the footwall of this structure (Figs 7E and 13B). UC2 is also locally offset, by up to 20 ms TWT, across normal faults adjacent to the Gjøa and Måløy fault zones (Fig. 4).

Origin and evolution. - UC2 document a second period of slope incision, forming over *ca*. 6.5 Myr in the early Albian, after a *ca*. 30 Myr period of UC1 canyon filling and broader slope onlap. Based on the criteria discussed above, and given its development within a fully deep-marine succession, it seems likely that UC2 also formed subaqueously, due to either slope failure and/or erosion by sediment gravity flows. Furthermore, the occurrence of a moderate unconformity (*ca*. 6.5 Myr) along the basin margin, coupled with a lack of seismic-scale erosion, implies that, like UC1, UC2 was not associated with major erosion of the shelf, but rather a protracted period of sediment bypass to the slope. The eastward extension of UC2 canyons upslope of those developed along UC1 (i.e. slightly in to the footwall of the Måløy Fault) suggest the Måløy Fault System was not as active and may have become inactive by the early Albian, resulted in a weakly fault-controlled shelf edge, and allowing canyons to propagate landward.

Downslope, UC2 canyons incised into and reworked sediments previously deposited within UC1 canyons. As argued above, the relatively thin UC1 succession on the footwall of the major intra-slope fault system (Gjøa Fault System) likely reflects decreased preservation of slope strata beneath UC2 canyons, rather than syn-UC2 (incision or filling) fault activity. Furthermore, local amalgamation of UC2 and UC3 in the footwall of the Måløy Fault System suggests this structure was also active at this time.

Unconformity 3 (UC3)

Stratigraphic and sedimentological expression. - UC3 is similar to the older unconformities in that its stratigraphic expression varies across the Måløy Slope. On the upper slope, in the immediate hangingwall of the Øygarden Fault Complex, UC3 locally merges with older (including UC2) and younger unconformities, capping and being overlain by late Albian and early Eocene deep-marine deposits, respectively; this defines a time gap of *ca*. 155 Myr (i.e. 36/7-2; Figs 6 and 7A; see also Table

1). Downslope to the west, UC3 is typically characterised by a correlative conformity defining a transition from Albian to Cenomanian deep-marine mudstone (i.e. 35/9-3, 36/7-1, 36/7-3; Figs 5 and 6; see also Table 1). The exception to this occurs in 35/9-2, in the immediate footwall of the Gjøa Fault System, where UC3 merges with UC2, thereby defining an unconformable upward transition from early Aptian deep-marine mudstone to early Turonian deep-marine marl, and a time gap of *ca*. 20 Myr (Figs 5A and 6; see also Table 1). Constraining the age of UC3 is difficult; both 36/7-2 and 35/9-2 penetrate UC3 where it forms part of a composite unconformity, whereas other wells penetrate it in a relatively distal position where it defines a correlative conformity (35/9-3, 36/7-1, 36/7-3). 35/9-2 at least constrains the possible oldest (i.e. early Albian) and youngest (i.e. early Turonian) age, and the maximum time gap (i.e. *ca*. 20 Myr) associated with UC3. However, UC3 must be younger than UC2 (early Aptian), suggesting it defines an unconformity of <20 Myr duration.

Seismic expression. - On the upper slope, on the northern part of the terrace bound by the Måløy Fault and the Øygarden Fault Complex, UC3 is typically expressed as a major angular unconformity; however, as described above, borehole data indicate that, on the southern part of this terrace, UC3 forms a composite unconformity with the much younger, base Pleistocene unconformity (Fig. 7A) (Martinsen et al., 2005). Slightly further downslope, UC3 is strongly erosional and represented by a discrete surface that marks the development of at least five canyons (Figs 7A and 7B), which pass north-westwards into a conformable, canyon-free surface on the lower slope (Fig. 13C). These upper slope canyons are relatively straight and trend E or SE, slightly oblique to those that developed slightly further downslope along UC2, and sub-parallel to those located even further downslope in UC1 (cf. Fig. 13C with Figs 13A and B). In cross-section, the UC3-related canyons have distinct 'V'-shaped geometries, and are up to 300 m deep, 3 km wide and have relatively steep margins (up to 10°). The observation that UC3 becomes conformable downslope is consistent with observations from borehole data (see above).

UC3 is rarely offset by any rift-related faults (Fig. 4), implying the majority of these structures were inactive before the early Cenomanian. An exception to this is observed in the north-eastern corner of the study area, where UC3 is offset across a NW-SE-striking segment of the Måløy Fault (Fig. 4B).

Origin and evolution. - UC3 documents a third and final period of slope incision, forming over a time period of a little under *ca*. 20 Myr in the early Turonian, following a *ca*. 20 Myr period of UC2 canyon filling and broader slope onlap. We again infer that UC3 formed in a fully marine setting, given its development within deep-marine strata; as such, we suggest that, like UC1 and UC2, UC3 also formed subaqueously due to either slope failure and/or erosion by sediment gravity flows. Because base Pleistocene erosion removes the stratigraphic record of UC3 in the immediate hangingwall of the Øygarden Fault Zone, it is not clear if, like UC1 and UC2, UC3 dies-out onto the outer reaches of the contemporaneous shelf, and if it therefore associated with shelf bypass with limited erosion. However, it is clear that UC3 extended further upslope than the older canyons, and that it died-out downslope to the west into a correlative conformity. These observations indicate continued landward propagation of the erosional surfaces and associated canyons, possibly in response to continued basin margin uplift, which tilted the slope and augmented erosion, and ongoing subsidence in the basin centre, which suppressed erosion and resulted in the formation of a correlative conformity. In contrast to UC1, and in a similar manner to UC2, intra-slope faults were inactive during the formation and filling of UC3 canyons.

DISCUSSION

Stratigraphic context and the origin of submarine canyons

We used 3D seismic reflection and borehole data to describe the geometry and stratigraphic context of several large (up to 700 m deep and 9 km wide), slope-confined canyons preserved in late Mesozoic strata of the northern North Sea, offshore western Norway. These canyons record relatively protracted periods (c. 2-17 Myr) of slope degradation, separated by >10 Myr periods of deposition and slope accretion. The geometry and scale of these canyons, 100's metres deep and kilometres wide, are comparable to others described from modern, outcrop and in seismic reflection data. Here we briefly describe some notable, relatively well-documented examples of submarine canyons, focusing on debates related to their stratigraphic context and genesis. We then consider the implications of these previous studies, in addition to our data presented here from the northern North Sea, for the slope evolution and stratigraphy.

The Wonoka canyons (Neoproterozoic; 570-550 Ma), South Australia are some of the largest and best-exposed, yet controversial canyons described from the rock record. The canyons are up to 1.5 km deep and 4 km wide, and thus of broadly comparable dimensions to those described here from the northern North Sea. The Wonoka canyons are also similar to the North Sea examples in that multiple (up to five) periods of incision and canyon formation are interpreted. In terms of their general stratigraphic context, most authors agree that the Wonoka canyons emanate from the lower, deep-water part of the Wonoka Formation and are underlain by deep-water rocks of the Bunyeroo Formation. The stratigraphic fill and thus origin of the Wonoka canyons remains highly contentious. For example, some authors interpret a deep-water canyon-fill succession, arguing the canyons formed in a fully submarine setting (e.g. von der Borch et al., 1982; Giddings et al., 2010). In contrast, other authors interpret the fill is fluvial-to-shallow marine (e.g. Eickhoff et al., 1988; von der Borch et al., 1989; Christie-Blick, 2001), arguing that the canyons formed as incised valleys (*sensu stricto*; Van Wagoner, 1995; Van Wagoner et al., 1998) that were filled during subsequent marine flooding (Christie-Blick et al., 1995). The latter model requires km-scale changes in relative sea-level, which lead these authors to invoke regional tectonic uplift or 'Messinian-style' drawdown of the marine waters. Giddings et al. (2010) recently rejected this interpretation, arguing that: (i) the basal conglomerate is marine, being deposited by strongly erosive, very coarse-grained sediment-gravity currents that carved the canyons; (ii) there is no evidence for subaerial exposure and related erosion of the canyon walls; and (iii) multiple large-magnitude changes (i.e. several hundreds of metres) in relative sea-level, which are required to drive incision and canyon formation, are highly unlikely.

The Baliste-Crécerelle canyon lies within the Upper Oligocene-Middle Miocene Mandarove Formation, offshore Gabon (Wonham et al., 2000). The canyon is up to 4 km wide and 500 m thick, slope-confined, and likely formed in a fully submarine setting over a >10 Myr; the scale and stratigraphic context and of the Baliste-Crécerelle canyon is thus broadly similar to that documented here from the northern North Sea. The base of the canyon is inferred to be diachronous, and six intra-canyon erosion surfaces are identified within the canyon itself. These erosion surfaces, and the stratigraphic packages they bound, record several phases of erosion and sediment bypass, and canyon filling, possibly related to relative sea-level change and related changes in sediment supply from the shelf. The Baliste-Crécerelle canyon is thought to have formed via retrogressive failure of the slope in response to uplift of the African continent, with upslope propagation of the canyon heads eventually cannibalizing the outer shelf. Erosion of the canyon base was augmented by sediment gravity-flows derived from rivers or longshore drift.

Submarine canyons are also described using 3D seismic reflection data from the Ebro Continental Margin, western Mediterranean (Bertoni & Cartwright, 2005). Although also slope-confined, these Plio-Pleistocene canyons are smaller (0.5-2 km wide, 10-15 km long, and incise >50 m) than those we describe from the northern North Sea. Several periods of canyon incision and filling are identified within a relatively short period (i.e. 1.7 Myr, based on study interval duration indicated in their fig. 5). The vertical extent of the canyons (>500 m), and their restriction to the upper-middle slope to the base-of-slope of well-developed clinoforms, is clear evidence for their fully submarine origin; critically, their vertical extent is far greater than the magnitude of any eustatic sea-level falls documented for the stratigraphic interval of interest (i.e. maximum of 100-150 m; Haq et al., 1987). Their shelf-detached location, and their linear geometry, points to an origin by internally slope-driven failure, with some contribution by erosion by shelf-sourced sediment gravity-currents (e.g. Bakley et al., 1990; Robb, 1990; Pratson & Coakley, 1996).

This brief synthesis of some well-documented examples of exposed and buried examples of submarine canyons indicate these features often form in fully deep-marine conditions, with limited or no evidence for subaerial exposure. This observation, coupled with the recognition that slope incision can occur at any stage in the relative sea-level cycle (e.g. Ebro; Bertoni & Cartwright, 2005), argues against relative sea-level fall (or at least for complete subaerial exposure of the entire canyon length/dip extent) as the main driver for canyon formation. Establishing the trigger for slope degradation remains challenging. In the case of the Baliste-Crécerelle and, potentially, Wonoka canyons, margin-scale tectonic uplift may have driven canyon formation; in the case of the northern North Sea examples we describe here, slope tilting may also reflect tectonically-driven uplift of the basin margin and simultaneous subsidence of the basin axis, or simply faster subsidence in the basin axis compared to the more slowly subsiding basin margin (cf. Artoni, 2013). In the case of the Ebro Continental Margin, much smaller, more clearly slope-confined, ‘gully-like’ canyons may have formed in response to downslope-eroding sediment gravity currents derived from the shelf edge, rather than major tectonic uplift and/or differential uplift (e.g. Spinelli & Field, 2001; Jobe et al., 2011; Lonergan et al., 2013; Prélat et al., 2015). The duration and pacing of degradational events likely reflects the factors controlling slope instability and incision; in the case of the northern North Sea and offshore Gabon, this is the pulsed nature of slope tilting, whereas in the Ebro example this would be the magnitude and timing of sediment delivery to the shelf edge and upper slope.

Stratigraphic development of submarine canyons

Our ability to read the stratigraphic record of submarine canyon formation, evolution, and abandonment is poor because: (i) field exposures are spatially limited and contain the stratigraphic expression of only one or a few cycles of slope aggradation and degradation, and/or lack detailed chronostratigraphic constraints (e.g. Giddings et al., 2010; Hodgson et al., 2011, 2016); and (ii) bathymetric maps of the present seabed and/or near-seabed geophysical studies do not permit analysis of the longer-term (10^3-10^4 Myr) stratigraphic development of submarine canyons (e.g. McGregor et al., 1982; Twichell & Roberts, 1982; Spinelli & Field, 2001; Jobe et al., 2011; Lonergan et al., 2013). Using our high-quality 3D seismic reflection and borehole dataset, we are able to show that spatially varying patterns of canyon-related erosion and deposition lead to the development of a complex stratigraphic record. For example, slope-confined submarine canyons in the northern North Sea are underlain by erosion surfaces that pass downdip into correlative conformities. Updip towards the basin margin, canyon bases may pass into cryptic stratigraphic surfaces that document non-deposition and/or erosion, but which lack evidence for seismic-scale incision (e.g. UC1). Furthermore, canyon-driven erosion, transport and (re)deposition leads to spatially complex patterns of sedimentation both above and downdip of the main areas of canyon incision. For example, after the canyon has formed, its downdip reaches are filled before more proximal areas, although in some cases the late-stage record is removed by younger canyons (see also Jackson et al., 2008). The transition from erosion to deposition in post-rift systems likely reflects the position of the ‘fulcrum’; i.e. the approximate point around which the slope rotates (Fig. 14). The stratigraphic record of areas updip of the fulcrum is principally controlled by relative sea-level fall and net-erosion, whereas the record of those downdip of this position are more strongly influenced by relative sea-level rise and net-deposition. However, this process will be highly time transgressive and the location of the fulcrum may migrate, leading to a complex distribution of related deposits and their bounding surfaces (e.g. erosion and flooding surfaces; Fig. 14). The ultimate distribution of erosion surfaces and preservation of overlying deposits are likely controlled by the rate of slope tilting, the magnitude of incision, and the position and movement of the fulcrum point; in the case of the Måløy

Slope, there is a systematic migration of the canyons upslope towards the basin margin suggesting that the fulcrum of basinward tilt also migrated landward through time (Fig. 14). However, this configuration likely varied along strike, reflecting lateral changes in the rate and magnitude of slope rotation and sediment supply.

