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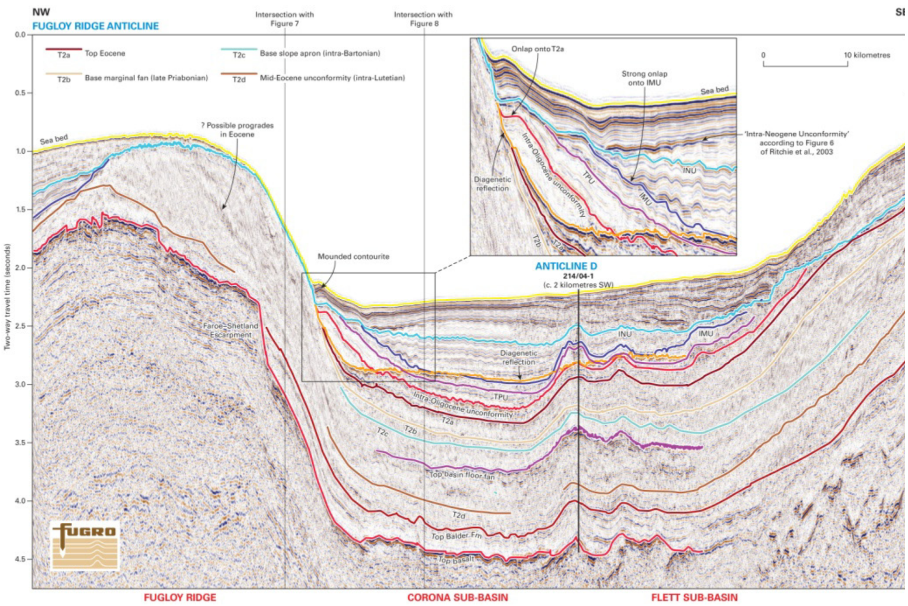
Cenozoic pre- and post-breakup compression in the Faroe-Shetland area, within the context of the NE Atlantic

Marine Geoscience Programme

Commissioned Report CR/12/017



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Cenozoic pre- and post-breakup compression in the Faroe-Shetland area, within the context of the NE Atlantic

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Front cover

Geoseismic profile illustrating tie to well 214/04-1: see Figure 5 for details.

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Foreword

This report is the result of a joint study by the British Geological Survey (BGS) and Jarðfeingi (the Faroese Earth and Energy Directorate) on behalf of the Faroe-Shetland Consortium (FSC) and presents a regional analysis of Cenozoic pre- and post-breakup compression in the Faroe-Shetland area. The study is based mainly upon 2D and some 3D seismic data provided by the Consortium members and other supporting organisations. The work was conducted in parallel with an FSC study of Eocene (Stronsay Group) tectonostratigraphy (Stoker et al. 2012). A new seismic interpretation of both the UK and the Faroese sectors is presented in the form of geoseismic profiles, structure contour maps and isochore maps. These outputs are utilised to consider the spatial and temporal development of Cenozoic compression in the Faroe-Shetland area within the context of the tectonic framework of the NE Atlantic margin.

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H Johnson	Reporting, seismic interpretation, task and project management
M F Quinn	Major contribution to map outputs
G S Kimbell	Gravitational potential energy section and contribution to integration and analysis of seismic mapping and 3D gravity model
M S Stoker	General contribution to understanding Faroe-Shetland Cenozoic development and major contribution to Eocene tectonostratigraphy
K Smith	Major contribution to Eocene tectonostratigraphy
J Ólavsdóttir	General contribution to understanding of Cenozoic of the Faroese sector, and provision of initial Faroese seismic interpretation
T Varming	General contribution to understanding of Cenozoic of the Faroese sector, and provision of initial Faroese seismic interpretation

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Summary

This report is primarily based upon the interpretation of oil industry 2D seismic data, and aims to elucidate aspects of Cenozoic tectonostratigraphic development in the Faroe–Shetland region, especially with regard to post-breakup compression. Evidence of Cenozoic and Late Cretaceous pre-breakup compression and deformation is briefly reviewed. We have utilised established seismo-stratigraphic frameworks and a recently updated scheme for the post-breakup Eocene (Stronsay Group) succession, which are largely based upon the recognition of units bounded by regional unconformities. The seismic expression, extent and thickness of the seismo-stratigraphic units are illustrated by geoseismic profiles, structure contour maps and isochore maps, which are used to analyse the spatial and temporal development of post-breakup compression and deformation within the Faroe-Shetland region.

The Faroe-Shetland region records a complex spatial and temporal pattern of departures from the thermal subsidence normally associated with passive margins, including broad uplifts and accelerated basinal subsidence together with fold development up to kilometre scale. The phases of latest Eocene / earliest Oligocene ‘sagging’ (accelerated subsidence) and early Pliocene uplift and exhumation (tilting) appear to be coeval with compression. Indeed, compression appears to have been active throughout post-breakup times, although the loci of deformation have varied both spatially and temporally. Conceivably, some of the large scale sagging, tilting and uplift may be associated with lithospheric folding.

Much of the intra-Eocene folding appears to be focused in the southwestern part of the Faroe-Shetland region, around the Munkagrunnur Ridge and Judd area, where phases of shelf progradation are preserved and may be associated with contemporaneous uplift. However, there also appears to be evidence of episodic intra-Eocene and younger uplift in the area around the northern Fugloy Ridge. The overall shaping of the Faroe-Shetland Channel appears to have been initiated at the end of the Eocene, associated with uplift on the Fugloy Ridge and Faroe Platform areas, and with accelerated subsidence in the Faroe-Shetland Basin; this shaping was further developed during the Neogene. A Neogene opening of the ‘Faroe Conduit’ oceanic gateway is favoured on the basis of regional evidence of faunal isolation and restricted environment of deposition together with uncertainty regarding the nature of the ‘Southeast Faroes drift’.

A significant phase of Miocene folding is associated with the Intra-Miocene Unconformity (IMU), whereas the Mid Miocene Unconformity (MMU) represents a relatively minor break with a restricted distribution in the NE Faroe-Shetland region. Seabed relief on some folds and late Neogene seismic onlaps may indicate that fold development persisted into Recent times. Lateral offsets and local basin inversion associated with the folding, suggest a strong structural inheritance from the underlying rift architecture.

A broad coincidence between the timing of formation of the unconformities and plate reorganisation events in the adjacent Norway Basin and wider region may suggest that these events made important contributions to the forces shaping the margin. The development of Miocene and younger folds may have been influenced by gravitational potential energy / body forces associated with the density structure of the Iceland Insular Margin and the Southern Scandes, or with modulations to ridge-push resulting from transient changes in ridge elevation associated with plume-related temperature (buoyancy) variations in the underlying asthenosphere. Far field stresses associated with, for example, collision between Eurasia and Iberia may also have exerted significant influence on deformation within the Faroe-Shetland region.

1 Introduction

The post-rift development of the NE Atlantic margins is generally classed as passive. However, the configuration of the NW European margin has been significantly modified by late Palaeogene and Neogene epeirogenic and compressive movements, indicating an evolution that has been anything but passive (e.g. Doré et al., 1999, 2002; Ritchie et al., 2003; Stoker et al. 2005a, b). An understanding of Cenozoic pre- and post-breakup compression and associated tectonic movement, within the Faroe-Shetland region has considerable relevance to the spatial and temporal aspects of the petroleum prospectivity. The impacts on exploration are potentially profound and may include uplift and the cessation of active maturation, embrittlement of seals, creation or destruction of overpressures and creating or amplifying traps (Levell et al., 2011).

1.1 SCOPE AND OBJECTIVES

This project was designed primarily to provide an overview of the distribution, orientation and timing of growth of compressional and related structures within the Faroe-Shetland study area within the context of a joined-up seismic interpretation from both the UK and Faroe Islands national sectors. Key aspects of our analysis are demonstrated through geoseismic profiles together with associated structure contour and isochore maps (in two-way-travel time). A diagenetic horizon (Davies and Cartwright, 2002), which cuts across and locally obscures imaging of the stratigraphic horizons was also interpreted on seismic profiles, but no structure contour map was generated. The maps associated with this study will be incorporated into the ArcGIS database that is being created on behalf of the Faroe-Shetland Consortium. The age of post-breakup fold development is assessed by conventional seismic-stratigraphic methodology (e.g. Mitchum et al., 1977), constrained where possible by available released well and borehole data. We have also utilised criteria for the recognition of syndepositional compression on seismic profiles (Pereira et al., 2011).

The study has benefited from integration with a parallel FSC project regarding the tectonostratigraphic development of the post-rift Eocene succession (Stoker et al., 2012). Results from 3D gravity modelling (Kimbell et al., 2010) are used to aid in the evaluation of the contribution that regional gravitational potential energy variations may have made to the post-breakup compressional deformation and to review the importance of post-breakup structural reactivation.

1.2 DATA AND METHODOLOGY

This seismic interpretation study covers both the UK and Faroe Islands national sectors and is primarily based upon 2D seismic datasets (Figure 1), although some 3D seismic data were also utilised. Unfortunately, a large 3D seismic data set located in Quadrants 204 and 205, was withdrawn from the FSC project, because it may include some non-oil company proprietary data. Well data were made available to the study via CDA, although for many parts of the Cenozoic succession velocity logs / seismic calibration logs are not available (Robinson 2004).

A BGS / Jarðfeingi collaborative workshop, held in Torshavn during January 2011, highlighted the need for a complementary study of Eocene tectonostratigraphy in parallel with the FSC ‘Compression Project’. In May 2011, the FSC Steering Committee prioritised such a study (Stoker et al. 2012). Integration of seismic interpretations from the UK and Faroe Islands national sectors was facilitated by the exchange of data and interpretations (e.g. the OF94 and OF95 surveys) and through a collaborative workshop to agree the interpretation, which was held in Torshavn during August 2011.

1.3 REGIONAL SEISMO-STRATIGRAPHIC FRAMEWORK

A number of stratigraphical frameworks for the Cenozoic of the Faroe-Shetland Basin have been proposed; this study is primarily based upon the seismo-stratigraphical scheme described by Stoker and Varming (2011), but with significant revision of the post-breakup Eocene (i.e. post-Balder Formation) (Stronsay Group) stratigraphic framework following Stoker et al. (2012) (see Table and section 4).

2 Regional setting

2.1 PLATE TECTONIC SETTING

It is generally accepted that a triple junction between the Eurasia, Greenland and North American plates developed in the Late Paleocene (c. 54 Ma) when spreading initiated between Greenland and Eurasia, while seafloor spreading was still active in the Labrador Sea, between Greenland and North America (e.g. Chalmers and Laursen, 1995) (Figure 2). Spreading on the two arms of the North Atlantic drove Greenland north relative to the Barents Sea and Canadian Arctic margins, and resulted in the Eurekan and West Spitsbergen orogenic belts (Doré et al., 2008). The triple junction was active until about 33 Ma (chron 13, earliest Oligocene), when sea-floor spreading in the Labrador Sea ceased completely (Roest and Srivastava, 1989). This event has been considered as a trigger of major changes in the Arctic-North Atlantic region and led to the establishment of a continuous plate boundary linking the NE Atlantic and the evolving Eurasia Basin. However, detailed analysis of geophysical data suggests that several events affected the NE Atlantic between breakup and the final reorganisation at chron 13 when the triple junction became extinct and Greenland became part of the North American plate (Gaina et al., 2009). On a regional scale, the plate boundary between Eurasia and Greenland seems to be the result of a two-plate system, but closer inspection of geophysical data and plate geometry has shown the existence of short-lived additional plate boundaries (Gaina et al., 2009) (Figure 3).

Many workers (e.g. White 1988; White and Lovell, 1997; Jones et al., 2002; Smallwood and Gill, 2002) interpret the Iceland hot spot as a mantle plume beneath the lithosphere, which provided support until continental breakup. The coeval extensive volcanism has also been linked to the interaction between rifting and the Iceland mantle plume (e.g. Smallwood and White, 2002). An alternative view is that the hot spot anomaly is an upper mantle response to plate breakup (Doré et al., 1999; Foulger and Anderson, 2005). Episodes of magmatic underplating associated with mantle plume activity have been postulated as an important mechanism for driving permanent regional surface uplift, denudation and the shedding of large amounts of clastic sediments into surrounding basins (e.g. White and Lovell, 1997). Estimates of the magnitude and duration of transient Cenozoic uplift have been used to calculate the temperature and velocity of spreading ‘ripples’ of hotter plume material (e.g. Hartley et al., 2011).

Post-rift thermal subsidence within the Faroe-Shetland Basin and surrounding area was interrupted by significant phases of accelerated subsidence, uplift and compression (e.g. Stoker et al., 2005a, b). The driving mechanisms causing such departures from normal passive margin subsidence have been the topic of much research. In particular, the relative control of plate boundary forces associated with changes in the developing ocean basin and surrounding plates versus mantle plume influences continues to be debated. Commonly, the observed deformation has been attributed to a combination of number of these factors, and the timing of events is regarded as key to evaluating their relative importance.

2.2 PREVIOUS RESEARCH ON PASSIVE MARGIN DOMES

Doré et al. (2008) reviewed the development of domes on the NE Atlantic margin and distinguished two classes. Firstly, domes of Late Cretaceous to Paleocene age of tectono-magmatic origin and related to intrusion or to the emplacement of remobilised crust or magmatically underplated material at depth. An example within the Faroe-Shetland region may be the dome associated with the Brendan Volcanic Centre (Hodges et al., 1999; Rohrman, 2007) (Figure 1). Secondly, domes of Early Eocene to Recent age, which comprise compressional-compaction domes.

Among the first workers to report the importance of basin inversion and / or transpression within the Rockall-Faroe region were Roberts (1989) and Earle et al. (1989). Roberts (1989) noted that the Palaeogene lavas of the Wyville Thomson Ridge are strongly folded and considered a phase of Oligocene inversion to be important in basins of the Western British Isles. Earle et al. (1989) described Late Cretaceous to Oligocene inversions concentrated in and near a cross-cutting NW-SE-trending tectonic element lying north-west of the Orkney Islands, which they termed the Orkney Faroe alignment. They inferred that these inversions were generated by transpression associated with strike-slip on a fault which is part of the Orkney-Faroe alignment and suggested that inversion took place in several minor episodes during the Late Cretaceous to Paleocene and one major episode in the mid-Oligocene.

On the basis of interpreted seismic profiles, Boldreel and Andersen (1993) postulated the importance of compression rather than transpression in the Faroe-Rockall area, including the Wyville Thomson and Ymir ridges, which they interpreted as ramp anticlines. They postulated at least three, *post-basalt* phases of compression and considered their development to be linked to pronounced changes in sea floor spreading geometries in the north Atlantic. They reported a late Paleocene / early Eocene compressional phase, which possibly related to NE-SW to ENE-WSW-oriented stress. An Oligocene phase was interpreted to result from N-S directed compressional stress and was possibly associated with movement on conjugate shear zones. The final phase of compressional stress was from the NW and was interpreted to be of mid or late Miocene age.

Doré and Lundin (1996) and Doré et al. (1997) described the regional extent of Cenozoic compressional and inversion structures in the NE Atlantic and focused on structures in the Norwegian Sea, where they recorded important growth phases in the mid Eocene to early Oligocene and in the Miocene. They postulated that development of the inversion structures was linked to both NE Atlantic spreading and Europe-Africa convergence. The compressive structures in mid-Norway were believed to have a transpressive character, conforming to minor sinistral movement along pre-existing NW-SE transfer zones. They also noted that the compressional structures are frequently expressed in the present-day seafloor relief and in the case of the Faroe Islands are probably responsible for their present subaerial exposure.

Blystad et al. (1995) suggested that some of the major anticlines in the mid-Norway region are mainly the result of tectonic reactivation along cross-cutting lineaments such as the Jan Mayen Lineament, but that the Helland-Hansen Arch may be mainly the result of differential subsidence between the eastern and western Vøring Basin and Late Pliocene loading. However, the Vema and Naglfar Domes and part of the Gjallar Ridge were considered to be inversions due to magmatic underplating (e.g. Skogseid and Eldholm, 1989 and Skogseid et al., 1992). In a detailed study of the Helland Hansen Arch and Vema Dome in Mid-Norway, Gómez and Vergés (2005) combined backstripping and unfolding methods to quantify the relative contributions of tectonic growth versus compaction in the development. They concluded that a small but protracted phase of compression affected the Mid-Norway margin and accounted for a minimum of 27-37% respectively of total dome amplitude the Helland Hansen and Vema Dome structures. Brekke et al. (2001) commented that two main phases of compression are associated with the formation of intrabasinal domes and arches in the Vøring Basin and around the Faroe Islands, dated to latest Eocene/ earliest Oligocene and Mid Miocene. In contrast, Vågnes et al. (1998) proposed that the compression was continuous from Eocene to Miocene without discrete phases.

Lamers and Carmichael (1999) postulated that plate reorganisations in the NE Atlantic throughout the Cenozoic reactivated major NW-SE trending transfer zones as strike-slip faults and produced spectacular inversion structures. They suggested that the first compressional event occurred in the Early Eocene and that major inversions took place in the Mid Eocene, Oligo-Miocene and in the Pliocene. In contrast, Tate et al. (1999) interpreted seismic, well, borehole and potential field data to interpret the WNW-ESE-trending Wyville Thomson Ridge as a late Eocene to Oligo-Miocene fault propagation fold, which developed above a ramp-flat detachment during N-S compression. Tate et al. (1999) noted that the earliest signs of uplift are signalled by some progradational clinoforms on the south side of the ridge and speculated that uplift of the Ridge has possibly continued until Recent times. On the basis of structural interpretation of commercial seismic exploration data, Andersen et al. (2000) postulated a late Eocene to early Oligocene phase of doming of the Faroe Platform, which caused post-depositional tilting of Eocene strata; they also recognised a Mid Miocene phase of compression and a possible Pliocene uplift of the Faroe Islands. Using similar data, Sørensen (2003) postulated Mid Eocene uplift within the Faroes part of the NE Atlantic, which was followed by a phase of uplift and sea level fall in the Late Oligocene, and a final phase of Neogene folding and uplift of the Fugloy Ridge.

Until the late 1990's, the stratigraphy of the Faroe-Shetland Basin was only known from petroleum industry wells drilled on the south-eastern flank of the basin and from a limited number of shallow research boreholes located on the Faroese and Shetland margins (Davies and Cartwright, 2002). Consequently, early interpretations of the history of compression within the Faroe-Shetland region were relatively poorly calibrated. Davies and Cartwright (2002) drew attention to the importance of Mobil well 214/04-1 for establishing the Eocene to Recent stratigraphy within the centre of the Faroe-Shetland Basin and recognised four seismic-stratigraphic units separated by a Late Eocene Unconformity, a Middle Miocene Unconformity and an Early Pliocene Unconformity. Subsequently, the STRATAGEM Partners (2002, 2003) provided the first regional calibration of seismic reflection data with the thinly scattered key boreholes and wells on the NE Atlantic margin, from offshore Norway to offshore Ireland, including well 214/04-1, and presented a new regional seismo-stratigraphic scheme for the Neogene. Ritchie et al. (2003) utilised this regional seismo-stratigraphic framework and described a suite of mainly early to mid-Miocene, NE- to NNE-trending, compressional folds within the NE Faroe-Shetland Basin, and local evidence of growth folding through early Pliocene to Recent times. Davies and Cartwright (2002) and Davies et al. (2004) described the asymmetric Miocene folds within the NE Faroe-Shetland Basin and measured amplitudes of up to 600 m. Davies et al. (2004) attributed the fold development to contractional reactivation of Mesozoic extensional faults.

In the southern part of the Faroe-Shetland Basin, around the Judd / Westray High, Smallwood (2004) utilised 3D and 2D seismic reflection data to characterise Tertiary compression, inversion and erosion. In particular, he identified four phases of compressive deformation during Latest Ypresian, Latest Lutetian, Late Lutetian and Pre-Top Palaeogene unconformity, although some of these phases have been recalibrated (Stoker et al., 2012). Smallwood (2004) reported that the latest Ypresian and the late Lutetian phases of compression are associated with unconformity development and speculated that time-dependent convection variations in the Iceland plume may be influencing the compressional stresses which form the inversion structures.

Ritchie et al. (2008) noted that the scale and orientation of the Cenozoic compressional folds within the Faroe-Shetland Basin and surrounding areas varies significantly, with fold amplitudes up to 2 km and axial traces ranging up to 250 km or more and trends including east, NE, NNE, NW, NNW and WNW. The NE-trending folds are the most numerous, although they are mainly restricted to the NE Faroe-Shetland Basin, where an inherited Caledonian structural grain has been inferred. Ritchie et al. (2008) postulated that compression has resulted in local fold development since the early to mid-Eocene until Pliocene or Recent times. Working principally in the Rockall-Faroe area, but also including the Wyville Thomson Ridge and Judd areas, Tuitt et al. (2010) presented a contrasting interpretation of compression-related folds in the Rockall-Faroe and Judd areas and postulated that compression effectively ceased in mid-Oligocene times.

Doré et al. (2008) reviewed the timing and potential mechanisms for the genesis of the compressional Cenozoic domes and concluded that the structures developed from early Eocene to Recent times, but underwent episodes of greater activity, including a marked compressive phase in the Mid Miocene. Doré et al. (2008) considered the role of various driving mechanisms, such as sedimentary loading, far-field stress, divergent asthenospheric flow, ridge push and the growth of the topographically high Iceland Insular Margin, together with factors which may have served to localise the strain, such as structural inheritance and the reactivation of zones of weakness and weakening related to hotter basement. Lundin and Doré (2011) postulated that long-lived lithospheric weakening and proneness to deformation may be associated with crustal hyperextension.

Interestingly, Davies and Cartwright (2002) suggested that there was little or no compression in the NE Faroe-Shetland Basin during the Eocene and Oligocene, but did remark that the most striking element of the structural architecture in the Faroe-Shetland Basin is the present-day eastward tilt of Palaeogene strata away from the Faroes and the westward tilt of the same succession away from the Shetland Isles toward the central axis of the present-day deep-water channel. Davies et al. (2004) noted that this phase of tilt/differential uplift and subsidence has been observed along the length of the Irish, UK and Norwegian Atlantic margins (e.g. Riis and Fjeldskaar, 1992; Vanneste et al., 1995) and is also associated with km-scale uplift. Japsen and Chalmers (2000) postulated that a late Cenozoic episode of uplift of basin margins and concomitant accelerated subsidence of basin centres adjacent to uplifted landmasses may be related to processes such as emplacement of magma at the base of the crust leading to isostatic uplift, flow of asthenospheric material into active diapirs (e.g. Rohrman and van der Beek, 1996), isostasy associated with glacial erosion, phase changes in the lithosphere due to pressure relief or regional compression of the lithosphere. Stoker et al. (2005a, b) remarked that the NW European margin has been ‘anything but passive’, and recognised three main tectonic phases:

- Late Eocene-early Oligocene sagging or rapid differential subsidence
- Early-mid-Miocene compressive doming
- Early Pliocene tilting / differential subsidence

They inferred that plate forces were the driving mechanism for the compressive phase, but considered these could not account for the larger amplitude of the mid-Cenozoic sagging and the early Pliocene tilting.

More recently, it has been suggested that elevated passive continental margins may represent anticlinal lithospheric folds formed under compression and similar to those discussed by Cloetingh and Burov (2011). If so, the vertical movements are thought to be related to compressive stress in the lower crust and/or mantle and are produced where there is an abrupt, lateral change in crustal or lithospheric thickness (Japsen et al., 2011). It is interesting to speculate that some of these abrupt lateral changes in crustal or lithospheric thickness may relate to conjugate concave downward ‘stretching / thinning / exhumation faults’, which have been predicted from numerical modelling studies and from geological observations from the Alpine Tethys (e.g. Lavier and Manatschal, 2006; Péron-Pinvidic and Manatschal 2010).

3 Pre-breakup compression

Earliest Paleocene transpressional reactivation of many of the major ENE-trending previously normal fault zones, west of Shetland was reported by Booth et al. (1993). They suggested that this tectonism did not result in general inversion of the Solan Basin, as had a previous mid-Cretaceous event in earliest Turonian times. In contrast, this later reactivation phase created several localised, but very ‘intense’ inversion structures, especially where the ENE faults are offset by transfer zones and relays. Booth et al. (1993) illustrated an example in the West Solan Basin (Figure 4) where a ruck fold is developed over two splays of the Solan Fault, which have an overall normal displacement. The fold was created by reverse motion on these splays. The Top Cretaceous is folded, while the Top Balder Formation reflection is unaffected. Furthermore, the first seismic reflection above the Top Cretaceous is interpreted to be folded and onlapped by the overlying sequence, which is undeformed. Given this relationship, the transpression was interpreted to have occurred during a short pulse during earliest Paleocene times. In contrast, Roberts et al. (1999) suggested that there is evidence of Late Cretaceous tectonism on the margins of the Faroe-Shetland Basin, as shown by the anticlinal subcrop to the Turonian along the Clair Ridge and also by compressional structures in the Solan Basin, as reported by Booth et al. (1993). Thus, Roberts et al. (1999) postulated that the Late Cretaceous tectonism created a series of large inversion structures that acted as a regional focus for hydrocarbon migration (e.g. the Clair and Westray Ridges). These structures had significant submarine topography, as shown by basinal onlap. Roberts et al. (1999) noted that seismic profiles within the Faroe-Shetland area also show clear evidence of Late Cretaceous extension and suggested that this pattern of co-eval extension and compression is consistent with regional strike-slip associated with transtension and transpression. They speculated that this pattern of strike-slip swings south-westwards through the Hatton-Rockall Basin, where Neish (1993) reported pre-Paleocene deformation albeit on poor quality seismic data. Similarly, Johnson et al. (2005) reported fold structures in probable Mesozoic strata within the Mid Hatton Bank Fold Complex, the formation of which appears to pre-date the extrusion of Palaeogene lavas and could possibly be Late Cretaceous in age.

Lundin and Doré (2011) also recognised mounting evidence of pre-breakup Late Cretaceous compression in the NE Atlantic margin. They described the mid-Norwegian Cretaceous succession, which does not show simple post-rift subsidence, but is overprinted by a series of Late Cretaceous broad-wavelength regional folds (e.g. Blystad et al., 1995), implying compression. Anticlines, such as the Nyk and Utgard Highs, are faulted at their crests and, while they have been viewed in terms of extensional footwall uplift, they have also been interpreted as compressional (Brekke, 2000). The onset of the broad-scale folding is constrained to latest Turonian. The compression appears to have been interrupted by a phase of Campanian extension (e.g. Ren et al. 2003) followed by renewed compression in Maastrichtian-Late Paleocene time (Lundin and Doré,2011), after which further extension led to Early Eocene breakup. Stoker at al. (2010a) also described similar fold and inverted Cretaceous geometries in the Faroe-Shetland Basin area, and considered these to have developed during Late Cretaceous transpression and transtension.

Roberts et al. (1999) postulated that a hiatus between the Lower Series (Beinisvørð Formation) and Middle Series (Malinstindur Formation) basalts of the Faroes coincides with a first phase of inversion over the Foinaven oilfield area and may be related to the reorientation of compressive stress and sill intrusion reported by Geoffroy et al. (1996). They reported evidence of pre-Balder inversion in the Faroe-Shetland Basin, as indicated by strong onlap, and suggested this may be related to the reorientation of stress and compression. Dean et al. (1999) also interpreted a transpressional ‘pop-up’ structure along the Victory Lineament, which they confused with the Clair transfer zone. However, Moy and Imber (2009) have convincingly reinterpreted this feature to be related to an igneous intrusion / hydrothermal vent.

4 Post-breakup compression: geoseismic profiles

Our analysis of post-breakup compression, deformation and uplift within the Faroe-Shetland Region is largely developed from an interpretation of seismic data from both the Faroese and UK sectors, which is demonstrated through the illustration of a number of geoseismic profiles (Figures 5-17) together with structure contour maps of horizons (Figures 18-28) and associated isochore maps (Figures 29-35). The geoseismic profiles are described first; followed by descriptions of the structure contour and isochore maps. The key seismic horizons are briefly described in Table 1. A number of anticlines have been mapped within the 'carapace' overlying the block-faulted syn- and pre-rift rocks. Some of these anticlines have been informally named in this report and are labelled in blue font on the illustrated profiles. The larger anticlines tend to follow the mapped traces of the recognised structural highs (Keser Neish, 2003) (red font on the illustrated profiles).

The interpretation of syndepositional compression from seismic data has been a topic of recent study. Davies and Cartwright (2002) noted that Miocene sediments onlap the folds within the NE Faroe-Shetland Basin, indicating that either bathymetry was being passively infilled or that this infill took place during the folding episode. They noted that in some cases, the onlap converges towards the fold crest suggesting growth during the Miocene, and in others it is a passive onlap-fill configuration, suggesting early arrest of fold growth. Similarly, Pereira et al. (2011) described post-rift compression on the SW Iberian margin and developed a number of criteria to estimate the onset of compression on the basis of seismic profiles (Figure 36). We have utilised the criteria of Pereira et al. (2011) to evaluate evidence of post-breakup syndepositional compression within the Faroe-Shetland region, with the caveat that estimations of the timing of compression are hindered by the seismic expression of some contourites and/or distal turbidites. Interestingly, Pereira et al. (2011) also noted that compressive phases on the SW Iberian margin rarely comprise a discrete event, but occur during relatively prolonged periods of convergence.

