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Journal Pre-proofs

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PII: S1367-9120(21)00042-0
DOI: https://doi.org/10.1016/j.jseaes.2021.104704
Reference: JAES 104704

To appear in: *Journal of Asian Earth Sciences*

Received Date: 27 August 2020
Revised Date: 20 January 2021
Accepted Date: 1 February 2021

Please cite this article as: Korniyenko-Sheremet, Y., Sosson, M., Murovskaya, A., Gobarenko, V., Yegorova, T., Palaeo-stress regimes and structural framework during the Mesozoic-Cenozoic tectonic evolution of the Crimean Mountains (the northern margin of the Black Sea), *Journal of Asian Earth Sciences* (2021), doi: https://doi.org/10.1016/j.jseaes.2021.104704

Palaeo-stress regimes and structural framework during the Mesozoic-Cenozoic tectonic evolution of the Crimean Mountains (the northern margin of the Black Sea)

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Abstract

This paper presents a chronology of tectonic events in the Crimean Mountains (CM) based on a micro- (palaeo-stress) and macro-scale (tectonic structures) structural analysis since the Cretaceous. Recent studies have attempted to fit the geology of the CM into the geodynamic context of the subduction of the Neotethys plate beneath the Eurasian margin; however, these attempts remain difficult and controversial as they are based dominantly on stratigraphic analysis. New structural analysis results indicate that structural development within the CM was largely determined by the influence of pre-existing structures that occurred long before the Cretaceous rifting in the Black Sea (BS). This could explain the features of the geological structure in the Western and Eastern CM separated by the Alushta–Simferopol Fault Zone (ASFZ). New palaeo-

stress analysis allows us to identify the trends and timing of the Mesozoic–Cenozoic tectonic deformations with greater precision. Superimposing the low-magnitude seismicity ($M \leq 3$) with structures defined on the established geological cross-sections permit the localization of currently active structures. Considering both recent scientific achievements concerning the geology of the CM and the results of this study, we propose a qualitative model of the formation of the BS Basin and its inversion in the context of the BS-Greater Caucasus (GC). In this model, the Shatsky Ridge (ShR), moving to the north, acts as an indenter deforming the CM and the GC. The results are of practical importance for geodynamic modelling and hydrocarbon exploration.

Keywords: Crimean Mountains, Black Sea; palaeo-stress; tectonic evolution; structural analysis

1. Introduction

The long-term subduction of the Tethys oceanic plate beneath the Eurasian margin was a geodynamic process that influenced the development of the GC–BS region, of which the CM occupies an integral part (Fig. 1A) (Adamia et al., 1974, 1981; Angelier et al., 1994; Barrier et al., 2018; Barrier and Vrielynck, 2008; Dercourt et al., 1986; Ershov et al., 1999, 2003; Finetti et al., 1988; Khain, 1974; Letouzey et al., 1977; Milanovski, 1991; Nikishin et al., 1998, 2003, 2015c; Robinson et al., 1996; Saintot & Angelier, 2002; Saintot et al., 2006; Sengör et al., 1980; Sosson et al., 2016; Stampfli et al., 2002; Stephenson & Schellart, 2010; Vincent et al., 2016, Zonenshain & Le Pichon, 1986). The CM is a relatively small geological area, ~160–180 km long with a maximum width of 50 to 60 km, and its highest summit reaches 1500 m. However, determining the tectonic stages of the CM during the tectonic evolution of the GC–BS domain remains difficult and controversial (e.g. Saintot et al., 2006). Often, the proposed scenarios of the tectonic evolution of the CM are either limited to a description of stratigraphic features, making casual remarks about

the tectonic structures (e.g. Nikishin et al., 2015c), or focus on only one tectonic stage in the history of the CM (e.g. Hippolyte et al., 2018); moreover, the tectonic evolution of the CM in many studies did not consider the tectonic evolution of the regional context of the BS (e.g. Yudin, 2011).

Nevertheless, recent geological studies devoted to new age determination (Ivanik et al., 2013; Nikishin et al., 2017; Oszczypko et al., 2017, Popadyuk, 2011, Popadyuk et al., 2013a, b; 2014; Sheremet et al., 2014, 2016a) and analysis of tectonic structures of the CM (Hippolyte et al., 2018; Nikishin et al., 2017; Sheremet et al., 2016a, b) allowed the correlation of tectonic stages in both regions between the CM and the GC since the Early Cretaceous. The micro-tectonic analysis of the CM facilitated the determination of the directions of principal stresses (palaeo-stress fields). Detailed works of structural analysis at the micro- and macro-scale by Saintot et al. (1999) and Saintot and Angelier (2002) have been conducted for the CM and Kerch Peninsula (KP) (Fig. 1B) before the new dating information was obtained. Therefore, the stress-field obtained by these authors within the Triassic–Middle Jurassic age flysch rocks, specifically for the Eastern part of the CM, should be reconsidered given their newly defined Cretaceous age (Ivanik et al., 2013; Oszczypko et al., 2017; Popadyuk et al., 2011, 2013a, b; 2014; Sheremet et al., 2014, 2016a). Therefore, this paper is about presenting new palaeo-stress results for the Eastern part of the CM. These new results are going to presented in the context of new geological cross-sections (built taking into account the recent results of the stratigraphy) and localization data of low magnitude ($M \leq 3$) seismicity along the geological cross-sections, which are described and presented in section 3 and 4 before documenting new plaeo-stress results.

As the CM outcrops reveal a much older geological history of this region, this work focuses on structures that formed long before the Cretaceous rifting of the BS (e.g. the Late Triassic rifting and the Early–Middle Jurassic extension) but were observed to have a significant impact on

tectonic development. These structures, deep crustal faults, have been reactivated over the late Mesozoic and Cenozoic tectonic stages of deformation (Sheremet et al., 2019). This study was conducted with specific consideration to the reactivation timing of pre-existing structures and the trends and timing of the tectonic deformations during the Mesozoic–Cenozoic geological history of the CM. This paper demonstrates why there is a difference in geological structures between the western and eastern parts of the CM, and proposes a tectonic evolution of the CM in the context of the GC–BS domain.

2. Tectonic setting and morphology of the Crimean Mountains

The Crimean Peninsula is represented by a mountain domain (the CM) in its south; to the north, it is covered by Cenozoic deposits forming a northwest-dipping monocline in the Crimea relief—the Northern Crimea Lowland (NCL) (Fig. 1B). The continental crust of Crimea and adjacent offshore shelf areas of the BS and the Azov Sea belong to the southern margin of the Eastern European Platform, commonly called the Scythian Platform or the Scythian Plate (SP). The structural heterogeneity of the Crimean crust, revealed by tomography images (Yegorova et al., 2018), can be explained by various tectonic events that occurred during the tectonic evolution of the Eurasian Plate. These include 1) the late Proterozoic–Early Palaeozoic tectonic stage (Saintot et al., 2007, Stephenson et al., 2004, Yegorova et al., 2018; Ziegler, 1988) during which the SP most likely became the southern rigid margin of the East European Plate (or Platform) and 2) the Late Palaeozoic Variscan orogeny that resulted in the occurrence of slightly metamorphic rocks, which are interpreted as the basement rocks of Crimea (e.g. Muratov et al., 1969, Muratov et al., 1984; Okay & Topüz, 2017). Moreover, since the late Permian, the SP has served as a part of the overriding plate (Laurasia) in the subduction process of the Tethys oceanic plate (e.g. Saintot

et al., 2006). For example, the thickness of the crust beneath the Eastern BS and the Mid Black Sea Ridge (MBSR) and the ShR (Fig. 1B) is different: the Moho is at 20–22 km and ~32 km depth under the Eastern BS and under the ridges, respectively (Yegorova & Gobarenko 2010, Yegorova et al., 2010). This difference in thickness most likely indicates a significantly extended continental crust of the Eastern BS (Yegorova & Gobarenko 2010, Yegorova et al., 2010).

The relief of the CM formed as a result of the collisional tectonic events related to the Tethys closure and represents an inverted part of the northern BS shelf due to the Cenozoic compressional stages (Gobarenko et al., 2016; Sheremet et al., 2016 b; Stephenson et al., 2004; Stephenson & Shellart, 2010; Sydorenko et al., 2017; Ushakov et al., 1977). The CM are organized in three almost parallel and very large hogbacks ('yaylas') with very steep southern frontal slopes and gentle northern slopes. The highest elevations around 1200–1500 m comprise the mountain chain of the CM. The heights of the CM gradually decrease to the north to 250–350 m where they meet transition to the NCL (or the 'Steppe-like' Crimea).

3. The geology of the Crimean Mountains

To the geology of the CM, undoubtedly, numerous studies are devoted. But purely for the sake of integrating and interpreting the new results of this paper (the paleo-stress results) we present the summary of the geological features of the CM taking into account new dating results, obtained recently and documented in Popadyuk et al. (2013a, b), Sheremet et al. (2014, 2016a) and Oszczypko et al. (2017). As a visual aid we propose a figure 2, which is composed of two maps juxtaposed: 1) a structural map of the Eastern part of the CM from Sheremet et al. (2016a) and 2) a geological map of Muratov et al. (1969), which was updated here with results of new dating of rocks (Sheremet et al., 2016a). To show stratigraphic features of the CM we appeal to a

figure 3 published earlier for the eastern part of the CM in Sheremet et al. (2016a) but supplemented for this study with stratigraphic columns built for the western part of the CM, taking into account new datings of rocks published in Sheremet et al. (2016a) and Hippolite et al. (2018).

The structural framework of the CM presented in section 3.3 is illustrated with two pairs of geological cross sections, shown on figure 4: the cross section for the western part of the CM were built taking into account the compiled map on a figure 2, while the cross sections for the eastern part of the CM, which we use in the current study for the comparison, are reproduced fragments of the transects from Sheremet et al. (2016b). Additionally, before presenting data on paleostress, it is necessary to show a map of tectonic structures (fig.5), which was built on a base of existed geological data (Hippolite et al., 2018; Muratov et al., 1969; Murovskaya et al., 2018; Sheremet et al., 2016 a,b) and our structural data gathered in the field.

3.1. Pre-Cretaceous stratigraphy and structural forming in the CM

The oldest rock outcropping in the CM, the Tauric Formation (TFm) or Tavrica (Muratov et al., 1984; Nikishin et al., 2015c; Sheremet et al., 2016a; Slavin 1986) is penetrated by several wells around the Crimean Peninsula (e.g. Khriachtchevskaya et al., 2010). This is a succession of thick, dark-coloured shales with intercalations of siliciclastic turbidites, siltstones, and conglomerates in its upper part containing blocks (olistoliths) of Permian and Carboniferous limestones (Fig. 2) (Muratov, 1960, Nikishin et al., 2015c, Sheremet et al., 2016a and references therein). TFm contains magmatic bodies of the Late Triassic age, interpreted as sills (e.g. Spiridonov et al., 1990b). The TFm is very similar to the same flysch-like deposits in the Northern Dobrogea tectonic province (Romania), to the west of the CM (Fig. 1A). This area is located within the same latitudes as the CM, where the Upper Triassic flysch rocks seal the surface of the Middle Triassic limestones and contain pillow lavas, debris flow, and slumps in their lower part (Gradinaru, 1995).