Implications for tectono-stratigraphic models of marine rifts

Marine rift-basin tectono-stratigraphic models indicate syn-rift sediment dispersal is intimately linked to the growth of normal faults (Gawthorpe & Leeder, 2000). During the early stage of rifting (so-called 'rift initiation'; *sensu* Prosser, 1993), relatively small volumes of sediment are derived from the low-relief scarps of numerous, short, low-displacement faults. In contrast, during the latter stages of rifting (so-called 'rift climax'; *sensu* Prosser, 1993), large volumes of sediment are sourced from the high-relief scarps formed in the footwalls of a few, long, large-displacement faults that accommodate the majority of ongoing rift-related strain. Strain localisation onto a few large faults causes increasing topographic segmentation of the rift and the formation of wider, deeper graben and half-graben; as a result, sediments sourced from relatively large antecedent systems are trapped in proximal depocentres (rift-margin), leading to sediment starvation in more distal areas (rift-axis).

Our subsurface study of the Mesozoic succession the Måløy Slope, northern North Sea shows that the formation of large canyons during the latter stages of rifting can result in the spatially focused bypass of sediment towards the rift-axis. During the very earliest phase of their development, these canyons may represent major sediment conduits that cross-cut still-active normal faults at a high angle, establishing a transverse supply system that links and feeds sediment to otherwise isolated depocentres. The volume of sediment transported through these conduits may be substantially greater than those supplied by relatively small drainage systems formed in response to fault-driven uplift of intra-rift structural highs (see Fig. 7D in Gawthorpe and Leeder, 2000).

The model we present for the Måløy Slope should be applied with caution to other rifts. This is because: (i) other rifts likely evolve over different time- and length-scales depending on, for example, variations in extension rate, and crustal rheology and structure (see review by Peron-Pinvidic et al., 2019); and (ii) the temporal and spatial scales over which submarine slopes prograde and degrade vary, reflecting differences in, for example, sediment supply (e.g. Olariu and Steel, 2009), and the presence and vigour of oceanographic currents (e.g. Brackenridge et al., 2020). For example, a sub-basin located near the axis of a marine rift, detached from a large, basin-margin sediment source, will likely be characterised by broadly conformable, late syn-rift to early post-rift succession that lacks widespread erosion and canyon formation. Indeed, this description characterises the Albian-to-Cenomanian succession on the distal, lower-slope part of the Måløy Slope, which evolved basinward and beyond the influence of the UC3-related canyons (Fig. 13C). Despite these variations in the tectono-sedimentary setting of individual rifts, we maintain that existing marine rift-basin tectono-stratigraphic models should be updated to capture late syn-rift canyon formation and transverse sediment supply systems, especially in the case of margin-attached sub-basins located near major sediment sources.

7. Conclusions

We integrated 3D seismic reflection and borehole data to determine the geometry and origin of ancient (Late Mesozoic) slope canyons, and their infills, on the Måløy Slope, offshore western Norway. We show that the initial phase of slope degradation (UC1) started in the Late Jurassic (late Kimmeridgian), during a period of rifting and active normal faulting. Two subsequent periods of slope degradation and canyon formation and infilling occurred during the post-rift in the Aptian-to-Albian (UC2) and Albian-to-Cenomanian (UC3). We constrain the timescales over which slope progradation and degradation occur, showing that the canyons record relatively protracted periods (c. 2-17 Myr) of slope degradation, separated by >10 Myr periods of deposition and slope accretion. Boreholes indicate that the canyons bases are defined by sharp, erosional surfaces, across which we observe an abrupt upward shift from shallow marine to deep marine (UC1), or deep marine to deep marine facies (UC2 and 3). The canyons are relatively straight, up to 700 m deep and 10 km wide on the upper slope, and die-out downdip onto the lower slope. All the canyons trend broadly perpendicular to, and crosscut most of, the rift-related Late Jurassic normal faults, although syn-filling fault slip resulted in the local preservation of thicker canyon-fill successions. The updip extent of the oldest, late Kimmeridgian canyons is defined by a fault-controlled shelf-edge, whereas the younger, Cretaceous canyons overstepped the now-inactive fault and incise the shelf. In the middle slope, UC2 canyons either erode into the older, UC1 canyon-fills, forming a complicated set of unconformities; in contrast, UC3 canyons die-out downslope into correlative conformities. Slope degradation and canyon formation likely reflects some combination of basinward tilting and over-steepening of this tectonically active rifted margin, augmented by slope incision by erosive sediment gravity-flows. We show that the geophysical and geological (i.e. stratigraphic) expression of slope degradation-related features (i.e. canyons) in the rock record is complex, and that their formation can drive a major reorganisation of rift-related drainage patterns and sediment dispersal; this is not currently predicted by existing marine rift-basin tectono-stratigraphic models.

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Figure captions

Fig. 1. (A) Map showing the location of the study area (dashed black line) and major hydrocarbon exploration quadrants. Major structural elements, including key rift-related normal faults, are shown, as is the location of the cross-section shown in (B). Inset map shows the regional geographic location of the study area. CVG=Central Viking Graben; NVG=North Viking Graben; SG=Sogn Graben; LT=Lomre Terrace; UT=Uer Terrace; HP=Horda Platform; MS= Måløy Slope; TS=Tampen Spur; StF=Statfjord Fault; SnF=Snorre Fault; MFC=Mokkurkalve Fault Complex; SGF=Sogn Graben Fault; KF=Kinna Fault; VF=Vette Fault; GF=Gjøa Fault; GG=Gjøa Graben; MF= Måløy Fault; ØFC=Øygarden Fault Complex. (B) Schematic cross-section showing the regional geological setting of the study area.

Fig. 2. Stratigraphic column showing the interval of interest. We focus on the major Mesozoic unconformities (UC1-3). The maximum flooding surface (MFS) ('J-sequence') nomenclature is after Underhill & Partington (1993) (see also Fraser et al., 2003). The seismic horizon and stratal unit (SU) colour legend shown here applies to the geoseismic profiles in Figs 4 and 7, and the synthetic seismograms in Fig. 8.

Fig. 3. (A) Time-structure map of the top basement seismic horizon, illustrating the rift-related structure of the study area. Locations of seismic profiles in Figs 4a-b and 7a-e are shown, as are the locations of key boreholes used in the study (see Figs 5, 6, 8 and 9-12). (B) Dip-map of the top basement time-structure map highlighting the key structural elements within the study area. The locations of stratigraphic correlations in Fig. 5 are shown. GFN= Gjøa Fault North; GFS= Gjøa Fault South; MFN= Måløy Fault North; MFS= Måløy Fault South; ØFC=Øygarden Fault Complex; GG= Gjøa Graben.

Fig. 4. (A) SE- and (B) SSE-trending seismic (above) and geoseismic (below) profiles across the study area, showing the rift-related structure of the deep part of the Måløy Slope, and the structural and stratigraphic context of the Mesozoic unconformities (UC1-3; see Fig. 2). Abbreviations for the structural elements is in Fig. 3. The locations of intersecting seismic and geoseismic profiles in Fig. 7 are shown. 1C, 2G, etc, refer to specific unconformity-related canyons referred to in the text and highlighted on the stratigraphic correlations shown in Fig. 5.

Fig. 5. (A) Regional stratigraphic correlation between key boreholes on the Måløy Slope showing the overall structural and stratigraphic context of the Mesozoic unconformities (UC1-3; see Fig. 2). Note that 36/7-4 only contains lithostratigraphic tops data; because of this, the position of the biostratigraphically-constrained, chronostratigraphic surfaces that are clearly expressed in the nearby 36/7-3 borehole are unknown and can only be crudely estimated. (B) Local stratigraphic correlation between boreholes located in the footwall (35/9-3) and hangingwall (36/7-3) of the GFN. (C) (B) Local

stratigraphic correlation between boreholes located in the footwall (35/9-1) and hangingwall (36/7-1) of the GFS. The overall lithology of the material eroded into by and filling the Mesozoic canyons, and overlying with intra-canyon bypass surfaces, is shown. Stars on the left-hand side of the boreholes indicate the locations of the biostratigraphic samples that constrain the chronostratigraphic framework and surface correlation. Hachured areas in the 'chronostratigraphy' column indicate areas lacking age diagnostic fauna. The geometry of UC1-related canyons between 35/9-1 and 35/9-2, and in the immediate hangingwall of the GFS and GFN, is constrained by observations from seismic reflection data (e.g. Figs 4 and 7). Location of correlation shown in Fig. 3B. 1C, 2G, etc, refer to specific unconformity-related canyons referred to in the text and highlighted on the seismic profiles shown in Figs 4 and 7.

Fig. 6. Wheeler-style diagram showing the stratigraphic context and expression of Mesozoic unconformities on the Måløy Slope. Stratigraphically continuous sections are shown in white; hachured areas indicate major time gaps, some of which define seismic-scale canyons (see Figs 4 and 7). Error bars are based on uncertainties related to biostratigraphic sample spacing (see stars in Fig. 5). 36/7-4 lacks biostratigraphic data and is therefore not shown. Note that, away from intra-slope structural highs, UC3 represents a correlative conformity (indicated by brown dashed line; 35/9-3, 35/9-1, 36/7-1, and 36/7-3). See text for full discussion.

Fig. 7. N-trending seismic (above) and geoseismic (below) profiles across the study area, showing the rift-related structure of the deep part of the Måløy Slope, and the structural and stratigraphic context of the Mesozoic unconformities (UC1-3; see Fig. 2). (A) is located in the most proximal/upslope position; (E) is located in the most distal/downslope position. Abbreviations for the structural elements is in Fig. 3. The locations of intersecting seismic and geoseismic profiles in Fig. 4 are shown. 1C, 2G, etc, refer to specific unconformity-related canyons referred to in the text and highlighted on the stratigraphic correlations shown in Fig. 5.

Fig. 8. Synthetic seismograms for (A) 36/7-3 and (B) 36/7-1. See Figs 3-5 and 7 for location of boreholes. The overall lithology of the material eroded into by and filling the Mesozoic canyons is shown. See Fig. 2 for colour legend.

Fig. 9. Core description from 36/7-1, showing the sedimentary facies and depositional environment immediately below and above UC1. Note the sharp upward transition from shallow- (shoreface) to deep-marine (slope canyon) facies. See text for full description. See Figs 3 and 5 for location of the borehole. Locations of photos shown in Figs 11 and 12 are labelled.

Fig. 10. Core photograph showing the sedimentary facies and depositional environments encountered below UC1 in 36/7-1. Note the upwards transition from highly bioturbated, silt-rich, lower shoreface sandstone to bioclastic, medium-grained, upper shoreface sandstone; the contact between these two facies is sharp, occurring across a regressive surface (RS). Location of photo shown in Fig. 9.

Fig. 11. Core photograph showing the sedimentary facies and depositional environments encountered immediately below and above UC1 in 36/7-1. Note the sharp upward transition from shallow- (shoreface) to deep-marine (slope canyon) facies. See text for full description. Location of photo shown in Fig. 9.

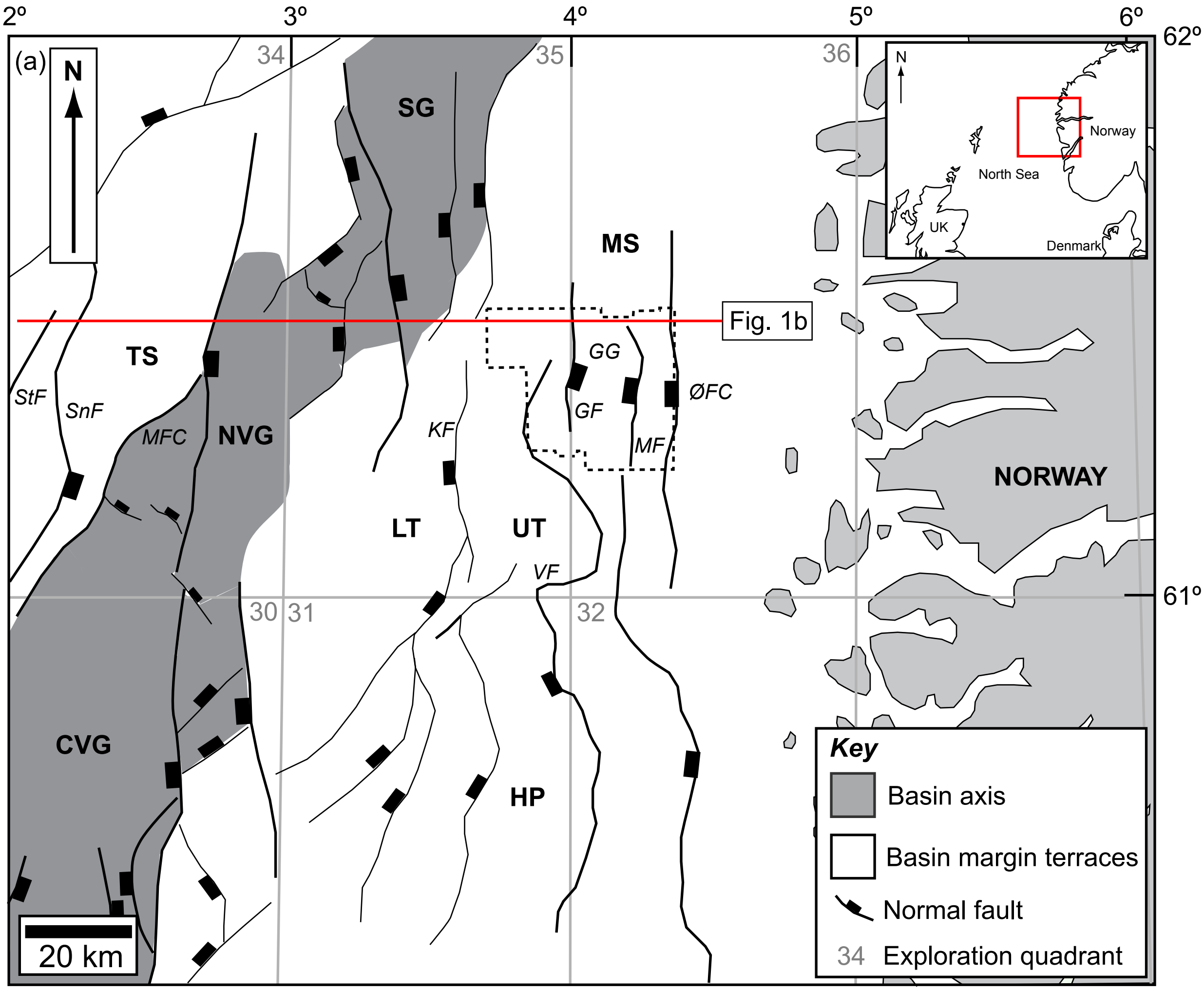
Fig. 12. Photographs showing details of the deep-marine facies encountered in the UC1-related canyon-fill succession. (A) deformed, thin-bedded, mud clast-rich, fine-grained turbidites, erosively overlain by mud- and clast-rich debrite; (B) thin-bedded, very fine-grained turbidites, erosively overlain by sand- and clast-rich debrite; (C) strongly deformed, thin- and thick-bedded turbidites; (D) sandy conglomerate debrite containing abundant extrabasinal clasts.