4.1 TIES WITH WELL 214/04-1 (FIGURES 5, 6 AND 7)

Figure 5 comprises a dip section that illustrates the correlation of well 214/04-1 to commercial seismic reflection data and broadly equates to figure 111 of Stoker and Varming (2011). The profile demonstrates the present-day eastward tilt of Eocene and older rocks away from the Fugloy Ridge and the westward tilt of the Palaeogene succession away from the Shetland Isles toward the central axis of the deep-water channel. Whereas Davies et al. (2004) considered this configuration to be characteristic of the entire Palaeogene succession, the onlap of Oligocene and younger deposits onto the SE flank of the broad Fugloy Ridge anticline, suggests a distinct and important phase of Late Eocene tectonic activity. Over the crest of the Fugloy Ridge, the Eocene section, which is truncated beneath a composite intra-Neogene Unconformity (INU), may even include units with prograding character, possibly indicative of proximity to a contemporary uplifted basin margin.

Figure 5 demonstrates strong onlap onto an intra-Miocene Unconformity on the flanks of Anticline D (Ritchie et al., 2003), which was drilled by the 'Tobermory discovery' well 214/04-1, and also onto the SE flank of the Fugloy Ridge, which is indicative of an important phase of Miocene compression. In the basin centre, the Top Palaeogene (TPU) and Late Eocene/Early Oligocene (T2a) unconformities are not marked by angular breaks, but on the SE flank of the Fugloy Ridge there appears to be an indication of truncation below and onlap above both of these seismic horizons, suggesting successive growth phases.

Figure 5 also demonstrates a prominent seismic horizon, which can be observed lying approximately mid-way between the Top Palaeogene (TPU) and Late Eocene/Early Oligocene (T2a) unconformities. When traced back to well 214/04-1, this horizon can be assigned an intra-

Oligocene age. In some published accounts, this intra-Oligocene horizon has been confused with the TPU (e.g. Ritchie et al., 2003, figure 6; Stratagem Partners 2003, figure 9c), but it is demonstrably older. On the Fugloy Ridge Anticline, the intra-Oligocene horizon appears to correspond to an angular unconformity characterised by deep ravinement (see below).

The INU on Figure 5 equates to the INU recognised by Stoker and Varming (2011) and the Early Pliocene Unconformity of Davies & Cartwright (2002). However, Ritchie et al. (2003) also picked out another, younger Pliocene unconformity, that they erroneously termed the INU (see their figure 6), which appears to be characterised, at least locally, by a convex upward surface and stratigraphic truncation, possibility indicative of a post-INU phase of gentle flexure. At the southeast end of Figure 5, the Palaeogene and Neogene successions are truncated by the angular INU, which reflects the Neogene tilting described by Stoker et al. (2005a, b).

Figures 6 and 7, which are also located in the area around well 214/04-1, provide an opportunity to consolidate the seismic interpretation in an area where stratigraphic control and seismic resolution are reasonably good. A notable feature of the Faroe-Shetland region is a deterioration of seismic resolution below the diagenetic horizon. However, on the East Faroe High, a reflective seismic package rises above the diagenetic horizon and can be mapped and tied back to well 214/04-1 via Figures 6 and 7. Thus, these profiles, and their relative intersections, help to demonstrate some of the three-dimensional characteristics of the regional unconformities and the megasequences they bound.

Figure 6, is a dip section located on the SE flank of the East Faroe High and reveals the INU to be a spectacular angular unconformity, which is characterised by truncation below and thinning and onlap above. Interestingly, the INU unconformity over the East Faroe High Anticline has an asymmetrical, monoclinal disposition, with dipping SE limb, and this form is also expressed at the seabed as a bathymetric step. Figure 6 also images dramatic truncation below, and onlap above, the IMU, which defines a tighter monocline / anticline than that expressed at the level of the INU. The anticlinal forms of seismic horizons beneath the IMU, such as T2a and Top Palaeogene lavas, are broadly similar to that expressed by the IMU, indicating that the main growth phases of this fold were associated with the development of the IMU, the INU and possibly post-INU. Conceivably, growth of the monocline may have continued into Recent times, as reflected by the bathymetric step and the post-INU onlap which converges towards the fold crest.

Figure 7 is an approximately strike section, located in the East Faroe High / Steinvør Sub-basin area, and reveals the characteristic fold development and heavy onlap associated with the IMU. Thus the profile also constrains the 3D interpretation such that the IMU is cut out by the INU before the tie with Figure 6. Figure 7 also indicates that the intra-Oligocene horizon in the East Faroe High area represents an angular unconformity characterised by deep ravinement. Towards the north-eastern end of Figure 6, the bathymetry reflects the anticlinal form of the exhumed Eocene strata.

4.2 RATIONALISATION OF THE IMU AND MMU (FIGURES 8 AND 9)

Figure 8 is a strike section that is located in the Guðrun Sub-basin / Corona Sub-basin area and demonstrates a rationalisation of previous seismic interpretations regarding the relative importance of the Intra-Miocene Unconformity (IMU) and a Mid Miocene Unconformity (MMU) (Ritchie et al., 2003). In this study, a consistent view has emerged that identifies the IMU as the more significant unconformity, whereas the MMU is relatively minor and mainly manifested in the northern part of the Faroe-Shetland region. Typically, the IMU is folded and associated with strong onlap, as demonstrated by Figure 8 (and by Figure 9). This interpretation is consistent with that of Davies and Cartwright (2002 e.g. their figures 14 b and c) and with the STRATAGEM Partners (2003 e.g. their figure 9e), but at variance with some aspects of the interpretation of Ritchie et al. (2003) (e.g. their figure 6). It should also be noted that a considerable amount of differential compaction appears to have taken place during Miocene times either side of the mound of Mid-Eocene basin floor fan deposits. Figure 9, which is located in the Erlend Sub-basin, demonstrates that the MMU (Ritchie et al., 2003) represents a relatively minor angular break, which is characterised by some tilting and onlap. It remains open to debate whether the MMU reflects a minor phase of compression or is the result of other mechanisms, such as a change in bottom current activity.

4.3 EVIDENCE OF EPISODIC UPLIFT, NORTHERN FUGLOY RIDGE (FIGURE 10)

Figure 10 is a dip section, which is located at the northern end of the asymmetric Fugloy Ridge Anticline and provides an interesting comparison with Figure 5, to the SW. The lowermost part of the Stronsay Group (i.e. below T2d) is banked against the Faroe-Shetland Escarpment (Figure 10). Within the upper part of the Stronsay package, however, there appears to be evidence of onlap onto intra-Eocene seismic horizon T2d and for truncation below and onlap above T2c. Possible onlap onto T2b is poorly resolved. Onlap onto T2a, the Intra-Oligocene Unconformity and the TPU is also imaged. This pattern of repeated onlap and truncation shows some similarity to Figure 5, confirming that Oligocene and older strata are banked against the Fugloy Ridge Anticline. However, there are also some contrasting indications of earlier, mid-Eocene, phases of fold growth on the northern segment of the Fugloy Ridge Anticline.

4.4 EVIDENCE OF RECENT FOLDING (FIGURES 11 AND 12)

Figure 11 is located in the Guðrun Sub-basin / Corona High area and images a gentle anticline / ‘structural nose’, here informally termed the ‘Guðrun Anticline’, which is expressed at the level of the INU and at the seabed. Neogene deposits above the INU, become thinner towards the axis of the anticline, and are also tilted and onlap the parallel-bedded Palaeogene strata within the core of the anticline. Comparison of this reflection pattern with the criteria for the identification of syntectonic compression of Pereira et al. (2011) suggests the possibility of Neogene syndepositional compression both during and after the development of the INU, and which may even be persisting to the present day. Figure 11 also illustrates the relative stratigraphic positions of some of the Eocene tectonostratigraphic units described by Stoker et al. (2012), including the Mid-Eocene Portree, Caledonia and Strachan basin-floor fans (Stronsay Group phase 2 of Stoker et al., 2012) and the overlying slope apron deposits (Stronsay Group phase 3 of Stoker et al., 2012).

Figure 12 is a dip section, which is located in the East Faroe High / Guðrun Sub-basin area, which provides further evidence to suggest that Neogene deformation may have persisted into relatively recent times. Towards the south-eastern part of Figure 12, equivalents of the Mid-Eocene basin-floor fan deposits (Stronsay Group phase 2 of Stoker et al., 2012) are overlain by the slope apron package (Stronsay Group phase 3 of Stoker et al., 2012). These Palaeogene strata thin towards, and successively onlap, the limb of the large asymmetrical East Faroe High Anticline / Monocline imaged on the NW part of the profile. Within the Palaeogene succession, seismic reflection continuity is disrupted by polygonal faulting, but there appears to be evidence of onlap above the folded T2b unconformity. However, the thick Palaeogene deposits within the syncline flanking the East Faroe High Anticline have been subsequently deformed into a broad anticline (the ‘Guðrun Anticline’), and in a style which is detached from ‘basement’, which may be similar to the Palaeogene ‘Onika Anticline’ in the NE Rockall Basin (Tuitt et al., 2010). Growth of this anticline has restricted the distribution and thickness of late Neogene strata and may also have resulted in a subdued positive bathymetric relief over the structure. A seismic pattern of thinning and a further onlap surface above the INU (Figure 12) may provide evidence for continued and episodic rise of the anticline during late Neogene times.

4.5 SEISMIC EXPRESSION OF STRONSAY GROUP TECTONOSTRATIGRAPHIC UNITS (FIGURES 13, 14, 15 AND 16)

A number of profiles are illustrated here in order to demonstrate the seismic expression of the provisional Stronsay Group tectonostratigraphic units described by Stoker et al. (2012).

Figure 13 is a dip line, which is located in the Heri High / East Faroe High / Mid Faroe High area. The south-eastern part of this profile images the slope apron fan deposits (Stronsay Group phase 3 of Stoker et al. 2012), which downlap equivalents of the basin-floor fan succession (Stronsay Group phase 2 of Stoker et al., 2012) beneath the T2c unconformity. Above T2b, there is a suggestion of onlap onto the slope apron fan, although the seismic data quality in this area is poor. The position of the T2a unconformity and Intra-Oligocene unconformity on this profile are poorly constrained. As in the nearby Figure 12, the Palaeogene strata appear to be gently uplifted and deformed into a broad anticline against which the Neogene strata are ponded, and possibly subsequently uplifted to the NW. Once again, the stratal patterns may suggest Palaeogene and Neogene syntectonic development, which persisted after the early Pliocene, and may even be continuing to the present day.

Figure 14 is a dip line, which is located in the East Faroe High / Grimhild/ Guðrun Sub-basin area, and demonstrates aspects of intra-Eocene folding. Onlap onto the folded T2c surface in the Grimhild / Guðrun Sub-basin area suggests a phase of compression associated with the T2c unconformity. Similarly, onlap at T2b and onto the inverted slope apron succession at the southern end of the Corona High / Judd Anticline may indicate a further phase of compression.

Figure 15 is a strike line, which is located in the Corona High / Flett Sub-basin area. Mounded units representing the three channelized developments of Mid-Eocene basin floor fans (Stronsay Group phase 2 of Stoker et al., 2012); from south to north the Portree, Caledonia and Strachan fans are imaged in the central and NE parts of the profile. The stratigraphically higher, slope apron deposits (Stronsay Group phase 3 of Stoker et al. 2012) are clearly imaged on the south-western part of the profile. Neogene strata are notably thin in this part of the Faroe-Shetland Basin.

Figure 16 is located in the southwest of the study area and images the downlapping prograded units of the Lower Eocene (Ypresian) Munkagrunnur Ridge Delta (Stronsay Group phase 1 of Stoker et al., 2012), which has been described by Ólavsdóttir et al. (2010). Onlap onto the gently folded T2d unconformity surface indicates some mid-Eocene compression on the Judd Anticline, as described by Smallwood (2004) and Ritchie et al. (2008).

4.6 'SOUTHEAST FAROES DRIFT' (FIGURE 17)

Figure 17 demonstrates the feature known as the 'Southeast Faroes drift'. This feature exhibits an elongate planform geometry, parallel to the basin axis, and this aspect together with the possible occurrence of moats, and the interpretation of an internal 'upslope migrating downlap configuration' led Davies et al. (2001) to propose that it comprises a contourite drift. The body is dated as Early Oligocene on the basis of seismic correlation with well 214/04-1. However, it should be noted that the identification of a pronounced upslope migrating downlap system (Davies et al., 2001) remains uncertain, due to confusion with the cross-cutting diagenetic horizon (see Davies et al., 2001, figure 2b). Furthermore, the internal seismic character of the body appears to be cut by many small faults and thus the recognition of moats remains equivocal. In view of the regional evidence regarding the timing of initiation of North Atlantic Deep Water formation, it may be appropriate to consider some alternative explanations for the nature of the 'Southeast Faroes drift' (see Discussion section).

5 Post-breakup deformation; structure contour maps

Structure contour maps, in two-way-travel-time, have been produced for a number of interpreted seismic horizons. The maps include (in red) the positions of structural elements derived from figure 7 of Ritchie et al. (2011), and the location of the Faroe-Shetland Escarpment derived from figure 120 of Ritchie et al. (2011); also displayed are the locations geoseismic profiles illustrated in this report. The geographical extents of some of the mapped seismic horizons are restricted; this can be due to a number of factors, such as truncation of the surface due to the development of composite unconformities, or the limit of mapping and/or the limit of data, as noted below.

5.1 TOP PALAEOGENE LAVAS

The lavas are widely distributed and cover most of the Faroese sector and the northeast part of the UK area (Figure 18). The Faroe-Shetland Escarpment (FSE) forms a primary lava escarpment marking a former coastline; to the south and east of the escarpment the lavas are thinner and reach a feather edge.

A number of anticlines are depicted on the Top Palaeogene lavas structure contour map, many of which have been recognised and informally named by previous authors (e.g. Ritchie et al. 2003; Davies et al., 2004). It is clear that there is a wide range of fold dimensions and fold trends, including NNW-trending (e.g. Munkagrunnur Ridge), ENE-trending (e.g. Fugloy Ridge), NE-trending (e.g. East Faroe High Anticline) and NNE-trending (e.g. Anticline E and Pilot Whale Anticline). The larger anticlines tend to follow the mapped traces of recognised structural highs (Keser Neish, 2003). The characteristics of selected anticlines (e.g. Fugloy Ridge Anticline, East Faroe High/Heri High Anticline, Pilot Whale Anticline and Anticline D), some of which feature on the illustrated geoseismic profiles, are described below.

The Fugloy Ridge Anticline is a large, ENE-trending, anticline, with bathymetric expression that is located to the east of the Faroe Platform, and has been described by a number of authors (e.g. Boldreel and Andersen, 1995 and 1998; Ritchie et al., 2003). The anticline is asymmetrical, with a steeper WSW limb (Figures 5 and 10). Towards the western part of the Fugloy Ridge Anticline, the Palaeogene lavas rise to outcrop at seabed and are uplifted above sea level on the Faroe Islands. The anticline plunges towards and terminates to the ENE, possibly at the intersection with a cross-cutting lineament. Two large, separate, structural culminations are mapped along the Fugloy Ridge Anticline and this morphology may also reflect the influence of cross-cutting lineaments.

The East Faroe High/Heri High Anticline is a NE-trending, anticline (Figure 6) with bathymetric relief, which is located to the southeast of the Faroe Platform (Boldreel and Andersen, 1995; Ritchie et al., 2003). The anticline is asymmetrical with a steeper SW limb and plunges to the NE, where it converges towards the Fugloy Ridge. Separate structural culminations are mapped along the axis of anticline ridge and this morphology may reflect segmentation by cross-cutting lineaments.

The Pilot Whale Anticline is a large, NNE-trending symmetrical anticline located near the margins of the Erlend Sub-basin / Møre Marginal High (Figure 9), which is associated with seabed / subsurface diapirs (e.g. Holmes et al., 2003). The Pilot Whale Anticline terminates sharply to the south, where it appears to be dextrally offset from Anticline E. Anticline E is a NNE-trending slightly asymmetrical fold, located within the Erlend Sub-basin. The line of offset between these anticlines coincides approximately with the trace of the Faroe-Shetland Escarpment, which may indicate a common structural inheritance from an underlying rift structure.

Anticline D (e.g. Figure 38) is a NE-trending anticline located within the Erlend Sub-basin. Davies et al. (2004) recognised a number of separate structural culminations along Anticline D, to which they assign separate names, which may reflect segmentation associated with cross-cutting lineaments (e.g. Erlend Lineament).

The Munkagrunnur Ridge Anticline (Figure 18) appears to be a large NNW-trending fold that merges northwards with the Faroe Platform. Over much of the anticline the Palaeogene lavas crop out and are eroded at the sea bed. The availability of FSC seismic data is relatively restricted over the Judd and Westray anticlines. However, these folds are well described by Smallwood (2004) and Ritchie et al. (2008), but it should be noted that the age of Eocene unconformities has been amended in the light of new biostratigraphic data (Stoker et al., 2012).

5.2 TOP BALDER FORMATION

The distribution of the Balder Formation (Figure 19) is limited in the northern part of the study area, due to onlap onto the Faroe-Shetland Escarpment (Figures 9 and 10), but elsewhere the mapped extent generally reflects a limit of interpretation. The Balder Formation seismic unit does appear to overlap the south-western part of the Faroe-Shetland Escarpment, where the escarpment presumably formed a smaller topographic feature. The Top Balder map displays the distribution of most of the significant anticlines / domes within the Faroe-Shetland region, including parts of the east-west-trending Judd Anticline and the NW-trending Westray Anticline in the south and east, beyond the limit of lavas.

5.3 T2D

The mapped extent of seismic horizon T2d (Figure 20) largely reflects a limit of interpretation, but in the NE of the study area, equivalents of Stronsay Group phase 1 deposits (Stoker et al., 2012) appear to onlap contemporaneous structural highs, including the Faroe-Shetland Escarpment (Figures 9 and 10). Towards the SE corner of Faroese Quadrant 6104, a NW-trending structural nose, informally termed the ‘Guðrun Anticline’ has been identified. Figure 12 suggests this feature may relate to a detached style of inversion of the Eocene and Oligocene sedimentary basin fill.

5.4 T2C AND T2B

The mapped extent of seismic horizons T2c (Figure 21) and T2b (Figure 22) largely reflects a limit of interpretation, but an area in the south where these horizons are absent may reflect the growth of the east-west-trending Judd Anticline.

5.5 T2A

The mapped extent of seismic horizon T2a (Figure 23) largely reflects truncation of the Eocene by later unconformities, such as the IMU and the INU. However, in the south of the study area, a NE-trending area where horizon T2a is absent picks out an apparent NE continuation of the Judd Anticline as postulated by Stoker et al. (2012) and may reflect growth of that structure.

5.6 INTRA-OLIGOCENE UNCONFORMITY AND TPU

The Intra-Oligocene Unconformity (Figure 24) and Top Palaeogene Unconformity (Figure 25) seismic horizons have relatively restricted distributions, largely reflecting uplift and erosion of the Eocene and Oligocene successions. The Intra-Oligocene Unconformity and TPU commonly merge into composite unconformities with the IMU and INU, especially in the southern part of the Faroe-Shetland Basin and on the flanks of the surrounding structural highs (e.g. Fugloy Ridge Anticline).

5.7 IMU

The IMU (Figure 26) forms a folded surface whose mapped extent is limited by truncation at the INU. Erosion at separate structural culminations of the East Faroe High Anticline may partly reflect segmentation of the anticline by cross-cutting lineaments.

5.8 INU

The INU (Figure 27) is a widely distributed horizon whose mapped extent near the Faroe Platform, Munkagrunnur Ridge and Wyville Thomson Ridge largely reflects truncation at the seabed, but to the north and east of these areas is a limit of interpretation. A NE-trending dome in the SE part of Quadrant 6104, which is here informally termed the 'Guðrun Anticline' corresponds to an uplifted zone of thick Palaeogene deposits, as imaged on Figure 12. A possible dextral offset of the Fugloy Ridge, the Erlend Basin and Anticline E / Pilot Whale Anticline approximates to the mapped trace of the Faroe-Shetland Escarpment, possibly indicating a common structural inheritance / reactivation of an underlying rift structure.

5.9 SEABED

The morphology of the seabed (Figure 28) generally reflects the large scale fold structures, which appear to control the shape of the Faroe-Shetland Channel, such as the Fugloy Ridge, Munkagrunnur Ridge, and the SE flank of the basin. A 'structural nose' feature is apparent in south-eastern part of Q1604 and corresponds to uplifted Palaeogene sediments within the informally termed 'Guðrun Anticline' (Figure 12).

6 Post-breakup deformation: isochore maps

6.1 TOP BALDER FORMATION – T2D

This map (Figure 29) reveals thick Lower to Middle Eocene deposits of the Munkagrunnur Ridge Delta (Ólavsdóttir et al., 2011) in the south of the report area. As noted by Ólavsdóttir and Ziska (2009) and by Stoker et al. (2012), thick prograding deposits of Stronsay Group phase 1 also extend to form a NE-trending shelf in Quadrant 205. This phase of shelf progradation has been interpreted to reflect contemporaneous uplift and erosion of source areas southeast of the study area, including part of the Munkagrunnur Ridge. Some thickening of this stratigraphic interval is also apparent adjacent to the Erlend and Brendan centres; however, no significant thickening of the unit has been identified along the northern Fugloy Ridge, nor along the East Faroe High, which may indicate that no significant sediment source areas were developed in these areas at that time. Locally, the lower part of this seismic unit onlaps the Faroe-Shetland Escarpment (Figures 8 and 9).

6.2 T2D – T2C

This map (Figure 30) reveals the thick development of Middle Eocene basin-floor fan deposits that is largely located within the Corona High / Corona Sub-basin / Erlend Sub-basin area (Stronsay Group phase 2 of Stoker et al., 2012). There is a noticeable thinning of this stratigraphic interval against the northern segments of the Fugloy Ridge and the Faroe-Shetland Escarpment (Figures 9 and 10), which may reflect contemporaneous growth and/or differential thermal subsidence across these structures.

6.3 T2C-T2B

This map (Figure 31) reveals a NE-trending area of shelf and associated slope apron deposits (Stronsay Group phase 3 of Stoker et al., 2012), which prograded into the Flett Basin/Corona High area, possibly in response to a contemporaneous phase of uplift to the southeast. Notably, the Flett High appears to correspond to a local area of thin deposits.

6.4 T2B-T2A

The map of this rather poorly-defined seismo-stratigraphic unit (Figure 32) shows a number of areas with local thickening, including a NE-trending development apparently infilling a growth syncline in the Guðrun Sub-basin area (Figure 12).

6.5 T2A-IMU

This map (Figure 33) reveals, at least in part, a thickening of the Oligocene succession ascribed by Davies et al. (2001) to the development of the ‘Southeast Faroes drift’ (Figure 17). The morphology of this body appears to display a NE-trend, which is broadly aligned with that of the East Faroe High and the Fugloy Ridge (see Discussion section).

6.6 IMU-INU

This map (Figure 34) clearly demonstrates the control exerted on early Neogene sediment distribution by the developing Miocene folds (e.g. Figure 8). A number of depocentres are shown in the ‘warm’ colours and correspond to the infill of synclines, whereas the thin deposits over the anticlines are picked out in ‘cool’ colours. An en echelon aspect to the depocentres/synclines and anticlines can be observed, for example, at the northern end of the East Faroe High, which may reflect the long-lasting influence of cross-cutting offsets in the underlying rift architecture (see Discussion section).

6.7 INU-SEABED

This map (Figure 35) demonstrates aspects of the sedimentary response to change in Early Pliocene times, with a shift in the focus of sedimentation from the basinal areas (IMU-INU, Faroe-Shetland Neogene 2 [FSN-2]) to the shelf and slope (Faroe-Shetland Neogene 1 [FSN-1]) as described by Stoker and Varming (2011). Thick deposits of the glacially-fed North Sea Fan are prominent across the Møre Basin and Møre Marginal High. However, the area around the Brendan Volcanic Centre corresponds to relatively thin deposits, possibly indicating an area of relative buoyancy. The main instigation of the late Neogene change has been related to the onset of epeirogenic movements (tilting) from about 4 Ma, which modified the patterns of contourite sedimentation and facilitated the onset of rapid shelf progradation, including the Foula Wedge (Figure 35) (e.g. Stoker and Varming, 2011). To the SE of the East Faroe Wedge, late Neogene sediments appear to infill a growth syncline developed on the NW flank of the 'Guðrun Anticline' (Figure 12).

7 Gravitational potential energy

7.1 BACKGROUND

The vertical distribution of mass within the lithosphere affects the gravitational potential energy (GPE) and lateral differences in GPE result in intraplate stresses which can be large enough to cause lithospheric deformation (e.g. Frank, 1972; Artyushkov, 1973; Coblentz et al., 1994). For example, in the context of mountain belts, gravitational collapse can be seen as a process driven by the GPE imbalance developed during the preceding orogenic deformation: areas of high topography (and thus high GPE) undergo extensional collapse and the surrounding lower GPE regions are subject to compressional deformation (Rey et al., 2001). Ridge-push arises from the GPE difference between the mid-ocean ridges and the adjacent regions, and is commonly considered one of the dominant forces acting on passive margins. A shortcoming of a GPE (ridge-push) explanation for Cenozoic deformation on the NE Atlantic margin is the apparent episodicity of that deformation. Doré et al. (2008) suggested that, although ridge-push forces associated with a normal mid-ocean ridge are relatively uniform with time, the situation in the NE Atlantic differs because of the presence of the Iceland Insular Margin. They argued that, if this topographic feature developed during the Miocene, it could explain the enhanced development of compressional structures at that time and also the arcuate distribution of such structures around Iceland.

There are two ways of estimating lateral variations in gravitational potential energy. If its vertical lithospheric density structure is known, the GPE of a rock column is given by the vertical integral of the product of density, elevation and gravitational acceleration (i.e. the integral of lithostatic pressure). Alternatively, it can be shown that, in regions that are in isostatic equilibrium, the geoid height anomaly provides an independent measure of GPE (Coblentz et al, 1994). The relationship is linear: at present latitudes, multiplying the geoid height anomaly (in m) by 2.3×10^{11} will provide an approximate measure of GPE (in N/m).

Doré et al. (2008, their fig. 9) illustrated the geoid height anomaly variation, and the GPE variation based on this, across the north Atlantic region. They noted the contrast between the high GPE values over the Iceland Insular Margin and the low values over the adjacent Norwegian continental margin and argued that this is conducive to the development of compressional stresses in the latter area. In an important qualification, they recognised the limitations of their use of unfiltered geoid height anomalies in such calculations. The unfiltered geoid includes long-wavelength components due to deep sources that are unlikely to transmit stresses to the lithosphere directly. Doré et al. (2008) considered that this may have led to an overestimate of stress magnitude but nonetheless concluded that body forces generated in this way were a primary mechanism for the development of Neogene compressional structures on the NE Atlantic margin.

Pascal and Cloetingh (2009) calculated GPE for the Southern Scandes region by the two independent methods (from inferred lithospheric density structure and from geoid anomalies). The longest wavelengths were removed from the geoid by truncation at degree and order 12 of the spherical harmonic series from which it was constructed. They achieved a reasonable correspondence between modelled GPE and the filtered geoid, with local mismatches attributed to limitations in the simulation of lithospheric density structure in the modelled version (Pascal and Cloetingh, 2009, their fig. 6). There was a clear contrast between high GPE in the mountainous Southern Scandes and lower GPE in adjacent offshore areas and they argued that this made a significant contribution to the regional stress pattern, helping to explain the distribution of seismicity and the rotation of observed maximum horizontal stress directions between the Norwegian Margin and Northern North Sea. The influence of the Southern Scandes is much less evident in the unfiltered geoid of Doré et al. (2008).

Rather than a sharp cut-off of the harmonic series at particular degree and order, other authors have used a more 'gentle' truncation to isolate the relevant component of the geoid with less risk of distortion. Both Flesch et al. (2000) and Coblentz et al. (2007) employed a one-sided cosine taper between degree and order 7 and 11. According to Coblentz et al. (2007), such a filter removes all anomalies with wavelengths greater than 3520 km (at the equator), corresponding to depths greater than about 620 km.

7.2 NEW GPE RESULTS FOR THE FAROE-SHETLAND BASIN REGION

Figure 37 illustrates new GPE calculations for the north Atlantic region. In the version based on geoid height anomalies (Figure 37a) the longest wavelength components have been removed by a filter similar to that advocated by Flesch et al. (2000) and Coblentz et al. (2007). The filtered geoid anomalies were extracted using a Java Applet at: http://icgem.gfz-potsdam.de/ICGEM/potato/Service.html. The Eigen_gl04c geoid (Förste et al., 2006) was employed, with the 'gentle-cut' filter provided within the applet applied to the spherical harmonic series between degree and order 7 and 11.

