The geometric and temporal evolution of fault-related folds constrains normal fault growth patterns, Barents Sea, offshore Norway

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Acknowledgement

The results presented here are part of Ahmed Alghuraybi's PhD research at Imperial College London, funded by Saudi Aramco. We thank the Norwegian Petroleum Directorate (NPD) for making the seismic and well data publicly available on the Norwegian national data repository for petroleum data (NDR). The Fruholmen 3D seismic reflection dataset can be downloaded from the Diskos NDR (https://portal.diskos.cgg.com/whereoil-data/) by searching for "Fruholmen 3D" and the well data can be found by searching for "7124/4-1 S". We thank Schlumberger for providing access to Petrel software and Geoteric for access to their software. We also thank Amir Joffe, Bailey Lathrop, and Sophie Pan from Imperial College's Basins Research Group (BRG) for their input and helpful discussions throughout this research. The authors extend their thanks to Hongdan Deng, John Conneally and Giovanni Camanni for their constructive reviews that helped improve the initial version of this paper. The journal editor-in-chief Atle Rotevatn and the entire

Basin Research editorial staff are thanked for editorial handling and assistance with paper production.

Data Availability Statement

The data that support the findings of this study are openly available in Norwegian national data repository for petroleum data at https://portal.diskos.cgg.com/whereoil-data/.

Abstract

Extensional growth folds form ahead of the tips of propagating normal faults. These folds can accommodate a considerable amount of extensional strain and they may control rift geometry. Fold-related surface deformation may also control the sedimentary evolution of syn-rift depositional systems. Thus, by examining the stratigraphic record, we can constrain the four-dimensional evolution of extensional growth folds, which in turn provides a record of fault growth and broader rift history. Here we use high-quality 3D seismic reflection and borehole data from the SW Barents Sea, offshore northern Norway to determine the geometric and temporal evolution of extensional growth folds associated with a large, long-lived, basement-rooted fault. We show that the fault grew via the linkage of four segments, and that fault growth was associated with the formation of fault-parallel and fault-perpendicular folds that accommodated a substantial portion (10 – 40%) of the total extensional strain. Several periods of fault-propagation folding occurred in response to periodic burial of the fault, with individual folding events (c. 25 Myr and 32 Myr) lasting a considered part of the c. 130 Myr rift period. Our study supports previous suggestions that continuous (i.e., folding) as well as discontinuous (i.e.,

faulting) deformation must be explicitly considered when assessing total strain in extensional setting. We also show that changes in the architecture of growth strata record alternating periods of folding and faulting, and that the margins of rift-related depocentres may be characterised by basinward-dipping monoclines as opposed to fault-bound scarps. Our findings have broader implications for our understanding of the structural, physiographic, and tectonostratigraphic evolution of rift basins.

Key words: Normal Fault Growth, Extensional Growth Folds, Continuous Strain, Strike-Projections , Barents Sea

Introduction

Basin-bounding normal fault systems have a complex three-dimensional geometry, consisting of variably linked or unlinked segments that branch along both the strike and dip directions (e.g. Walsh et al., 1999, 2002, 2003; Childs et al., 2002; van der Zee & Urai, 2005; Schöpfer et al., 2006; Long & Imber, 2010; Giba et al., 2012; Jackson & Rotevatn, 2013; Fossen & Rotevatn, 2016; Freitag et al., 2017; Camanni et al., 2019; Deng & McClay, 2021; Roche et al. 2021). These geometries reflect the fact that these systems evolve via the growth, interaction, and linkage of smaller segments (e.g. Jackson, 1987; Schlische, 1992; Morley, 1999; Gawthorpe & Leeder, 2000; McLeod et al., 2000; Peacock, 2002). Therefore, how faults grow has significant implications for the structural, physiographic, and tectonostratigraphic evolution of rift basins, as well as their potential for hosting energy resources and for sequestering CO_2 (e.g. Gawthorpe & Leeder, 2000; Childs et al., 2017; Jackson et al., 2017; Michie et al., 2021).

Fault growth patterns also control the development of growth folds, which are also known as fault-propagation folds. These structures form ahead of the propagating normal fault tips and typically form during the early phases of extension, during which time they may accommodate a considerable amount of extensional strain (e.g. Withjack et al., 1990; Allmendinger, 1998; Cosgrove & Ameen, 1999; Hardy & McClay, 1999; Gawthorpe & Leeder, 2000; Withjack & Callaway, 2000; Coleman et al., 2019; Jackson et al., 2020). fault-propagation folds play a major role in controlling rift geometry and the sedimentary evolution of syn-rift systems as they alter sediment-transport pathways and may act as sediment sources if eroded (e.g. Laubscher, 1982; Maurin & Niviere, 1999; Gawthorpe & Leeder, 2000; Sharp et al., 2000; Lewis et al., 2015; Jackson & Lewis, 2016; Coleman et al., 2019). Geometrically complex hanging wall fold and fault geometries can develop within a single regional extensional stress regime (e.g. Khalil & McClay, 2018), although multiphase extension, stratigraphic and rheological variability in the host rock, and temporal and spatial changes in regional stress regime can all result in additional, four-dimensionally complexity (e.g. Schöpfer et al., 2007; Conneally et al., 2017; Deng & McClay, 2019; Jackson et al., 2020). For example, along-strike variations in displacement along segmented normal faults may mean that breached and un-breached growth folds are spatially and temporally related (e.g. Gawthorpe et al., 1997; Gupta et al., 1999; Corfield & Sharp, 2000; Sharp et al., 2000; Corfield et al., 2001; Khalil & McClay, 2002; Willsey et al., 2002; White & Crider, 2006; Lewis et al., 2013; Lewis et al., 2015; Khalil & McClay, 2017; Coleman et al., 2019; Jackson et al., 2020). These structural dynamics are often recorded in the stratigraphic architecture of and facies distributions within fault- and fold-bound hanging wall depocenters (Gawthorpe et al., 1997; Corfield & Sharp, 2000; Sharp et al., 2000; Kane et al., 2010; Duffy et al., 2013; Coleman et al., 2019). Fault-propagation folds also vary in shape and size across the fault surface as a function of, for example, the depth at which faults nucleate, changes in host rock lithology and rheology, dip linkage between initially isolated segments, and spatial changes in fault dip (e.g. Mansfield & Cartwright, 2000; Rykkelid & Fossen, 2002). Despite containing a record of fault growth and broader rift history, few studies have attempted to analyse fault-propagation folds in four dimensions using borehole and 3D seismic reflection data (e.g. Gawthorpe et al., 1997; Corfield et al., 2001; Patton, 2004; Lewis et al., 2015; Coleman et al., 2019; McHarg et al., 2020; Long & Imber, 2010; Conneally et al., 2017; Deng & McClay, 2019). Such data are optimal for this purpose, given they reveal the present basin structure and stratigraphic architecture, and allow inferences to be made regarding the underlying kinematics.

In this study we use high-quality 3D seismic reflection and borehole data from the SW Barents Sea, offshore northern Norway to provide a detailed analysis of the geometric and temporal evolution of fault-propagation folding associated with part of a large, basement-rooted fault that is c. 30 km long, c. 5 km tall , and that has up to 2 km of displacement. This fault forms part of the Troms-Finnmark Fault Complex (TFFC), a basement-rooted, rift-related structure that accommodated several phases of Palaeozoic to Cenozoic extension (e.g. Gabrielsen, 1984; Indrevær et al., 2013). Combining both 3D seismic reflection and borehole data enables us to study the temporal evolution of this fault system using age-constrained synkinematic stratigraphy. By doing this we show how large, crustal-scale faults, and their associated folds, evolve and control rift structure and sedimentation over timescales of c. 100 Myr.

Geological Setting

Regional Tectonics

The Barents Sea is a shallow continental shelf located in the northwest corner of the Eurasian tectonic plate, between the Arctic Ocean to the north and the Russian and Norwegian coastlines to the south (Gabrielsen, 1984) (Fig. 1). The SW Barents Sea is presently a passive margin that consists of numerous predominately NNE-trending rift basins (e.g. Nordkapp Basin (NB), Hammerfest Basin (HB), and Sørvestsnaget Basin (SB); Fig. 1a) and basement highs (e.g. Fedynsky High (FH), Loppa High (LH), and Stappen High (SH); Fig. 1a) (Faleide et al., 1984, 1993, 2008). Multiple phases of rifting occurred in the Barents Sea, initiating in the Devonian in response to the collapse of the Caledonian orogenic belt, and concluding with the opening of the Norwegian and Greenland seas and the onset of seafloor spreading in the Eocene (Faleide et al., 1984, 1993, 2008; Gabrielsen, 1984; Gabrielsen et al., 2016). The late Cambrian to mid-Devonian Caledonian Orogeny deformed the crystalline basement rocks of the SW Barents Sea (Faleide et al., 1984). The large-scale fabric of these Caledonian igneous and metamorphic rocks played a major role in shaping the evolution and present structural framework of the area (see below; e.g., Faleide et al., 1984; Ritzmann & Faleide, 2007).

Seismic reflection and wide-angle refraction data, and potential field (e.g., gravity and magnetic) data indicate that at least four distinct rift phases controlled the long-term evolution and large-scale structure of the SW Barents Sea (Gabrielsen et al., 1984; Faleide et al., 1993).

. These rift phases occurred in the Late Devonian – Carboniferous , Late Permian, Middle Jurassic – Early Cretaceous, and Paleocene – Eocene (Faleide et al., 2008). Crustal extension in the Late Devonian – Carboniferous created N- to NE-trending half-grabens and resulted in the formation of the TFFC (e.g. Gabrielsen, 1984; Faleide et al., 2008; Indrevær et al., 2013). The TFFC is a basement-rooted normal fault system that represents a major structural element of the SW Barents Sea, separating recent shelf sediments offshore from onshore crystalline basement rocks (e.g. Gabrielsen, 1984; Indrevær et al., 2013). The TFFC runs parallel to the present-day coastline of Norway, striking NE-SW in its southernmost part and NW-SE in its northern part (Gabrielsen, 1984). Previous studies of the TFFC suggest that it was continually reactivated until the Eocene (Gabrielsen et al., 1984; Faleide et al., 1993).

In the latest Carboniferous to Permian and following the first phase of rifting, the newly formed narrow basins were filled with evaporites and carbonates, which were later covered by clastic rocks (Faleide et al., 1984). Clastic sedimentation continued during the second phase of rifting in the Late Permian, which created considerable accommodation (e.g. Johansen et al., 1993; Larssen et al., 2002). The Triassic was a period of moderate tectonic activity that saw high rates of subsidence and clastic sediment accumulation , which ended with a regional marine transgression (e.g. Glørstad-Clark et al., 2010; Harishidayat et al., 2015; Mattos et al., 2016).

During the Middle Jurassic – Early Cretaceous, a third, intense phase of rifting occurred in the SW Barents Sea, resulting in the formation of multiple large basins bounded by basement-cored structural highs (e.g. Gabrielsen, 1984; Doré, 1995; Mattos et al., 2016). This third phase of rifting was also marked by notable across-fault thickening of clastic (growth) strata (Faleide et al., 1993;

2008). The base of the Cenozoic succession is marked by a regional unconformity that separates Cretaceous and Paleogene strata (Faleide et al., 2008). The fourth and last reported rifting phase in the Palaeocene – Eocene is linked to the opening of the Norwegian and Greenland seas (e.g. Eldholm & Thiede, 1980; Faleide et al., 2008; Harishidayat et al., 2015). This phase of rifting coincided with increased magmatic activity, followed by a period of glaciation, uplift and erosion (e.g. Eldholm & Thiede, 1980; Faleide et al., 2008; Harishidayat et al., 2015).

Stratigraphy

The sediments filling the SW Barents Sea basins are dominated by clastic rocks overlying Caledonian crystalline basement (e.g. Doré, 1995; Harishidayat et al., 2015) (Fig. 2). The clastic-dominated succession can be divided into six megasequences (e.g. Glørstad-Clark et al., 2010). The 1st megasequence comprises Late Devonian-Early Carboniferous fluvial deposits at its base and marginal marine deposits towards its top (e.g. Larssen et al., 2002). The 2nd megasequence contains Middle Carboniferous-Late Permian clastics in its lower part and carbonates in its upper part, with evaporites present in basinal areas (e.g. Larssen et al., 2002). The very top of the 2nd megasequence is dominated by bioclastic limestones that pass upwards into cherts and siliceous limestones (e.g. Larssen et al., 2002). A subaerial unconformity separates the underlaying late Paleozoic carbonates from the overlaying Early Triassic-Middle Jurassic strata of the 3rd megasequence (e.g. Glørstad-Clark et al., 2010). This megasequence boundary marks a transition from carbonate-dominated deposition to siliciclastic-dominated sedimentation. The base of the 3rd megasequence comprises predominantly fine-grained shelf clastics, whereas the rest of the unit is dominated by shallow marine and coastal clastic rocks (e.g. Glørstad-Clark et al., 2010). This megasequence is unconformably overlain by a succession of Late Jurassic to Early Cretaceous marine mudstones and thin sandstones, which together form the 4th megasequence (e.g. Glørstad-Clark et al., 2010). The uppermost, Late Cretaceous part of the 4th megasequence is generally composed of mudstones with thin intervals of limestones and sandstones (Faleide et al., 1984, 1993, 2008; Gabrielsen, 1984; Gabrielsen et al., 2016). The 5th and 6th megasequences are Cenozoic and consist mainly of marine mudstones with minor siltstones and sandstones (Faleide et al., 1984,1993, 2008; Gabrielsen, 1984; Gabrielsen et al., 2016).

From the description above, it is clear that strong mechanical competency contrasts exist throughout the sedimentary succession in the SW Barents Sea, with relatively weak mudstone-rich intervals alternating with relative strong, carbonate- and sandstone-dominated intervals. This is especially clear between the 3rd and 4th megasequences, where the sandstone-rich upper part of the 3rd megasequence passes upwards into the mudstone-dominated base of the 4th megasequence (Fig. 2). These mechanical competency contrasts are important in terms of the formation and evolution of the normal faults and their associated folds that form the focus of our study.

Data

We used the Fruholmen 3D seismic reflection data, which was acquired and processed by WesternGeco in 2007 and retrieved from the DISKOS database in late 2019 (https://portal.diskos.cgg.com/whereoil-data/). It is a post-stack time migrated (PSTM) seismic reflection cube that covers an area of c. 533 km^2. The survey in-lines trend broadly ENE, whereas the cross-lines trend broadly NNW. The studied fault (i.e., the TFFC) changes strike within the study area (see below), so we use arbitrary seismic lines, locally oriented normal to strike, to conduct our quantitative analysis of fault structure and throw (see below). The total record length of this survey reaches 5.5 seconds two-way time (TWT). The data were acquired with a group interval of 12.5 meters (m), a shot point interval of 18.75 m, and a streamer length of 5000 m. The original sampling interval for this data was 2 milliseconds (ms), which was resampled during data processing (by WesternGeco) to 4 ms. The data was time migrated using Kirchhoff migration, and an amplitude gain was applied at the last stage of the data processing sequence. A full processing report can be accessed from the DISKOS database (https://portal.diskos.cgg.com/whereoil-data/) by searching for "Fruholmen 3D". The data have a dominant frequency of 40 Hz (see Fig. 1 in supplementary material for complete frequency spectrum) and a vertical resolution (over the depth range of interest; c. 1000 m – 3500 m) of c. 12.5 – 25 m, based on an average velocity of 2000 m/s – 4000 m/s, values which are derived from sonic log data. This depth interval covers the syn-kinematic stratigraphy and fault-related folds studied here.

In addition to the 3D seismic reflection data, one wellbore (7124/4-1 S) is available in the study area. The wellbore has basic lithostratigraphic data that was retrieved from the Norwegian Petroleum Directorate (https://factpages.npd.no/en/wellbore/pageview/exploration/all/6678# However, no checkshot surveys were available from the well to tie it to the seismic reflection data. Sonic log data (DT) were available for depths of c. 500 m to the well total depth (2730 m), but other logs, like the density log (RHOB), were only available for depths from c. 1200 m to c. 2700 m. We therefore used a modified seismic-well tie workflow to generate a reliable TDR and to establish an age-constrained seismic-stratigraphic framework for our specific study area (see supplementary materials for a detailed description of our seismic-well tie workflow).

Methods

Seismic interpretation

Our seismic interpretation comprised two main stages. The first stage represented an initial regional interpretation to define the overall basin structure and context of the TFFC. The second, more detailed stage of interpretation focused on the TFFC in the southwest part of the study area. During the regional interpretation, we mapped seven key horizons that represented surfaces that mark major changes in seismic facies; these surfaces represent the top of the acoustic basement (Caledonian), top Permian-Carboniferous, top Lower Triassic, top Upper Triassic, top Jurassic, top Cretaceous, and base Quaternary (Fig. 2). Our detailed interpretation was performed on a cropped seismic volume around the TFFC and included 25 additional horizons that represent two intra-Paleogene, 10 intra-Cretaceous, five intra-Jurassic, and eight intra-Triassic surfaces. These interpreted horizons were used in the geometric and kinematic analysis of the fault, in particular the construction of strike-projections (Walsh and Watterson, 1991) and the timing of key periods of fault and fold growth. Given the poor seismic resolution at depth and the lack of well penetrations, it is difficult to distinguish between the 1st and 2nd

megasequences, meaning they were combined into one seismic package (Permian – Carboniferous; Fig. 2).