Fig. 13. Isochrons (time-thickness maps) (left) and sketches (right) illustrating the geometry and distribution of Mesozoic unconformity-related canyons on the Måløy Slope. (A) SU3 (related to UC1); (B) SU4 (related to UC2); and (C) SU5 (related to UC3). Solid red lines=mapped canyon thalwegs; dashed red lines= inferred canyon thalwegs. See text for full discussion.

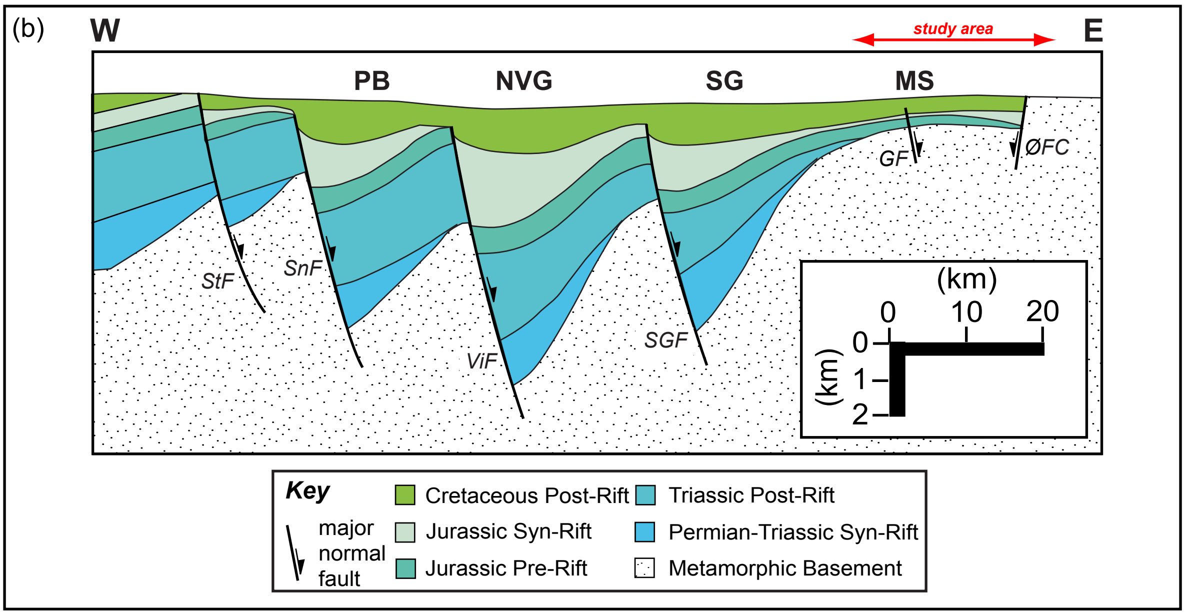
Fig. 14. Cartoon to illustrate the response of basinward tilting of a basin margin. Numbers are time steps from oldest (1) to youngest (3). (A) The fulcrum point about which the basin margin rotates is fixed with erosion on the landward side, and deposition on the basinward side. (B) The fulcrum point moves basinward through time, such that the point of erosion also moves basinward with an offlap configuration. (C) The fulcrum point moves landward through time, such that the updip pinchout moves landwards resulting in an onlap configuration.

Table 1. Table showing the stratigraphic context and sedimentological expression of Mesozoic unconformities on the Måløy Slope.

Fig. 1

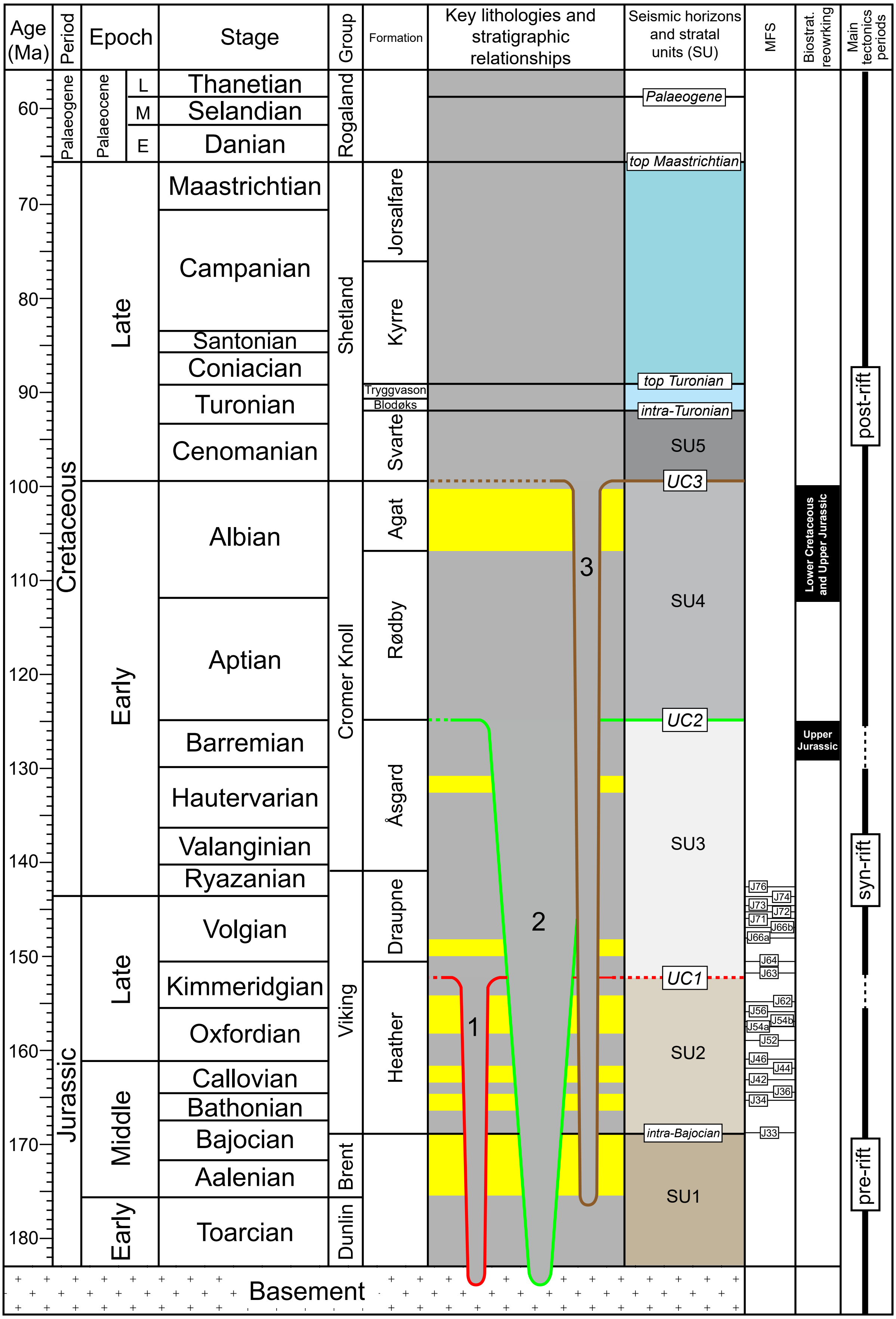


2°
3°
4°
5°
6°
62°
61°
(a)
N
34
35
36
SG
MS
TS
NVG
CVG
LT
UT
HP
NORWAY
StF
SnF
MFC
KF
GG
GF
MF
ØFC
VF
30
31
32
Fig. 1b
Norway
North Sea
UK
Denmark
20 km
Key
Basin axis
Basin margin terraces
Normal fault
34 Exploration quadrant



(b)
W
E
study area
PB
NVG
SG
MS
GF
ØFC
StF
SnF
ViF
SGF
(km)
0
10
20
0
1
2
(km)
Key
major normal fault
Cretaceous Post-Rift
Jurassic Syn-Rift
Jurassic Pre-Rift
Triassic Post-Rift
Permian-Triassic Syn-Rift
Metamorphic Basement

Fig. 2



Lithological key

= mudstone-dominated

= sandstone-dominated

Fig. 3

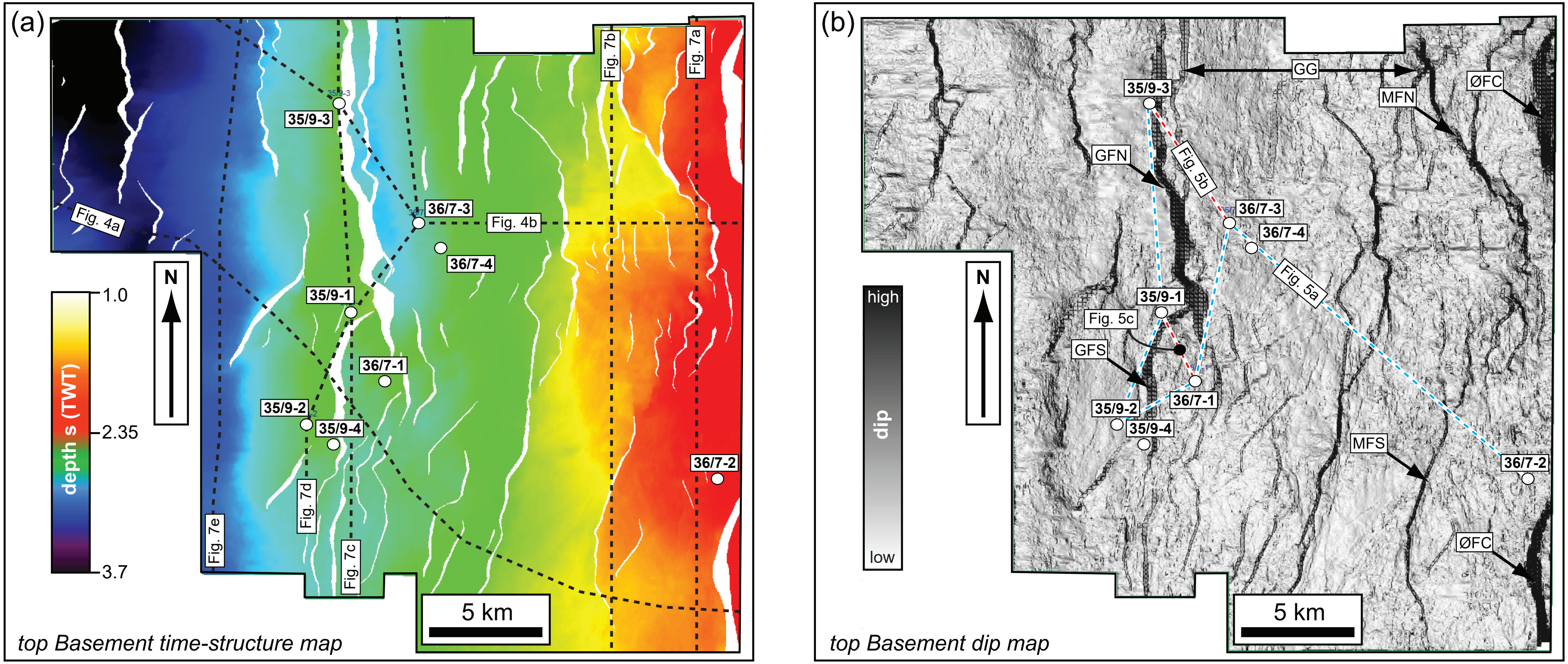
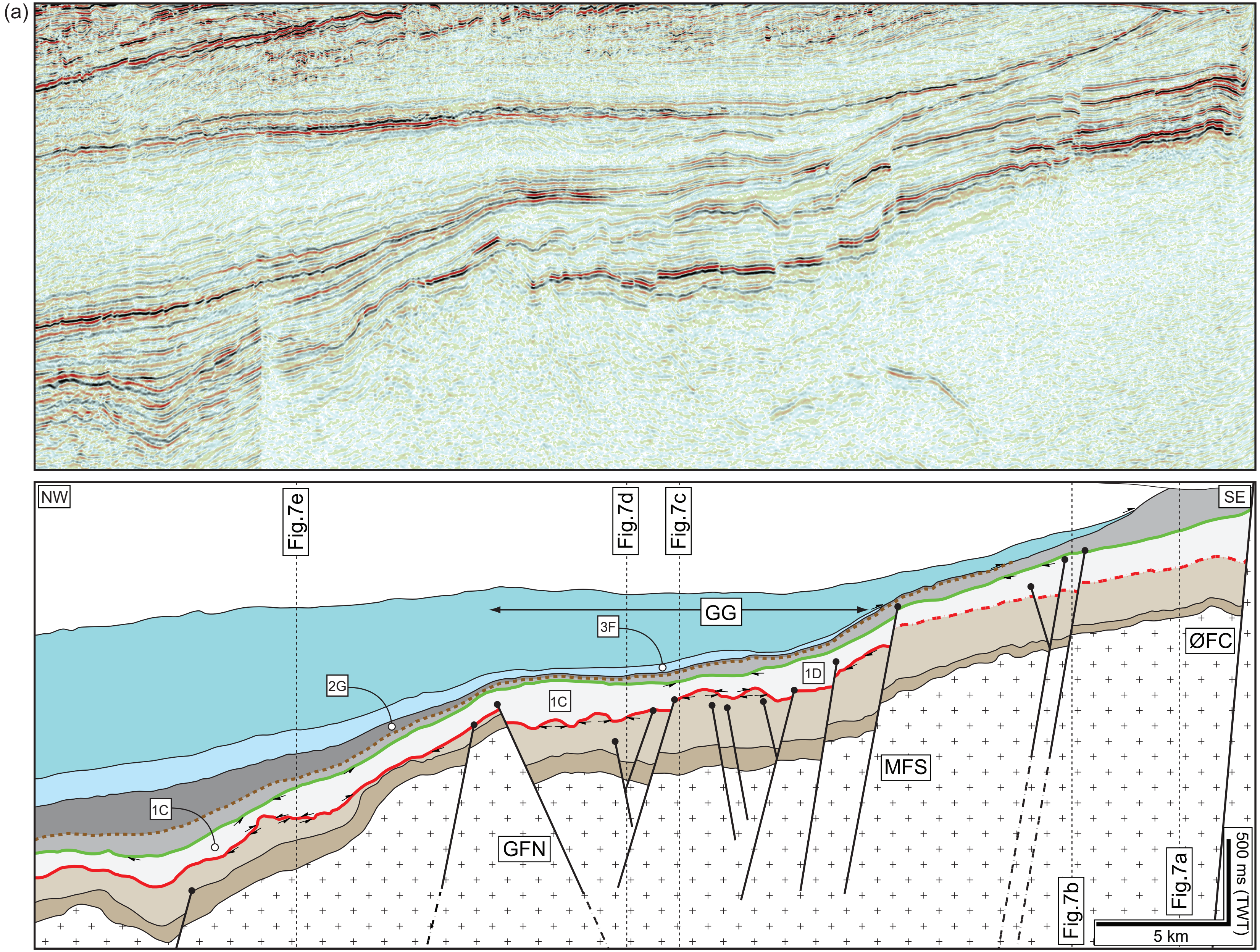