The occurrence of turbiditic formations with olistostromes (deposited on the marine slope) and active magmatism in both areas could be related to a period of subsidence by normal faulting in the Palaeozoic basement of the overriding plate (Laurasia) above the subduction of the Tethys plate during the Late Triassic. The trend of normal faulting was likely parallel to a subduction front interpreted to be of a ~E–W trend, according to the palaeogeographic reconstruction of the Tethys Ocean by Barrier & Vrielynck (2008) and Barrier et al. (2018). Thus, these ~E–W faults could be understood as a first generation of normal faults within the southern Laurasian margin when the subduction starts. Many faults that border the North Azov Trough (Golitsin and Azov faults), which continues as the North Crimean Trough to the west and extends as Karkinit Trough in the Odessa Shelf, as well as the following tectonic units of the western BS shelf such as Gubkin and Kalamit Ridge, Zmeiniy uplift, and Krayova step also elongated in this ~E-W direction. Despite the fact (e.g. Khriachtchevskaya et al.,2010) that all of those tectonic structures were active during the Cretaceous, some of them are just reactivated early Mesozoic structures that attested their pre-Cretaceous tectonic activity by thick accumulated lower Mesozoic deposits north of Zmeiniy Uplift (in the Dobrogea Depression) and south from the Azov Fault (on the Azov Ridge), what is visible from the interpreted seismic lines by Khriachtchevskaya et al., 2010. These ~E–W normal faults, attested from wide-angle refraction and reflection (WARR) velocity model superimposed on the NE–SW seismic line DOBRE-2 crossing the Azov sea, KP, and BS (Starostenko et al., 2017) were also determined by the morphotectonic analysis of the CM (Saintot et al., 1999). During the Latest Triassic–Earliest Jurassic compression, known as the Cimmerian orogeny (Eo or Early Cimmerian), these faults were reactivated and the TFm of the CM as well as their analogues around the BS (Kure Fm in the central Pontides, Dizi Fm in the GC) were deformed (Barrier & Vrielynck, 2008; Mileev et al., 2009; Nikishin et al., 2015c; Okay, 2000; Okay &

Nikishin, 2015; Okay et al., 2015; Robinson et al., 1996; Saintot et al., 1999; Sengör, 1979, 1984, 1987; Sengör et al., 1985; Tari et al., 1997; Ustaömer & Robertson, 1994; Yudin, 2008, 2009 , 2011; Zonenshain & Le Pichon, 1986).

The evidence of the Cimmerian tectonics in the CM is only present westward of the valley joining Alushta and Simferopol (A–S) (Fig. 1B), where the compressional south verging structures (folds and reverses faults) in the TFm are covered with unconformity by the Hauterivian sandy limestones (unconformity U2) (Fig. 3, Fig. 4) (e.g. Murovskaya et al., 2012, 2014; Saintot et al., 1999; Sheremet et al., 2016a).

The Cimmerian deformations were also described to the east of A–S (Saintot et al., 1999). However, recent studies (Ivanik et al., 2013; Popadyuk, 2011, Popadyuk et al., 2013a, 2014; Oszczypko et al., 2017; Sheremet et al., 2014, 2016a;) have shown the Cretaceous age of the flysch rocks, which were previously described as TFm rocks. The position of the Cretaceous rocks under the Jurassic limestones in the Eastern CM has been explained by thrusting and demonstrated in Sheremet et al. (2014, 2016a). The compressional deformations to the east of the A–S probably occurred in post-Cretaceous timespan and are therefore Alpine in age (Sheremet et al., 2014, 2016a).

The middle Jurassic flysch-rocks are more calcareous in the upper part of its succession and unconformably cover the TFm (Unconformity U1, Fig. 3 and Fig. 4) (Muratov, 1960; Muratov et al., 1969). The TFm is intruded by Middle Jurassic magmatic rocks (Fig. 3, Fig. 4) (e.g. Dovgal' et al., 1991; Kazantcev et al., 1989; Lebedinskij & Shalimov, 1967; Leshchuh et al., 1999; Meijers et al., 2010; Muratov, 1960; Okay et al., 2015; Solov'ev & Rogov, 2010; Spiridonov et al., 1990a, b; Teslenko & Yanovskaya, 1990; Nikishin et al., 2015c), which originated in magmatic arc settings related to northward subduction below the Eurasian margin based on 40Ar/39Ar dating

and the geochemical analysis (Meijers et al., 2010). Thus, some tectonic structures that formed earlier (e.g. during the Triassic) were reactivated during the Middle Jurassic back-arc extension of the overriding plate; new micropalaeontological findings of detritic deposits along the ~E–W fault south of Simferopol ('Bitak formation' mostly consisting of conglomerates and with coal seams in some places) show that they were deposited during the Middle Jurassic (Nikishin et al., 2017) (Fig. 1B, Fig. 2, Fig. 3).

Additionally, beginning from the Early–Middle Jurassic, the GC area experienced a rifting stage related to the formation of the GC back-arc basin, followed by the occurrence of normal faults parallel to a NW–SE strike of the GC (the 'Caucasian' trend) (Adamia et al., 1981, Kaz'min & Tikhonova, 2006 a,b; McCann et al., 2010; Nikishin et al., 1998, 2001, Saintot & Angelier, 2002, Saintot et al., 2006; Sydorenko et al., 2017). The system of NW–SE trend faults could be attributed to a second generation of normal faults, which became dominant for the BS-GC domain from the Early–Middle Jurassic. In the vicinity of the future CM, this system of faults (fault zone) played an important role in the tectonic evolution of the CM and influenced the stratigraphy features and structural patterns of the areas termed Western Crimean Mountains (WCM) and Eastern Crimean Mountains (ECM) (Fig. 1B). This wide fault zone (>14 km wide) is referred to as Krivorozhsko–Samsunskiy Fault Zone (Lebedev & Orovetzky, 1966), Salgir–Oktyabrskaya (Gintov, 2005; Gintov et al., 2009; Gonchar, 2005; Patalakha et al., 2003), or ASFZ (Saintot et al., 2006; Sheremet et al., 2016a; Spiridonov et al., 1990a, b) (Fig. 1B). This zone has been active several times since at least the Middle Jurassic (Gintov, 2005; Lebedev & Orovetzky, 1966; Okay & Nikishin, 2015; Saintot et al., 2006; Sheremet et al., 2016a). It justified its activity by (a) sharp changes in the stratigraphy of both regions (Fig. 3) (Abashin et al., 1982; Himshiashvili 1967; Lysenko & Janin 1979; Mileev et al., 2004, 2009; Muratov 1960; Tkachuk 1970; Sheremet et al.,

2016a), (b) the distribution of the Middle Jurassic magmatism (Afanasenkov et al., 2007; Dubois de Montpereaux, 1843; Muratov, 1960; Leshchuh et al., 1999; Lebedinskij & Shalimov, 1967; Nikitina et al., 1979; Sludskij, 1917; Yamnychenko, 1969; Zaika-Novackij, 1981; Zolotarev, 1968; among many others), and (c) the location of seismicity (e.g. Gobarenko et al., 2016). The offset of the NW–SE trending normal faulting of the 'Caucasian strike' (Saintot et al., 2006) bounds the ShR in the Eastern BS, while to the west, they bound the MBSR (Western Crimean Fault, WCF) (Fig. 1B).

The Middle Jurassic rocks of the CM are overlain transgressively by various marine facies (limestones and basinal marine facies) of Late Jurassic age (Karlov, 1963; Kazantcev et al., 1989; Mileev et al., 2004; Robinson & Kerusov, 1997), which is likely due to the activation of the aforementioned ~E–W and NW–SE faults. For example, the age of Upper Jurassic limestones of the ECM and WCM (Nikishin et al., 2015 a, b, Oszczypko et al., 2017; Rudko et al., 2017 Sheremet et al., 2014, 2016a) changes from west to east; in the WCM, the carbonate-rich rocks are dated as late Kimmeridgian–Tithonian (based on 87Sr/86Sr isotope measurements: Rudko et al., 2017), while to the east from ASFZ, the eastern termination of the main ridge is composed of limestones of the Tithonian–Early Berriasian age (Andruhovich & Turov 2002). It means that the area of WCM was elevated compared to the ECM during the Late Jurassic, as the WCM had a more suitable condition for the formation of the Upper Jurassic platform. The wide distribution of carbonate rich rocks could be an evidence of relatively stable tectonic activity in the Late Jurassic (mostly in the Tithonian), while the occurrence of flysch rocks at the end of Tithonian, normal faulting, and back-arc magmatism indicate a new tectonic stage in the geological history of the BS–GC Domain (Meijers et al., 2010; Sheremet et al., 2016a).

3.2. The stratigraphy of the CM since the Cretaceous.

The Early Cretaceous began abruptly in the CM by a period of back-arc magmatism (volcanic massif Karadag in the ECM (Fig. 2), based on 40Ar/39Ar dating and geochemistry (Meijers et al., 2010; Shniukova, 2019) due to the subduction of the Neotethys beneath the Eurasian margin (Fig. 2, 3). Lower Cretaceous rocks, mainly of Valanginian–Albian age, are composed of a large variety of slope-basin deposits (carbonate dominated breccias, conglomerates, siliciclastic sediments, turbidites, shales, and a rhythmic flysch-like sequence) and olistostromes containing olistoliths of Upper Jurassic–Berriasian limestones (Fig. 3) (Dobrovolskaya et Salman 2008; Ivanik et al., 2013; Nikishin et al., 2015c; Oszczypko et al., 2017; Popadyuk et al., 2014; Rudko et al., 2017; Sheremet et al. 2016a;). Hauterivian reef limestones and sandy limestones unconformably cover Upper Triassic rocks in the WCM and the top of the Upper Jurassic–Berriasian rocks of the main ridge and northern hills in the ECM (Unconformity U2 on Fig. 3). Albian shales unconformably cover the Barremian series in the WCM and Hauterivian limestones west of the ECM (Unconformity U3, Fig. 3).

The Cenomanian rocks in the ECM lie unconformably on the Albian rocks (Unconformity U4, Fig. 3). There are several erosional surfaces inside the Upper Cretaceous formations (Muratov, 1960; Zaika-Novatckij et al., 1976; Alekseev et al., 2005; Khriachtchevskaia et al., 2010; Datcenko, 2012), and some stratigraphic horizons pinch out eastwards in the ECM (Datcenko, 2012). The Upper Cretaceous exhibits lithological differences: in the WCM, mostly calcareous marls and chalk limestones are present, whereas in the ECM, they are more terrigenous (Alekseev et al., 2005; Sheremet et al., 2016a).

The volcanic ash material that could be found in the matrix of the Lower Cretaceous olistostromes (Sheremet et al., 2016a), in the Cenomanian and Campanian strata (up to 2–3 cm

thick) (Nikishin et al., 2013, 2015a) and the presence of volcanoclastic sandstones of Late Albian age in the WCM (Nikishin et al., 2015c) indicate magmatic activity during the Cretaceous rifting stage in the BS area (Krezsek et al., 2017; Okay & Nikishin, 2015; Sosson et al., 2016; Tari, 2015).

The palaeocene succession, which is composed by shallow marine limestones, sandy limestones, siltstones, and sandstones (in the upper horizons), is significantly thicker in the ECM (up to 280 m) than in the WCM (~85–90 m) (Stratigraphy of the Ukrainian SSR, 1963) and covers the erosional surface of the Cretaceous strata.

The Lower–Middle Eocene deposits include small-sized nummulite-rich marls (Nikishin et al., 2015c) covered by nummulitic limestones of the Middle Eocene. The thickness of the Eocene nummulitic limestones changes from 10 m to 40–45 m (Lygina, 2010; Datcenko, 2012). The Eocene unconformity (U6 on Fig. 3), overlying Lower Cretaceous to Palaeocene rocks, is well known in the CM (Datcenko, 2012; Geology of USSR, 1969, Kazantcev, 1982; Kazantcev et al., 1989; Muratov, 1960; Nikishin et al., 1999, 2003, 2015c; Pivovarov & Derenyuk, 1984; Popadyuk et al., 2013a, b, 2014; Yudin, 2008, 2009, 2011) and can be identified on the seismic lines off-shore the CM (Sheremet et al., 2016b). The pre-Eocene erosion in the CM ranges from 100 m to 500 m (Nikishin et al., 2015b, c).

The Oligocene–Early Miocene deposits (Maikopian Formation, MFm) are composed of grey, brown, or reddish clays with siderite concretions and gypsum (Vernyhorova & Ryabokon, 2020). The thickness of the MFm strata reaches 110 m in the WCM foothills and 280 m in the ECM foothills (e.g. Muratov, 1960). These series, interpreted as syntectonic terrigenous deposits during the inversion of the GC basin (e.g. Adamia et al., 2010, 2011), unconformably cover the Middle Eocene nummulitic limestones or Upper Eocene marls (Fig. 2) (e.g. Muratov, 1960). The MFm sediments are locally present in the WCM (Pivovarov and Derenyuk, 1984) but they are

absent (possibly eroded; Datcenko, 2012) in the area near Simferopol. In the ECM, the MFm appears only in the north-eastern part of the Kubalach Hills directly on the Middle Eocene limestones (Fig. 2, Fig. 4) (Afanasenkov et al., 2007; Datcenko, 2012 and references therein; Geology of USSR, 1969; Muratov, 1960; Nikishin et al., 2015a, b).