We also used seismic attributes and colour blending techniques to help highlight and map fault networks. First, we used a variance seismic attribute that highlights discontinuities in the seismic signal by returning low values for continuous events and high values for discontinuous events (Randen et al., 2001). Second, we colour-blended dip, tensor, and semblance attributes to create a volume that images faults as relatively dark features (for detailed description and definition of these seismic attributes, please refer to Iacopini et al. (2016) and references therein). Guided by the seismic attributes and colour blended volume, we mapped the fault segments in the seismic reflection volume using composite lines taken perpendicular to the strike of the fault segments as seen on the attribute slices. We mapped the fault segments on lines that were spaced every 150 m.

Fault kinematic analysis

Our kinematic analysis of the TFFC involved the analysis of the distribution and style of strain along and between its constituent segments (e.g. Peacock & Sanderson, 1991; Childs et al., 1995; Childs et al., 2019). In this study, we drew on the methods summarised in Jackson et al. (2017). The first step was to study the temporal and spatial evolution of the fault segments using time-structure and time-thickness (isochron) maps. Isochron maps record changes in subsidence and accommodation related to fault and fold growth (e.g. Gawthorpe et al., 2003; Morley, 2002;

Schlische, 1995; Young et al., 2001; Jackson & Rotevatn, 2013). Isochron map analysis also helps highlight temporal and spatial trends in across-fault thickening, which helps determine fault growth style (e.g. Jackson et al., 2017). Next, we performed throw analysis by creating throw-length (T-x) profiles (e.g. Baudon & Cartwright, 2008; Cartwright et al., 1995; Dawers & Anders, 1995; Gupta & Scholz, 2000; Mansfield & Cartwright, 1996; Rykkelid & Fossen, 2002; Jackson & Rotevatn, 2013; Jackson et al., 2017). We collected throw values on composite seismic lines oriented perpendicular to fault strike every c. 150 m. We could then visualise how throw varies across the fault surface ('strike-projections'; Walsh and Watterson, 1991; see also Duffy et al., 2015). The strike-projections were used to further understand the geometric and kinematic evolution of faults and adjacent folds (e.g. Jackson & Rotevatn, 2013; Jackson et al., 2017; Collanega et al., 2019; Deng & McClay, 2021).

In our study, we measured two types of throw values at each (150 m-spaced) position along the fault (Fig. 3). These were labelled *observed throw* and *projected* throw to denote the difference between values that did or did not include so-called *continuous deformation* (i.e. folding), respectively (Walsh & Watterson, 1991; Walsh et al., 1996; Fig. 3). We could then calculate and display the difference between observed and projected throws to give a measure of the extent and magnitude of folding across the fault surface, given the difference between these values is essentially a proxy for fold amplitude. Throw data collected using so-called *projected horizons* mitigated the local effects of fault scarp erosion, which mainly impacted the post-Triassic succession (Fig. 3b).

Results

Structural framework

There are three major normal fault systems in the study area (Fig. 4a, b). The first one is a basement-rooted normal fault system (the TFFC; Fig. 4c). The TFFC strikes NW-SE and dips ENE. However, in the NW corner of the study area, the TFFC strikes E-W and dips N. In detail, the TFFC exhibits significant lateral variation of plane geometry with some pronounced strike bends (Fig. 4a.iii). The fault plane exhibits a planar geometry in some areas (Fig. 5a, b, e, f), while in others the fault plane shows more of a 'ramp-flat-ramp' fault geometry (Fig. 5c, d) (e.g. Rotevatn & Jackson, 2014). Based on its relatively large throw (locally >2 km; see below) and the seismic facies of the deepest, seismically imaged material it displaces, the TFFC is likely rooted in crystalline basement (i.e., Caledonian) rocks. This observation is also confirmed by previous studies using seismic reflection, gravity, and magnetic data (e.g. Gabrielsen, 1984; Faleide et al., 2008; and Indrevær et al., 2013). The upper tip of the TFFC is typically located within Cretaceous rocks (Figs. 4c and 5). The maximum throw on the TFFC is c. 670 ms (c. 1120 m) (across horizon T5) and c. 1045 ms (c. 1742 m) (across horizon J4) for observed and projected throws, respectively (Fig. 5). Maximum throw values for horizons T5 and J4 are observed at the same location, towards the NW end of the TFFC. When examined closely using seismic attribute time slices, the TFFC appears to be comprised of four distinct geometrical segments (Fig. 4a). These segments are 6-12 km long and have maximum throws that range from c. 339-669 ms (c. 565-1115 m) and c. 634-1044 ms (c. 1057-1740 m) for observed (across horizon T5) and projected (across horizon J4) throws, respectively. The TFFC generally lacks any clear abandoned hanging wall or footwall splays, which suggests that relay ramps did not develop between these segments (Fig. 4ai, ii). Instead, the TFFC is associated with clear along-strike bends (Fig. 4a). The location of these bends corresponds with the approximate tips of the segments comprising the larger TFFC (Fig. 4aiii). Fault-normal anticlines also define paleo-segment boundaries, with synclines developed at the centres of the four flanking segments (Fig. 4a.iii, b.iii).

The second fault system exists in the hanging wall of the TFFC and consists of curvilinear, ENE-WSW striking faults that in some cases are >40 km long (Fig. 4a.iii). These faults tend to dip N and die-out to the WSW before linking with the TFFC (Fig. 4a, b). The bottom tip lines of these faults are typically located in the base of the Lower Triassic succession, whereas the upper tip lines are in uppermost Jurassic-lowermost Cretaceous strata (Fig. 4c). Compared to the TFFC, these E-W striking faults have relatively low throws (c. 40 – 60 ms). The ENE-WSE-striking fault system terminates against or just before the folds located in the hanging wall of the TFFC (Fig. 4b.ii, iii) suggesting the former are not responsible for the formation of the latter.

The third fault system is present in the northern part of the study area and consists of numerous relatively short faults (c. 10 – 30 km) that strike NE-SW and dip N-NW (Fig. 4a, b). Like the second fault system, these faults have their upper tips located in the Upper Jurassic-Lower Cretaceous strata; however, their basal tips are typically located at deeper depths (i.e., in Upper Permian strata; Fig. 4c). The faults in this system tend to have modest throws (up to 200 ms).

Fault-related folding

Fault-parallel folding

Strike-perpendicular seismic sections show extensive deformation of hanging wall strata immediately adjacent to the TFFC (Fig. 5). The main structure is a fault-parallel hanging wall syncline that is best-developed in Triassic and Jurassic strata (Fig. 5a). Underlying Paleozoic stratigraphy and overlying Cretaceous stratigraphy are not or are only very gently folded, i.e. Palaeozoic strata are simply offset across the TFFC, whereas Lower Cretaceous strata onlaps onto underlying, folded Jurassic unit (Fig. 5). Along the NW portion of the hanging wall of the TFFC, antithetic and synthetic faults occur, in particular within the gently folded Triassic interval (Fig. 5c, d, f).

Fault-perpendicular folding

We also observe significant fault-perpendicular folding along the TFFC, with major hanging wall synclines and anticlines forming along strike of the fault system, especially along its south-eastern portion (Fig. 6). These large, broad folds are up to 5 km wide and have a maximum amplitude of c. 180 ms (c. 300 m) (Fig. 6). A composite seismic line running parallel to the TFFC shows that the pre-Jurassic succession is tabular, albeit folded. In contrast, the overlying Jurassic interval thins across the anticlines and thickens into the flanking synclines (Fig. 6). However, we cannot clearly see similar thickness changes in the Jurassic sequence on seismic lines trending perpendicular to the fault, a key observation that we return to below (Fig. 5a, b).

Towards the NW part of the TFFC, folding is less apparent (Fig. 6). This decrease in folding spatially coincides with an area where the TFFC is physically linked to the NE-SW-striking fault system (i.e. the third fault system described above; see also Fig. 4a).

Spatial distribution of folding

We can visualise the distribution of folding across and along the fault surface in two ways. The first and simplest way is by looking at folding at specific structural levels (Fig. 7). The second way is by using strike-projections to display how folding varies across the entire fault surface, i.e., areas of enhanced folding are marked by large differences between the observed and projected throw on the strike-projection (Fig. 8). The latter method reveals that the folding along the TFFC seems to be restricted to the sedimentary cover next to the fault, being most prominent in Upper Triassic to Upper Cretaceous strata (Fig. 8c). In detail, there are three distinct areas or patches of folding, which are all greatest at the structural level of the Jurassic (Fig. 8c). The sedimentary sequence that appears to be most folded corresponds to a highly stratified interval where differences in rheology and mechanical properties may be expected to be the highest (i.e. megasequences 3-4; Figs. 2 and 8c). Less prominent folding is also observed in the deeper Lower Triassic to Permian-Carboniferous intervals, typically near the centres of the segments comprising the TFFC (Fig. 8c); this is clearly seen in the throw-length plot for the Lower Permian horizon, where the projected and observed throw lines are almost equal along the length of the structure (Fig. 7d).

The intensity of folding in the Lower Triassic to Lower Jurassic interval appears to increase towards the NW, with the maximum observed throw difference of c. 300 ms (c. 500 m) being observed near the centre of fault segment D (Figs. 7b, c and 8c). Whereas both discontinuous (i.e. faulting) and continuous (i.e., folding) strain appear to generally increase towards the NW (Fig. 8a, b), on the *projected* throw strike-projection we note four clear throw maxima that are spatially linked to and define the centres of the constituent segments of the TFFC (Fig. 8a). In contrast, the *observed* throw strike-projection surface shows only two clear throw maxima that spatially correlate to the centres of segments B and D, and a less prominent throw maximum towards the SE that corresponds to the centre of segment A (Fig. 8b). This is supported by the throw-length plots in Fig. 7, which show that significant changes in throw or fold magnitude correlate with fault branchlines. More specifically, we notice an abrupt decrease in throw magnitude in areas where strain is partitioned between the main fault and intersecting faults (Fig. 8).

The northwesternmost segment of the TFFC (segment D) has notably higher throw than segments to the SE (up to c. 670 ms (c. 1120 m) and c. 1100 ms (c. 1835 m) for observed and projected throws, respectively, compared to c. 500 ms (c. 835 m) and 700 ms (c. 1170 m) for the other segments; Figs. 5f; 7b-d and 8). A potential explanation for the difference in throw between segment D and other segments is that the throw on the former reflects combined slip on at least two faults; the TFFC itself, plus a series of intersecting fault segments likely related to the NE-SW fault system imaged on the fault attribute slices in Fig. 4a.i, ii. The impact of splay faulting on these locally high throws becomes clearer when we examine the hanging wall and footwall cut-off points on the strike-projection surfaces (Fig. 8) and the throw-length plots (Fig. 7b-d); on these we can see an increase in the magnitude of folding and fault displacement at fault branchlines. The fact that the NW ends of the TFFC are not imaged within the study area makes it difficult to determine the geometric link between segment D and adjacent structures.

Origin of fault-related folding

Different processes can account for the folding, and specifically fault-parallel folding, next to the TFFC. First, the folding could be a result of horizontal compression related to basin inversion. However, this is not the most likely scenario, given the lack of regional evidence for an Early Cretaceous compressional event, and the fact that the observed fold geometries and onlap relationships are not consistent with inversion, e.g., the syn-kinematic strata we observe here is not folded, and onlaps onto the fault-perpendicular anticlines and onto the steep, basinward dipping limbs of the fault-parallel syncline, whereas syn-inversion growth strata typically onlaps onto the inversion fold, with the earliest or oldest growth strata being folded (e.g. Coleman et al., 2019). Another possible causal mechanism is frictional drag folding (e.g. Schlische, 1995). However, the folds formed along the TFFC are wide (>100s meters), and this is not likely to be the result of drag folding, which typically forms far narrower folds (10s – 100s meters) (Coleman et al., 2019) (Figs. 5a, f and 9a, b). We therefore interpret that the fault-parallel folds are fault-propagation folds, formed ahead of the propagating (upper and lower) tips of normal faults (e.g. Withjack et al., 1990; Allmendinger, 1998; Cosgrove & Ameen, 1999; Hardy & McClay, 1999; Gawthorpe & Leeder, 2000; Withjack & Callaway, 2000; Coleman et al., 2019; Jackson et al., 2020). In fact, vertical fault propagation can also produce frictional drag folds, which might be synthetic dip panels (i.e. layers dipping in the same direction) that are remnants of fault tip folding (see Ferrill et al., 2012, 2017 for more details).

Changes in fault plane geometry might also be responsible for the formation of small-scale fault-perpendicular and fault-parallel folds (Ehrlich & Gabrielsen, 2004). For instance, areas where the fault plane is curved (convex or concave) seems to coincide with some of the folding we noted in the Jurassic and Triassic intervals (Fig. 5c-e and 6c). These folds can potentially superimpose smaller structures onto larger ones, or simply augment and amplify the larger structures. However, we interpret that the fault-perpendicular folds (Fig. 6)represent now-breached segment boundaries between the lateral tips of precursor faults given their prevalence near the mapped lateral tips (i.e. displacement minima) of TFFC fault segments (e.g. Gawthorpe et al., 1997; Corfield & Sharp, 2000; Sharp et al., 2000; Willsey et al., 2002; White & Crider, 2006; Kane et al., 2010; Duffy et al., 2013; Khalil & Mcclay, 2017; Tavani et al., 2018; Coleman et al., 2019). These changes in fault plane geometry might have led to the formation of different fold styles along-strike of the TFFC. For example, where the fault plane is planar (Fig. 5a, b, e, f), we tend to observe a relatively simple, monocline-style of folding in the hanging wall of the TFFC (Fig. 9a, b, e, f). In contrast, in areas where the fault plane shows more of a 'ramp-flat-ramp' fault geometry (e.g. Rotevatn & Jackson, 2014), an anticline-syncline pair is present (Figs. 5c, d and 9c, d). The former folding style (i.e. fault-parallel monocline) is consistent with our earlier interpretation of extensional folding due to upward propagation of the fault tip (e.g. Khalil and McClay, 2002; Ferrill et al., 2007; Conneally et al., 2017; Smart and Ferrill, 2018). Although differential compaction can also form monoclines above planar faults (e.g. Fig. 5 in Skuce, 1996), it would not result in the syn-kinematic strata onlapping onto the monocline's basinward-facing steep limb like we observe here (Fig. 9a, b, e, f). This onlap relationship indicates the fold grew whilst at-surface rather than after some burial as required by a differential compaction model. For the latter style of folding (i.e. anticline-syncline pairs) we note a correlation between the location of these anticline-syncline pairs and changes in fault plane geometry, and we interpret the origin of these folds as a result of changes in fault (e.g. Gibbs, 1984; McClay & Scott, 1991; Rotevatn & Jackson, 2014; Deng and McClay, 2019; McHarg et al., 2020).

Temporal evolution of the TFFC

Having established the present structural style of the TFFC and described its geometry, we now focus on how this fault system developed through time. We do this by studying thickness variations in syn-kinematic stratal units along and across fault arrays (e.g. Cowie et al., 2000; McLeod et al., 2000; Gawthorpe et al., 2003), as well as the spatial distribution of the fault-propagation folds described in the previous section; together these observations help us determine the position of now-breached segment boundaries and the style of propagation (surface breaching vs. buried) of the TFFC (e.g. Gawthorpe et al., 1997; Corfield & Sharp, 2000; Sharp et al., 2000). For each interpreted stratigraphic unit, we present our observations first and then provide possible interpretations that could explain how the TFFC evolved during that time period.

Carboniferous – Permian (350-250 Ma)

The first seismic package above the crystalline basement is assigned a Carboniferous to Permian age, which makes it lithostratigraphically equivalent to the 1st and 2nd megasequences (Fig. 2). No clear thickness variations are observed across the TFFC during this period (Fig. 10a). However, these Paleozoic strata seems to thicken regionally towards the NE, away from the TFFC, as well as thickening (by c. 70 ms or c. 150 m) across relatively small faults in its footwall (Fig. 10a).

The lack of thickness variations across the TFFC suggests that the fault was not active at this time, although smaller, E-W striking faults in its footwall may have been (Fig. 10a). This style of strain partitioning may have arisen because the smaller faults were more optimally oriented to accommodate N-S-directed stretching associated with this initial period of relatively mild extension (Rift Phase 2; Fig. 2). Regional NE thickening of this seismic package might be related to the long wavelength variations in basin subsidence related to regional tectonic events.

Early Triassic (251-245 Ma)

The second stratal unit is Early Triassic and represents the Klappmyss Formation (Fig. 2). This mudstone-dominated unit is generally well-imaged in seismic data and shows no significant thickness variations across the TFFC, or across the entire area of interest (Fig. 10b). The only exception to this is the clear across-fault thickening seen in association with the N-dipping fault segment defining the NW portion of the TFFC (e.g. segment D; Figs. 5f and 10b). These thickness variations (up to 200 ms, c. 360 m thick) suggest that this segment was active during this period, while others were inactive.

Middle Triassic (245-227 Ma)

The third stratal unit is Middle Triassic and represents the Kobbe and Snadd formations (Figs. 2 and 5c). This unit displays significant thickness variations (43 to 346 ms, c. 80 to 620 m) across all segments of the TFFC, and more minor variations along a NE-SW-striking fault segment located in the NE part of the study area (Figs. 5c and 10c). These thickness variations suggest that segments A-D of the TFFC, along with a NE-trending fault in its hanging wall, were all active during this period.

