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Review of the main Black Sea rifting phase in the Cretaceous and implications for the evolution of the Black Sea lithosphere

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ABSTRACT

The Black Sea is a deep marine basin formed by lithosphere extension and active rifting in a back-arc tectonic setting, by general consensus, in the Cretaceous. Its present structural architecture, however, is mainly defined by compressional tectonics during the Cenozoic when large scale "basin inversion" reactivated extensional fault systems formed in the Cretaceous. Rifting during the Cretaceous is usually taken to represent the main process forming the present-day basin (that is, producing crustal thinning and concomitant subsidence prior to its modification during Cenozoic inversion). Rifting at this time took place within continental lithosphere that had been accreted to and, by the Cretaceous, formed part of the Eurasian lithospheric plate. The precise history of how and when pre-Cretaceous aged tectonic domains were accreted to Eurasia forming the continental lithosphere underlying the Black Sea is poorly known. A critical issue to the tectono-thermal evolution of the Black Sea basin with important implications for paleogeography and sedimentary depositional environments is the degree of crust (and lithosphere) thinning during Cretaceous rifting and whether oceanic or "sub-oceanic" crust was formed at that time. The main focus of this paper, in order to illuminate this issue, is on kinematic observations related to the Cretaceous (Albian-Cenomanian) rifting phase, including subsidence analysis, as well as the immediate post-rift sedimentation and stratigraphy. The results suggest that rifting during the Cretaceous was insufficient in its own right to reveal exhumed mantle or to promote ocean crust formation beneath the deep basins of the Black Sea. It is concluded that an important contribution to observed present-day crustal and lithosphere architecture of the Black Sea area are legacy extensional tectonic events affecting the area in pre-Cretaceous times, with implications for the Late Palaeozoic-Mesozoic paleogeography and paleotectonic evolution of this area.

1. Introduction

The Black Sea is a deep marine basin on the northern periphery of the Alpine-Tethys convergence belt at the southern margin of Eurasia (Fig. 1). It formed by lithosphere extension and active rifting in a back-arc tectonic setting (Neprochnov et al., 1970; Letouzey et al., 1977; Zonenshain and Le Pichon, 1986; Görür, 1988; Finetti et al., 1988; Okay et al., 1994, 2018; Robinson et al., 1996; Spadini et al., 1997; Nikishin et al., 2015a,b; Starostenko et al. 2004; Stephenson and Schellart, 2010; Shillington et al., 2009, 2017; Scott et al. 2009; Graham et al., 2013; Okay and Nikishin, 2015; Tari et al., 2015; Sosson et al., 2016; Munteanu et al., 2011, 2018; Monteleone et al., 2019). The present structural architecture of the Black Sea, however, is mainly defined by Alpine compressional tectonics during the Cenozoic when large scale "basin inversion" reactivated extensional fault systems formed during the rifting stage. Most of the structures formed during the rifting and subsequent inversion stages are buried by a later (Plio-Pleistocene) influx of a huge volume of sediments, such that the present-day bathymetric architecture of the Black Sea basin is one of a largely undeformed, continuous flat depositional surface at a depth of about 2000 m in its deepest part (e.g., Tugolesov et al., 1985; Finetti et al., 1988; cf. isobaths in Fig. 1).

Although the physiography of the present-day Black Sea is expressed as a single deep marine basin, the architecture of the Cretaceous and younger sedimentary successions therein expresses two distinct segments, western and eastern, each with its own depocentre where the sedimentary succession is thicker than elsewhere. These areas are delineated in Fig. 1 as the WBSB (West Black Sea Basin) and EBSB (East

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Black Sea Basin) that are usually considered independent, central, rift segments during the initial rifting stage (e.g., Stephenson and Schellart, 2010).

The initial rifting stage in the WBSB is generally considered to have taken place in the Early (to early Late) Cretaceous (e.g., Görür, 1988; Finetti et al., 1988; Okay et al., 1994, 2018; Robinson et al., 1996; Khriachtchevskaia et al., 2010; Hippolyte et al., 2018; Nikishin et al., 2012, 2015a, 2015b; Stovba et al., 2013, 2020; Tari et al., 2015) although Munteanu et al. (2011, 2018) considered it to be middle Cretaceous until the Eocene. Rifting in the EBSB took place simultaneously with the WBSB, hence Early (to early Late) Cretaceous, according to some authors (Stephenson and Schellart, 2010; Nikishin et al., 2012, 2015a, 2015b; Stovba et al., 2013, 2017a, 2017b, 2020) or later than in the WBSB, namely, from the Palaeocene until the Early Eocene according to others (Robinson et al., 1996; Shillington et al., 2009, 2017; Hippolyte et al., 2015) or Oligocene (Monteleone et al., 2019).

Cretaceous (and possibly Palaeogene) rifting is usually taken to represent the main process forming the present-day basin (that is, producing crustal thinning and concomitant subsidence prior to its modification during Cenozoic inversion). Rifting at this time took place within continental lithosphere that had been accreted to and, long

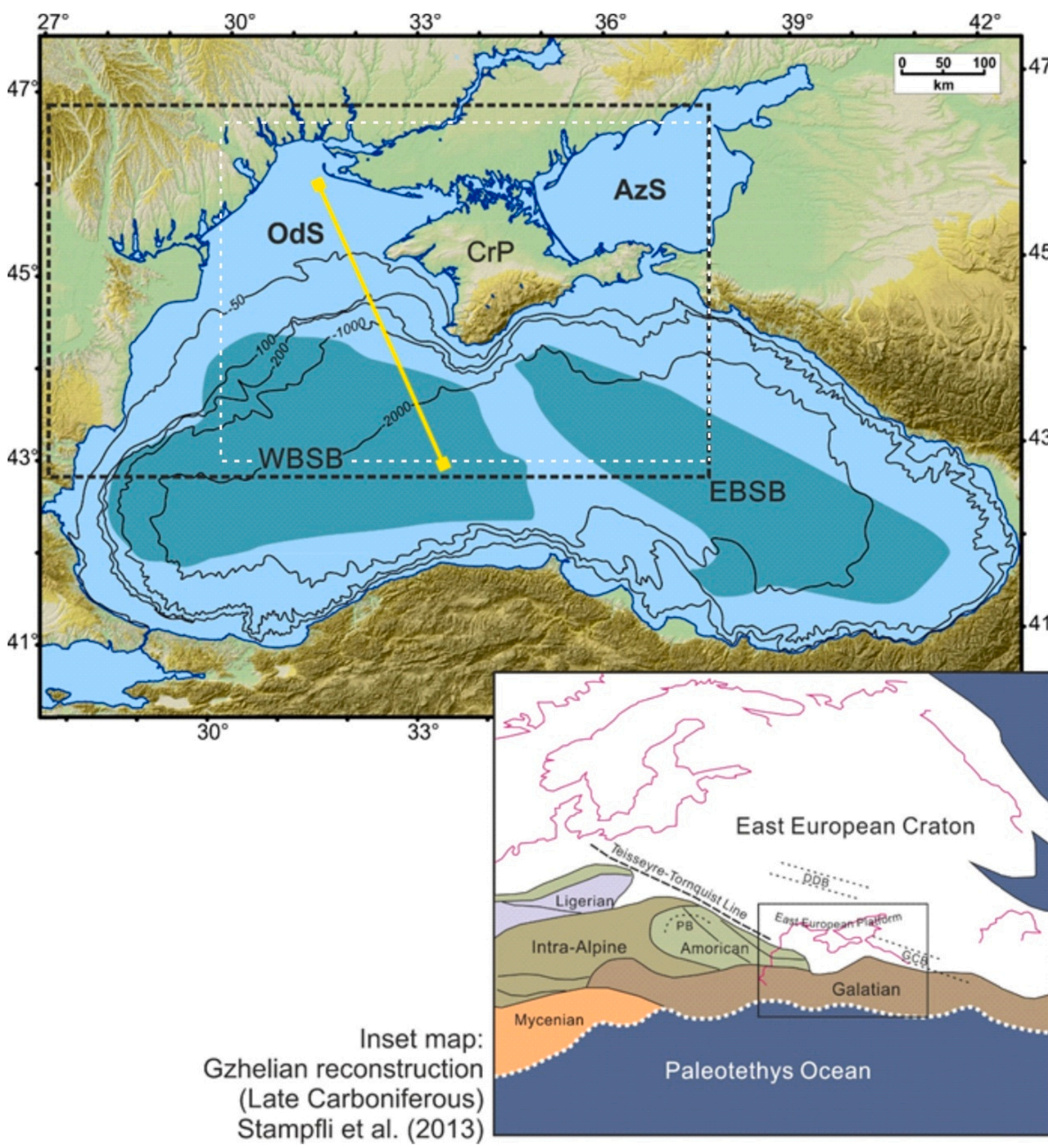

Fig. 1. The present-day Black Sea, showing bathymetry (contours) and outlines of the West Black Sea Basin (WBSB) and East Black Sea Basin (EBSB), with deeper marine blue-green colour, as commonly delineated for the Black Sea (e.g., Okay and Tüysüz, 1999; Stephenson and Schellart, 2010). The black dashed rectangle indicates the location of the data distribution map shown in Fig. 2 and the white dashed rectangle inside it indicates the coverage of maps shown in Figs. 4 and 5. The yellow line is the schematic location of the crustal-lithosphere cross-section shown in Fig. 6. The inset map shows the Black Sea in the context of a paleotectonic reconstruction of the southern Tethys margin of Europe in the Late Carboniferous (Gzhelian), simplified and slightly modified from Stampfli et al. (2013) The black box indicates the implied extent of the main map. The coloured and labelled units are the tectonic elements of the Galatia Superterrane (as named and according to Stampfli et al.; cf. von Raumer et al., 2012) that were amalgamated during the Variscan Orogeny prior to their accretion to the European margin in the Carboniferous. The Precambrian cratonic lithosphere of Europe lies to the east of the Teisseyre-Tornquist Line; Variscan terranes and earlier accreted Caledonian terranes lie to its west. Minor dashed lines with labels on the inset map indicate the approximate locations of basins mentioned in Section 6: DDB – Dniepr-Donets Basin (late Palaeozoic); GCB – Greater Caucasus Basin (Jurassic); PB – Pannonian Basin (Cenozoic).

before the Cretaceous, formed part of the Eurasian lithospheric plate (e.g., Saintot et al., 2006a; Okay and Nikishin, 2015; Sosson et al., 2016). The precise history of how and when pre-Cretaceous aged tectonic domains were accreted to Eurasia forming the continental lithosphere underlying the Black Sea, however, is poorly known. The inset map on Fig. 1 shows the present-day geographic limits of the Black Sea according to the Late Carboniferous paleotectonic reconstruction of Stampfli et al. (2013).

One critical issue that remains, however, one that has not been the subject of a great deal of debate in the published literature, is the degree of crust (and lithosphere) thinning that occurred solely as a result of Cretaceous(-Palaeogene?) rifting and, accordingly, the nature of what forms the basement to the subsequently deposited, thick Cretaceous-Cenozoic sedimentary pile that now lies within the WBSB and EBSB. This is clearly of importance from a tectono-thermal evolution point of view and also has critical implications for paleogeography and sedimentary depositional environments because of the direct link between intensity of rifting and syn-rift subsidence, crustal thickness and water depth.

It has generally been considered that crustal thinning during the main Black Sea rifting phase was responsible for the evidently thin crust underlying the deep basins (e.g., Starostenko et al., 2004; these and other results will be described in greater detail in Section 3.1, below). This has often been interpreted as crust with "oceanic" or "sub-oceanic" affinity (e.g., Neprochnov et al., 1970; Nikishin et al., 2015a,b) and taken as evidence for assuming the sedimentary environment during deposition of Cretaceous strata was one of a deep marine basin. To illuminate this issue, the main focus of this paper will be on kinematic observations related to an inferred Cretaceous (Albian-Cenomanian) rifting phase, as well as the immediate post-rift sedimentation and stratigraphy.

The main study target of the present work is the western segment of the Black Sea – specifically the Odesa Shelf and the adjacent deep-water WBSB – in part because of the availability of a comprehensive dataset with a full set of densely-spaced seismic reflection profiles integrated with borehole and dredge data and fully correlated with geological observations onshore the Crimea Peninsula. The stratigraphic record of the WBSB is more complete than in the EBSB and, importantly, there has been less disturbance by subsequent post-rift compressional deformation and basin inversion than in the EBSB (Stovba et al., 2020). The results of the study, nevertheless, may have general implications for the tectonic history of the initiation and evolution of the whole of the Black Sea and both its constituent sedimentary basins, including the role played by pre-rift regional history of the Black Sea lithosphere (e.g., Stovba and Stephenson, 2019).