Three-dimensional gravity modelling of the sort described by Kimbell et al. (2004, 2005) provides a lithospheric density model from which GPE variations can be calculated directly. Figure 37b illustrates this in the case of the original model of Kimbell et al. (2004, 2005), and Figure 37c shows the pattern derived from the 3D model of the Faroe-Shetland area constructed in a recent Faroe-Shetland Consortium project (Kimbell et al., 2010). In both cases, the GPE was calculated for a column extending from surface down to 125 km below datum. The GPE of the reference lithosphere employed in the original gravity modelling was then subtracted in order to express the values as deviations from the norm. A Butterworth filter with a central wavelength of 300 km was applied to suppress shorter wavelength features which may be held out of isostatic equilibrium by the flexural strength of the lithosphere and are not appropriate for analysis of this sort (cf. Pascal and Cloetingh, 2009).

The two independent methods of calculation provide very comparable results, with features that are similar in both shape and amplitude identifiable in the GPE patterns derived from the geoid and from the 3D gravity models (Figure 37). The Faroe-Shetland Consortium 3D model (Figure 37c) clearly covers too small an area to contribute usefully to the analysis of regional stress patterns, but it does corroborate the results obtained from other sources in that area. The results over the Southern Scandes are similar to those of Pascal and Cloetingh (2009) although the match between the geoid and model calculations is better in the new versions, perhaps because of more accurate density structure and fewer edge-effects in the model and a more effective filter in the geoid.

The seismic Moho beneath northern Scotland and the Hebrides Shelf is about 2 km shallower than predicted by regional gravity modelling, and this may be due to the influence of a relatively low density underlying mantle (Kimbell et al., 2004). It follows that GPE in this area may be underestimated, because the lithospheric column is more 'top heavy' than is assumed in the model. The sensitivity trials of Pascal (2006) suggest that the underestimate could exceed 0.5 x 1012 N/m, although there is no evidence of a discrepancy this large between the present model and geoid predictions. The depth at which the mantle density anomaly occurs will, however, affect its impact on GPE: for example if the anomaly lies just beneath the Moho its influence on GPE will be relatively small.

The maps reveal that the northern part of the Faroe Shetland Basin forms an area of low GPE lying directly between the high GPE areas of the Iceland Insular Margin and the Southern Scandes. Providing there is no impediment to the lateral transmission of stress, this configuration predicts a present-day broadly WNW-ESE compressive regime within this part of the basin. It would be instructive to examine the evidence from breakouts in wells in the area to see if this is corroborated. A similar compressive regime may have existed in Middle Miocene times, since the Southern Scandes were an upland area at that time (shedding the Utsira Formation into the Viking Graben; Jordt et al., 2000) and, according to Doré et al. (2008), the Iceland Insular Margin was also exerting a key influence.

Further north, the main compressive structures on the Norwegian margin can also be correlated with a zone of relatively low GPE between the Iceland Insular Margin and the Norwegian mainland. The gap in the belt of such structures that coincides with the Møre area may be in part related to the hiatus in the GPE pattern that is observed there, although Doré et al. (2008) argue that late ultra-slow spreading on Aegir Ridge may have 'protected' this part of the margin by absorbing compression directed from the Iceland Insular Margin. The compressive structures further south-west, on the Hatton Margin, do not show a close correspondence with the GPE pattern, although the Reykjanes Ridge (to the west of the area shown) has high GPE and will be responsible for a compressive influence at this margin. Early deformation may have been influenced by divergent asthenospheric flow at or close to the time of breakup (Doré et al., 2008).

The Munkagrunnur High and Wyville-Thomson Ridge are oriented in a WNW-ESE direction, at a high angle to the trend that prevails in the compressional structures further north in the Faroe-Shetland Basin. The present-day GPE pattern does not provide an obvious explanation for this: the GPE trends are deflected in the area, but these structures lie in a 'saddle' rather than an area of low GPE. At the time they were initiated the relative influence of the south-eastern end of the Iceland-Faroe Ridge may have been greater and thus contributed to a rotation of stress orientations in this area. It is also possible that reactivation of pre-existing structures was a significant factor.

Although the difference in GPE between two areas provides an approximate measure of the horizontal deviatoric stress generated, numerical modelling is required in order to achieve a realistic assessment of the orientation and magnitudes of such stresses. The rheological structure of the lithosphere will affect factors such as the dissipation and deflection of stresses and whether deformation is likely to occur. Weak zones can have a major influence. For example, the presence of a rift along the axis of the mid-ocean ridge fundamentally changes the stress distribution, dissipating extensional stress near the ridge crest and increasing compressional stress within the adjacent plate interiors (Bott, 1993).

Although important, gravitational potential energy is not the only factor generating stress within the lithosphere and is likely to be supplemented by processes such as shear drag (i.e. lithosphere-asthenosphere interaction). In fact, the combination of GPE with other factors could help to explain the episodic nature of the deformation. GPE may have maintained a compressive regime such that only a small additional 'push' from another source was required to initiate a phase of deformation. Such sources might include: (i) stress variations associated with plate reorganisation; (ii) modulations to ridge-push resulting from transient changes in ridge elevation associated with temperature (buoyancy) variations in the underlying asthenosphere; (iii) temporal variation in shear drag; and (iv) far-field influences from the Alpine orogen.

8 Discussion

8.1 MARGIN SEGMENTATION AND REACTIVATION

Large-scale segmentation of the continental margin of NW Europe along NW-trending offsets has been described by a number of authors (e.g. Kimbell et al., 2005). However, a finer level of segmentation has also been proposed within the Faroe-Shetland region (e.g. Rumph et al., 1993; Keser Nesh, 2003; Ellis et al., 2009; Ritchie et al., 2011). Indeed, post-breakup transpressional reactivation or 'shuffling' of continental basement blocks separated by such cross-cutting lineaments has been postulated as a local mechanism that controlled the location and growth of some compressional domes in mid-Norway (Doré and Lundin, 1996). Similarly, Ritchie et al. (2003) noted the close spatial relationship between the cross-cutting NW-trending lineaments and lateral offsets observed between NNE-trending anticlines E and F (Pilot Whale Anticline) and postulated that may have been formed in response to sinistral strike-slip motion along the postulated Magnus Lineament. However, they also noted that other factors, such as buttressing by pre-existing structures may have been significant. Seismic mapping for this study confirms the offset nature of anticlines E and F (Pilot Whale Anticline) (Figure 38) and also suggests that cross-cutting structures may separate structural culminations along individual structural highs. Generally, the cross-cutting lineaments facilitating the offsets are presumed to be NW-trending. However, within the Faroe-Shetland region there is also evidence of important WNW-trending and east-west-trending faults. For example, on the basis of field mapping and correlations on the Faroe Islands, an offshore WNW-trending lineament with about 5 km of post-lava, dextral strike-slip movement, has been postulated separating the islands of Sandoy and Hestur (Passey, 2009). Furthermore, Mid to late Paleocene E-W and NE-SW tensional faults are thought to be related to right-lateral oblique-slip movements on the NW-SE oriented transfer faults (Ellis et al. 2009). Interestingly, Cooper et al. (2012) report Palaeogene NNW-SSE-trending dextral and ENE-WSW-trending sinistral conjugate strike-slip faults in Northern Ireland, on the basis of extensive high-resolution aeromagnetic data. Within the area around the Pilot Whale Anticline, comparison with the results from 3D gravity modelling (Figure 38) may favour segmentation by WNW-trending lineaments. Reactivation or inversion of individual pre-existing structures is generally difficult to demonstrate in the Faroe-Shetland region, due to the widespread masking effects of the Palaeogene lavas and associated intrusive rocks. However, in general terms, the 3-D gravity modelling suggests that the central part of the Pilot Whale Anticline directly overlies a small inverted sub-basin (Figure 38). Similarly, further south, parts of the Corona/Erlend Sub-basin, the northern East Faroe High/ Steinvør Basin and the Guðrun Sub-basin correspond to inversion anticlines (Figure 38).

8.2 'SOUTHEAST FAROES DRIFT'

To the southeast of the Faroe Islands, part of a thick Oligocene section lying between the TPU and the Intra-Oligocene Unconformity has been termed the Southeast Faroes drift (SEFD) (Davies et al., 2001). The elongate external form, internal reflection geometry and stratigraphic context, together with a 'pronounced upslope migrating downlap system' has been considered to be sufficient to characterise this package as a contourite drift. However, this interpretation may be compromised by the nature of the proposed downlap system, which appears to rely upon the cross-cutting diagenetic horizon as the putative downlap surface (Davies et al., 2001, figure 2d). Davies et al. (2001) considered the SEFD to indicate that breaching of the Greenland-Scotland Ridge (GSR) oceanic gateway and initiation of deep-water circulation from the Norwegian Sea into the North Atlantic took place much earlier than some models suggest (e.g. Schnitker, 1980; Thiede and Myhre, 1996). However, there is a wealth of data from DSDP and ODP sites to suggest that the establishment of a significant deep-water connection across the gateway is predominantly a Neogene phenomenon (Stoker et al., 2005a, b). For example, ^{13}C and ^{18}O data show that bottom-current circulation in the Atlantic Ocean prior to the Mid Miocene was driven by Tethyan Outflow water, as far north as the GSR and including DSDP site 610 (Ramsay et al. 1998). Furthermore, comparison of Oligocene deep-water agglutinating foraminifera from ODP site 985 (north of the GSR) with ODP site 647 (south of the GSR) reveals major taxonomic differences indicative of faunal isolation of the deep Norwegian Basin (Kaminski and Austin 1996). Therefore, we favour Neogene gateway development (Faroe Conduit) and suggest that it may be appropriate to consider alternative explanations for the nature of the SEFD, such as whether it could represent a relatively undeformed slump (internal seismic facies not chaotic), or alternatively the fill of a contemporaneous growth syncline.

8.3 NEW SEISMO-STRATIGRAPHIC CHART

In order to compare the spatial and temporal distribution of the observed deformation in the Faroe-Shetland region with the timing of tectonic events within the wider region and to seek new insights into the relative importance of the possible regional controls on deformation, we have combined the results of the seismo-stratigraphic analysis developed in this study with results from selected regional studies, including Stoker et al., (2012), to construct a new stratigraphic chart summarising key aspects of mainly post-breakup compression and deformation within the Faroe-Shetland Basin and the wider NE Atlantic region (Figure 39). It should be noted that the dating of some unconformities has been recalibrated in the light of new biostratigraphic information (Stoker et al., 2012), but significant areas of uncertainty do remain with regard to the accurate dating of some of the unconformities (e.g. intra-Oligocene Unconformity, Mid-Eocene Unconformity (T2d)). The unconformities depicted on the chart are recognised from stratal configurations displayed on seismic profiles such as tilting, folding, truncation and onlap (e.g. Figure 5). The chart indicates the positions of passive margin unconformities, which are considered to be associated with tectonically driven uplift, differential/accelerated subsidence / sagging / tilting, and folding / doming and not solely due to the lowering of relative sea level caused by the interactions of thermal subsidence and eustacy. Seismic evidence from this study, and perhaps more regionally, suggests that the differential / accelerated subsidence / sagging / tilting episodes were coeval with phases of compression, which seems to have occurred at a variety of scales (Stoker et al., 2012) and were also coeval with important phases of uplift and exhumation (e.g. Holford et al., 2010; Stoker et al., 2010b; Japsen et al., 2011; Robinson et al., 2004) (Figure 39). Poly-phase compressional deformation may be a common feature of many passive margins and rifts and the resulting folds are characterised by a spectrum of spatial wavelengths spanning several tens of kilometres up to several hundred kilometres (Cloetingh et al., 2008). Furthermore, the interplay of plume and intraplate compressional deformation may be associated with temporal transitions from basin inversion to lithospheric folding (Cloetingh et al., 2008). Consequently, it remains a possibility that lithospheric folding may have contributed to end-Eocene sagging and early Pliocene tilting in the manner postulated by Japsen et al. (2011).

The chart highlights some areas of contrasting interpretation when compared to previous studies. For example, Tuitt et al. (2010), working mainly in the Hatton-Rockall region, but also extending into the Wyville-Thomson and Judd areas, postulated a ‘switch off’ of compression in Oligocene times, followed throughout the remainder of Cenozoic times by the development of unconformities associated solely with bottom currents. In contrast, this study suggests that post-breakup compression does not appear to be restricted to a single interval of time, rather the margin has been deforming near-continuously since soon after breakup in the earliest Eocene (Figure 39; Stoker et al., 2012), although the loci of deformation have varied both spatially and temporally. Indeed, some structures may have been active into relatively Recent times (e.g. Figure 12). However, there does appear to be some key periods of deformation / peaks in tectonic activity, which coincide with regional unconformities, such as especially T2a, IMU and INU.

This study has also highlighted some inconsistencies with regard to published correlations of post-breakup unconformities within the Faroe-Shetland region. For example, we recognise the IMU as the major Miocene unconformity within the Faroe-Shetland region, whereas the MMU is a relatively minor break, which is limited in extent to the NE of the region. We have also identified seismic evidence of a previously little documented intra-Oligocene Unconformity, which is characterised by deep ravinement on the East Faroe High. On the basis of ties with well 214/04-1, the age of this stratigraphic break is considered to be intra-Oligocene, although the precise age remains uncertain. In the Mid Norway offshore, Lundin and Doré (2002) postulated an Early Oligocene relative fall in sea level and an associated north-trending deltaic unit to reflect mild uplift of the shelf in response to mid-Cenozoic doming. However, this progradational unit has subsequently been shown to be of Late Miocene to Early Pliocene age and is thought to be deposited as a result of the compression and uplift of mainland Norway in mid Miocene time ((Eidvin et al., 2007).

In any assessment of the tectonostratigraphic significance of the mapped unconformities within the NE Atlantic margin, the relative control of tectonism and eustacy can be difficult to evaluate. However, an early Mid-Eocene break (the T2d reflection) associated with channelized incision of contemporaneous shelf deposits does not appear to coincide with a significant eustatic or glacio-eustatic fall in sea level (Stoker et al., 2012), but rather to compression and uplift over a wide area, including the formation of the Judd and Westray anticlines and intra-Lutetian uplift of the Flett High (Robinson et al., 2004). A late Mid-Eocene progradation of the West Shetland margin was associated with the development of the T2c reflection, and may be a response to uplift of the margin (Stoker et al. 2012). Intra-Eocene tectonism appears to have been particularly active in the southern parts of the Faroe–Shetland region (Figure 38), although there also appears to be evidence of intra-Eocene tectonism affecting the Stronsay Group around the northern part of the Fugloy Ridge (Figure 10).

On the basis of seismic interpretation, Andersen et al. (2000) postulated that the structural high associated with the Faroe Platform was formed at the end of the Eocene and did not provide any significant sedimentary input into the Faroe-Shetland Basin until the Oligocene. Indeed, the shaping of the Faroe-Shetland Channel does appear to have been instigated at the end of the Eocene, coeval with similar development along the margin as a whole, and was enhanced during late Palaeogene–Neogene times. However, inspection of Figure 5 reveals a number of SE-dipping and converging reflectors within the Eocene succession on the crest of the Fugloy Ridge Anticline, which might represent south-easterly progradation possibly derived from an area to the west of the current ridge crest. However, more detailed regional mapping would be required to assess this issue (Stoker et al., 2012).

Any consideration of controls on tectonostratigraphic development in the Faroe–Shetland region must also take account of the wider tectonic setting. The initiation of spreading to the north and west of the Faroe–Shetland region, combined with the effects of continental collision to the north, such as the Eurekan and West Spitsbergen orogenies, and to the south, such as the collision between Eurasia and Iberia and associated Alpine tectonism (Hibsch et al., 1995; Sissingh, 2001), and developing GPE body forces associated with growth of the Iceland Insular Margin from Miocene times (Doré et al., 2008) would have placed this developing passive margin and intra-plate region into compression. There may also have been a contribution to the development of post-breakup regional unconformities from plume-related transient convective uplift (Hartley et al., 2011). However, the proximity of the Faroe–Shetland region to protracted Palaeogene breakup and spreading events within the oceanic Norwegian Basin may be especially significant. The mid-ocean ridge propagating from the southern NE Atlantic appears to have failed to join the active ridge in the Norway Basin, resulting in the development of a wide zone of extension and/or transtension to the south and SE of the Jan Mayen microcontinent (JMMC) (Gaina et al., 2009) (Figure 3). This led to counter-clockwise rotation of the JMMC between chron 25 (c. 56 Ma) and chron 24 (c. 53.5 Ma). Kinematic reconstructions suggest that extension occurred in the SE part of the JMMC at about chron 21 (c. 48 Ma), followed by further extension of the southwestern margin of the JMMC at chron 18 (c. 40Ma) associated with a final, westward ridge jump of the southern spreading ridge. The two episodes of extension in the southern part of the JMMC resulted in counter-clockwise rotation and local compression on the east or SE margin of the JMMC (Figures 3 and 39). At about 30 Ma, the Aegir Ridge became extinct and the ridge propagating from the southern NE Atlantic managed to completely detach the southern part of the JMMC by chron 6 (c 20 Ma). The complex breakup history of the JMMC is reflected by the presence of major unconformities that may reflect the various ridge jumps before the final ridge jump, which led ultimately to the microcontinent rifting from East Greenland at some time after 30 Ma (Gaina et al., 2009). The broad coincidence between the timing of plate reorganisation events in the adjacent Norway Basin and the formation of the Eocene unconformities in the Faroe–Shetland region may suggest a causative linkage.

9 Conclusions

There is regional seismic evidence of localised, pre-breakup, folding and inversion, probably related to mainly Late Cretaceous transpression and transtension. Post-breakup compression in the Faroe–Shetland region has been analysed primarily through interpretation of oil industry seismic data. The seismic expression, extent and thickness of unconformity-bounded units are illustrated by geoseismic profiles, structure contour maps and isochore maps. Although further seismic mapping and characterisation is warranted, on the basis of this analysis our key conclusions are:

1. Post-breakup compression, folding and associated uplift may have been initiated in the early Eocene and may have persisted into late Neogene / Recent times, as evidenced by seabed domes with associated late Neogene onlap configurations.

2. There is a considerable variety of scales to the folds and the associated broad uplifts, and previously recognised phases of end Eocene 'sagging' (accelerated subsidence), early Pliocene tilting and uplift/exhumation were coeval with compression. Conceivably, some of the large scale sagging and tilting may be associated with lithospheric folding.

3. Intra-Eocene folding was particularly active in the southwestern part of the Faroe-Shetland region, around the Munkagrunnur Ridge and Judd area. However, there is also evidence of episodic intra-Eocene and later uplift/deformation in the area around the northern Fugloy Ridge.

4. Neogene gateway development (Faroe Conduit) is favoured primarily on the basis of regional evidence of basin restriction and faunal separation and uncertainty concerning the nature of the 'Southeast Faroes drift'.

5. A significant phase of Miocene folding is associated with the Intra-Miocene Unconformity; the Mid Miocene Unconformity forms a relatively minor break with a restricted distribution in the NE of the Faroe-Shetland region. Lateral offsets and basin inversion associated with the Miocene folds indicate a strong structural inheritance from the underlying rift architecture.

6. Regional unconformity development appears to correlate key periods of deformation or peaks in tectonic activity within the wider region. There is a broad coincidence between the timing of formation of the unconformities in the Faroe–Shetland region and plate reorganisation events/changes in plate movement in the adjacent Norway Basin. Variations in gravitational potential energy associated with growth of the Iceland Insular Margin from Miocene times and with the transient changes in ridge elevation associated with temperature (buoyancy) variations of the Iceland plume, together with far field stresses associated with, for example, collision between Eurasia and Iberia may also have exerted significant influence on deformation within the Faroe-Shetland region.

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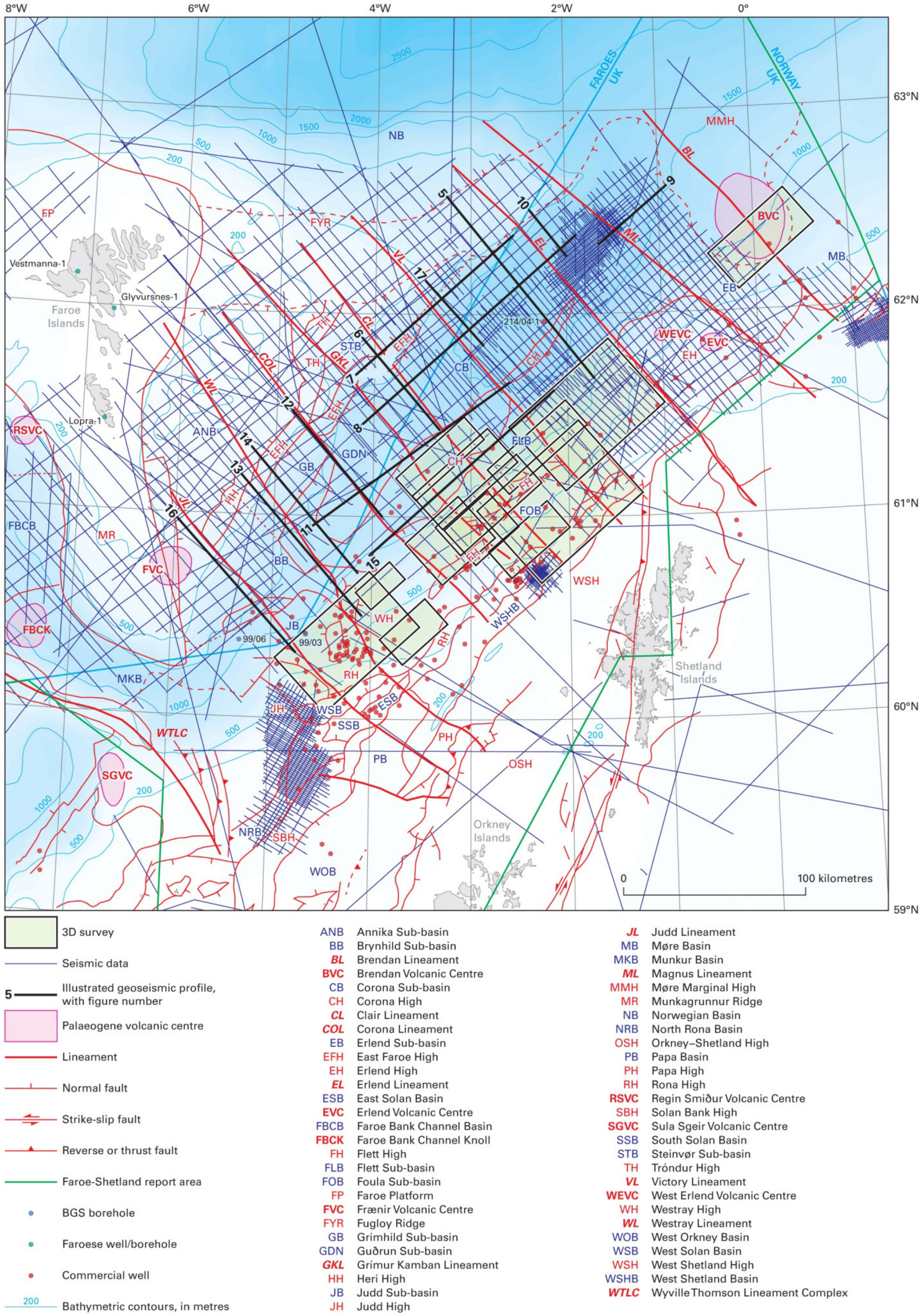

Figure 1 Seismic data location map; locations of structural elements, Palaeogene volcanic centres and bathymetry from Ritchie et al. (2011). Numbers on the illustrated geoseismic profiles refer to figure numbers. Positions of wells from CDA.

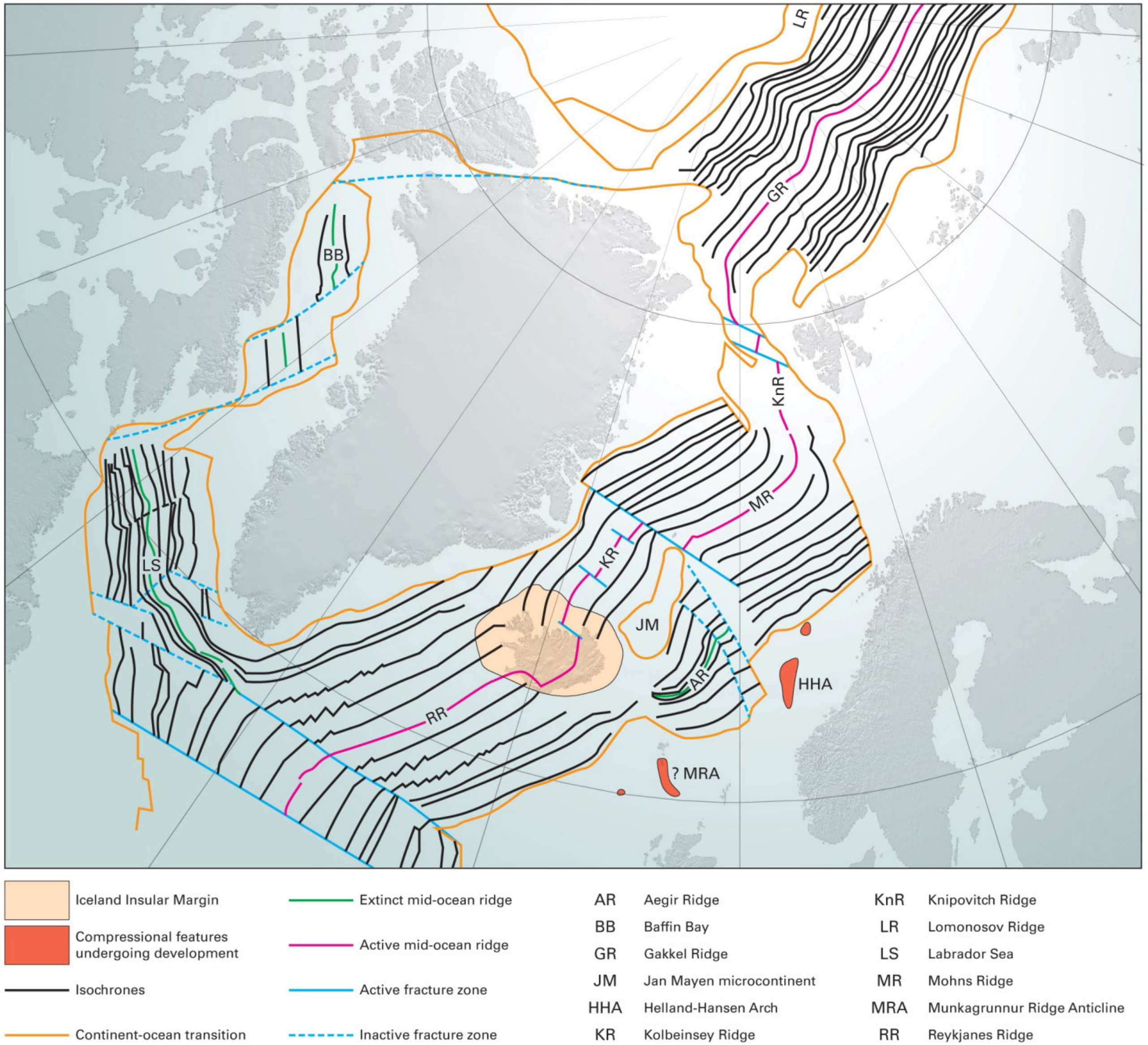

Figure 2 Present day plate configuration in the NE Atlantic (modified after Doré et al., 2008).