(a)
Fig. 4a
Fig. 4b
Fig. 7a
Fig. 7b
Fig. 7c
Fig. 7d
Fig. 7e
35/9-3
36/7-3
36/7-4
35/9-1
36/7-1
35/9-2
35/9-4
36/7-2
N
1.0
2.35
3.7
depth s (TWT)
5 km
top Basement time-structure map
(b)
GG
35/9-3
MFN
ØFC
GFN
Fig. 5b
36/7-3
36/7-4
Fig. 5a
35/9-1
Fig. 5c
GFS
36/7-1
35/9-2
35/9-4
MFS
36/7-2
ØFC
high
dip
low
N
5 km
top Basement dip map

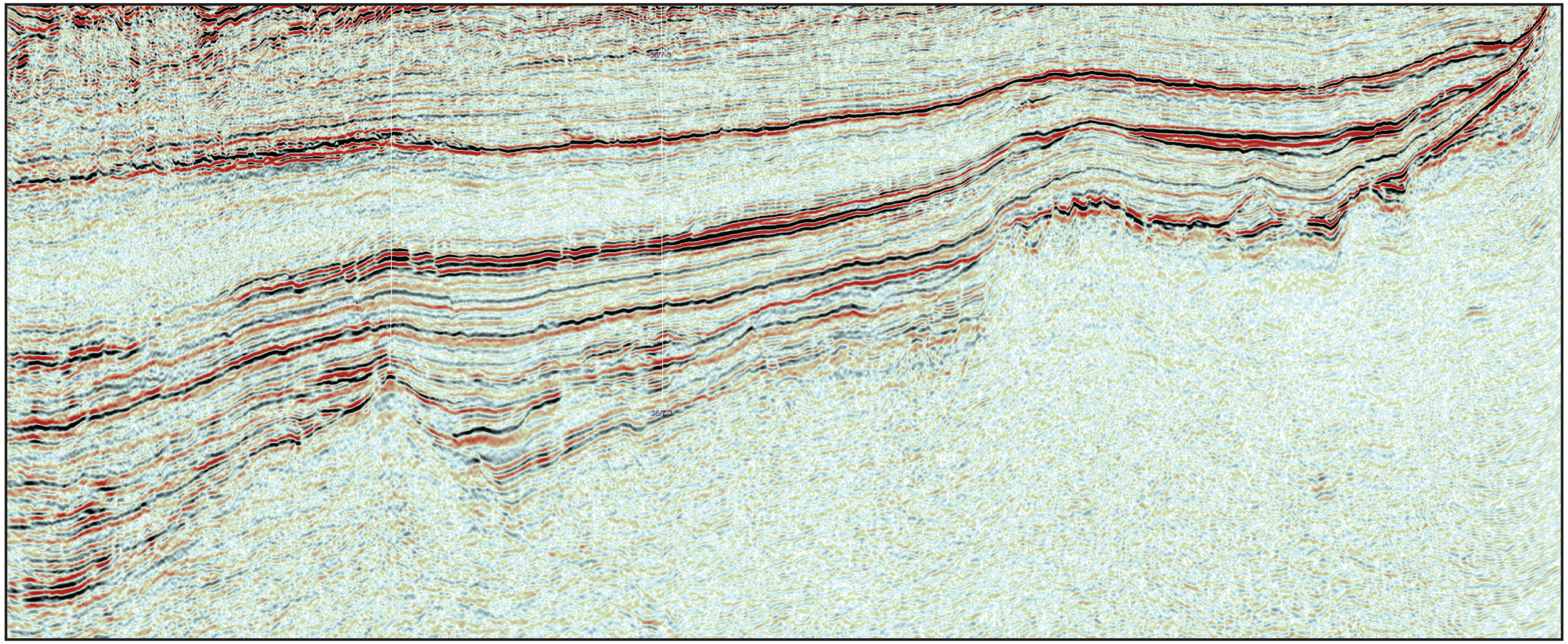
Fig. 4

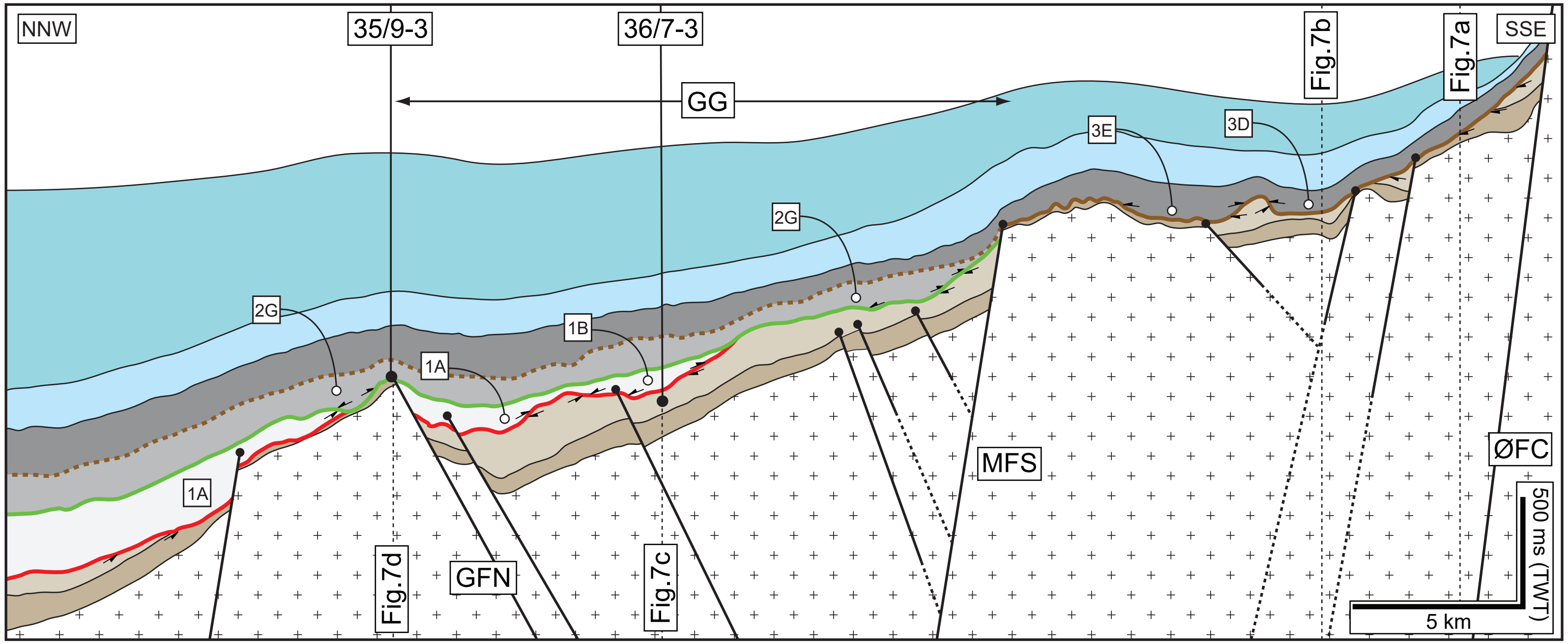


(a)
NW
SE
Fig.7e
Fig.7d
Fig.7c
Fig.7b
Fig.7a
GG
3F
2G
1C
1D
1C
GFN
MFS
ØFC
500 ms (TWT)
5 km

Fig. 4

(b)



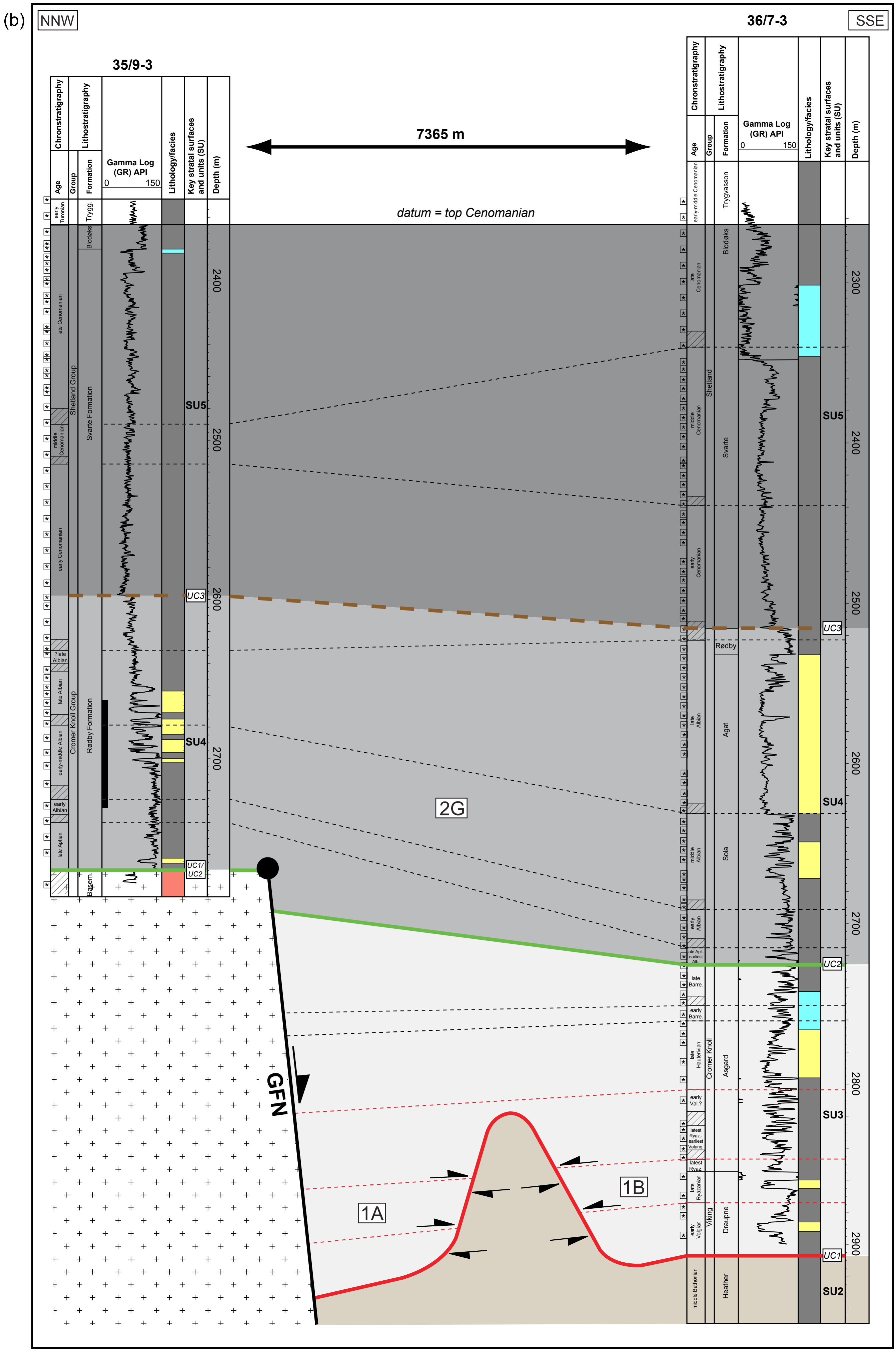


NNW
35/9-3
36/7-3
SSE
GG
Fig.7b
Fig.7a
3E
3D
2G
2G
1B
1A
1A
MFS
ØFC
Fig.7d
GFN
Fig.7c
500 ms (TWT)
5 km