Late Triassic (227-201 Ma)

The fourth stratal unit is Late Triassic and represents the Fruholmen Formation (Figs. 2 and 5c). This unit shows thickening across the NE- trending fault segment (segment D), and minor thickness variations across segments A-C of the TFFC (Fig. 10d). In fact, the strata appear to thin towards the TFFC, suggesting that the fault was likely buried and expressed by a fault-propagation fold during this period (Fig. 10d).

Jurassic (200-150 Ma)

The fifth stratal unit is Jurassic and mostly includes the upper part of the 3rd megasequences (Fig. 2). This unit includes all the Jurassic formations and is well-imaged in seismic data, except for areas immediately next to the TFFC where bedding is relatively steep and rocks may be highly faulted (Fig. 5c, f). Despite being relatively thin, this unit shows significant thickness variations across and along the TFFC (Fig. 10e). More specifically, we see significant synclinal depocenters (up to 170 ms, c. 302 m thick) near the centres of all fault segments, with Jurassic strata thinning across and onlapping onto fault-perpendicular anticlines *and* onto the relatively steep, basinward-dipping limbs of the fault-parallel syncline (Figs. 9d and 10e). However, the Jurassic stratal unit, having thinned onto the steep dipping limb, then thickens towards the surface of the TFFC (Fig. 9d). The overall across-fault thickening also differs *along strike* of the TFFC, with the maximum across-fault thickening (130 ms or c. 230 m) occurring along segment D (Fig. 10e).

The observed onlap relationships and thickness variations across and along-strike of the TFFC suggest that many of the fault-propagation folds described above formed during the Early Jurassic (start of Rift Phase 3; Fig. 2). However, this stage of fault-propagation folding might have been restricted to segments B-D of the TFFC. Therefore, we tentatively suggest that segments B-D rapidly formed and linked laterally during the Late Triassic to Early Jurassic, before linking to segment A. The fact that the fault-propagation folding was the prominent at-surface deformation style in the Jurassic, when earlier, during the Triassic, the fault was a surface-breaching structure, requires that the fault was reburied between the Late Triassic to Early Jurassic. Having breached the surface, it is clear from thinning of the footwall strata and the presence of hanging wall depocenters that by the end of the Jurassic, the TFFC was a single, through-going fault system (Fig. 10e).

Cretaceous (132.6-97.5 Ma)

The youngest stratal unit is Cretaceous and is lithostratigraphically equivalent to the 4th megasequence (Fig. 2). This relatively thick unit (up to 870 ms, c. 1240 m) displays an overall wedge-shaped geometry on seismic data, a geometry that is classically associated with syn-kinematic strata (e.g. Fig. 5a) (Prosser, 1993). However, we further subdivide the unit into three sub-units based on intraformational onlap relationships (Fig. 5). The lowermost sub-unit onlaps onto and thins towards the underlaying Jurassic unit forming the steep, basinward dipping limb of the TFFC hanging wall syncline. This sub-unit is absent in the immediate hanging wall of the fault and from the fault footwall (Figs. 5 and 9). These geometries suggest that the TFFC was acting as a blind, fully linked fault at this stage, leading to fault-propagation folding of cover strata and the formation of an at-surface, basinward-facing monocline (e.g. Sharp et al., 2000) (Figs. 6 and 10f). Thus, having broken surface again during the Jurassic, the fault was reburied, such that during the Early Cretaceous the fault was expressed at the surface as a fault-propagation fold. In the north of the study area, in contrast to that seen along the main part of the TFFC, the lower sub-unit thickens across the NE-SW-striking segment in the hanging wall of the larger structure (Fig. 10f). Thickening of this sub-unit along the entire strike length of this fault suggests that this was actively accruing displacement as a single, continuous, relatively long structure (c. 35 km) (Fig. 10f).

The middle Cretaceous sub-unit, in comparison to the lowermost one, thickens (rather than thins) towards the NE-dipping segments of the TFFC, defining several clear, fault-bound depocenters (i.e., A-C). This unit is partially absent in the footwalls of these segments (Fig. 10g). At its base, this sub-unit locally onlaps the underlying, basinward-dipping, lowermost sub-unit; this onlap is not, however, as pronounced as that observed at the base of the lowermost sub-unit, which onlaps onto basinward-tilted Jurassic strata (Figs. 5a-c and 9a-c).

The distribution of the uppermost Cretaceous sub-unit defines several large, fault-bound depocenters, with basal and intraformational onlap towards the TFFC being absent (Figs. 9 and

10h). There is also a notable across-fault thickening along a NE-SW-striking segment that is likely part of the ENE-WSW fault system, as opposed to the TFFC (Figs. 4a and 10h).

The presence of discrete fault- rather than fold-bound depocenters in the middle and upper parts of the Cretaceous, as defined by the distribution and seismic-stratigraphic architecture of the related sub-units, suggest that, by this time, the TFFC had breached its overlying fault-propagation fold to form a classic, half-graben-style depocenter. We interpret that this depocenter was segmented along-strike due to the along-strike variations in accommodation created by differential compaction of underlying strata across earlier-formed (i.e., Triassic and Jurassic) fault-propagation folds . This is supported by the lack of any clear fault-perpendicular, syncline-anticline pairs along-strike of the TFFC (Fig. 6) and the overall decrease in throw magnitude at the Cretaceous structural level (Figs. 7a and 8).

Summary

In summary, we suggest that the TFFC initiated during the Early – Middle Triassic, with fault segments B, C and D rapidly forming and linking laterally during the Middle Triassic(Fig. 10). During this period (Middle Triassic), the TFFC was a surface-breaching structure. However, the TFFC was subsequently reburied between the Late Triassic and Early Jurassic, with continued upward propagation of the fault forming a fault-propagation fold at the free surface. By the Late Jurassic the TFFC broke the surface again and was a single, through-going fault system. The fault was subsequently reburied again, such that during the Early Cretaceous it was expressed at the surface as a fault-propagation fold. The TFFC again breached its overlying fault-propagation fold to form a classic, half-graben-style depocenter at the time of deposition of the middle and uppermost Cretaceous stratal units. Therefore, the TFFC has experienced two distinct phases of fault-propagation folding during its lifetime.

Discussion

Current models show that extensional fault-propagation folds are formed during the early stages of extension, with initially intact monoclines subsequently being breached by their causal faults (e.g. Gawthorpe & Leeder, 2000; Sharp et al., 2000; Jackson & Lewis, 2016; Coleman et al., 2019; Deng & McClay, 2019). Our work on the TFFC shows that the start of fault-propagation folding is marked by the thinning and onlap of the Lower Jurassic strata towards the fault onto the NE-dipping limb of the Upper Triassic hanging wall syncline (Figs. 5 and 9). However, a second phase of fault-propagation folding likely occurred during the Early Cretaceous as marked by the onlap of growth strata onto the underlaying Jurassic unit, and the overall thinning of the Lower Cretaceous towards the TFFC (Figs. 9 and 10f). The thickening of overlying Cretaceous units towards the fault indicates that the fault had breached the fault-propagating fold at this time (Fig. 10g, h). Therefore, the fault-propagation folds we observe did not simply form during the earlier phase of extension, but seemingly are best-developed during the Jurassic and, especially, the Cretaceous, after the fault had been active for some time. Based on our borehole-constrained seismic-stratigraphic framework, we show that fault-propagation folds can form over periods of c. 25.7 Myr (1st phase of folding) and c. 32 Myr (2nd phase of folding). Multiple phases of superimposed fault-propagation folding can therefore occur if a fault is subsequently reburied during times of: (i) relatively low tip propagation rates (relative to a constant, long-term, sediment accumulation rate); and/or (ii) relatively high sediment accumulation rate (relative to a constant, long term, tip propagation rate).

Previous studies have shown that changes in fault dip direction and the presence of multiphase extension results in complex hanging wall fold and fault geometries (Coleman et al., 2019; Deng & McClay, 2019). Our results from the TFFC show that despite the presence of multiphase extension and associated changes in fault geometry as a function of growth patterns, the geometry of the studied fault-propagation folds remained relatively simple. This is probably because the area experienced coaxial rifting and did not undergo dramatic changes in extension direction. However, our study confirms the complicated nature of syn-kinematic hanging wall strata and fault geometries as a result of folding before the fault reaches the surface and forms a scarp. The fault propagation folding and secondary structures observed in the area have led to the formation of complex syn-kinematic hanging wall geometries, with stratal units thinning onto the fold limb and other units thickening towards the fault. These different observed geometries suggest that the nature of syn-kinematic hanging wall strata can be more complex than the classic wedge-shaped cross-sectional geometry often described from extensional basins (e.g. Prosser, 1993).

Differences in host rock rheology, fault propagation rate, and throw magnitude can lead to spatial variability in the size and distribution of fault-propagation folds (Mansfield & Cartwright, 2000; Coleman et al., 2019). The fault-propagation folds we studied appear to have preferentially developed above the upper tip line of the TFFC within mechanically weak/incompetent layers (Fig. 8c). As discussed above, the constituent segments of the TFFC fault segments likely linked relatively rapidly during the Late Triassic – Early Jurassic (Figs. 7c and 10d, e). Rapid linkage, which may have been associated with high displacement rates, might account for relatively poor development of fault-propagation folds at the Jurassic structural level (Figs. 5d, f and 6). In contrast, the well-developed Early Cretaceous fault-propagation folds probably reflect a period of relatively low displacement rates on the TFFC. This period of low displacement rates during the Early Cretaceous is consistent with our regional understanding of rift phases as it correlates with very end of Rift Phase 3 (Fig. 2). Additionally, the Lower Cretaceous interval is dominated by mudstone, which we infer is relatively weak and deforms in a more ductile manner, whereas the underlying Jurassic section is more heterogenous, comprising relatively strong and brittle sandstone and siltstone layers at its base, and weaker and more ductile mudstone near its top. We therefore infer that well-developed fault-propagation folds formed in the weaker and more ductile Early Cretaceous mudstones and a relatively poorly-developed fault-propagation folds in the more the mechanically layered stratigraphic Jurassic sequence (e.g. Conneally et al., 2017; Lăpădat et al., 2017; Roche et al., 2020). Despite differences in how well-developed the fault-propagation folds are at the various structural levels, they appear to be confined to the upper, relatively mechanically weak or incompetent units (i.e. megasequences 3-5) capping the more competent, deeper units (i.e. megasequences 1-2). Similar cases of folding being restricted to upper units with lower deeper units acting as forcing members are described in literature (e.g. Maurin & Niviere, 1999; Sharp et al., 2000; Jackson et al., 2006).

Even though the importance of considering ductile deformation when assessing normal fault growth has long been recognized (e.g. Walsh & Watterson, 1991), this notion has only relatively recently seen a broader support due to the increasing use of high-quality 3D seismic reflection data, and the study of exceptional field exposures that reveal the complex three-dimensional geometry of normal faults and related folds (e.g. Childs et al., 2002; Walsh et al., 1999, 2002, 2003; van der Zee & Urai, 2005; Schöpfer et al., 2006; Long & Imber, 2010; Giba et al., 2012; Jackson & Rotevatn, 2013; Duffy et al., 2015; Fossen & Rotevatn, 2016; Conneally et al., 2017; Freitag et al., 2017; Camanni et al., 2019; Jackson et al., 2020; Roche et al., 2020, 2021; Deng & McClay, 2021). Our study further highlights the value of considering both discontinuous (i.e., faulting) and continuous (i.e., folding) deformation when studying normal fault growth. We show that a significant proportion of strain (10 – 40%; see Fig. 4 in supplementary material for complete data distribution) along the TFFC is expressed in a ductile manner as (fault-propagation) folding, as opposed to a brittle shear fracturing or faulting. Therefore, not including the continuous component of the strain field will most likely result in erroneous structural restorations or an incomplete understanding of normal fault growth. Our work on the TFFC also suggests that the construction of strike-projections including both discontinuous and continuous strain, and plotting the difference between the two, can be a quick and powerful tool to illustrate the three-dimensional variability of fault-related folds across normal fault surfaces. When integrated with isochron analysis, this can help us determine the patterns and products of normal fault growth, and the origin and evolution of fault-related folds, both of which have major implications for rift morphology and stratigraphic development.

Conclusions

We used high-quality 3D seismic reflection data from SW Barents Sea, offshore Norway to analyse the along-strike variations and geometric evolution of fault-propagation folds associated with a large, basement-involved fault. We showed that this fault system consists of four, hard-linked segments that are 6-12 km long. Fault-perpendicular anticlines, which are flanked by fault-perpendicular synclines, occur at relict segment boundaries (i.e. displacement-gradient folds), whereas fault-parallel folds (i.e. fault-propagation folds) occur along-strike and down-dip of the flanking fault. Based on the fault-fold relationships and the architecture of the nearby stratigraphic record, we suggest that the fault underwent a phase of relatively rapid lateral linkage in the Late Triassic to Early Jurassic. We also show that fault-propagation folding was protracted (c. 57 Myr), occurring in multiple phases due to periodic burial of the fault by syn- and intra-kinematic strata. Our findings further highlight the value of considering both discrete (i.e., fault-related) and continuous (i.e., fold-related) strain when assessing the processes of fault growth, and the geometry, tectono-stratigraphic evolution, and resource potential of rift basins. Using strike-projections of fault surface is particularly powerful for highlighting 3D variations in faulting and folding, something that may be missed or unclear when using throw-length and throw-depth plots alone.

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Figure Captions

Figure 1. (a) Regional structural elements of SW Barents Sea. The green outline highlights the location of the study area. The location of the studies used to further constrain seismic facies and age relationships is noted by S1 (Mohammedyasin et al., 2016) and S2 (Harishidayat et al., 2015 and Torabi et al., 2019). The red dashed line shows the location of the regional 2D cross section shown in Fig 1b. The line brown line shows the present-day coastline of Norway. Structural elements abbreviations (FH: Fedynsky High, VD: Veslekari Dome, NB: Nordkapp Basin, NH: Norsel High, SG: Swaen Graben, MH: Mercurius High, SD: Samson Dome, LH: Loppa High, HB: Hammerfest Basin, BF: Bjørnøyrenna Fault Complex, VH: Veslemøy High, SR: Senja Ridge, TFFC: Troms-Finnmark Fault Complex, MF: Måsøy Fault Complex, SB: Sørvestsnaget Basin, SH: Stappen High, BP: Bjarmeland Platform, TB: Tromsø Basin, FB: Finnmark Platform). The map is modified after information found in the Norwegian Petroleum Directorate fact page http://www.npd.no/en/. (b) Regional 2D seismic cross section showing the basin scale settings across The Finnmark Platform, Hammerfest Basin and Loppa High (modified from Gabrielsen,

1984 and Mohammedyasin et al., 2016).

Figure 2. Stratigraphic column for SW Barents Sea showing major tectonic events (modified after Gac et al., 2018 and Edmundson et al., 2019). The figure shows the major seismic horizons picked in the area and near the well location.

Figure 3. Cartoon illustrations showing (a): how projected and observed throw cut-off points are picked to distinguish between throw due to tectonic subsidence alone and throw due to tectonic subsidence and deformation (modified after Duffy et al., 2015). (b): a degraded fault scarp.

Figure 4. (a.i): time slice at depth of 1400 ms that reveal the major fault systems in the study area using colour blending that combines tensor (yellow), dip (cyan) and semblance (magenta). (4.a.ii): a variance horizon slice extracted at the J4 level that shows the fault segments comprising the TFFC structure. (4.a.iii): interpreted fault systems using seismic attribute data above that shows the three fault systems within the study area. The red dotted line shows the composite seismic line location (Fig 4.c). The TFFC fault segments are labeled A-D. (4.b) Time structure maps of the base Permian-Carboniferous (4.b.i) and base Jurassic (4.b.ii) horizons. These horizons represent base syn-rift surfaces that corresponds to the 2nd and 3rd rift phases. The Blue dashed lines show the location of the seismic sections shown on Figs 5 & 6. The black dashed area represents the extent of the detailed horizon interpretation used in the fault kinematic analysis and Fig 10. (4.b.iii): interpreted structures from the base-Jurassic horizon.

(4.c) Composite seismic line that highlights the structural variability between the three fault systems in the study area. Faults are colour-coded following the map legend in Fig 4a (Black: TFFC, Blue: E-W fault system, Maroon: ENE-WSW fault system). **Figure 5.** Interpreted seismic sections taken perpendicular to the strike of TFFC. These sections show the changes in fault patterns and deformation along the TFFC. Black arrows represent observed reflection terminations and red arrows highlight key onlap relationships in the sedimentary cover approaching the fault surface. The location of these sections is shown on Fig 4.b.i. The location of the sections in Fig 9 are shown on top of each respective section here.

Figure 6. (a) Composite seismic line taken along strike of the TFFC. The line location is given in Fig. 4b.i. Location of seismic sections perpendicular to TFFC (Fig 5) are shown as black dashed lines and the location of the strike bends in the section are highlighted by black triangles above the section. TFFC fault segments are labelled A-D and shown above the interpreted composite seismic section (b). The location where the TFFC intersects this composite seismic section is shown on (b) by a red dashed line. (c) Time-structure maps for the P-C4 (i), T4 (ii), J4 (iii) and C4 (iv) horizons are shown below the interpreted composite seismic line to show the fold development and distribution across different structural level.