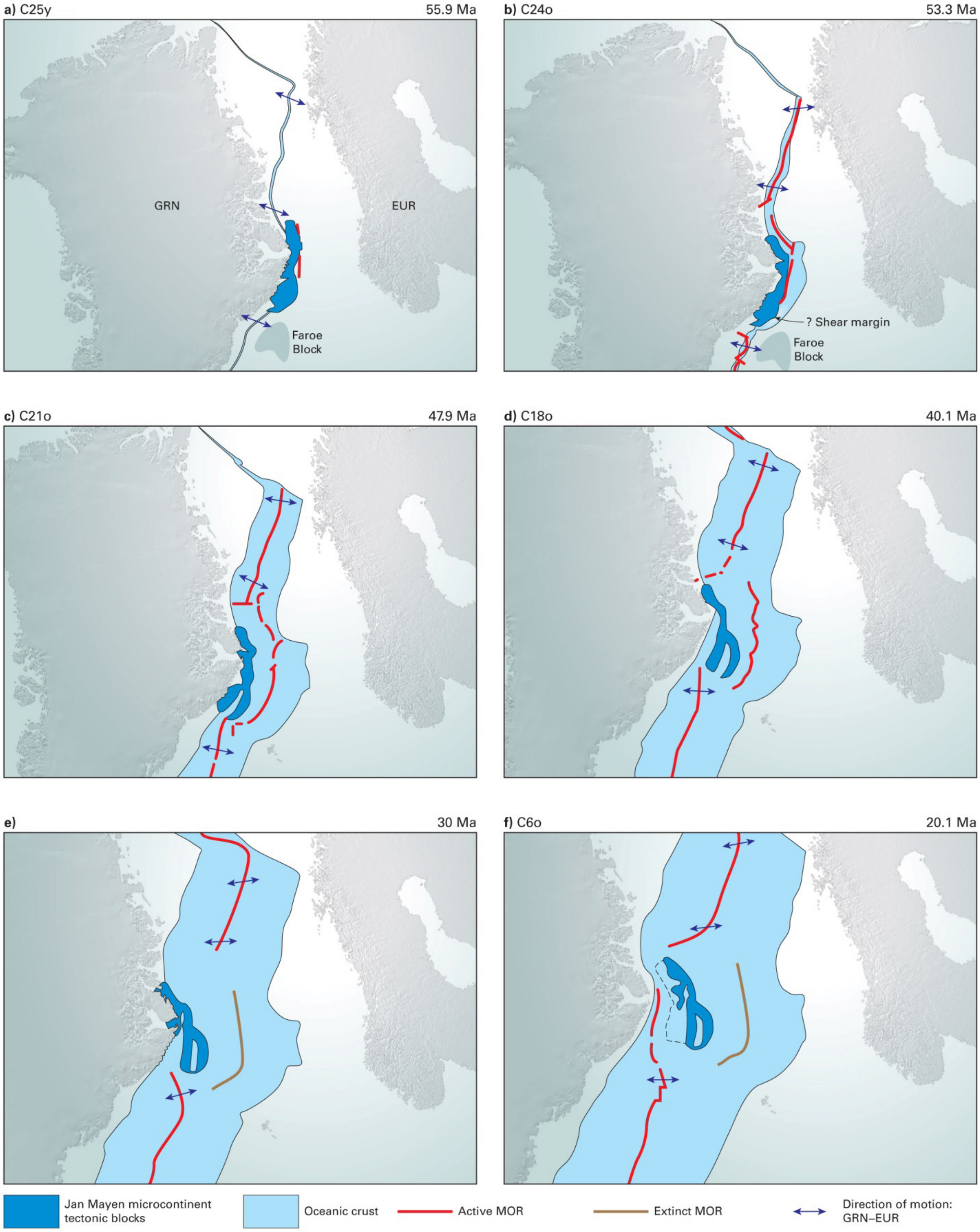

Figure 3 Tectonic reconstructions illustrating evolution of NE Atlantic plate boundaries and kinematic evolution of the Jan Mayen microcontinent (modified after Gaina et al., 2009).

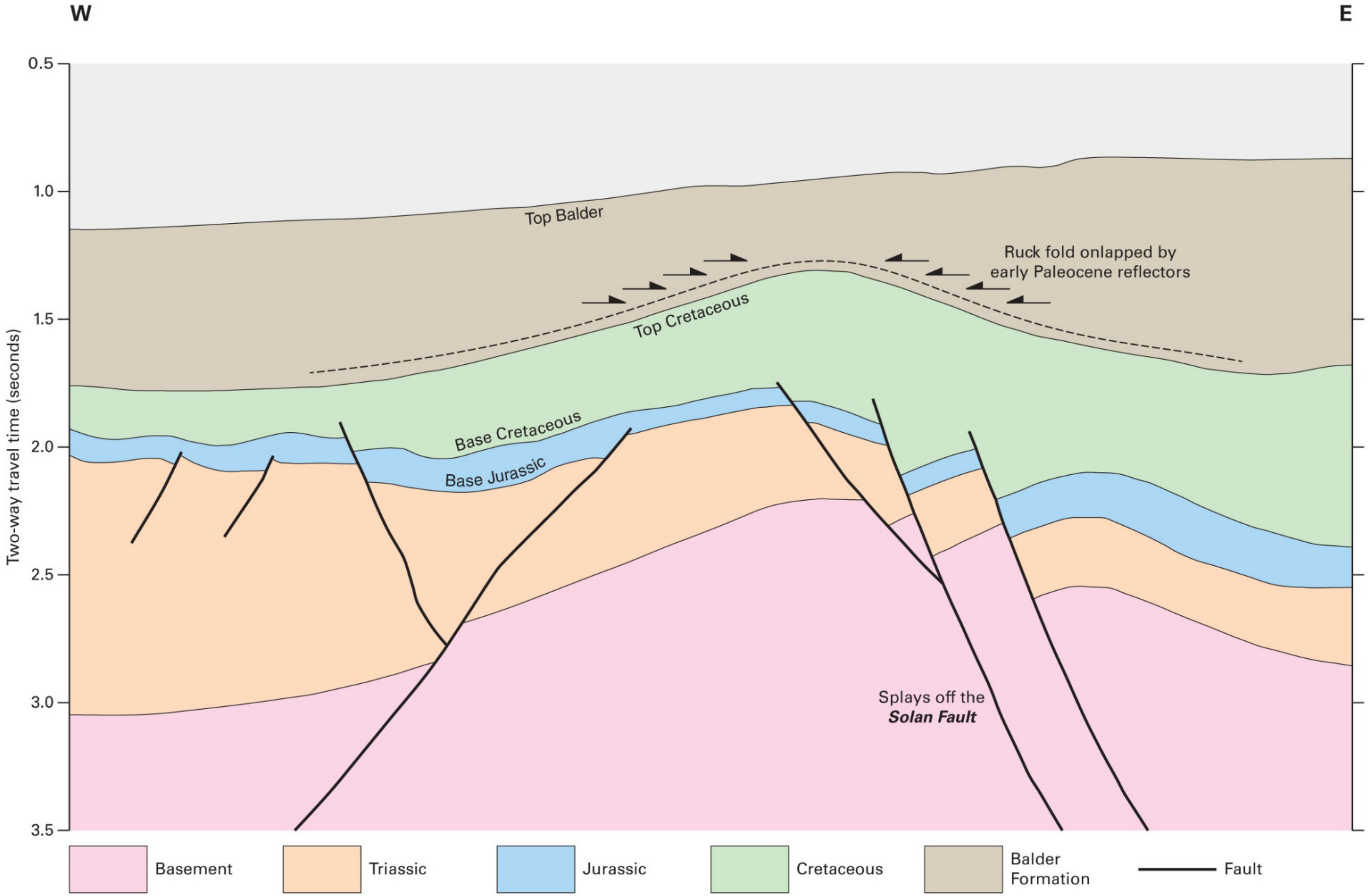

Figure 4 Geoseismic profile illustrating pre-breakup folding in the West Solan Basin (modified after Booth et al., 1993).

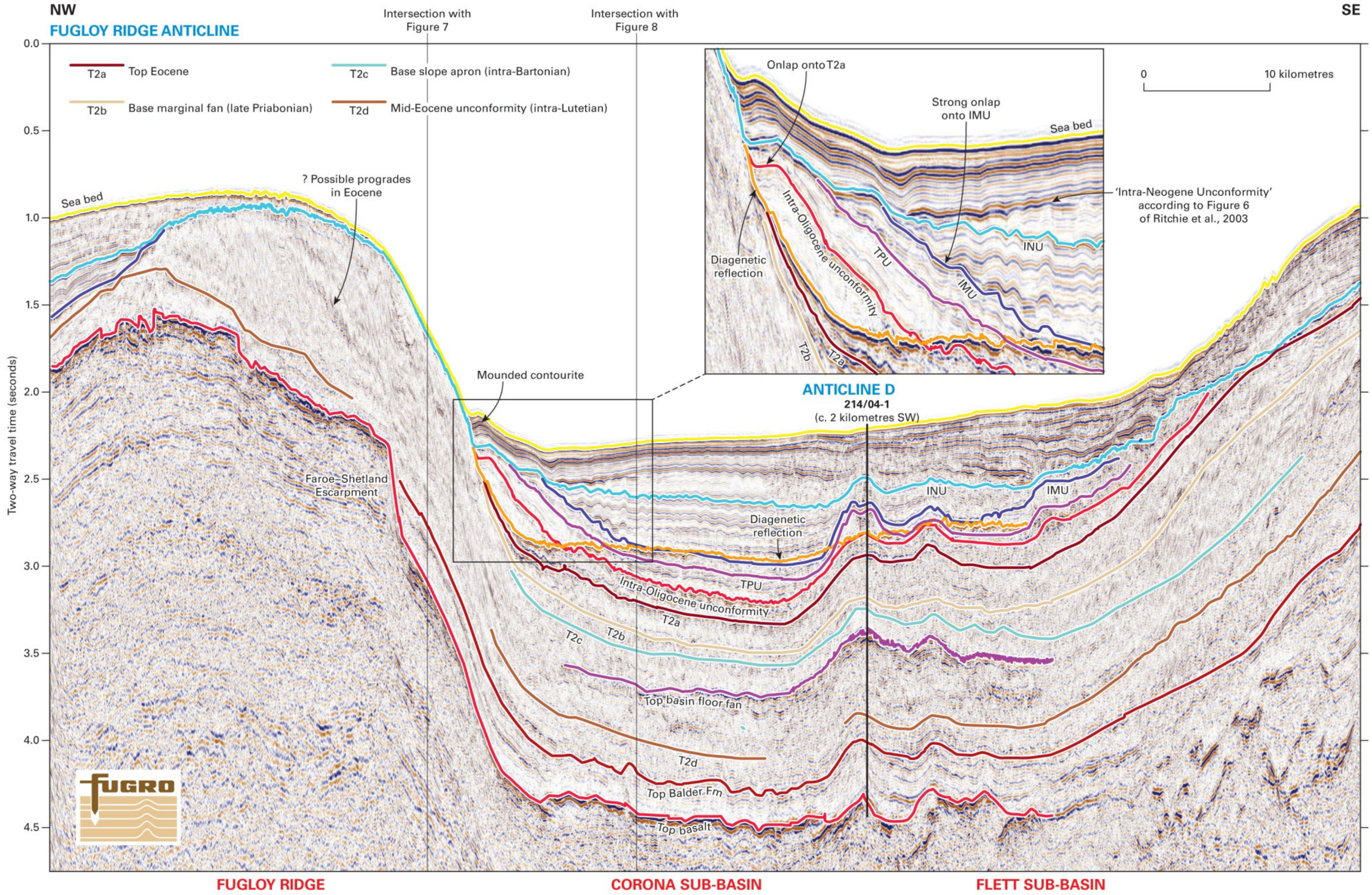

Figure 5 Geoseismic profile illustrating well tie 214/04-1.

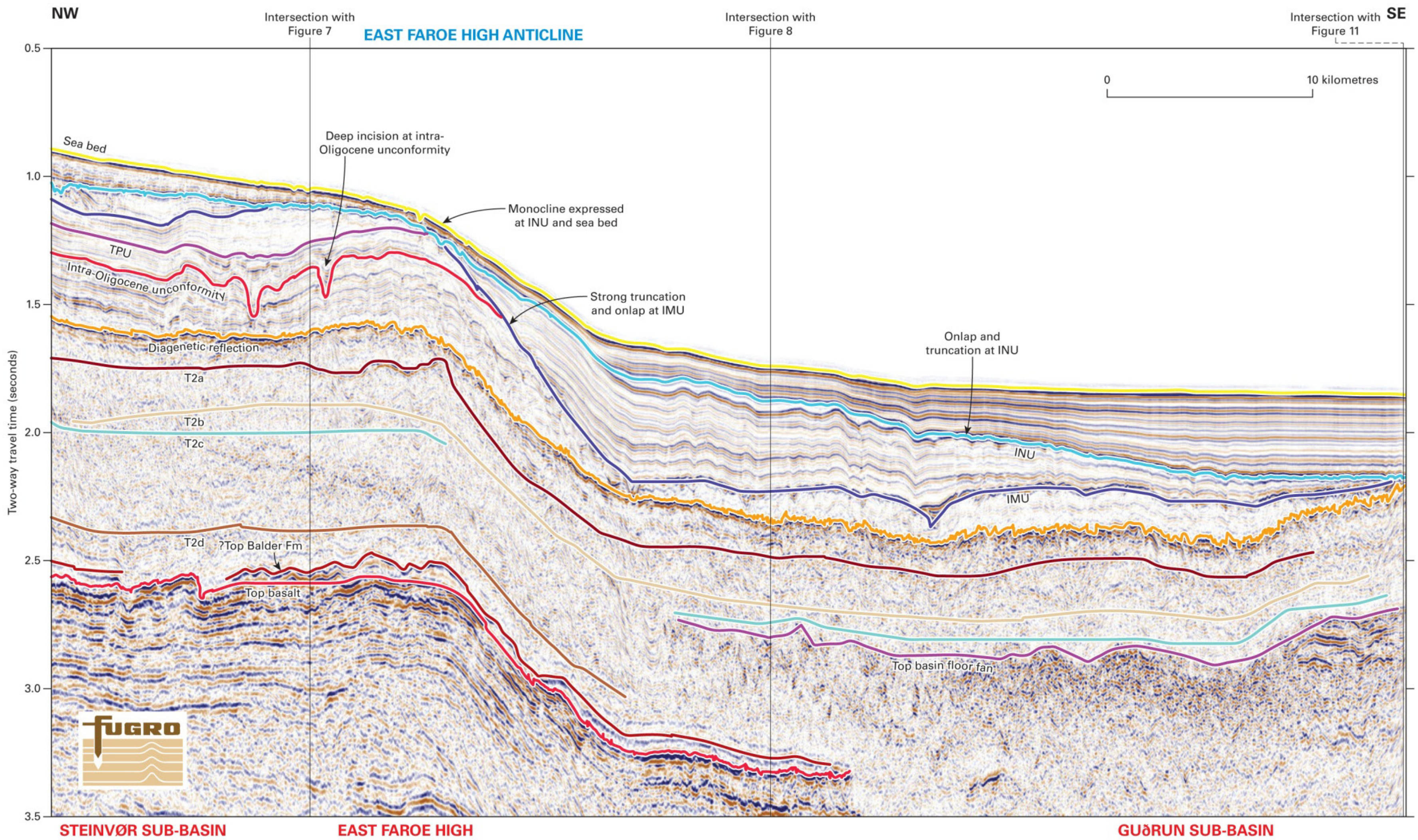

Figure 6 Geoseismic profile illustrating dip section near well 214/04-1.

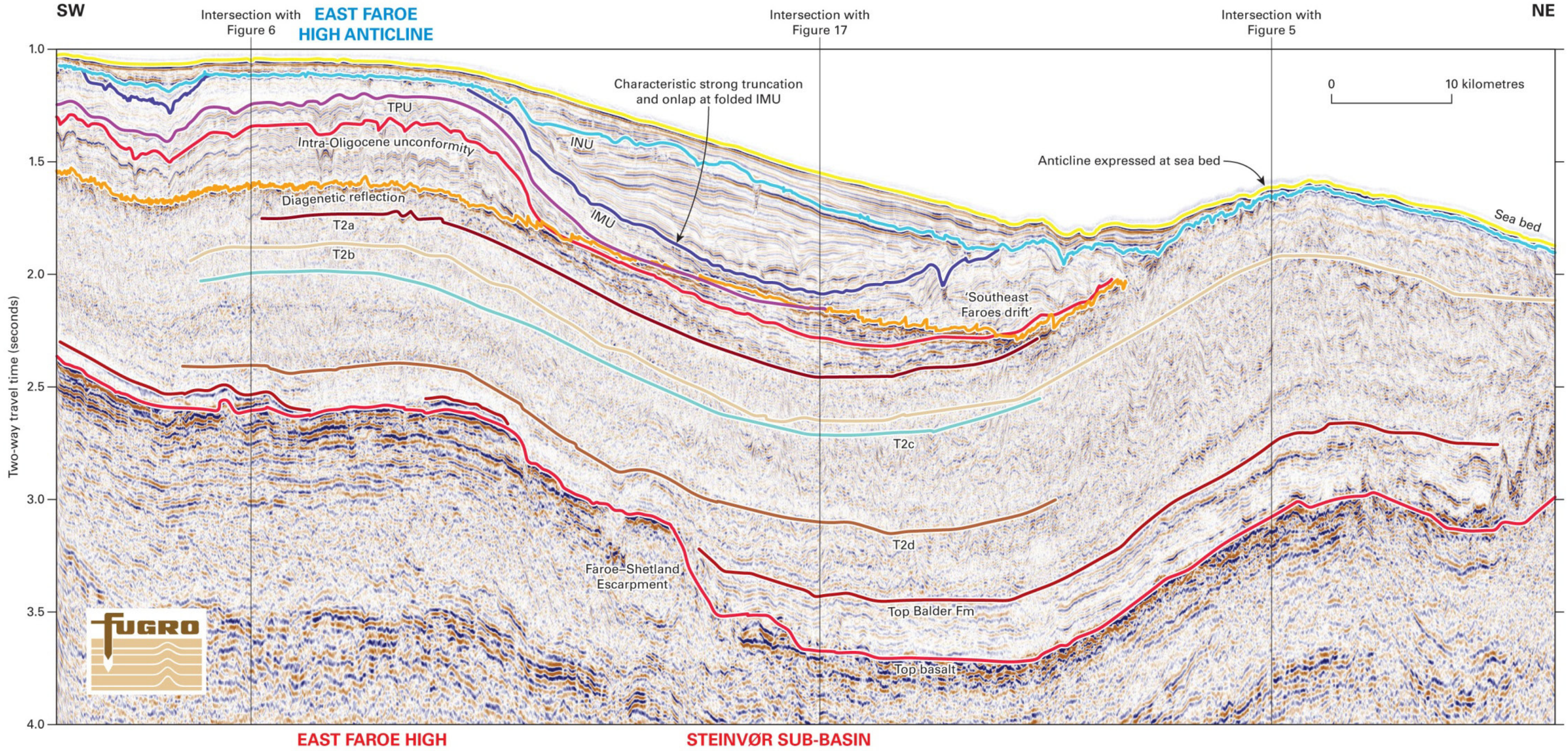

Figure 7 Geoseismic profile illustrating strike section near well 214/04-1.

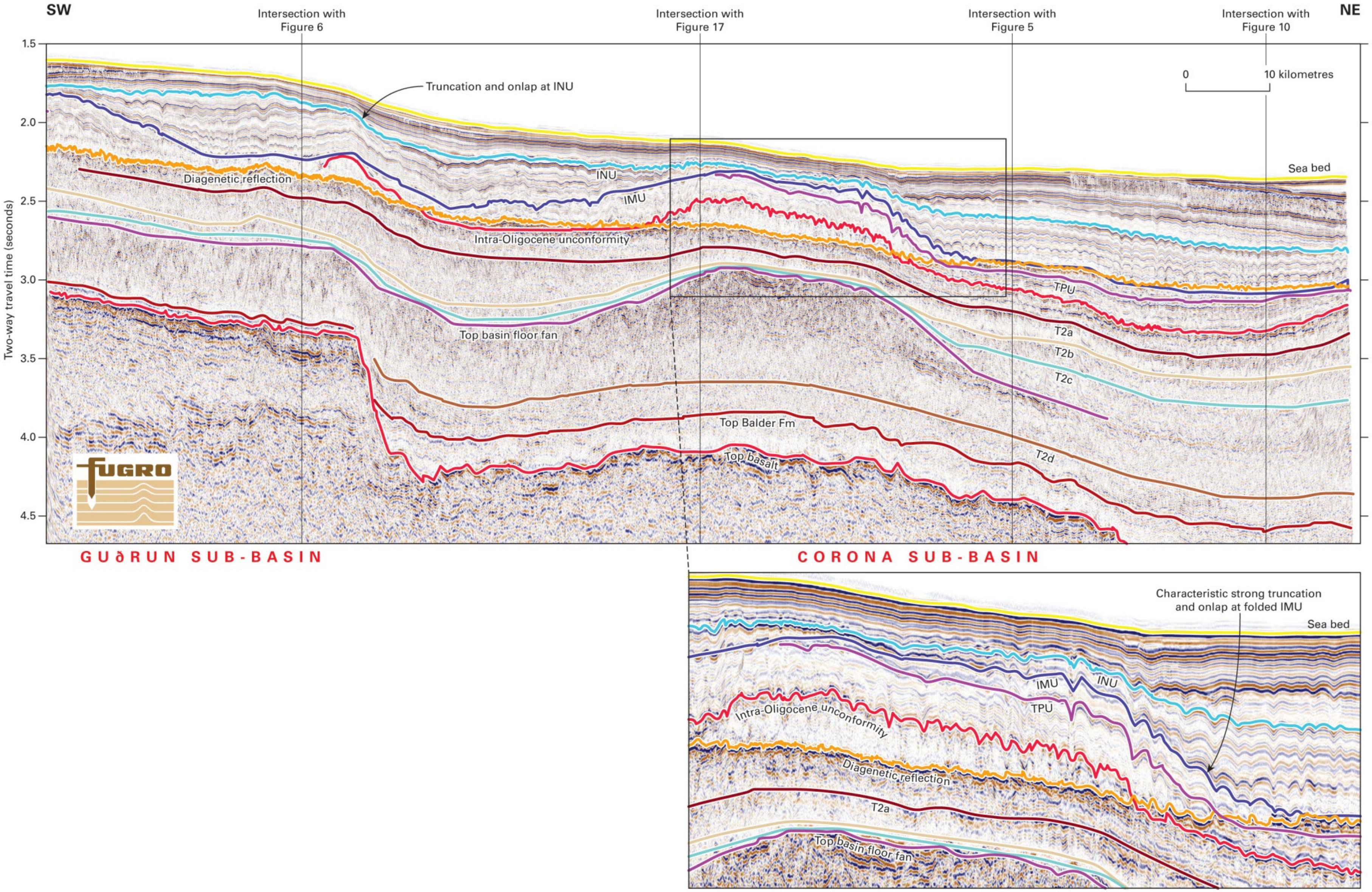

Figure 8 Geoseismic profile illustrating the Intra-Miocene Unconformity.

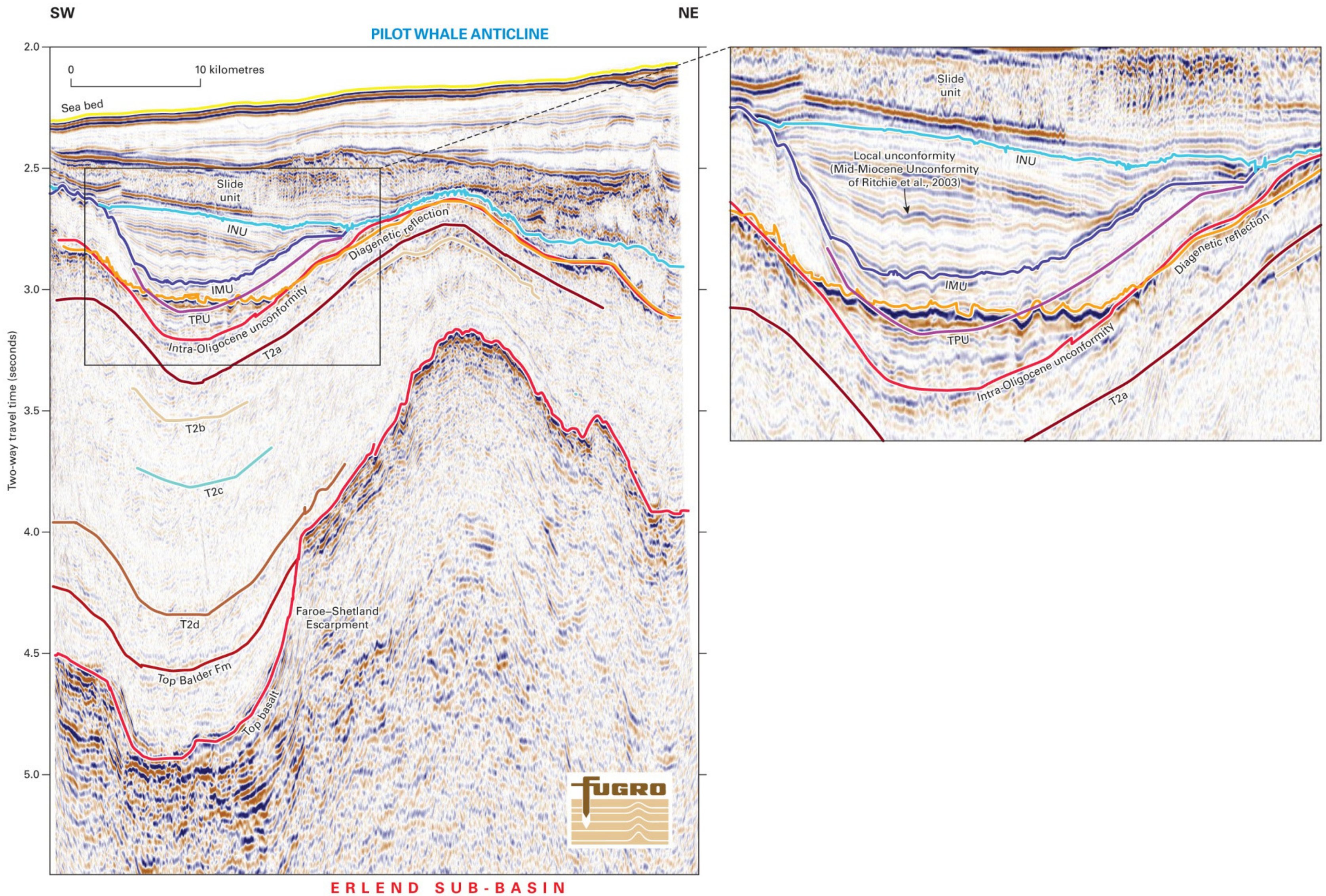

Figure 9 Geoseismic profile illustrating the Mid-Miocene Unconformity.

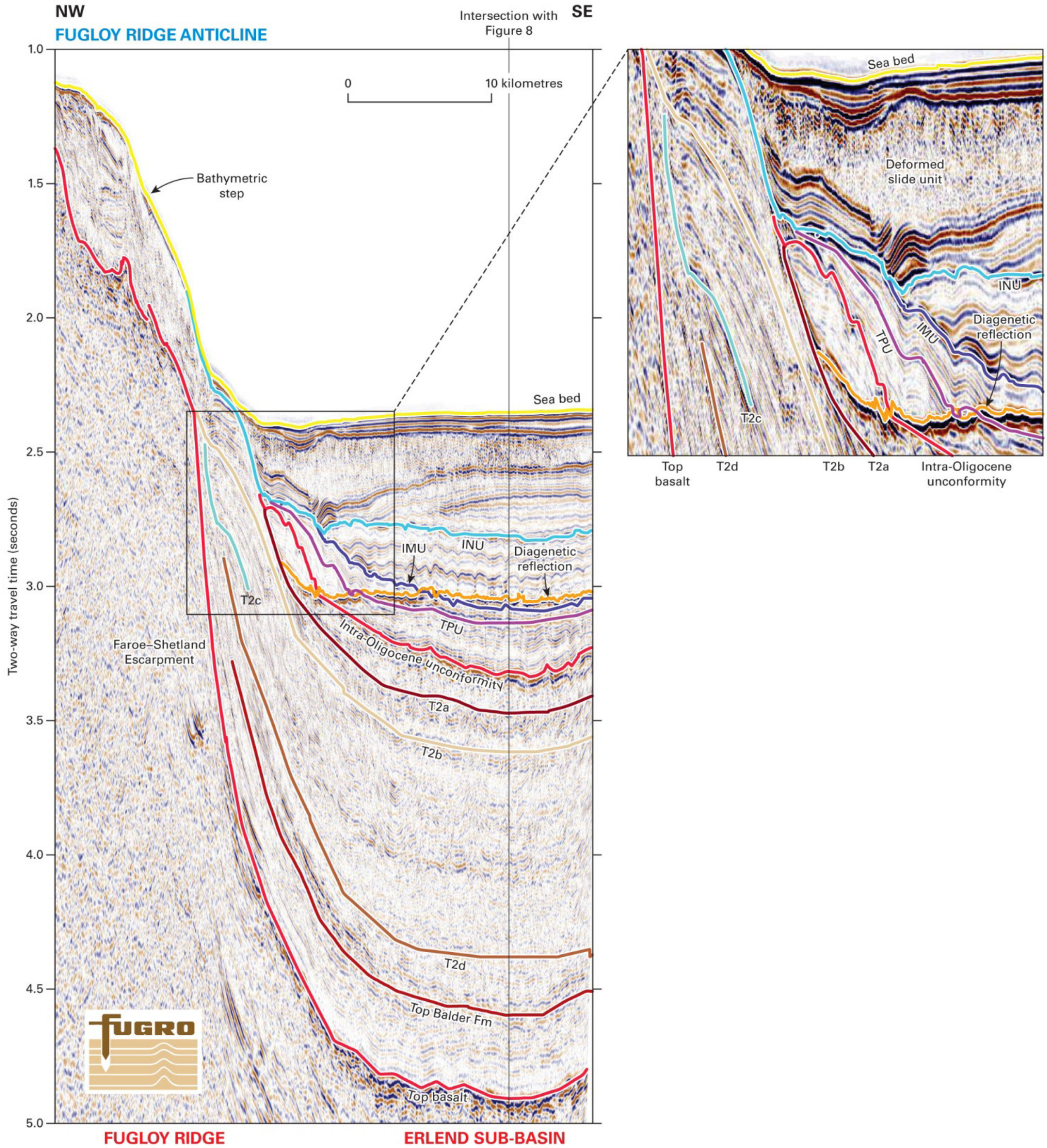

Figure 10 Geoseismic profile illustrating evidence of episodic uplift of the northern Fugloy Ridge.