Fig. 5

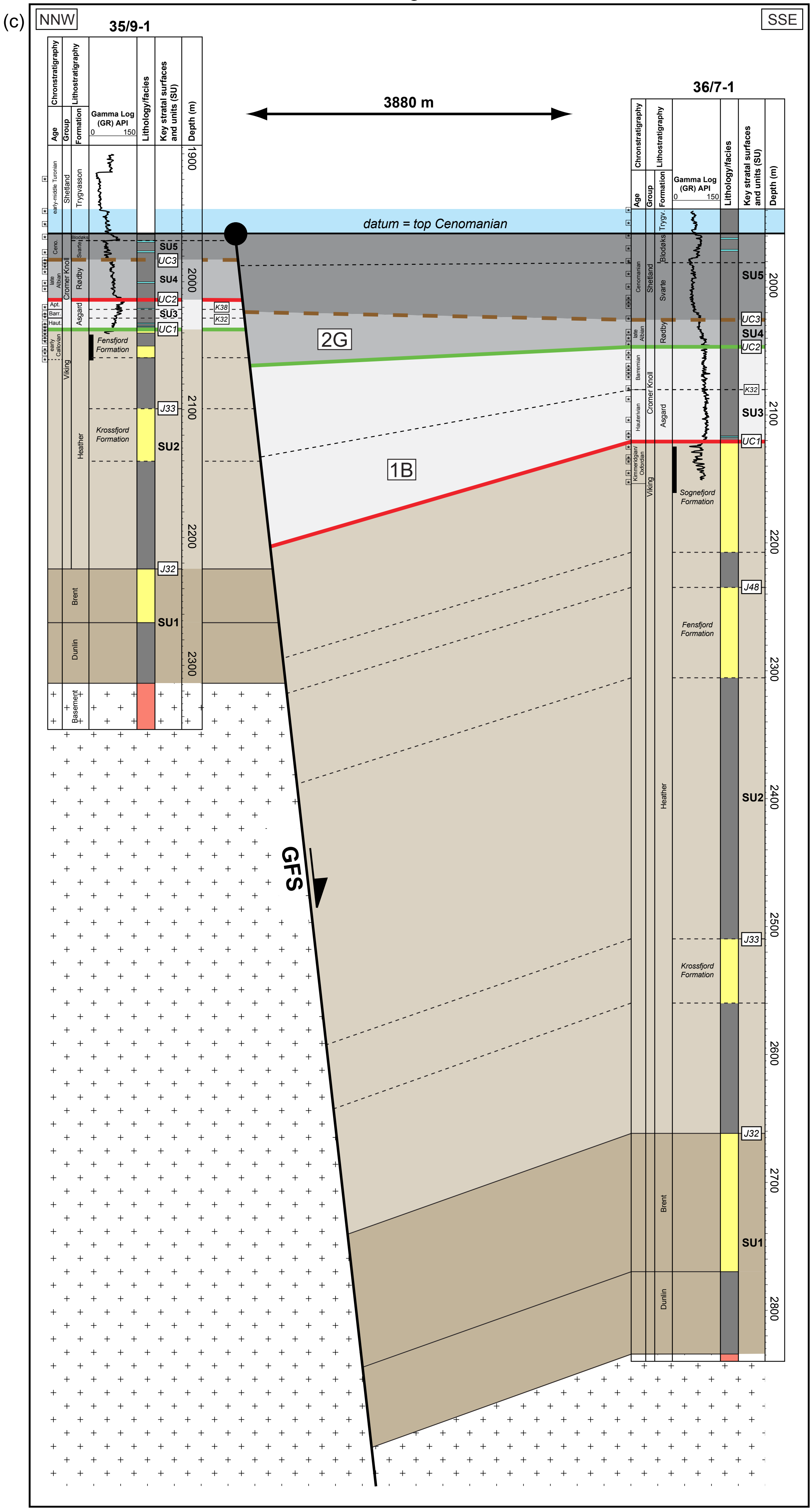
(a)
N-trending section in footwall of GF; perpendicular to canyons
NE-trending section across the southern tip of the southern segment of the GF
NNE-trending section across southern tip of the northern segment of the GF
NW-trending section from centre of the GG to the eastern basin margin
35/9-3
35/9-1
35/9-2
36/7-1
36/7-3
36/7-4
36/7-2
N
S
1069 m
NNE
SSW
6112 m
SW
NE
4616 m
SSW
NNE
8214 m
NW
SE
c. 1500 m
NW
SE
19977 m
datum = top Cenomanian
top Svarte Fm
top Cromer Knoll Gp/Rodby Fm
top Agat Mbr
top Sola Fm
top Asgard
Cretaceous-filled canyon
1A+2G
2
1C
Jurassic-filled canyon
1C
Jurassic-filled canyon
Jurassic/ Cretaceous-filled canyon
2G
1B
UC3 highly erosional east of the Gjøa Graben and dies out downslope to the west
UJUNC not strongly erosion in SE
200 m
GFS
GFN
MFS
ØFC
Gjøa Graben

Fig. 5



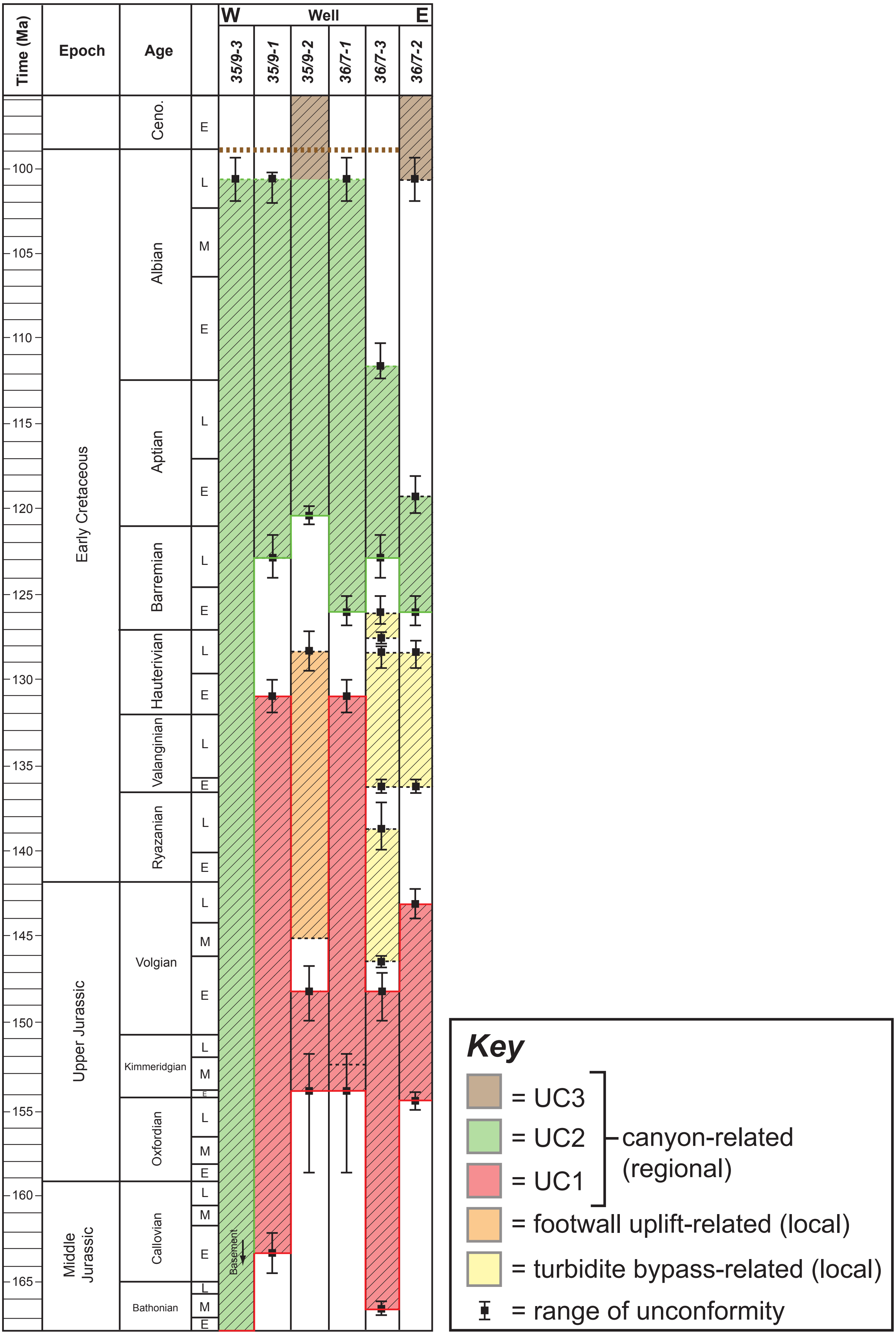
(b)
NNW
SSE
35/9-3
36/7-3
7365 m
datum = top Cenomanian
Chronostratigraphy
Lithostratigraphy
Age
Group
Formation
Gamma Log (GR) API
0
150
Lithology/facies
Key stratal surfaces and units (SU)
Depth (m)
Shetland Group
Svarte Formation
Cromer Knoll Group
Rødby Formation
Trygg.
Blodøks
Basem.
late Cenomanian
middle Cenomanian
early Cenomanian
?late Albian
late Albian
early-middle Albian
early Albian
late Aptian
early Turonian
SU5
SU4
UC3
UC1/ UC2
2400
2500
2600
2700
2G
GFN
1A
1B
Trygvasson
Blodøks
Svarte
Rødby
Agat
Sola
Asgard
Draupne
Heather
Shetland
Cromer Knoll
Viking
early-middle Cenomanian
late Cenomanian
middle Cenomanian
early Cenomanian
late Albian
middle Albian
early Albian
late Apt.-earliest Alb.
late Barre.
early Barre.
late Hauterivian
early Val.?
latest Ryaz.-earliest Valang.
latest Ryaz.
late Ryazanian
early Volgian
middle Bathonian
UC2
UC1
SU3
SU2
2300
2800
2900

Fig. 5



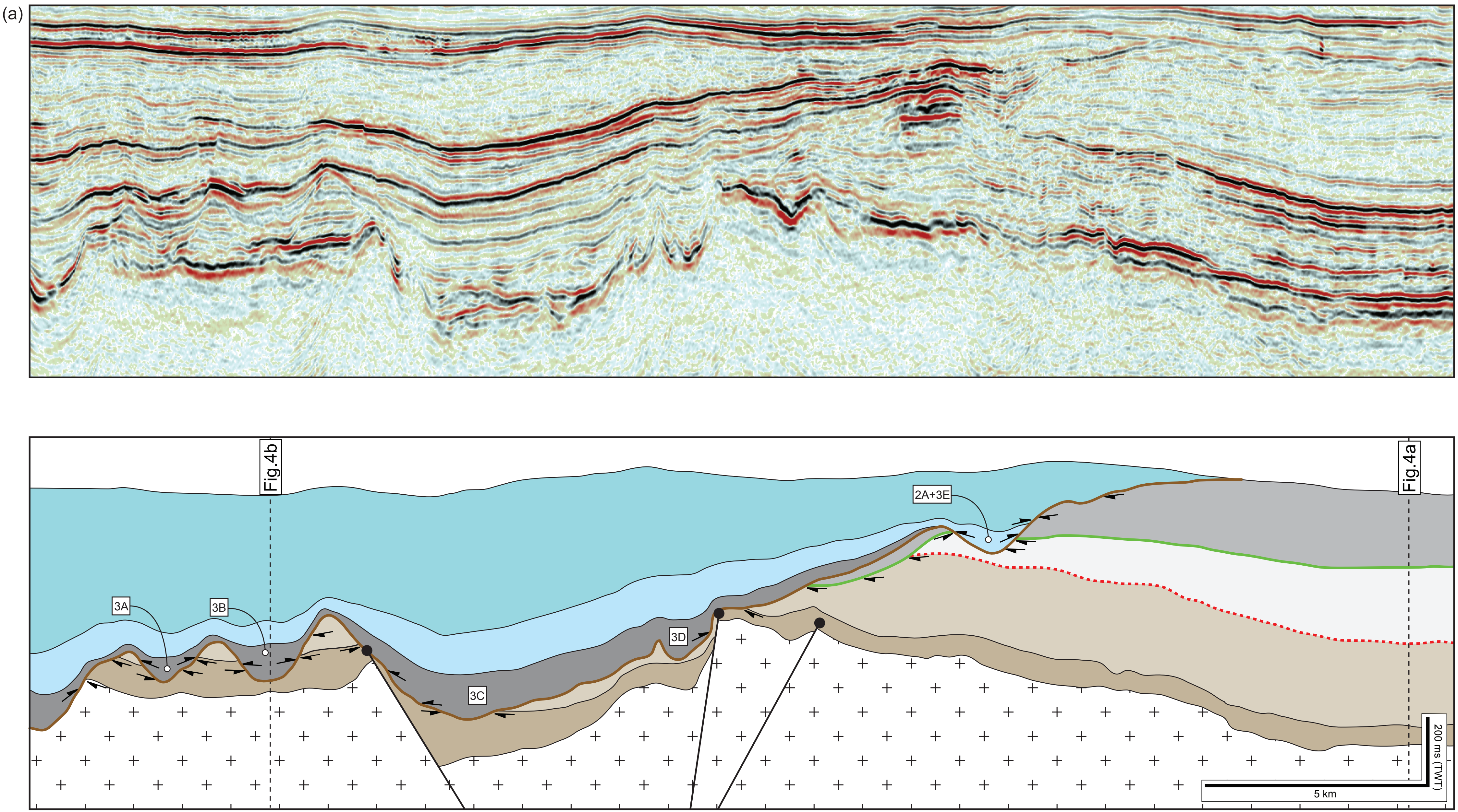
(c)
NNW
SSE
35/9-1
36/7-1
3880 m
datum = top Cenomanian
2G
1B
GFS
Chronostratigraphy
Lithostratigraphy
Age
Group
Formation
Gamma Log (GR) API
0
150
Lithology/facies
Key stratal surfaces and units (SU)
Depth (m)
1900
2000
2100
2200
2300
2400
2500
2600
2700
2800
SU5
UC3
SU4
UC2
SU3
UC1
J33
SU2
J32
SU1
J48
K38
K32
Shetland
Trygvasson
Cromer Knoll
Rødby
Asgard
Svarte
Blodøks
Viking
Heather
Brent
Dunlin
Basement
Fensfjord Formation
Krossfjord Formation
Sognefjord Formation
Cenomanian
late Albian
Barremian
Hauterivian
Kimmeridgian/ Oxfordian