2. Seismic interpretations and syn- and post-rift geometry of the Odesa Shelf/western Black Sea Basin

2.1. Data and methodology

Stovba et al. (2017a, 2017b) analysed some thirty thousand kilometres of offshore seismic profiles, constrained by well data from some forty marine boreholes as well as samples dredged on the offshore continental slope. This was further integrated with the onshore geology, which in part was remapped and recorrelated (Popadyuk et al., 2013a, 2013b; Stovba et al., 2017b), and interpreted to produce an atlas of structure maps and isopach maps for ten key horizons and sedimentary units in the study area and, from these to produce a set of simplified paleogeographic and paleotectonic maps (Stovba et al., 2020). These cover the present study area seen in Fig. 2, the offshore limit of which is based on the border of the Ukrainian segment of the Black Sea as established prior to its relocation in 2009 by UNCLOS (United Nations Convention on the Law of the Sea). Two key interpreted N-S regional

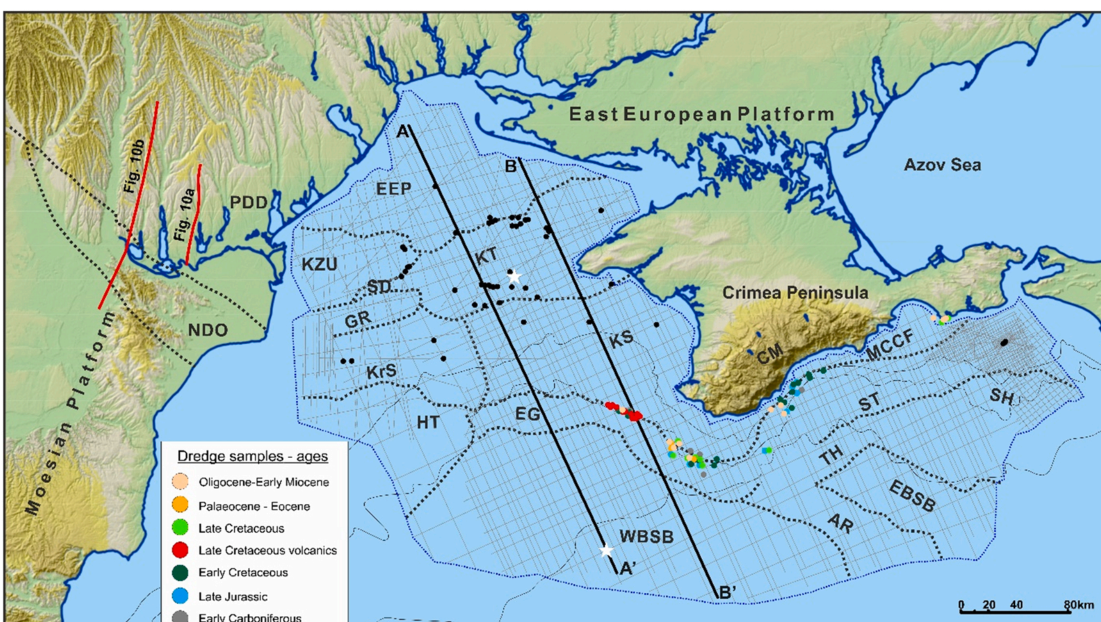

Fig. 2. The study area, which comprises about one-quarter of the area of the Black Sea, is enclosed by a light blue line; the exposed geology on the Crimea Peninsula is incorporated into interpretations of the offshore geology. Grey lines indicate the grid of 2-D seismic profiling used in the study; wells providing borehole data used in the study are indicated by black dots. The locations of bottom dredged rock samples used in the study are indicated by coloured dots and their ages indicated in the legend adopted from Stovba et al. (2020). The white stars represent approximate locations where "pseudo-wells" for subsidence analysis presented in this paper were compiled; the KT pseudo-well location coincides with the actual Archangelska-1 well location. Interpreted seismic profiles A-A′ and B-B′ are shown in Fig. 3. Seismo-geological cross-sections presented in Fig. 10 are located in the PDD profiles in red with black labels. Bathymetric contours are as in Fig. 1. Tectonic elements: AR – Andrusov Ridge; CM – Crimea Mountains; EBSB – East Black Sea Basin; EEP – East European Platform (offshore); EG – Euxinian Graben; GR – Gubkin Ridge; HT – Histria Trough; KrS – Krayova Step; KS – Kalamit Swell; KT – Karkinit Trough; KZU – Krylov-Zemiiny Uplift; MCCF – Marine Continuation of Crimean Folds; NDO – North Dobrogea "Orogen"; PDD – Pre-Dobrogea Depression; SD – Sulina Depression; SH – Shatskiy High; ST – Sorokin Trough; TH – Tetyaev High; WBSB – West Black Sea Basin.

seismic profiles located on Fig. 2 (A-A′ and B-B′) and shown in Fig. 3 provide examples from the integrated dataset. Additional details and documentation of the seismic interpretations and stratigraphic correlations controlling these are available in Stovba et al. (2020).

The published seismic interpretations used in the present work, like those in Fig. 3, are comparable to other extant interpretations such as those of the excellent Geology Without Limits (GWL) dataset (Nikishin et al., 2015a, 2015b) except that the present data allow more detailed analysis in places. Profiles A-A′ and B-B′ in Fig. 3 are almost coincident with the northern segments of profiles BS-50 and BS-60 interpreted in Nikishin et al. (2015a). Though the interpretations are largely compatible there are important differences. For example, the Stovba et al. (2020) dataset allows recognition of the syn-rift succession underlying the Upper Cretaceous post-rift succession, which was not identifiable in the GWL dataset interpreted by Nikishin et al. (2015a). In turn, this has allowed delineation of the structure of the important syn-rift Euxinian Graben between the Odesa Shelf and the deeper WBSB (Figs. 2 and 3).

Stovba et al. (2020) have documented a structural architecture for the study area and its tectonic history that is essentially in keeping with inferences made by others using less discriminating datasets (e.g., Tugolesov et al., 1985; Finetti et al., 1988; Nikishin et al., 2015a, 2015b). The Stovba et al. (2020) datasets permit a greater degree of precision on the timing of the tectonic phases (at least within the study area, though the regional nature of the tectonic processes involved would suggest that such events would be more or less synchronous throughout the contiguous parts of the Black Sea). It was demonstrated that the tectonic history of the Odesa Shelf/West Black Sea Basin in the present study area comprises three distinct tectonic phases: (i) active crustal extension and rifting from the late Early Cretaceous (Albian) to the early Late Cretaceous (Cenomanian) followed by (ii) the passive post-rift (thermal) subsidence phase until the middle Eocene and, thereafter, (iii) the continuation of post-rift subsidence interrupted by a series of four, short-lived, compressional deformation (inversion) events at the ends of the middle and late Eocene and at the ends of the early and late Miocene. The most significant of these short-lived, compressional deformation events are the first and fourth, at the end of the middle Eocene and at the end of the late Miocene (cf. Khriachtchevskaia et al., 2007, 2010; Stovba et al., 2017a, 2017b, 2020). Classic basin inversion structures are readily observed on graben-bounding faults formed initially during the syn-rift phase. There are also fold-like structures visible in the otherwise little deformed deep WBSB, for example, as indicated by the downwards pointing arrow in Fig. 3(b). This structure is of Late Miocene age but suggests pre-rifting basement involvement.

The primary focus of the present work is the main syn-rift period and its tectonic expression, including its intimate relationship with the immediately subsequent and genetically linked post-rift "sag" basin. Both these successions and their structural architecture are strongly modified during the inversional tectonic phase that begins in the middle Eocene. What follows is a description of the sedimentary successions and their depositional settings formed during the syn- and post-rift phases of Cretaceous Black Sea formation outlining the key observations relevant to the question of depositional environment at the end of Cretaceous rifting in the Black Sea.

2.2. Main Black Sea rifting phase in the Cretaceous

The study area was one of marine deposition during the main Cretaceous rifting phase with minor exceptions where the footwalls of some active faults were uplifted above sea level (Stovba et al., 2020). The presence of volcanic strata of some hundreds of metres thickness within the syn-rift succession has been inferred from earlier seismic interpretations (Finetti et al., 1988; Nikishin et al., 2015a,b).

The maps in Fig. 4 show offshore structure at the base of the syn-rift sedimentary succession and its thickness (syn-rift isopach). The faults displayed on Fig. 4 were mapped from the seismic dataset offshore and adapted from Stovba et al. (2017a, 2017b) onshore the Crimea Peninsula. A widely distributed system of grabens and half-grabens that formed during this time is mappable throughout the study area. The vertical offsets on the bounding faults ranged from several tens of metres to more than 2–3 km (e.g., Fig. 3). The footwalls of some of these faults emerged above sea level and were subject to erosion resulting in an absence of syn-rift strata in places (blue outlined white areas in Fig. 4). The whole of the study area was undergoing tectonic extension and concomitant tectonic subsidence synchronously (from a geological-tectonic point of view) during this period.

Faults formed and mapped on what is now the Crimea Peninsula have strikes and affinity that are compatible those in the contiguous Odesa Shelf to the west, bounding structures such as the Gubkin Ridge-

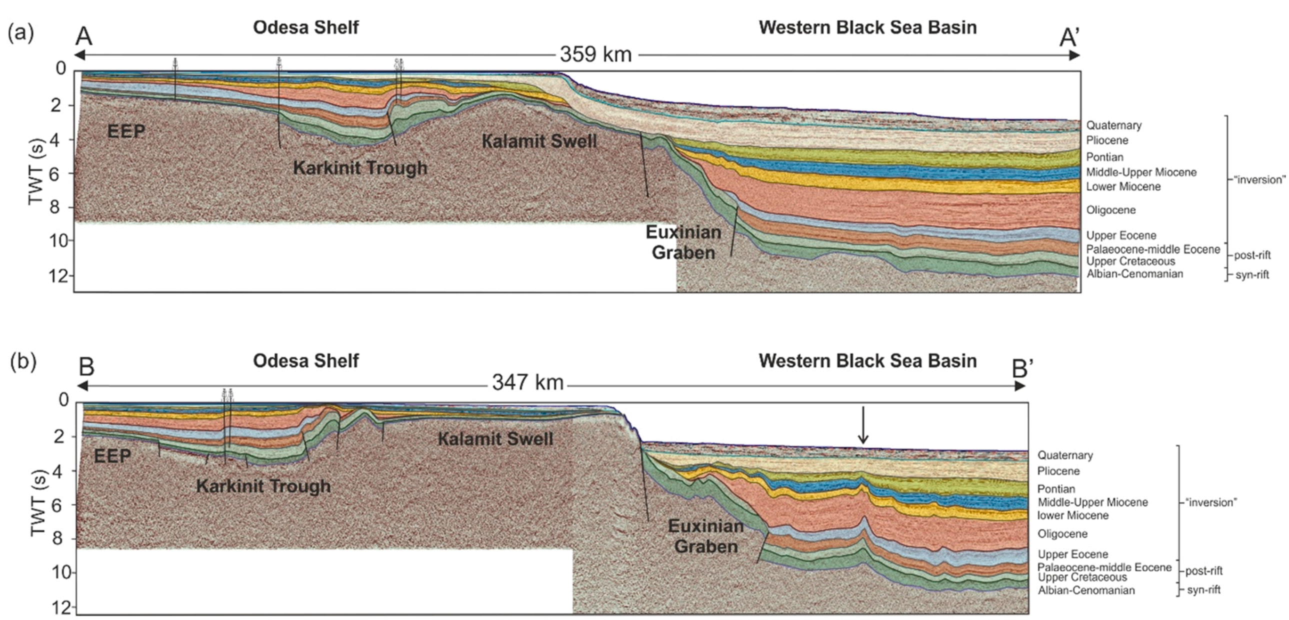

Fig. 3. – Interpreted regional seismic profiles (a) A-A′ and (b) B-B′ (located in Fig. 2). The downwards pointing arrow in panel (b) indicates the location of the deeply seated Miocene-aged structure mentioned in the text. Ages of interpreted units and the three "tectonic" phases discussed in the text are indicated to the right of each panel. EEP – East European Platform.