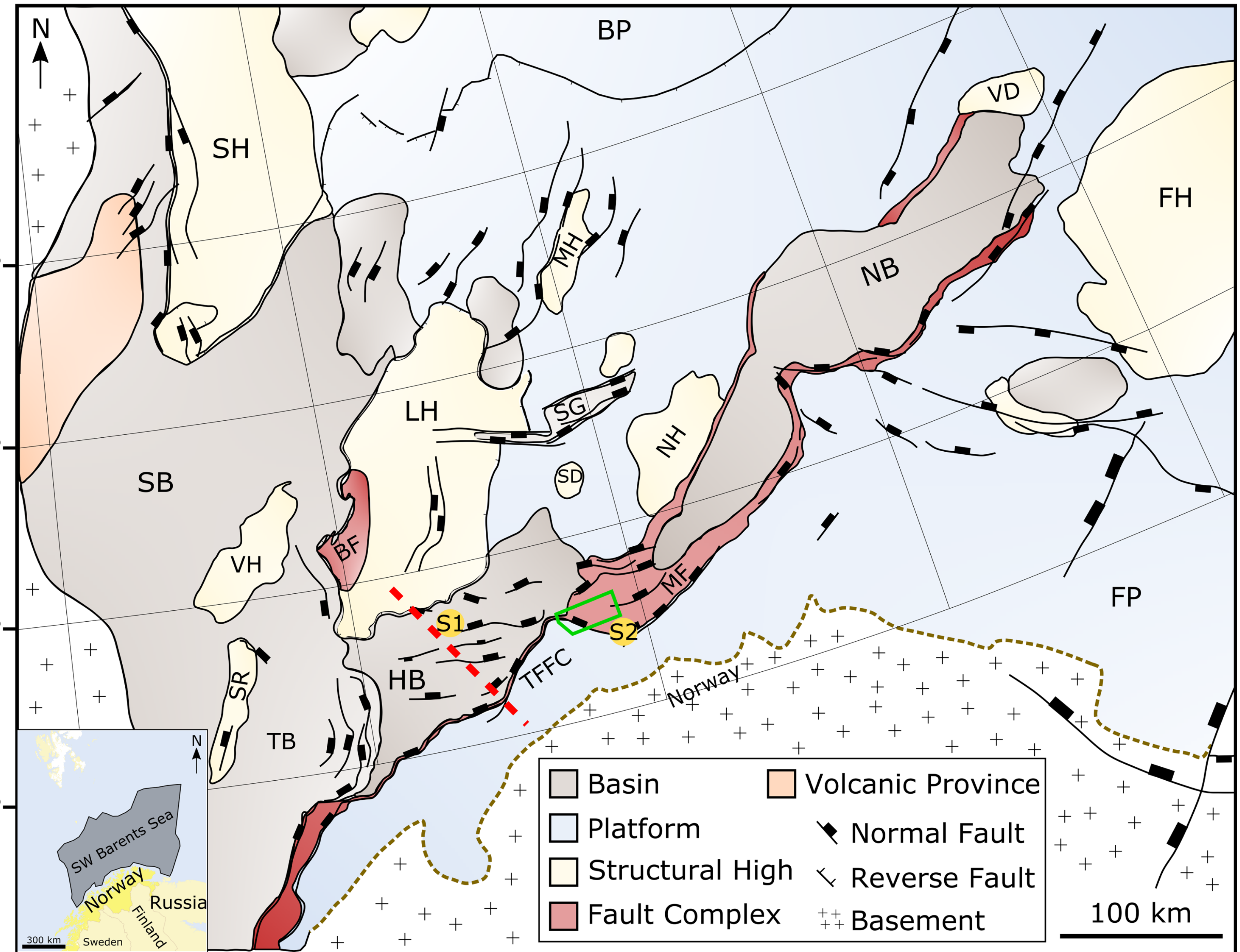
Figure 7. Throw-length profiles for horizons C4, J4, T4 and P-C4 (Lower Cretaceous, Lower Jurassic, Lower Triassic and Lower Permian of Upper Carboniferous). The blue line represents projected throw values while the red line shows the observe throw values. The difference between the two throw types is shaded in light yellow to represent ductile deformation. TFFC fault segments are marked by the green lines separating segments A – D from left to right.

Figure 8. Strike-projected throw distributions along the TFFC surface for projected (a) and observed (b) throws. The two strike-projections show increased throw towards the NW part of the fault and local throw maximums along strike of the fault. The location of the local throw maximums coincides with the centre of each fault segment making the TFFC structure. The throw maximum the NW corresponds with a branchline that is interacting with the TFFC. The image on the right (c) shows the throw difference between projected and observed throws and represent the folding component along the TFFC, which is shown to be restricted in the sedimentary cover above basement. TFFC fault segments are marked by the green lines and annotated A – D respectively.

Figure 9. Un-interpreted and interpreted, vertically exaggerated (x10) versions of the seismic sections shown in Fig.5 that highlight the folding and onlap relationships in the U. Triassic – u. Cretaceous stratigraphy.

Figure 10. Isochrone maps superimposed on variance attribute surfaces (left) and interpretive sketches (right) for key stratigraphic units across the TFFC. The isochrone maps and accompanying sketches illustrate the thickness variations across the stratigraphic unit and constrain the timing of fault activity in the TFFC and adjacent faults. The location of individual TFFC fault segments is marked by the labelled circles A – D.

(a)


20°
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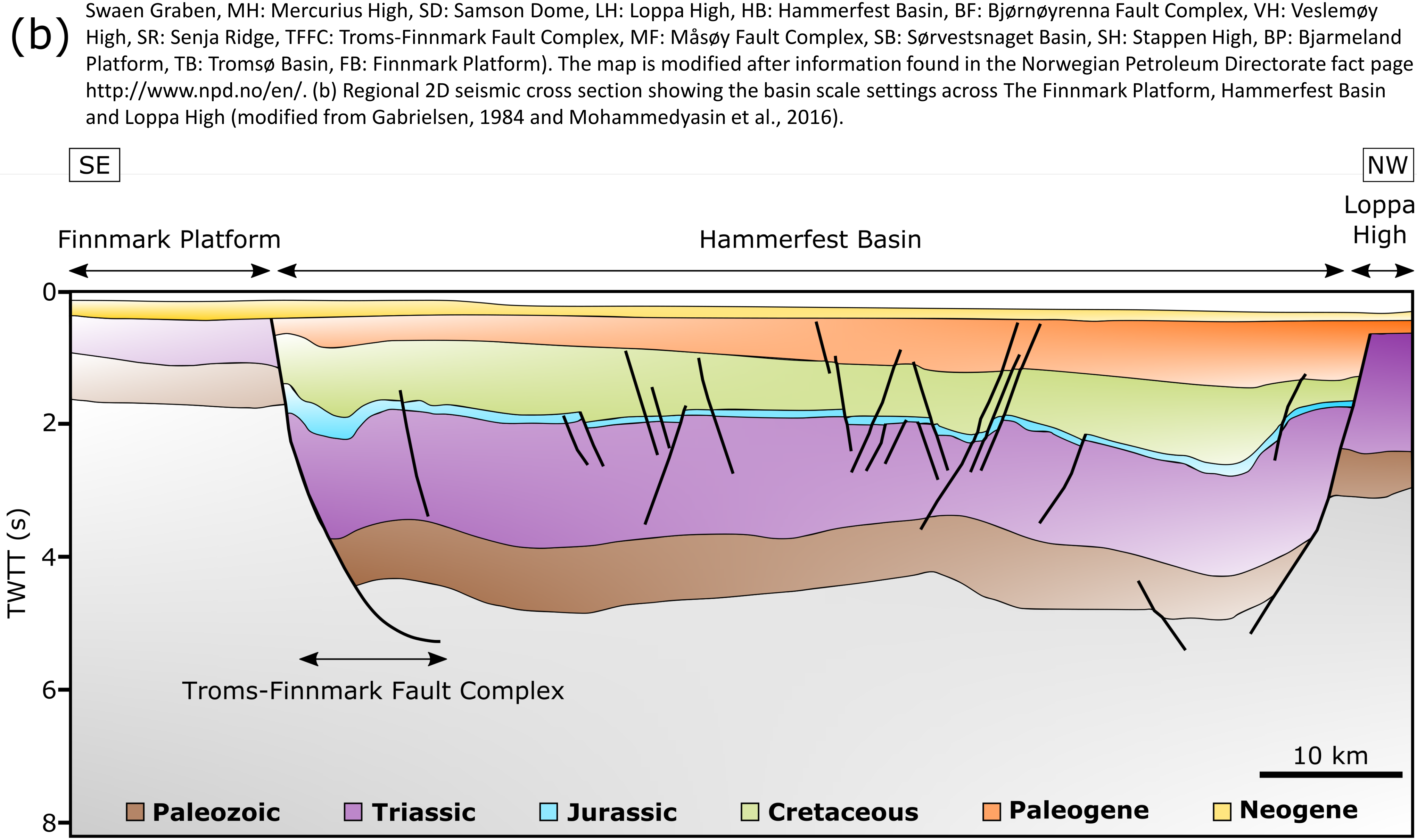
TFFC
Norway
SR
TB
SW Barents Sea
Norway
Russia
Finland
Sweden
300 km
Basin
Platform
Structural High
Fault Complex
Volcanic Province
Normal Fault
Reverse Fault
Basement
100 km

Figure 1. (a) Regional structural elements of SW Barents Sea. The green outline highlights the location of the study area. The location of the studies used to further constrain seismic facies and age relationships is noted by S1 (Mohammedyasin et al., 2016) and S2 (Harishidayat et al., 2015 and Torabi et al., 2019). The red dashed line shows the location of the regional 2D cross section shown in Fig 1b. The line brown line shows the present-day coastline of Norway. Structural elements abbreviations (FH: Fedynsky High, VD: Veslekari Dome, NB: Nordkapp Basin, NH: Norsel High, SG: Swaen Graben, MH: Mercurius High, SD: Samson Dome, LH: Loppa High, HB: Hammerfest Basin, BF: Bjørnøyrenna Fault Complex, VH: Veslemøy High, SR: Senja Ridge, TFFC: Troms-Finnmark Fault Complex, MF: Måsøy Fault Complex, SB: Sørvestsnaget Basin, SH: Stappen High, BP: Bjarmeland Platform, TB: Tromsø Basin, FB: Finnmark Platform). The map is modified after information found in the Norwegian Petroleum Directorate fact page http://www.npd.no/en/. (b) Regional 2D seismic cross section showing the basin scale settings across The Finnmark Platform, Hammerfest Basin and Loppa High (modified from Gabrielsen, 1984 and Mohammedyasin et al., 2016).



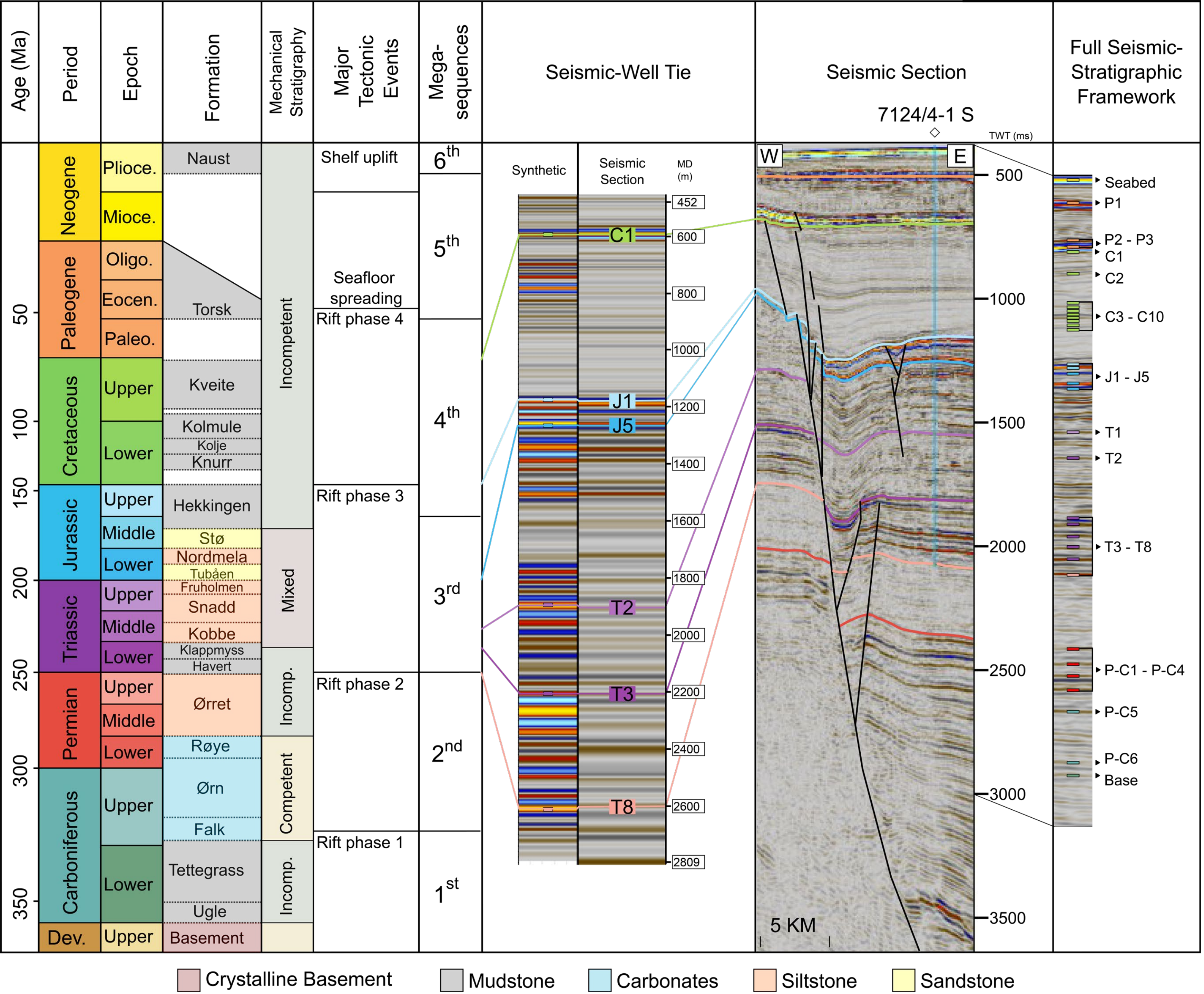


Figure 2. Stratigraphic column for SW Barents Sea showing major tectonic events (modified after Gac et al., 2018 and Edmundson et al., 2019). The figure shows the major seismic horizons picked in the area and near the well location.

