

Spatial and temporal evolution of rifting and continental breakup in the Eastern Black Sea Basin revealed by long-offset seismic reflection data.

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Key Points:

- Rifting in the Eastern Black Sea Basin continued up to Oligocene, based on seismo-stratigraphic evidence for ongoing extension in the NW.
- An early-Tertiary breakup was initiated in the SE part of the basin, synchronous with extension in the NW.
- Sedimentary sequences record the along-strike change from syn-kinematic to breakup processes in the Eastern Black Sea Basin.

Abstract

The age and distribution of the syn-rift and early post-rift infill records the spatial and temporal distribution of extension and breakup processes in a rift basin. The Eastern Black Sea Basin (EBSB) is thought to have formed by back-arc extension during Cretaceous to Early Cenozoic time. However, a lack of direct constraints on its deep stratigraphy leaves uncertainties over the time, duration and location for rifting and breakup processes in the basin. Here we use the enhanced imaging provided by 2D long-offset seismic reflection profiles to analyze the deep structural and stratigraphic elements of the EBSB. Based on these elements, we infer the presence of two distinct Late Cretaceous syn-rift units, recording initial extension (rift stage 1) over the continental highs (Shatsky Ridge and the Mid Black Sea High), followed by strain localization along the major basin-bounding faults and rift migration towards the basin axis (rift stage 2). Overlying these units, Palaeocene(?)–Eocene and Oligocene units show a syn-kinematic character in the NW, with evidence for ongoing extension until Oligocene time. Towards the SE, these sequences are instead post-kinematic, directly overlaying a basement emplaced during breakup. We interpret the Palaeocene(?)–Oligocene units to record the time spanning from the initiation of breakup (Late Cretaceous–Palaeocene, in the SE) to the end of extension (Oligocene, in the NW). The first ubiquitously post-rift infill is the Lower Miocene Maykop Formation. Our results highlight the along-strike temporal variability of extension and breakup processes in the EBSB.

1. Introduction

Classical rifting models imply that rifting and continental breakup occur as processes well defined in time and space and recorded by distinct sedimentary sequences (e.g., Driscoll et al., 1995). However, these models are based on the interpretation of isolated transects across margins, so cannot be expected to explain the 3D distribution of structures and sedimentary sequences along a rift system. More recently, the availability of more comprehensive datasets has allowed the documentation of the along-strike evolution at rifted margins, showing that continental rifting and breakup processes are more complex and variably distributed in time and space (e.g., Alves & Cunha, 2018; Muirhead et al., 2016; Nixon et al., 2016; Péron-Pinvidic et al., 2007; Soares et al., 2012).

In the context of intra-continental back-arc rifting, the Eastern Black Sea Basin (EBSB) is one of the most studied and yet most controversial basins in the world (Figure 1). It consists in a narrow NW trending linear trough underlain by oceanic or proto-oceanic crust, bounded by the continental domains of the Mid Black Sea High (MBSH) and the Shatsky Ridge (Figure 1a). The thickness of the sedimentary infill in the EBSB is approximately 8–9 km (Scott et al., 2009), including Cenozoic and upper Mesozoic sediments (Shillington et al., 2008). However, the deep part of the EBSB is yet to be drilled and little is known about the age and distribution of its syn-rift and early post-rift infill. Early insights into the geological evolution of the Black Sea came from regional seismic lines (Belousov et al., 1988; Finetti et al., 1988; Tugolev et al., 1985). Since then, researchers have attempted to unravel the kinematics, mechanisms and timing for the opening of the EBSB (e.g., Finetti et al., 1988; Görür, 1988; Graham et al., 2013; Nikishin et al., 2003; Okay et al., 1994; Rangin et al., 2002; Robinson et al., 1996; Shillington et al., 2008; Stephenson & Shellart, 2010). In addition, crustal-scale industry seismic datasets have been acquired due to the growing interest in the EBSB as a region with high hydrocarbon potential,

related to the presence of a regionally distributed source-rock, the Maykop Formation (e.g., Graham et al., 2013; Robinson et al., 1996; Stovba et al., 2009).

The identification of syn-rift stratigraphic intervals is essential to understand the evolution of rifted margins, as they record the early stages processes in a rift system. However, the determination of these syn-rift sequences is often complicated by a lack of data and of direct constraints (e.g., Péron-Pinvidic et al., 2007). In the EBSB, wells penetrating the entire Cenozoic and part of the Mesozoic sections are located along the margins or onshore, while strata of this age lie at depths ≥ 8 km in the center of the basin. Therefore, stratigraphic sections at well locations are comparatively condensed, and there are uncertainties in tracing major horizons from well locations into the basin center (Shillington et al., 2008). Also, most of the wells have been drilled in the Western Black Sea Basin (WBSB), and tracing horizons from the WBSB to the EBSB is challenging because of the presence of the MBSH, which prevents direct correlation of horizons older than Late Eocene (Banks et al., 1997).

Seismic data have also been used to define the presence and distribution of the syn-rift in the EBSB, but imaging is often limited by attenuation through the thick basin infill (e.g., Finetti et al., 1988; Rangin et al., 2002; Shillington et al., 2008). In fact, in most studies, syn-rift geometries (e.g., fanning reflector geometry toward fault planes) are not clearly observed in seismic data (e.g., Nikishin et al., 2015a; Shillington et al., 2008). Therefore, syn-rift units may be mistaken for pre-rift or post-rift strata because of the lack of obvious growth beds (Withjack et al., 2002). This ambiguity may explain the small thickness of syn-rift deposits identified in the EBSB (e.g., Görür & Tüysüz, 1997; Görür et al., 1993).

Here, we interpret and discuss the deep seismo-stratigraphic framework of the EBSB using long-offset seismic reflection profiles acquired in 2011 by Geology Without Limits (GWL) and ION GXT, which provide enhanced imaging at depth, combined with published litho-stratigraphy and well correlated seismic data. Isopach maps, seismic facies and examples of seismic profiles are presented to illustrate the characteristics, areal distribution and thickness of each unit. The aim of this study is to present the distribution and ages of the syn-rift, early post-rift and post-rift sequences, to analyze their stratigraphic contacts and their interaction with the structural features of the basin. These elements are used to define rift onset, propagation and breakup stages in the EBSB.

Unravelling the rifting and breakup processes controlling EBSB formation has a key role in understanding the geodynamic evolution of this area and of similar rift systems. Furthermore, this study provides fundamental insights into (i) how deformation is accommodated in space and time, and (ii) how stratigraphic and structural elements track strain distribution and localization during the process of continental extension.

2. Geological Framework

2.1. Geological evolution of the EBSB region

The EBSB is a deep sedimentary basin lying between the Eastern Pontides and the Great Caucasus mountain ranges (Figure 1a). Rifting is the first-order geodynamic process responsible for the tectonic evolution of the EBSB and is related to the northwards subduction of the Neotethys Ocean below the Eurasian Plate (Finetti et al., 1988; Görür, 1988; Nikishin et al., 2003; Okay et al., 1994; Robinson et al., 1996; Stephenson & Schellart, 2010; Zonenshain & Pichon, 1986). The basin opened in a back-arc setting behind the Pontides magmatic arc (Finetti et al., 1988; Görür, 1988; Okay et al., 1994; Zonenshain & Pichon, 1986). Back-arc extension was generated by the counter-clockwise rotation of the Shatsky Ridge away from the MBSH, as evidenced by paleo-magnetic data (Westphal et al., 1986). Compressional forces and tectonic inversion, due to the northward movement of the Arabian Plate, started along the Pontides in the Late Cretaceous, reaching the Greater Caucasus possibly in the Palaeocene and becoming regionally widespread during the Late Eocene and Oligocene (Espurt et al., 2014; Hippolyte et al., 2017; Munteanu et al., 2011; Robinson et al., 1996; Vincent et al., 2016). The syn-collisional (orogenic) stage is generally considered to have started in the Oligocene, forming the Caucasus mountain range and causing the uplift of some areas within tectonic units as well as the reactivation of older fault systems and the formation of inverted structures (Stovba et al., 2009). Compressive deformation is concentrated around the edges of the EBSB, as shown by seismic reflection data, earthquake locations, and subsidence analysis (e.g., Edwards et al., 2009; Shillington et al., 2008).

2.2. Available data on the basin stratigraphy

Detailed analyses of the thicknesses and distribution of the sedimentary deposits in the EBSB are based on the most updated interpretation of seismic data (e.g., Kazmin et al., 2000; Nikishin et al., 2001, 2003, 2015a; Scott et al., 2009; Sydorenko et al., 2017) and estimates on the age and lithology of the sedimentary units generally come from seismic stratigraphic horizons tied to well control at the edges of the basin and to onshore outcrops (e.g., Greater and Lesser Caucasus, and Crimean Mountains - Nikishin et al., 2003, 2015a, 2015b; Okay et al., 2013; Robinson et al., 1996; Saintot et al., 2006; Stephenson & Schellart, 2010; Stovba et al., 2009). More reliable information comes from dredged sequences from the Archangelsky Ridge (Rudat & Macgregor, 1993) and from a complete Mesozoic to Lower Palaeocene stratigraphy in wells drilled over the Shatsky Rise in Georgia (Robinson et al., 1996). These stratigraphic interpretations have fed into paleo-geographic reconstructions of the mechanisms and timing of basin opening (e.g., Egan & Meredith, 2007; Nikishin et al., 2015b; Okay et al., 1994; Robinson et al., 1996; Stephenson & Schellart, 2010).

Extensive deep-water drilling has occurred in the WBSB, whereas only four deep wells have been drilled in the EBSB to date (e.g., Stovba et al., 2009; Sydorenko et al., 2017; Tari & Simmons, 2018) (Figure 1b). Subbotina-403 reaches 4300 m below seabed and terminates in the Lower Eocene succession (Stovba et al., 2009; Sydorenko et al., 2017) (Figure 1b). Sinop-1, 5531 m deep, was drilled on a structural high of Andrusov Ridge, targeting a syn-rift of Aptian to Albian age (e.g., Tari & Simmons, 2018) (Figure 1b). The sediment cover above the MBSH and the Shatsky Ridge has been described by using extensive academic and industry seismic

reflection data (e.g., Belousov et al., 1988; Nikishin et al., 2015a, 2015b; Rangin et al., 2002; Robinson et al., 1996; Tari et al., 2018), crustal scale wide-angle seismic data (e.g., Shillington et al., 2017; Yegorova et al., 2010) and seafloor dredging (Rudat & MacGregor, 1993). Chronostratigraphic charts have been compiled to illustrate the different ages and litho-stratigraphy elements of the Black Sea region and EBSB (Adamia et al., 2017; Shillington et al., 2008; Sydorenko et al., 2017; Tari & Simmons, 2018).

Figure 2 and Table 1 summarize previous attempts to estimate the timing of rifting in the EBSB. Note that the definition of rift duration changes from author to author: it can be either the time from rift initiation to breakup (e.g., Nikishin et al., 2015a), or the time from rift initiation to the end of oceanic spreading (e.g., Okay et al., 1994). Some studies assume that the EBSB open at the same time as the WBSB, and provide age estimates which are mostly based on data coming from the WBSB region (e.g., Hsü et al., 1977; Tüysüz, 1999). Based on different data, rifting has been considered to be Jurassic (e.g., Zonenshain & Pichon, 1986), Cretaceous (e.g., Adamia et al., 1974; Görür, 1988; Nikishin et al., 2003; Okay et al., 1994), Early Cretaceous to Palaeocene (e.g., Finetti et al., 1988; Sydorenko et al., 2017), Palaeocene to Early Eocene (e.g., Banks et al., 1997; Robinson et al., 1995b), or Eocene (e.g., Kazmin et al., 2000; Yilmaz et al., 2000) in age. From the analysis of the Crimean Mountains onshore outcrops, and from micro-paleontological dating (nanno-plankton) in the same region, studies infer similar timing for the EBSB rift (starting in Early Cretaceous), breakup (Albian-Cenomanian) and oceanic spreading (Cenomanian to Santonian/Campanian) (e.g., Hippolyte et al., 2018; Nikishin et al., 2003; Sheremet et al., 2016). Conversely, studies based on Georgia and Eastern Pontides onshore outcrops generally infer a Late Cretaceous-Early Cenozoic basin opening (e.g., Okay et al., 1994; Yilmaz et al., 2000). Similarly, studies integrating wells data from the Shatsky Ridge and dredging from the Archangelsky Ridge, infer a Palaeocene-Eocene age for rifting (e.g., Banks et al., 1997; Rudat & Macgregor, 1993; Spadini et al., 1996). Depending on the location of the sampling, there are different outcomes in terms of timing for rifting and breakup in the EBSB.