The terrigenous sediments (marine sandy and clay deposits) of the Middle Miocene (40–100 m thick) are marked in the CM by their angular unconformity on the erosional surface of the MFm in the west end of the ECM, while in the far western WCM (Cape Fiolent), they cover the Middle Jurassic volcanic rocks (including pillow lavas) (Datcenko, 2012; Geology of USSR 1969; Nikishin et al., 2006) (Fig. 2).

The Middle Miocene series is capped by the Upper Miocene (Sarmatian) shallow marine facies (limestones, clays, sandstones, and sands), indicating an overall regressional trend. The Upper Miocene succession varies in thickness from 35 to 120 m (Geology of USSR, 1969). It unconformably lies on Middle Eocene limestones and sometimes overlies the Upper Cretaceous chalky limestones (north of Zuya Village, Fig. 2), covered by continental deposits of the Middle Pliocene.

3.3 Main structures of the CM

Comparing the structures of the CM to the east and west from the deep NW–SE ASFZ (Fig. 4), it is possible to notice similarities in the tectonic style of deformations within the turbiditic units. However, the ages of the deformed rocks in WCM and ECM are different (Upper Triassic–Middle Jurassic and Cretaceous, respectively). The eastern sections (Fig. 4) are characterized by the widespread occurrence of Upper Mesozoic, mostly Cretaceous, sediments (see Sheremet et al., 2016a for the details) and represent structures formed by Cenozoic shortening. In contrast, marked by the Hauterivian unconformity U2 (Fig. 3), the WCM sections (Fig. 4) expose the Upper

Triassic–Middle Jurassic slope sediments and compressional deformations of the Cimmerian orogeny in the TFm (e.g. Nikishin et al., 2015c; Okay, 2000; Okay et al., 2013; Okay & Nikishin, 2015; Robinson et al., 1996; Sengör, 1979, 1984, 1987; Sengör et al., 1985; Ustaömer & Robertson, 1993, 1994; Yudin, 2009, 2008, 2011; Zonenshain & Le Pichon, 1986). However, considering the scale of Cenozoic deformations in the ECM, it is clear that the Cimmerian deformations in the WCM underwent a second compression during the Cenozoic (Murovskaya et al. 2014). This interpretation proposed on the cross-sections view (Fig. 4) indicates that pre-existing structures played an important role in determining the location of the Cenozoic thrust.

The main thrust zone of the ECM (Sheremet et al., 2016a), traced west from the Chatyr-Dag massif ('yayla') (WCM) (Fig. 2 and Fig. 5) is displaced by strike-slip system faults (Gintov et al., 2003; Gintov 2005; Gonchar, 2005, 2013), which cross the elevated part of the WCM with TFm (termed the 'Crimean structural high') (Fig. 5) (Byzova 1980; Saintot et al., 2006). The results of the structural analysis (Hippolyte et al., 2018; Murovskaya et al., 2018) allow the definition of these structures as Cretaceous normal faults reactivated during the Cenozoic compressional tectonic stage as strike-slip faults (Balavadze et al., 1968; Gobarenko et al., 2016; Hippolyte et al., 2018; Murovskaya et al., 2018; Sheremet et al., 2016a;). The thrust zone could be traced to the west of the Cape Ay-Todor (Fig. 1B, Fig. 5) where the Jurassic limestones overthrust the latest Jurassic–Early Cretaceous flysch sediments (Oszczypko et al., 2017; Sheremet et al., 2016a).

In addition to the compression structures formed in the Cenozoic, in the nummulite limestones of the Middle Eocene, we noted numerous synsedimentary normal faults with a ~SW–NE trend (Fig. 6), which evidence extension during the deposition of the Eocene succession.

4. Localization of deep structures, active at present

4a. Low magnitude seismicity

Usually, the microseismicity (low magnitude, M ≤ 3) can lead to imprecise location data and is dependent on the type of recording array used, but it is possible to employ it for the approximate trace in-depth the structures (fault zones), determined by geological field data. By cumulations of seismic records in the vicinity of the CM we can evaluate the tectonic activity of these structures at present.

The study area (Southern Crimea and adjusted northern margin of the BS) is characterized by seismicity within the Crimean Seismic Zone (CSZ) located northwestward of the seismically active Greater Caucasus orogen. The CSZ is defined by the earthquakes of M=3–5 with foci in the crust and upper mantle and abundant weak seismicity (M ≤ 3) (Yanovskaya et al., 2016; Gobarenko et al., 2016).

We studied the seismicity of the CSZ using data from weak earthquakes registered on- and off-shore in this region (the Crimean Seismic Zone, CSZ) during the period of 1970-2013 by nine permanent seismic stations of Crimea seismic network (shown by black triangles on Fig. 7A) on the Crimea Peninsula (at Feodosiya (FEO), Yalta (YAL), Simferopol (SIM), Sevastopol (SEV), Alushta (ALU), Sudak (SUDU), Kerch (KERU), Tarhankut (TARU), Donuzlav (DON)) and by the Anapa station on the Caucasian coast (Pustovitenko, 1991; 2010).

To specify the parameters of the earthquake foci (location and depth), relocation procedure has been used for the P- and S-wave arrivals at all permanent stations of Crimea seismic network. The relocated seismic data on weak earthquakes were used not only to study the seismicity distribution (of the study region) and its relation with recent tectonics, but also (and primarily) to

restore the velocity structure of the crust of Southern Crimea and adjusted northern part of the BS by local seismic tomography (Gobarenko et al., 2017, Gobarenko and Yegorova 2020 a, b).

Within the study region (CM and adjusted northern part of the BS) two major subzones are recognized in the CSZ: Yalta–Alushta (Yuzhnoberezhnaya) and Sudak (subzones 2 and 3 in Fig.7a). The Yalta–Alushta subzone is known for generating a majority of densely located earthquakes with foci depth at 10-40 km, arranged by several stripes of N-S and NNW-SSE onshore in the CM and offshore on the BS continental slope within the area with $\lambda \approx 34°–34.7°$ (Fig. 7A; Gobarenko et al., 2016). The Yalta–Alushta subzone is limited from the west by an orthogonal subzone (Sebastopol subzone (labeled as subzone 1 in Fig. 7a)) characterized by scattered seismicity down to a depth of 40 km. The Sudak subzone, with a small number of earthquakes clustered at the continental slope to a depth of 35 km, shows a tendency to plunge northward (below the CM). The deepest earthquakes (with focal depths in the range 70–90 km depth) occur below the Kerch Peninsula shelf and Caucasus coast (subzone 4 in Fig. 7a) with foci deepen northwards (Gobarenko et al., 2016; 2017).

We used the data of weak seismicity ($M \leq 3$) (Fig. 7a) to interprete the location of tectonic structures depending on the location of cumulative clouds of hypocentres (Fig. 8) placed along the geological cross-sections and transects we constructed (Fig. 4). As it is seen in Fig 7a, two active tectonics provinces (Gobarenko et al., 2017, 2016), Yalta–Alushta (Yuzhnoberezhnaya) and Sudak, are separated by the ASFZ. Transects a and b (Fig. 8) illustrate the distribution of earthquake loci in the eastern part of the Sorokin Trough. It is possible to see domains with ruptures that could correspond to the south verging fault zones (Fig. 8B). These structures are diagnostic for proposing underthrusting along the Crimea–GC deep thrust fault (Gobarenko et al., 2017, 2016). The faults along the external zone of GC are interpreted as Cenozoic reverse faults and

thrusts (Mosar et al, 2010; Tan & Taymaz, 2006). A sub-horizontal dashed line on Fig. 8 could correspond to a decollement level that marks the underthrusting between the BS and the CM, interpreted by Yegorova et al. (2018) based on DSS data, as an inversion zone in the upper crust with decreased velocities.

The images of the distribution of earthquake loci along the western geological cross-sections (Fig. 8A) do not allow the definition of the vergence of structures. The hypocenters seem to be distributed uniformly along the represented cross-sections, which could correspond to vertical structures (faults) parallel to the cross-section to be viewed, possibly, strike-slip faults. This complex spread of loci could account for the impact of NW–SE Early–Middle Jurassic and Cretaceous normal faults (Hippolyte et al., 2018), which were reactivated as strike-slip faults during the Cenozoic compression (Fig. 5; Gobarenko et al., 2016). One of these faults, the WCF that bounds the MBSR from the west, is well traced by loci as a NW–SE trending zone off-shore western Crimea (Fig. 8) (Krezsek et al., 2017; Nikishin et al., 2015a, b; Okay & Nikishin 2015; Tari et al., 2015, Yegorova et al., 2018). Its activation explains the widespread occurrence of earthquakes with strike-slip mechanisms westward of the CM (Gobarenko et al., 2016). Additionally, the character of loci distribution in the western BS area could be explained by the complex composition of the crust of the CM, revealed by seismic methods to be high-velocity heterogeneities and deep intrusive bodies, with the largest one located within the ASFZ (Yegorova et al., 2018).

4b. Focal mechanisms of the strongest earthquakes

The existence of the southverging fault zones is supported by the focal mechanisms of the strongest earthquakes (Mw>3) defined during 1927–2016 (Gobarenko et al., 2016 Pustovotenko and Eredzhepov, 2017) for CSZ (see *Table* 1, Fig. 7, and Fig. 8: A on profiles SR and NM, B on profile BA).

Events 7, 14, 31 correspond to reverse, 1, 15 – to normal, 6, 8 – to thrust mechanisms (Fig. 7). Events 4, 13, 21, 26, 27 show reverse and strike-slip components. The Andrusov ridge is characterized by a compression deformation regime, revealed in foci 20, 23, and 29. The focal mechanisms of Kerch-Taman foci cluster correspond mainly to reverse faulting. Mechanisms with a significant strike-slip component are located in the western part of the CSZ. The orientations of the compressional axes in the foci indicate shortening with transpressive component for the current tectonic stage.

We projected 7 foci of the earthquakes (Table 1) located within north BS continental margin and the CM on the closest geological cross sections (Fig. 8) and plotted their focal mechanisms. Within ECM three earthquakes (8, 14, 16) are in the continental crust with depths ranging from 20km to 40km (Fig. 8B). One of nodal planes of 14 is steeply (60°) northward dipping, and the focal mechanism evidence a reverse component on this plane which could correspond to the rupture along a main thrust (see figure 8B, a). The mechanism of the earthquake number 8 with a low angle nodal plane of landward dipping is located almost at the bottom of foci cluster's distribution and possibly corresponds to a detachment within the lower crust. In WCM, the projected hypocenters (6, 7, 31) are shallower and are located along the northward dipping rupture zone of low angle (Fig. 8A).

The focal mechanisms of the strongest earthquakes defined in the CM evidence an active southward thrusting, with some transpressive component, within the continental crust. The

geometry of the compressional structures in the vicinity of the CM that correspond to an underthrusting process within the southern margin of Crimea (Gobarenko et al, 2015, 2016) allows us to define the ShR as the indenter. Considering crustal heterogeneities, mentioned earlier, the ShR could be responsible for the formation of compressional structures of the CM and of its off-shore.

5. New Palaeo-Stress Data of the Eastern Crimean Mountains

As the results of structural analysis of the CM presented in works of Sheremet et al. (2014, 2016a, b) and Hippolyte et al. (2018) allowed synchronizing the tectonic stages defined for the CM and the GC, the use of the palaeo-stress analysis can allow the clarification of these tectonic shortening/extension event trends. A new fault kinematic analysis was applied to rock-complexes of the ECM using new age dating from the TFm in this region (Ivanik et al., 2013; Nikishin et al., 2017; Oszczypko et al., 2017; Popadyuk, 2011, Popadyuk et al. 2013a, b; 2014; Sheremet et al., 2014, 2016a); the outcrops with the flysch rocks of the TFm in the ECM are limited in space, because they are covered by the Lower Cretaceous flysch deposits. Therefore, compressional deformations that some authors (e.g. Saintot et al., 1999) have attributed to Cimmerian orogeny are corresponding to Cenozoic compression and must be reconsidered in light of the collisional processes south of the BS, which are linked to the closure of the Neotethys oceanic plate from the Late Cretaceous to Early Palaeocene in the east (region of Lesser Caucasus; Rolland et al., 2012; Sosson et al., 2010) and from the Palaeocene to Eocene in the west (region of Pontides; Espurt et al., 2014; Lefebvre et al., 2013; Okay & Nikishin, 2015; Okay & Tüysüz, 1999; Robertson et al., 2014; Sengör et al., 2003).