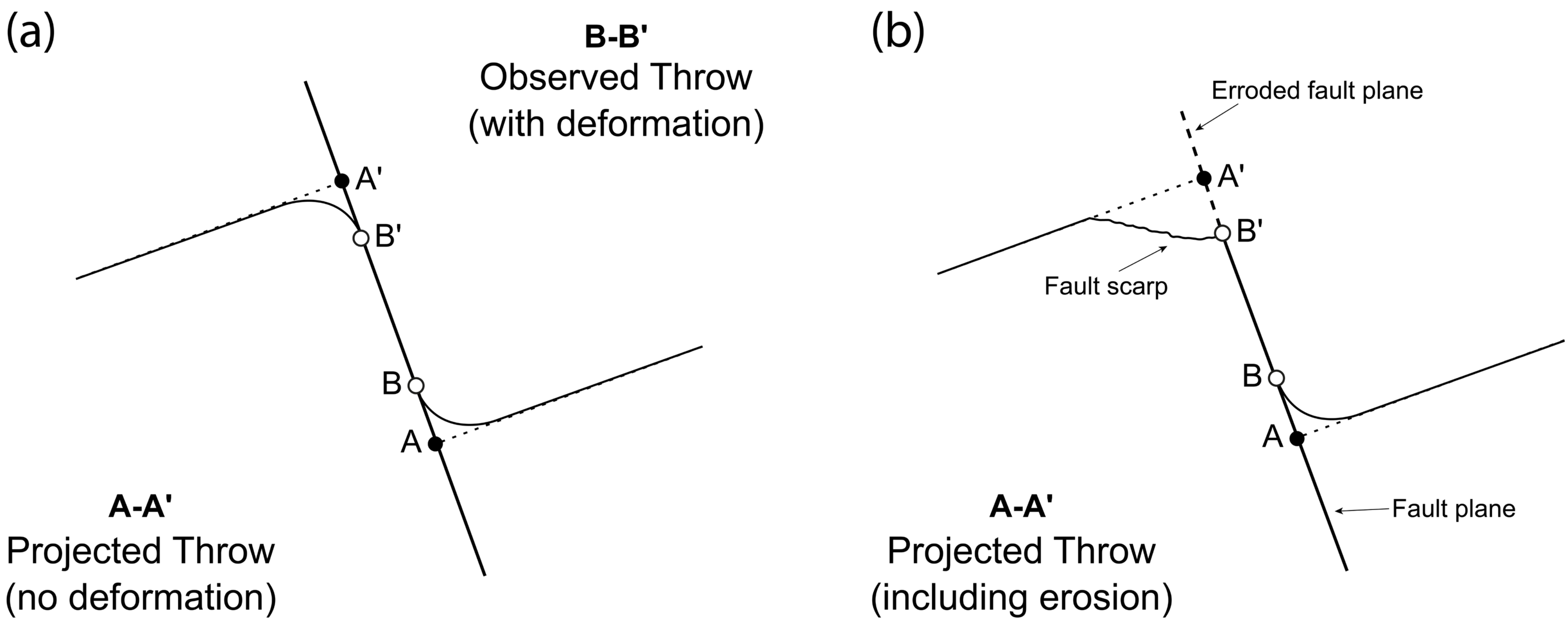
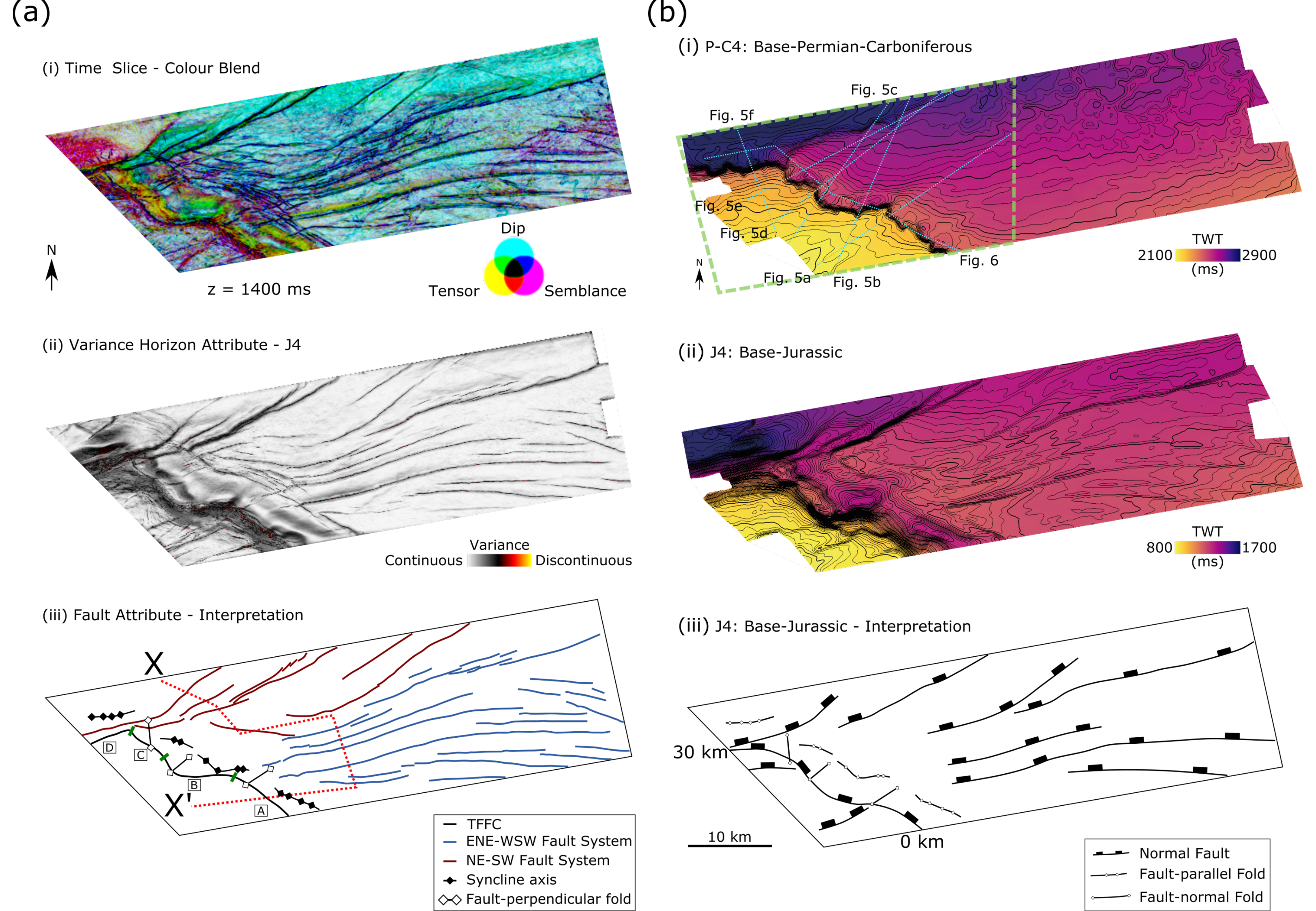


Figure 3. Cartoon illustrations showing (a): how projected and observed throw cut-off points are picked to distinguish between throw due to tectonic subsidence alone and throw due to tectonic subsidence and deformation (modified after Duffy et al., 2015). (b): a degraded fault scarp.



(a)
(b)
(i) Time Slice - Colour Blend
N
z = 1400 ms
Dip
Tensor
Semblance
(ii) Variance Horizon Attribute - J4
Variance
Continuous
Discontinuous
(iii) Fault Attribute - Interpretation
X
X'
A
B
C
D
TFFC
ENE-WSW Fault System
NE-SW Fault System
Syncline axis
Fault-perpendicular fold
(i) P-C4: Base-Permian-Carboniferous
Fig. 5c
Fig. 5f
Fig. 5e
Fig. 5d
Fig. 5a
Fig. 5b
Fig. 6
TWT
2100
2900
(ms)
(ii) J4: Base-Jurassic
TWT
800
1700
(ms)
(iii) J4: Base-Jurassic - Interpretation
30 km
0 km
10 km
Normal Fault
Fault-parallel Fold
Fault-normal Fold

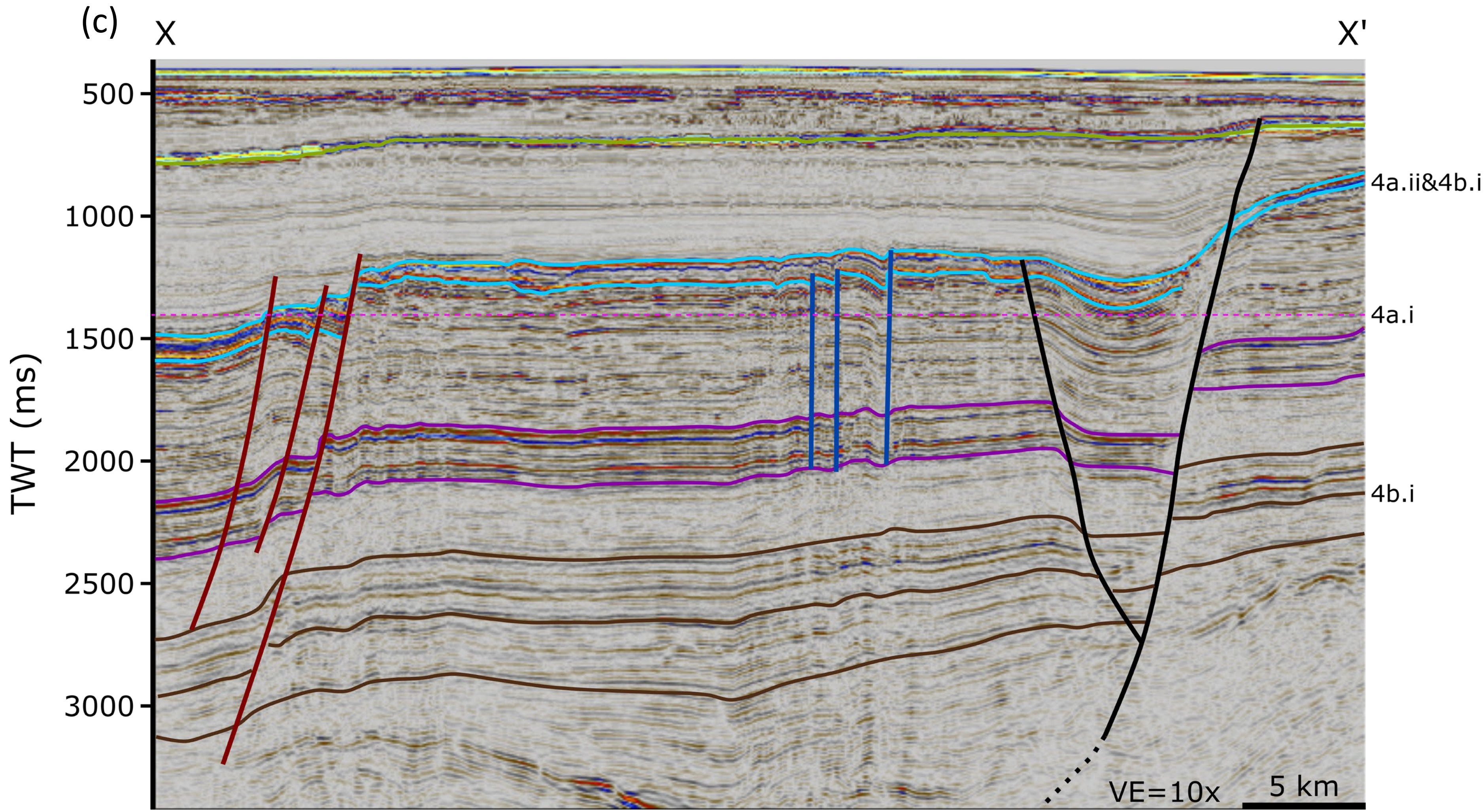


Figure 4. (a.i): time slice at depth of 1400 ms that reveal the major fault systems in the study area using colour blending that combines tensor (yellow), dip (cyan) and semblance (magenta). (4.a.ii): a variance horizon slice extracted at the J4 level that shows the fault segments comprising the TFFC structure. (4.a.iii): interpreted fault systems using seismic attribute data above that shows the three fault systems within the study area. The red dotted line shows the composite seismic line location (Fig 4.c). The TFFC fault segments are labeled A-D. (4.b) Time structure maps of the base Permian-Carboniferous (4.b.i) and base Jurassic (4.b.ii) horizons. These horizons represent base syn-rift surfaces that corresponds to the 2nd and 3rd rift phases. The Blue dashed lines show the location of the seismic sections shown on Figs 5 & 6. The black dashed area represents the extent of the detailed horizon interpretation used in the fault kinematic analysis and Fig 10. (4.b.iii): interpreted structures from the base-Jurassic horizon. (4.c) Composite seismic line that highlights the structural variability between the three fault systems in the study area. Faults are colour-coded following the map legend in Fig 4a (Black: TFFC, Blue: E-W fault system, Maroon: ENE-WSW fault system).

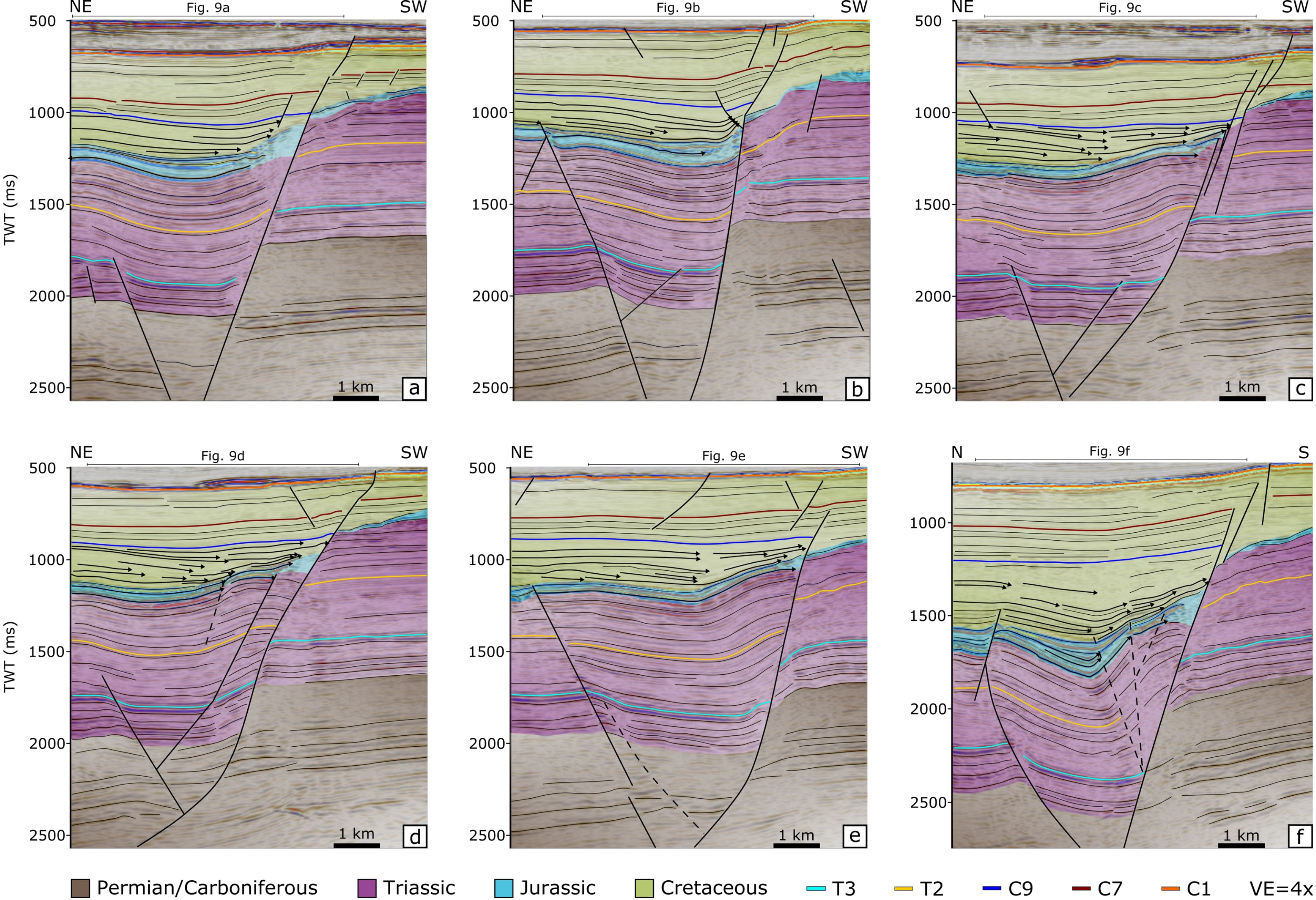


Figure 5. Interpreted seismic sections taken perpendicular to the strike of TFFC. These section show the changes in fault patterns and deformation along the TFFC. Black arrows represent observed reflection terminations and red arrows highlight key onlap relationships in the sedimentary cover approaching the fault surface. The location of these sections is shown on Fig 4.b.i. The location of the sections in Fig 9 are shown on top of each respective section here.

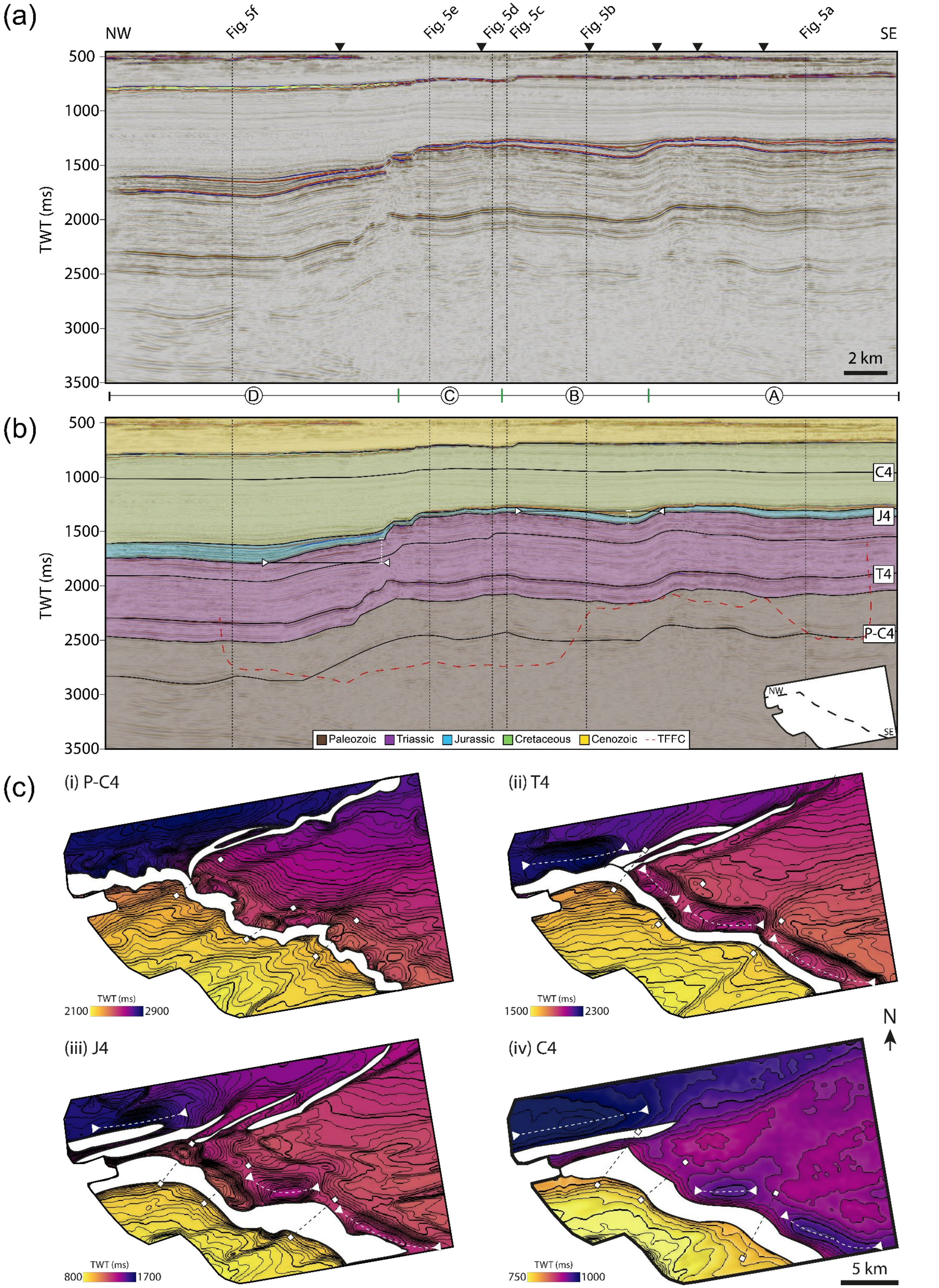


Figure 6. (a) Composite seismic line taken along strike of the TFFC. The line location is given in Fig. 4b.i. Location of seismic sections perpendicular to TFFC (Fig 5) are shown as black dashed lines and the location of the strike bends in the section are highlighted by black triangles above the section. TFFC fault segments are labelled A-D and shown above the interpreted composite seismic section (b). The location where the TFFC intersects this composite seismic section is shown on (b) by a red dashed line. (c) Time-structure maps for the P-C4 (i), T4 (ii), J4 (iii) and C4 (iv) horizons are shown below the interpreted composite seismic line to show the fold development and distribution across different structural level.