3. Data and Methods

3.1. Seismic Data

We used twelve post-stack time-migrated long-offset seismic reflection profiles, acquired in 2011 by Geology Without Limits (GWL) and ION GXT (Figure 1b). Seismic profiles were acquired using an airgun array with total volume of 5680 cu. in, 816 channels, a 10.200 m long streamer and an 18-second (TWTT) record length. Processing was carried out by GWL and consisted in editing, low-cut filtering, spherical divergence correction, multiple suppression (predictive deconvolution and radon de-multiple), amplitude balancing, normal move-out correction, muting and stacking, post-stack Kirchhoff time migration and post-stack/post-migration band-pass filtering, coherent filtering and amplitude balancing. Long-offset profiles are integrated with six time-migrated multichannel seismic reflection profiles, acquired by BP before 2005 (Figure 1b). These profiles were acquired with streamer lengths of 3600 to 6000 m, allowing 8 to 15-second (TWTT) record length, and they were used to constrain the shallower seismo-stratigraphic interpretation. We converted the time migrated profiles to depth using interval velocities derived from the stacking velocities provided by GWL, combined with velocities from wide-angle seismic data (Shillington et al., 2009, 2017). A comparison between the GWL velocities and those from wide-angle seismic profiles, which have a sufficiently large aperture to constrain the velocities of deep sediments, showed that the two velocity functions are very similar in the shallow section (above basement reflection) whilst wide-angle seismic velocities are faster in the deeper section. In order to provide the most accurate velocity model for time-depth conversion, we combined these two velocity fields. Further information on the time-depth conversion steps is given in Supporting Information.

3.2. Seismic Interpretation: Method and Terminology

Regionally distributed prominent reflectors, characterized by discordant stratigraphic contacts and variations in seismic character, extend across the EBSB. These reflectors were used to identify the top and base of separate sedimentary sequences (S1-S7). We used the internal geometries and seismic facies variation of these sequences (Figures 3-7), as well as their interaction with structural elements across the basin, to identify their pre-, syn- or post-rift character (e.g., Franke, 2013; Williams, 1993). Seismic horizons interpreted in depth were gridded using a Convergent Interpolation algorithm (e.g., Haecker, 1992) and a cell size of 250x250m, to generate horizon structural surfaces. Thickness maps for each sedimentary sequence (isopach maps) were calculated from the difference between the horizon surfaces representing top and base of that sequence. Isopach maps are presented to illustrate the depo-center distribution and their spatial and temporal migration (Figure 8).

Normal fault systems were mapped and interpolated across seismic profiles, where possible. The resulting fault polygons were overlain on isopach maps to help to define fault-related depo-center development (Figure 8). Fault planes have been picked from discontinuities in reflections, from residual diffractions and from sedimentary infill geometries.

In this paper we classify the identified unconformities based on the stratigraphic models of Bond et al. (1995), Falvey (1974) and Franke (2013), which recognise two key unconformities that develop during rifting and continental breakup: the *rift-onset unconformity*, which separates pre-rift and syn-rift strata, and the *breakup unconformity*,

which separates syn-rift from post-rift strata at the time of crustal breakup and onset of oceanic crust accretion (e.g., Falvey, 1974; Withjack et al., 1998). The *breakup unconformity* is defined as an erosional surface characterizing the uplifted margins during rifting stage, often truncating wedge-shaped syn-rift sediments and separating them from younger post-rift sequences that show little/no evidence of extension controlling their deposition (Franke, 2013). Basinward, this unconformity merges with an angular unconformity and terminates onto the top of the igneous crust (Franke, 2013). However, a clear definition of this unconformity as a single stratigraphic surface, together with the identification of distinct syn-rift and post-rift stratigraphic sequences, is complicated by the fact that rifting and continental breakup processes are variably distributed in time and space (e.g., Péron-Pinvidic et al., 2007). In this study we define the presence of a *breakup unconformity* of regional extent, which is the principal stratigraphic feature related to continental breakup along the continental highs. Downslope and in the central basin this unconformity becomes a stratigraphic contact between distinct sedimentary sequences and/or the contact between the top of the inferred oceanic crust and the overlying sequence. Due to the complex definition of rift and breakup processes, previous studies introduced the idea of a *transition sequence* recording the gradual transition from syn-rift to post-rift stage (e.g., Moore, 1992). This idea has been further developed by more recent studies, acknowledging breakup as a prolonged event identified by a discernible unconformity-bounded stratigraphic sequence of regional extent, showing distinct depositional architectures compared to older (syn-rift) strata and younger (post-rift) units: the *breakup sequence* (e.g., Alves & Cunha, 2018; Soares et al., 2012). Here, we refer to a recent definition of *breakup sequence* by Soares et al. (2012), which describes the character of some sedimentary units deposited from the onset of lithospheric breakup to the establishment of thermal relaxation as the main process controlling the subsidence of the basin.

4. Seismic Data Analysis

4.1. Structural elements

The main rift-related structural elements of the EBSB are illustrated in profile bs-90 (Figure 4a). To the NE, the Shatsky Ridge is a basement high bounded to the south by one or more large normal faults (Shatsky Ridge fault), with a northern flank that is flexed beneath the Greater Caucasus, generating a small foreland basin, the Tuapse Trough (Figure 4a). On the western and south-western side of the basin are the Archangelsky and Andrusov Ridges (MBSH). Various evidence for extensional tectonics is visible, with normal faults generally showing a NW-SE to WNW-ESE trending direction and bounding graben and half-graben structures (Figures 4a, 4b, 5a-c, 6a, 6b, 8). Basinward, normal faults arrays are organized in closely spaced domino faults (Figure 5c). All the identified faults cut the basement (pre-rift) and the deep sedimentary sequences (S1 and S2). The main fault systems, interpolated across the seismic profiles, are the basin-bounding faults also representing the flanks of Shatsky Ridge and MBSH (Figures 4a, 4b, 8). The sedimentary sequences immediately overlying faulted basement blocks allow us to estimate the age and duration of fault activity. The faults affecting the basement highs generally show a variable but mostly high dip angle (30-70°) (Figure 5a), whereas basinward faults arrays have quite a uniform and low dip angle (15-25°) (Figure 5c). The velocity model used for seismic depth conversion affects the dip angle of the interpreted faults. Below basement, time-depth conversion using wide-angle velocities results in faults with dip angles up to 4-5° steeper than those from using stacking velocities. This difference in fault dip angle is similar to the 1-2° dip angle uncertainty from the standard deviation in wide-angle velocities (see Supporting Information). Another change from the structural highs to the central basin, is represented by the different stratigraphic contacts with the top of the faulted basement blocks. Over the ridges, S1 is the dominant infill, whilst S2 is the prevalent infill within basinward half-graben structures (Figures 4a, 5a, 5c, 6a, 6b). In the central and SE part of the basin, there is no evidence for normal faults (Figures 4a, 4b). The top basement is a corrugated surface becoming progressively smoother to the SE (Figures 7a, 7b). In this area, faults affecting the basement generate basement scarps that have been described in previous studies as the result of interpreted ENE-WS trending trans-tensional faulting (Shillington et al., 2009).

4.2. Seismo-stratigraphic elements

We mapped the acoustic basement and seven sedimentary sequences across the EBSB, based on their seismic facies, internal geometries and interaction with structural elements (Figure 3). Some of the deepest and thus oldest, fault-bounded sequences are discontinuous throughout the basin. This lack of continuity, together with the presence of gaps between seismic profiles, prevents a clear correlation of these sequences at a regional scale. This point has to be considered when looking at isopachs for these sequences (especially S1 and S2), particularly in areas where seismic coverage is low. With this point in mind, we describe the interpreted sequences below, starting from the oldest, S1 (deepest), and moving to the youngest, S7 (shallowest):

S1 is a low amplitude seismic unit with isolated higher-amplitude reflections (Figure 3). Despite the general lack of coherency of its seismic facies, it can be present as a well layered infill of graben and half-graben structure (Figures 3, 4a, 5a). The S1 isopach map shows that S1 is present over the tops of the MBSH and the Shatsky Ridge, where it is 250 to 1500 m thick (Figure 8a). It then rapidly thins downslope, to disappear towards the central basin (Figures 4a, 4b, 5c). S1 is bounded by two unconformities (horizons U1 and U2), marking the discordant contact at its base and top with the underlying basement and the

overlying S2 sequence. Horizon U1 is, in some places, a very high reflectivity boundary (Figure 5a). In contrast, horizon U2 is a low amplitude reflection (Figure 5a). S1 is a tectonically controlled sedimentary infill, with strata thickness changes and wedged geometries bounded by normal fault planes (Figures 4a, 5a).

S2 is a medium to low amplitude sequence, showing layered packages organized into sedimentary wedges bounded by normal faults (Figures 3, 4a, 5a, 5c). Along the main basin-bounding faults, S2 is sometimes a chaotic and semi-transparent package resembling a detrital material deposited downslope and subsequently overlapped by chaotic to well layered younger sedimentary sequences (S3-S4) (Figures 4a, 5b, 6a, 6b). S2 is up to ~ 1500 m thick, and is often confined within small depo-centers (Figure 8b). S2 occurs as secondary infill above S1, on the top of Shatsky Ridge and the MBSH, whilst it is a primary infill of basinward half-graben structures (Figures 4a, 5c). Within these half-graben structures, S1 may be still present below S2, although not properly imaged. The top of S2 is defined by unconformity U3 (Figures 4a, 4b, 5a, 5c, 6a, 6b). The progressive basinward thinning of S1 and S2 causes the related unconformities U1 and U2 to merge in the basin center, and here only the top U2 horizon is interpreted (Figure 5c). Further basinward thinning of S2 causes the coalescence of horizons U2 with U3, leaving horizon U3 as the contact between S3 and the top basement (Figures 5c, 7a, 7b). The S2 isopach map shows the uneven distribution of this unit through the basin (Figure 8b). Compared to S1, which is mostly interpreted over the Shatsky Ridge, the MBSH and the Tuapse and Sorokin Troughs, S2 exists further basinward, in the NW (Figure 8b). In the SE, S2 is interpreted along the basin-bounding faults, but it is absent in the central basin (Figures 4b, 8b).

S3 is generally a well layered sequence, with low amplitude reflections at its base and higher amplitudes towards its top (Figure 3). In the NW, down the Shatsky Ridge slope, this sequence has the character of a chaotic detrital deposit (Figures 5b, 6a, 6b). Seismic profiles and isopach map show that this unit is preferentially distributed in the central basin (with maximum thickness up to 1200 m), whilst it is < 500 m thick over the MBSH and the Shatsky Ridge (Figure 8c). Thickness changes are partly a response to the underlying basement topography (Figures 4a, 7a), affecting the distribution of S3 in the central basin, but also a response to tectonic activity visible in the NW part of the basin and along the main basin-bounding faults (Figures 4a, 5c, 6a, 6b). In these areas normal faults cross-cut S3, indicating a syn-kinematic deposition (Figures 5c, 6a, 6b).

S4 is characterized by sub-parallel, high to low amplitude reflections (Figure 3). The top and base of this sequence are identified by high acoustic impedance contrasts with the base of S5a (lower Maykop Formation) and the top of S3, respectively (Figure 3). Tectonic control is less visible within this sequence, but some normal faults propagate through S4 generating small thickness changes (Figures 5c, 6a, 6b). S4 is deposited mainly in the central basin (500 to 1250 m thick), whilst it is a drape of sedimentary cover over the continental highs (\leq 250 m thick) (Figure 8d). S4 is generally more uniformly distributed in the central basin compared with the older sequence S3.

S5a, S5b and S5c units are associated to the regionally distributed Maykop Formation characterized by a semi-transparent seismic layer (e.g., Robinson et al., 1996; Sydorenko et al., 2017; Zonenshain & Pichon, 1986). S5 is indeed a low amplitude package with homogeneous appearance, generally linked to uniformity in its lithological character. This sequence is up to ~ 4500 m thick in the central basin, thinning over Shatsky Ridge and the MBSH (\leq 500 m thick), and thickening again along the compressed margins of the EBSB (Tuapse and Sorokin Troughs), where it can reach thicknesses of 2500 to \geq 5000 m (Figures 8e, 8f, 8g). Although the Maykop Formation is usually interpreted as a single sedimentary

package, we subdivide it into three layers: S5a, S5b and S5c (Figures 3, 4a, 4b, 5b, 6a, 6b). This sub-division is marked by the presence of weak but laterally continuous reflectors, probably linked to lithological variation within the unit. Within the lower Maykop Formation (S5a) some high reflectivity layers are probably related to the intercalation of more sandy interbeds (e.g., Stovba et al., 2009). Extensional faults propagate through S5a sequence generating small thickness changes (Figures 5c, 6a, 6b). Furthermore, the lateral termination of S5a on the Shatsky Ridge fault shows evidence of normal-drag (the hanging-wall beds dip away from the normal fault – Withjack et al., 2002) along the fault plane (Figures 5b, 6b). In contrast, S5b and S5c show no particular evidence for extensional tectonics, and their strata onlap in reverse-drag (the hanging-wall beds dip towards the normal fault – Withjack et al., 2002) onto the basin-bounding faults (Figures 5b, 6b). S5a, S5b and S5c isopach maps show depo-center migration through time (Figures 8e, 8f, 8g). S5a has quite a uniform distribution in the central basin, with visible thickness changes controlled by normal fault activity along the Shatsky Ridge fault, and thrust fault activity along the NE margin of the EBSB (Figure 8e). The compressional tectonics is associated with the folding and thrusting visible in the seismic data, also affecting the shallower units S5b and S5c (Figures 4a, 5b). The S5b main depo-center is located in the NW corner of the basin (Figure 8f), whereas S5c main depo-center migrates towards the SE (Figure 8g). No clear tectonic control is visible within these units, so depo-center migration is probably due to a change in the main source of sediment supply. Furthermore, small thickness changes along fault planes, visible in the S5b and S5c isopach maps, may be related to the compressive reactivation of the basin-bounding faults (Figures 5b, 6b, 8g).