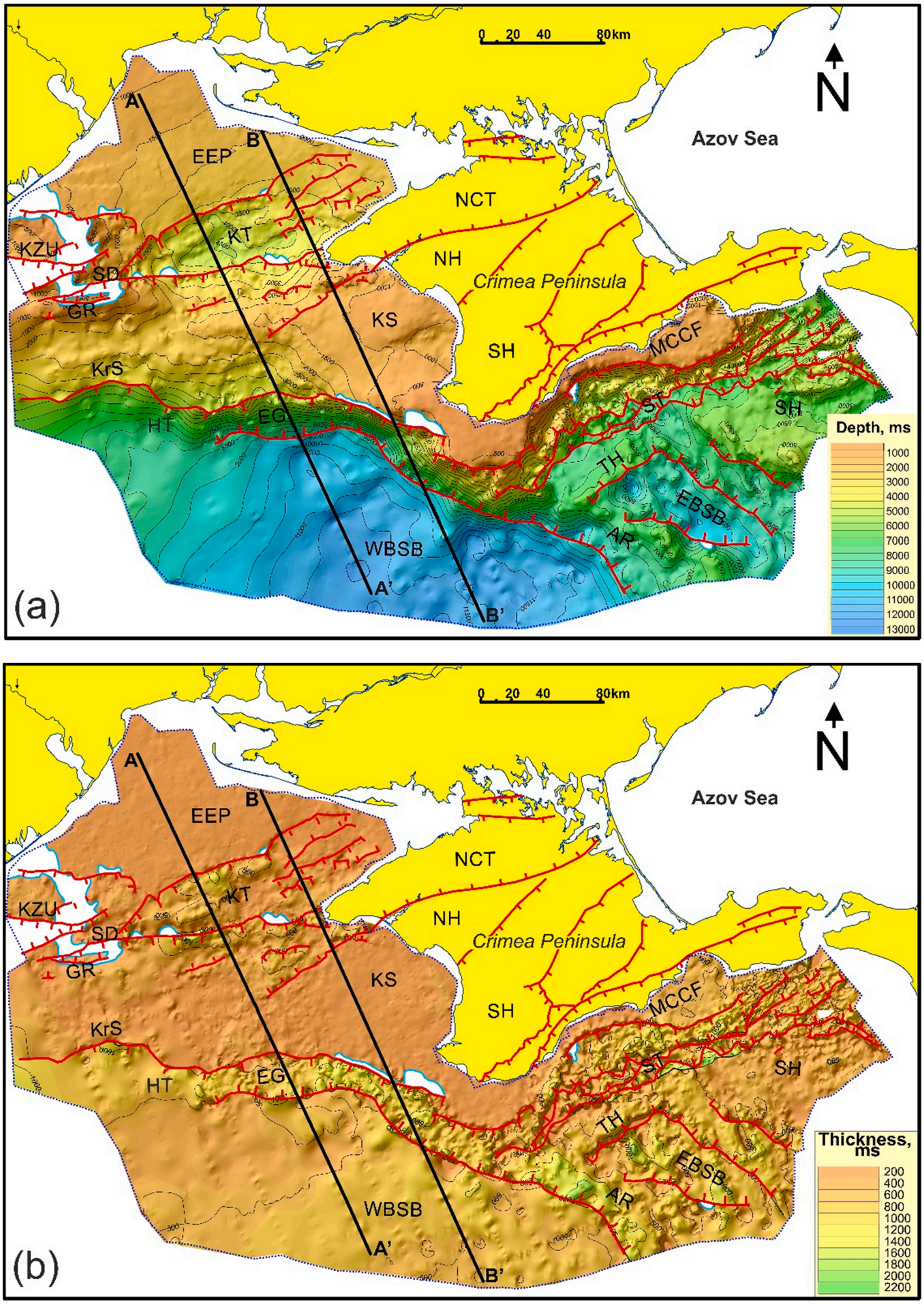

Fig. 4. (a) TWT structure map at the base of the Cretaceous syn-rift sediments and (b) TWT thickness (isopach) map of the syn-rift sediments for the study area. Ornaments on rift faults (ornamented red lines) refer to their original planes; some of these were severely inverted during later compressional events and now present as reverse faults even at the base of the syn-rift sequence. Blue-outlined white zones offshore represent areas of no deposition except in the northern part of the Sorokin Trough (ST) and on the margin of the Andrusov Ridge (AR) where they correspond to areas of complete erosion of the syn-rift sediments. Faults on the Crimea Peninsula are from Stovba et al. (2017a, 2017b). Note that these maps refer only to the preserved syn-rift sequence offshore and do not incorporate wide occurrences onshore on the southern Crimea Peninsula. A-A′ and B-B′ are the seismic profile seen in Fig. 3. Tectonic elements are as in Fig. 2, with the following onshore features added: NCT – North Crimean Trough; NH – Novosalivka High; SH – Simferopol High. (For interpretation of the references to colour in this figure, the reader is referred to the web version of this article.)

Karkinit Trough-North Crimea Trough and Kalamit Swell-Novosalivka High-Simferopol High (GR-KT-NCT and KS-NH-SH; Fig. 4) as well as in the Crimea Mountains (CM; Fig. 2) and tectonic elements in its contiguous offshore to the south, including the "Marine Continuation of Crimean Folds", Sorokin Trough and Tetyaev High (MCCF, ST and TH; Fig. 4). These faults trend roughly E-W to ENE-WSW. The southernmost structure in the study area west of Crimea is the southern bounding fault of the Euxinian Graben (EG; Fig. 4), the trend of which changes

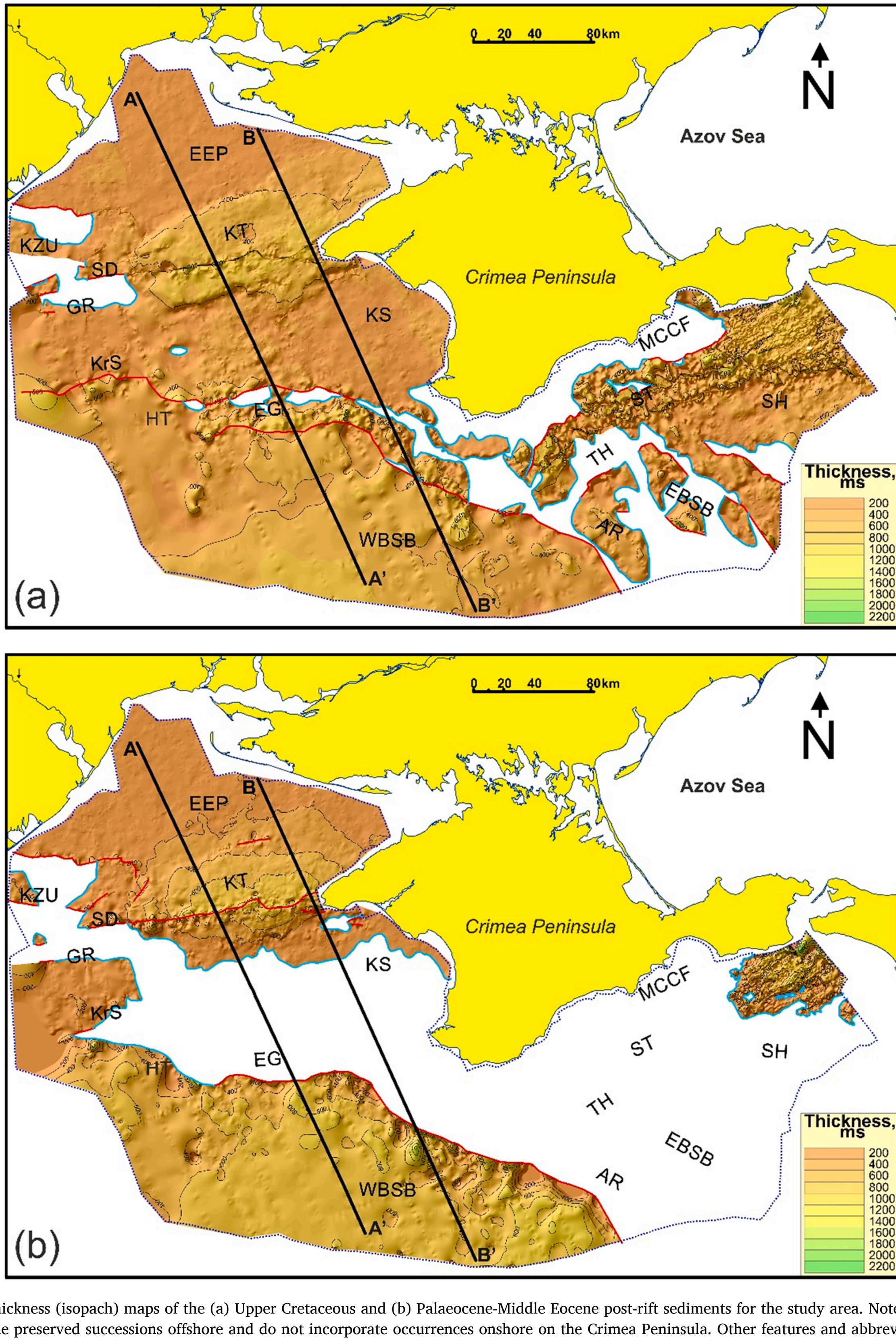

Fig. 5. TWT thickness (isopach) maps of the (a) Upper Cretaceous and (b) Palaeocene-Middle Eocene post-rift sediments for the study area. Note that these maps refer only to the preserved successions offshore and do not incorporate occurrences onshore on the Crimea Peninsula. Other features and abbreviations are as for Figs. 2 and 4.

eastwards from E-W bordering the Odesa Shelf to more NW-SE becoming sub-parallel to the trends of the system of normal faults associated with the eastern limit of the West Black Sea Basin (WBSB; Fig. 4), the Andrusov Ridge, East Black Sea Basin (AR and EBSB; Fig. 4) and the Shatskiy High (SH; Fig. 4), together forming one extensional system during Cretaceous rifting. The complex zone south of Crimea where the structural trends associated with Cretaceous rifting change orientation may be related to changes in inherited basement trends. The present-day base of the syn-rift sequence is deepest in the central part of the WBSB, where it lies at about 12 s TWT (Fig. 3a), which Stovba et al. (2020) estimated to be 15.5–16.5 km. In the EBSB its maximum depth is 12.5–13.5 km (~10 s; Stovba et al., 2020).

Many if not all of the rift-related faults displayed in Fig. 4 were subsequently inverted during the late Eocene and younger inversion events, accompanied by folding of strata within syn-rift grabens and half-grabens. Cumulative vertical displacements during inversion range from several hundred metres to as much as 4–5 km in many parts of the study area, as seen, for example, in the seismic cross-sections around the Karkinit Trough and Euxinian Graben (Fig. 3). Erosion on the hanging walls on some inverted faults has resulted in places of the partial or complete absence of syn-rift and younger strata at present (cf. Fig. 4). Rock samples dredged from the sea bottom along the flank of the Euxinian Graben and south of Crimea (Fig. 2) confirm the seismic interpretations. Tectonic elements such as the Andrusov Ridge and Shatskiy High only became the positive structures their names imply in the late Eocene. The most profound of these positive structures is the Crimean Mountains themselves. Remnants of the uplifted Cretaceous syn-rift sequence crop out widely in the Crimean Mountains (Popadyuk et al., 2013a; Stovba et al., 2013a).

2.3. Late Cretaceous-Eocene post-rift “sag” phase

Fig. 5 shows isopach maps of the preserved post-rift but pre-inversion sedimentary succession, separated into Upper Cretaceous (Fig. 5a) and Palaeocene-middle Eocene (Fig. 5b) sequences, according to the resolution of the seismic interpretations seen on profiles A-A′ and B-B′ in Fig. 3. Passive thermal (post-rift) subsidence occurred in the whole of the study area and sedimentation was characterised by a quiet depositional environment without any evidence of active tectonic processes affecting basin subsidence. The post-rift sequences covered the entire northern part of the Black Sea and the Crimea Peninsula. Areas of zero isopach thickness on Fig. 5 are the result of erosion during the subsequent late Eocene-Miocene period of inversion. Accordingly, the present boundaries of these sequences (Fig. 5) are predominantly erosive ones. In areas where these sediments have been preserved from erosion, in whole or in part, the corresponding seismic sequences, whether in the deep-water part of the Black Sea (WBSB) or in the shallower parts of the offshore, indicate sub-horizontal sedimentary bedding with no evidence of disconformities or angular unconformities or abrupt thickness changes. Specifically, the thickness and seismic appearance of the post-rift sequences in the axial zone of the Karkinit Trough on the Odesa Shelf are comparable with those in the WBSB (Figs. 3 and 5), with conformable bedding and uniform seismic character occurring over significant distances in the WBSB (Fig. 3). The maximum TWT to the base of the Upper Cretaceous post-rift sequence in the WBSB is about 11.5 s (Fig. 3a), some 14.5–15.5 km.

The whole Upper Cretaceous through middle Eocene post-rift succession has been preserved in the WBSB, which was not strongly deformed during the Cenozoic compressional phases. Elsewhere, however, the present-day architecture of the preserved post-rift succession was strongly affected by compressional tectonic processes. The previously basinal area that emerged to become sub-aerial during the end of the Middle Eocene compressional event (Fig. 5b) was thereafter reduced by erosion and ongoing subsidence to again become submerged below sea level. Structural relief formed in the study area at this time was transgressively overlapped by younger sub-parallel sedimentary strata at the margins of basins (Fig. 3).

Fig. 3 clearly shows that the bulk of the post-Early Cretaceous sedimentary succession in the Black Sea was deposited after the middle Eocene, during the period of its evolution that occurred in a generally (though punctuated) compressional tectonic environment. This is true for secondary depocentres like the Karkinit Trough as well as for the WBSB deep-water basin. The lithofacies character and details of depositional environments as well as tectonic subsidence mechanisms driving basin evolution (e.g., the competition between sediment supply and accommodation space supply; e.g., Schlager, 1993) after the middle Eocene is not the main focus of this paper and will not be addressed in any detail further. What is notable is that the seismic character and thicknesses of syn- and post-rift successions on the Odesa Shelf within in the Karkinit Trough and within the deep WBSB until the onset of compression in the middle Eocene are very similar, suggesting similar depositional environments. They do not suggest that the former is in shelf conditions and that the latter is in deep basin conditions at this time.

3. The crust and lithosphere of the Black Sea

3.1. Background and tectonic setting

There is little directly known about the lithosphere in which Black Sea rifting took place other than it was lithosphere of continental affinity and that it was lying in a back-arc setting with respect to the active convergent southern margin of Eurasia (e.g., Stephenson and Schellart, 2010). Certainly, the lithosphere below the East European Platform as well as the northernmost Black Sea, including what is now in part the Odesa Shelf, Crimean Peninsula and Azov Sea (OdS, CrP, AzS; cf. Fig. 1, inset), was already part of Eurasia by the Palaeozoic (Saintot et al., 2006a; cf. Okay and Nikishin, 2015). The latter is sometimes referred to as the “Scythian Plate” (though this term is actually derived from the geomorphological term Scythian Platform (Fig. 2), onto which Neoproterozoic and Palaeozoic platform sediments were deposited and its implied history as an independent “plate” fragment or “microplate” has never been clearly established (e.g., Saintot et al., 2006a).