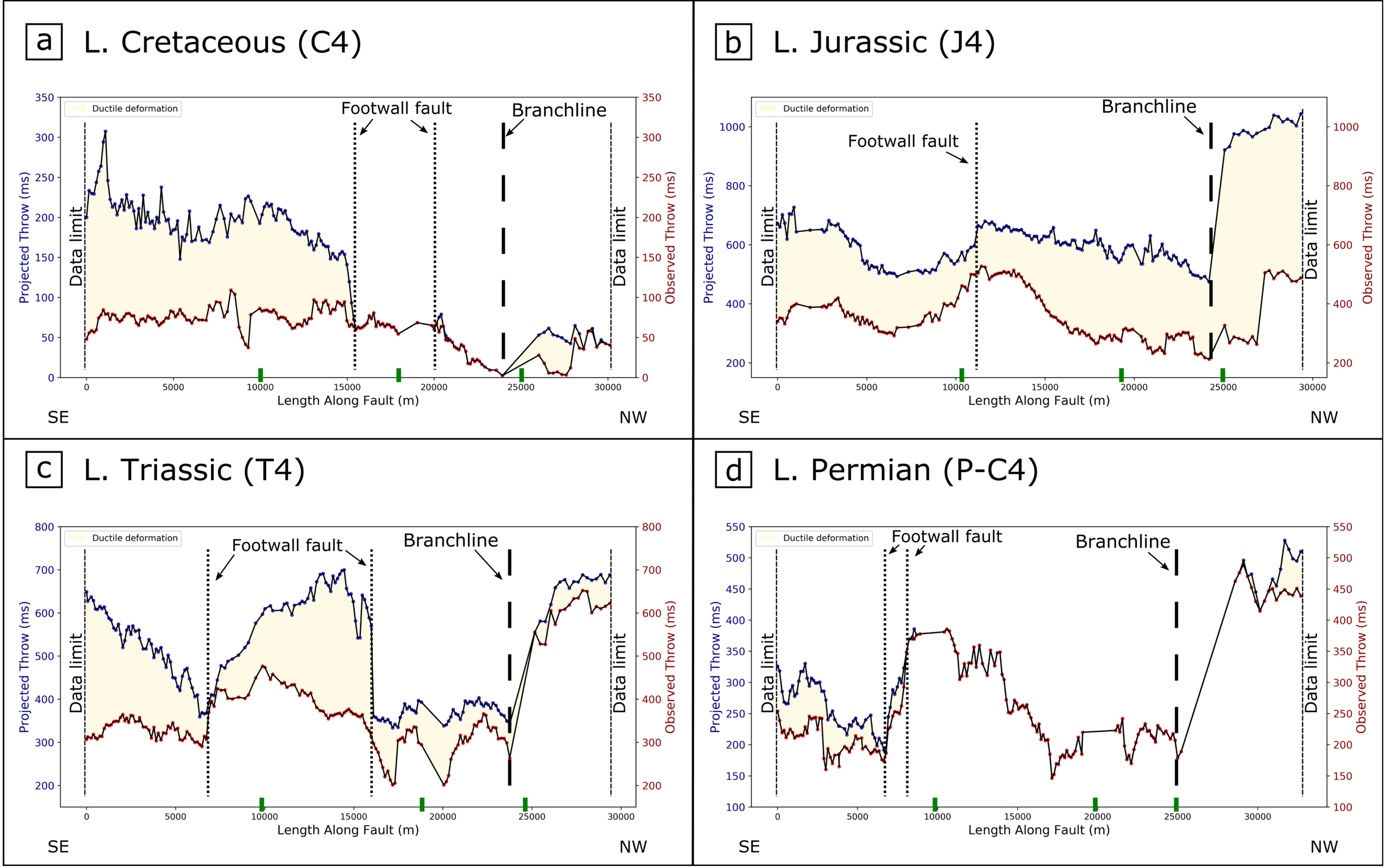


Figure 7. Throw-length profiles of TFFC at stratigraphic levels of C4, J4, T4 and P-C4 (Lower Cretaceous, Lower Jurassic, Lower Triassic and Lower Permian of Upper Carboniferous). The blue line represents projected throw values while the red line shows the observe throw values. The difference between the two throw types is shaded in light yellow to represent ductile deformation. TFFC fault segments are marked by the green lines separating segments A – D from left to right.

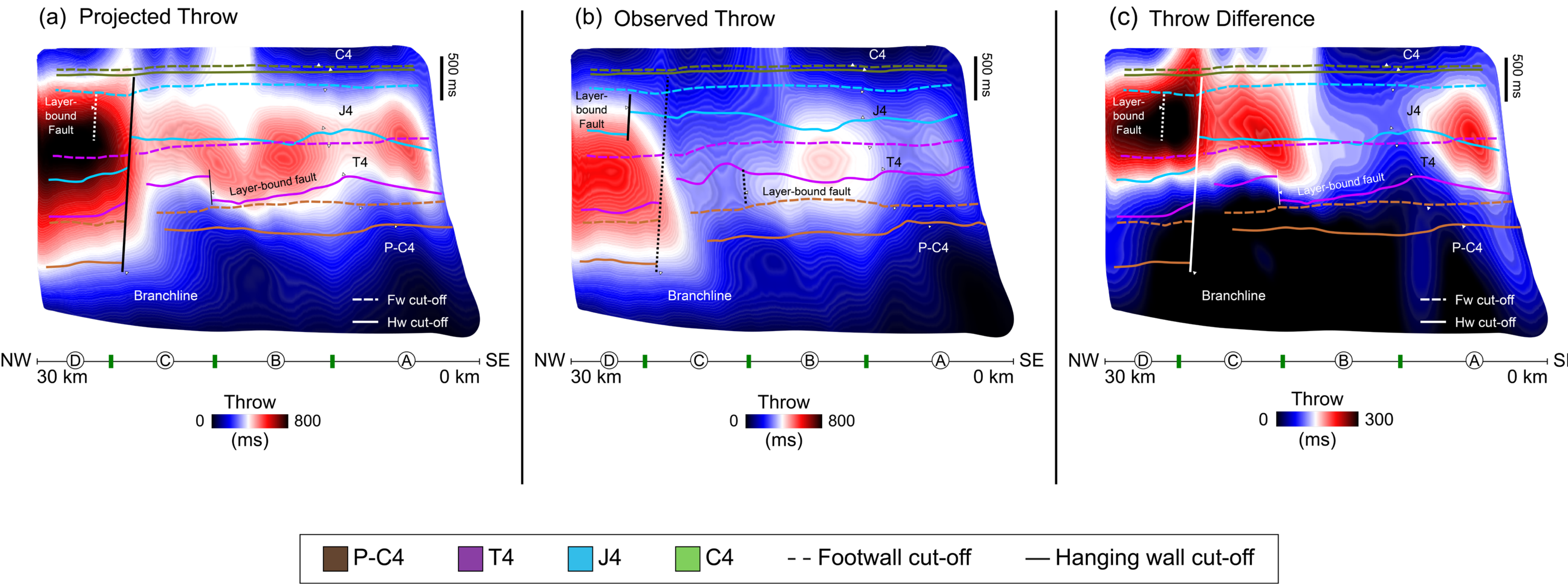


Figure 8. Strike-projected throw distributions along the TFFC surface for projected (a) and observed (b) throws. The two strike-projections show increased throw towards the NW part of the fault and local throw maximums along strike of the fault. The location of the local throw maximums coincides with the center of each fault segment making the TFFC structure. The throw maximum the NW corresponds with a branchline that is interacting with the TFFC. The image on the right (c) shows the throw difference between projected and observed throws and represent the folding component along the TFFC, which is shown to be restricted in the sedimentary cover above basement. TFFC fault segments are marked by the green lines and annotated A – D respectively.

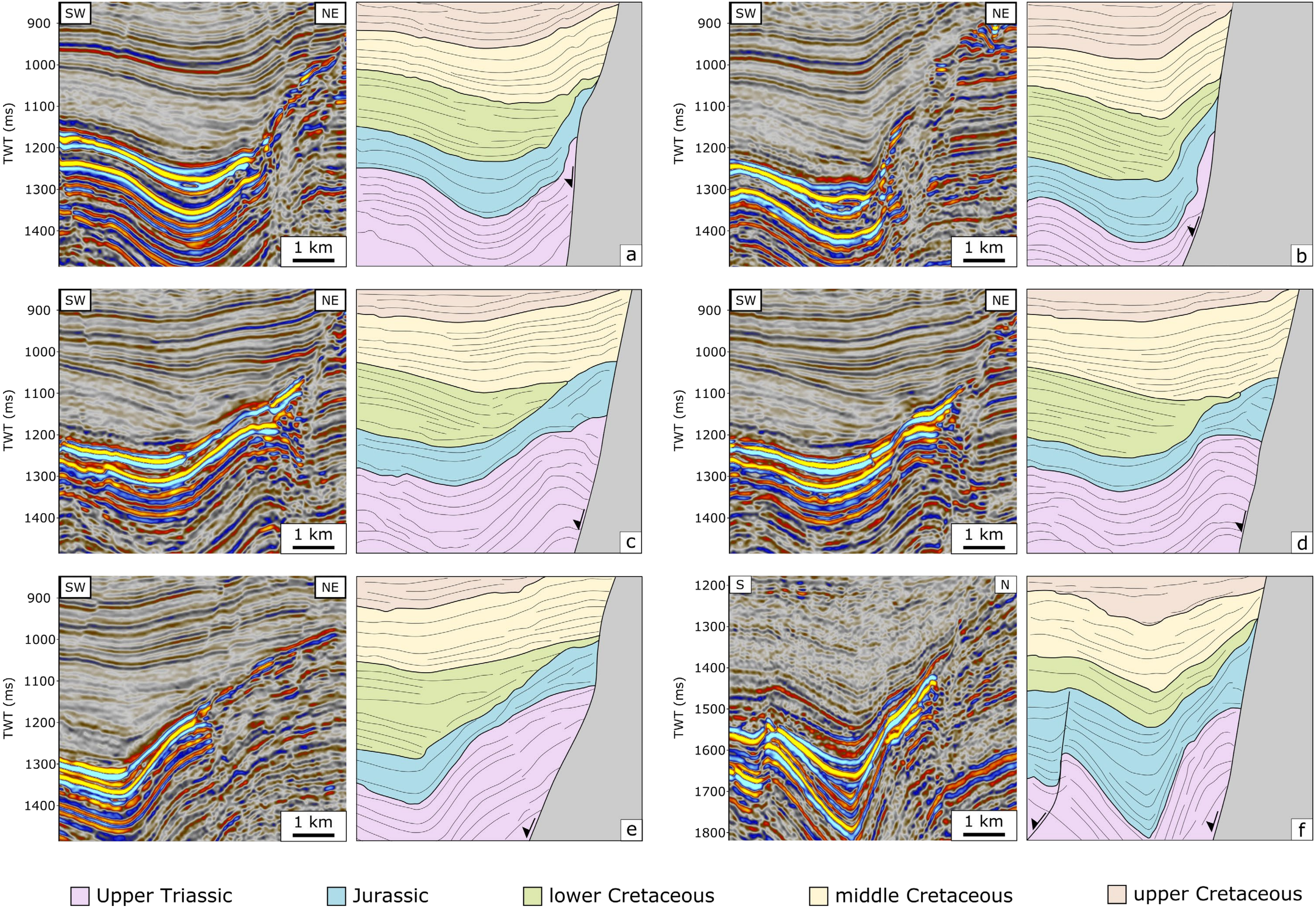
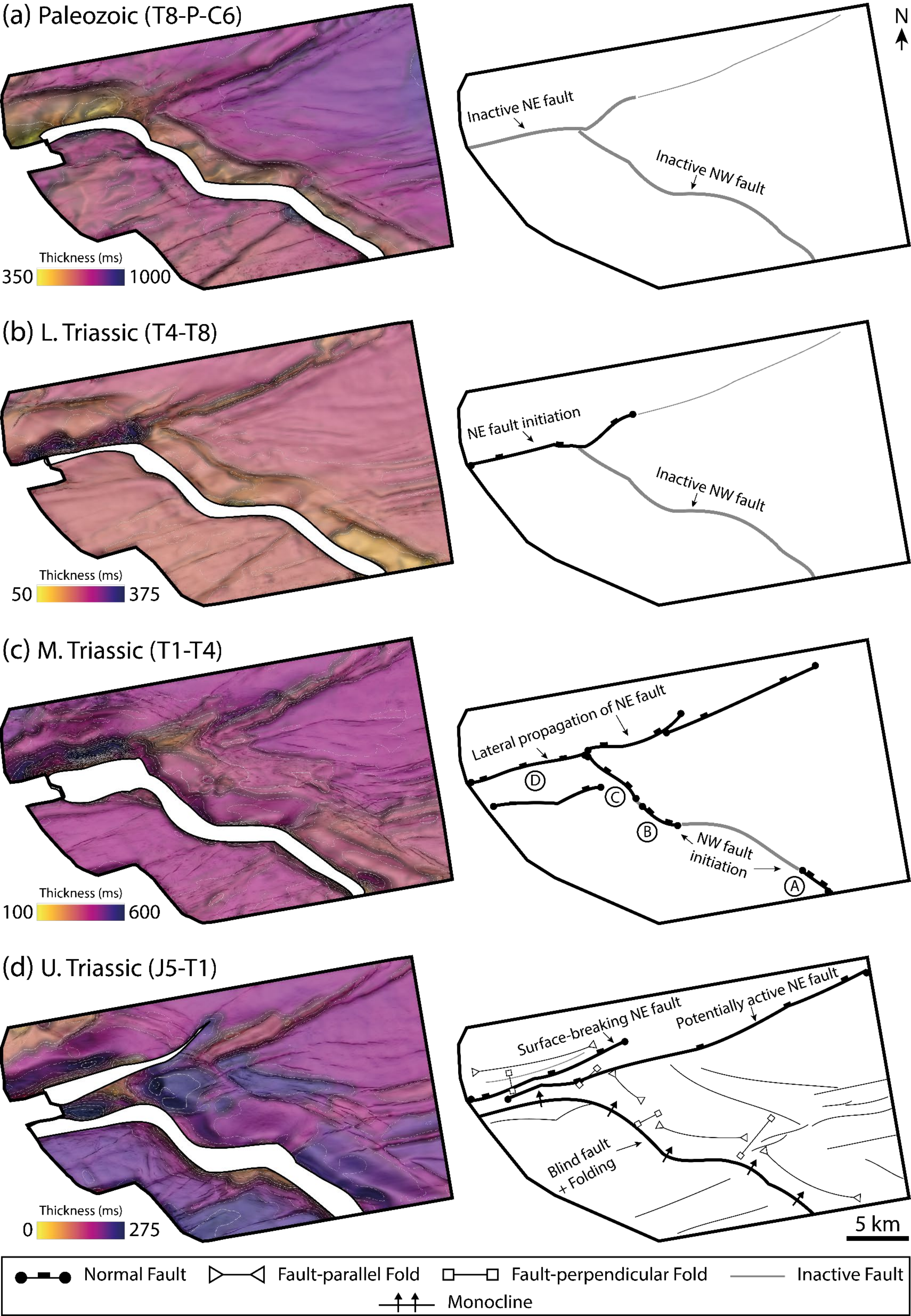


Figure 9. Un-interpreted and interpreted, vertically exaggerated (x10) versions of the seismic sections shown in Fig.5 that highlight the folding and onlap relationships in the U. Triassic – U. Cretaceous stratigraphy in the hanging wall of the TFFC.