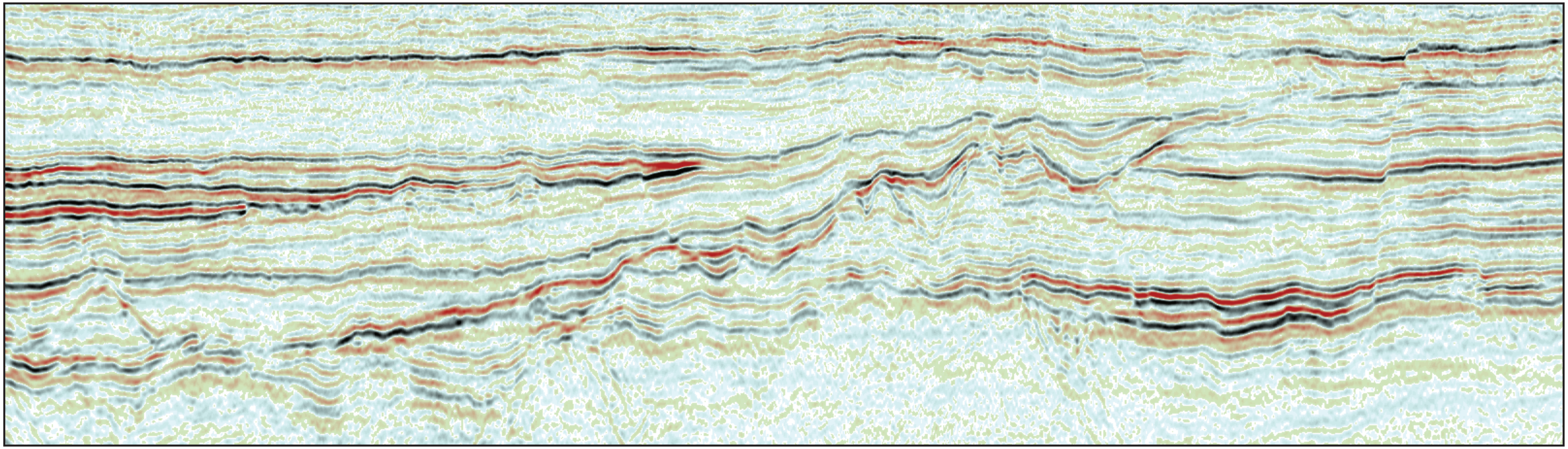
Fig. 6



Time (Ma)
Epoch
Age
W
Well
E
35/9-3
35/9-1
35/9-2
36/7-1
36/7-3
36/7-2
Ceno.
Albian
Aptian
Barremian
Hauterivian
Valanginian
Ryazanian
Volgian
Kimmeridgian
Oxfordian
Callovian
Bathonian
Early Cretaceous
Upper Jurassic
Middle Jurassic
100
105
110
115
120
125
130
135
140
145
150
155
160
165
Basement
Key
= UC3
= UC2
= UC1
canyon-related (regional)
= footwall uplift-related (local)
= turbidite bypass-related (local)
= range of unconformity

Fig. 7





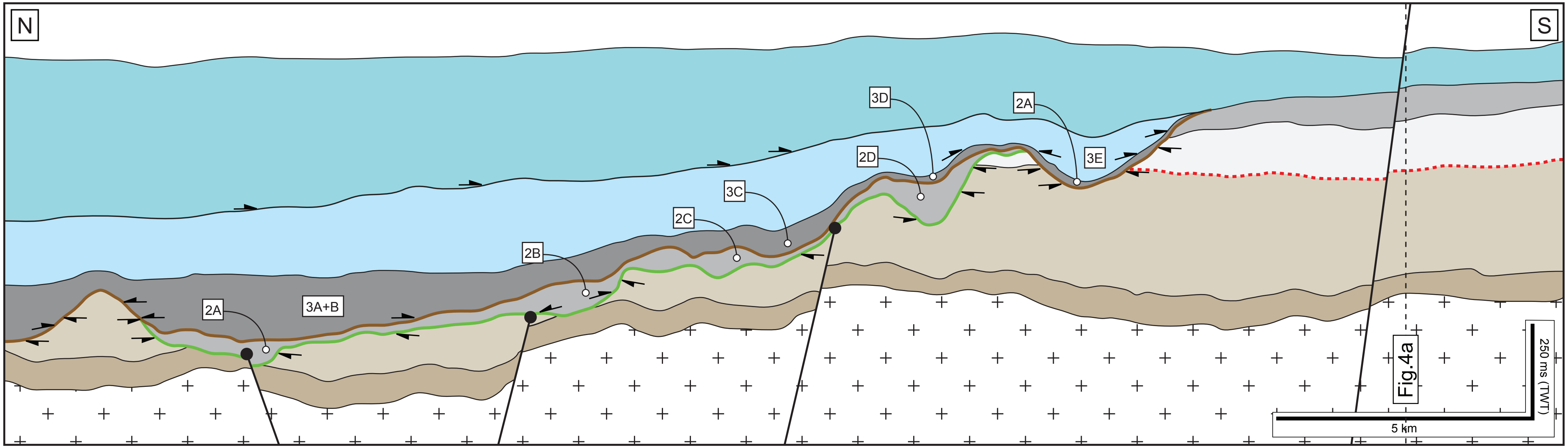

N
S
3D
2A
2D
3E
3C
2C
2B
2A
3A+B
Fig.4a
250 ms (TWT)
5 km

Fig. 7

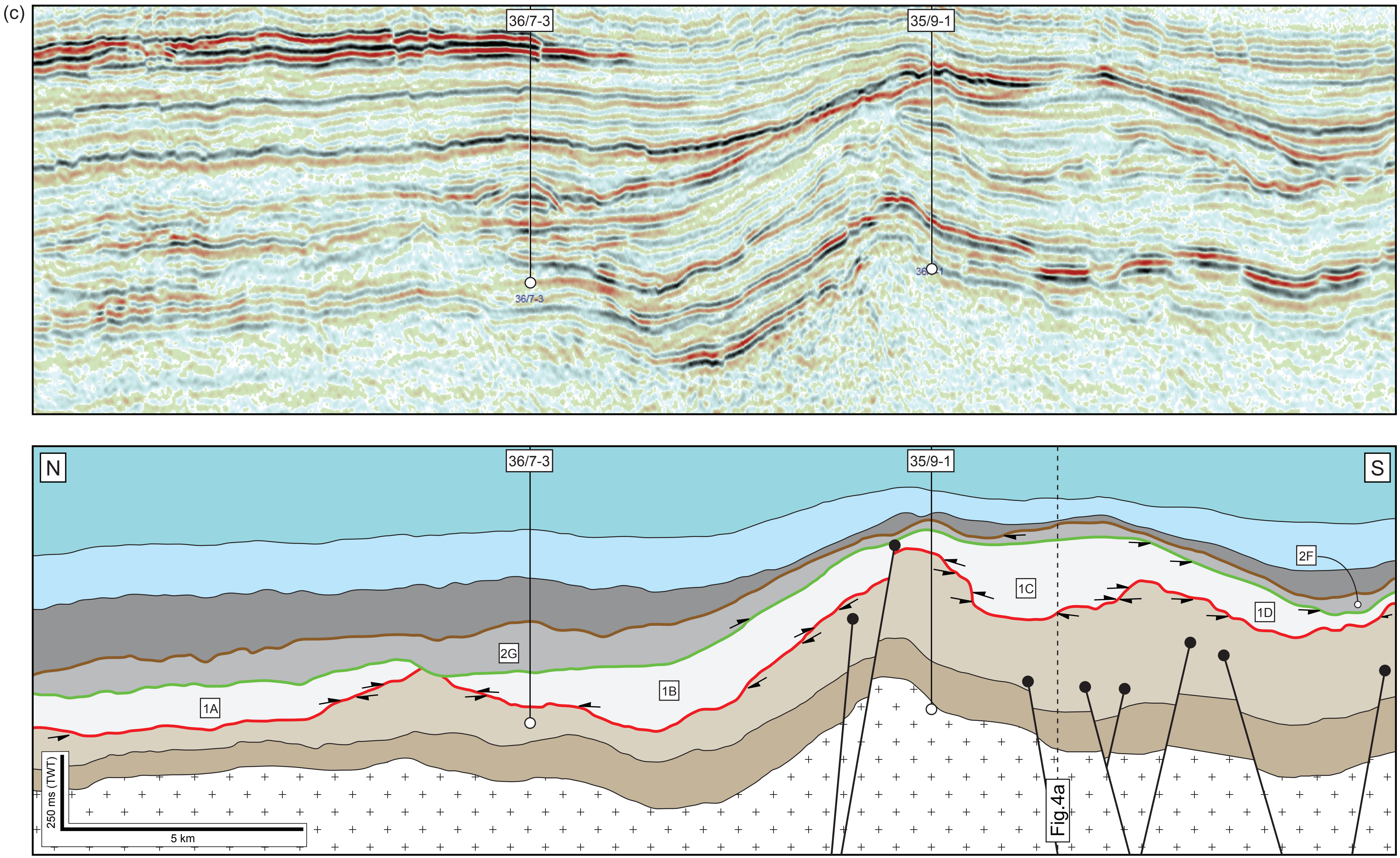
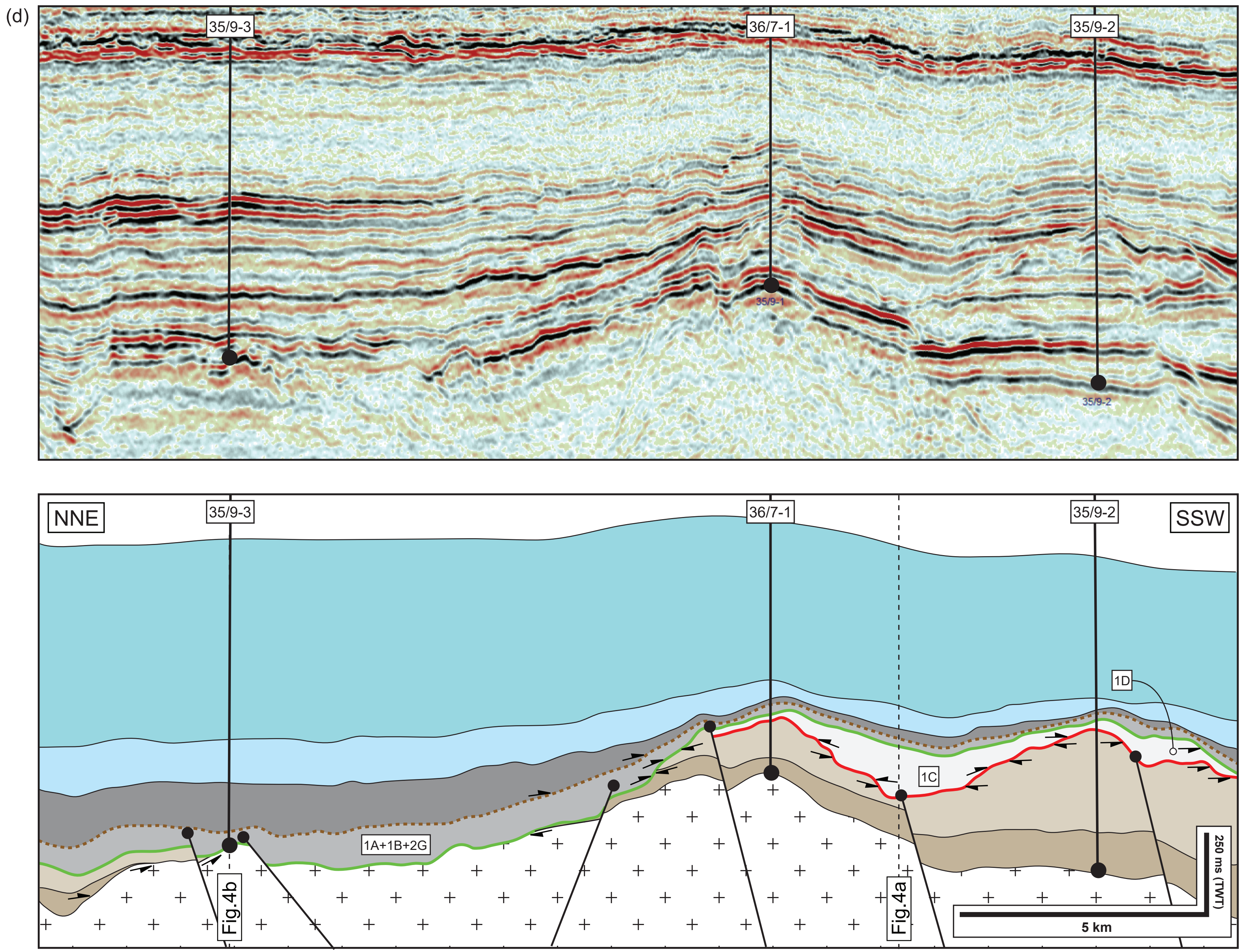


Fig. 7



(d)
35/9-3
36/7-1
35/9-2
35/9-1
35/9-2
NNE
SSW
35/9-3
36/7-1
35/9-2
1D
1C
1A+1B+2G
Fig.4b
Fig.4a
250 ms (TWT)
5 km

Fig. 7

(e)