Rifting leads to thinning of the crust and the continental lithosphere, potentially to the point of continental “break-up” with the subsequent onset of formation of a new ocean basin, floored by oceanic crust and lithosphere and bordered by new continental margins with thinned continental crust and lithosphere. Many authors have invoked the presence of oceanic crust beneath the sediments of the deep basins of the Black Sea on the basis of generally poorly constrained estimates of crustal thickness and that the Black Sea deep basins lay in a back-arc tectonic setting. This was then linked to the general view of back-arc basins in the geological literature, as well as according to some modern marine geophysical observations, that they represent small oceanic basins, which, by definition, would be underlain by newly formed oceanic lithosphere.

A recent example of this kind of possibly flawed logic is exposed by Monteleone et al. (2020), who argue for oceanic crust in the south-easternmost EBSB, their diagnostic criterion being the smoothness of the basement surface in that area imaged by seismic reflection data. However, this is opposite of the criterion used by Nikishin et al. (2015a) in the WBSB and elsewhere, in marginal basins where oceanic crust is established by the actual presence of a sea-floor spreading signature in potential field data such as the Canada Basin and the South China Sea. In these last two settings, the oceanic crust identified by sea-floor spreading anomalies is characterised by a rough and irregular top of basement horizon (indicative of the igneous, dyking and faulting processes that form oceanic crust) whereas it is transitional, highly extended continental crust or exhumed continental mantle basement that is characterised as “smooth” (e.g., Chian et al., 2016; Zhang et al., 2021).

Besides these kinds of first-order ambiguities related to crustal affinity in the Black Sea, very little attention has been paid to the nature of the lithosphere as a whole that underlies the Black Sea specifically and what the implications of this might be to its tectonic history and role of Cretaceous rifting in its formation. The present-day crustal structure in the Black Sea as well as models of the present-day lithosphere of the Black Sea are, however, highly pertinent for evaluating formation of rift basins like the Cretaceous Black Sea basin. This section provides a brief review of what has been published in this regard. The results presented below pertaining to Moho depth and depth to the lithosphere-asthenosphere boundary (LAB) for the Black Sea are summarised in Fig. 6 for a cross-section, the schematic location of which is seen in Fig. 1 (subparallel to and equidistant from the seismic profiles located in Fig. 2.

3.2. Crustal thickness/Moho depth

Numerous estimations of crustal thickness and/or Moho depth beneath the Black Sea and its margins have been published in the last decades, many based primarily on gravity data, including the most recent by Bilim et al. (2021), although these authors seem to have neglected the gravity effects of the thick sedimentary layer above crystalline crust in the Black Sea. The Moho depth map of Starostenko et al. (2004) was mainly derived from gravity data but was calibrated by a considerable number of controlled-source seismic surveys and took into account "unconsolidated", "semiconsolidated" and "consolidated" sedimentary layers as well as an inferred mantle contribution to the gravity field. It shows Moho depths beneath the Odesa Shelf and the WBSB to be > 40 km and < 20 km, respectively. These are generally greater than and less than, respectively, those estimated by earlier authors writing in the Russian language and based on a tectonic integration

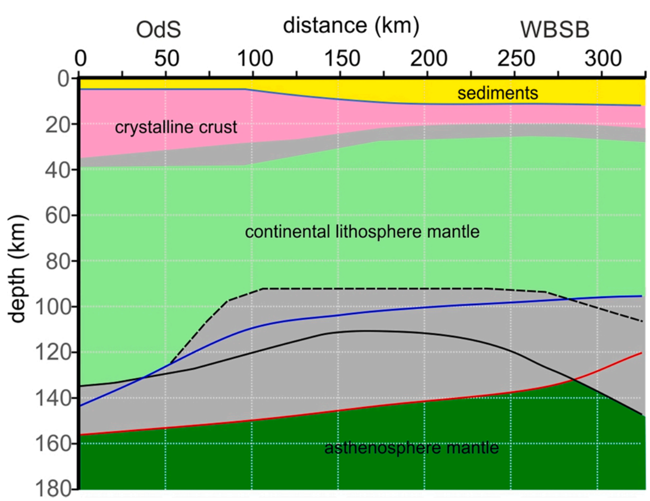

Fig. 6. Summary of geophysical data mentioned in the text and constraining the disposition of the Moho (base of the crust) and the LAB (lithosphere-asthenosphere boundary) on a N-S cross-section of the WBSB parallel and adjacent to seismic lines shown in Fig. 3. The grey envelope at the base of the crystalline crust shows the range of values for the various published works mentioned in the text (including papers cited by Starostenko et al., 2004). The dashed black line at the base of continental lithosphere mantle is the (approximate) LAB from the integrated geophysical model of Starostenko et al. (2015b); the blue line is the (approximate) LAB derived from the passive seismological study of Bijwaard and Spakman (2000); the red line is the (approximate) LAB from the thermal model of Tesauro et al. (2009), defined as the 1200° isotherm; and the black line is the model LAB of Entezar-Saadat et al. (2020) adjusted towards the south by the model of Motavalli-Anbaran et al. (2016). As such, the grey zone defined by these lines represents a LAB zone of uncertainty. The architecture of the layer labelled "sediments" is purely schematic and represents only a very approximate view of published data constraining this horizon (cf. Fig. 3). (For interpretation of the references to colour in this figure, the reader is referred to the web version of this article.)

of geological and geophysical data available at the time (1970s–1980s; see Starostenko et al., 2004, for further references).

Geophysical data considered by Starostenko et al. (2004) included a series of DSS (Deep Seismic Sounding) profiles in the Black Sea acquired in the 1960s. Later, Yegorova et al. (2010) recorrelated and remodelled some of these data, including one profile crossing from the Odesa Shelf into the central WBSB, partly in the present study area (on the Odesa Shelf), producing a crustal velocity model with a Moho at 39 km and 18–20 km depth, respectively, similar to Starostenko et al. (2004). There is also an E-W running composite onshore-offshore wide-angle reflection-refraction profile crossing the Odesa Shelf, approximately along the southern margin of the Karkinit Trough (KT; Fig. 3) called DOBRE-5 (Starostenko et al., 2015a). The onshore segments, across Crimea and onto the Pre-Dobrogea Depression (PDD; Fig. 3), are modern and the offshore segment comprises DSS of similar vintage as the Yegorova et al. (2010) line recorrelated and modelled collectively with the modern data segments. The resulting velocity model shows a slightly shallower Moho (~35 km) beneath the Odesa Shelf in this area than that of Yegorova et al. (2010).

The Geology Without Borders deep seismic reflection dataset does not image Moho (or any horizon interpreted as Moho) in the WBSB although one profile (BS-90) in the EBSB shows a faint event at ~12 s that Nikishin et al. (2015a) tentatively interpreted as Moho, in part because it correlates with a wide-angle reflection event recognised by Minshull et al. (2005). This lies some 3–4 s (two-way travel-time) below the base of the presumed sedimentary succession, which is interpreted by Nikishin et al. (2015a, 2015b) to be the top of oceanic crust. The age of the directly overlying sedimentary succession is unknown although Nikishin et al. (2015a, 2015b) consider it to be Upper Cretaceous, since it is evidently post-rift. The basement horizon itself is quite ambiguous and is actually shallower in Nikishin et al.'s interpretation than what has been identified as Cretaceous strata by Stovba et al. (2020). In the Starostenko et al. (2004) model, Moho is at a depth of 19–20 km in the centre of the WBSB where the base of immediately overlying "consolidated" sediments is 15–16 km, implying crystalline crust as thin as 3–4 km.

This last point highlights one of the significant issues with determining present-day crystalline crustal thickness beneath the deep basins of the Black Sea, this being the ambiguity intrinsic to defining the base of the overlying sedimentary layer and the age of the oldest sediments at the base of that layer. In the case of Starostenko et al. (2004), the "consolidated" sedimentary layer was defined on the basis of seismic velocities from DSS-style observations. In seismic reflection images (e.g., Stovba et al., 2020; Nikishin et al., 2015a,b) it may be possible to identify the presence of sediments on the basis of laterally coherent reflecting horizons and evident stratification, such as seen in Fig. 3, but there are no nearby wells with stratigraphic control for firm age correlation. Accordingly, it is not possible in the deeper sedimentary successions of the deep basins of the Black Sea to assign stratigraphic ages diagnostically and, therefore, with confidence. A Cretaceous age for the deepest sedimentary strata, at least in some instances where correlation involves significant distances and possibly ambiguous lateral coherence of reflection horizons, is an assumption allied with a tectonic model that assigns all of the sedimentary succession to the main Black Sea rifting event in the Cretaceous.

The depth of the sediment-crystalline basement boundary is difficult to know with certainty in the deep basins of the Black Sea and, further, there exists some variability, both in amplitude and spatial pattern, in the various models proposed for depth to Moho in the Black Sea as a whole. Nevertheless, almost without exception, these models document that there is a shallowing of Moho (almost certainly accompanied by a thinning of the overlying crystalline crust), coincident with the main WBSB and EBSB sedimentary depocentres, as illustrated in Fig. 6 for the former.

3.3. Lithosphere-asthenosphere boundary (LAB)

There are few direct geophysical observations – meaning those comprising mainly seismological models such as receiver functions or tomographic inversions of body and surface waves – of lithosphere thickness of the Black Sea at the scale of the Black Sea itself. Artemieva et al. (2006) suggested a LAB depth, based on a filtered P-wave tomographic model of Bijwaard and Spakman (2000), who probably had fairly poorly distributed data for the Black Sea region given its scale, of 200 km decreasing to less than 100 km southwards along the section depicted in Fig. 6. Similarly, the results of Shapiro and Ritzwoller's (2002) surface wave tomographic model were interpreted to indicate a LAB of 150–100 km along this cross-section also decreasing from north to south. Recent work from Alkan and Çınar (2021), from joint inversion of receiver functions and surface waves, places the LAB at 90–100 km along the Anatolian coast south of the Black Sea, results not dissimilar to those mentioned above.

Besides these seismologically based models there do exist several models of Black Sea lithosphere based on thermal parameters (such as surface heat flow and temperatures within boreholes) and, more recently, thermal parameters inverted or forward-modelled simultaneously with other geophysical constraints such as crustal structure (as described above in Section 3.1) as well as gravity, topography and geoid observations. Tesauro et al. (2009), shows the LAB ranging from 160 km to 130 km north to south along the cross-section in Fig. 6, rather compatible with the seismological estimates. This is based on the assumption of a LAB temperature of 1200° in a 3-D temperature model of Europe. Limberger et al. (2018) presented more detailed thermal modelling using more complex methodologies, using some of the same input parameters as Tesauro et al. (2009), and this work also suggests a moderately cool to cold Black Sea lithosphere. Temperatures at 100 km depth were computed to be in the range of only 700° – 900° beneath the Black Sea, constraining the LAB to be clearly much deeper. Artemieva (2019) presented a new method for modelling thermal lithosphere that she called thermal isostasy. The results for the Black Sea include a narrow, anomalous shallowing of LAB in the vicinity of the northern and western shelf edges of the western Black Sea (to about 100 km compared to >180 km elsewhere). However, the model results in this region are highlighted as being "poorly constrained".

A multidisciplinary analysis of geophysical fields was undertaken by Starostenko et al. (2015b) to determine the structure of the lithosphere in the Black Sea; however, only the thermal data and its modelling, given the derived crustal structure, are utilised for defining the LAB. This was done allowing the temperature in the lithosphere to be non-stationary (i.e., a transient thermal conduction governing equation was used), requiring the assumption that the present crustal structure of the Black Sea and any thermal perturbation of the lithosphere linked to its formation occurred during the Cretaceous, not earlier and not later. It was determined that the LAB beneath the WBSB lies at a depth of 90 km with little variability, increasing to 100–120 km under the Odesa Shelf to the north. Finally, something similar was done by Entezar-Saadat et al. (2020), also including satellite gravity and topography; results were presented along a series of cross-sections and one of these (WSB1), projected along tectonic strike onto the cross-section shown in Fig. 6, show the LAB shallowing to about 110 km from 130 to 140 km beneath the Odesa Shelf. A study using similar methodologies by Motavalli-Anbaran et al. (2016), with the Black Sea in the periphery of its model domain, suggest LAB depths greater than 140 for the area of the transect seen in Fig. 6 and much shallower values along the adjacent Anatolian domain to the south. Entezar-Saadat et al. (2020) provide a comprehensive discussion of various LAB results in this area, including many of the works mentioned here. Both the Starostenko et al. (2015b) and the Entezar-Saadat et al. (2020) results depend on a top-down modelling strategy and both, therefore, have to some extent been pre-ordained by input constraints from basin and crustal architecture constraints, which, as mentioned above (Section 3.1), retain a degree of ambiguity.

4. Subsidence analysis of rifting in the Black Sea

4.1. Introduction: measuring and modelling of tectonic subsidence

Continental rifting disrupts the crust and lithosphere by stretching and thinning them and the resulting architecture is basically controlled by isostasy. The mechanics of extension leads to a depression at the surface, which fills with low density water and sediment that is compensated by a shallowing of the Moho and (isostatically) the replacement of lower crust with higher density upper mantle. At the lithosphere scale this process is accompanied by a thermal perturbation that is usually considered to be equivalent in effect to the thinning of the lithosphere as a whole, not just the crust, and the resulting reconfiguration of the lithosphere geotherm. Changing temperatures means changing densities so this also has an isostatic effect, as explained by McKenzie (1978), and the lithosphere's tendency to remain in (approximate) isostatic equilibrium as the rift-perturbed geotherm relaxes back to its steady-state, during but also long after active extension has ceased, gives rise to the classic paired syn-rift/post-rift sedimentary succession observed in basins world-wide.