S6 and S7 are well layered and sub-parallel, medium to high reflectivity units with alternating semi-transparent strata (Figure 3). They represent the shallowest and youngest sequences filling the basin. Our interpretation matches previous seismic interpretation and well data correlation (e.g., Nikishin et al., 2015a; Shillington et al., 2008). The thickness of sequence S6 is up to 2500 m in the central basin, whilst sequence S7 reaches a thickness of 100-2000 m across the EBSB (Figures 8h, 8i). S6 is thicker towards the SE part of the basin, perhaps related to the main source of sediment supply in that area (Figure 8h).

5. Discussion

5.1. Character and distribution of the EBSB sedimentary infill

5.1.1. Age of the interpreted sequences

There are no new well data available for this study, so we rely on published information to define the age and litho-stratigraphy of the interpreted sequences. The Palaeocene(?)–Eocene to recent EBSB infill is generally well known and considered as a post-rift, based on onshore and offshore drilling (e.g., Ross, 1978; Sydorenko et al., 2017), whereas the age and distribution of the syn-rift sequences is much less constrained. Syn-rift units interpreted on seismic data rely mostly on onshore observations (e.g., in Turkey, Crimea and Georgia). However, correlation of the syn-rift laying in the central basin with its approximate onshore equivalent, requires the assumption that the timing of rifting is synchronous across this region.

The age of the interpreted sequences S1 and S2 is difficult to constrain, for the reasons stated above. Sequences overlying S2 are instead within depths investigated by some of the offshore drilling, especially in areas where those sequences have been uplifted by Eocene compression (e.g., Hippolyte et al., 2018; Sydorenko et al., 2017). In the N-NE part of the EBSB, offshore Crimea, the Subbotina-403 and -1 wells reached the Neogene and Oligocene–Eocene sequence (Stovba et al., 2009; Tari & Simmons, 2018). The oldest sequence identified consists in sediments of Lower Eocene age. Above them, lies the Oligocene–Miocene Maykop Formation (corresponding to S5 units). Seismic data from the northern EBSB, correlated to Subbotina-403 well, are presented by Stovba et al. (2009) and Sydorenko et al. (2017). These profiles cross some of our seismic lines (bs220, bs180, bs210 and bs80), so we can correlate our interpreted sequences with this well stratigraphy. Sydorenko et al. (2017) interpreted a single sequence beneath the Maykop Formation, which includes the Palaeocene–Eocene deposits and corresponds to our S3–S4 units. Below that sequence, they interpreted an undifferentiated Cretaceous sequence, probably including our S1 and S2 units, which overlies an Upper Jurassic pre-rift in the northern part of the Shatsky Ridge.

Comparing the stratigraphic information from Subbotina-403 with the seismic facies of our interpreted sequences, we further subdivide the Palaeocene–Eocene unit. The S3 medium to low reflectivity facies, with high reflectivity towards the top (Figure 3), corresponds to the Lower Eocene sequence litho-stratigraphy sampled by Subbotina-403, which is made of marls interlayered with clay and sandy layers towards the top (Sydorenko et al., 2017). The seismic character of sequence S4 (Figure 3) matches well with the Middle–Upper Eocene sequences sampled by Subbotina-403, showing alternating shale and sandy units, with sandy layers dominating top and base of the sequence, which would generate strong acoustic-impedance contrasts. Based on this evidence, we define S3 to be Palaeocene(?)–Lower Eocene (45–65 Ma) unit and S4 Middle–Upper Eocene (45–33.9 Ma) unit. Consequently, the underlying sequences S1 and S2 represent an undifferentiated Upper Cretaceous infill, in agreement with other studies (e.g., Shillington et al., 2008; Sydorenko et al., 2017; Tari & Simmons, 2018).

From published litho-stratigraphy it is possible to define sequences S5a, S5b and S5c as the Oligocene and Lower–Middle Miocene part of the Maykop Formation, while S6 and S7 correspond to the Upper Miocene and Plio–Quaternary, respectively (e.g., Hippolyte et al., 2018; Sydorenko et al., 2017). Here, we subdivide the Maykop Formation into three sequences (S5a–

S5b-S5c). This subdivision is due to the presence of regional, medium-low reflectivity horizons that we interpret as the top and base of further packages into which this sequence is subdivided (Figures 3, 4a, 4b, 6a, 6b). The presence of these reflective boundaries within the Maykop Formation may be related to a change in the sedimentary character and physical properties of the sequence. The Maykop Formation is considered to be an over-pressured layer (e.g., Scott et al., 2009), so this subdivision may imply a compartmentalisation of the overpressure within the sequence. A previous subdivision of the Maykop Formation is described by Jones & Simmons (1997). This layering within the Maykop Formation is not only representative for the potential variability in its physical and lithological properties, but it also associates to a different syn-kinematic (S5a) to post-kinematic (S5b-S5c) character of the sequence.

5.1.2. Rift Stage

Although some studies have interpreted the presence of a syn-rift infill in the EBSB (e.g., Shillington et al., 2008; Sydorenko et al., 2017; Tari & Simmons, 2018), most authors tend to agree on the absence/near absence of syn-rift deposits in the central EBSB (e.g., Görür & Tüysüz, 1997; Görür et al., 1993), due to the lack of strong evidence for syn-rift sedimentary geometries on seismic profiles (e.g., Nikishin et al., 2015a).

In our seismic profiles, syn-tectonic growth structures and wedge geometries bounded by normal faults are clear features over the Shatsky Ridge and the MBSH (Figures 4a, 5a). Here, overlying a faulted pre-rift (basement), sequences S1 and S2 (Upper Cretaceous) are interpreted as two distinct syn-rift deposits with a combined thickness up to 3 km (Figures 4a, 5a, 5c, 8a, 8b). Studies based on onshore observations also identified the presence of two syn-rift sequences in the region (e.g., Hippolyte et al., 2018; Nikishin et al., 2015a; Tari & Simmons, 2018). We define the contact between pre-kinematic (basement) and S1 as unconformity U1, which we interpret as a *rift-onset unconformity* (Franke, 2013; Withjack et al., 2002), marking the initial rifting event affecting the structural highs (Rift stage 1 - Figure 9a). This unconformity progressively merges basinwards with the stratigraphically younger unconformity U2 (base of S2 unit) forming a composite unconformity (Figures 4a, 5a, 5c). The merging of these unconformities is due to the gradual thinning of S1 towards the basin center, where half-graben structures are mostly filled by the younger syn-rift (S2) instead (Figures 4a, 5a, 5c). The presence of an unconformity like U2, separating the early syn-rift strata from the later syn-rift strata, is a common feature in other rift basins (e.g., Buckley et al., 2015; Olsen, 1997; Lei et al., 2019). This unconformity shows the basinward migration of extension in the EBSB, associated with a progressive younging of the syn-rift infill towards the rift axis (Rift stage 2 - Figure 9b). This two-stage rifting is visible in S1 and S2 isopach maps, where there is a clear basinward propagation of the main depo-center location (Figures 8a, 8b).

Rift migration is also recorded by a change in the character and age of the extensional faulting. Faults interpreted over the structural highs generally show higher dip angles (30°-70°). Here, the faulted basement is covered by the older S1 unit, and fault activity continues through the S1 and S2 units, stopping at the base of S3 unit (Figures 4a, 5a). Downslope, faults are organized into more closely spaced domino-fault arrays, characterized by lower dip angles (15°-25°) (Figures 4a, 5c). In this area, the faulted basement is directly overlain by the younger syn-rift sequence (S2), and faults propagate sometimes into shallower units (up to S5a) steepening upwards (Figures 5c, 6a, 6b). Fault initiation and duration can be constrained by the age of the syn-kinematic infill, so basinward faults formed later than those interpreted over the ridges.

Similar evidence for syn-kinematic units recording the formation of faults that young oceanward is observed along other rift basins and rifted margins (e.g., Nixon et al., 2016; Péron-Pinvidic et al., 2007; Ranero & Pérez-Gussinyé, 2010; Reston, 2005). Various observational studies and numerical models show that rifting starts within a network of isolated normal fault segments generating a moderate crustal thinning (low β) and distributing deformation across the area (e.g., Cowie et al., 2000; Nagel & Buck, 2004; Péron-Pinvidic et al., 2017). As stretching continues, some faults grow laterally and coalesce into dominant basin-bounding single faults (e.g., Cowie et al., 2000). Deformation focuses along these few active, larger faults, whereas smaller faults along the less thinned rift flanks become inactive (e.g., Cowie et al., 2000; Ranero & Pérez-Gussinyé, 2010). As strain progressively migrates and focuses towards the future basin center, sequentially active normal faults form to accomplish rift migration (e.g., Brune et al., 2014; Nagel & Buck, 2004; Ranero & Pérez-Gussinyé, 2010). During this stage, further crustal thinning (high β) and increasing accommodation space occurs. In the EBSB, we consider high angle normal faults to record the initial stage of distributed faulting (Figure 9a), followed by subsequent fault linkage and generation of the master faults bounding the structural highs (Figure 9b). Continued extension then causes the border faults to experience more strain and basinward focusing of crustal thinning accompanied by the generation of new, lower angle sequential faults accommodating the basinward propagation of stretching (Figures 9b, 9c). The seaward decrease in fault dips is predicted by the rolling hinge model of Buck (1988) and Lavier et al. (1999). Evidence for normal faults initiating at higher dip angles and then rotating to lower angles during progressive extension is visible at other rifted margins (e.g., Ranero & Pérez-Gussinyé, 2010; Reston, 2005), and back-arc basins (e.g., Lei & Ren, 2016). This model may explain the progressive change in fault dip observed in the EBSB. Based on both stratigraphic and structural elements in the EBSB, we recognize a two-stage rifting process, initiating by widespread extension (Figure 9a), and followed by rift migration and strain focusing towards the basin axis (Figure 9b). Our observations from the EBSB are similar to those from other rift basins and rifted margins such as the Bay of Biscay (e.g., Tugend et al., 2015), the Iberia-Newfoundland margin (e.g., Brune et al., 2017; Ranero & Pérez-Gussinyé, 2010; Reston, 2005), the Brazil-Angola margin (e.g., Brune et al., 2017), the Gulf of Corinth (e.g., Nixon et al., 2016), and the northern margin of the South China Sea (e.g., Lei & Ren, 2016; Lei et al., 2019).

5.1.3. Breakup Stage

In our seismic interpretation, S3 lies in angular unconformity over the tilted and faulted blocks, covering the S1-S2 wedged geometries on the structural highs and cross-cutting, basinward, the tops of the S2 wedges (Figures 5a, 5c). The angular contact between S3 (Palaeocene(?)-Eocene) and S2 (Upper Cretaceous) defines unconformity U3 (Figures 5a, 5c). Based on the geometrical relationship of the contact between S2 and S3, U3 can be interpreted as a *breakup unconformity* (Falvey, 1974; Franke, 2013). Basinward, U3 becomes instead the stratigraphic contact between S3 and the underlying basement (Figures 4b, 7a, 7b). Previous studies interpreted the Palaeocene(?)-Eocene units as a post-rift infill (e.g., Nikishin et al., 2015a; Sydorenko et al., 2017). However, there are evidence for normal faults propagation within these units in the NW part of the basin (Figures 5c, 6a, 6b), together with evidence for strata terminations dragged upward along the main basin-bounding fault planes, indicating extensional activity during their deposition (Figures 5b, 6b). Similarly, S5a (Oligocene Maykop Formation) shows evidence for active extension in the NW (Figures 5c, 6a, 6b). Isopach maps highlight the irregularity of the S3 and S4 infills (Figures 8c, 8d). These irregularities partly result from the

topography of the underlying basement affecting the distribution of the infill (Figure 7a), and/or may be related to differential compaction processes. However, observations along NW seismic profiles show clear evidence for active faulting during S3-S4 deposition (Figures 5c, 6a, 6b). The S5a sequence appears to be more uniformly distributed across the basin, but fault activity observed along NW seismic profiles (Figures 5c, 6a, 6b), together with thickening towards the Shatsky Ridge fault visible on the isopach map, suggest fault activity during the time of S5a deposition (Figure 8e).