5.1 Method and data

We based our analysis on outcrop scale fault slip data (919 slickensides and calcite slickenfibres measured in 50 outcrops) (Table 2, Fig.9). For the processing and interpretation of the slickensides, the kinematic method was applied using the 'Win-Tensor' program (Delvaux & Sperner, 2003) along with 'Stereonet' (Allmendinger et al., 2012) to build the stereograms, considering the methodology of Sperner & Zweigel (2010), Hippolyte et al. (2012), and Lacombe (2012).

For each site, we determined the palaeo-stress regime (reverse, normal, or strike-slip). Since the main tectonic events are identified quite precisely according to the results of structural analysis (macroscale) (Hippolyte et al., 2018; Sheremet et al., 2014, 2016a,b), the main purpose of using microtectonic analysis (palaeo-stress) was to clarify the directions of the principal stresses (palaeo-stress field) of known tectonic stages. Considering that our data were selected exclusively from the Mesozoic rocks, it is not possible to determine the relative age of existing tectonic regimes, for example, in the Cenozoic solely by the method of fault slip data inversions. Therefore, to extract Cenozoic palaeo-stress, we compared our results with those in published materials (e.g. Gintov, 2005; Hippolyte et al., 2018; Murovskaya et al., 2012, 2014, 2018; Saintot et al., 1999; Saintot and Angelier, 2000).

5.2 Resulting palaeo-stress regimes

5.2.1 Reverse regime

For the majority of sites in the ECM, we obtained a large display of reverse regimes with σ_1 (greatest compressive stress) trending NW–SE (western part of the sector) and NNW–SSE (eastern part of the sector) (Fig. 9A, Table 2). This supports the results of the structural analysis both onshore and offshore, including the trends of thrust zones and fold axes, S_0-dipping analyses on

stereograms, and the orientation of the thrust front (Sheremet et al., 2016a, b) (Fig. 9A). The trends of compression vary locally; in the area of Sudak city, it is a N–S compression (site 24, Fig. 9A), and in the vicinity of Ordzhonikidze, the trend of compression was defined as NNE–SSW by the fold axis trend analysis (Fig. 6A).

5.2.2 Strike-slip regime

The analysis of the structural patterns in the ECM (Sheremet et al., 2016a) (Fig. 2, Fig. 5) revealed several NE–SW and NNE–SSW trending faults with left-lateral strike-slip motion. These strike-slip faults cut and displaced several thrusts and thrust fronts in several localities (Fig. 2, Fig. 5) (Sheremet et al., 2016a). In the rocks of the Miocene–Pliocene units, we defined a right-lateral displacement along the NW–SE strike-slip faults. A strike-slip regime with a N–S orientation of σ_1 and an E–W orientation of the σ_3 was also detected in the very eastern part of the ECM (Fig. 9B; Table 2). According to the structural data, these strike-slip faults postdate the main thrust of the CM. We suggest that the N–S trend of the compression characterizes the youngest tectonic stage of the ECM.

At several locations (stereodiagrams 17, 57, and 65; Fig. 9B) we obtained strike-slip regimes with an almost E–W orientation of compressional axis σ_1. Sites 17 (open pit in Grushevka village) and 57 (limestone close to the Karadag region) (Fig. 9B) belong to a very large Upper Jurassic limestone massif, which could possibly be a block. Site 65 is located near Hauterivian limestones close to Mezhgorye village.

The strike-slip regimes with a NW orientation of σ_1 were obtained mostly in the western part of the ECM (stereodiagrams 43, 37; Fig. 9B, Photo in Fig. 6B). A strike-slip regime with a clear N–S orientation of σ_1 and an almost E–W orientation of the σ_3 was detected locally for the eastern part of the ECM (stereodiagram 3; Fig. 9B).

5.2.3 Normal regime

North of Alushta, the Demerji Mountain (Fig. 1B, 2) (site 50; Fig. 9C, Table 2) is located in the area of the ASFZ, which is considered to be one of the major deep structures of Crimea (Byzova 1980; Gintov, 2005; Khain 1984; Koronovskiy 1984; Morgunov et al., 1979; Saintot et al., 1999; Sollogub 1977; Volfman 2015) it shows a NE–SW direction of extension. The same trend of extension was identified close to Dachnoe village (site 22, north of Sudak city) (Fig. 9C, Table 2).

We identified an E–W extensional regime (sites 61 and 56; Fig. 9C, Table 2) in several places of the ECM. The structural data observed around the village of Druzhnoe reveal a normal fault with an almost N–S strike near the Dolgorukovskaya Yayla (Fig. 6C). The hanging wall is Upper Jurassic–Berriasian limestone whereas the western foot-wall of this fault is red breccia. In the ECM, the red calcareous conglomerate-breccia of Early Cretaceous age overlies the Upper Jurassic limestones north of Karabi Yayla (Sheremet et al., 2016a), which is likely related to the normal faulting that occurred during Cretaceous rifting.

The results also indicate extensional regimes with a NW orientation for σ_3 (Fig. 9C), which is perpendicular to the orientation of the compressional structures.

6. Meso-Cenozoic stress history of the CM: integrated results and discussion

The structural frame within the CM during the late Mesozoic and the Cenozoic is pre-determined by pre-existing structures (Saintot et al., 1999; Sheremet et al., 2016b.). The chronological summary of the tectonic stages and stress regimes obtained for the whole CM in this section based on the structural and stratigraphic control of the macro scale data, taking into account published studies containing new results on the dating of rocks and structures of the CM (Hippolyte

et al., 2018; Murovskaya et al., 2012, 2014, 2018; Saintot et al., 1999; Saintot & Angelier, 2000) with references to the subduction–collisional geodynamic settings of the BS–GC domain.

Extensional deformations during the Cretaceous.

A large quantity of data on the palaeo-stress field in the CM is related to the normal-faulting. In the WCM, the stress fields are characterized by N–S and NNE–SSW extensions (Hippolyte et al., 2018; Saintot et al., 1999;), whereas in the ECM, the orientation of σ_3 corresponds to a NE–SW trend (Fig. 9C), which is supported by Saintot et al. (1999). In that study, the authors identified NW–SE normal faults (N140) parallel to the strike of the ASFZ and active during the Cretaceous. For the WCM, other than the NW–SE strike of the faults, Saintot et al. (1999) also pointed out the WNW–ESE (N105 and N110) strike as important Cretaceous faults with N–S σ_3 orientation.

We suggest that the difference in the structural pattern in the WCM and the ECM during the Cretaceous could be explained by the specific position that the territory of the future peninsula occupied. As mentioned, the eastern part of the peninsula was influenced by the extensional NW–SE faults of the 'Caucasian' trend since the Early–Middle Jurassic, the ASFZ is one of these faults that separates the ECM and WCM (Byzova, 1980; Lebedev & Orovetskiy, 1966; Saintot et al., 2006). In the backdrop of the GC basin's loading in the vicinity of the ECM, an eastward deepening slope was formed during the Hauterivian–Aptian, which was characterized by the distribution of olistostromes with the olistoliths of the Upper Jurassic–Berriasian limestone, flysch, grainstones, conglomerates, and coarser turbiditic material (Dobrovolskaya & Salman 2008; Ivanik et al., 2013; Nikishin et al., 2015c; Oszczypko et al., 2017; Popadyuk et al., 2014; Sheremet et al., 2016a; Sosson et al., 2016) (Fig. 10). The results of DSS and local seismic tomography (Yegorova et al., 2018) show an eastward dipping Mesozoic surface in the vicinity of Feodosia, which could be

interpreted as a normal fault; this occurred due to the deepening of the GC (Fig. 10). The WCM seem to be uplifted (Afanasenkov et al., 2007; Byzova, 1980; Dubois de Montpereaux, 1843; Lebedev and Orovetskiy, 1966; Lebedinskij & Shalimov, 1967; Leshchuh et al., 1999; Muratov, 1960; Nikitina et al., 1999; Saintot et al., 2006; Sludskij, 1917; Zaika-Novackiy, 1981; Zolotarev, 1968; Yamnychenko, 1969;) compared to the ECM (see stratigraphic logs Fig. 3). The data indicates a later onset (Valanginian–Barremian) of the rifting stage in the WCM (as opposed to the late Jurassic rifting in the ECM), which has been related to the reactivation of the NW–SE ASFZ (Murovskaya et al., 2014; Hippolyte et al., 2018; Saintot et al., 1999).

The Cretaceous normal faults with the ~E–W strike in the WCM (Murovskaya et al. 2012; 2014; Saintot et al., 1999) may have occurred owing to the reactivation of pre-existing Triassic normal faults; the major extensional features in the overriding plate may have formed due to the subduction of the palaeo-Tethys (Barrier et al., 2018). These structures extend to the west from the CM up to the Northern Dobrogea where they are limited by another deep crustal NW–SE fault zone: the Trans European Suture Zone (e.g. Hippolyte et al., 2002; Săndulescu et al., 1978; Seghedi et al., 2005; Winchester et al., 2006). These similar ~E–W structures have also been observed in the Balkanides (Bergerat, et al., 2010; Doglioni et al., 1996, Georgiev et al., 2001). Regarding the CM, the ~E–W faults are notable on the DOBRE-2 CDP profile (Sydorenko et al., 2016) and are revealed by seismic interpretation to the north of the CM (e.g. Khriachtchevskaia et al., 2010). The reheating of the Eurasian margin during the Cretaceous (Aptian–Albian) as a result of the subduction of the spreading centre of the Ankara–Erzincan–Sevan–Akera (AESA) oceanic back-arc basin (Sosson et al., 2016) (Fig. 1a) and/or the Tethys's slab roll-back (Stephenson & Schellart, 2010) reactivated these Triassic ~E–W and middle-Jurassic NW–SE structures as normal faults. It is likely that the ShR and MBSR became significant morphological 'Caucasian' units of the BS's

palaeorelief and were separated during the aforementioned period (Sheremet-Korniyenko, 2016). This could explain why two oblique directions of faults formed during the Cretaceous within the CM (Fig. 11).

Early Cenozoic compression in the CM

Since the end of the Maastrichtian, some areas of the BS domain underwent compressional tectonics due to the final closure of the northern branch of the Neotethys (Barrier and Vrielynck, 2008; Dercourt et al., 1986; Okay & Tüysüz, 1999; Sosson et al., 2010; 2016; Yilmaz et al., 1997; Zonenshain & Le Pichon, 1986), however, the CM underwent the shortening since the Palaeocene (Sheremet et al., 2016a,b). A reverse regime with σ_1 trending NW–SE is the main one that we defined for the CM. The NW–SE direction of the σ_1 compressional axis prevailed during the formation of the main compressional structures (thrusts, folding) onshore and offshore the CM and in the south-western part of the KP (fig.5) (Gerasimov, 1992; Saintot & Angelier, 2000; Sheremet et al., 2016a, b).

By the time when shortening begins, the lithosphere and crust of the Eastern BS was not cooled enough to had sufficient mechanical strength to deform the continental Crimean margin: after the rifting ("lithosphere stretching") it, was composed of horsts and grabens represented by thick and stretched continental crust with different rheological behaviour. Under the compressional settings, deformation of more brittle behaviour were performed between the blocks with thick continental crust (e.g. the Crimean margin and the ShR), while the very streched parts of the BS lithosphere could experiense ductile deformations. It could facilitate the inversion of the thick continental blocks of the crust, including their displacement along the faults, that consequently, could contribute to an extension, especially in the central parts of the BS's sub-basins. The latter,

however, were interpreted as the real opening of the BS during the Cenozoic (e.g. Maynard and Erratt, 2020).