The intensity of rifting – or the degree to which crust and lithosphere have been thinned by tectonic extension – is often stated in terms of a stretching factor (usually written as β), which is a measure of how much the crust and lithosphere are thinned during extension and rifting according to the 1-D rifting model of McKenzie (1978). Stretching factor β is the reciprocal of the ratio of the thinned post-rift crustal thickness to its initial pre-rift thickness. That is to say, if β were two then the crust after rifting is half of its original thickness. For $\beta = 1.1$, the thinned crust would be 1/1.1 or some 90% of its original thickness. If extension is uniform throughout the lithosphere, then the same β-factor applies to the thinning of the lithosphere as a whole as well. If rifting leads to continental break-up, with the formation of a new plate boundary and formation of an oceanic spreading centre, such as at an incipient continental margin, then β (in terms of McKenzie's mathematical formulation) is infinity. In practice, where the thermo-mechanical processes and lithosphere rheology are obviously much more complex than in the mathematical formulation and may involve feedback mechanisms (e.g., Huismans et al., 2001), it has been generally assumed that a model of $\beta > 4$–5 in effect leads to continental lithosphere rupture (e.g., Keen and Beaumont, 1990).

Observed crustal thickness, determined geophysically, can be an indication of stretching factor although this requires knowledge of initial conditions (e.g., the thickness of the crust prior to rifting). For this reason and other intrinsic uncertainties, it is not the primary way of measuring β. Rather, McKenzie's (1978) model predicts β in terms of pre-rift basement position from the onset of rifting and related subsidence through time, which can be extracted from knowledge of the stratigraphic succession preserved within the rift basin. Of interest, in particular, is the component of the basement subsidence that is purely tectonic in origin rather than related to changes in accommodation space produced by the compaction and isostatic loading effect of already deposited sediments, as well as to extraneous sea level changes. This is called tectonic subsidence and is determined by making corrections for the non-tectonic factors just mentioned. The calculation and application of these corrections is called "backstripping" and requires knowledge of the age and depth of a number of sedimentary horizons within the sedimentary pile, empirically derived compaction parameters for expected sediment lithologies within the pile and, where relevant, knowledge of paleowater depths throughout basin subsidence and possible eustatic sea level changes. Stretching factors are then determined from the shape of tectonic subsidence curves as predicted by McKenzie's (1978) model or variants thereof (cf. Allen and Allen, 2013; pp. 78–88).

Tectonic subsidence analysis can provide insights into the tectonic affinity and structure of the pre-rift Black Sea lithosphere, which are the stated objectives of the present study and, furthermore, the geological and seismic datasets outlined above provide a much-improved observational framework in which this can be done. New results are presented later in this section, following a short overview of relevant work published in the last 25-years.

4.2. Previous work in the Black Sea

A number of authors addressing the rift history of the Black Sea, in the WBSB or EBSB, or both, have made inferences on the basis of the preserved sedimentary successions in these basins and, in some cases, have estimated β-factors from the implied magnitude and rate tectonic subsidence. All such work would have been limited by the uncertainties intrinsic to defining the age and depth of sedimentary horizons, as well as the top of crystalline basement and the age of the sediments directly upon it. Some was carried out in the absence of the current knowledge regarding the time of onset of the compressional tectonic setting that evolved in the Palaeogene and its implied effects on subsidence and basin architecture.

Spadini et al. (1996; cf. Robinson et al., 1995) 2-D forward modelled the regional stratigraphic packages along three profiles crossing the Black Sea, roughly north to south. They assumed the presence of oceanic lithosphere in the WBSB and EBSB as an initial condition in their analysis, adopted maximum $\beta = 5.5$ on this basis, and then constrained other values from the available crustal thickness information (such as mentioned in Section 3.1). As such, the set of model parameter constrained by matching with the observed stratigraphy had to do with lithosphere "strength" rather than rifting intensity and, from this, the authors argued that the pre-rift WBSB lithosphere was thicker and stronger than the EBSB lithosphere (as mentioned in Section 3.2, above).

Later, Meredith and Egan (2002) also did a 2-D forward modelling exercise in the Black Sea, but limited to profiles in the EBSB and assuming that extension started in the Palaeogene and continued throughout the Cenozoic. They found quite low values for their β-factors (<2) from analysing fault displacements and thickness of sediments but considered the excessive accommodation space required for the post-Eocene succession as the biggest issue in their analysis. They argued that this (largely post-rift) succession implied sub-crustal (what they called "deep lithosphere") processes, without explicitly recognising the possible effects of compressional tectonics during the latter part of the Cenozoic.

Nikishin et al. (2003) calculated "basement subsidence" as well as "water-loaded tectonic subsidence" at several points along interpreted and depth-converted legacy seismic reflection profiles (Beloussov and Volvovsky, 1989). These authors did not express their subsidence estimates in terms of β-factors but focused more on the timing of rifting, which they inferred was from late Barremian until Cenomanian (so not dissimilar to the present study). They qualitatively asserted that, accordingly, oceanic crust formed during this time and that a deep marine depositional environment persisted until the Late Miocene (although arguments in support of these assertions are poorly expressed). Like Meredith and Egan (2002), considerable attention was given to the thick post-Eocene succession, which they qualitatively explained as sedimentary infill of a deep-water basin followed by a bending downwards of the entire lithosphere by tectonic compression in the Pliocene-Quaternary.

Shillington et al. (2008) inverse modelled tectonic subsidence based on the EBSB sedimentary succession published by Robinson et al. (1996; cf. Robinson et al., 1995). While recognising the already considerable uncertainty associated with the age-depth characteristics of the sedimentary succession, they structured their modelling approach in terms of testing three schematic scenarios for paleodepth of sedimentary deposition during the Cenozoic, with arguably even greater attendant uncertainty. The three model cases gave cumulative β-factors in the range 2–14 (the latter, in effect, being infinity) although the authors' preferred result was β up to 5 in the southeastern EBSB and 3–4 in the northwestern EBSB and the southeasternmost WBSB, both for a deep marine basin throughout the Cenozoic. The model was applied only to the Cenozoic succession and it assumed that extension continued until the early Miocene (20 Ma) meaning the model parameterisation was inappropriate (being one of tectonic extension rather than compression) during much of the time it was applied.

Khriachtchevskaia et al. (2007) backstripped a series of wells drilled in the Karkinit Trough (KT; Figs. 2–5) area on the Odesa Shelf (OdS; Fig. 1) and using a 1-D tectonic subsidence model based on van Wees et al. (1996), which, in turn, was an elaboration of the McKenzie (1978) model, these authors estimated a β-factor of 1.08–1.13 for the Cretaceous rifting phase. These are the only β estimates in the Black Sea derived directly from well data, wells with detailed information regarding stratigraphic ages and correlation, thickness and lithologies for determining decompaction parameters. Khriachtchevskaia et al. (2010) presented additional tectonic subsidence curves from the same area showing similar tectonic subsidence results but did not publish any β calculations. Yamasaki and Stephenson (2011) applied an inverse modelling approach to Khriachtchevskaia et al.'s (2010) tectonic subsidence curves in which the main model goal functions were extensional force and strain rate through time but where initial crustal thickness (given the present-day crustal thickness as an input parameter) was also inferred as part of the model. Comparison of these model inferred initial and observed present-day crustal thicknesses suggest β-factors of 1.14–1.24, slightly larger but not dissimilar to the earlier forward-modelling results of Khriachtchevskaia et al. (2007).

None of the studies described above found stretching factors β that are representative of continental break-up and oceanic crust formation (when this has not been an a priori assumption), in the main WBSB and EBSB depocentres, the only exception being the result preferred, among several variants proposed, by Shillington et al. (2008) in which subsidence during the Cenozoic compressional phase of Black Sea evolution was modelled as the result of extension. There is no convincing evidence – poorly constrained as it is – that the degree of Cretaceous rifting in the Black Sea has proceeded to oceanic lithosphere formation; rather, it is no more profound than that representative of a typical "failed" rift, one failing to achieve continental break-up.

5. Tectonic subsidence analysis of the Odesa Shelf (OdS) and West Black Sea Basin (WBSB): new results

5.1. Methodology

The litho- and chronostratigraphic data from the carefully calibrated seismic and well-based subsurface correlations presented in this paper (drawn from Stovba et al., 2020) permit a more confident assessment than previous studies of tectonic subsidence and, therefore, the degree of rifting in the western Black Sea. This is particularly true for the WBSB, where such studies were made on the basis of highly speculative interpretations of very deeply buried and poorly resolved seismic stratigraphy. In the present case, the WBSB sedimentary succession is based on more detailed seismic interpretations and, most importantly, seismic interpretations that are much more tightly correlated with those on the adjacent OdS. In terms of tectonic subsidence analysis, there also exists the earlier backstripping and tectonic subsidence modelling on the Odesa Shelf by Khriachtchevskaia et al. (2007, 2010), which will provide an independent calibration of results derived from the present stratigraphic dataset.

The main objective of the present tectonic subsidence analysis is to test a single main hypothesis, which is whether Cretaceous rifting in the WBSB was severe enough such that current basement beneath Cretaceous and younger sediments is "oceanic" crust or, indeed, was severe enough such that it is exhumed (continental) mantle. Accordingly, the estimation of tectonic subsidence by backstripping and its modelling to determine stretching factor β is carried out in a first-order manner

Table 1
Stratification and depths of the bases of sedimentary sequences used in the backstripping calculations. TWT are observed from seismic data. Average velocities of the strata overlying a horizon in the WBSB are from velocity analysis during seismic processing (Tugolesov et al., 1985) and are prolongated below the Oligocene horizon with interval velocities from Finneti et al. (1988). VSP determinations were used in the Archangelska-1 well (Khriachtchevskaia et al., 2007; 2010). Depths and interval velocities are calculated from these except for depths in boldface, which are measured directly in the Archangelka-1 well. WBSB interval velocities in parentheses are set arbitrarily higher than those from Tugolesov et al. (1985) and Finneti et al. (1988) to provide upper limit thickness estimates (also in parentheses). Final column shows the three-layer stratigraphic column that was backstripped: (1) syn-rift, (2) early post-rift and (3) late post-rift plus superimposed compression (see text and Table 3). The water layer was not considered when backstripping. Depths and thicknesses have been rounded off to the nearest 10 m.

	Horizon (base of unit)	TWT (sec)	Average velocity overlying unit (m/sec)	Interval velocity of unit (m/sec)	Depth (m)	Thickness of unit (m)	Thickness of backstripping units (m)
WSBS – Deep water basin of the western Black Sea	sea (bottom)	2842	1500	1500	2130	2130	0
	Middle Miocene	6402	2356	2360 (2900)	6330 (7290)	4200 (5160)	(3) 10,420 (13,020)
	Oligocene (Maykopian)	9302	2800	3338 (4200)	11,170 (13,380)	4840 (6090)	
	Upper Eocene	10,040	2895	3745 (4800)	12,552 (15,150)	1380 (1770)	
	Palaeogene	10,797	2995	3931 (4800)	14,040 (16,970)	1490 (1820)	(2) 2690 (3170)
	Upper Cretaceous (post-rift)	11,360	3078	4263 (4800)	15,240 (18,320)	1200 (1350)	
	Cretaceous syn-rift (Albian-Cenomanian)	12,090	3175	4301/(4800)	16,810 (20,070)	1570 (1750)	(1) 1570 (1750)
KT – Odesa Shelf prolongated Archangelska-1 well	sea (bottom)	53	1500	1500	**40**	40	0
	Middle Miocene	748	1871	1900	**700**	660	(3) 2240
	Oligocene (Maykopian)	1635	1958	2029	**1600**	900	
	Upper Eocene	2225	2046	2305	**2280**	680	
	Palaeogene	2843	2356	3485	**3360**	1080	(2) 2430
	Upper Cretaceous (post-rift)	3520	2674	3994	4710	1350	
	Cretaceous syn-rift (Albian-Cenomanian)	4020	2851	4100	5730	1020	(1) 1020

designed to test this first-order hypothesis.

The Cretaceous rift-related stratigraphic succession, as discussed above, can be readily subdivided into three temporal, tectonic, phases: (1) the Albian-Cenomanian (113–94 Ma) syn-rift succession, (2) the first 56 My of the Late Cretaceous-Eocene (94–38 Ma) post-rift, thermal sag succession and (3) the later Cenozoic succession post-rift, thermal sag succession, deposited since 38 Ma (~end of middle Eocene) during which time a series of compressional pulses significantly enhancing tectonic subsidence can be recognised. Tectonic subsidence during the last of these is presumed to be driven by thermal relaxation of rift-related changes to the lithosphere geotherm as well as by other processes related to the response of the lithosphere to tectonic compression and shortening. Only these three successions are considered during the backstripping process. Greater resolution than this is unwarranted for the first-order approach that is being used. In turn, the first-order modelling approach (described below) is fully sufficient to test the hypotheses being posed in the study.

The ages of the base and top of each of three key successions are defined; their thicknesses are measurable or can be accurately estimated from the seismic sections (Table 1); and their bulk lithologies, for decompaction estimates, are known where drilled or can be assigned with confidence where they are not (Table 2). Although only three units are used for backstripping, rather than multiple, thinner units, there is minimal to no additional error associated with sediment loading and decompaction corrections using only these because both corrections are cumulative within the observed, present-day "complete" succession.