Moving towards the central and SE part of the basin, there is no clear evidence of active extension in the S3 to S5a units. Here, these sequences appear like a typical post-rift infill overlying a basement reflection (Figures 4b, 7a, 7b). The crustal velocity structure, determined from ocean bottom seismometer data, suggests that the SE part of the EBSB is underlain by thick oceanic crust (Shillington et al., 2009). This interpretation agrees with the lack of evidence for extension, and with the post-rift character of S3-S5a sequences in this part of the basin. Therefore, S3-S5a probably consist of marine sediments deposited during and after the breakup and spreading in the SE of the basin. The S3-S4-S5a units show an along-strike change in stratigraphic character that appears to be intermediate between syn-rift and post-rift sequences. The seismic character of these units shows similarities with the definition of *breakup sequences* given by Soares et al. (2012) and Alves & Cunha (2018). Similarly to this definition, S3 to S5a units record the transitional period between the onset of breakup (SE) and the end of extension (NW), followed by the establishment of thermal relaxation as the main process controlling subsidence in the basin. Based on seismic evidence along the west Iberia-Newfoundland margins, Alves & Cunha (2018) identified within the *breakup sequence* distinct seismic intervals of chaotic to partly continuous seismic reflections terminating against the slope fault system and interpreted these intervals as mass-wasting and turbiditic deposits associated to the breakup stage. These units are then gradually draped by wavy to sub-parallel, low amplitude seismic reflections interpreted to reflect the spreading conditions on a fully developed margin. Similarly, we recognize that S3 and S4 units are locally organized in chaotic deposits linked to slope degradation processes, overlain by the younger and semi-transparent units of the Maykop Formation (Figures 5b, 6a, 6b). This may be seen as a further evidence of the comparable character between our S3, S4 and S5a units and the *breakup sequence* defined by Soares et al. (2012), both recording the diachronous evolution of breakup processes at extensional settings. Based on these considerations, and on similarities with observations coming from other studies (e.g., Péron-Pinvidic & Manatschal, 2009; Soares et al., 2012), S3 to S5a units are interpreted to record a diachronous breakup stage across the EBSB (Figure 9c).

These considerations imply that the defined *breakup unconformity* (U3) cannot be considered representative for a widespread end of rifting and breakup initiation across the EBSB, as we need to acknowledge for the temporal and spatial variability of these processes across the basin (e.g., Péron-Pinvidic et al., 2007). The breakup stage (Figure 9c) is therefore time constrained by the ages of S3 to S5a infills, which show that extension continued up to Oligocene time in the NW, whereas breakup started before the deposition of S3 (Palaeocene(?)-Early Eocene time) in the SE, synchronous with the NW ongoing extension. The total thickness of S3 to S5a sequences ranges from 1.8 km to 4-4.5 km in the central basin. Because these units form part of the syn-kinematic infill in areas where extension was active up to Oligocene time, they are considered to contribute to the overall thickness of the syn-rift deposits of the EBSB.

5.1.4. Post-rift Stage

Seismic data show no evidence for extensional tectonics within S5b, S5c, S6 and S7 units (Figures 6a, 6b). As these sequences all show a ubiquitous post-rift character, we consider the onset of thermal relaxation to be represented by the Lower Miocene Maykop Formation (S5b) (Post-rift Stage – Figure 9d). Isopach maps show that these post-rift sequences progressively infill the pre-existing topography, while compression starts deforming the sediments accumulated along the NE margin (Tuapse and Sorokin Troughs) (Figures 8f, 8g, 8h). Thickness changes within these sequences are associated with depo-center migration related to a change in the source of sediment supply through time, probably partially controlled by compressional deformation along the peripheral regions of the basin.

5.2. Final considerations

The identification of syn-rift stratigraphic intervals is a very important aspect in the study of rifted margins (e.g., Péron-Pinvidic et al., 2007). In the EBSB, we defined the presence of two clear syn-rift units (S1, S2), which are ascribed to two separate Late Cretaceous extensional stages (Figures 9a, 9b). S3 to S5a units (Palaeocene(?) to Oligocene) are also interpreted to represent part of the syn-kinematic infill along the NW part of the basin, where active tectonics affects their deposition (Figures 6a, 6b). Previous studies suggested various timings for the syn-rift and thus for the basin opening, such as Aptian-Albian (e.g., Görür, 1988; Hippolyte et al., 2018), Cenomanian-Turonian-mid Santonian (e.g., Tari & Simmons, 2018; Tüysüz, 1999), or Palaeocene-Early Eocene or Eocene (e.g., Banks et al., 1997; Kazmin et al., 2000; Robinson et al., 1995a, 1995b), based on onshore data and well data acquired at different locations across the EBSB region. Similarly, breakup process has been estimated to be either Albian-Cenomanian (e.g., Görür, 1988; Hippolyte et al., 2018; Nikishin et al., 2015b), Turonian-Coniacian (e.g., Tari & Simmons, 2018), mid-Santonian (e.g., Tüysüz, 1999), or Maastrichtian-Danian (e.g., Okay et al., 1994, 2013; Shreider, 2005). Our study highlights the variable distribution of rifting and breakup processes at different locations and times in the basin. This distribution is recorded within the variable architecture and stratigraphic character of the identified sequences, showing either a syn-kinematic or post-kinematic character. With this observation in mind, the definition of rifting and breakup ages would be strictly dependent on the location of the study, and not really representative of the timing of the process on a regional scale. This complexity may explain such a variety of results and the general disagreement between different studies.

The EBSB is a multi-stage narrow rift formed by initial widespread stretching (Figure 9a), followed by rift migration towards the basin axis (Figure 9b), and progressive strain localization into the area of the future breakup (Figure 9c). Once breakup and spreading initiated along the more extended SE segment, extension continued along the less extended NW segment. This is well expressed by the Palaeocene(?)–Eocene and Oligocene sequences, which have both a syn-kinematic character in the area where extension continued up to Oligocene (NW) (Figure 9c), and a post-kinematic character in the area where breakup and “new” basement was generated (SE) (Figure 9d). Together with the along-axis variability in sedimentary ages and architecture, the change in character of the extensional faulting records the different rifting stages (e.g., Lavier & Manatschal, 2006; Moore, 1992; Péron-Pinvidic et al., 2007; Sutra et al., 2013). Starting as isolated high angle normal faults, some of these systems merged to form the main basin-bounding faults. Strain then migrated towards the basin axis forming younger, low angle domino-fault arrays, associated with the progressive localization of rifting into the area of the

future breakup (Figures 9a, 9b, 9c). This along-strike variability in fault geometry and sedimentary architecture reflects a specific kinematics of basin opening. Several studies have proposed models to explain formation and evolution of V-shaped basins (e.g., Le Pourhiet et al., 2018; Lundin et al., 2014; Nirrengarten et al., 2018; Taylor et al., 1995; Zhou et al., 1995). Some models refer to a “scissor-like” opening (e.g., Hey et al., 1980), whereas other invoke a rift propagation or “zipper-slider” opening style, where the tip point of extension initiates in one area to then “jump” (e.g., Taylor et al., 1995) or progressively migrate along with further rift propagation (e.g., Martin, 1984; Vink, 1982). The analysis of oceanic crust magnetic isochrons provides key information for the identification of propagation mechanisms, as it may reveal the diachronous development of oceanic crust younging towards the pole of rotation, thus allowing the reconstruction of its evolutionary pattern (e.g., Franke, 2013; Lundin et al., 2014). The three-dimensional modelling of Le Pourhiet et al. (2018) simulates various scenarios for the opening mechanism of the V-shaped South China Sea basin, showing that a “scissor-like” model contradicts evidence from magnetic isochrons and conjugate margin structure, whereas previous studies invoked such a mechanism for this back-arc opening (e.g., Zhou et al., 1995). Recent studies have investigated the Southern North Atlantic basins to understand the partitioning and propagation of deformation, proposing a modified version to the model of oceanic propagation (e.g., Martin, 1984) by introducing the idea of segmented propagation controlled by the presence of transform faults (Nirrengarten et al., 2018).

Our observations of along-strike variations in temporal and spatial patterns of extension show clear evidence for rift propagation, associated with progressive basin opening. Although it is generally assumed that the EBSB opened around a “fixed” rotation pole located in Crimea (e.g., Okay et al., 1994), thus implying a “scissor-like” opening kinematic, sediment distribution and fault patterns seem to indicate that the tip of the propagator could have migrated in time from SE to NW. To test this idea, detailed three-dimensional mapping of the fault distribution and of the deep sedimentary sequences is required in order to date the migration of deformation across the basin (e.g., Péron-Pinvidic et al., 2007). The sparse coverage of our dataset in some areas precludes such mapping. A lack of clear seafloor spreading anomalies over the inferred oceanic domain (e.g., Graham et al., 2013; Spadini et al., 1996) also limits our ability to constrain the timing of propagation in this area. Our observations provide initial spatiotemporal markers to constrain models of EBSB rifting and continental breakup, but ultimately more closely spaced profiles of similar quality, or even 3D seismic datasets, will be required to determine exactly how the rift propagated.

This study highlights the similarity of EBSB evolution to that of rifted margins (e.g., Iberia-Newfoundland and Brazil-Angola margins - Brune et al., 2017; Reston et al., 1996; Tucholke & Sibuet, 2007), back-arc basins (e.g., Tyrrhenian basin - Prada et al., 2015; Aegean Sea, Japan Sea and Yamato Basin - Jolivet et al., 1994, 1999; South China Sea - Lei & Ren, 2016; Woodlark Basin - Taylor et al., 1995), and rift basins elsewhere (e.g., Gulf of Suez - Gawthorpe et al., 2003; North Sea - Cowie et al., 2005), making the EBSB a case study for understanding the processes driving the evolution of extensional basins on a global scale.

6. Conclusions

Based on the interpretation of long-offset seismic reflection profiles, we conclude:

- 1) Overlying the EBSB rifted basement, up to 3 km of Late Cretaceous syn-rift deposits (S1 and S2) record an initial localised extension over the structural highs (S1), followed by rift migration towards the basin axis (S2).
- 2) Strain initiated along isolated high angle normal faults (30-70° dip) over the structural highs. Some faults coalesced forming the large systems bounding the ridges (Rift Stage 1). Strain then propagated downslope towards the basin axis, forming closely spaced domino-style fault arrays characterised by lower dip angles (15°-25°). Stratigraphic contacts show that these fault arrays are younger and they sometimes affect shallower stratigraphic levels (up to S5a), recording the process of strain localisation and basinward propagation (Rift Stage 2).
- 3) Three main unconformities are identified: i) the *rift-onset unconformity* (U1) represents the initiation of Late Cretaceous extension, ii) an intermediate rift unconformity (U2) records the migration of strain and the basinward younging of the syn-rift infill and normal faulting, iii) the *breakup unconformity* (U3) (Upper Cretaceous-Palaeocene(?)) records breakup initiation and spreading in the SE.
- 4) Above the *breakup unconformity* (U3), up to 4-4.5 km of Palaeocene(?)-Eocene and Oligocene deposits (S3-S4 and S5a) show an intermediate character between syn-rift and post-rift, recording the diachronous distribution of syn-kinematic and breakup processes across the EBSB. Stratigraphic and structural evidence within these units shows extension up to Oligocene time, in the NW area and along the main basin-bounding faults. In the central and SE part of the EBSB, the post-rift character of these units confirms that localised breakup may have occurred before their deposition, and thus at the Late Cretaceous-Palaeocene (?) boundary, synchronous with extension in the NW.
- 5) The diachronous evolution of the EBSB is typical of well-studied rifted margins worldwide, and provides further evidence for the need to consider continental breakup as a fully three-dimensional process that is poorly represented by a single transect.

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Zonenshain, L. P., & Pichon, X. (1986). Deep basins of the Black Sea and Caspian Sea as remnants of Mesozoic back-arc basins. *Tectonophysics*, 123(1-4), 181-211.
[https://doi.org/10.1016/0040-1951\(86\)90197-6](https://doi.org/10.1016/0040-1951(86)90197-6)

#	List of Authors	Data used to constrain timing of rift
(1)	<i>Adamia et al. (1974)</i>	Structural relationships from EBSB coastline, ages of magmatism, plate reconstruction.
(2)	<i>Letouzey et al. (1977)</i>	Geological observations, seismic reflection data analysis.
(3)	<i>Hsü et al. (1977)</i>	Plate reconstruction.
(4)	<i>Tugolesov et al. (1985)</i>	Seismic data interpretation.
(5)	<i>Zonenshain & Pichon (1986)</i>	Seismic evidences and subsidence history analysis.
(6)	<i>Finetti et al. (1988)</i>	Seismic reflection data and onshore observations.
(7)	<i>Görür (1988); Görür & Tüysüz (1997); Görür et al. (1993)</i>	Paleogeography based on geological observations and stratigraphy Central Pontides.
(8)	<i>Golmshtok et al. (1992)</i>	Seafloor heat-flow measurements and multichannel data.
(9)	<i>Rudat & Macgregor (1993)</i>	Wells data from Shatsky Ridge in Georgia and dredging from Archangelsky Ridge.
(10)	<i>Okay et al. (1994); Okay & Şahintürk (1997); Okay & Tüysüz (1999)</i>	Structural/stratigraphic evidences, age of magmatic products of Eastern Pontides.
(11)	<i>Bektas & Güven (1995)</i>	Age of magmatism in Eastern Pontides.
(12)	<i>Robinson et al. (1995a), (1995b), (1996)</i>	Seismo-stratigraphic interpretation and well data.
(13)	<i>Spadini et al. (1996)</i>	Kinematic modelling and thermal structure analysis.
(14)	<i>Banks et al. (1997)</i>	Onshore outcrops, well data and seismic interpretation onshore Georgia.
(15)	<i>Shreider et al. (1997); Shreider (2005)</i>	Magnetic anomaly studies.
(16)	<i>Tüysüz (1999); Tüysüz et al. (2012)</i>	Analysis of volcanogenic rocks in north Turkey.
(17)	<i>Yilmaz et al. (2000)</i>	Structural correlations (Transcaucasus and Eastern Pontides), and age of magmatism.
(18)	<i>Kazmin et al. (2000)</i>	Thickness and distribution of Palaeocene-Eocene sedimentary units from seismic data.
(19)	<i>Verzhbitsky et al. (2002)</i>	Heat-flow analysis and seismic data interpretation.
(20)	<i>Cloetingh et al. (2003)</i>	Thermo-mechanical modelling.
(21)	<i>Nikishin et al. (2003), (2011), (2013), (2015b), (2015c)</i>	Correlation of Crimean onshore geology and regional seismic reflection profiles.
(22)	<i>Shillington et al. (2008)</i>	Integration of onshore geological mapping, well data and seismic reflection data.
(23)	<i>Stovba et al. (2009)</i>	Seismic data interpretation and correlation with Subbotina-403, -1 wells.
(24)	<i>Stephenson & Schellart (2010)</i>	Geo-dynamic modelling of back-arc extension.
(25)	<i>Hippolyte et al. (2010), (2017), (2018)</i>	Structural and stratigraphic analysis in Crimea and Pontides, and nannoplankton ages.
(26)	<i>Sheremet et al. (2016)</i>	Micro-paleontological dating sampled on the Crimean Mountains.
(27)	<i>Vincent et al. (2016)</i>	Subsidence analysis.
(28)	<i>Sydorenko et al. (2017)</i>	Seismic interpretation offshore Crimea, and correlation with Subbotina-403 well.
(29)	<i>Tari (2015); Tari & Simmons (2018)</i>	Stratigraphic compilation of published information and well data.