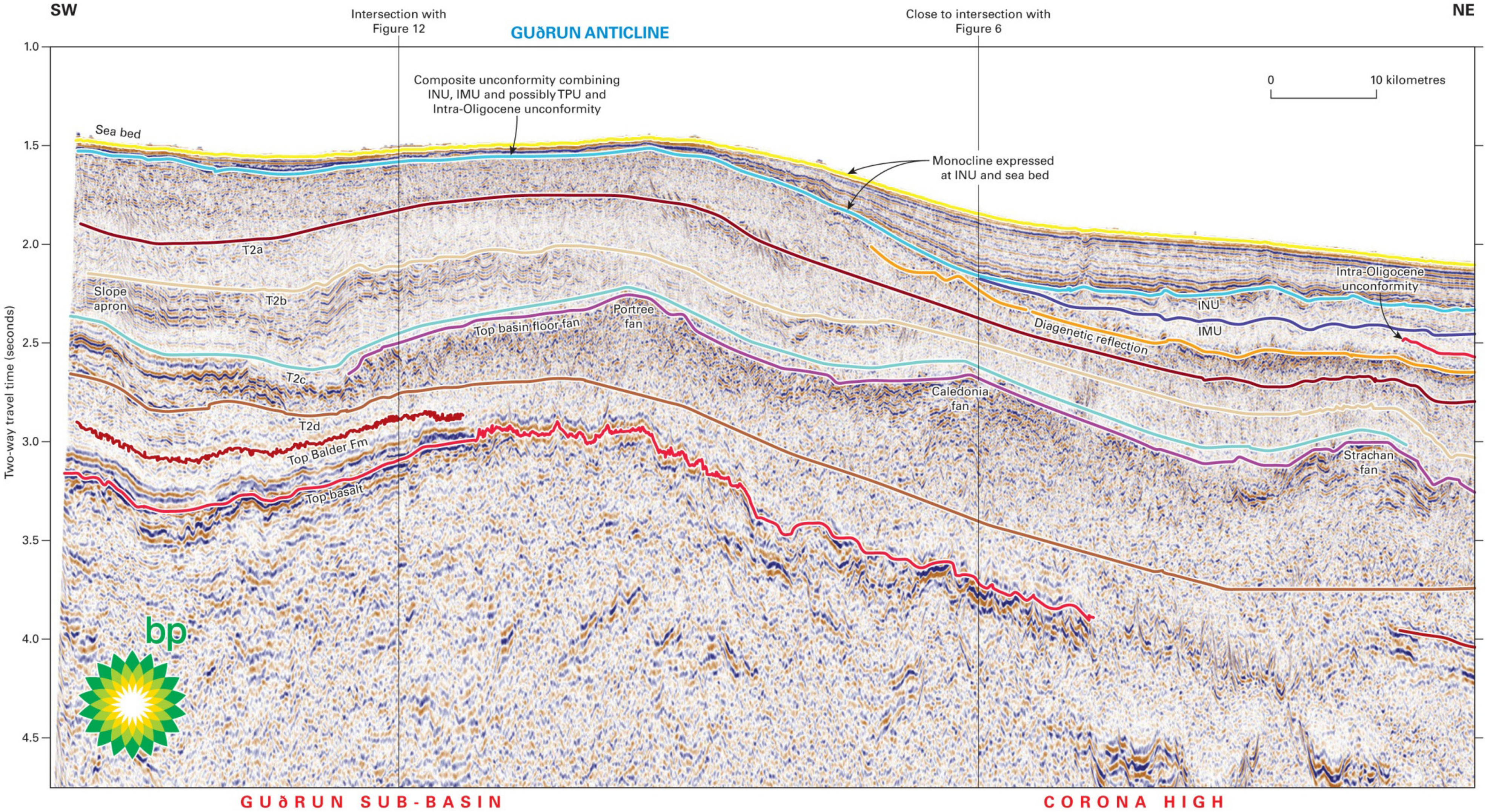

Figure 11 Geoseismic profile illustrating recent deformation and the seismic expression of Eocene tectonostratigraphic units.

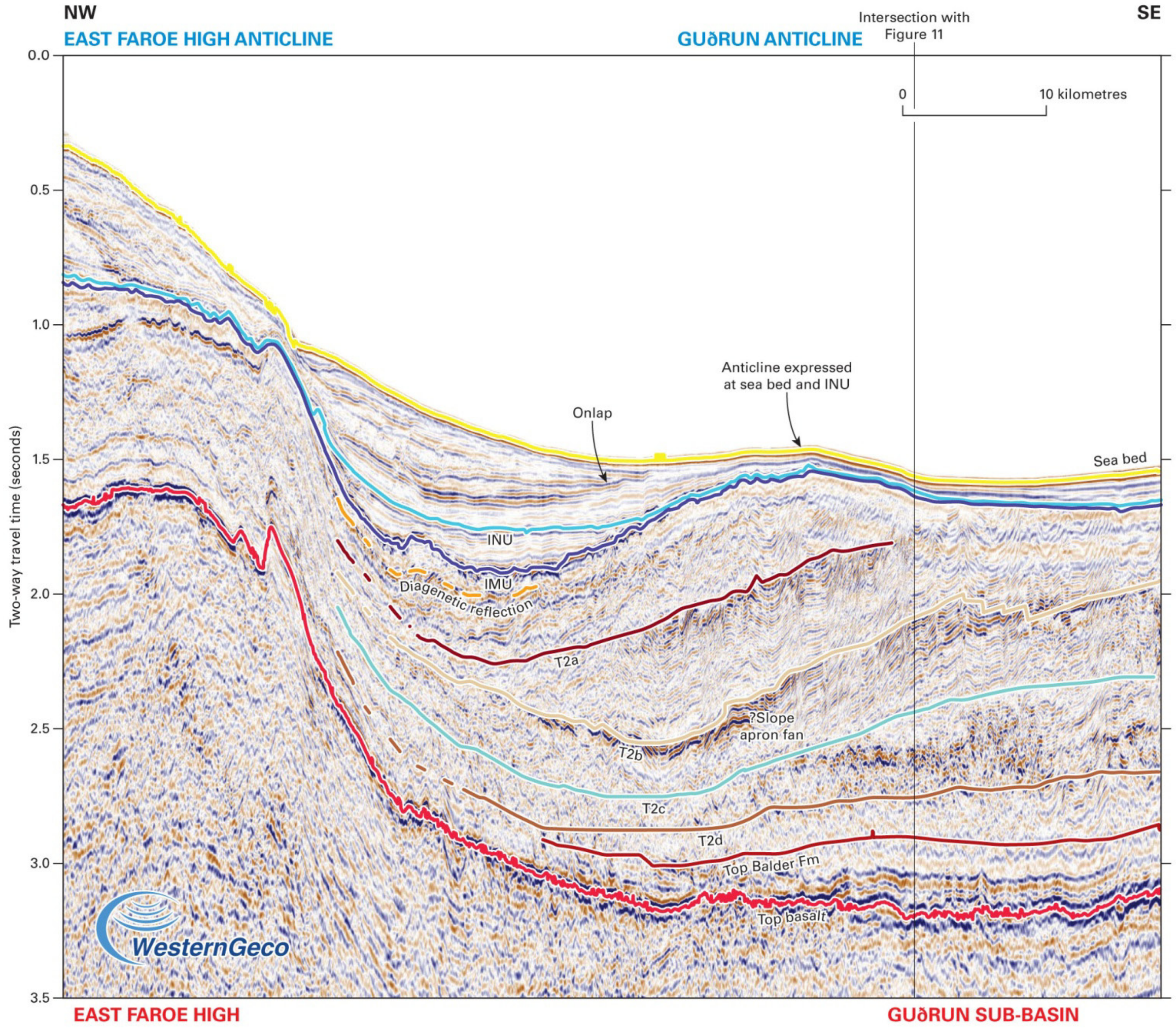

Figure 12 Geoseismic profile illustrating the 'Guðrun Anticline'.

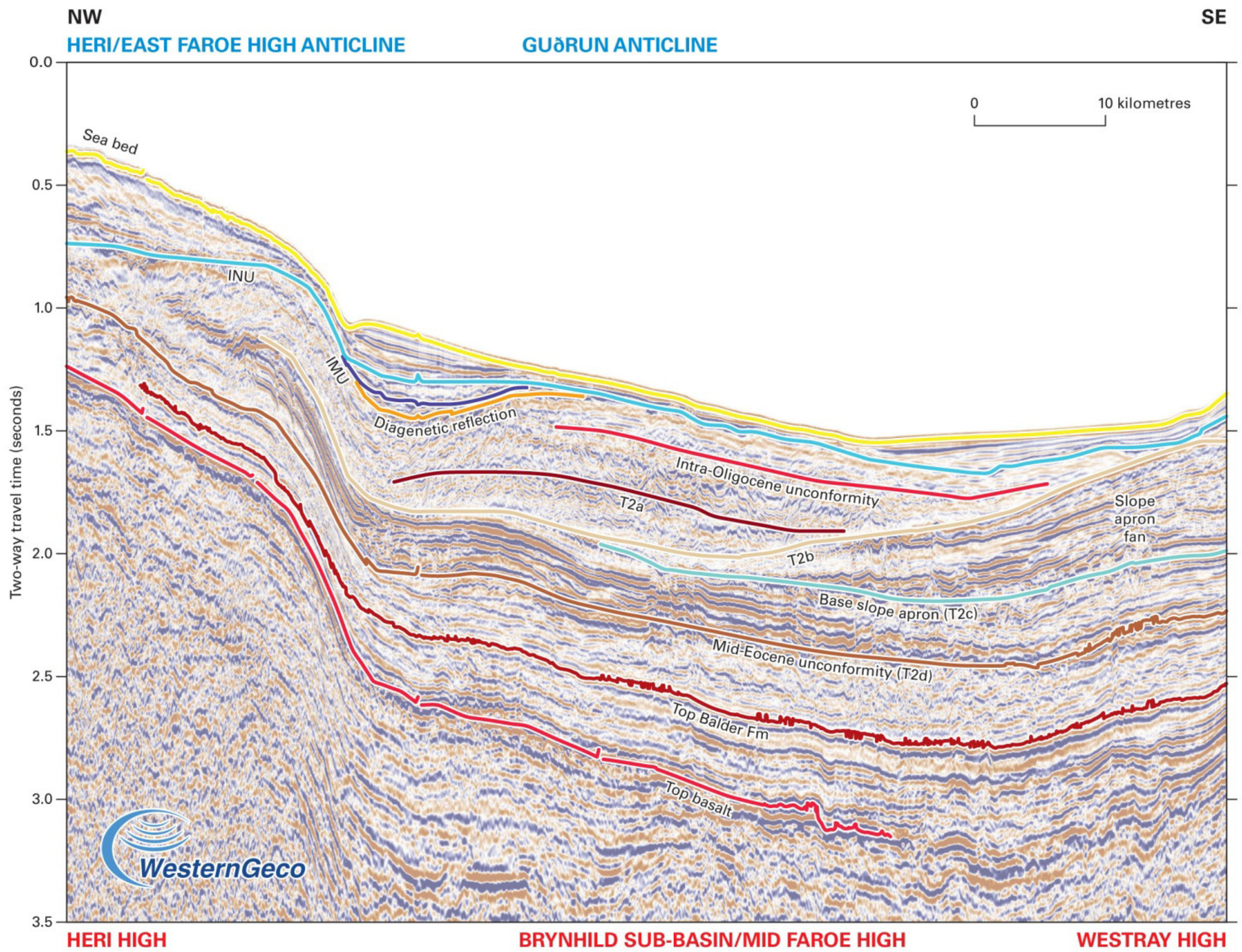

Figure 13 Geoseismic profile illustrating the seismic expression of Eocene tectonostratigraphic units, Brynhild Sub-basin/Mid Faroe High.

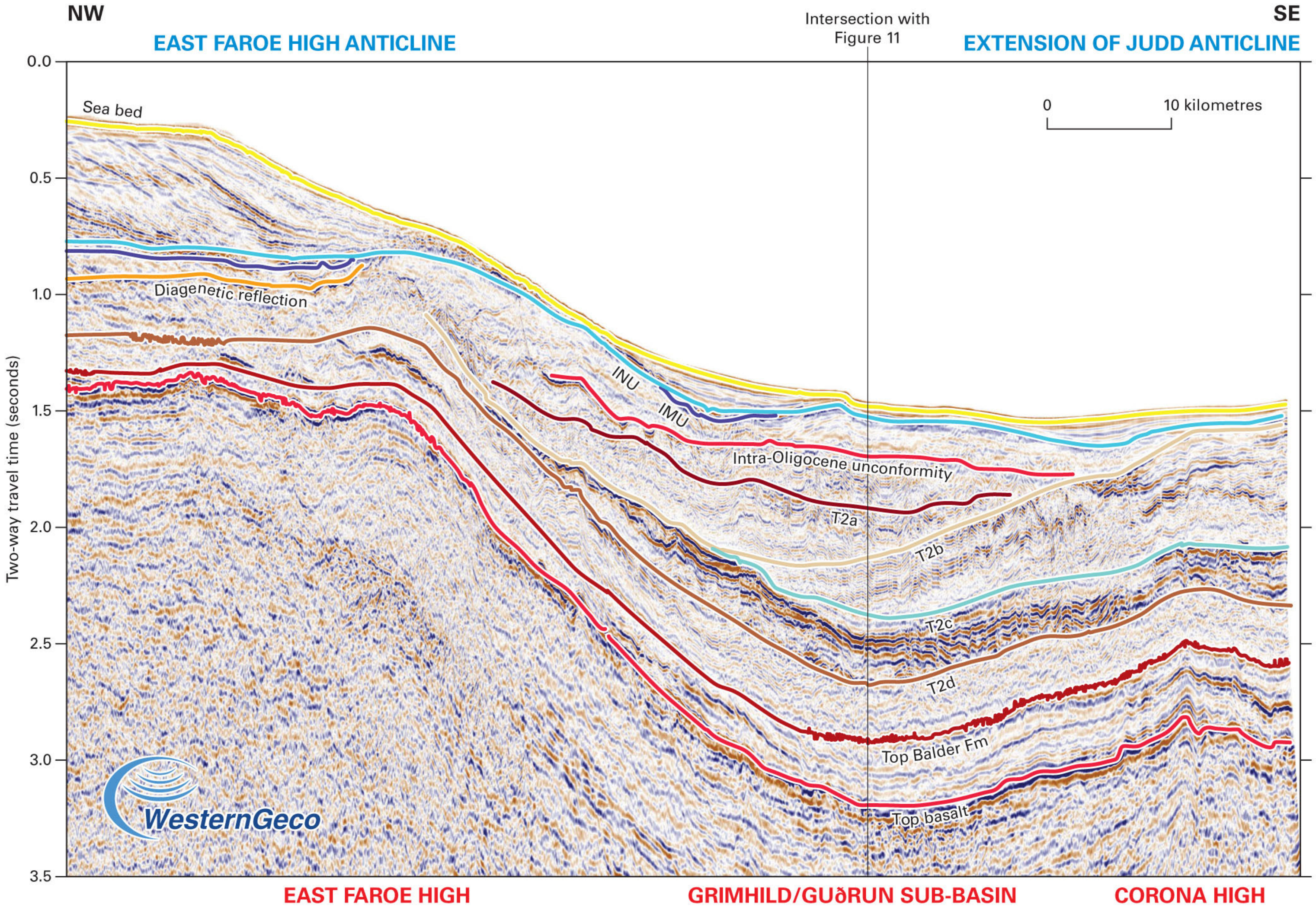

Figure 14 Geoseismic profile illustrating the seismic expression of Eocene tectonostratigraphic units, Grimhild / Gudrun Sub-basin.

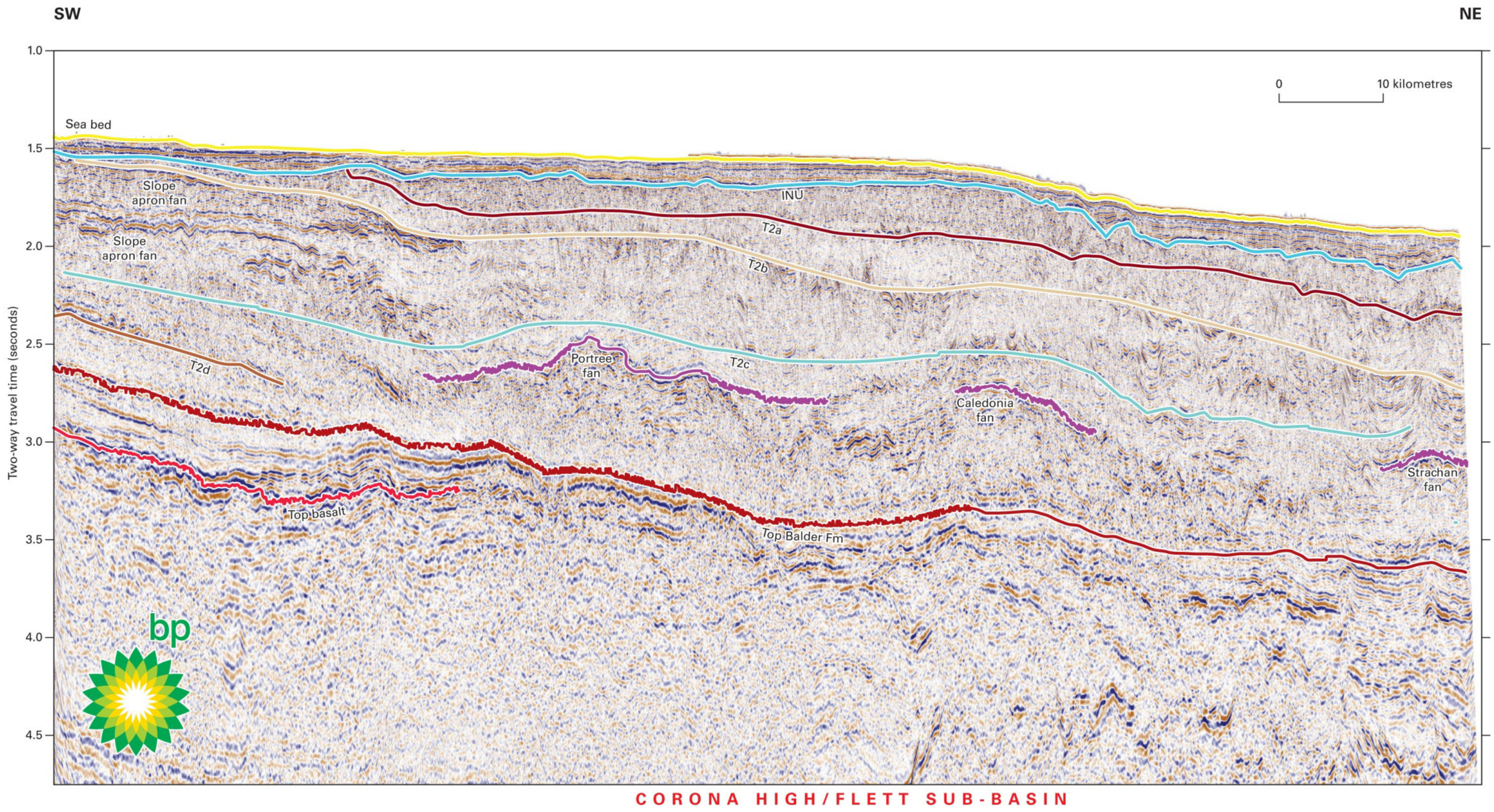

Figure 15 Geoseismic profile illustrating the seismic expression of Eocene tectonostratigraphic units, Corona High / Flett Sub-basin.

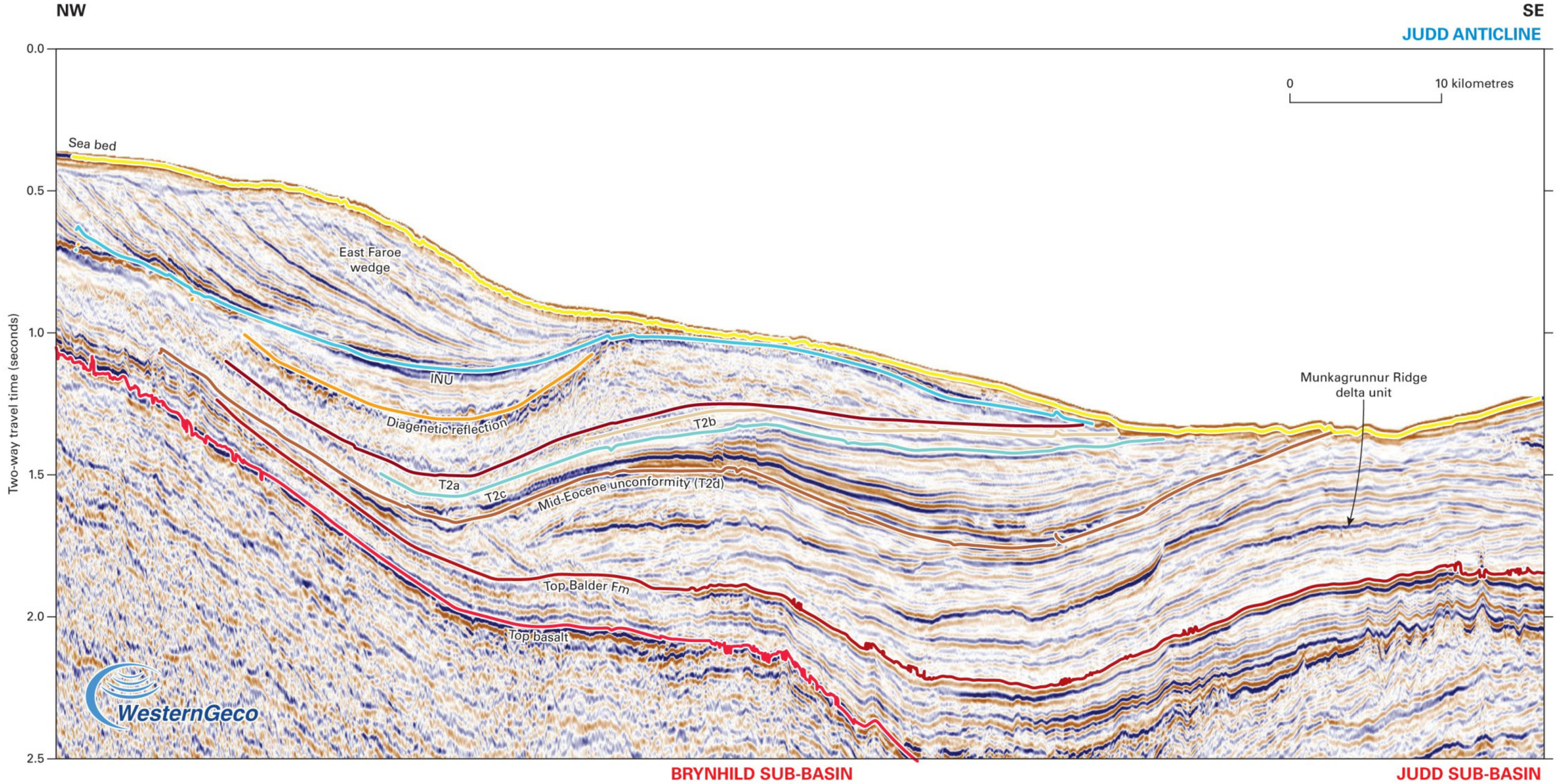

Figure 16 Geoseismic profile illustrating the Munkagrunnur Ridge Delta, Judd Sub-basin.

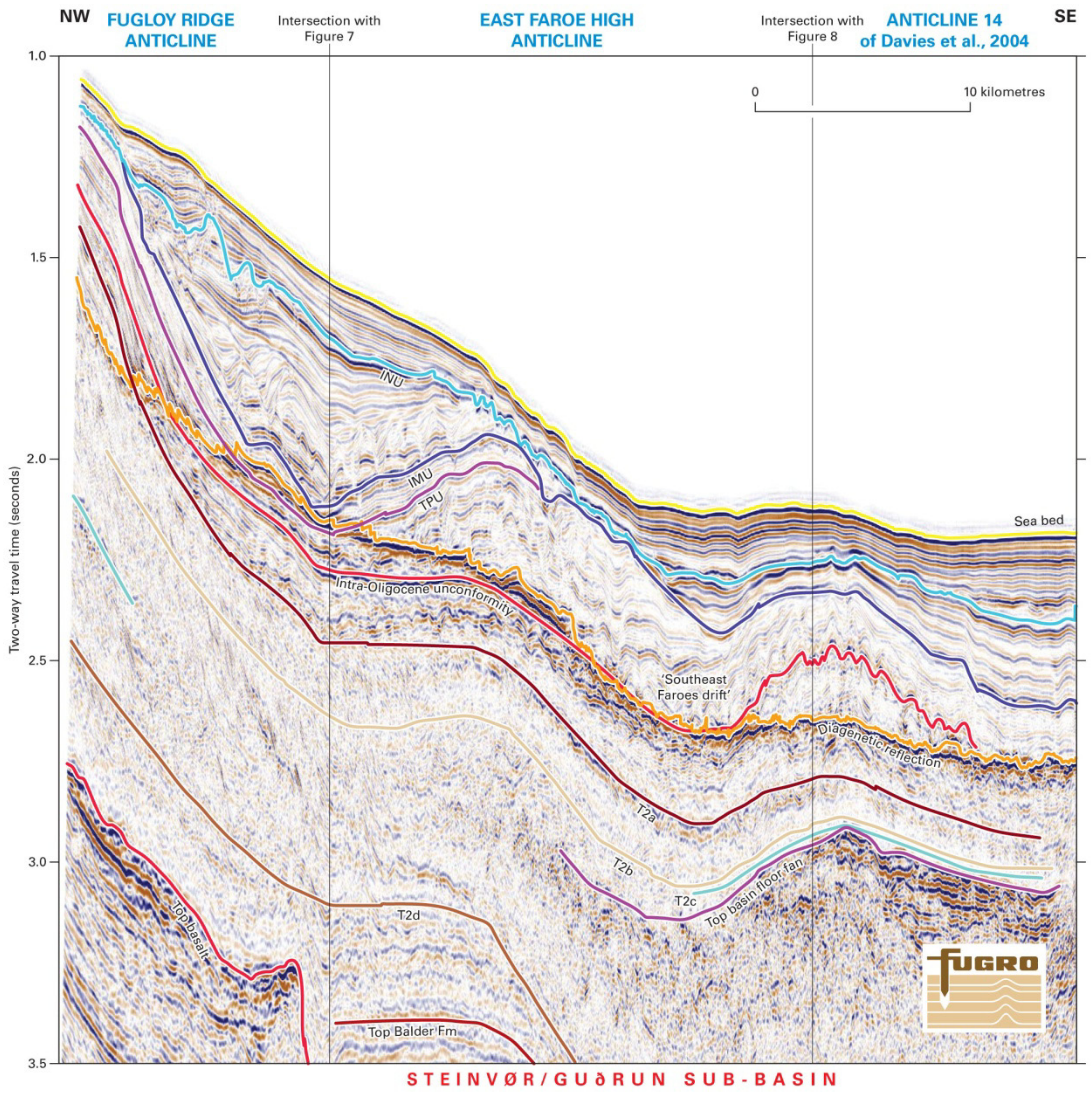

Figure 17 Geoseismic profile illustrating the 'Southeast Faroes drift', Steinvør / Guðrun Sub-basin.