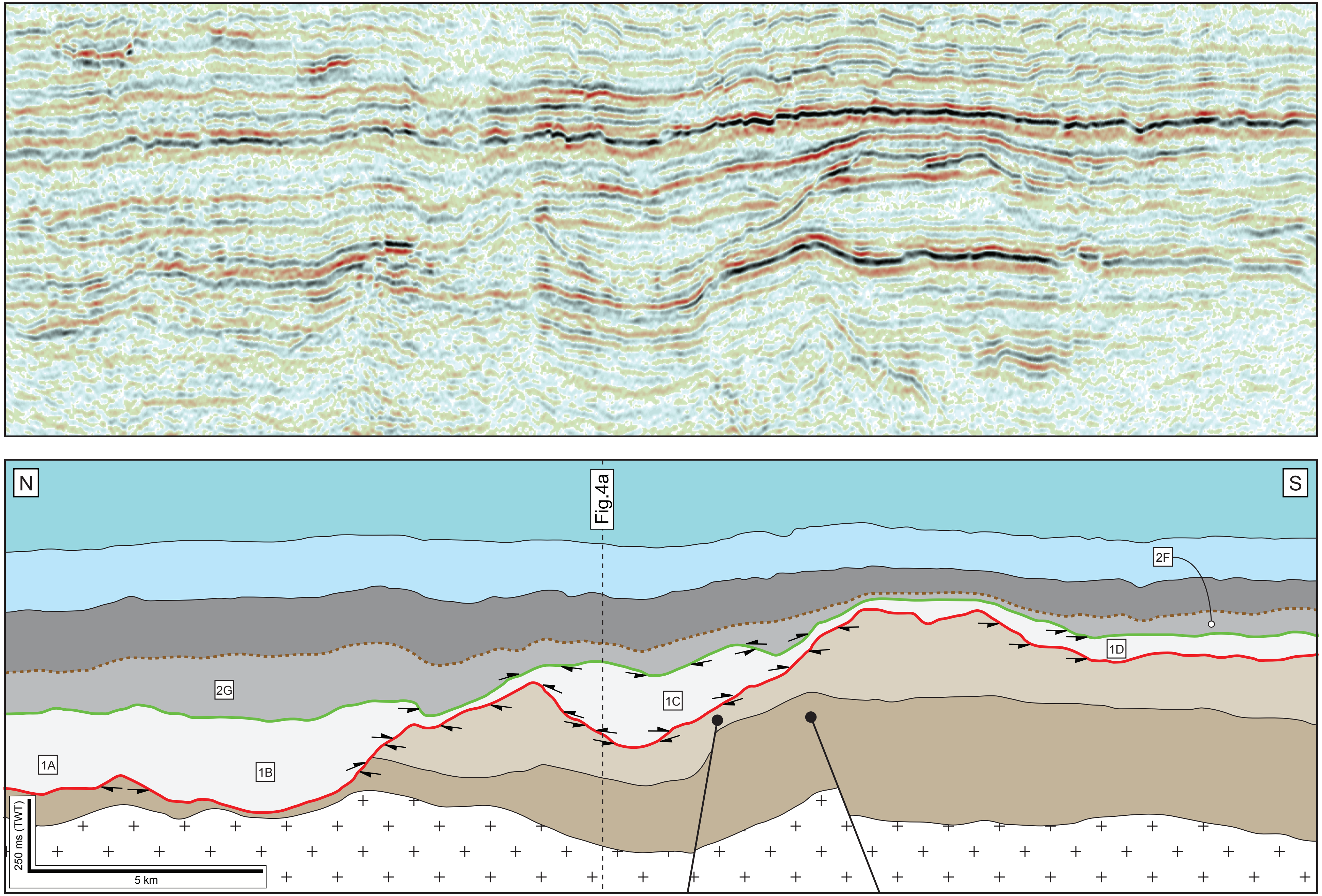
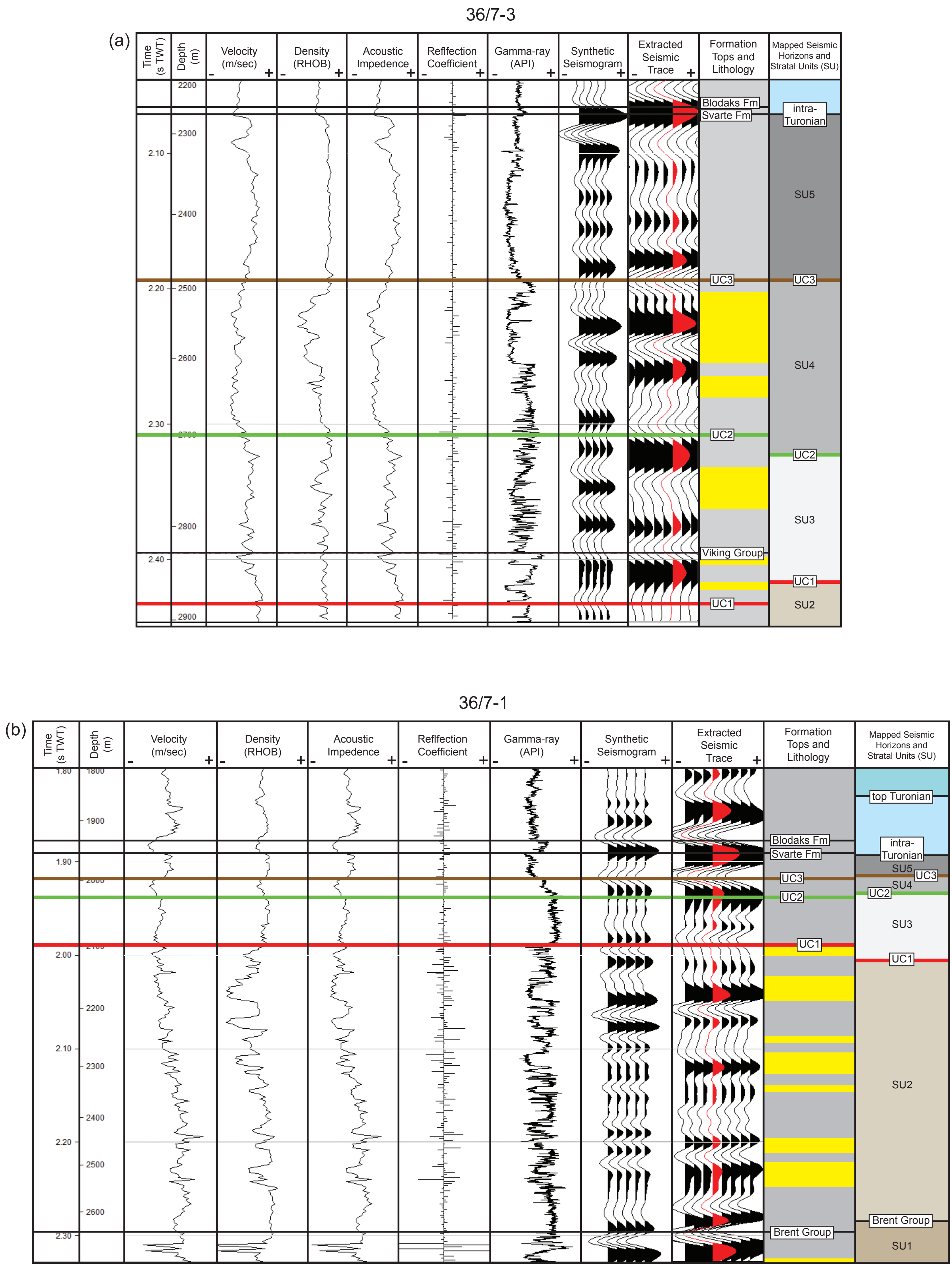


Fig. 8



36/7-3
(a)
Time (s TWT)
Depth (m)
Velocity (m/sec)
Density (RHOB)
Acoustic Impedence
Refflection Coefficient
Gamma-ray (API)
Synthetic Seismogram
Extracted Seismic Trace
Formation Tops and Lithology
Mapped Seismic Horizons and Stratal Units (SU)
Blodaks Fm
Svarte Fm
intra-Turonian
SU5
UC3
SU4
UC2
SU3
Viking Group
UC1
SU2
36/7-1
(b)
top Turonian
SU1
Brent Group

Fig. 9

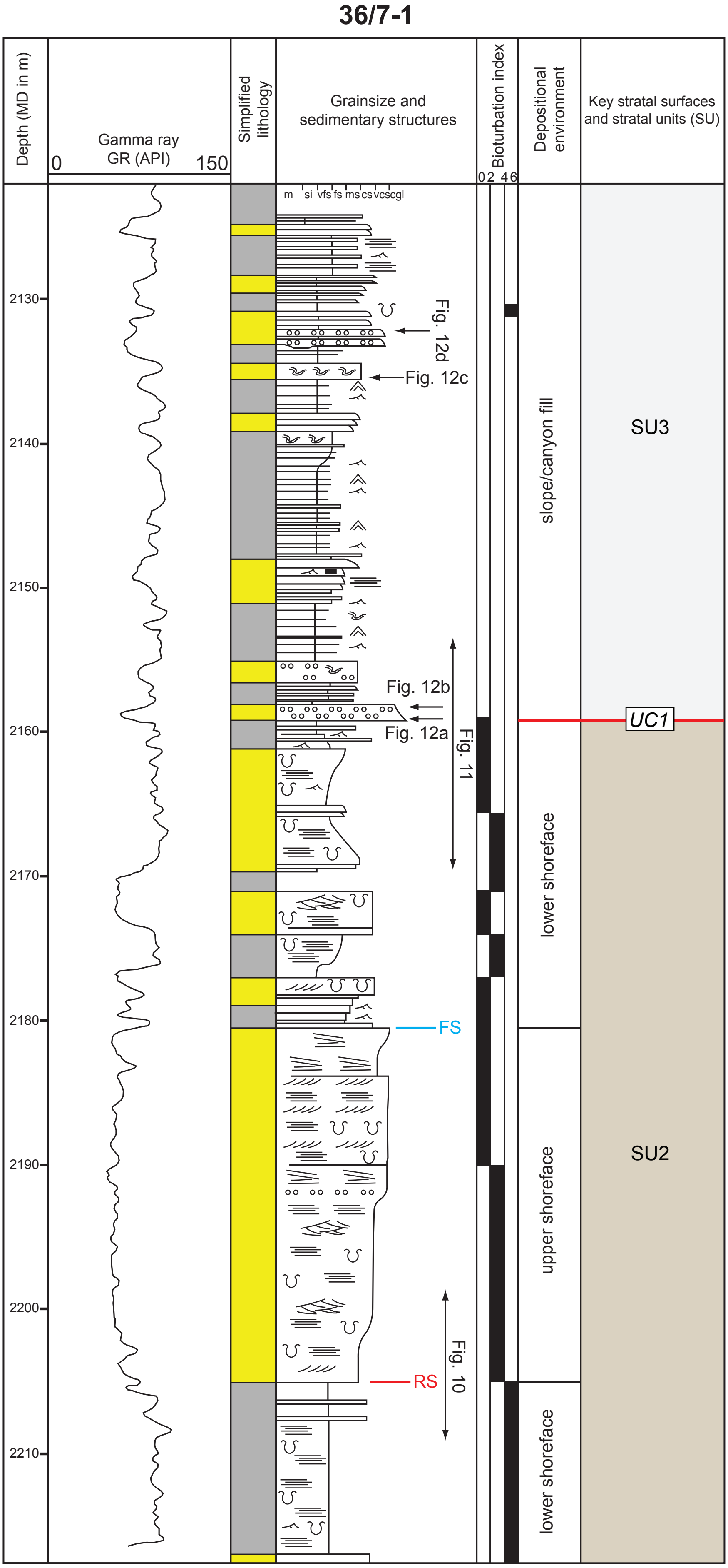

36/7-1
Depth (MD in m)
Gamma ray
GR (API)
0
150
Simplified lithology
Grainsize and sedimentary structures
m si vfs fs ms cs vcs cgl
Bioturbation index
0 2 4 6
Depositional environment
Key stratal surfaces and stratal units (SU)
2130
2140
2150
2160
2170
2180
2190
2200
2210
Fig. 12d
Fig. 12c
Fig. 12b
Fig. 12a
Fig. 11
FS
RS
Fig. 10
slope/canyon fill
lower shoreface
upper shoreface
lower shoreface
SU3
UC1
SU2

Fig. 10


NORSK HYDRO WELL: 36/7-1 CORE NO. 4
1 OF 28
TOP 2199.00
BTM 2200.00
2 OF 28
TOP 2200.00
BTM 2201.00
3 OF 28
TOP 2201.00
BTM 2202.00
4 OF 28
TOP 2202.00
BTM 2203.00
5 OF 28
TOP 2203.00
BTM 2204.00
NORSK HYDRO WELL: 36/7-1 CORE NO. 4
6 OF 28
TOP 2204.00
BTM 2205.00
7 OF 28
TOP 2205.00
BTM 2206.00
8 OF 28
TOP 2206.00
BTM 2207.00
9 OF 28
TOP 2207.00
BTM 2208.00
10 OF 28
TOP 2208.00
BTM 2209.00
RS

Fig. 11

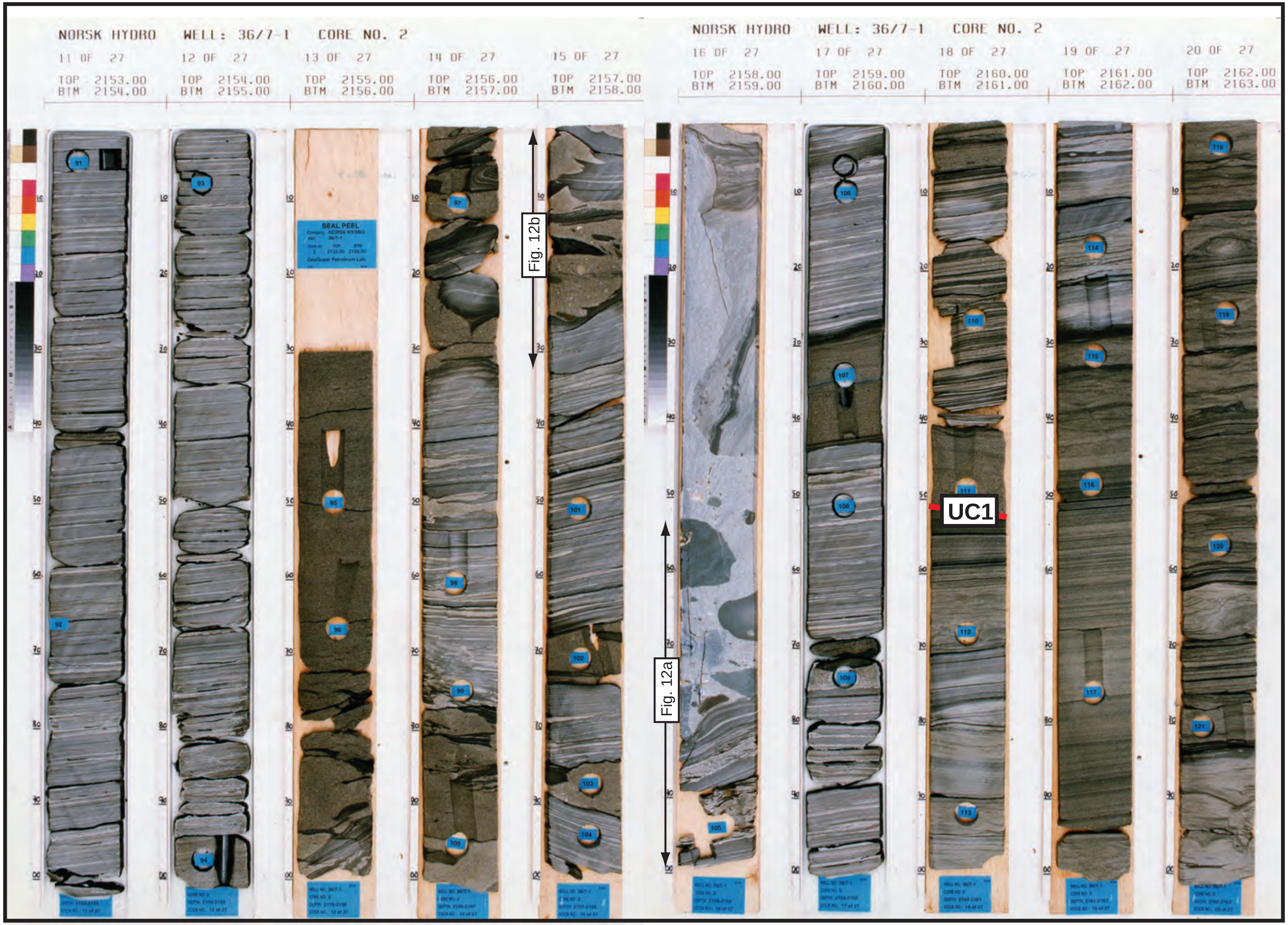

NORSK HYDRO WELL: 36/7-1 CORE NO. 2
11 OF 27 TOP 2153.00 BTM 2154.00
12 OF 27 TOP 2154.00 BTM 2155.00
13 OF 27 TOP 2155.00 BTM 2156.00
14 OF 27 TOP 2156.00 BTM 2157.00
15 OF 27 TOP 2157.00 BTM 2158.00
NORSK HYDRO WELL: 36/7-1 CORE NO. 2
16 OF 27 TOP 2158.00 BTM 2159.00
17 OF 27 TOP 2159.00 BTM 2160.00
18 OF 27 TOP 2160.00 BTM 2161.00
19 OF 27 TOP 2161.00 BTM 2162.00
20 OF 27 TOP 2162.00 BTM 2163.00
SEAL PEEL
Fig. 12b
Fig. 12a
UC1