Table 1. Table listing in chronological order the studies used as a reference for the previous estimated time of rifting and breakup in the EBSB. Numbers in the left column refers to the time-bar shown in Figure 2. For each of the studies, the methodology used to determine the time-constraints is briefly described.

Figure 1. (a) Topographic and structural map of the Black Sea and Caucasus region (www.emodnet-bathymetry.eu). The black box represents the study area (b). Structural elements highlighted are: EBSB/WBSB —Eastern/Western Black Sea Basins; SR—Shatsky Ridge; MBSH—Mid Black Sea High; TT/ST—Tuapse/Sorokin Troughs; GC/LC - Greater/Lesser Caucasus; TC—Transcaucasus; EP/WP—Eastern/Western Pontides; NAF/EAF—North/East Anatolian Fault. (b) Bathymetry and topography map of the EBSB. Black lines mark the long-offset seismic reflection profiles. Orange lines mark other profiles used. Yellow dots mark deep-water and onshore wells in the area. Surmene-1, Sinop-1, HPX-1 and Subbotina-403 wells locations are from Kitchka et al. (2014), Sydorenko et al. (2017) and Tari & Simmons (2018). The location for the onshore well Ochamchira is from Banks et al. (1997).

Figure 2. Timing of rift and breakup in the EBSB from previous studies. The black bar indicates rift duration. The grey bar indicates the age of a second syn-rift deposit identified by Tari & Simmons (2018). When available, timing of breakup (BU) and oceanic spreading are shown (hashed bar). Numbers correspond to the studies listed in Table 1. The geological time-scale is adapted from Cohen et al. (2016).

Figure 3. Summary of the interpreted seismo-stratigraphic units (S1—S7). From left to right: seismic examples for each sequence; litho-stratigraphy information from previous studies; sequence number and age estimated from correlation with previous studies. References: ¹Ross (1978); ²Robinson et al. (1995a); ³Nikishin et al. (2003); ⁴ Robinson et al. (1996); ⁵ Görür & Tüysüz (1997); ⁶ Sydorenko et al. (2017); ⁷ Robinson et al. (1995b).

Figure 4. (a) Profile bs-90. The MBSH and the Shatsky Ridge bound the central EBSB. Colors mark interpreted stratigraphic sequences (S1 to S7). The top basement is affected by normal faults (dashed lines) generating tilted crustal blocks. Normal faults over the structural highs generally show higher dip angle (30°-70°), whereas downslope they show lower dip angle (15°-25°). Syn-kinematic deformation is mainly visible within units S1 and S2, but it affects also shallower units S3-S4 and S5a, at locations. The main fault planes are interpreted in continuity across NE-SW trending profiles (red dotted lines): Shatsky Ridge fault and Andrusov Ridge fault. Black boxes show the locations of Figures 5b, 5c and 7a. (b) Profile bs-110. A major change in the character of the deep sedimentary infill is visible: no clear syn-kinematic deformation affects sequences S3 and S4, which have instead a clear post-rift character, directly overlying the central EBSB basement. The Shatsky Ridge fault is mapped across from profile bs-90, showing here a lower dip angle. The Andrusov Ridge fault is also mapped across nearby profiles. A volcanic body has been interpreted at the SW end of this profile by previous studies (e.g., Nikishin et al., 2015a). Black box shows the location of Figure 7b. (VE = 6:1).

Figure 5. (a) Part of profile bs-170, showing syn-rift wedged geometries (S1-S2). The contact between S1 and the basement is defined by unconformity U1. S2 overlies S1 in angular discordance, defining unconformity U2. Horizon U3 identifies the top of S2 and the base of sequence S3. Sequences S3 and S4 are a thinner sedimentary cover over the structural highs. (b) Part of profile bs-90. The Oligocene Maykop Formation (S5a) shows strata dragging along the Shatsky Ridge fault. The Miocene Maykop Formation (S5b-S5c) pinches-up over the same fault plane, indicating compressive reactivation along this fault. S3 and S4 units resemble the character of chaotic slope degradation deposits (VE = 4:1). (c) Part of profile bs-90. Downslope,

wedged S2 deposits represent the main syn-rift infill, overlying tilted fault-blocks. S1 is generally absent. The base of the well layered S3 unit truncates the S2 wedges, defining unconformity U3. (VE = 3:1).

Figure 6. Profiles showing fault activity through S3, S4 and S5a units: (a) part of profile bs-80; (b) part of profile bs-210. In both examples, faulting and tilting of continental blocks occur up to the Oligocene Maykop Formation (S5a). The resulting syn-kinematic topography is then progressively smoothed out by the Miocene Maykop Formation (S5b and S5c). Red dashed lines are the Shatsky Ridge fault (NE) and the Andrusov Ridge fault (SW). Mapped unconformities U1, U2, and U3 are also shown. Evidence for S5c pinching-up over the Shatsky Ridge fault (b) indicates compressive reactivation along this fault (see also Figure 5b). (VE = 4:1).

Figure 7. (a) Part of profile bs-90 (location in Figure 4a). There is no evidence for syn-rift deposits in this area. S3 directly overlies the top of a rough basement, which morphology affects S3 thickness distribution (Figure 8c). (b) Part of profile bs-110 (location in Figure 4b). No evidence for syn-rift deposits in this area. S3 directly overlies a smooth basement, which is considered to have formed due to breakup and spreading in this area.

Figure 8. Isopach maps. (a) S1 – visible over the structural highs and rapidly thinning basinward (Rift stage 1); (b) S2 – visible basinward (NW), records strain migration towards the basin axis (Rift stage 2); (c) S3 – Palaeocene(?)–Early Eocene sequence, progressively infilling the central EBSB; (d) S4— Middle-Upper Eocene; (e) S5a—Oligocene Maykop Formation. S3, S4 and S5a record ongoing extension in the NW, and a localized breakup in the SE; (f) S5b—Lower Miocene Maykop Formation, showing the end of extension and the beginning of a regionally spread post-rift subsidence; (g) S5c—Middle Miocene Maykop Formation. Thickness changes along basin-bounding faults are linked to compressional reactivation; (h) S6— Top Miocene, showing a greater amount of sediment supply in the SE, and (i) S7—Pliocene-Quaternary, progressively filling the all basin. Fault systems have been tentatively interpreted with a continuous black line (the fault was active during the time represented by that sequence), or with a dashed line (ceased/not started fault activity at that time).

Figure 9. A summary of the EBSB evolution, with sketches based on seismic reflection profiles from different parts of the basin (sketches not in scale). (a) Rift Stage 1: Crustal stretching (low β) forming isolated depo-centers (*Rift Onset Unconformity* - U1), bounded by high-angle normal faults (30-70° dip). Graben and half-grabens are filled by syn-rift (S1) (b) Rift Stage 2: Isolated faults coalesce forming main basin-bounding faults and extension migrates towards the basin axis, where increased crustal thinning (high β) is accommodated by lower angle (15-25° dip) fault arrays. Half-grabens are filled by the younger syn-rift (S2). Unconformity U2 records the basinward migration of rifting. (c) Breakup Stage: Its onset is marked by unconformity U3. Units S3, S4, and S5a show extension up to Oligocene time (S5a) in the NW, synchronous to localized breakup and spreading in the SE, where these units show a clear post-rift character (d). (d) Post-rift Stage: The Lower Miocene Maykop Formation represents the first widespread post-rift infill. Sketch profiles locations are plotted over the survey map on the top right corner.