The ShR (relic of the continental crust) was presumably perpendicular to the future CM at this time (Fig. 11). As a result of the collision to the south (Taurides–Anatolides–South Armenian continental microplate, fig. 1a)) with Eurasia, (e.g. Sosson et al., 2010, 2016), the ShR acted as an indenter and triggered the inversion of the pre-existing (Triassic–Cretaceous) extensional structures. Consequently, the pre-existing normal faults of different strikes underwent reactivation progressively (Hippolyte et al., 2014, 2018; Murovskaya et al., 2018; Sheremet et al., 2016b; Sosson et al., 2016). Shortening due to underthrusting of the Cretaceous extended crust beneath the CM caused intensive erosion of the raising CM and resulted in thick Palaeocene–lower Eocene sequences that filled the Sorokin Trough (Sheremet et al., 2016b; Sheremet-Korniyenko, 2016; Sydorenko et al, 2016). The presence of thick Palaeocene sediments in the Sorokin Trough is correlated with the Subbotina well log (Sheremet et al., 2016b) and attested by an isobaths map showing the top of the middle Palaeocene terrigenous sediments in the Sorokin Trough (Tari et al., 2011). The time estimation of the early shortening within the CM since the Palaeocene is in agreement with the results of the apatite fission-track thermochronology (Panek et al., 2009) that interpreted the uplifting of the CM within the Early–Middle Eocene. The absence of the Uppermost Cretaceous–Palaeocene sediments in the Simferopol area—likely the result of the uplifting of the CM—further supports this correlation (e.g. Zaika-Novackiy et al., 1976).

Normal regime with NW–SE orientation of σ_3

We obtained the normal regime with NW–SE orientation of σ3 in the Mesozoic rocks which coincides with the extension trend extracted from normal synsedimentary faults in the Eocene limestones (Fig. 6). The results show that their formation postdates the compression that

began in the Palaeocene. The NW–SE orientation of σ3 in the extensional regimes is parallel to the NW–SE direction of the σ_1 compressional axes responsible for the formation of initial folds and thrusts in the CM (Sheremet et al., 2016a). The NE–SW strike of the extensional structures is also noticeable in the scarp of the present day seafloor of the ECM's offshore slope (Saintot et al., 1999). We propose that the extension occurred synchronously to the uplifting of the CM. Data from the wells of the Azov and Black Seas and the Odessa shelf show terrigenous lithologies for Eocene sediments (shales, siltstones, and marls) (Khriachtchevskaia et al., 2010), and their age-analogue deposits within the CM are the nummulitic limestones. This indicates that the uplifted position of the CM may be due to shortening, thus causing some compression structures to become normal faults under a gravitational instability (relaxation of compressional stresses). The occurrence of synsedimentary normal faults in the Middle Eocene nummulitic limestones is related to the formation of a piggyback basin to the north of the rising folds and thrusts of the CM.

Since the Late Eocene, shortening in the BS has been reactivated (Bosworth and Tari, 2020; Khriachtchevskaia et al., 2010; Nikishin et al., 2015c; Sheremet et al., 2016a, b; Stovba et al., 2009; Tari et al., 2018).

Reverse regime with NS orientation of σ_1

The structures formed under the N–S shortening were observed in the Sudak-city area and KP (Saintot and Angelier, 2000). Previously, a N–S compression extracted from the flysch-rocks of the Sudak-city area was interpreted as shortening during the Cimmerian orogeny (Saintot et al., 1999). As the results of the new dating of these rocks show the Early Cretaceous age (Ivanik et al., 2013; Oszczypko et al., 2017; Popadyuk et al., 2013a, 2014; Sheremet et al., 2014, 2016a;), we suggest that the palaeo-stress reconstructions reflect post-Cretaceous deformations, which have superimposed on the early Mesozoic Cimmerian shortening. The structural data at the micro and

macro scale of the KP (Kazantcev & Bekher, 1988; Saintot et al., 1999; Saintot & Angelier, 2002) shows that the N–S trend for σ_1 characterizes the latest phase of compression. The folds and thrusts of the E–W strike in the Middle Pliocene marine sediments of the KP overthrust those formed under the NNW–SSE compression (i.e. ENE–WSW oriented folds of the Parpatch Ridge, south-west of the KP) (Saintot & Angelier, 2002). This means that changes in the trend of shortening (from NW–SE to N–S) post-date the Middle Pliocene deposition (Fig. 11). This conclusion is supported by the changes in the deformation regime within the CM (from reverse to strike slip); the σ1, extracted from strike-slip faults in the youngest marine sediments of Miocene–Pliocene of the CM, has a N–S trend (Gintov, 2005; Gonchar, 2013; Murovskaya, 2012; Saintot et al., 1999; Volfman, 2014).

Strike-slip regime postdating the reverse regime in the CM

As shown on the structural map (Fig. 2, Fig. 5) and by the data, the strike-slip faults cut several thrusts and displaced the thrust fronts of the CM.

The strike-slip regime in the WCM (Gobarenko et al., 2016; Murovskaya et al., 2014; 2018) is characterized by a NE–SW orientation of σ_1. Murovskaya et al. (2017, 2018) noted the presence of Cretaceous normal faults with a NW–SE strike in the WCM, which were reactivated during the Cenozoic shortening as dextral strike-slip faults (based on microtectonic measurements in outcrops). Some data correspond to a strike-slip regime with an almost E–W orientation of the compressional axis σ_1, for the ECM (stereodiagrams 17, 57, and 65; Fig. 9B). We derived two hypotheses from these results: first, these three sites may correspond to olistoliths within the Lower Cretaceous flysch, and therefore, the data are not representative due to rotation of the blocks during the deformation; second, these strike-slip faults may result from the reactivation of local structures.

We argue that the changes in the principal stresses of the regimes and trends during the Cenozoic can be explained by the northward plunging of the Eastern BS extended crust (Yegorova et al., 2010) that contain relics of the continental crust, such as the ShR, under the Crimean–Caucasian margin (Fig. 11). Because of the oblique position of the ShR (NW-SE orientation) to the Crimean margin, its gradual relocation to the north and gradual plunging trigger the uplifting of the CM and then, with a delay, the uplifting of the KP in the Late Pliocene (Muratov, 1960; Pivovarov, 1984; Saintot et al., 1999;). The northward plunging of the Eastern BS, containing the oblique relics of the continental crust as ShR, could explain a clockwise progressive trough formation south of the CM: the formation of the Sorokin Trough began in the Palaeocene as shown by the age of the terrigenous sediments that infilled the trough (Sheremet et al., 2016b; Tari et al., 2011). The Kerch–Taman Trough is infilled with deposits of the Palaeocene–Eocene (Sheremet et al., 2016 b; Tari et al., 2011), while the Tuapse Trough formed during the Eocene–Oligocene (Robinson et al., 1996) (Fig.11). Thus, due to this obliquity, the shortening within the CMs' margin also acquires to a clockwise progressive change in its direction (from NW to NNW and even to N–S). Probably, this northward relocation of the ShR is reflected as changes in the regime from reverse to strike-slip that was previously suggested by Saintot et al., 1999.

A second Cenozoic stage of compression was identified as having a somewhat periodic character (Datcenko, 2012; Khriechtchevskaya et al., 2010; Stovba et al., 2012; Stovba et al., 2013). The gradual uplift of the ECM onshore is exemplified by several discontinuities and/or unconformities (depending on location) throughout the Miocene (Datcenko, 2012; Geology of USSR, 1969; Muratov, 1960; Muratov et al., 1969, 1973; Nikishin et al., 2006). This interruption in the deposition probably corresponds to the gradual plunging of the ShR under the CM (underthrusting of the ShR). However, in the WCM, the transpressional regime identified by

Murovskaya et al. (2018) is the result of the reactivation of Cretaceous normal faults parallel to ASFZ as strike-slip faults (for example strike-slip WCF) (Murovskaya et al., 2018; Saintot et al., 1999;). A strike slip regime was also obtained by the study of local mechanisms of earthquakes in the Western CSZ (Gobarenko et al., 2016; Murovskaya et al., 2017, 2018) (Fig. 8). We suggest that their occurrence is linked to the stress field exerted on the MBSR by a strong sub-oceanic, or more probably, a highly extended continental crust of the Western BS under the ongoing compression, which occurred in the BS since the Tethys closure.

7. Conclusions

Tectonic evolution of the CM should be regarded as one of the consequences of the long-standing subduction of the Tethys oceanic plate under the Laurasian–European margin at least since the Early Mesozoic (e.g. Barrier et al. 2018; Barrier & Vrielynck, 2010; Stampfli & Borel 2002; Stampfli et al., 2001; van Hinsbergen et al., 2019). Throughout this process, in the overriding plate formed two systems of faults 1) of ~E–W trend, their origin could be linked to the extension most likely related to a Triassic rifting through the normal faults in the Eurasian basement, which were likely parallel to a subduction front (Barrier et al., 2018); and 2) ~NW–SE-trending faults originating from the Early–Middle Jurassic due to opening of the Caucasian domain in a back-arc setting (Zonenshain and Le Pichon, 1988). Both systems of faults were reactivated during the Early Cretaceous, which could explain the obliquity of cretaceous structures within the CM area. The pre-existed structures continue to impact the current seismicity distribution.

Our results of micro-/macro-scale structural analysis allow to identify the following periods of deformation and tectonic regimes with the palaeostress field orientations for the CM since the Cretaceous and to elucidated the possible mechanisms of tectonic development.

1. The Cretaceous extension is evidencing by the NE-SW and ENE-WSW orientation of σ3 obtained for the sites located along NW and NNW faults. Extensional deformations (the σ3 orientation of NE–SW) started earlier in the ECM (Earliest Berriasian) (Sheremet et al., 2016a); the structural framework was parallel or very close to a NW–SE strike of the ASFZ, since it was most likely influenced by the NW–SE 'Caucasian' faults, related to a back-arc basin formation of the GC initiated since the Early–Middle Jurassic;. then, during Valanginian–Barremian the rifting process propagated to the SW and reached the WCM (Hippolyte et al., 2014, 2018; Murovskaya et al., 2014,) along the NW–SE and WNW–ESE trending normal faults.

2. The shortening within the CM that started in the Palaeocene due to the collision of the Tauride–Anatolides–South Armenian microplate (Fig. 1, No *8 in legend*) and the Eurasian margin, exhibited a NW–SE trend, which produced the main compressional structures as folds and thrusts in the CM and KP of the NE–SW strike. Due to the thrust propagation to the SW (present offshore), the synsedimentary normal faults occurred on the back of the thrust unit (in the middle Eocene limestones). Thus, a piggy-back basin formed on top of the moving thrust sheets, producing a local extensional regime in the basin (within the CM) in the Middle Eocene (the NW trend of σ1 in the frontal part is parallel to the trend of σ3 in the piggyback basin). At one point during the Middle Pliocene, the trend of compression in the most eastern part of the CM and KP changed to N–S, which was then followed by changes in the tectonic regime from reverse to strike-slip in the CM.

We explain these trend changes, shortening, and tectonic regime evolution by the northward plunging of the extended BS crust, containing continental relics such as ShR in the oblique position to a general N-S direction of shortening, under the CM margin and, consequently, by gradual relocation of the ShR to the north. Owing to this, the Crimean Peninsula experienced

underthrusting and the KP uplifted after a delay. We suggest that underthrusting since the Palaeocene triggered the clock-wise formation of troughs south of the CM: the Sorokin, Kerch Taman, and Tuapse troughs in the Eocene–Oligocene.

Acknowledgments

This study is a final part of the PhD work of Yevgeniya Sheremet (defended in 2016) supported by the Embassy of France in Kyiv, Ukraine. We are grateful to the participation of the IRG project 'South Caucasus Geosciences' of the CNRS/INSU (GDRI 776), which made this research possible. We express special gratitude to the DARIUS programme for its funding. The authors also confirm that the data concerning the seismicity stored in the Supplemental Material_seismicity Data Set SI, were uploaded to Figshare. The data are contained in published sources cited in the References (Hippolyte et al., 2018; Murovskaya et al., 2014, 2018; Oszczypko et al., 2017; Popadyuk et al., 2013a, b; Sheremet et al., 2016a, b; Yegorova et al., 2018). We are also grateful to Marc Hässig for his comments on the earlier version of the manuscript and to reviewers: Alexandr Kitchka, Gabor Tari, Randell Stephenson and to anonymous reviewer for their efficient work and rational criticism leading to improve the manuscript, and mainly to Randell Stephenson who helped in its reorganisation.