(a) Paleozoic (T8-P-C6)
Thickness (ms)
350
1000
N
Inactive NE fault
Inactive NW fault
(b) L. Triassic (T4-T8)
Thickness (ms)
50
375
NE fault initiation
Inactive NW fault
(c) M. Triassic (T1-T4)
Thickness (ms)
100
600
Lateral propagation of NE fault
D
C
B
A
NW fault initiation
(d) U. Triassic (J5-T1)
Thickness (ms)
0
275
Surface-breaking NE fault
Potentially active NE fault
Blind fault + Folding
5 km
Normal Fault
Fault-parallel Fold
Fault-perpendicular Fold
Inactive Fault
Monocline

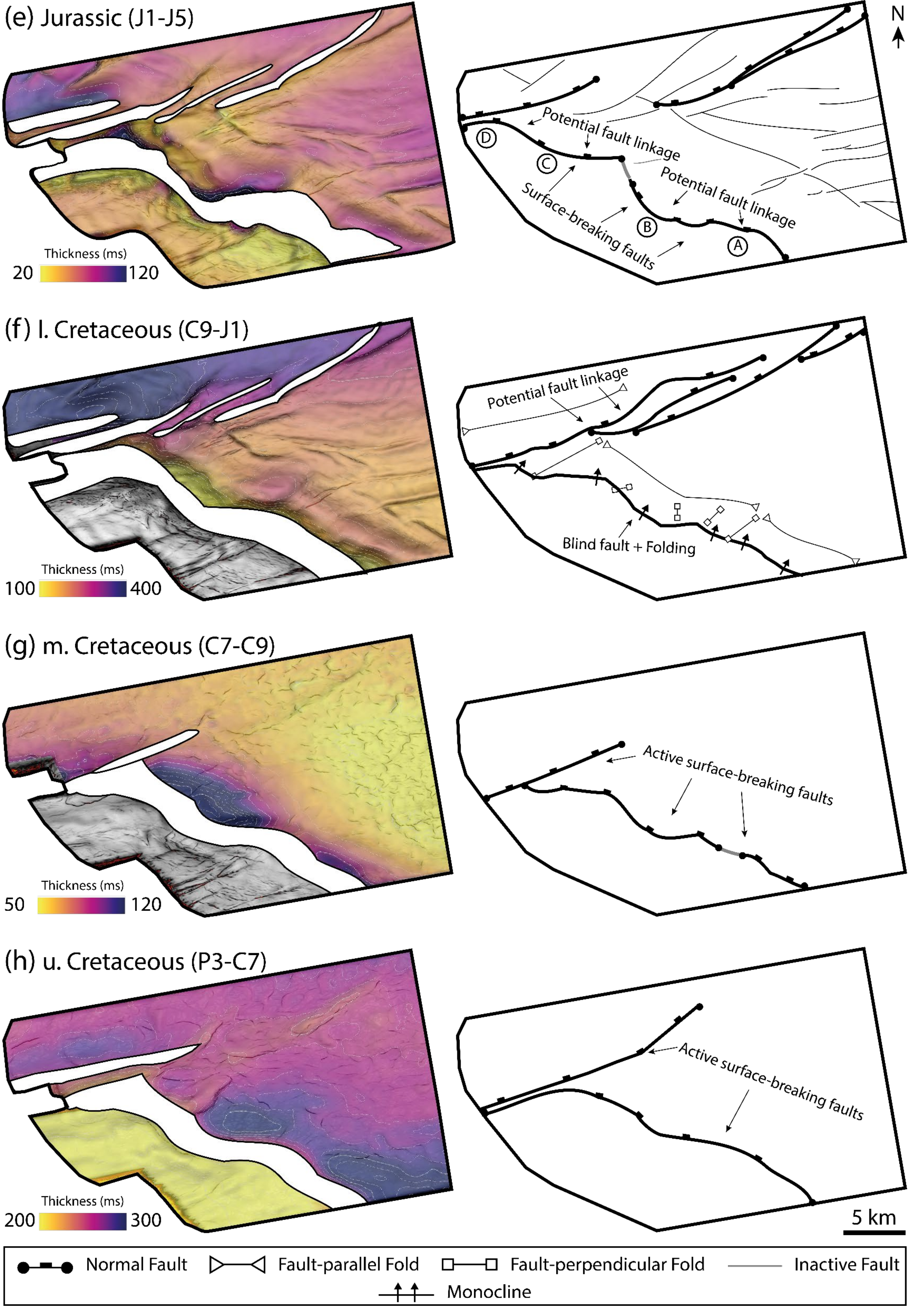


Figure 10. Isochrone maps superimposed on variance attribute surfaces (left) and interpretive sketches (right) for key stratigraphic units across the TFFC. The isochrone maps and accompanying sketches illustrate the thickness variations across the stratigraphic unit and constrain the timing of fault activity in the TFFC and adjacent faults. The location of individual TFFC fault segments is marked by the labelled circles A – D.

Seismic well tie workflow

In this study, we have a 3D seismic reflection survey and one wellbore. The wellbore has available wireline logs that include density and sonic logs that can be used for seismic well tie. The sonic log data cover a depth interval of c.440 m to total depth (2730 m). While the density log data were measured from c. 1200 m to c. 2700 m (Fig 2). However, the wellbore data did not include any checkshot data or a time depth relationship. Therefore, we followed the workflow described next to establish a time depth relationship between the wellbore and seismic reflection data at the well location.

The first step we did was to interpret key horizons in our seismic data that marked major changes in seismic facies or character. On this step, the horizon interpretation was done on a 64x64 grid (in-line, cross-line spacing), which was followed by 3D auto-tracking. This was followed by manual edits on complex areas or areas that the auto-tracking did not perform well. Next, we performed a qualitative seismic stratigraphic correlation with offset wells from nearby studies, where we tried to correlate major seismic units based on their overall attributes (i.e., amplitude strength, frequency, lateral continuity and geometry of reflectors). These studies are Harishidayat et al. (2015), Mohammedyasin et al. (2016) and Torabi et al. (2019). Wellbore 7125/4-2 used in Harishidayat et al. (2015) and Torabi et al. (2019) is located c. 34 km SE of our study area while wellbore 7121/4-1 from Mohammedyasin (2016) is situated c. 104 km away towards the west. Representative seismic sections used for this correlation are shown in Fig 3.

After establishing a seismic stratigraphic correlation with offset wells, we proceeded to determine a time-depth relationship at the wellbore location in our study area and calculate a pseudo seismic velocity log. This served as a first iteration towards constraining the age relationships in our study area and reaching a reliable seismic-well tie. We only used this pseudo velocity log as a comparative measure for other velocity estimates to make sure that we are using reasonable values and not completely off with our estimations. The second step we took was to generate a simplified layer cake velocity model using our key seismic surfaces that represented the top of each key seismic unit along with interval velocity values calculated from the average sonic log response the corresponds with each seismic unit. Where sonic log data are missing, we used geologically reasonable seismic velocity estimates that accounted for lithology and depth. From this velocity model, we derived a second time-depth relationship. We also calculated a third time-depth relationship using the previous approach but after applying a median filter to the sonic log data to remove any outlier data points. Together, these velocity modelling steps resulted in two time-depth relationships that we later used as pseudo checkshot data in the seismic-well tie process together with a reference time-depth relationship that was calculated from offset wells and used as a quality check measure.

We followed a two-step workflow to perform the seismic-well tie at the wellbore location in our study area. First, we applied a median filter to the density log data to remove any spikes or outlier data points. Then, we used the filtered density and sonic logs to generate a synthetic seismogram, which we combined with the time-depth relationship (derived from the filtered sonic log data) to tie the well tops to the seismic data. Given the limited coverage of the density log compared to the sonic log, this seismic-well tie step could only tie the well tops from c. 1200 m (Top Hekkingen Formation) to c. 2600 m (Top Havert Formation). Nonetheless, this was a useful step to get a more accurate time-depth relationship that is well-constrained for those deeper horizons. The second step in our efforts to get a reliable seismic-well tie at our area of interest was to use the time-depth relationship from the last step and combine it with a synthetic seismogram that was generated from sonic log data and a computed density log using sonic log data to perform a final seismic-well tie. This newly generated synthetic seismogram can be used to tie the shallower well tops (c. 440 m) in addition to the deeper well tops that were covered by the first seismic-well tie step.

As a result of this iterative velocity modelling and seismic-well tie method, we were able to tie all available well tops to our seismic data using a well-constrained time-depth relationship that integrates all available data. Lastly, to close the loop and increase our confidence with the seismic-well tie, we compared the final seismic-well tie results that were derived completely from our available data to offset wells from the nearby studies.

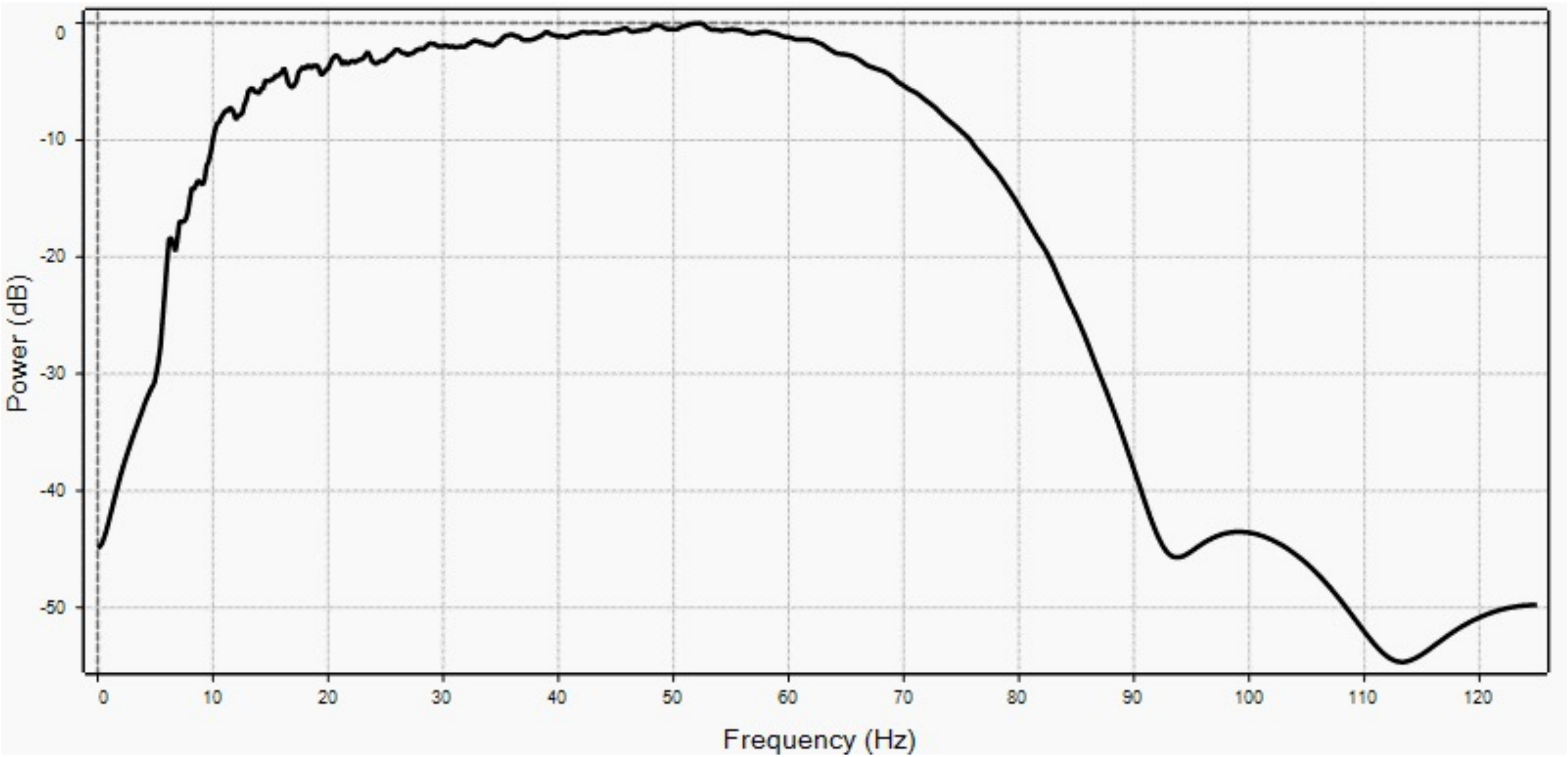


Figure 1. Frequency spectrum for the 3D seismic reflection dataset used in this study. The figure shows that the dominant frequency ranges between 40 – 60 Hz depending on the depth interval within the seismic survey.