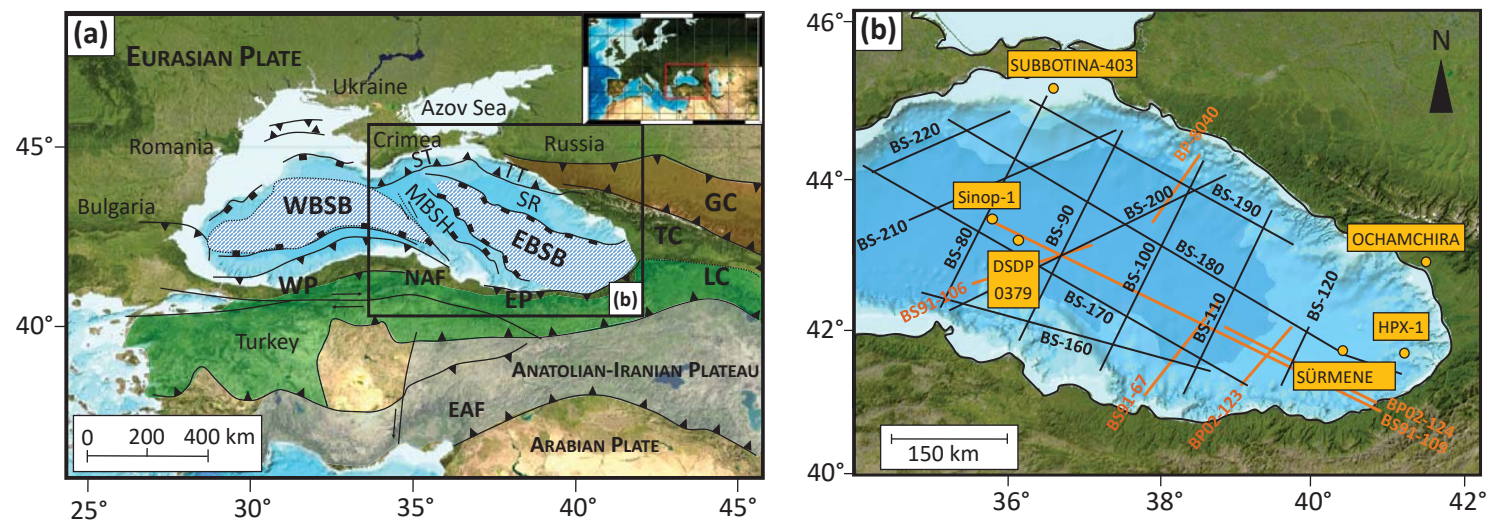


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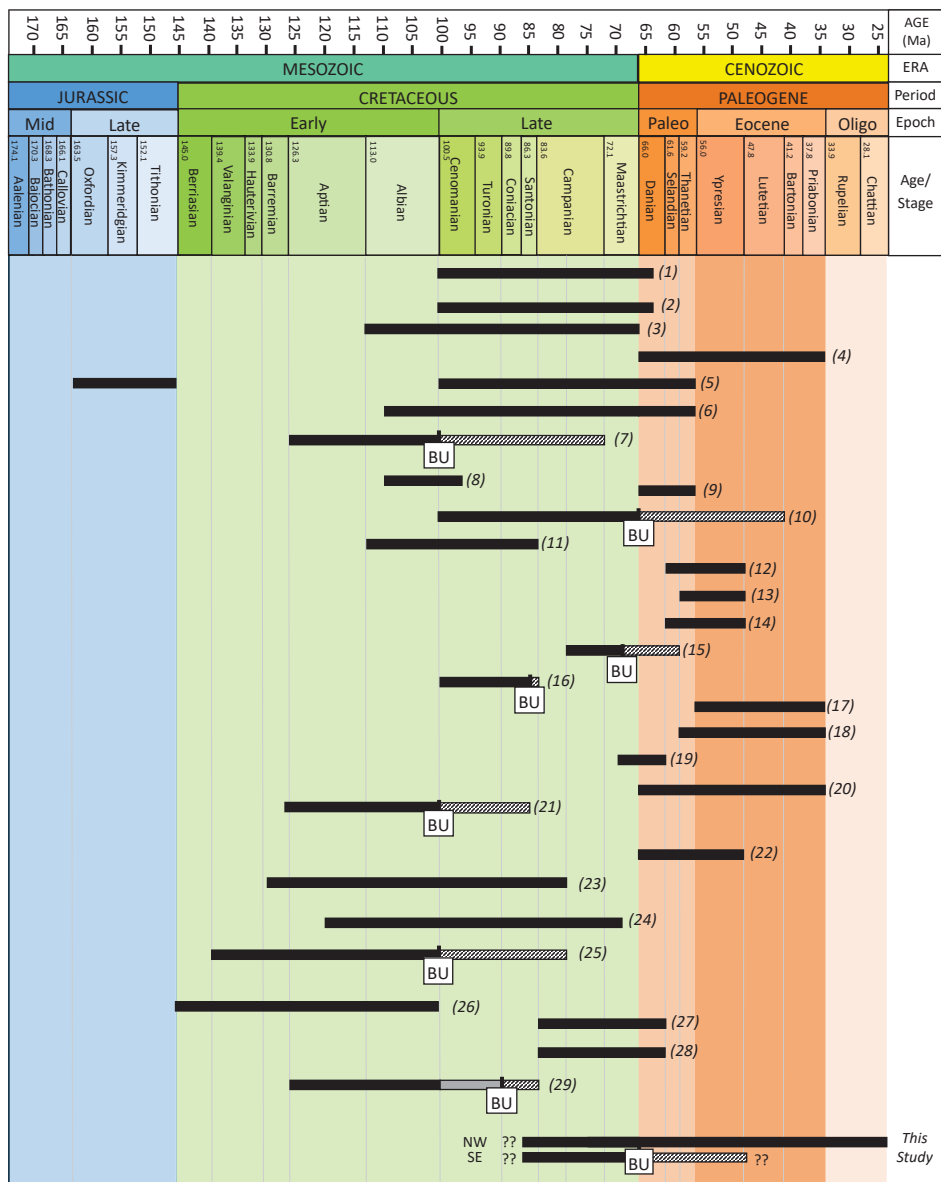


Figure 2.


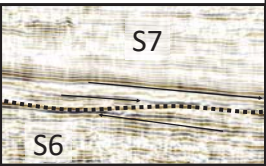
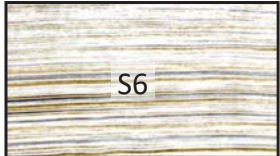
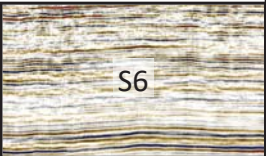
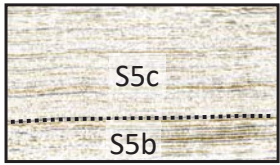
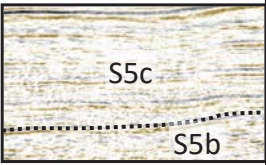
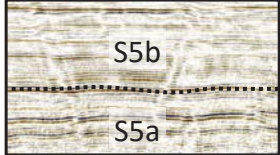
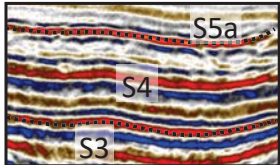
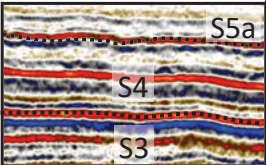
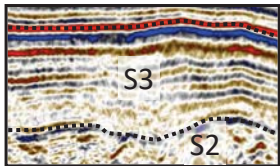
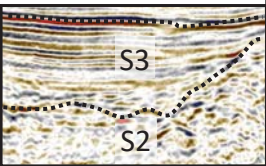
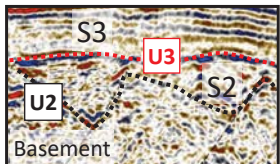
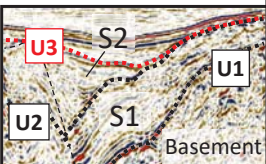
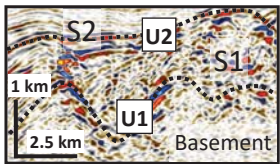
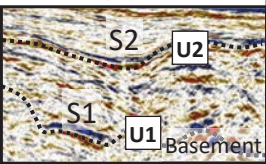
SEISMIC FACIES		LITHO-STRATIGRAPHY	SEQ #
		<ul style="list-style-type: none"> ♦ Plio-Quaternary sediments. ♦ Mostly clays, including marls and occasional turbidites¹. 	S7 (11-1.8 Ma)
		<ul style="list-style-type: none"> ♦ Upper Miocene interval². ♦ The shallower sequence (13-11 Ma) comprises terrigenous sediments³. 	S6 (20.5-11 Ma)
		<ul style="list-style-type: none"> ♦ Oligocene-Early Miocene Maykop Formation. ♦ Hydrocarbon source rock in the Black Sea and Caspian Sea. ♦ Muds rich in organic carbon with very little sand⁴. 	S5 (33.9-20.5 Ma)
			
		<ul style="list-style-type: none"> ♦ S3 and S4 associated to Palaeocene(?)–Lower Eocene and Middle–Upper Eocene formations, respectively. 	S4 (45-33.9 Ma)
		<ul style="list-style-type: none"> ♦ Eocene succession is composed by dominantly silicoclastic turbidites (sandstones and shales)⁵⁻⁶. 	S3 (45-65? Ma)
		<ul style="list-style-type: none"> ♦ The basement consists of Upper Jurassic to Cretaceous (over the MBSH) or Upper Cretaceous (over the Shatsky Ridge) limestones. ♦ The overlain sequences (S1-S2) are associated to the Upper Cretaceous interval⁷. 	S2 (65?–? Ma)
			S1

Figure 3.

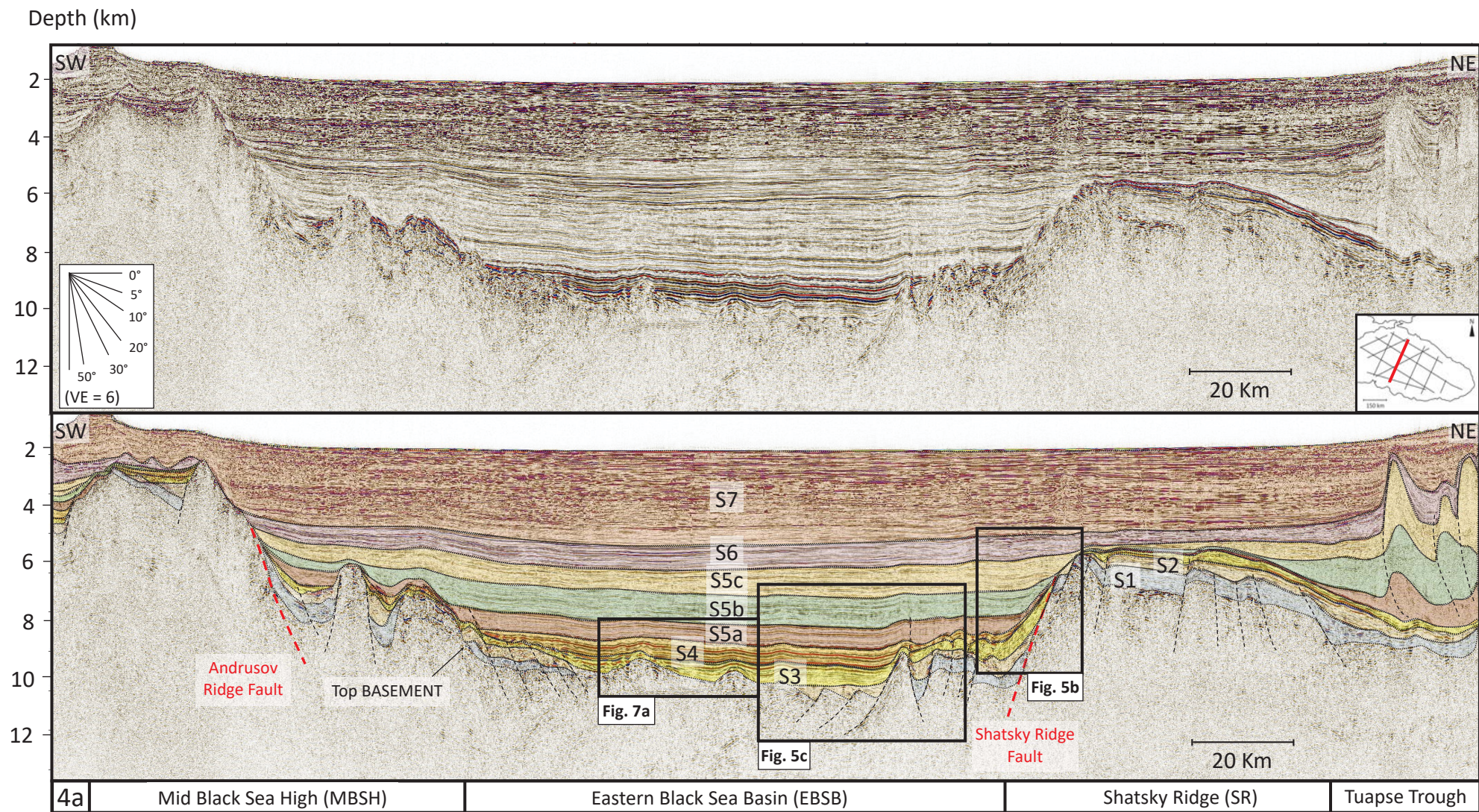


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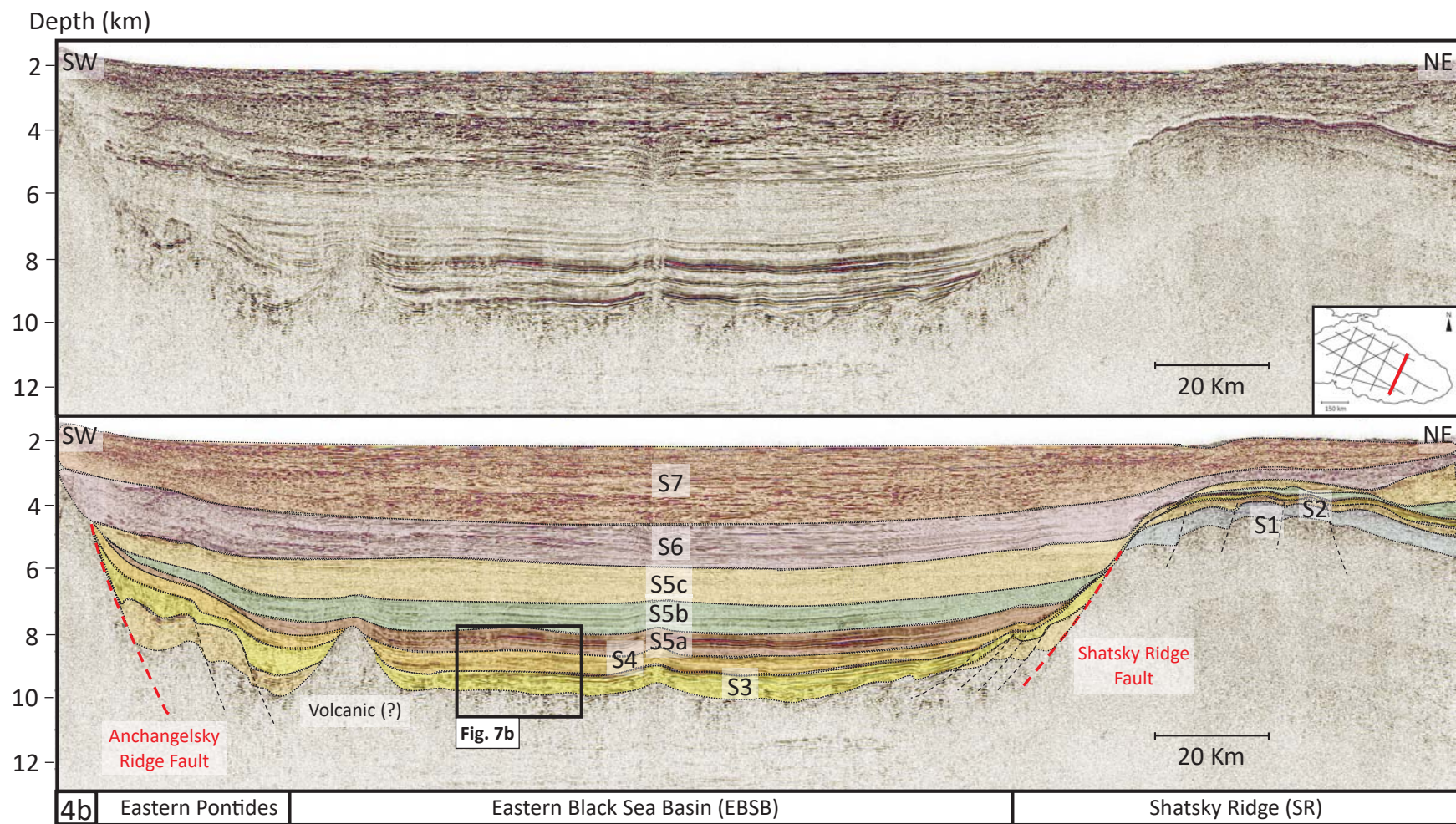