This approach provides estimates of tectonic subsidence at the end of the syn-rift phase (94 Ma) and at 38 Ma, at which time the post-rift subsidence thereafter is tectonically perturbed. The post-38 Ma part of the succession is, nevertheless, an essential part of the backstripping calculations because it has buried the older successions and the weight of its bulk has compacted the sediments within those underlying successions. However, it is not considered in the forward modelling approach. It provides nothing further in terms of testing the main hypothesis in this study and, in any case, it is not possible to distinguish the effects of its two tectonic constituents (post-rift thermal subsidence superimposed by subsidence caused by an unspecified mechanism linked to lithosphere compression).

The backstripping provides two independent measures for constraining the stretching factor β. These are tectonic subsidence at the end of the syn-rift phase (1) at 94 Ma and during the early post-rift phase (2) at 38 Ma, both of which can be easily modelled using McKenzie's (1978) methodology. The first of these gives a direct measure of β required to provide accommodation space for the syn-rift succession and is referred to as β_1. The second gives a direct measure of β required to provide accommodation space for the first 56 My of deposition of the post-rift succession and is referred to as β_2. Estimation of β_1 and β_2 will be done to closed one decimal place and is done using the simplest rift subsidence model possible, this being the pure shear model of McKenzie (1978), which assumes instantaneous stretching (the "syn-rift"phase) and uniform thinning of the lithosphere as a whole. Regarding the first of these, Jarvis and McKenzie (1980) noted that the effects on post-rift subsidence (caused by the conductive heat loss during a finite period of rifting) are not significant if the period of rifting is $\leq$ 20 My. For the present study this is 19 My. The calculations of Armitage and Allen (2010) further demonstrate that the differences in magnitude of post-rift subsidence calculated for an instantaneous rifting model versus one implementing a finite rifting period for $\beta < 2$ are considerably less than a β difference of 0.1, so less than the nominally required precision in the present study.

The backstripping and forward modelling was carried out using equations similar to those of Allen and Allen (2013), which were, in turn, based on the methodologies proposed by Steckler and Watts (1978) and McKenzie (1978), respectively.

5.2. Backstripping data

The tectonic subsidence analysis, first backstripping to estimate

Table 2
Lithological characterisation of stratigraphic units for backstripping. See text for further explanations. Bulk decompaction parameters are based on a weighted average of lithology-specific values from Allen and Allen (2013, p. 331, citing Sclater and Christie, 1980) for the given approximate bulk compositions of the three backstripping units: (1) syn-rift, (2) early post-rift and (3) late post-rift plus superimposed compression. Percent volcanics in unit (1) is highly spatially variable and generally uncertain and has been neglected for calculating bulk decompaction parameters. Siltstone has been aggregated with sandstone in unit (3).

Stratigraphic unit	Sand-stone (%)	Silt-stone (%)	Shale (%)	Carbonates (incl. chalk) (%)	Volcanics (%)	Approximate bulk composition		Bulk decompaction parameters		
								Surface porosity (%)	Depth constant (km^{-1})	Grain density ($kg\ m^{-3}$)
Quaternary-Middle Miocene	10	25	60	5	–	(3)	7.5% sand	0.63	0.52	2666
Lower Miocene-Oligocene (Maykopian)	10	10	80	–	–		80% shale			
Upper Eocene	5	10	50	35	–		12.5% chalk			
Middle Eocene-Palaeocene	5	–	45	50	–	(2)	30% shale	0.68	0.65	2695
Upper Cretaceous post-rift	–	–	10	90	–		70% chalk			
Cretaceous syn-rift (Albian-Cenomanian)	20	–	70	–	10	(1)	25% sand 75% shale	0.60	0.45	2658

tectonic subsidence and then 1-D forward-modelling to determine rift intensity in terms of a stretching β-factor, has been applied to the OdS and the adjacent, deeper WBSB. Results for the former are calibrated by the previous work of Khriachtchevskaia et al. (2007; 2010) and, with a common workflow and set of assumptions can be compared directly to those of the latter. The locations where backstripping data have been compiled are shown in Fig. 2 (white dots).

The bulk of the backstripping data for the OdS were derived from the Archangelska-1 well, which lies within the Cretaceous rift zone of the Karkinit Trough (white star in KT; Fig. 2) where Cenozoic inversions did not cause significant erosion of sediments. Archangelska-1 penetrated almost the whole of the Quaternary-Palaeocene succession. Data for deeper units, to the base of the Cretaceous syn-rift sediments, were derived from correlation of regional seismic lines to boreholes, which penetrated through the Cretaceous syn-rift succession, on the Crimea Peninsula and elsewhere on the OdS (Stovba et al., 2020).

For the WBSB, similar data have been compiled describing a "pseudo-well" located in its deepest part, where water depth reaches 2130 m (white star in WBSB; Fig. 2). Stratification is based on the correlation of the regional seismic dataset from the deep basin into areas of shallower water (Stovba et al., 2020). Conversion of seismic two-way travel-times (TWT) to depth as far as the base of the Quaternary-Oligocene succession was done with a velocity versus TWT function complied by Tugolesov et al. (1985), derived from velocity analyses carried out as part of the processing of seismic data for stacking. Interval velocities for the Eocene-Cretaceous sequences were taken from Finetti et al. (1988) who extrapolated velocities of sediments lying at shallower depths taking into consideration the dependency of velocity with lithostatic load.

Table 1 summarises the adopted stratigraphic backstripping data for the KT and WSBS locations. The final column shows the aggregated thicknesses of the three units (1−3) used for backstripping and forward-modelling as described in Section 5.1. The time-to-depth conversions to determine the thicknesses of these units, however, were carried out separately for each of the constituent sequences units.

Sedimentary facies and faunal studies suggest that depositional water depth on the OdS never exceeded 200 m, with much shallower water depth generally prevailing (Khriachtchevskaia et al., 2010; Stovba et al., 2020). Stovba et al. (2020) further concluded that there is no evidence from seismic facies comparisons, or from other considerations, suggesting a strongly different depositional environment for the WBSB from what is observed for the KT. In the absence of other directly observed information, it has been assumed that the lithological parameters for the WBSB are the same as those for the KT and these are summarised in Table 2. Bulk decompaction parameters in Table 2 are based on a weighted average of lithology-specific values (e.g., Sclater and Christie, 1980).

The present-day water column has not been considered further for backstripping or modelling nor have possible changes in depositional depth been taken into account. Both are considered inconsequential given the methodologies being utilised (described above) and this will be discussed further below.

5.3. Backstripping and forward modelling results

Fig. 7 shows the tectonic subsidence and basement subsidence curves for the data compiled in Table 1. Solid lines represent tectonic subsidence and dashed ones represent basement subsidence. As explained in the previous section, these curves are corrected for compaction effects using the lithological parameters in Table 2. There are two sets of curves for the WBSB. The mauve-coloured curves are based on the initial data as compiled in Table 1. The more teal-coloured curves are based on the initial data but with time-to-depth conversion velocities arbitrarily increased by up to almost 30% (cf. Table 1), so that they represent what is considered an upper limit of stratigraphic thicknesses, which is considered a "worst-case scenario" as regards the intensity of rifting and, accordingly, magnitude of computed stretching factors β. The numerical values of the curves seen in Fig. 7 are listed in Table 3.

During the syn-rift phase of rift basin formation, β describes the degree of thinning of the lithosphere as a result of its extension. The subsidence that takes place is basically the result of maintaining isostatic equilibrium as upper crustal materials are replaced by lower density air, water and sediment in a newly forming basin as the crustal surface is displaced downwards and lower crustal materials are replaced by higher density mantle at the base of the crust as it is thinned and the Moho displaced upwards. However, the amount of accommodation space produced for basin fill depends strongly on the initial crustal thickness for any given β-factor because, for example, thinning a 40 km thick crust by half (20 km of removed crustal column) requires much more isostatic balancing than thinning a 20 km thick crust by half, for which only 10 km of removed crust requires balancing. This is further complicated by the thermal isostatic effects of conformably thinning the lithosphere as a whole by the same β-factor proportionality as the crust.

Fig. 8 shows how this works for ranges of initial crustal and lithosphere thicknesses (expressed as the ratio of initial crustal thickness to the initial lithosphere thickness, referred to hereafter as R_{cL}) for the tectonic subsidence measured at the end of the syn-rift phase (94 Ma), listed in Table 1 and shown graphically in Fig. 7. Depending on R_{cL}, an

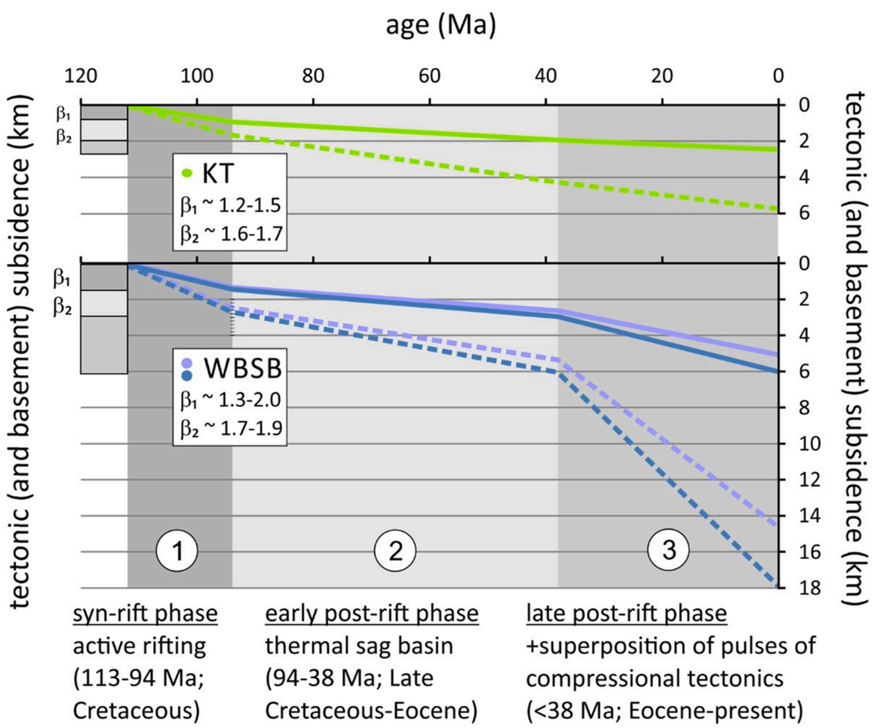

Fig. 7. Graphical representation of data tabulated in Table 3: decompacted depth of basement (dashed lines) and backstripped (water-loaded) tectonic subsidence (km) versus age (Ma). Green lines refer to the Karkinit Trough (KT) study location and mauve/blue lines refer to the West Black Sea Basin (WBSB) study location, based respectively on the published velocity models (mauve) and the arbitrarily higher velocities model (blue) as shown in Table 1. See text for explanation of the tectonic phases 1,2 and 3 and text and Figs. 8 and 9 for explanation of displayed β_1 and β_2 ranges. (For interpretation of the references to colour in this figure, the reader is referred to the web version of this article.)

Table 3
WBSB and KT basement subsidence (basement depth with time with overlying sediments decompacted) and tectonic subsidence (basement depth with time with overlying sediments decompacted as well as the isostatic effect of sedimentary load removed in a water-filled basin); these values define the curves plotted in Fig. 7. Values in parentheses (WBSB) refer to the thickened successions resulting from arbitrarily increased interval velocities (Table 1). Depths have been rounded off to the nearest 10 m.

Age (tectonics)	WBSB		KT	
	Basement subsidence (m)	Tectonic subsidence (m)	Basement subsidence (m)	Tectonic subsidence (m)
113 Ma (start of rifting)	0	0	0	0
94 Ma (end of syn-rift, start of post-rift)	2450 (2680)	1370 (1410)	1670	960
38 Ma (onset of compression, ongoing post-rift)	5350 (6000)	2480 (2550)	4250	1940
0 Ma (present-day)	14,680 (17,940)	5070 (6010)	5730	2410

array of different β_1 values predict tectonic subsidence amplitudes that satisfactorily match the observed tectonic subsidence. The green and blue dots in Fig. 8 represent the β_1 values for the KT and WBSB data, respectively, as a function of R_{cL} (over a range of crustal thicknesses of 20–45 km and lithosphere thicknesses of 80–150 km). For most continental crust and lithosphere, a range of something like 0.25 $<R_{cL}<$ 0.35 is typical and for this, in the absence of other information, Fig. 8 indicates β_1 ranges for KT and WBSB of ~1.2–1.5 and ~1.3–2.0, respectively. All numerical estimates of β are rounded-off to one decimal place only.