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Figure Captions

Fig. 1: A - Structural sketch map of the Alpine belt in the Middle East sensu lato (modified from Sosson et al., 2016). The study area is framed by the red rectangle. 1: Platform, 2: European margin (Pontides), 3: European margin including magmatic arc (Pontides, Somkheto–Karabakh), 4: Lesser Caucasus units including ophiolites, 5: Anatolian ophiolites, 6: Metamorphic massifs, 7: Sakarya accreted terrane, 8: Taurides Anatolides South Armenian continental microplate, 9: Iran accreted terrane during Eo-Cimmerian orogeny, 10: Taurides Anatolides with obducted ophiolites and Peri-Arabic (Lycian nappes) including ophiolites, 11: Highly extended continental crust and oceanic crust, and 12: Mesozoic inverted basins (Greater Caucasus–Crimean Mountain Belt)

CACC : Central Anatolian crystalline complex, KM : Kirsehir massif, MM : Menderes massif, SM : Sakarya massif, IAES : Izmir–Ankara–Erzinkan suture, AESA : Ankara–Erzinkan–Sevan–Akera suture, NAF : North Anatolian fault, EAF : East Anatolian fault, GC : Greater Caucasus, LC : Lesser Caucasus, ATB : Adjara–Trialeti basin, SA : South Armenian block, C: Carpathian, D: Dobrogea, M: Moesia, SR: Shatsky Ridge, TT: Tuapse Trough, MBSH:

Mid Black Sea High, K: Khoy ophiolites, V : Van Lake, S : Sevan Lake, and R : Rezaiyeh Lake

B: Structural schema containing major tectonic domains (coloured and labelled), regional tectonic units (labelled and separated by black lines) were compiled following Yegorova and Gobarenko (2010), Khriachtchevskaia et al. (2010), and Yegorova et al. (2018)

GC: Greater Caucasus, Lomonosov M: Lomonosov Massif, North Azov T: North Azov Trough, MBSR: Mid Black-Sea Ridge, CM: Crimean Mountain domain, TT: Tuapse Trough. The CM and the Kerch Peninsula (KP) are shown with main structures after Saintot et al. (2000), Sheremet et al. (2016), and Hippolyte et al. (2018) (see in details Fig. 5). Dashed lines correspond to fault zones: ASFZ: Alushta–Simferopol fault zone, WCF: Western Crimea Fault. White Line is a valley between the cities Simferopol (S) and Alushta (A). F: cape Fiolent, AT: cape Ay-Todor.

Fig. 2 Proposed compiled map of the CM based on a geological map (Ed. Muratov, 1967 from Popadyuk et al., 2013a) and structural geological map of the ECM (Sheremet et al. (2016a). The limit between them is shown by black dotted line. Red circles indicate locations of samples redated based on the analysis of foraminifera published in Popadyuk et al. (2013, 2014) as Albian in age. Black: the outcrop studied by Popadyuk et al. (2013). Yellow circles: locations of samples redated by nannofossils (from Sheremet et al., 2016a).

Fig. 3 Schematic stratigraphic logs of the Crimean Mountains supplemented from Sheremet et al. (2016a).

Fig. 4 Geological cross sections with interpretation through the WCM and ECM**.** Location of the cross-sections is shown in Fig. 5.

Fig. 5 Map of tectonic structures of the CM and of the KP juxtaposed on the shaded relief map of the CM and the KP and compiled map of the CM (Fig. 2). For the KP we used the map of Muratov (Ed. 1967) (the CMs' part of this map is shown in the side-bar). The cross-sections shown in Fig. 4.

Fig. 6 A: Example of the south vergence folding in the northern part of the Ordzhonikidze, cape between Koktebel and Feodosiya

B: Example of right-lateral strike slip fault in the Upper Jurassic limestones, Kokasan canyon, SW of Belogorsk, East of site 34 on Figure 4.

C: Normal fault in the westernmost part of the Dolgorukovskaya Yayla (Druzhnoye village). a: interpreted fragment of the general view b; f: fractures, accompanying the normal faulting, d: zoom-view on red carbonated material filled up the fractures.

D: Syndepositional faults in the Middle Eocene nummulitic limestones: a, b: Michurinskoe village (eastwards from Belogorsk); c-e: Belaya Skala (northwards of the Belogorsk).

Fig. 7 Hypocenters at different depths: (1) 0-15 km, (2)15-30 km, (3) more than 30 km; (4) CSZ subzones (1, Sevastopol; 2, Yalta-Alushta; 3, Sudak; 4, Kerch-Taman) (from Yegorova et al., 2018); (5) seismic stations (ALU, Alushta; ANN, Anapa; DON, Donuzlav; FEO, Feodosiya; KERU, Kerch; SEV, Sevastopol; SIM, Simferopol; SUDU, Sudak; TARU, Tarkhankut; YAL, Yalta); (6) isolines corresponding to depths of Cretaceous basement, km; (7) location of thrusts in the Sorokin Trough based on seismic line interpretation (Shermet et al., 2016b) and main structures of the CM (Fig. 5); (8) main fault zones: WCF and ASFZ; (9) location of the interpreted seismic

lines used for geological transects (Fig. 8) (from Sheremet et al., 2016b); (10) profiles of seismic records shown on Fig. 8 (number 4 in legend); and (11) focal mechanisms of strong earthquakes in the Black Sea northern coastal area (from Gobarenko et al., 2016). Colour of line (10) coincides with the colour of the numbers of focal mechanisms located along this line.

Fig. 8 Distribution of week seismicity records (M ≤ 3), within the Crimean Seismic Zone, CSZ) during 1970-2013 (from Yegorova et al., 2018) (**A**) along the geological transects through the ECM and the Sorokin Trough (from Sheremet et al., 2016b) and (**B**) along the geological cross sections through the WCM (from Fig. 4) and beneath the sea-bottom line as the prolongation of those cross-sections in the BS.

Legend 8A and B: 1: hypocenters of the seismic events; 2: sea-bottom line; 3: geological cross-sections (Fig. 4, Fig. 5), 4: profiles of seismic records, their location is shown on figure 7 (number 10 in legend), 5: presumable location of rupture zones. 6: focal mechanism in vertical cross-section, number of the seismic event located on the map of the figure 7.

Fig. 9. **A**: Local palaeo-stress tensor for NW-SE Palaeocene–Early Eocene and NS post-Miocene compressions imprinted in the ECM and Kerch Peninsula. The red plane in the stereo-diagrams corresponds to the bedding of the site. The characteristics of palaeo-stress are shown in the Table Different colours of the flashes mark different groups of the palaeo-stress regimes.

B: Local palaeo-stress tensors for the Pliocene–Quaternary time compression imprinted in the ECM and Kerch Peninsula. The characteristics of palaeo-stress is shown in the Table 2.

C: Local palaeo-stress tensors for the extensional stress-field of the ECM. The characteristics of palaeo-stress is shown in the Table 2.

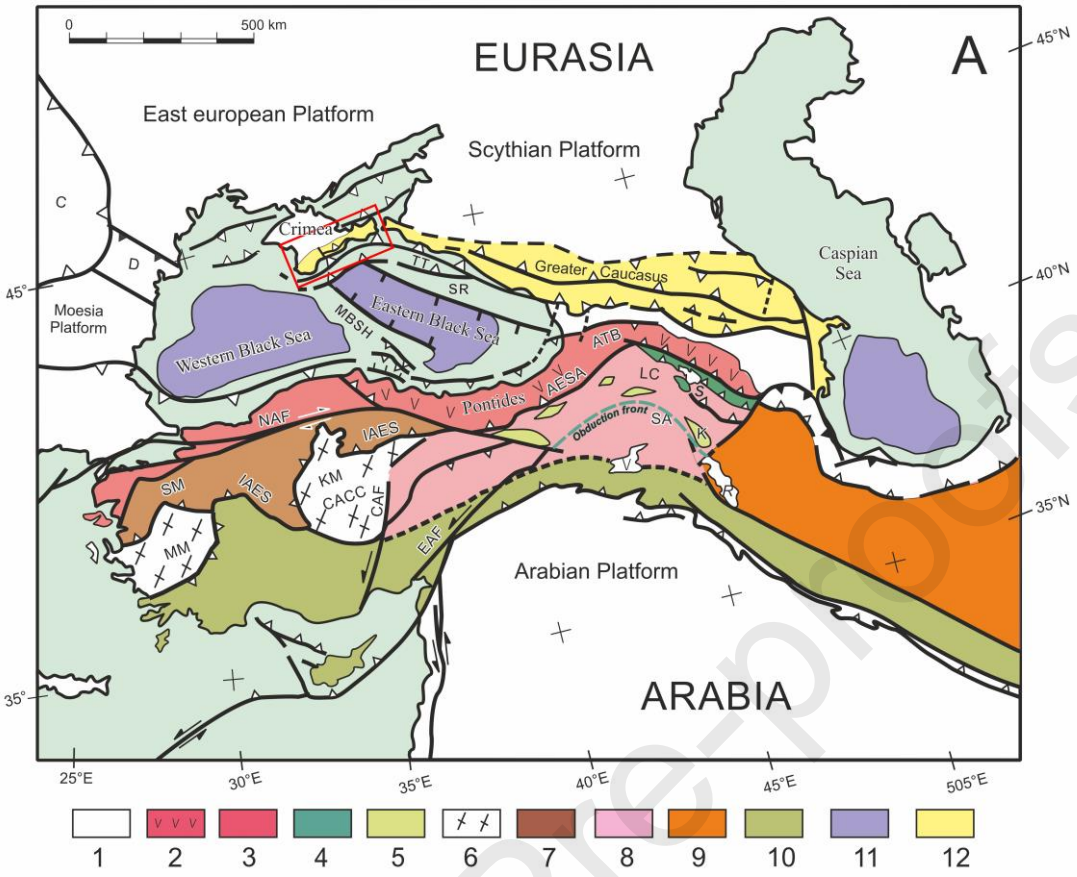
Fig. 10 Reconstruction of a WSW–ESE oriented cross section through the ECM for the Early Cretaceous.

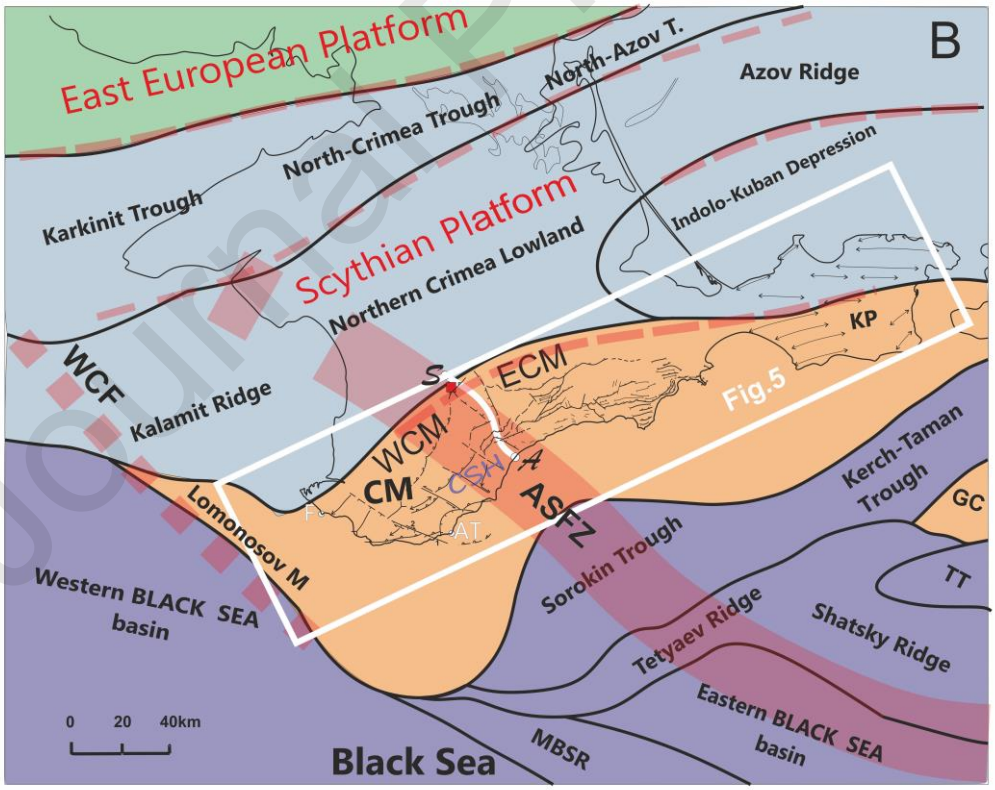
Fig. 11. Palaeoreconstructive sketch for Crimea during the Cenozoic N–S compression, showing the role of indenter in formation of the corresponding structures in the vicinity of the Crimean part of the northern Black Sea margin. Light grey faults correspond to the Triassic (or older) structures. Legend: SR: Shatsky Ridge, MBSR: Middle Black Sea Ridge, ASFZ: Alushta–Simferopol Fault Zone, PCFZ: Pecenega–Camena Fault Zone (prolongation of the Teisseyre–Tornquist zone), KT: Kerch-Taman Trough; TT: Tuapse Trough WCF: Western Crimea Fault, EBSB: Eastern Black Sea Basin; WBSB: Western Black Sea Basin. **B:** formation of the Sorokin Trough; **C:** Undethrusting beneath the northern margin of the Black Sea