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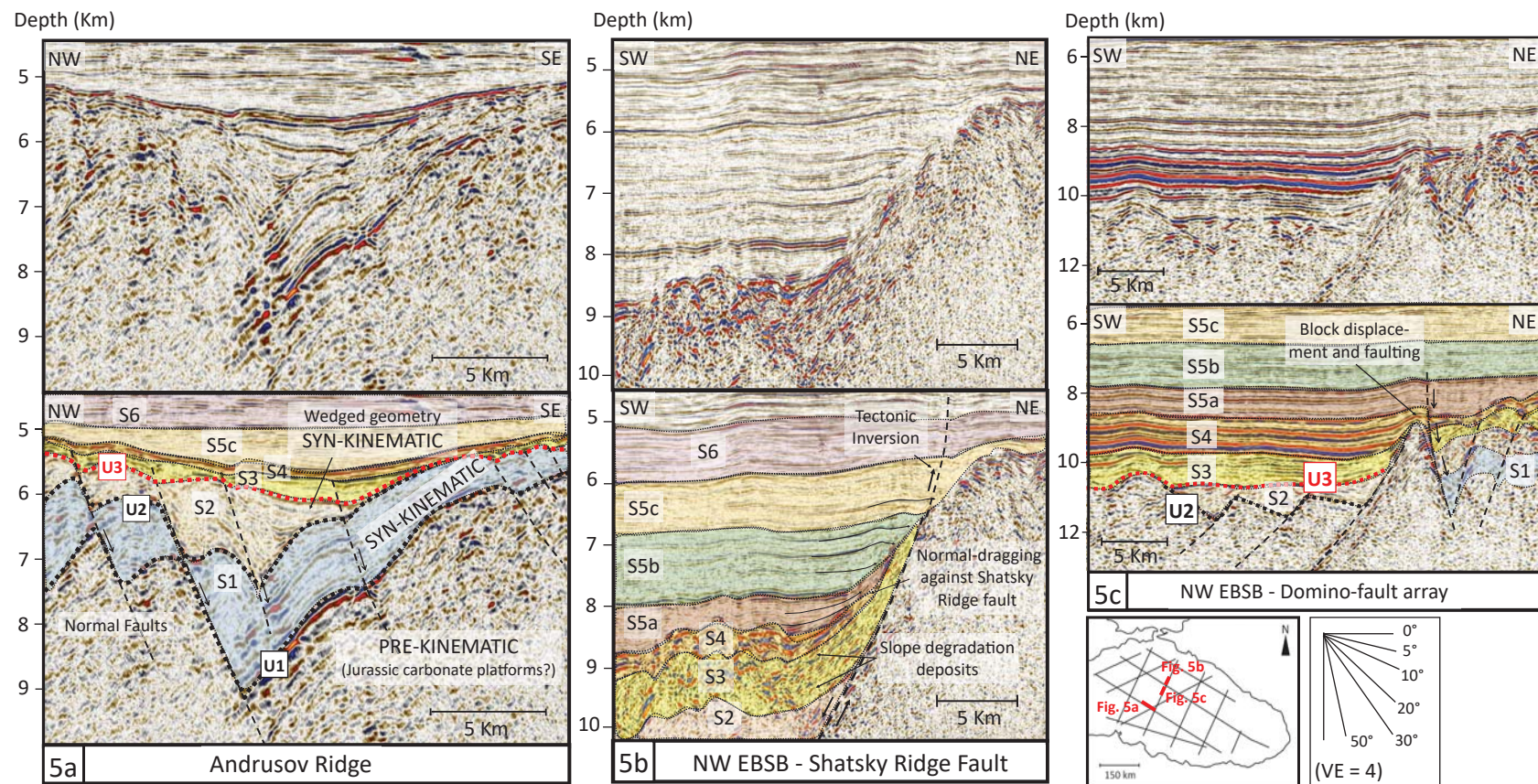


Figure 5.

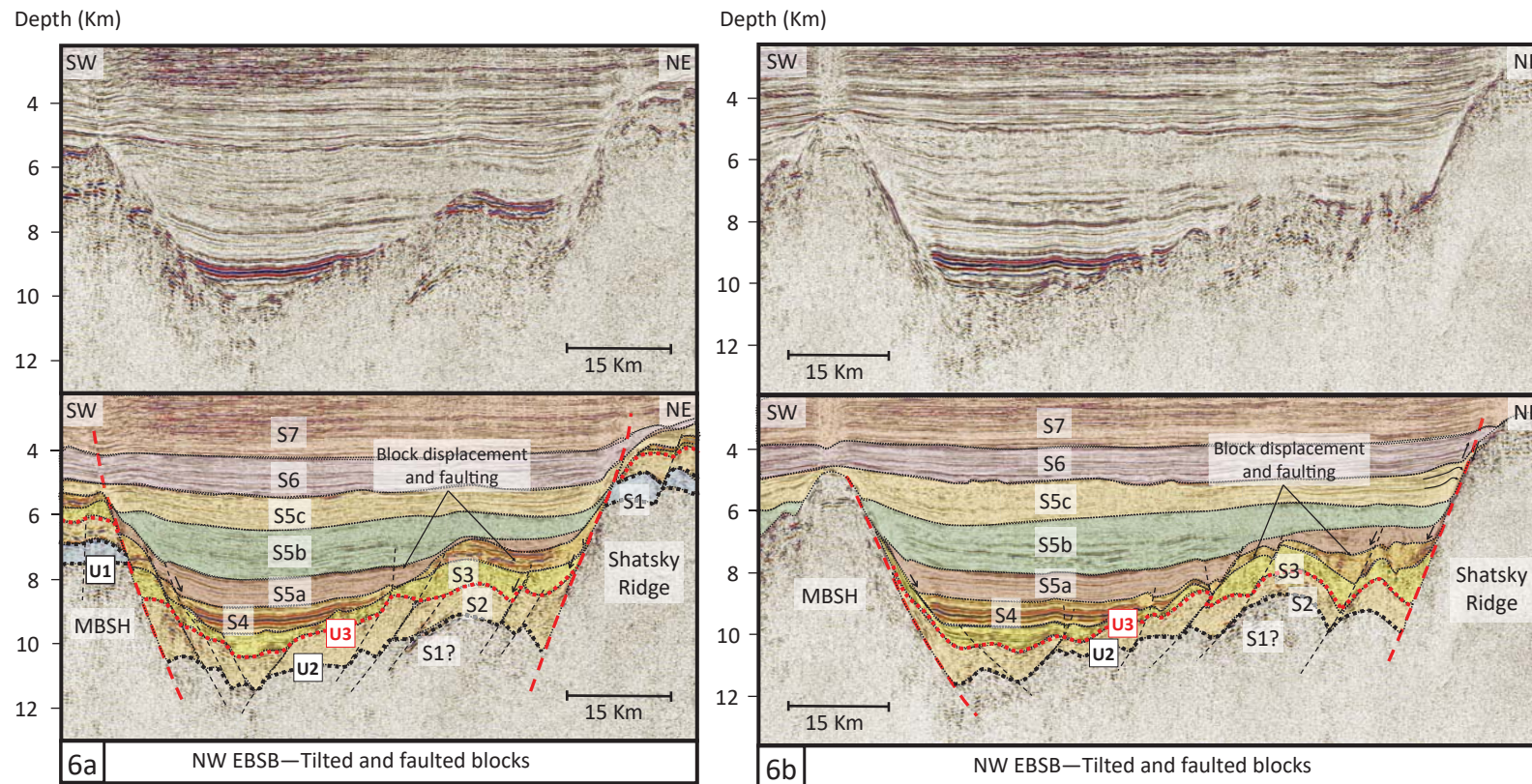


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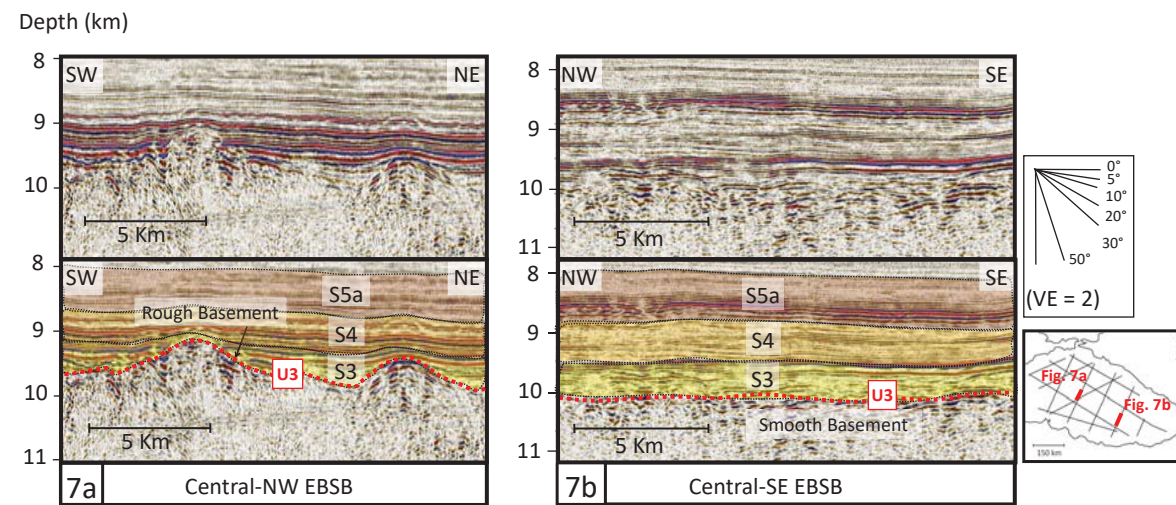


Figure 7.

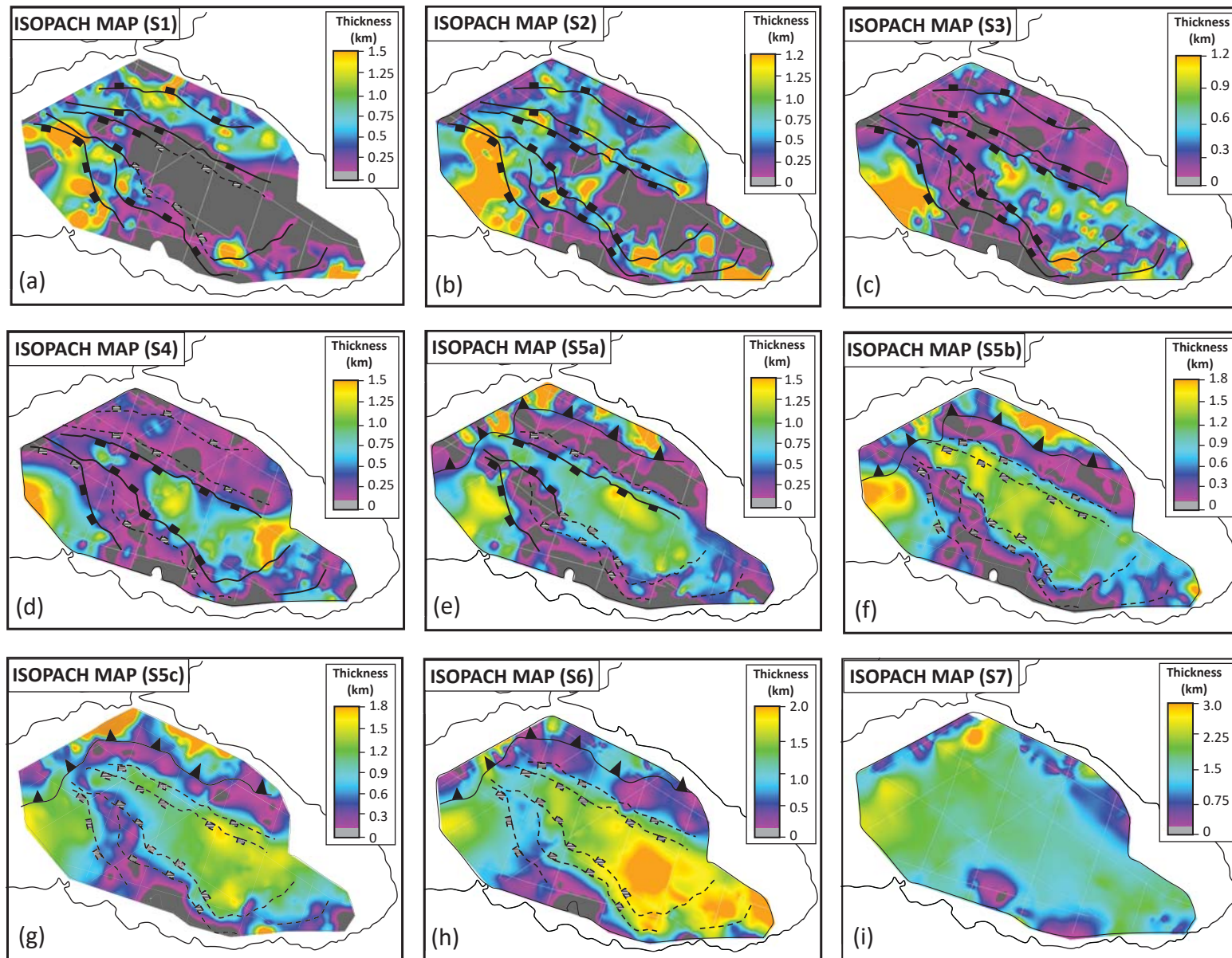


Figure 8.