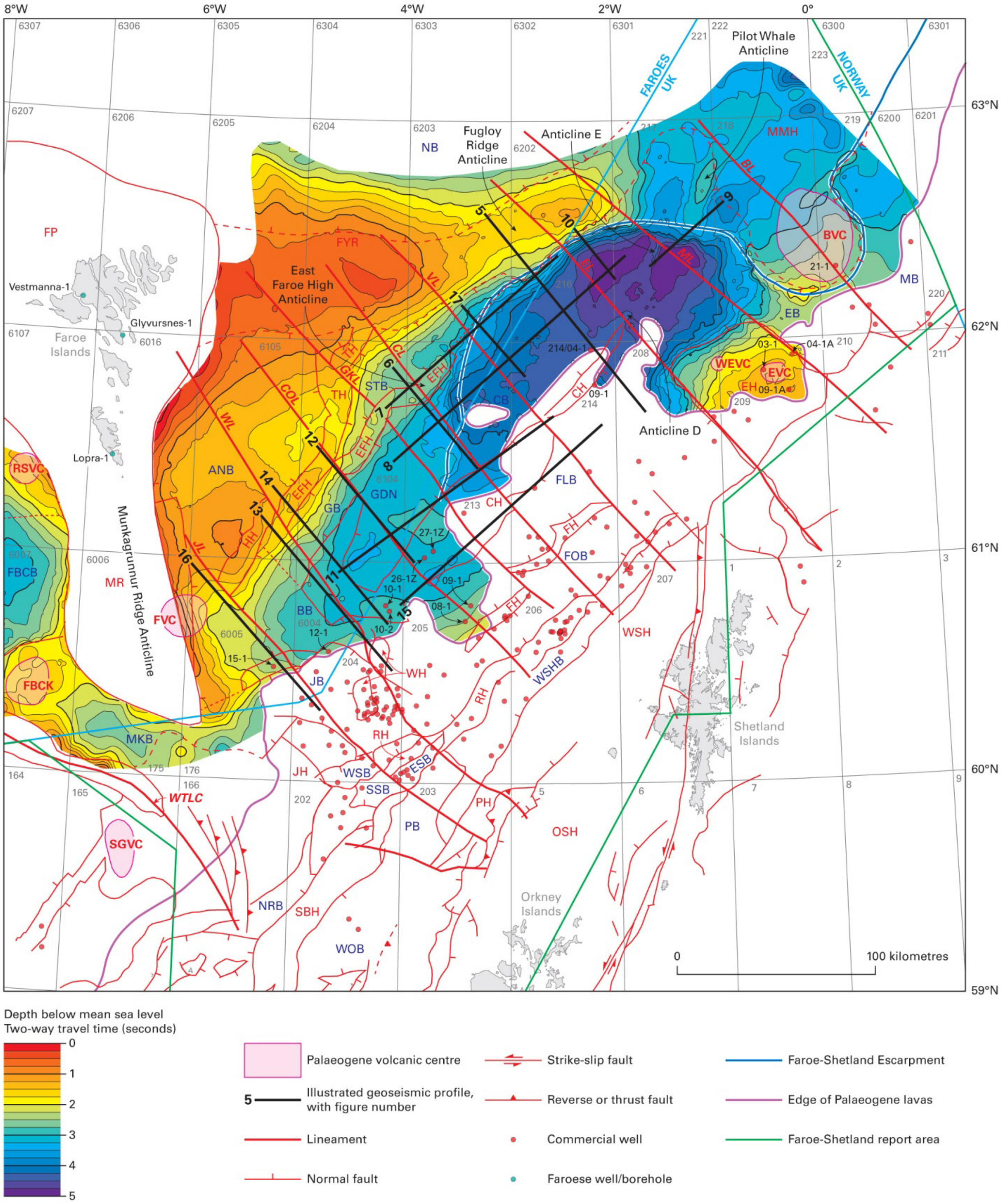

Figure 18 Structure contour map (two-way-travel-time in seconds): Top Palaeogene lavas. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

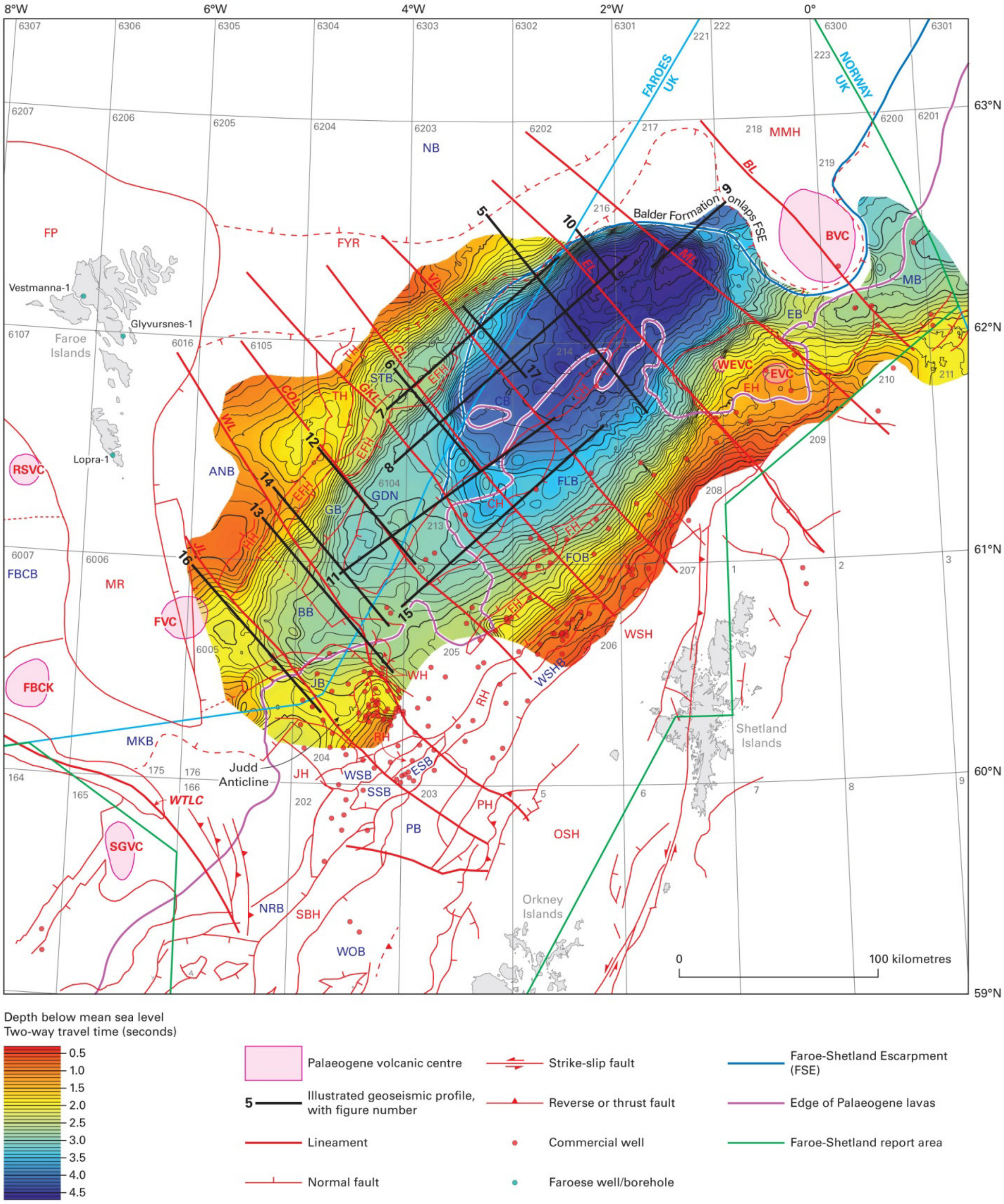

Figure 19 Structure contour map (two-way-travel-time in seconds): Top Balder Formation. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

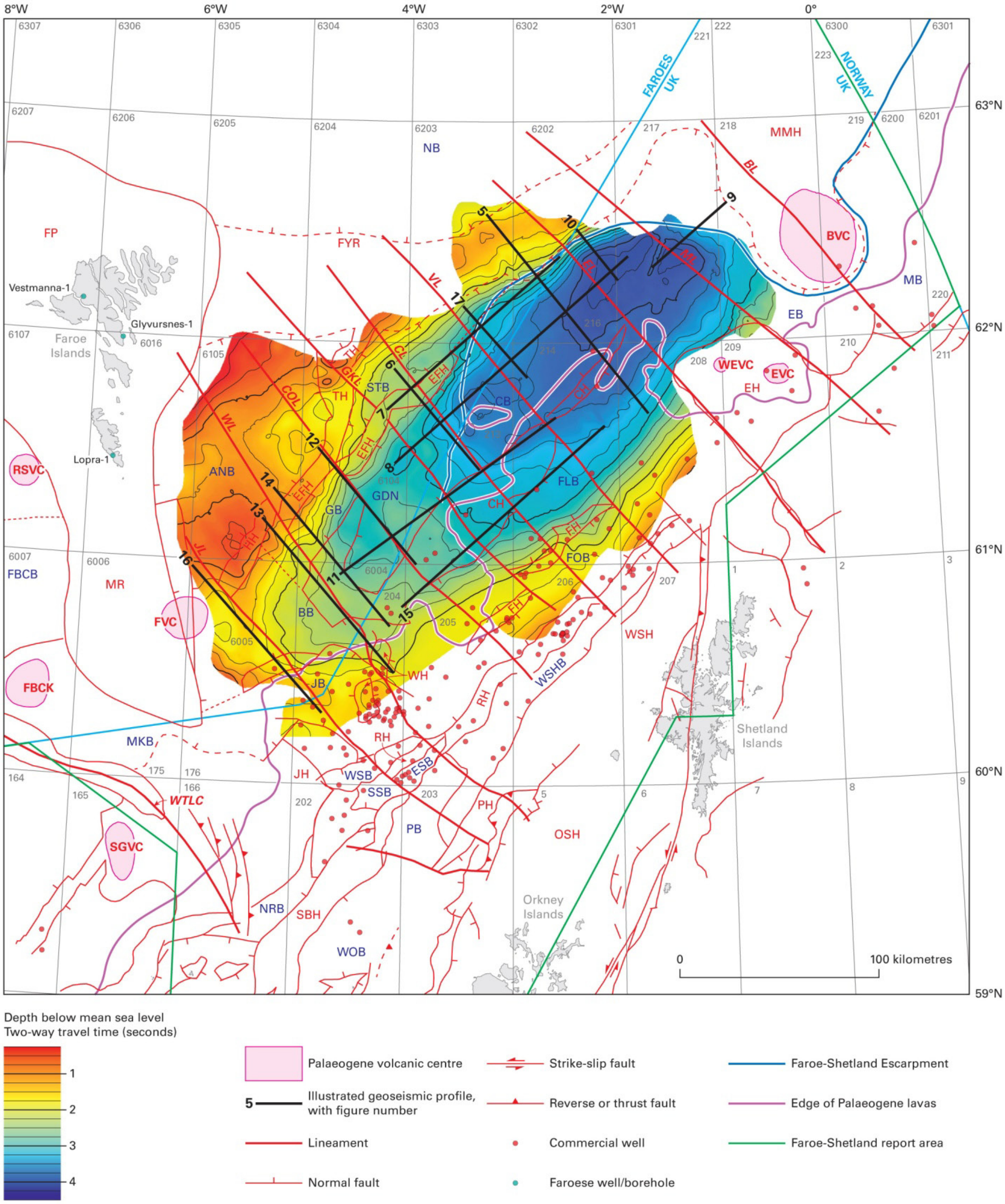

Figure 20 Structure contour map (two-way-travel-time in seconds): T2d (intra-Lutetian). Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

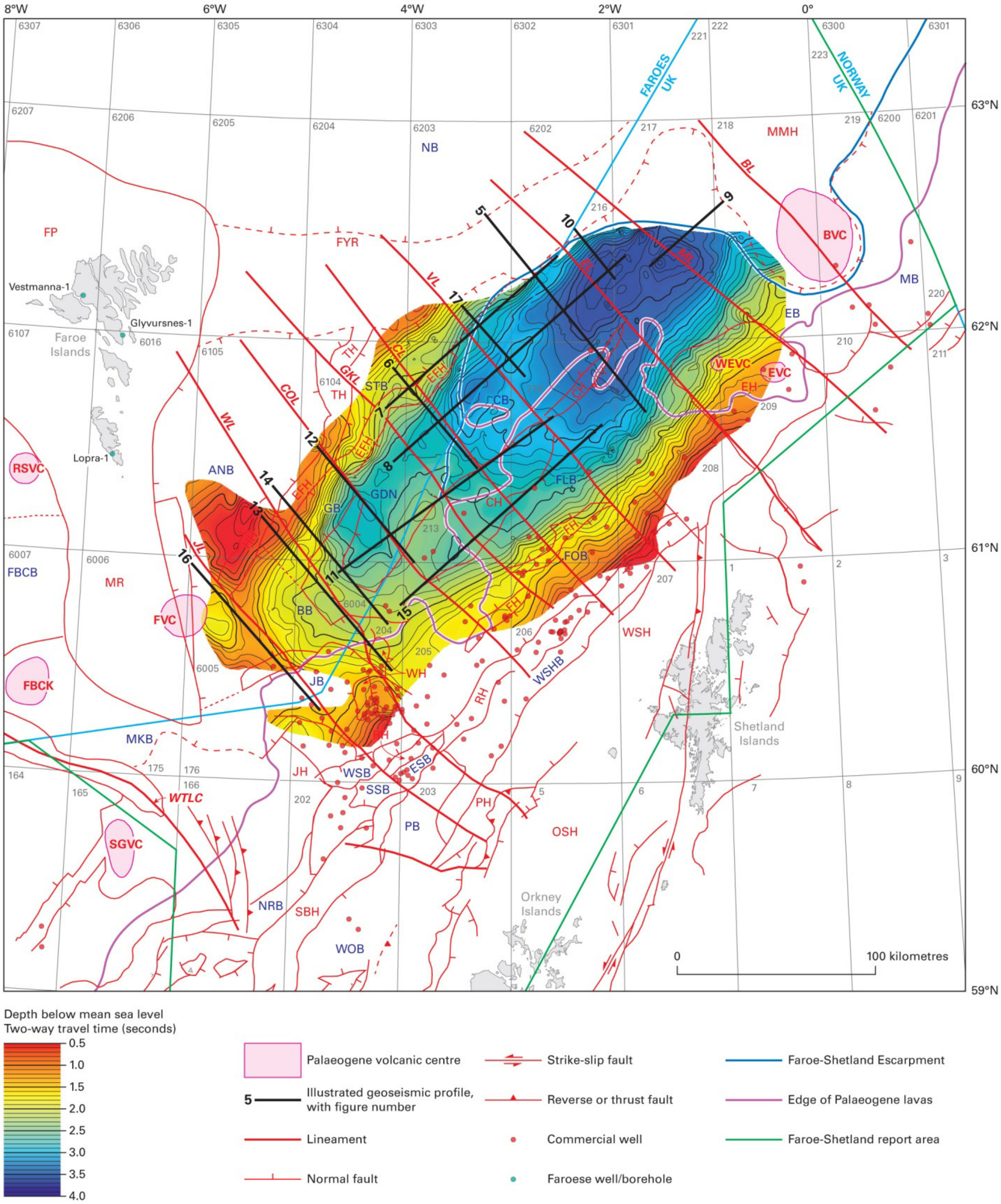

Figure 21 Structure contour map (two-way-travel-time in seconds): T2c (intra-Bartonian). Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

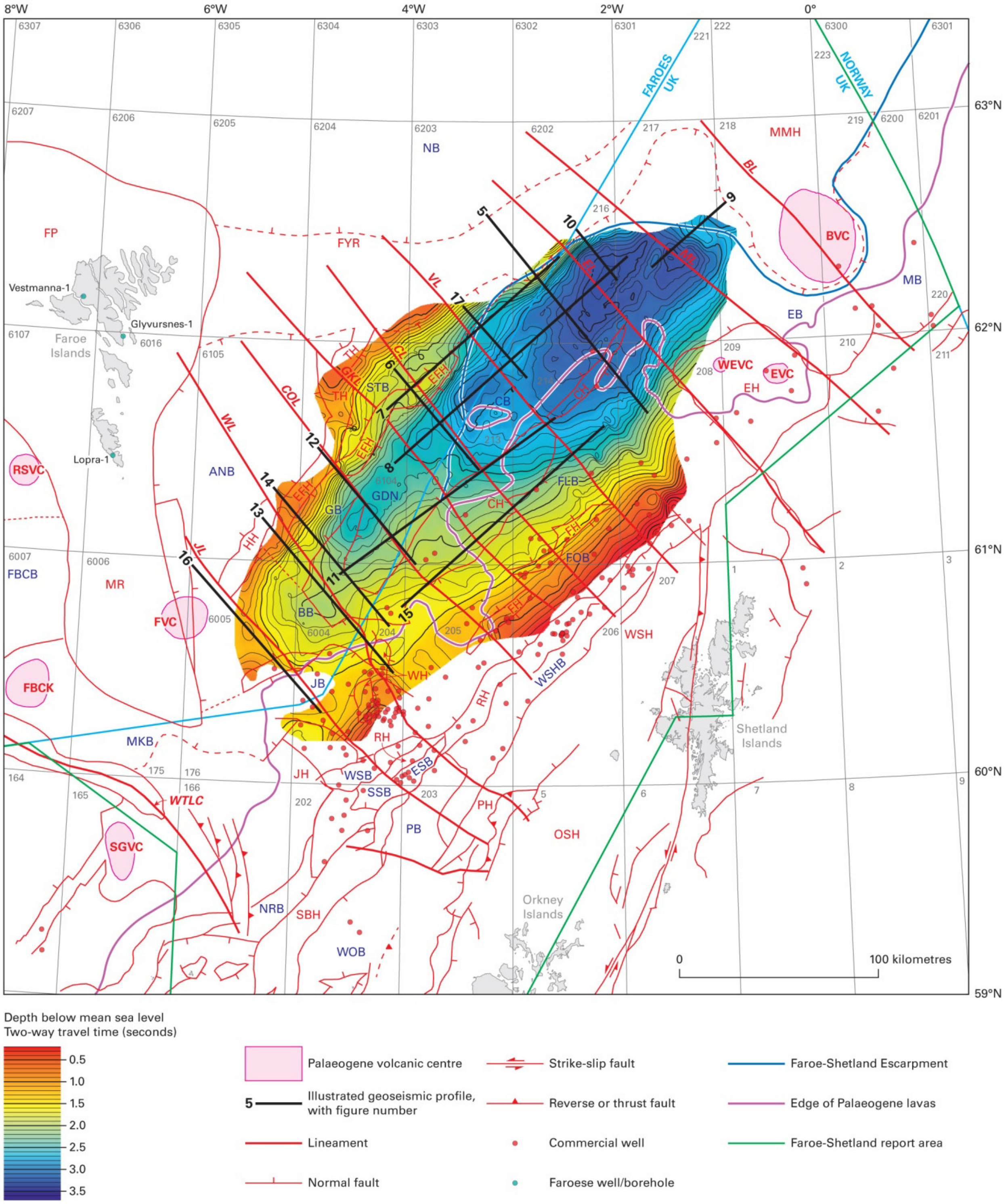

Figure 22 Structure contour map (two-way-travel-time in seconds): T2b (late Priabonian). Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

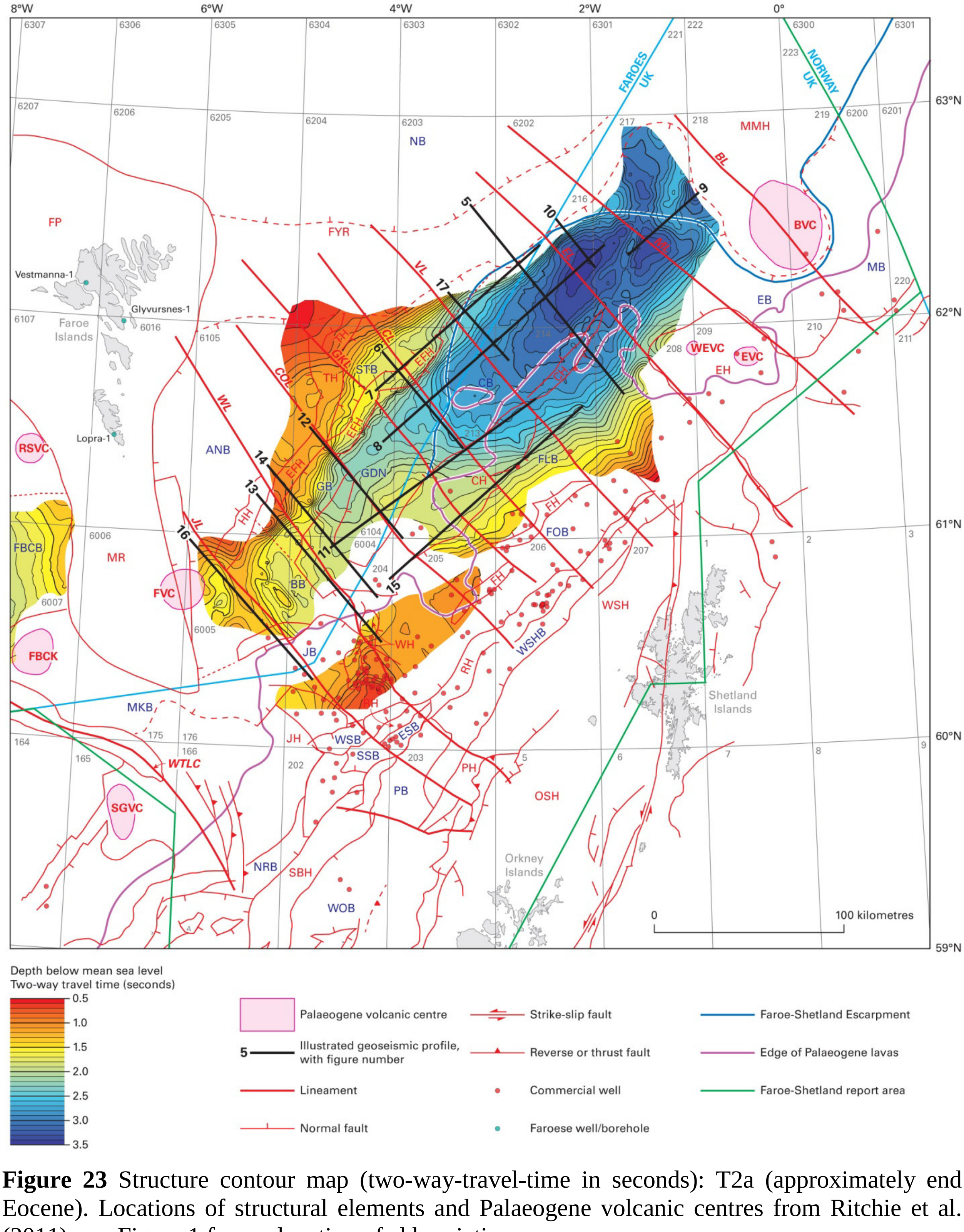

Figure 23 Structure contour map (two-way-travel-time in seconds): T2a (approximately end Eocene). Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

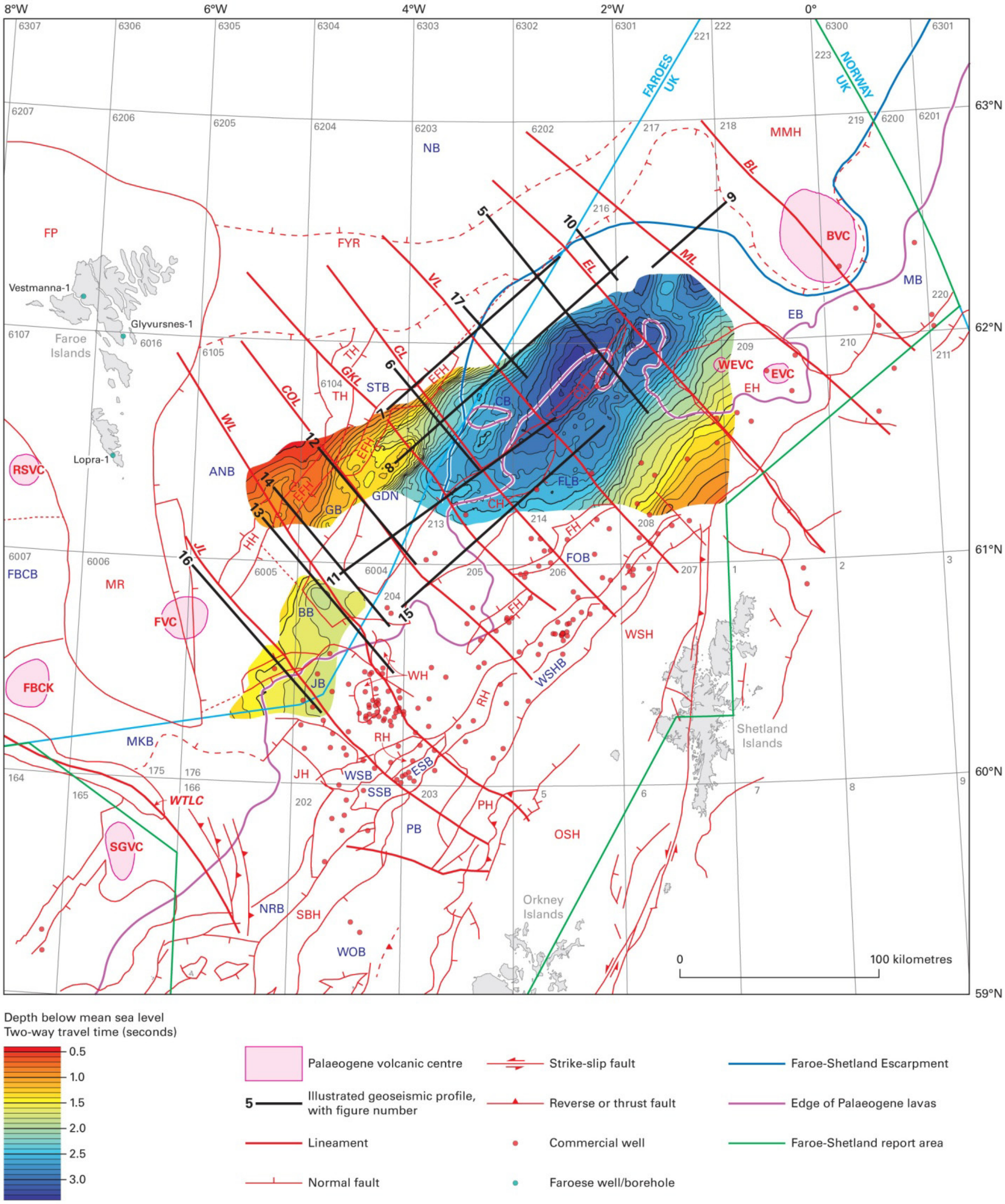

Figure 24 Structure contour map (two-way-travel-time in seconds): Intra-Oligocene Unconformity. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

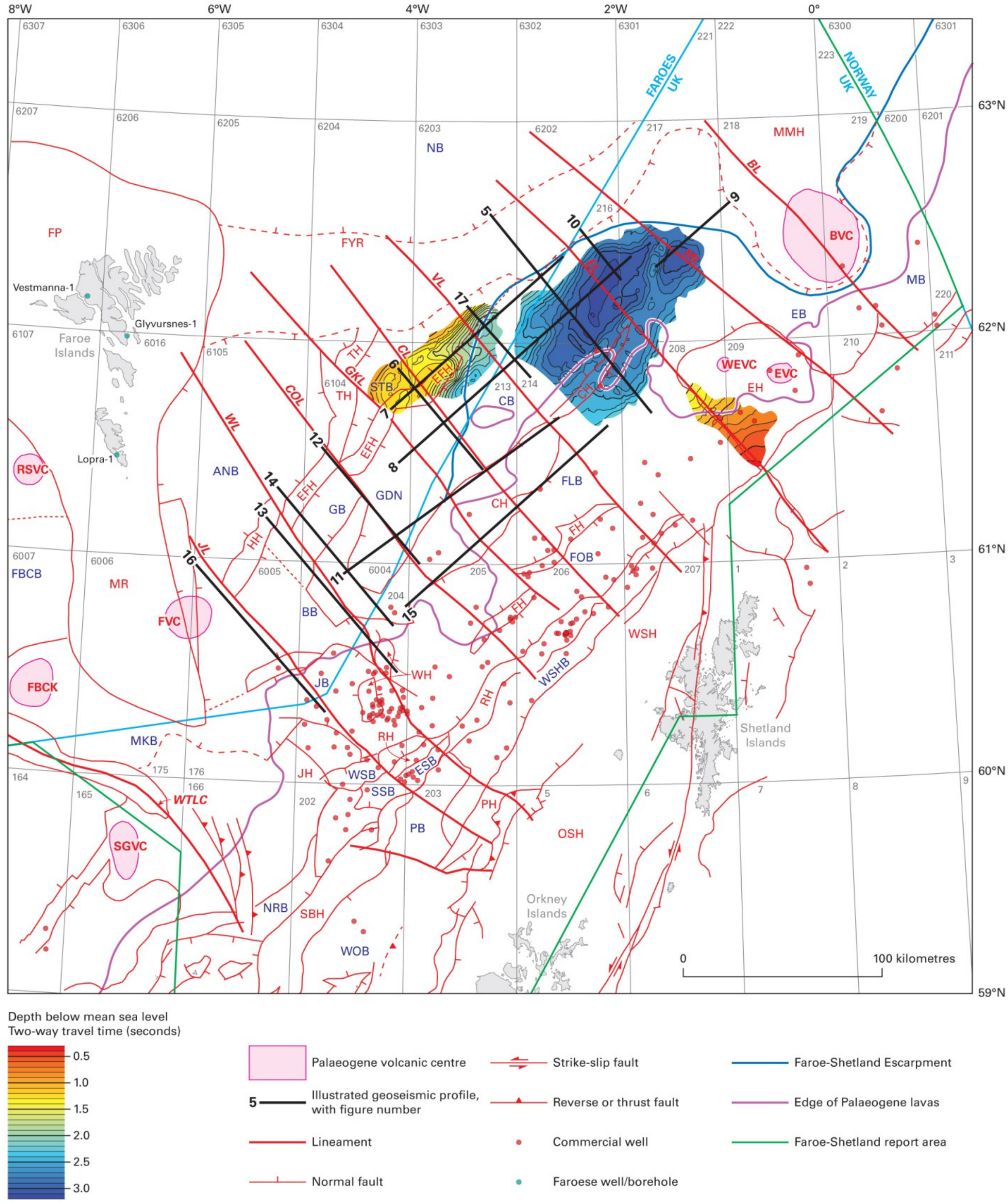

Figure 25 Structure contour map (two-way-travel-time in seconds): Top Palaeogene Unconformity. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