As explained above, estimates of β (β_2 in this case) are independently derived at the end of the unperturbed period of post-rift subsidence at 38 Ma. The modelling shows that β_2-factors required to produce the necessary tectonic subsidence after 56 My of post-rift lithosphere

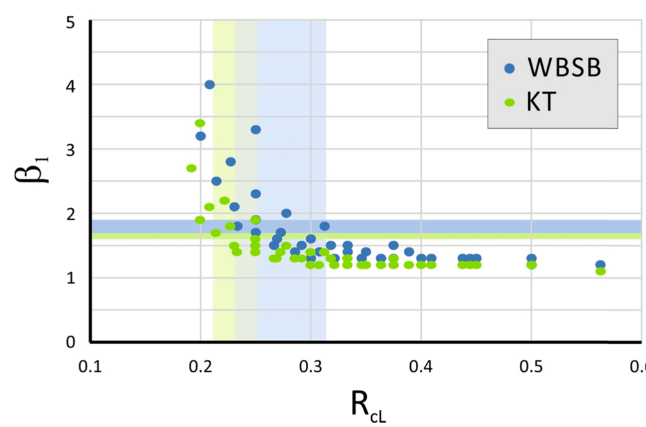

Fig. 8. Best-fitting β values for observed syn-rift subsidence (β_1), to one decimal place, as a function of the (unknown) ratio of initial crustal thickness to initial lithosphere thickness (R_{cL}). Green dots are for the Karkinit Trough (KT) study location and blue dots refer to the West Black Sea Basin (WBSB) study location, the latter computed using the arbitrarily higher velocities model, as shown in Table 1. The green and blue horizontal zones indicate the best-fitting β values for observed initial post-rift subsidence (β_2), to one decimal place, for KT and WBSB, respectively 1.6–1.7 and 1.7–1.9 (see text and Fig. 9). The vertical light green and light blue panels (with intermediate overlap colour) represent the R_{cL} zones where β_1 estimates most closely replicate β_2 estimates for KT and WBSB, respectively. (For interpretation of the references to colour in this figure, the reader is referred to the web version of this article.)

cooling are ~1.6–1.7 for the KT and ~1.7–1.9 for the WBSB. However, post-rift thermal subsidence is strongly dependent on the initial thickness of the lithosphere. The thicker is the lithosphere the greater will be its thermal perturbation for any given stretching factor β (i.e., a 200-km thick lithosphere thinned to half its original thickness will have 100-km of asthenosphere temperature material to "cool off" whereas an originally 100-km thick lithosphere will have only 50 km to "cool off") and it will take much longer for this to happen because of the greater distances over which heat must be diffused.

Fig. 9 shows the suites of model post-rift thermal subsidence curves through time for the inferred Black Sea β_2-factors (taken as precisely 1.6 and 1.8, in the figure, for KT and WBSB), enclosed by the green and blue envelopes, respectively, which overlap in the checkerboard regions. For each envelope the upper bounding curve represents post-rift thermal subsidence through time for a lithosphere with initial thickness 80 km and the lower bounding curve for one with 150 km thickness. The physics of the system, as formulated by McKenzie (1978), results in a limited degree of variability in the first 60 My or so and, notably, minimal variability at around 50 My regardless of initial lithosphere thickness. The dots indicate the observed post-rift tectonic subsidence at 56 My (Table 3) and it can be seen that these essentially fall within the zone of low sensitivity of model predicted tectonic subsidence to initial lithosphere thickness. The inferred β_2-factors are, therefore, moderately robust without further consideration of the unknown initial lithosphere thickness, whether below the Odesa Shelf or the deep basin of the western Black Sea.

6. Discussion

6.1. Intensity of Cretaceous rifting: no continental break-up and ocean crust formation in the WBSB

The tectonic subsidence analysis presented in Section 5 demonstrates that Cretaceous rifting in the deep basin of the WBSB was not necessarily profoundly greater in intensity than in subsidiary, contemporaneous marginal rifts like the KT. This is particularly well demonstrated by the β_2 estimates (1.6–1.7 and 1.7–1.9 for KT and WBSB, respectively); these estimates are not strongly affected by the essentially unknown parameters of initial crust and lithosphere thicknesses (e.g., Fig. 9). Furthermore, the initial post-rift (94–38 Ma) sedimentary succession throughout the Black Sea is itself far more confidently known than the syn-rift succession, both in terms of position and age of its base and top. The β_2 analysis clearly indicates that β was less than 2 in both Black Sea study locations and was likely not significantly higher in the WBSB than

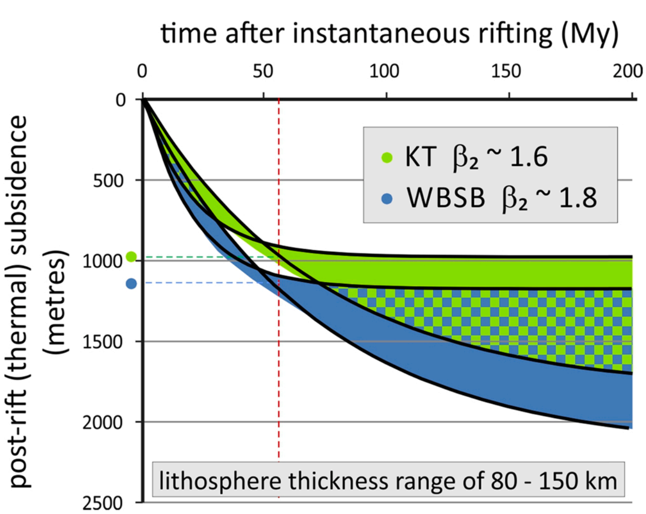

Fig. 9. Post-rift subsidence (m) versus time after instantaneous rifting (My) for β = 1.6 and β = 1.8, these being the best fitting β_2 estimates for the Karkinit Trough (KT) and West Black Sea Basin (WBSB) study locations, respectively (to one decimal place only; see text for further explanation). The green envelope shows the range of post-rift subsidence for β = 1.6 (e.g., KT) through a range of initial lithosphere thicknesses 80–150 km and the blue envelope the same for β = 1.8 (e.g., WBSB). The green and blue dots on the subsidence axis indicate the observed post-rift subsidence for KT and WBSB, respectively, during the first 56 My (marked by the vertical, red-dashed line) after the cessation of the syn-rift period (980 m and 1140 m, respectively, from 94 Ma until 38 Ma; cf. Table 3). (For interpretation of the references to colour in this figure, the reader is referred to the web version of this article.)

in the KT (only some ~10–20%).

β_1 estimates demonstrate a much greater degree of ambiguity because of their dependence on the ratio of crust to lithosphere initial thicknesses (R_{cL}; Fig. 8) but the robust β_2 estimates (Fig. 9) allow a sensible interpretation of the β_1 results in Fig. 8 even in the absence of any knowledge of R_{cL}. The coloured horizontal bands in Fig. 8 delimit the narrowly constrained ranges of β determined from the initial post-rift subsidence (i.e., β_2), which are essentially insensitive to R_{cL} after the first 50–60 My of post-rift subsidence and the respectively coloured dots represent values of β_1, as a function of R_{cL}, to which these are highly sensitive. According to McKenzie's (1978) first-order model, β_1 should be equal to β_2 because the model assumes uniform ("pure shear") thinning throughout the lithosphere. The R_{cL} ranges where such a criterion is satisfied by the present results are (approximately) 0.21–0.25 for KT and 0.23–0.31 for WBSB.

These kinds of R_{cL} values are not out of order for either the Odesa Shelf (KT) or the present deep basin of the Black Sea (WBSB). For the KT, the initial, EEC, crustal thickness was likely around 30–40 km and initial lithosphere thickness around 130–150 km (based on materials presented in Section 3 and Fig. 6). The WBSB is closer or at the margin of the EEC so its underlying lithosphere could be in part cratonic, highly thinned at its Paleotethyan margin, conjoined with superimposed or otherwise accreted Variscan terranes of uncertain origin (Fig. 1, inset). The crust and lithosphere along strike of the Variscan belt adjacent to the distal margin of the EEC (south of the Teisseyre-Tornquist line (Fig. 1, inset) could be considered as representative of the pre-rift Black Sea lithosphere. Limberger et al. (2018) following Tesauro et al. (2009), for example, report what could be taken as fairly typical continental lithosphere in this belt with initial crustal thickness of around 35–40 km (excluding the area of the Cenozoic Pannonian Basin; cf. Fig. 1, inset) and initial lithosphere thickness around 100–120 km.

The reason β becomes so large at small R_{cL} in Fig. 8 is that the relative effect of thermal uplift, as the thinning lithosphere becomes on average hotter and hotter and nearer and nearer the Earth's surface, eventually dominates the crust's diminishing capability (as it thins) to compensate isostatically any basin forming (or trying to form) on its surface. Eventually, at small ratios, the Earth's surface is uplifted above its original baseline so that no basin at all is formed. In other words, the relatively high β values at small R_{cL} (<0.25) seen in Fig. 8 are not directly caused by the necessity of generating sufficient accommodation space for a thick, syn-rift package of Cretaceous sediments. They represent a "losing battle" in the model domain rather than in the observation domain for creating accommodation space for any thickness of syn-rift sediments against the countervailing effects of thermal uplift caused by a sudden elevation of the lithosphere geotherm during extension.

This is a physical consequence of how McKenzie's (1978) model is formulated but has been rarely invoked, if ever, to explain any rift basin behaviour recorded in nature, for the simple reason that continental lithosphere with an anomalously thin crustal layer but with moderate overall thickness (say, 20 km and 100 km or R_{cL}=0.2) is itself very rare. In any case, it seems unlikely to be relevant to the present circumstances and certainly is not relevant, given the available geophysical information about present-day crustal and lithosphere structure, to the pre-rift Odesa Shelf (KT) region. The observed syn-rift succession for the WBSB is very similar to that of the KT except that it is a bit thicker and, furthermore, their respective depositional environments are similar as well (Stovba et al., 2020). The model behaviour expressed by the distribution of β_1 estimates plotted against R_{cL} in Fig. 8 is basically the same for each with, accordingly, slightly higher values for the WBSB, so these low R_{cL}, high β_2 results are unlikely to be significant.

On balance, the comparative values of β_1 and β_2 for KT and WBSB, which can be broadly summarised as implying Cretaceous stretching factors smaller for KT than WBSB and also less than 2 for both locations, demonstrate that it is permissible to accept the first-order modelling results presented in this work as reliably indicating that Cretaceous rifting in the Black Sea could not have achieved continental break-up and formation of oceanic or sub-oceanic crust and lithosphere beneath the deep basin of the WSBS.

6.2. Legacy of lithospheric extension

Some authors have interpreted the available geophysical observations to suggest that the present-day crustal thickness beneath the WBSB may be 10 km or less (e.g., Starostenko et al., 2004; Nikishin et al., 2015). Such values, however, are likely reconcilable with the present tectonic subsidence modelling results if the crust at the outset of Cretaceous rifting was already relatively thin. A β-factor of, say, maximum, 2 as inferred in the present study, is not contradictory if the continental crust prior to the Cretaceous rifting was already thinned to 20 km.

One possible explanation for such a circumstance is that the Black Sea formed atop lithosphere with a legacy of earlier tectonic extension, crustal thinning and concomitant basin formation. Indeed, there is abundant evidence that the lithosphere underlying the Black Sea could well have been subject to repeated episodes of post-Variscan extensional tectonics. Two examples of pre-Cretaceous extensional events affecting the crust northwest of the Black Sea are shown Fig. 10. These seismo-geological cross-sections demonstrate Jurassic (Fig. 10a), most recently, and, before that, Permo-Triassic (Fig. 10b) extensional tectonics.

The Jurassic half-graben imaged in Fig. 10a is in the subsurface of the Pre-Dobrogea Depression (PDD; cf. Fig. 2) near the northwestern margin of the Black Sea. The extensional faulting seen in Fig. 10a may be contemporaneous with the formation of the Greater Caucasus Basin (GCB; Fig. 1, inset), formed during an earlier stage of back-arc tectonics north of the Neo-Tethys oceanic system (Barrier et al., 2018) and now inverted to form the Greater Caucasus Mountains to the northeast of the Black Sea (e.g., Saintot et al., 2006b). What is interpreted as a remnant of the GCB has been imaged in wide-angle seismic velocity model (profile DOBRE2) to a depth up to 10 km, beneath Cretaceous and younger sediments, in the western prolongation of the Greater Caucasus between the Azov and Black seas (Starostenko et al., 2017). Furthermore, Liu et al. (2021) recently postulated, from the analysis of detrital zircons, that the "Eastern Black Sea", as they called it, had formed by the Middle Jurassic (although it is very likely that they were referring to the GCB as otherwise generally understood). Jurassic sediments are also preserved in some wells on the OdS and eastern Crimea Peninsula (CrP; Fig. 1) as well as in outcrop in the Crimea Mountains (CM; Fig. 2) and in the North Crimea Trough (NCT; Fig. 2) although there is no fixed evidence of Jurassic faulting per se (Muratov, 1969; Kruglov and Cypko, 1988; cf. Stephenson et al., 2004).

The Permo-Triassic faulting is possibly part of a widespread rift-wrench tectonic system affecting much of Variscan Europe (e.g., Ziegler, 1990, 2006) as well as reactivating peri-cratonic areas such as the Pre-Dobrogea Depression (PDD; cf. Figs. 2 and 10b) and the intra-cratonic, late Palaeozoic Dniepr-Donets Basin (DDB; Fig. 1, inset) at this time (e.g., Stovba et al., 1996; Saintot et al., 2003). Fault-bounded Permo-Triassic sediments are also reported in the NCT (Fig. 2) by Muratov (1969) and Kruglov and Cypko (1988).