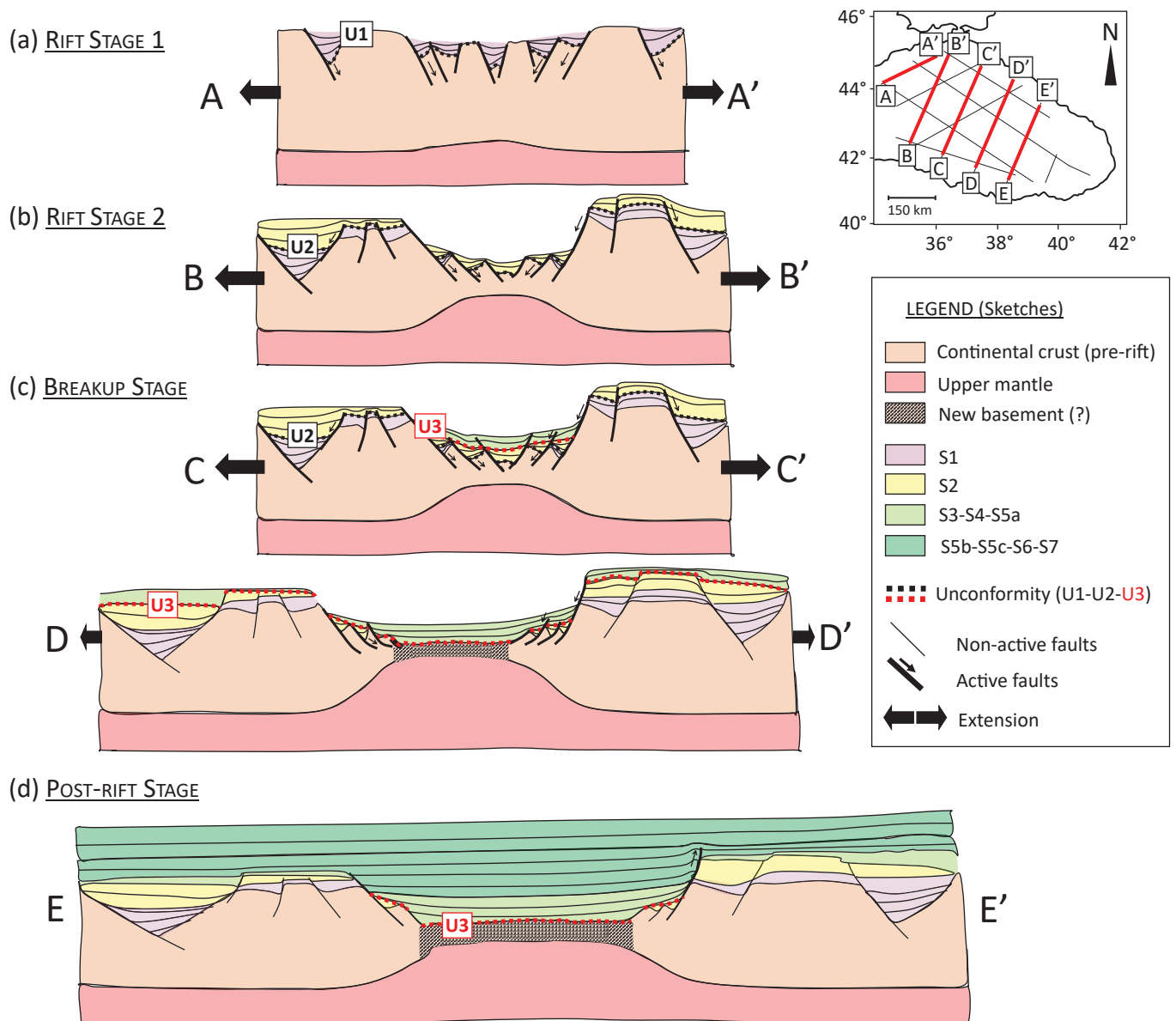


Figure 9.

Spatial and temporal evolution of rifting and continental breakup in the Eastern Black Sea Basin revealed by long-offset seismic reflection data.

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Figures S1 to S2

Introduction

This supporting material provides information on the time-depth conversion we applied to the GWL seismic dataset, and on the velocity model building used to generate a homogeneous velocity field to apply for the conversion.

Initially, we compared the available velocity information coming from GWL stacking velocities and wide-angle seismic data (Shillington et al., 2009) (Figure S1). The comparison shows that wide-angle seismic data have faster trends at depths below basement, whereas both velocity datasets are similar in the shallow section (Figure S1). Wide-angle velocities are considered more reliable at depth, so these data have been used to estimate a mean velocity trend below basement. Although shallow velocity trends are similar in both stacking and wide-angle velocities, stacking velocities have a better coverage across the study region, and they also show a more detailed and reliable match between velocity changes and geological variations. Therefore, stacking velocities are used to estimate a mean velocity trend above basement. The velocity model building was generated using Petrel software. The model is constrained by four time-interpreted seismic horizons (Water Bottom, Top Maykop, Base Maykop and Top Basement), and the mean velocities (V_0) estimated along these horizons from the combined stacking and wide-angle velocity fields. These horizons define a five-layer velocity model (Figure S2). A velocity gradient (k) has been introduced in the model generation, to allow V_0 for each layer to increase in a linear function (Figure S2). The resulting velocity model has been used to time-depth convert seismic profiles and interpretation.

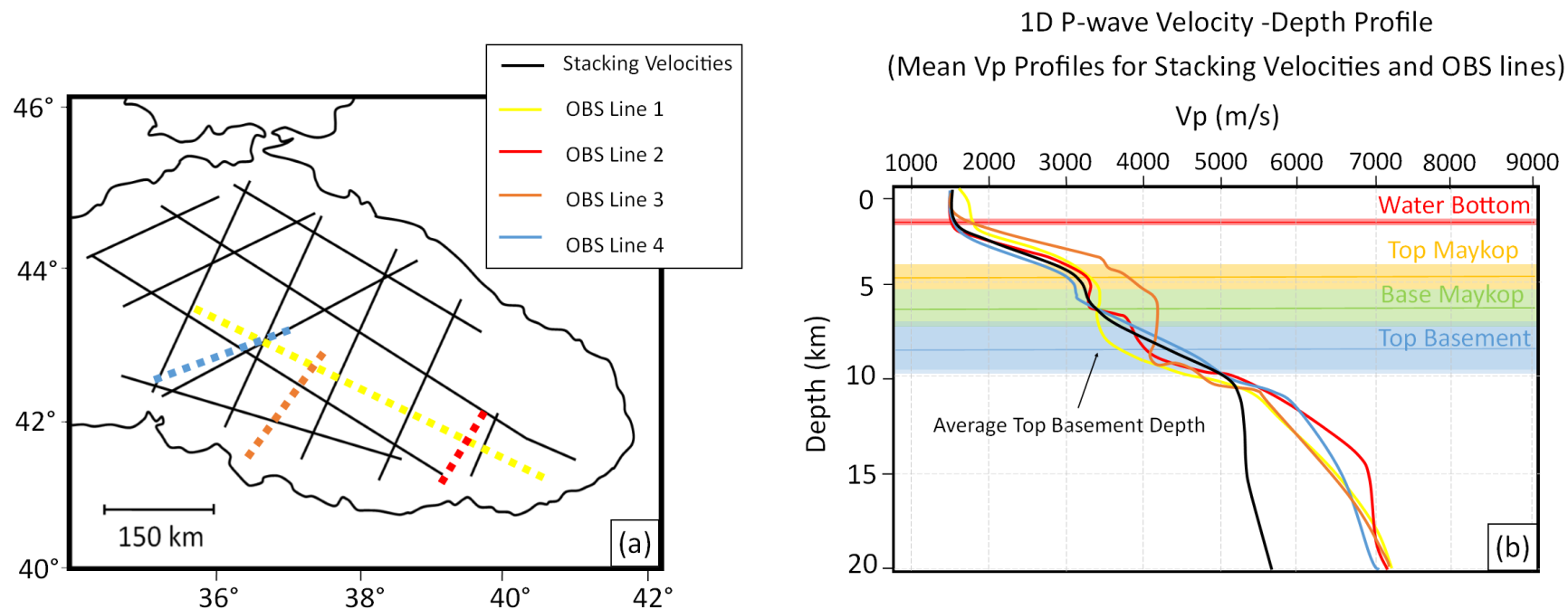


Figure S1. (a) Survey area map showing the location of the stack profiles used in this study and the OBS lines (Shillington et al., 2009). (b) 1D velocity-depth profiles comparing stacking velocities and OBS profiles. Black line— Mean velocity profile derived from stacking velocities. Coloured lines—Mean velocity profile derived from OBS lines (colour code for each line is shown in (a)). Mean depth ranges for the interpreted horizons and their standard deviations are highlighted in the coloured bands: Red - Water Bottom, Yellow - Top Maykop, Green - Base Maykop and Blue - Top Basement (see also Mean Depth Horizon (\pm error) column in Figure S2b). OBS velocities are faster below the basement compared to stacking velocities.

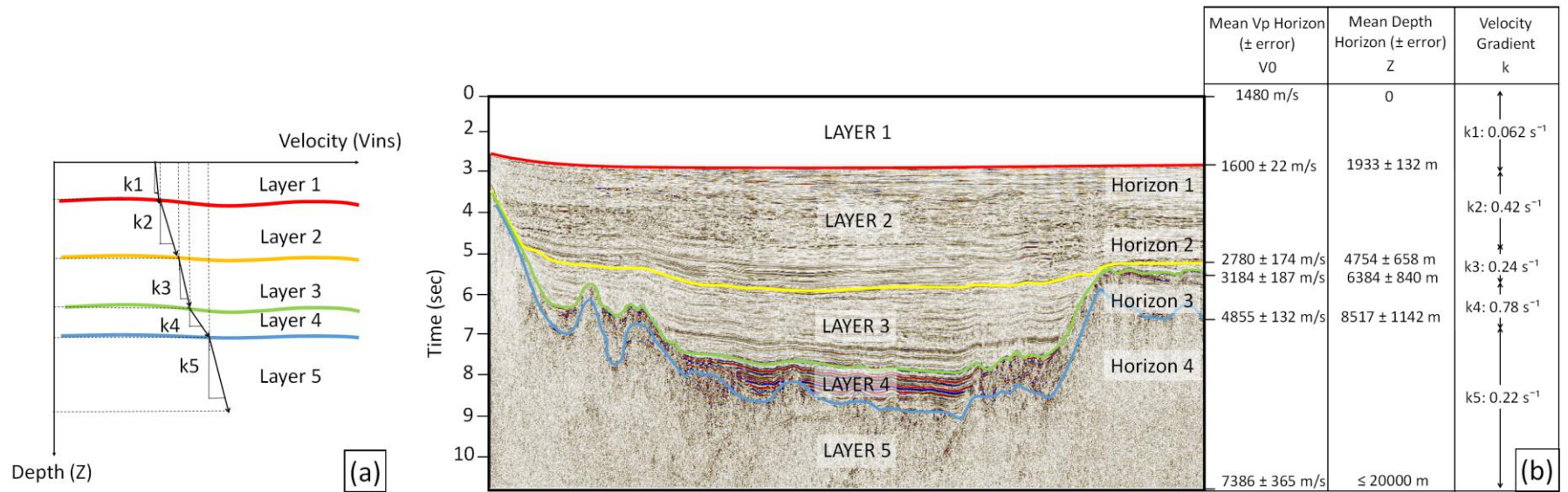


Figure S2. (a) Velocity model built in Petrel using an instantaneous linear function: $V_{inst} = V_0 + k(Z - Z_0)$ (Slotnick, 1936). Within a layer, this function assumes that velocity varies linearly with depth. Z and Z_0 are defined by time-picked horizons and their surfaces. For each layer, k is the velocity gradient (s^{-1}); V_0 is the velocity at the top of the layer, and $(Z - Z_0)$ is the increment in thickness of the layer with depth. (b) Seismic example from line bs-go. The horizons defining the velocity model layers are shown in overlay. On the right, table summarising values of V_0 , Z and k used for the modelling. The horizons used to estimate the V_0 values are: Horizon 1 - Water Bottom, Horizon 2 - Top Maykop, Horizon 3 - Base Maykop, Horizon 4 - Top Basement. These horizons also correspond to major velocity changes (Figure S1b). V_0 is calculated by lateral stacking of velocity information along each interpreted horizon ($\pm 200 \text{ m}$ from that horizon).