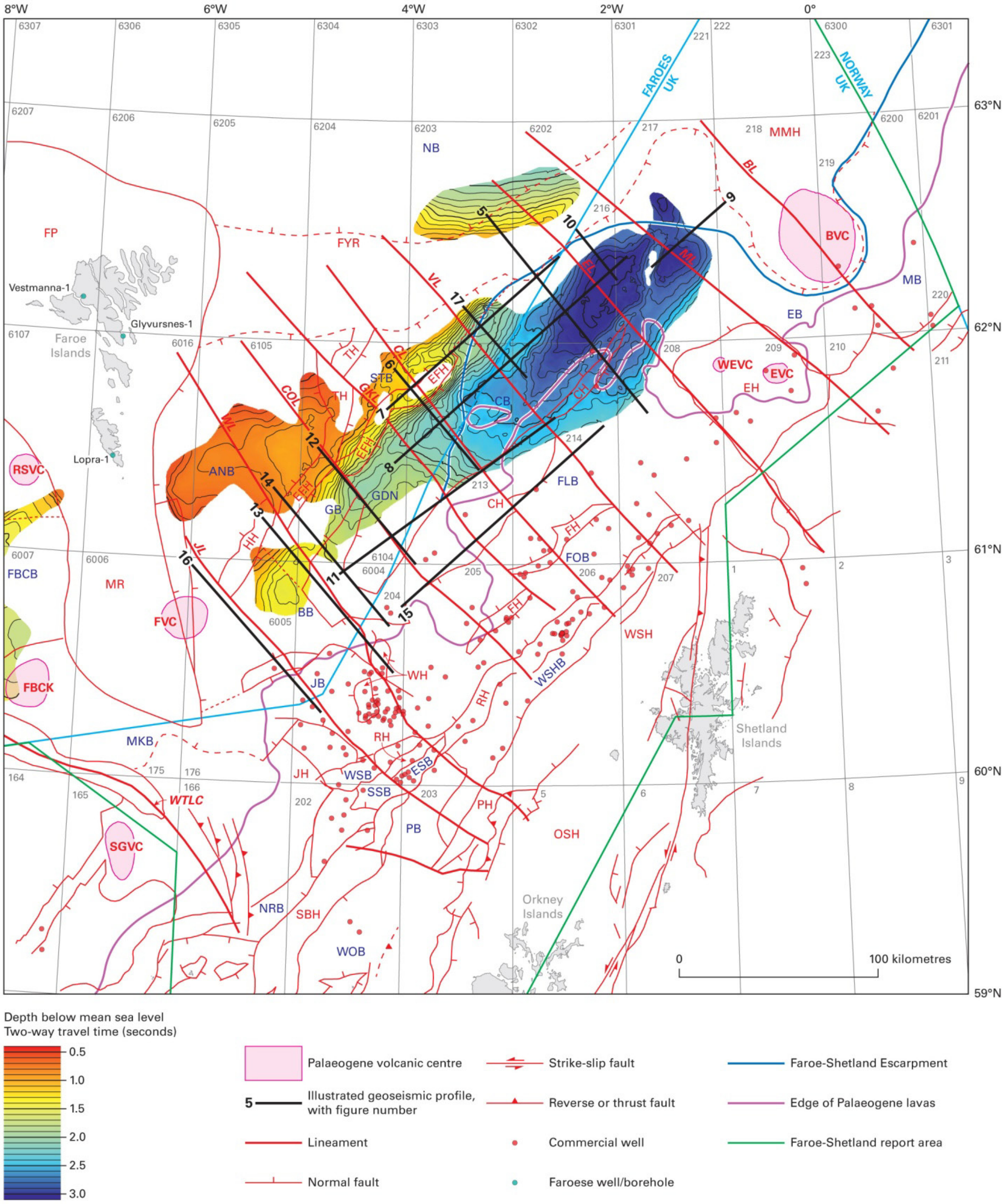

Figure 26 Structure contour map (two-way-travel-time in seconds): Intra-Miocene Unconformity. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

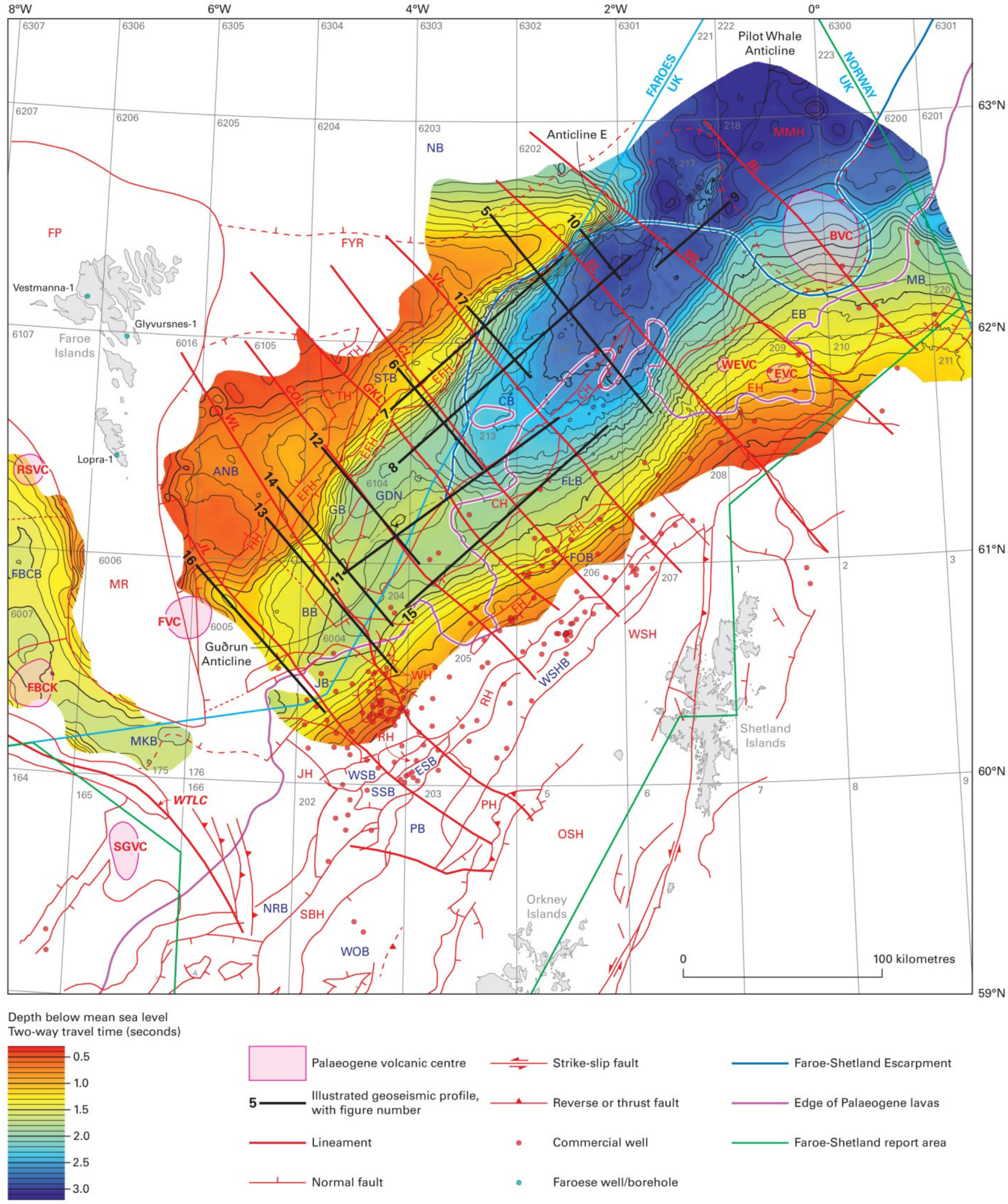

Figure 27 Structure contour map (two-way-travel-time in seconds): Intra-Neogene Unconformity. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

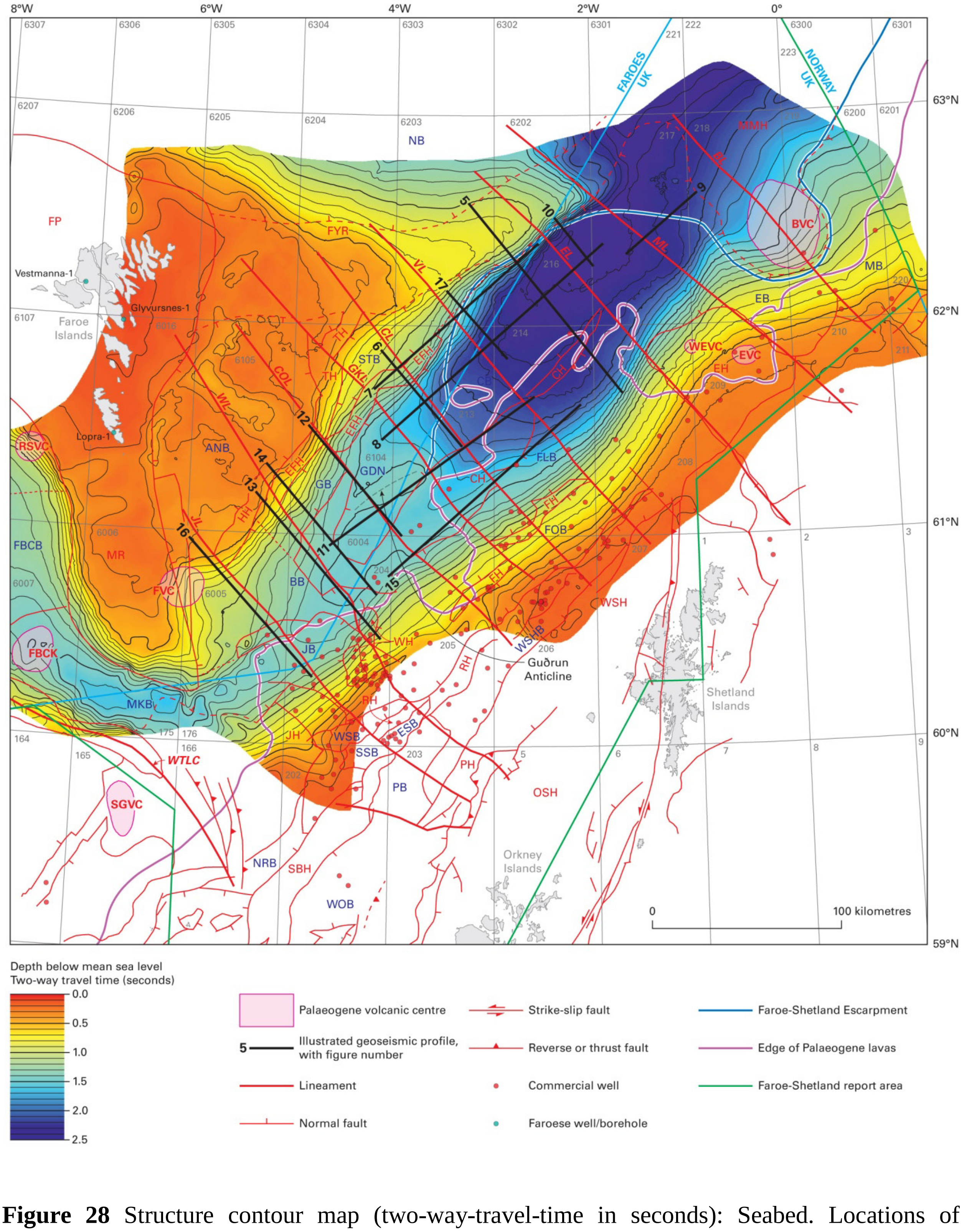

Figure 28 Structure contour map (two-way-travel-time in seconds): Seabed. Locations of structural elements from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

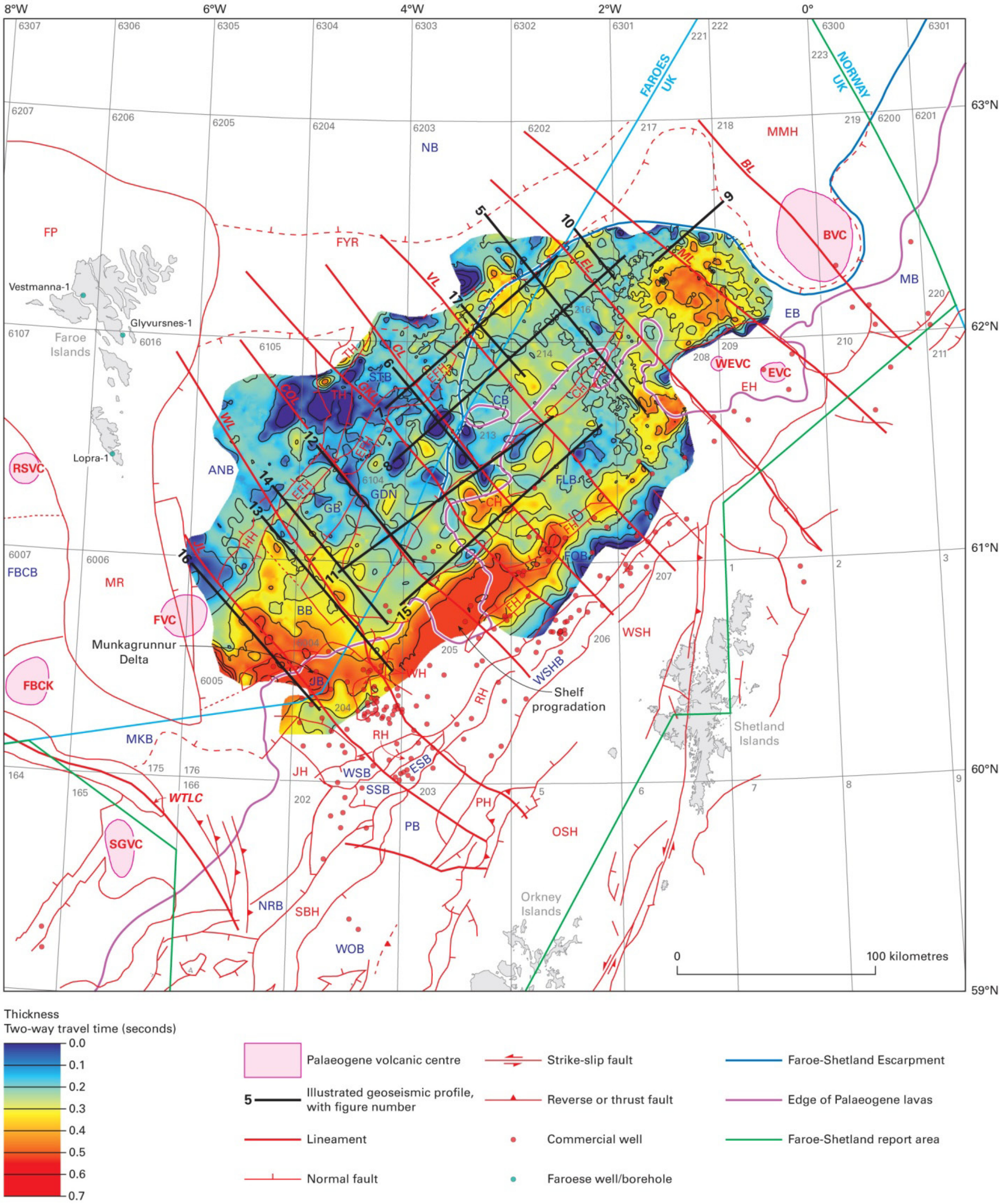

Figure 29 Isochore map (two-way-travel-time in seconds): Top Balder Formation - T2d. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

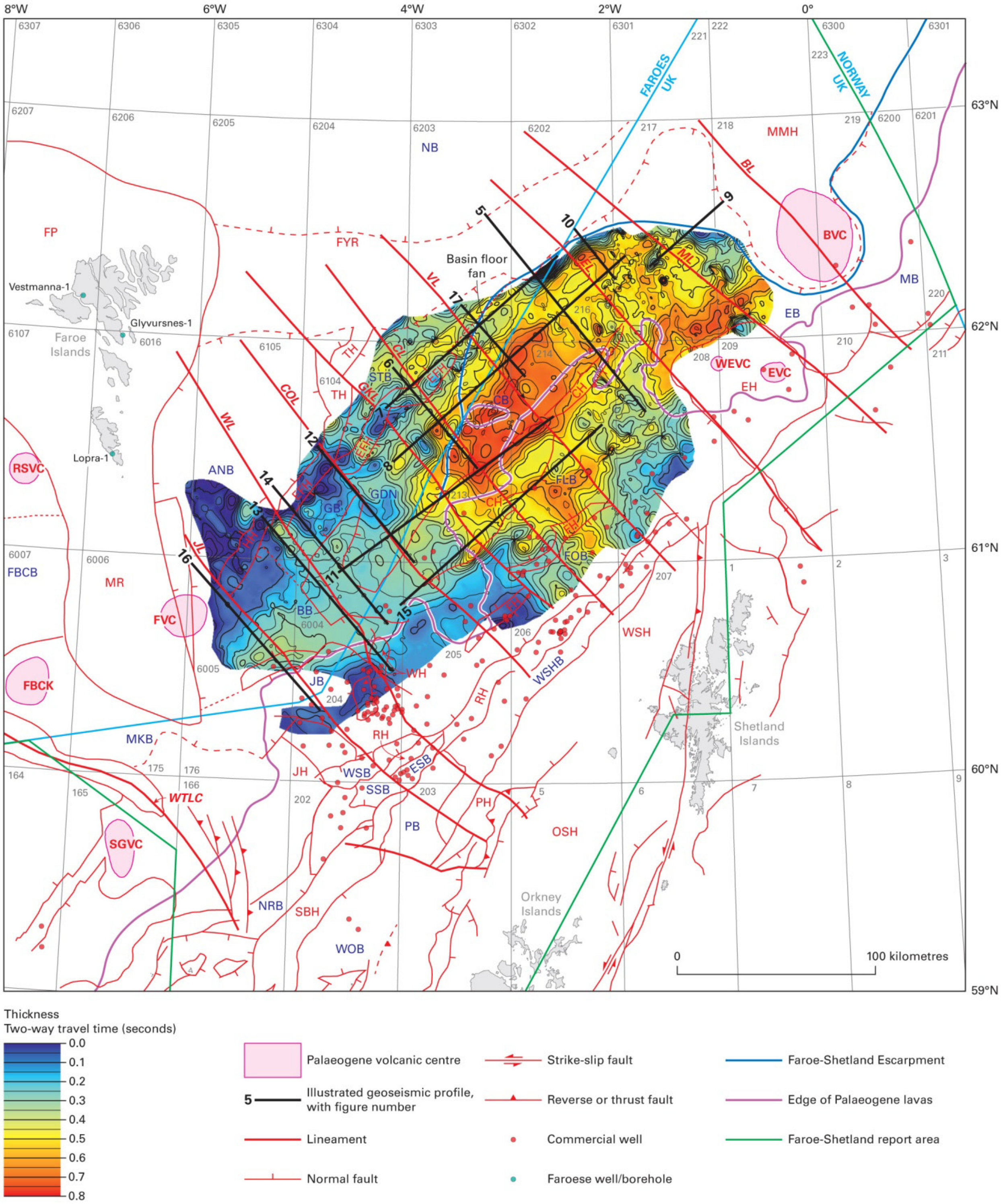

Figure 30 Isochore map (two-way-travel-time in seconds): T2d-T2c. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

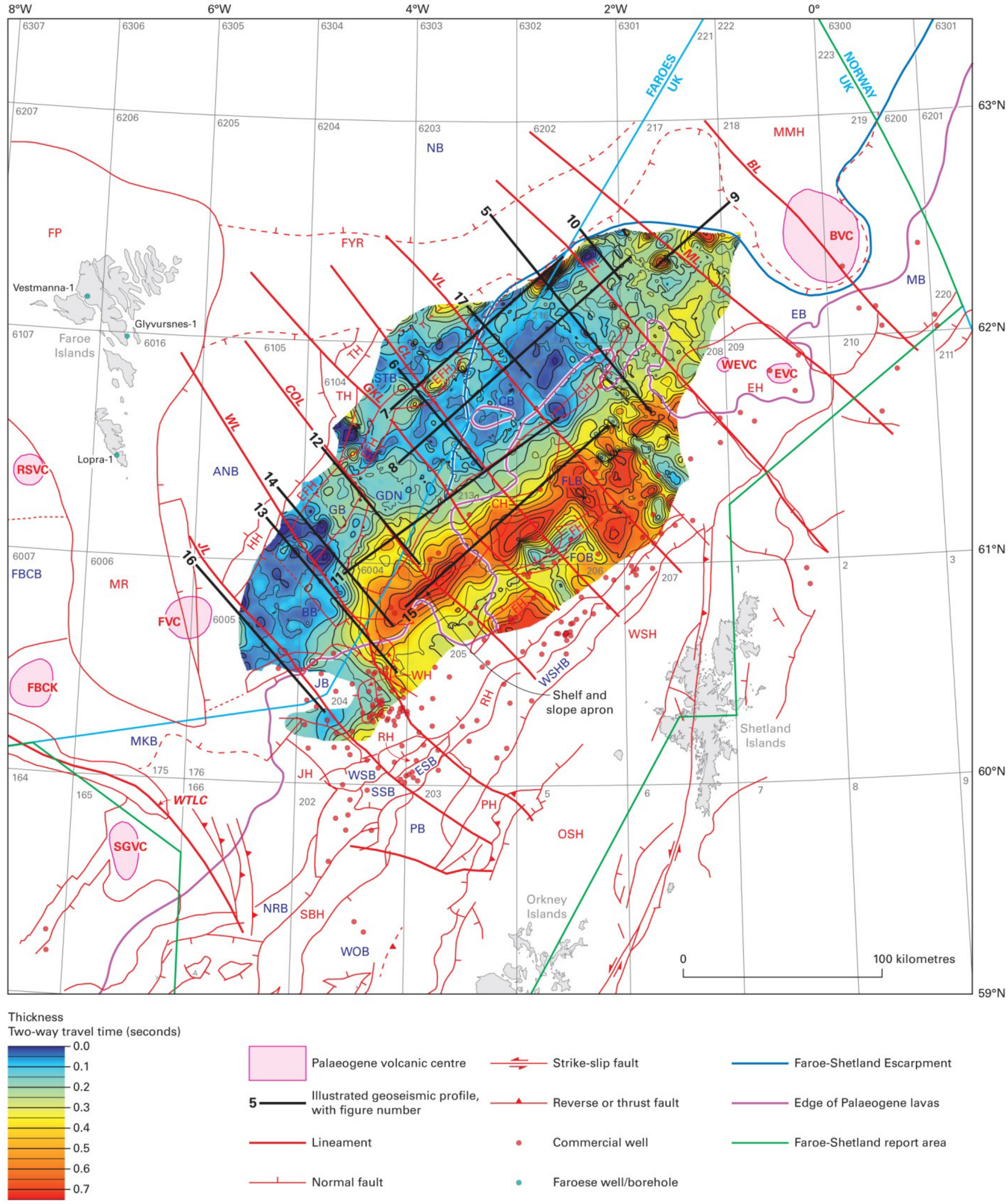

Figure 31 Isochore map (two-way-travel-time in seconds): T2c-T2a. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

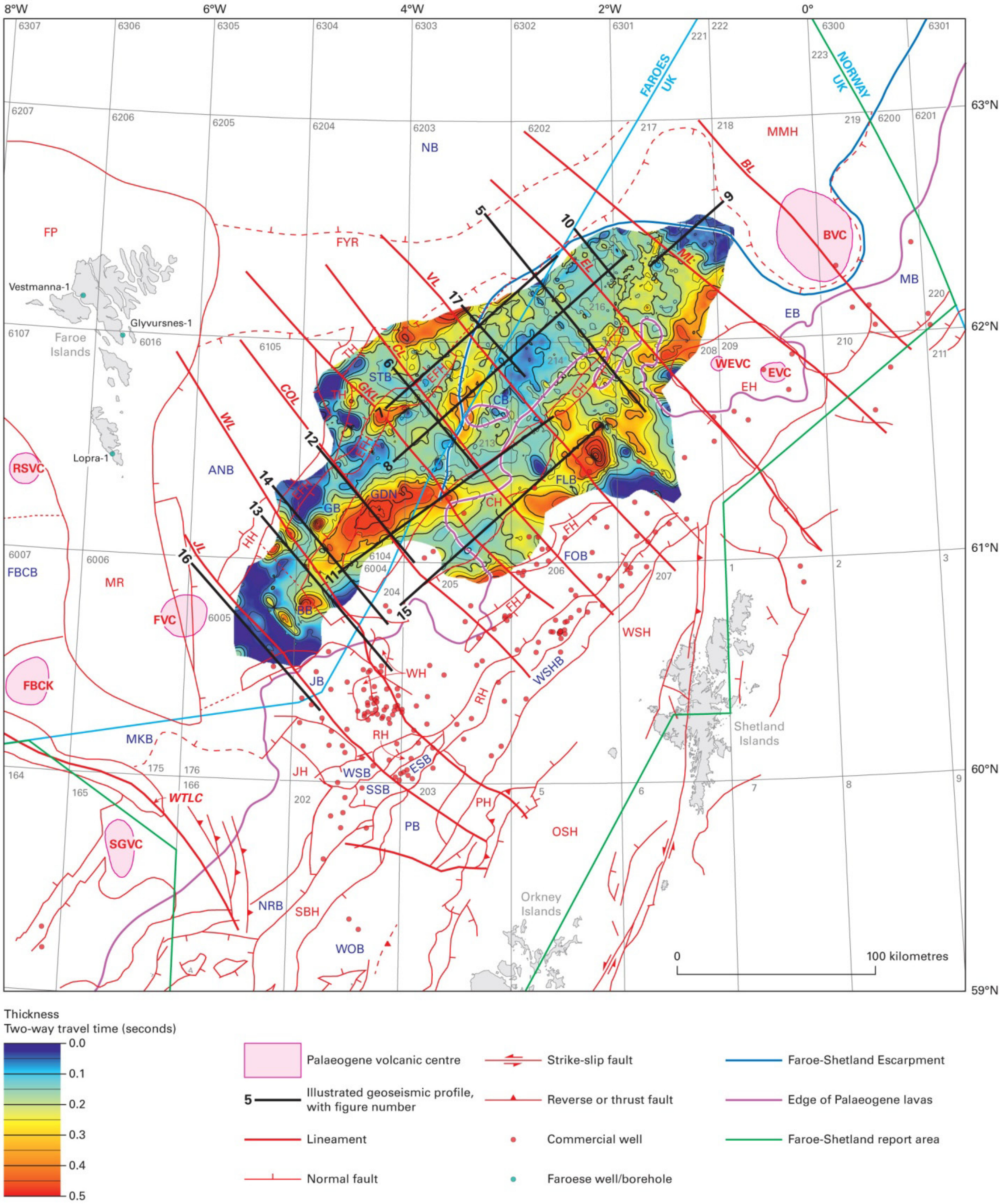

Figure 32 Isochore map (two-way-travel-time in seconds): T2b-T2a. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