Potential pre-Cretaceous extensional or rifting phases such as these would have thinned the lithosphere as a whole, so any consequences must necessarily be reconcilable with present-day lithosphere thickness as well as with concomitant crustal thinning. The resulting cumulative thinning, through geological time, would not be proportionally the same for the lithosphere as a whole as for the crust, even if McKenzie's (1978) worked perfectly in nature. This is because once active rifting was complete the thickness of the lithosphere would begin to recover by the thermal relaxation of the originally perturbed lithosphere geotherm. In contrast, the concomitantly thinned crust would remain as such (in the absence of other processes). If there were a series of legacy extensional events the effects would be cumulative for the crust (progressively greater thinning after each event) but, for the lithosphere, this would be a matter of the superposition of the effects of multiple thermal relaxation times and the length of the period over which these occurred. It follows that a tectonic legacy involving a series of extensional events could have the effect of producing lithosphere with a relatively small R_{cL}: thin crust but disproportionally thicker, partly thermally relaxed, lithosphere and – because mantle rocks are intrinsically stronger than crustal materials – a relatively stronger lithosphere rather than a weaker one.

A similar concept was developed and applied directly to the Ordovician-Jurassic evolution of Canning Basin in NW Australia by Braun (1992). Braun (1992) had limited crustal thickness information but focused on inferences that the strength of the lithosphere increased during this period of repeated extensional events, which he called "post-extensional mantle healing", rather than being weakened by a series of thermal events and structural reactivations.

Such a model for the Black Sea is in keeping with a number of studies suggesting the underlying lithosphere is stronger than what might be

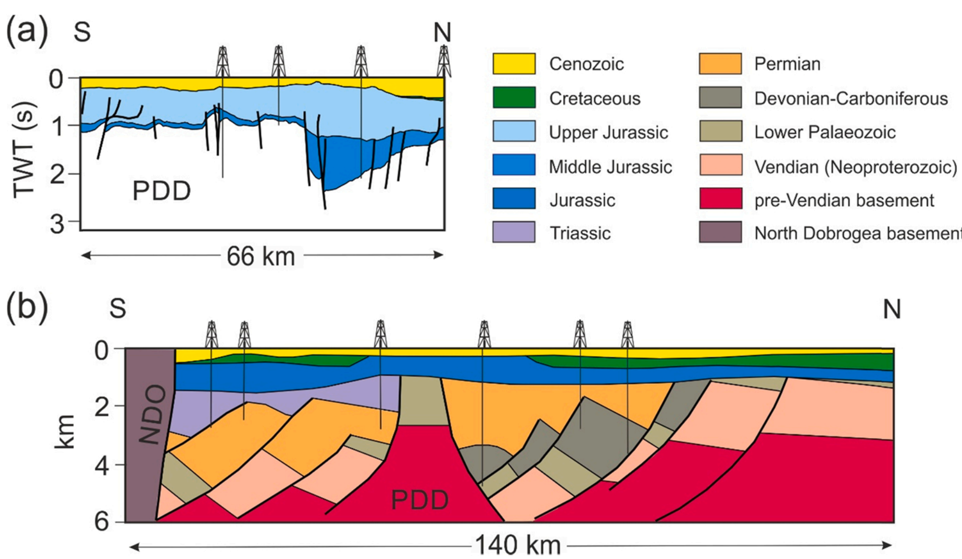

Fig. 10. (a) Interpreted legacy onshore seismic profile showing Jurassic half-graben above undivided Palaeozoic strata (modified from S.V Koltsov, pers. comm.; cf. Koltsov, 1999) (b) geological cross-section (from wells and seismic) showing Permian rifting (A. Seghedi, pers. comm.; cf. Seghedi, 2012), both in the Pre-Dobrogea Depression (PDD) and located in Fig. 2. NDO – North Dobrogea "Orogen". Geological legend pertains to both panels.

expected if it had been broken apart, as some authors have suggested (although the results of this work strongly suggest otherwise), to the point of asthenospheric diapirism producing oceanic crust at the surface as recently as the Late Cretaceous. Based on the thermal model presented in Section 3, Tesauro et al. (2009) computed integrated lithosphere strength profiles (i.e., Ranalli and Murphy, 1987) for European lithosphere and found that the Black Sea lithosphere is moderately strong, much stronger, for example, than Anatolian lithosphere to the south and generally stronger than European lithosphere to the north and northwest. Similarly, Yegorova et al. (2013), from gravity modelling and seismic tomography of the crust and upper lithosphere mantle of the Black Sea, inferred the presence of rheologically strong continental lithosphere in this area, similar in strength to the Precambrian lithosphere of the East European Craton north of the Black Sea. A strong, cold lithosphere for the Black Sea, in particular its western segment, was also inferred from rift/basin modelling results – numerical and analogue, respectively – by Cloetingh et al. (2003; following Spadini et al., 1996) and Stephenson and Schellart (2010).

This kind of scenario is schematically portrayed in Fig. 11. It is emphasised that the cartoon in Fig. 11 is not intended as a formal model of the evolution of Black Sea lithosphere since the Late Carboniferous. Rather, it is only an illustration of what could be the consequence of several rifting events over several hundreds of millions of years placed in the context of the present modelling results. It is semi-quantitative at best, based on simple McKenzie (1978) model-style β-factor calculations and thermal relaxation, as well as speculations about the geological history of the region as summarised above. Simply put, there is a typical initial continental lithosphere of 120 km thickness with a 35 km thick crust that is then affected by three periods of extensional tectonics, or rifting events: Permo-Triassic and Jurassic events, followed a Cretaceous rifting event. The first two are evidential in the Black Sea area (e.g., Fig. 10) but otherwise unconstrained in terms of model parameters. The third has model parameters based on the first-order modelling results presented in this work (duration 113–94 Ma, β of 1.8) for the WBSB. The Permo-Triassic event is taken to have occurred during the period 260–240 Ma and the Jurassic event 170–160 Ma. The chosen onset and

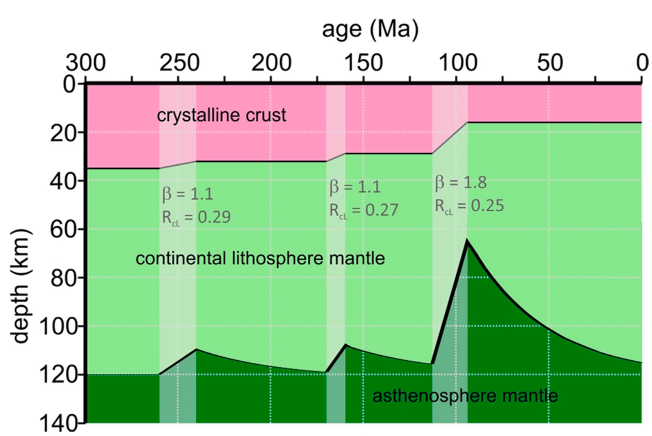

Fig. 11. Cartoon of the evolution of 120 km thick continental lithosphere with 35 km thick crust during 300 My incorporating the simplified results of a sequence of three independent rifting events (viz. McKenzie, 1978), broadly intended to represent Permo-Triassic and Jurassic events (260–240 Ma and 170–160 Ma, respectively), each with an imagined stretching factor β of 1.1, followed by a Cretaceous rifting event based on the results of the present work for the WBSB (113–94 Ma and β of 1.8). These are indicated by the vertical grey zones. The crustal thickness to lithosphere thickness ratio (R_{cl}, as discussed in the text) for each rifting event are also indicated. The present-day (age 0 Ma) thickness of the crustal layer represents the accumulated thinning imposed by each rifting event while the base of the lithosphere tends to relax back to its initial position (neglecting the effects of thermal blanketing of sediments). It is also rheologically stronger than it was initially because of a greater ratio of mantle to crust.

duration of these events are not incompatible with what is seen in Fig. 10 nor are they in any way tightly constrained. The β-factors for each have been taken as 1.1; this is based on abundant evidence that the vast majority of continental rift basins observed globally can be characterised by β-factors less than 1.2 (e.g., Newman and White, 1999), so can be considered a typical, modal value.

If the various parameters used to construct Fig. 11 were varied within reasonable limits the image would possibly change discernibly, though not drastically, but its message would remain exactly the same. This is simply that the results presented in this work, based on the well-constrained Cretaceous rifting event affecting the Black Sea, at least its western segment, are compatible with (1) what is known about the present-day structure and strength of the crust and lithosphere of the Black Sea and (2) the tectonic history of the Black Sea area recorded by the geology observable on its margins.

6.3. *The eastern Black Sea*

The eastern part of the Black Sea within the present study area (Fig. 2), including the East Black Sea Basin, Andrusov Ridge and Shatskiy High (EBSB, AR and SH; Figs. 2, 4 and 5), all of which continue into the adjoining parts of the Black Sea, consists of highly inverted syn-rift (half-)grabens (Stovba et al., 2020). The main depocentres of the syn-rift sedimentation and magmatism according to these authors took place on the AR and perhaps SH, as well as what are now the Crimea Mountains (onshore and off), Sorokin Trough and Tetyaev High (CM-MCCF; ST and TH, all within the present study area; Figs. 2, 4 and 5).

The syn-rift isopach map (Fig. 4b) shows generally thicker syn-rift sediments in the eastern Black Sea, even not taking into consideration that syn-rift sediments were partly eroded in areas like the AR and SH during the subsequent Cenozoic compressional events. In this regard, it is noted that Stovba et al. (2020), based on their detailed interpretations of seismic profiles, considered that the EBSB itself comprises mainly the footwall of an AR half-graben during the rift stage but was subsequently inverted less than other contiguous tectonic elements (cf. Fig. 5). This partly can explain why the thickness of the syn-rift sediments in the EBSB is comparable to that in the WBSB.

Nevertheless, even with the added uncertainty of the incomplete post-rift succession in the EBSB within the present study area, it can be speculated that the well-constrained inferences regarding rift intensity in the WBSB may not be substantially different for the EBSB, in particular as regards the unlikely presence of oceanic crust. However, it is cautioned that there is significant along-strike variability in the EBSB (particularly in its majority segment outside the present study area) according to some authors (e.g., Monteleone et al., 2019; 2020), including significant features possibly related to Eastern Pontide tectonics and magmatism that are of relevance to the tectonic record of the eastern Black Sea (e.g., Nikishin et al., 2015a). Further, Monteleone et al. (2019) and others (e.g., Robinson et al., 1996; Shillington et al., 2009, 2017; Hippolyte et al., 2015) have considered that active rifting in this part of the ESBS is younger than in the WBSB.

7. Summary and conclusions

The availability of a new, detailed and comprehensive interpretations of a dense set of seismic profiles (combined with onshore, sea bottom and borehole geological observations) published as detailed structure and isopach maps (Stovba et al., 2020) has allowed a new tectonic subsidence analysis to be performed in the western segment of the Black Sea. Compared to previous studies, the newly defined sedimentary basin architecture in the area of the present study permits a more confident assessment of tectonic subsidence and its implications for Cretaceous rifting not only on the Odesa Shelf margin of the western Black Sea but in also the main West Black Sea Basin (WBSB) depocentre. This is not to say that uncertainty can be reduced to nil, particularly in the context of the dating of horizons in the deepest part of the WBSB sedimentary succession, but it is significantly reduced and available constraints are much richer.

The tectonic subsidence analysis, which consists of backstripping of the observed sedimentary successions at key locations in the study area and forward modelling the resulting tectonic subsidence estimates, was designed to test the hypothesis that Cretaceous rifting in the main rift axis of the WBSB was insufficient to produce continental lithosphere break-up and formation of oceanic crust beneath sediments deposited from that time.

The geological input constraints to the tectonic subsidence analysis are reviewed, as is what is known of present-day crustal and lithosphere structure of whole of the Black Sea and its vicinity, the latter representing important geophysical boundary conditions for interpreting the tectonic subsidence results. A review of all previous modelling work aimed at defining the effects of Cretaceous rifting on the present-day basin architecture of the Black Sea provides a context for the present modelling analysis.

The main results and conclusions of the new tectonic subsidence results are summarised as follows:

(1) The tectonic subsidence results presented in this paper provide a successful, quantitative test of the geological implications of the seismic and other subsurface observations reported by Stovba et al. (2020) that the main rift structures, including the thickness of the syn-rift sequence, lithofacies distribution and offsets of normal faults bounding these rift structures, are similar throughout the present study area including the deep-water basin of the Black Sea.
(2) The magnitude of rifting, expressed as a stretching factor β, is less than 2 in the present-day deep-water West Black Sea Basin (WBSB), only 10–20% greater than within the Karkinit Trough on the Odesa Shelf of the northwestern Black Sea. A β-factor of 2 or less in the WBSB is significantly less than would be required to produce continental lithosphere break-up and ocean crust formation, or even through-going "oceanisation" of continental lithosphere in the Cretaceous. It follows that the crust and lithosphere underlying the sediments of the WBSB are continental in affinity.
(3) The tectonic subsidence results are compatible with what is known of present-day crust and lithosphere structure, including interpretations reporting a very thin crustal layer (<10–15 km) beneath the WBSB if the legacy structural effects of post-Carboniferous and pre-Cretaceous tectonic extension events affecting the Black Sea lithosphere, for which there is circumstantial evidence on the margins of the Black Sea, are taken into account.
(4) Basin architecture within the present study area suggests that similar conclusions, including those pertaining to tectonic legacy, may also apply in the northern sector of the deep-water East Black Sea Basin (EBSB). Although the present study does not provide direct constraint on the Mesozoic and Cenozoic tectonic evolution of the ESBS, which is likely more heterogeneous than the WBSB, any interpretation of the tectonic history of the ESBS as a whole should necessarily accommodate and be compatible with the implications of the present study.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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