Fig. 12

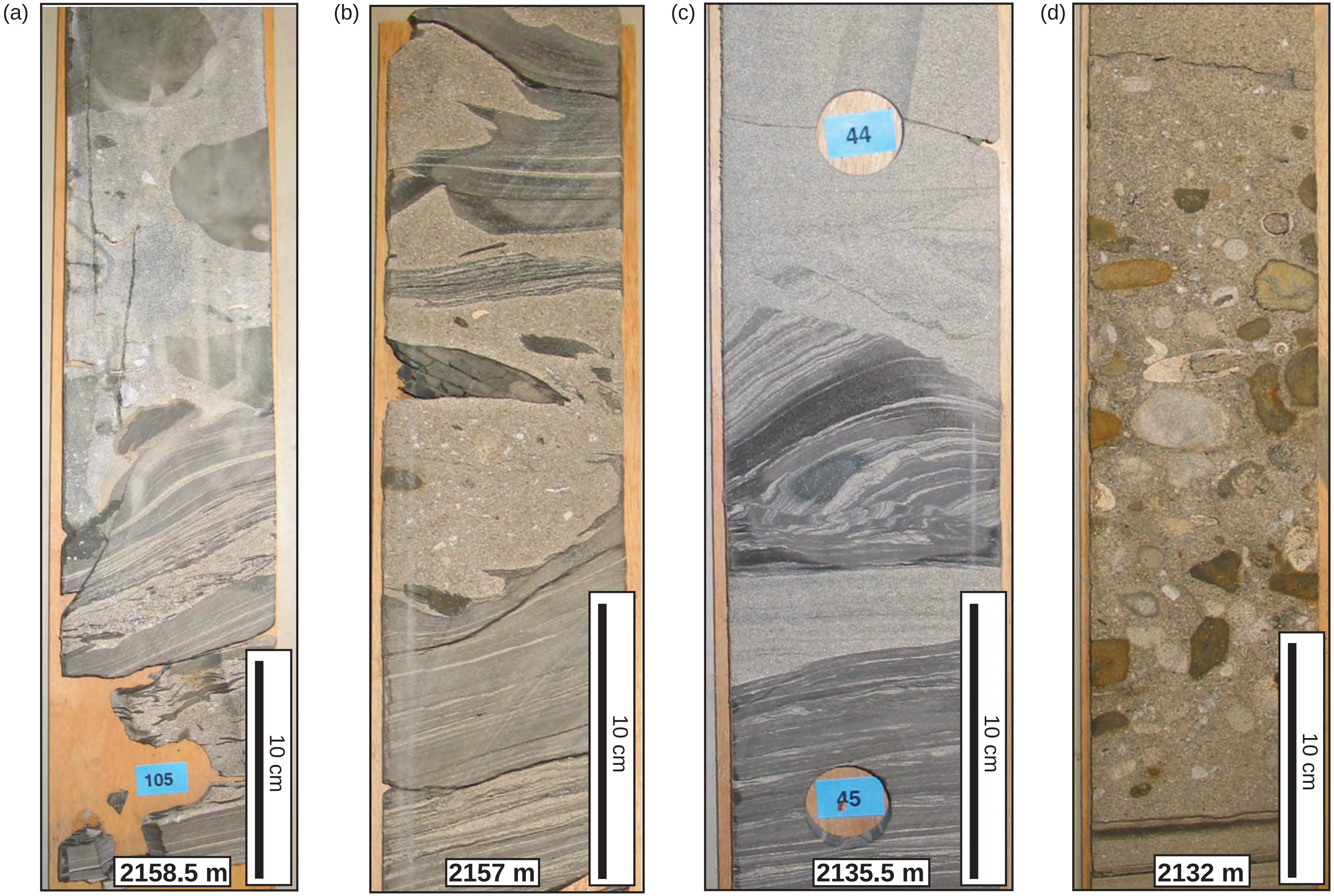


Fig. 13

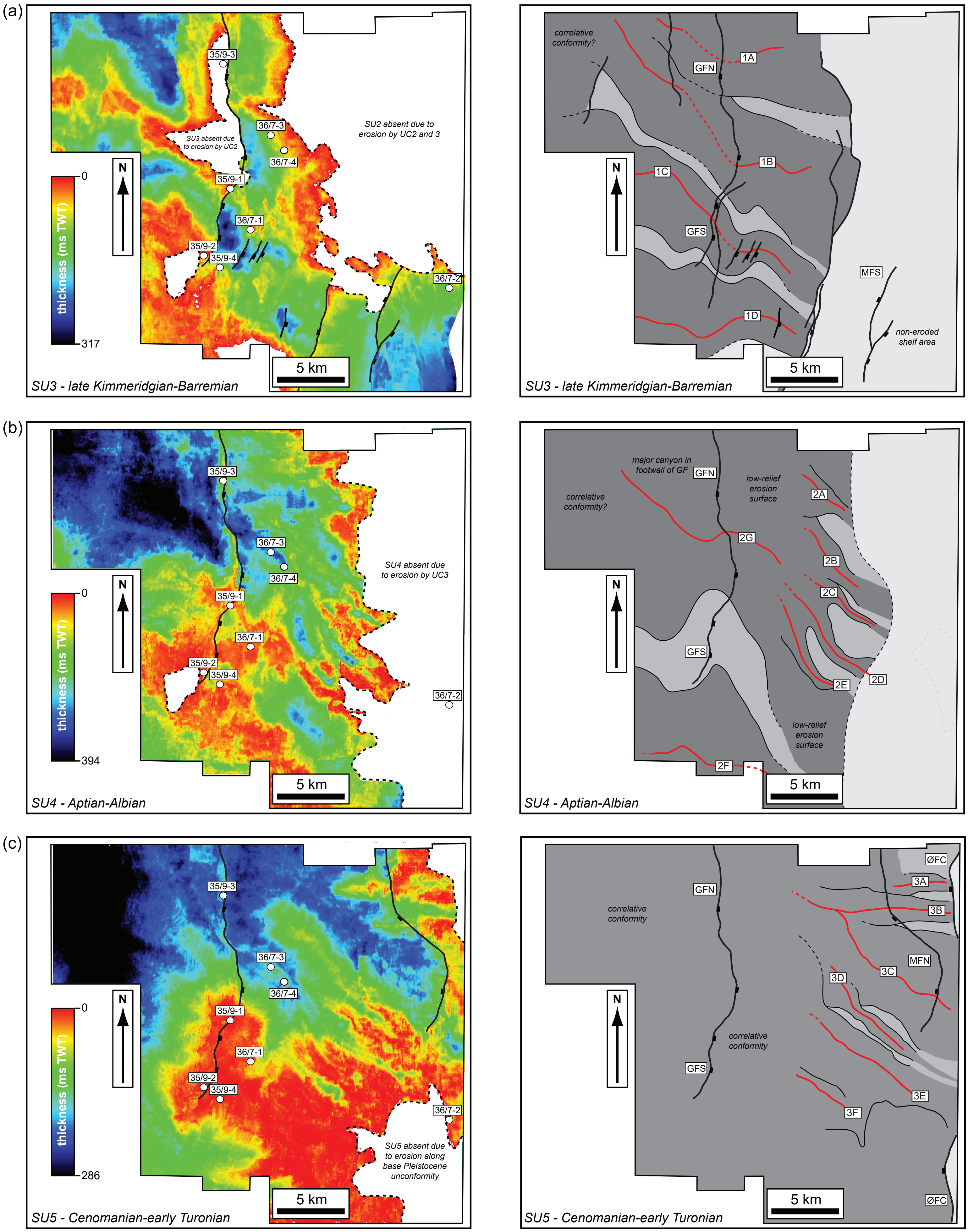
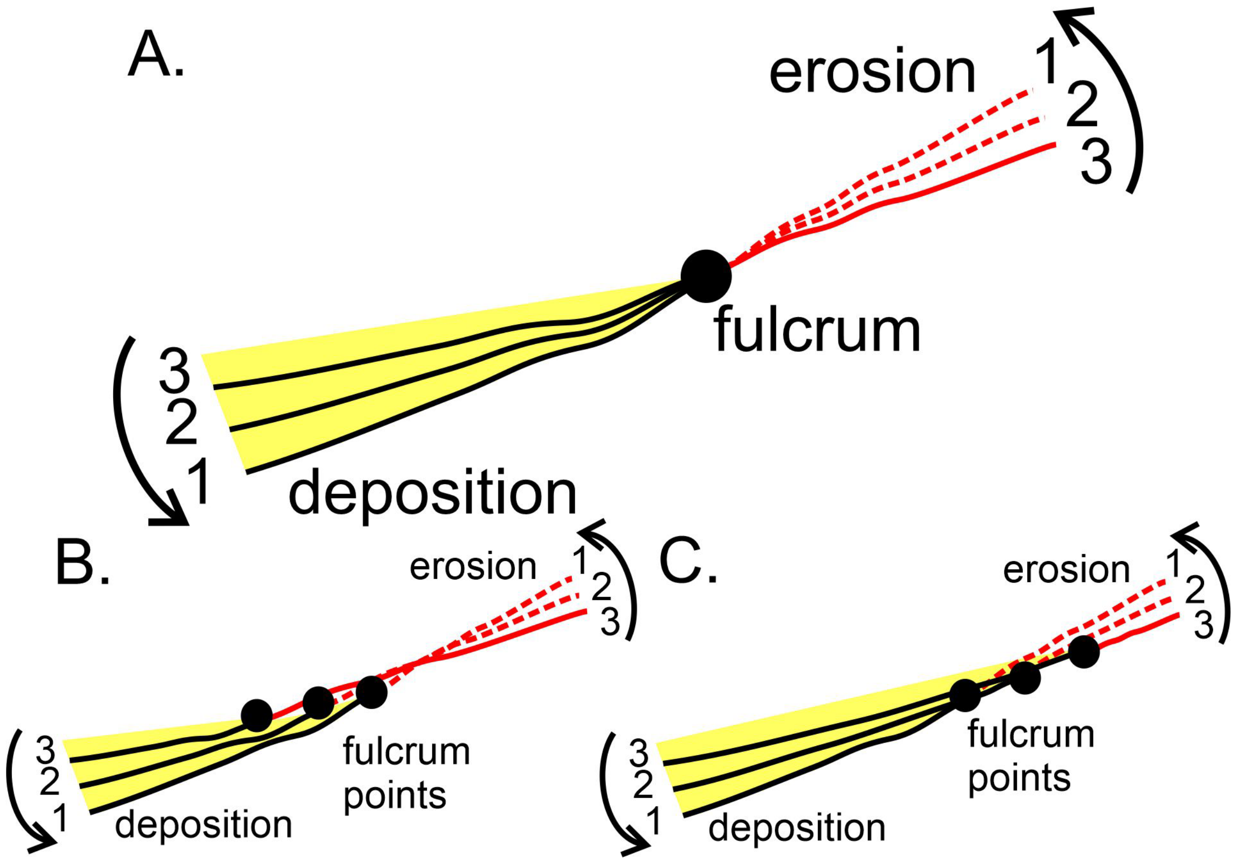

(a)
35/9-3
36/7-3
36/7-4
35/9-1
36/7-1
35/9-2
35/9-4
36/7-2
SU3 absent due to erosion by UC2
SU2 absent due to erosion by UC2 and 3
thickness (ms TWT)
0
317
N
5 km
SU3 - late Kimmeridgian-Barremian
correlative conformity?
GFN
1A
1B
1C
GFS
1D
MFS
non-eroded shelf area
SU3 - late Kimmeridgian-Barremian
(b)
SU4 absent due to erosion by UC3
0
394
SU4 - Aptian-Albian
major canyon in footwall of GF
low-relief erosion surface
correlative conformity?
2A
2G
2B
2C
2D
2E
2F
low-relief erosion surface
SU4 - Aptian-Albian
(c)
0
286
SU5 absent due to erosion along base Pleistocene unconformity
SU5 - Cenomanian-early Turonian
ØFC
3A
3B
MFN
3C
3D
3E
3F
correlative conformity
correlative conformity
SU5 - Cenomanian-early Turonian

Fig. 14


A.
erosion
1
2
3
fulcrum
3
2
1
deposition
B.
erosion
1
2
3
fulcrum
points
3
2
1
deposition
C.
erosion
1
2
3
fulcrum
points
3
2
1
deposition