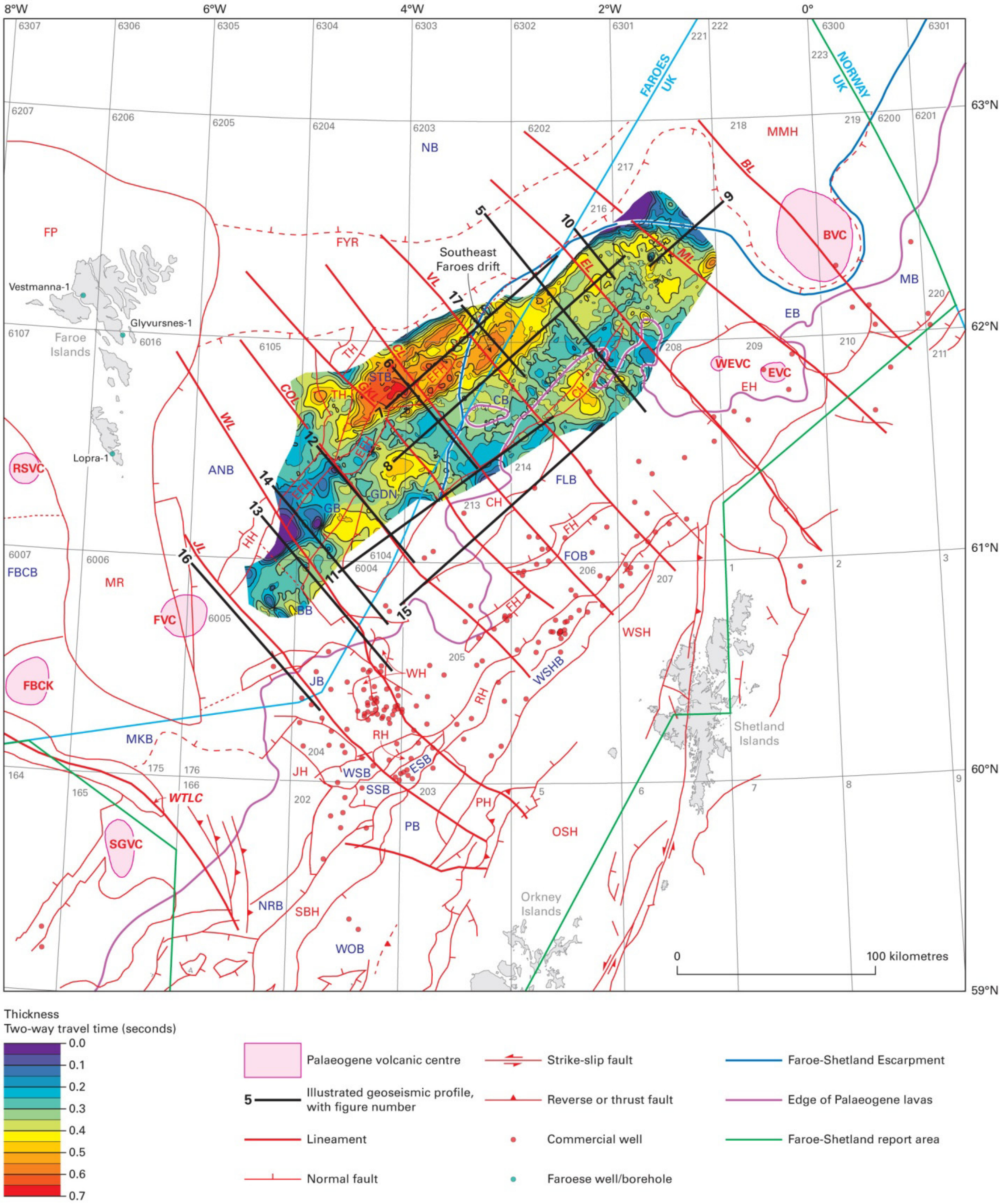

Figure 33 Isochore map (two-way-travel-time in seconds): T2a-IMU. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

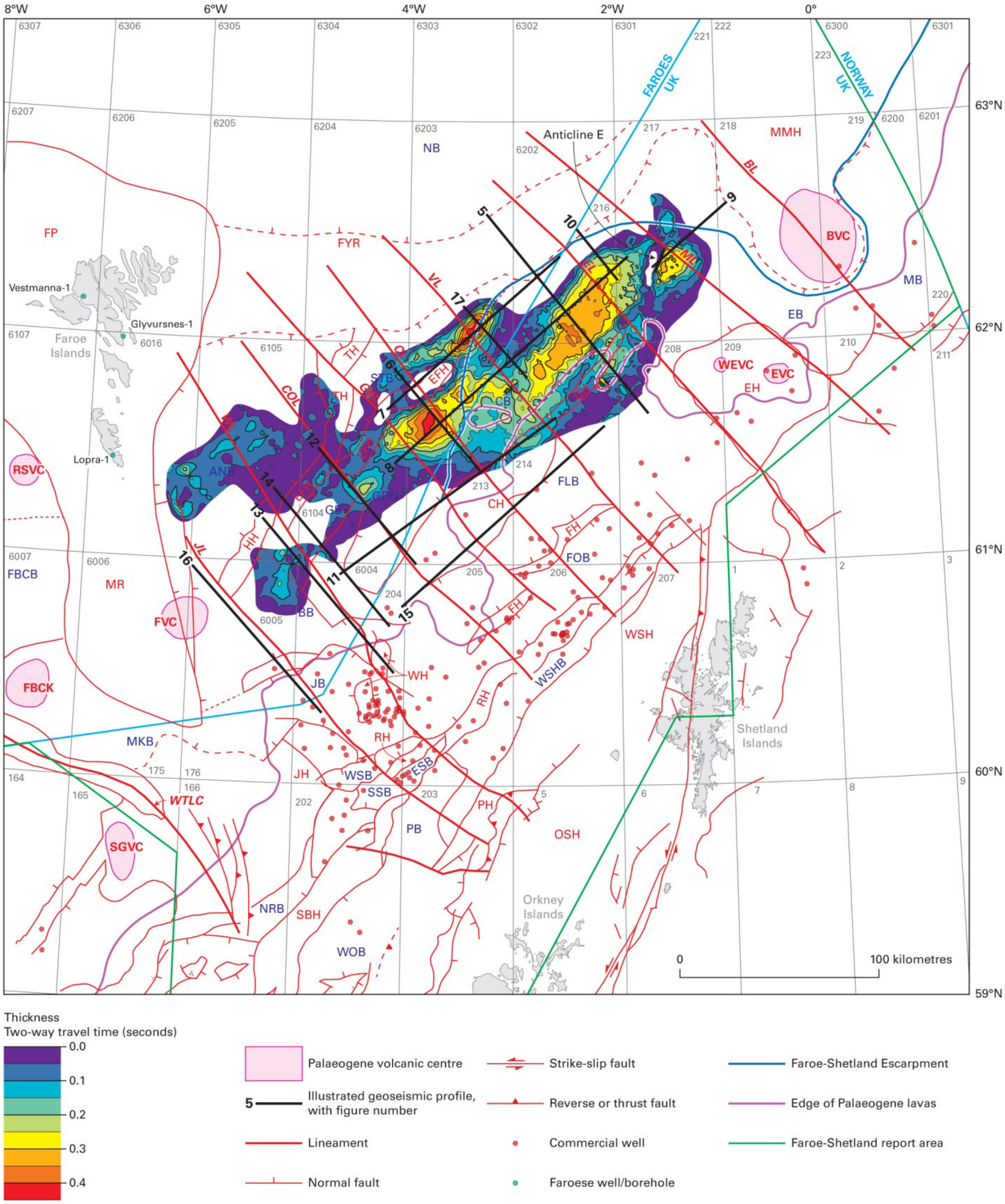

Figure 34 Isochore map (two-way-travel-time in seconds): IMU-INU. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

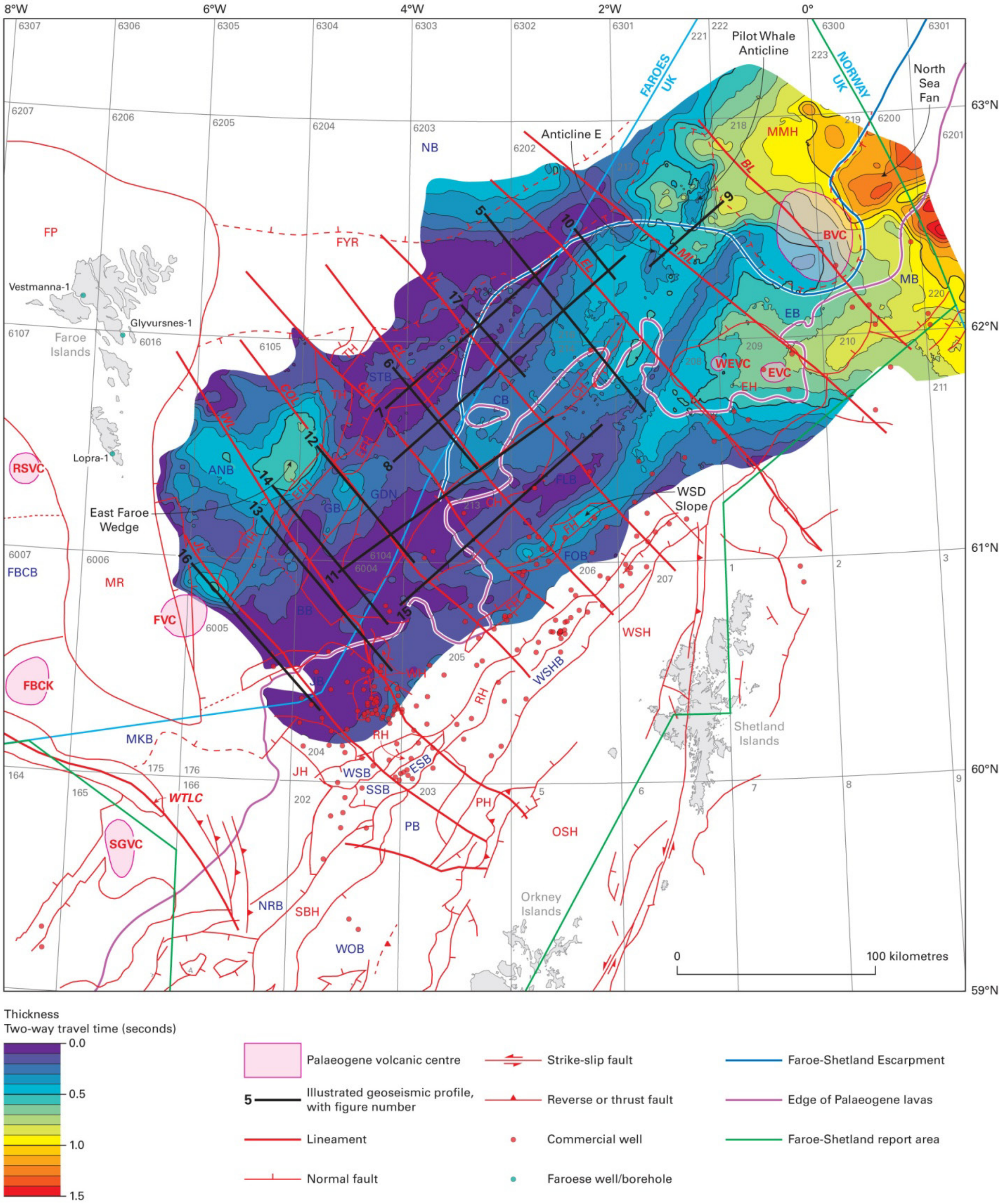

Figure 35 Isochore map (two-way-travel-time in seconds): INU-Seabed. Locations of structural elements and Palaeogene volcanic centres from Ritchie et al. (2011), see Figure 1 for explanation of abbreviations.

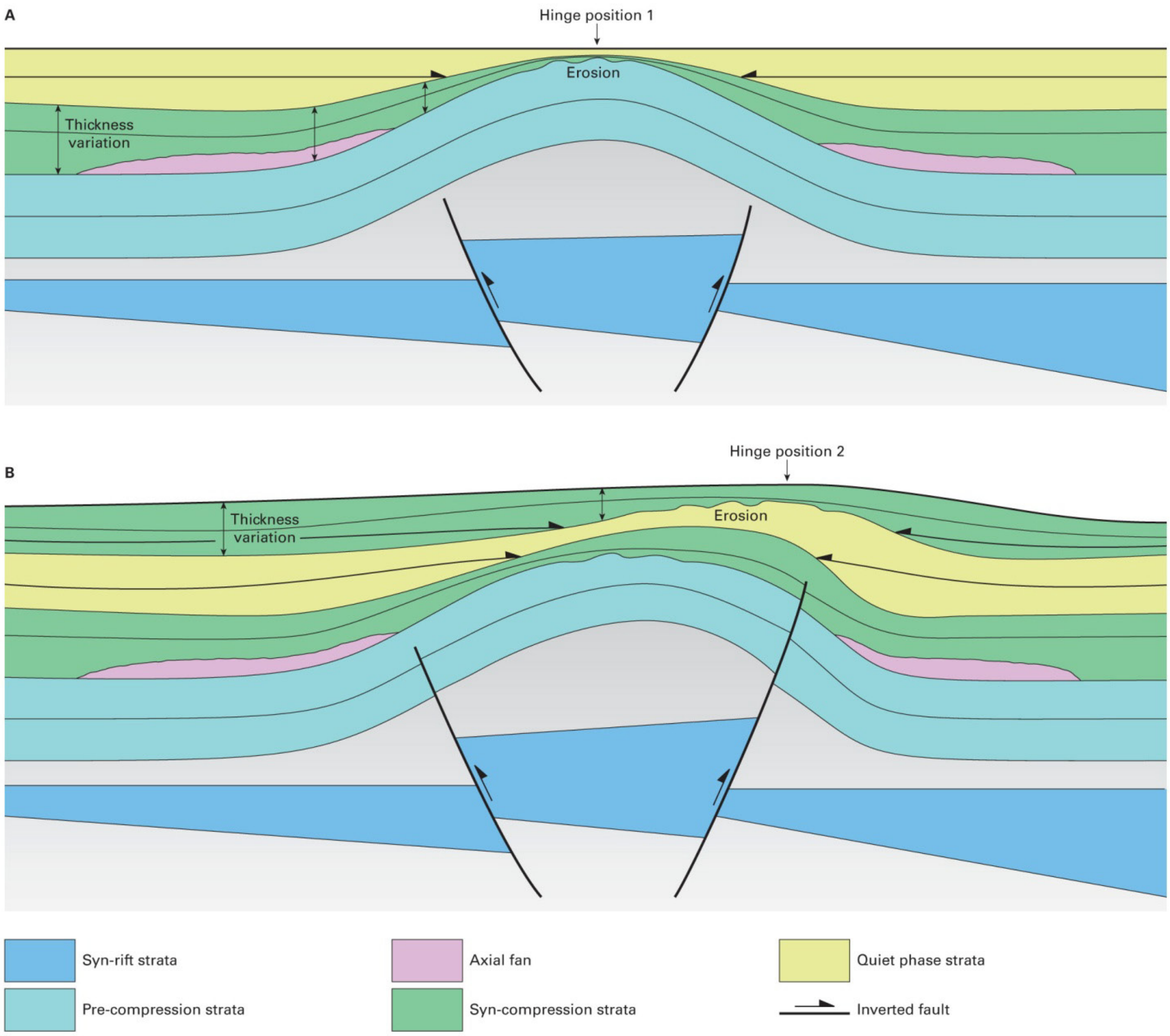

Figure 36 Criteria for identification of syndepositional compression on seismic profiles (modified after Pereira et al., 2011).

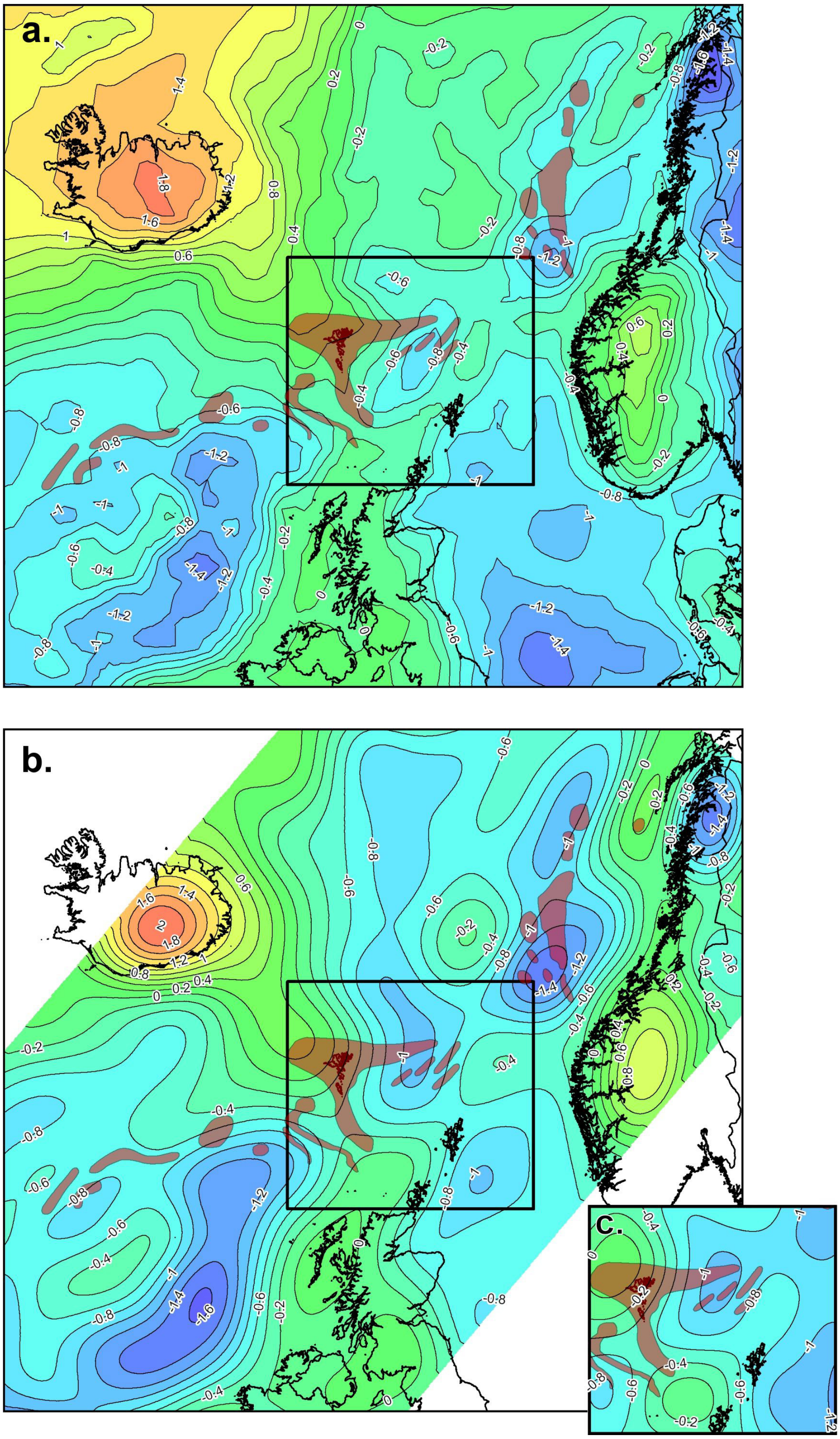

Figure 37 (a) Gravitational potential energy (GPE) in the north Atlantic region derived from geoid height anomalies; contours are at 0.2 x 10^{12} N/m intervals. Based on the Eigen_g104c geoid (Förste et al., 2006) with longer wavelengths removed by a filter tapered between degree and order 7 and 11. (b) GPE calculated from a 3D gravity model of the Faroe-Shetland area. The area covered by (c) is indicated by a black rectangle on (a) and (b). Dome outlines from Døre et al. (2008). See text for details.

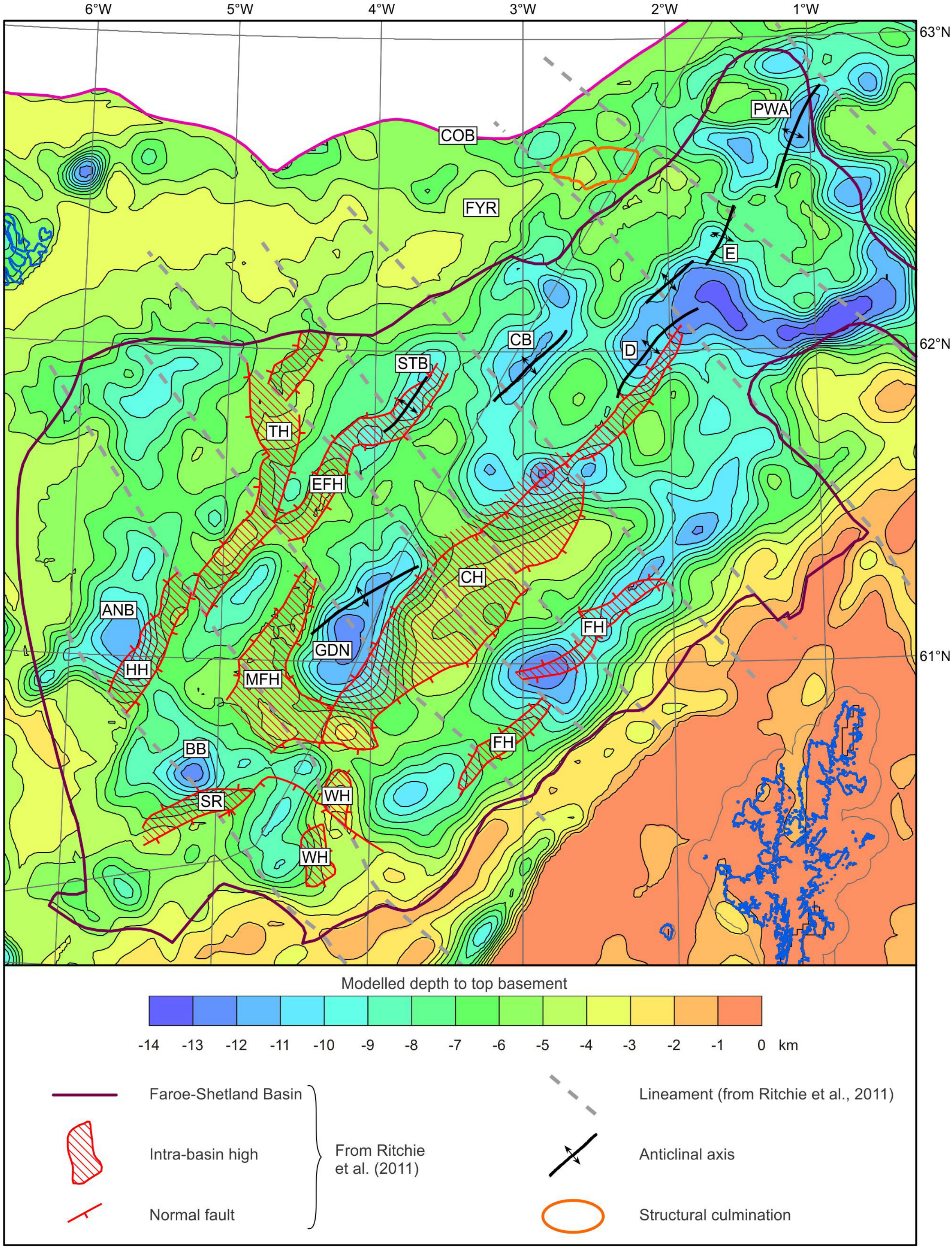

Figure 38 Structural elements in the Faroe-Shetland area (Ritchie et al., 2011) superimposed on the top basement surface from the 3D gravity model (Kimbell et al., 2010). The structural culmination at the east end of the Fugloy Ridge is represented by the 1.5 s two-way-travel-time closed contour from the top Palaeogene lavas structure contour map (Figure 18). Pilot Whale Anticline (PWA), Anticline D and Anticline E are terms applied by Ritchie et al. (2003). Abbreviations: ANB – Annika Sub-basin; BB – Brynhild Sub-basin; CB – Corona Sub-basin; CH – Corona High; COB – Continent-ocean boundary (after Kimbell et al. 2010); EFH – East Faroe High; FH – Flett High; FYR – Fugloy Ridge; GDN – Gudrun Sub-basin; HH – Heri High; MFH – Mid Faroe High; SR – Sjúrður Ridge; STB – Steinvør Sub-basin; TH – Tróndur High; WH – Westray High.

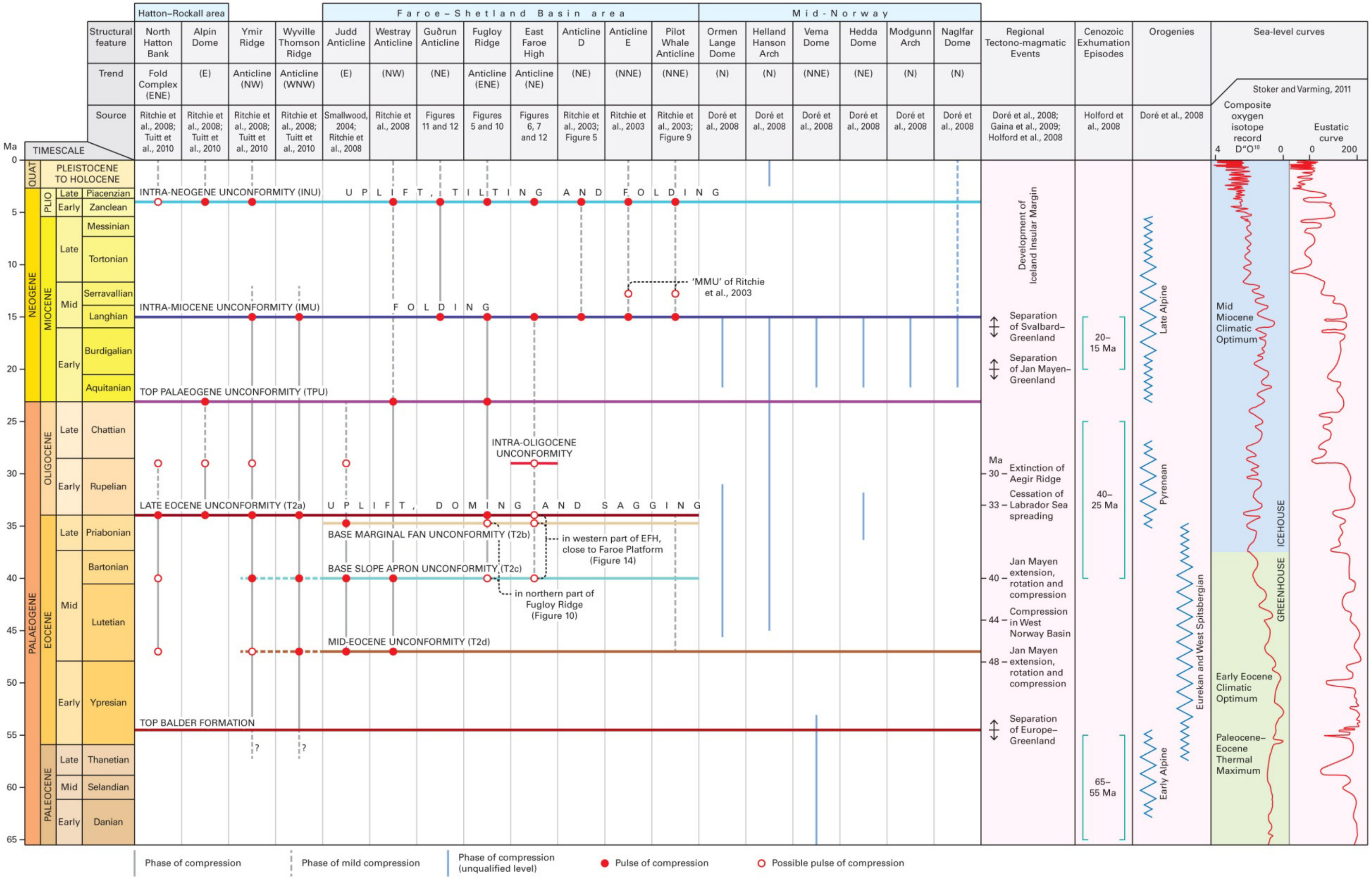

Figure 39 Seismo-stratigraphic chart summarising the timing of post-breakup folding / doming, unconformity development and accelerated subsidence / sagging / tilting and uplift within the Faroe-Shetland Basin area and surrounding NE Atlantic region. The timescale is from Gradstein et al. (2004) and Ogg et al. (2008). See text for discussion regarding uncertainty in the dating of some of the unconformities.

Mapped seismic horizon	**Characteristics of seismic horizon**
Intra-Neogene Unconformity (INU)	This is an angular unconformity typically characterised by truncation below and onlap above (e.g. Figure 6)
Intra-Miocene Unconformity (IMU)	This is an angular unconformity typically characterised by truncation below and strong onlap above the folded horizon (e.g. Figure 6).
Top Palaeogene Unconformity (TPU)	This regional unconformity has been recognised in only a limited part of the Faroe-Shetland region (e.g. Figure 5).
Intra-Oligocene unconformity	This seismic horizon corresponds to an angular unconformity on the East Faroe High; characterised by incision, although no angular break is evident in the adjacent Corona Sub-Basin (e.g. Figure 6).
T2a (Top Eocene)	This seismic horizon, at least locally, corresponds to an angular unconformity (e.g. Davies and Cartwright, 2002), characterised by onlap (e.g. Figures 5 and 10).
T2b (late Priabonian – base marginal fan)	This seismic horizon, at least locally, corresponds to an angular unconformity (e.g. Figure 14)
T2c (intra-Bartonian – base slope apron)	This seismic horizon, at least locally, corresponds to an angular unconformity (e.g. Figure 14)
T2d (intra-Lutetian) (Mid-Eocene Unconformity)	This seismic horizon, at least locally, corresponds to an angular unconformity (e.g. Figure 16)

Table 1 Summary of the characteristics of selected seismic horizons mapped in this study.