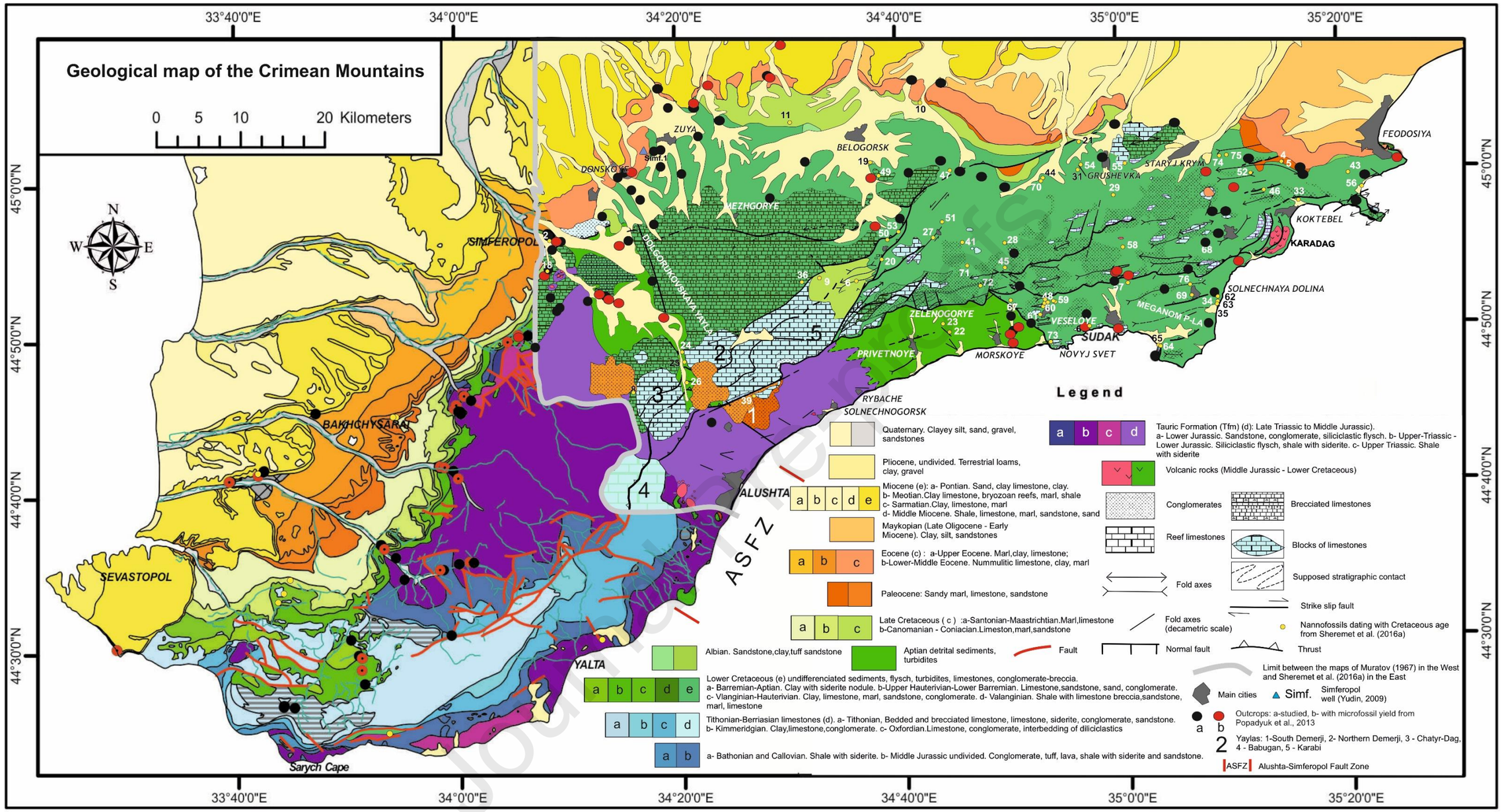
Tables

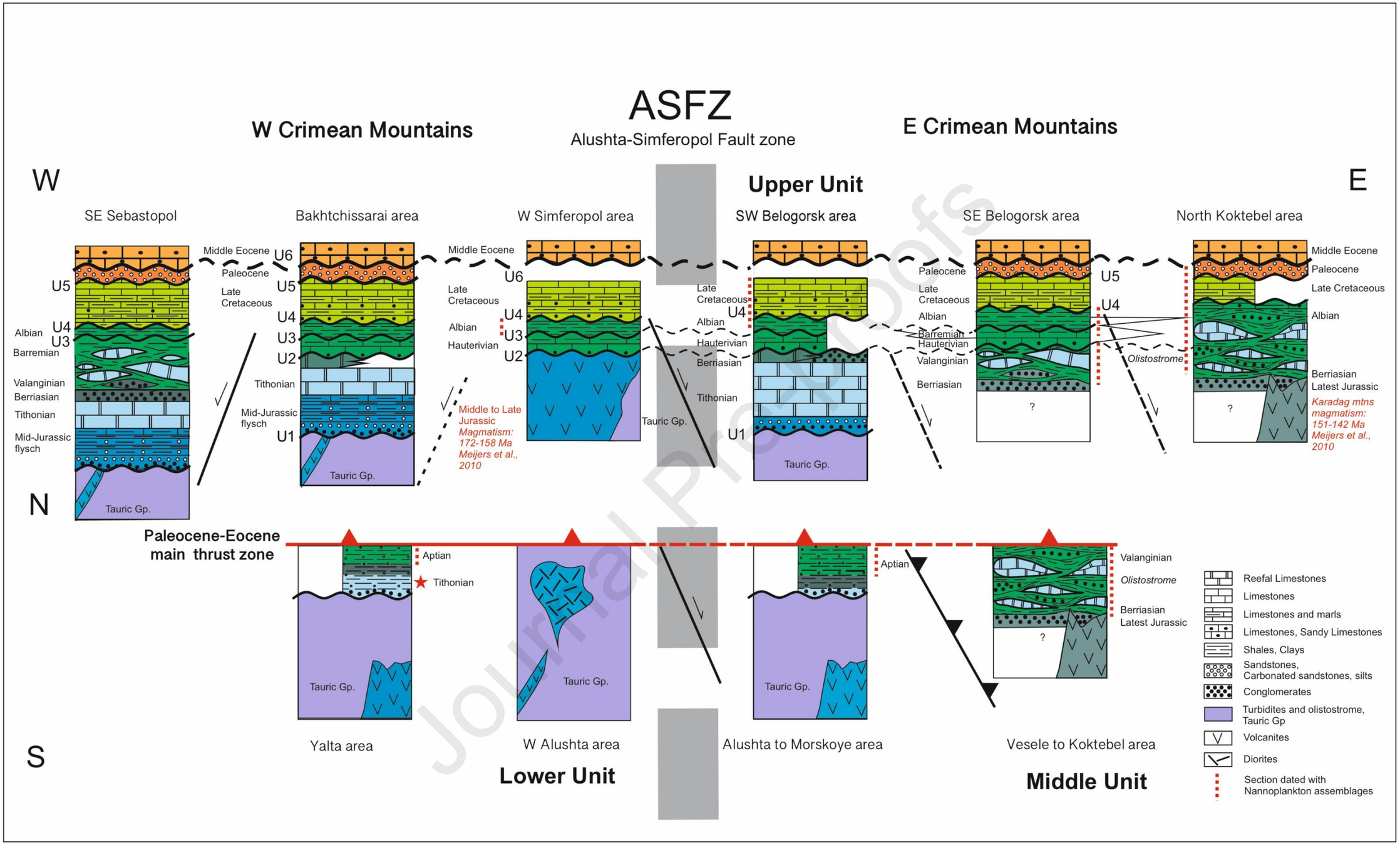
Table 1. Parameters of focal mechanisms of strong earthquakes in the Black Sea's northern coastal area (selected and supplemented from Gobarenko et al. 2016, Pustovitenko and Eredzhepov, 2017). Serial number corresponds to number of earthquake mechanism in Fig. 7 and Fig. 8.

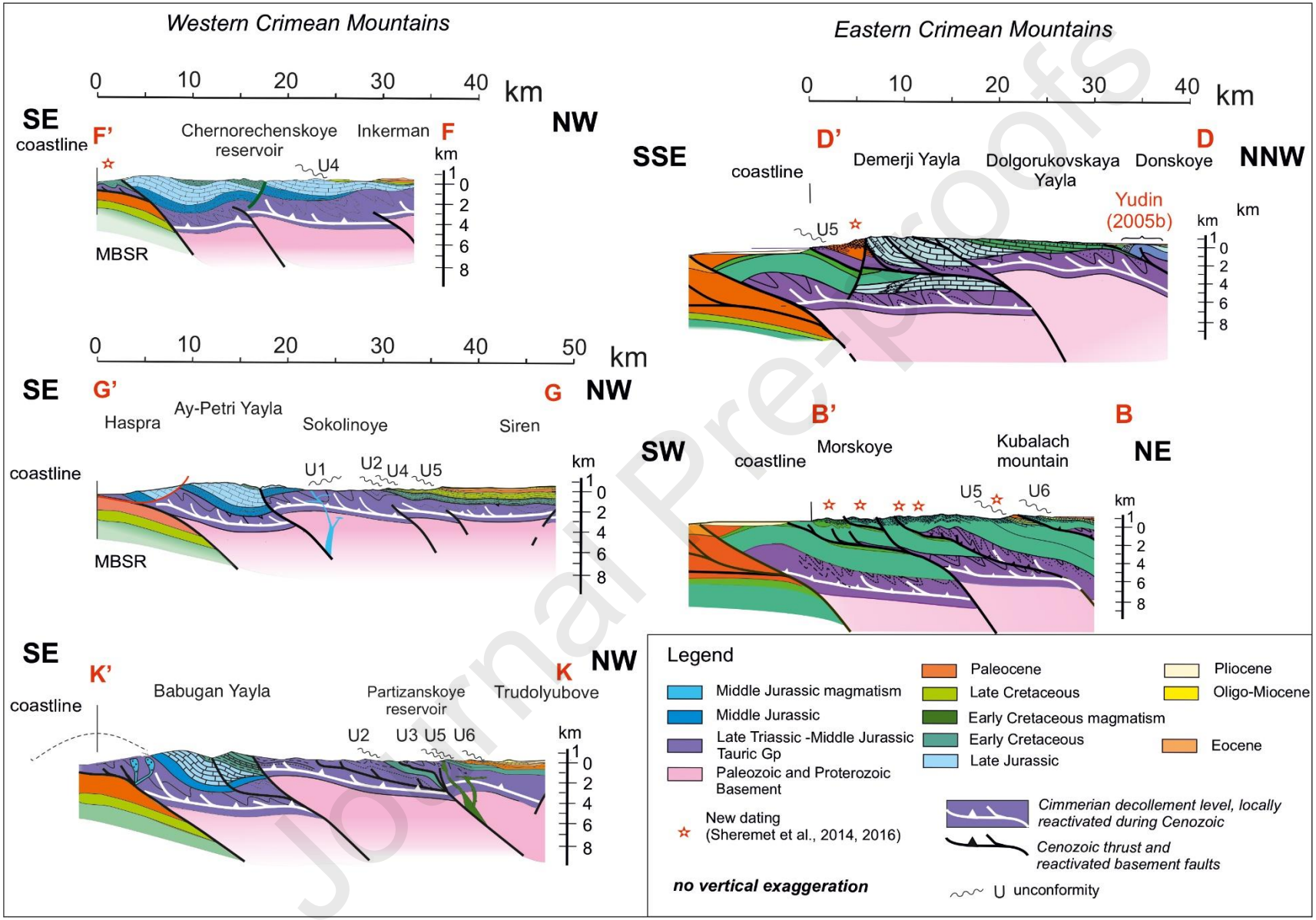
Table 2. Reverse, strike-slip, and extensional regimes obtained for the ECM, illustrated in Fig. 4. Nd: Number of fault data, R: Stress ratio $(\sigma 2 - \sigma 3)/(\sigma 1 - \sigma 3)$, CD: Counting Deviation in %, (Delvaux and Spencer, 2003).