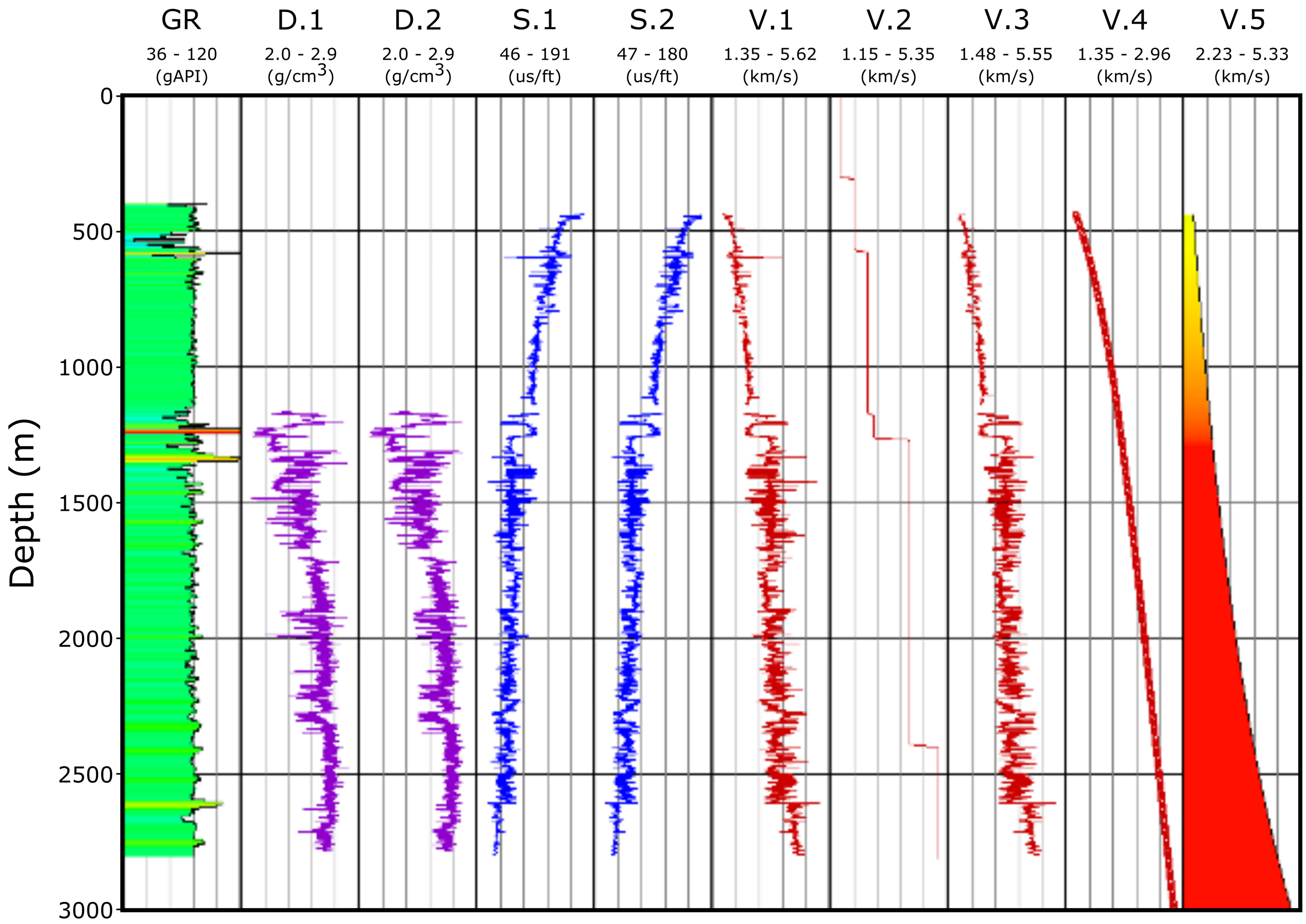
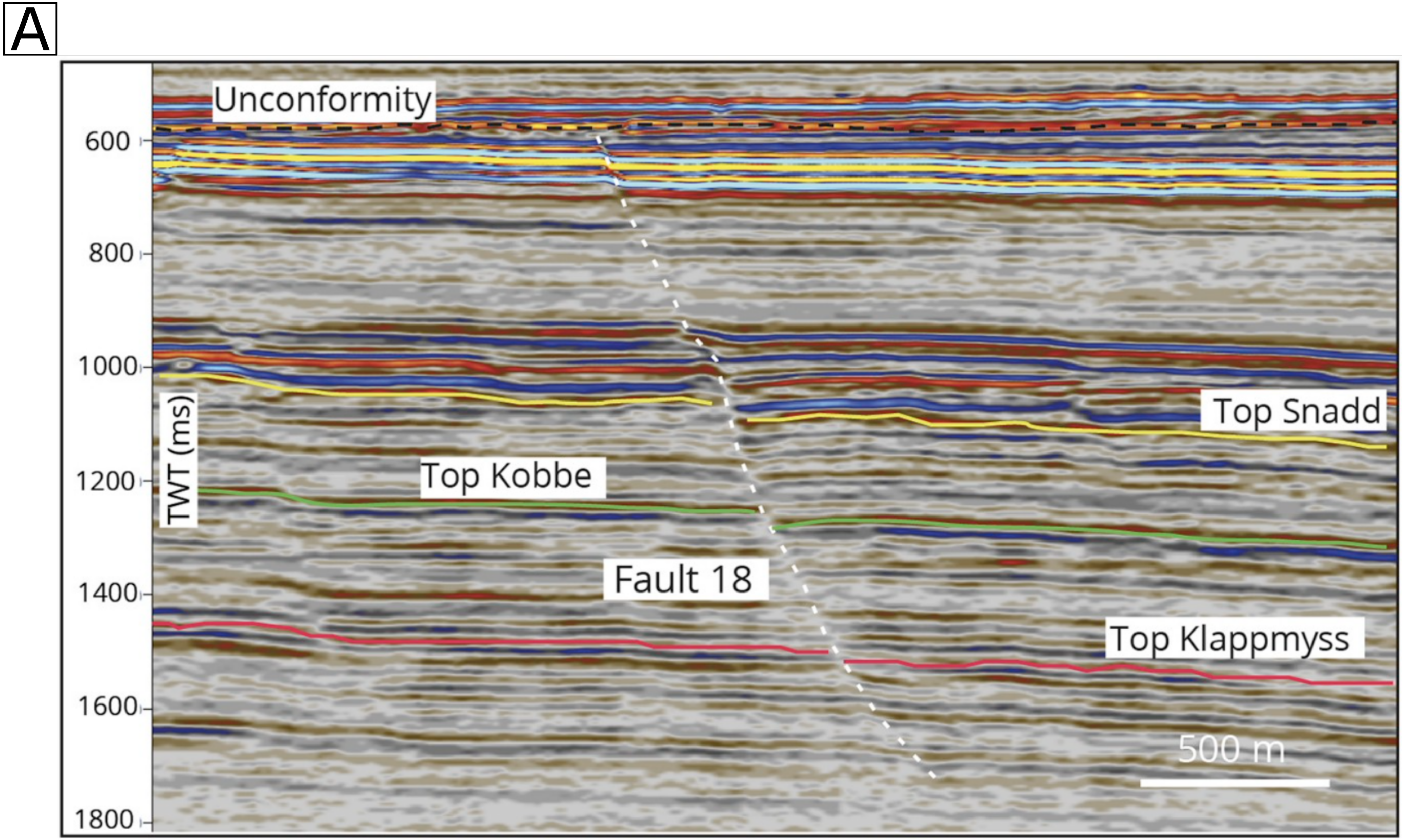
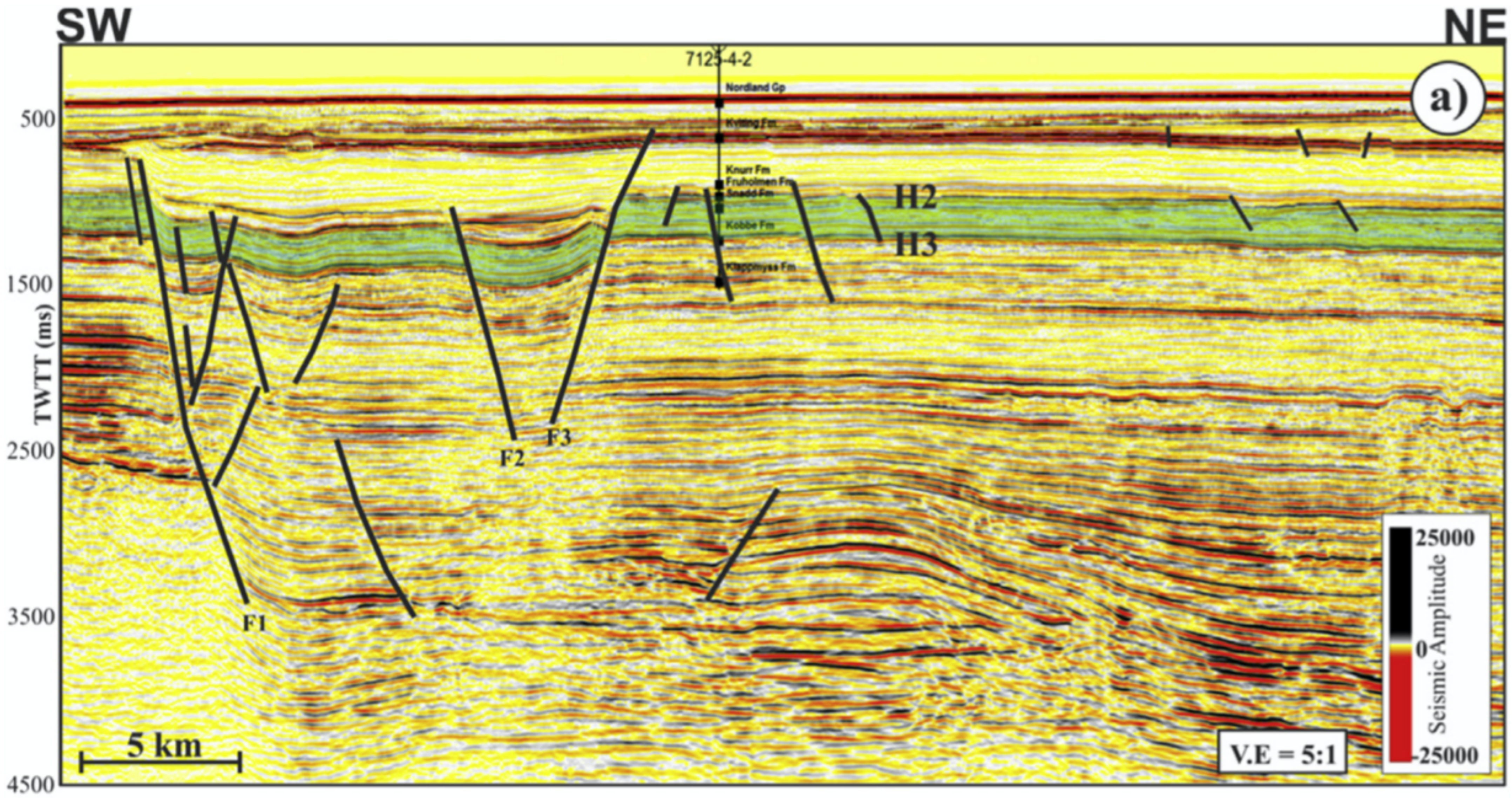


Figure 2. Well section view showing wireline and calculated logs for wellbore 7124/4-1S. The displayed data are for gamma ray (GR), density log (D.1), filtered density log (D.2), sonic log (S.1), filtered sonic log (S.2), calculated velocity from sonic log (V.1), modelled interval velocity using simplified geological model (V.2), calculated velocity from filtered sonic log (V.3), calculated average velocity using estimated time-depth relationship from first seismic well tie (V.4), calculated pseudo interval velocity using estimated time-depth relationship from second seismic well tie (V.5).

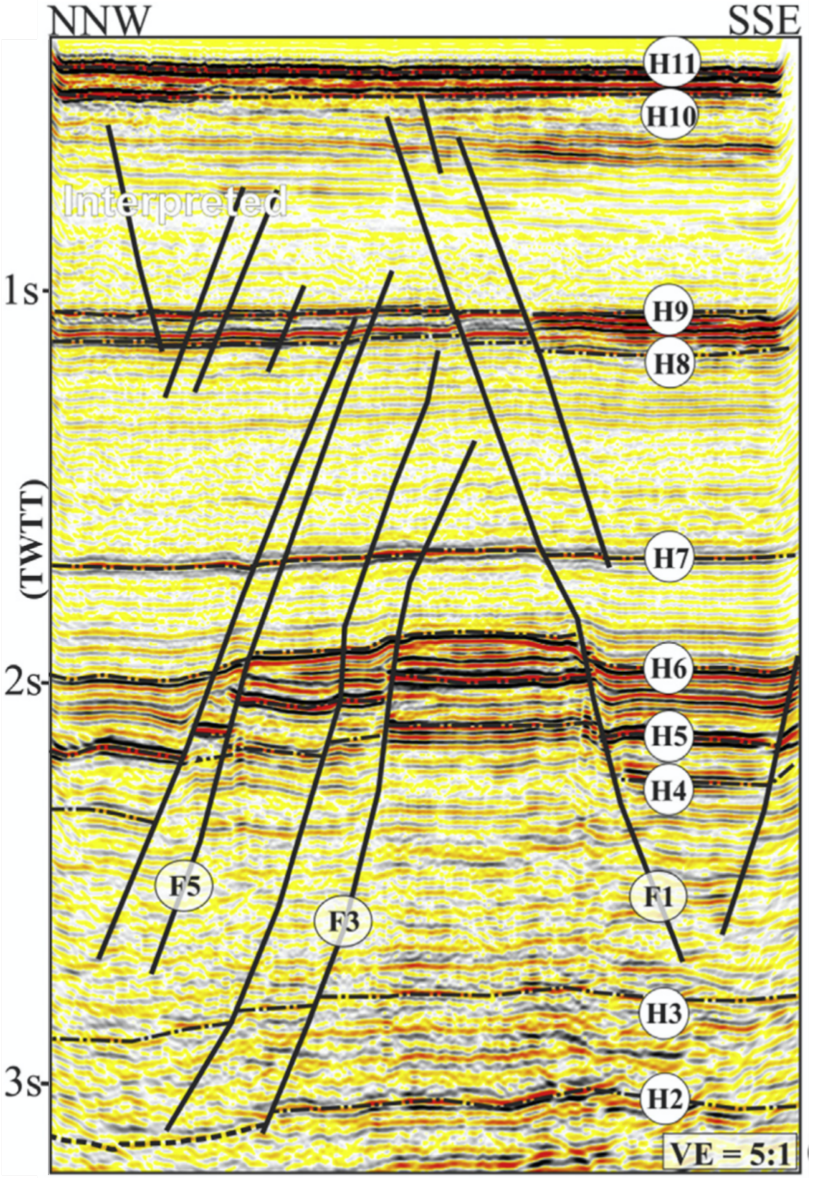

A
Unconformity
600
800
1000
1200
1400
1600
1800
TWT (ms)
Top Snadd
Top Kobbe
Fault 18
Top Klappmyss
500 m

B



SW
NE
7125-4-2
Nordland Gp
Kviting Fm
Knurr Fm
Fruholmen Fm
Snadd Fm
Kobbe Fm
Klappmyss Fm
a)
H2
H3
F1
F2
F3
500
1500
2500
3500
4500
TWTT (ms)
5 km
V.E = 5:1
25000
0
-25000
Seismic Amplitude

C


NNW
SSE
Interpreted
H11
H10
H9
H8
H7
H6
H5
H4
H3
H2
F5
F3
F1
1s
2s
3s
(TWTT)
VE = 5:1

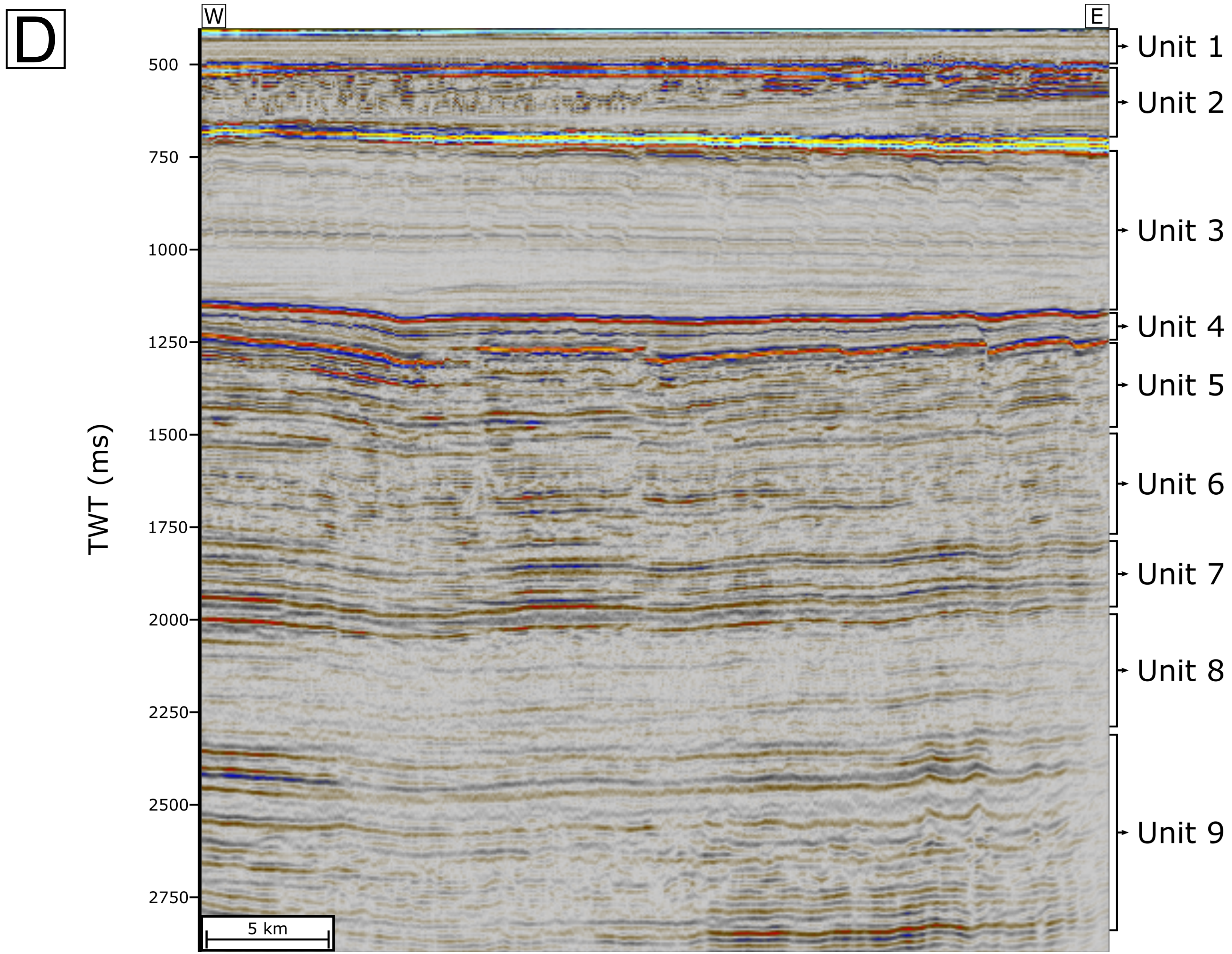


Figure 3. Representative seismic sections used for qualitative seismic stratigraphic correlation with offset well (A: Torabi et al., 2019; B: Harishidayat et al., 2015; C: Mohammedyasin et al., 2016; D: our study area).

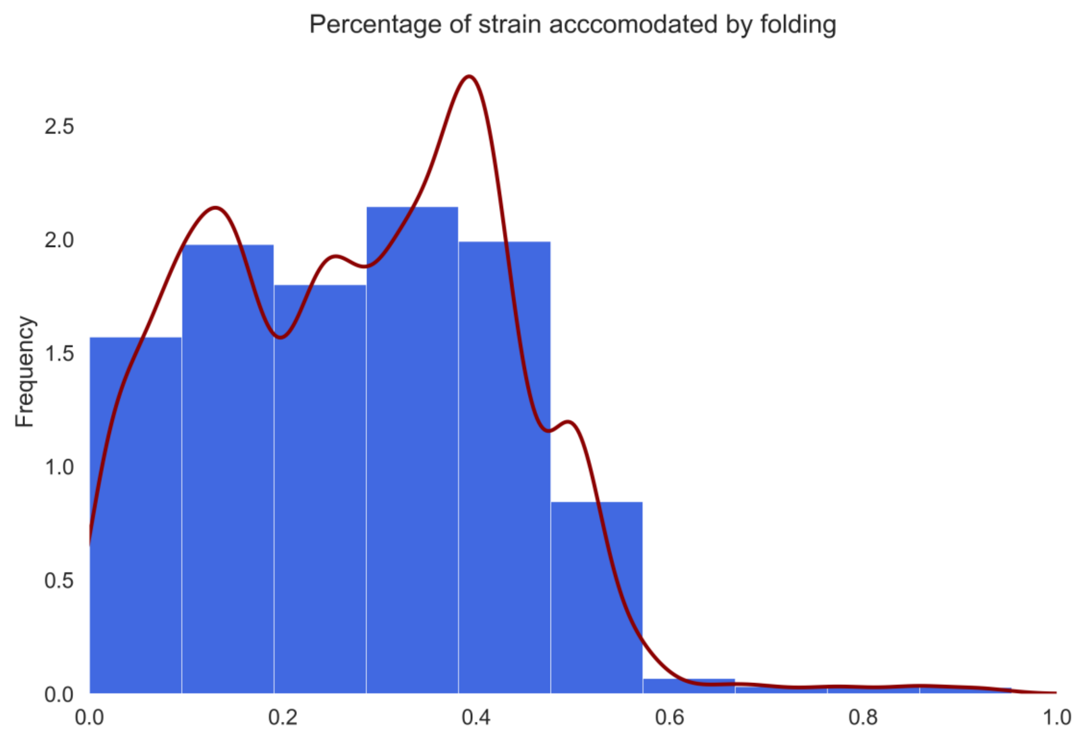


Figure 4. A histogram showing the distribution of strain accommodated by folding. The plot shows that 10 – 40% of total extensional strain is accommodated by the formation of fault-parallel and fault-perpendicular folding.