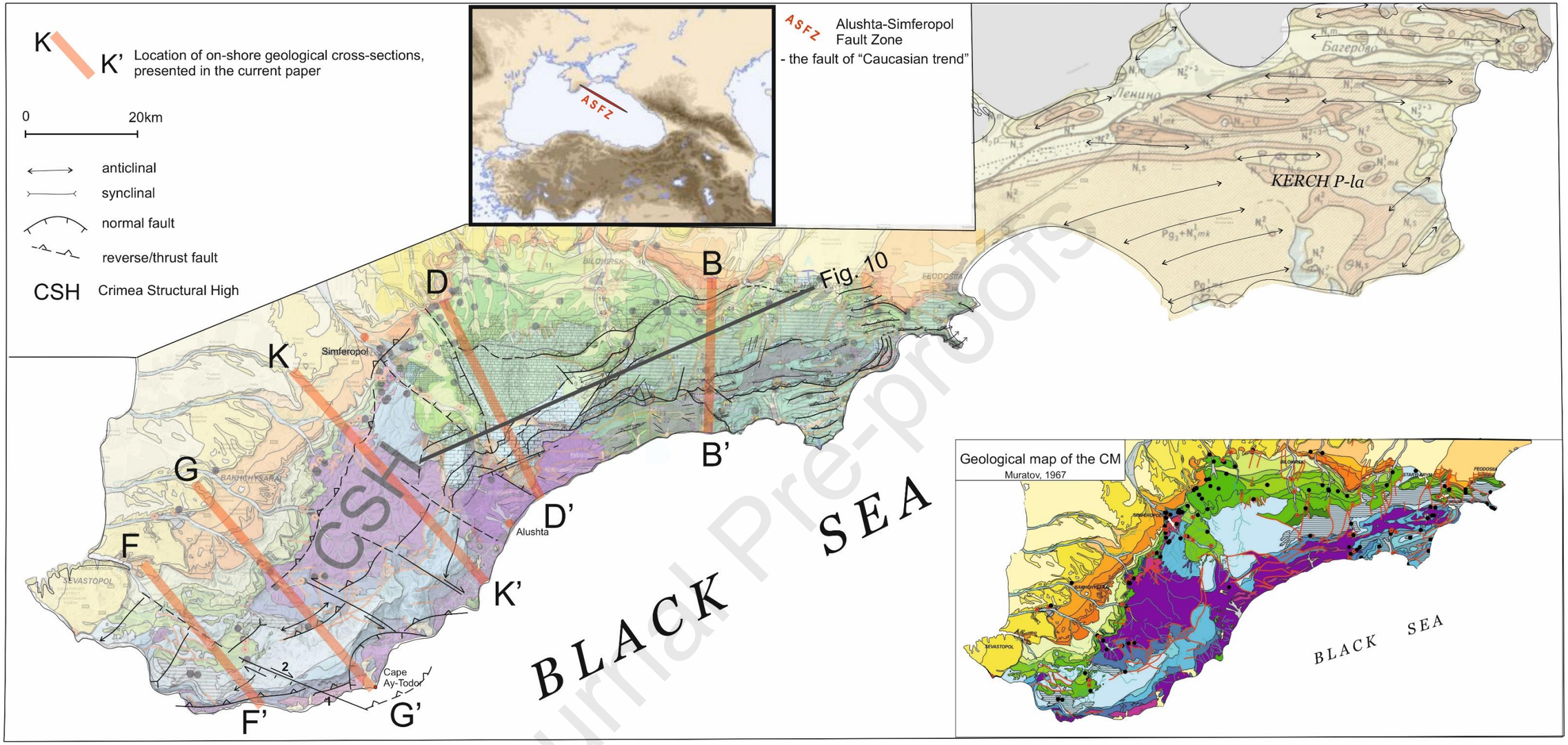


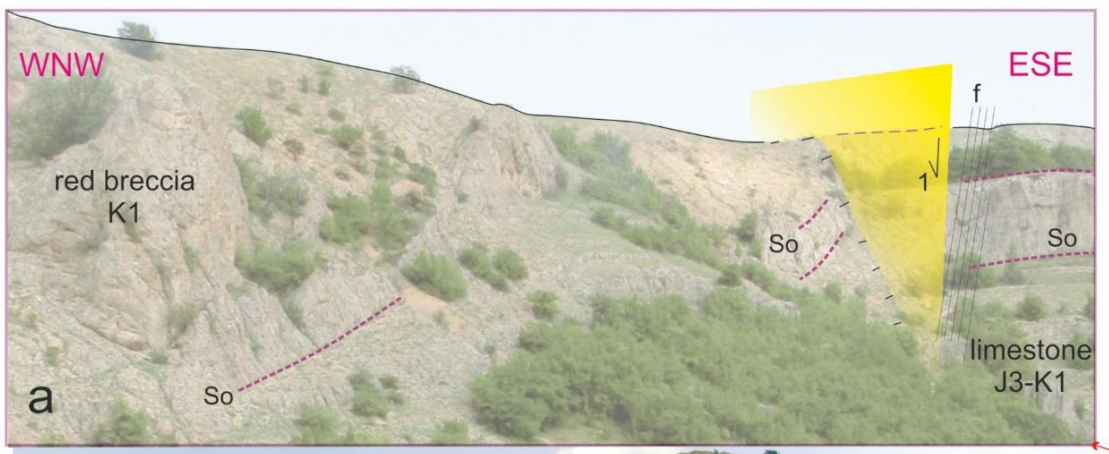


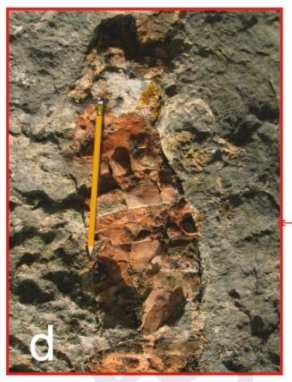



ASFZ
Alushta-Simferopol Fault zone
W Crimean Mountains
E Crimean Mountains
Upper Unit
W
E
N
S
SE Sebastopol
Bakhtchissarai area
W Simferopol area
SW Belogorsk area
SE Belogorsk area
North Koktebel area
Middle Eocene
Paleocene
Late Cretaceous
Albian
Barremian
Hauterivian
Valanginian
Berriasian
Tithonian
Mid-Jurassic flysch
Tauric Gp.
U1
U2
U3
U4
U5
U6
Middle to Late Jurassic Magmatism: 172-158 Ma Meijers et al., 2010
Olistostrome
Karadag mtns magmatism: 151-142 Ma Meijers et al., 2010
Latest Jurassic
Paleocene-Eocene main thrust zone
Aptian
Tithonian
Yalta area
W Alushta area
Alushta to Morskoye area
Vesele to Koktebel area
Lower Unit
Middle Unit
Reefal Limestones
Limestones
Limestones and marls
Limestones, Sandy Limestones
Shales, Clays
Sandstones, Carbonated sandstones, silts
Conglomerates
Turbidites and olistostrome, Tauric Gp
Volcanites
Diorites
Section dated with Nannoplankton assemblages



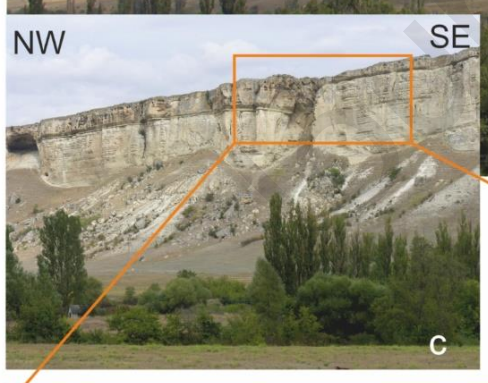




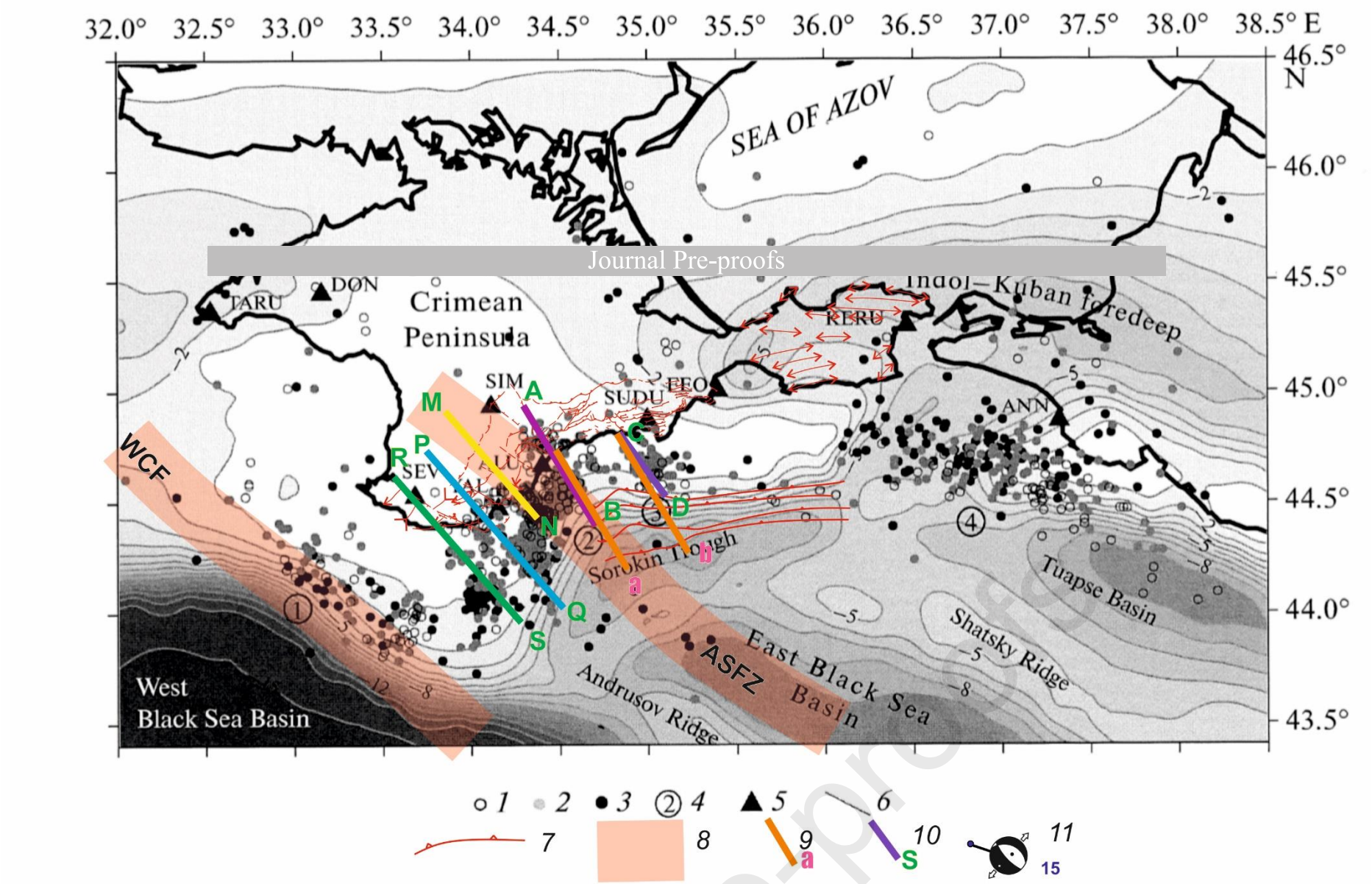


C

D



A



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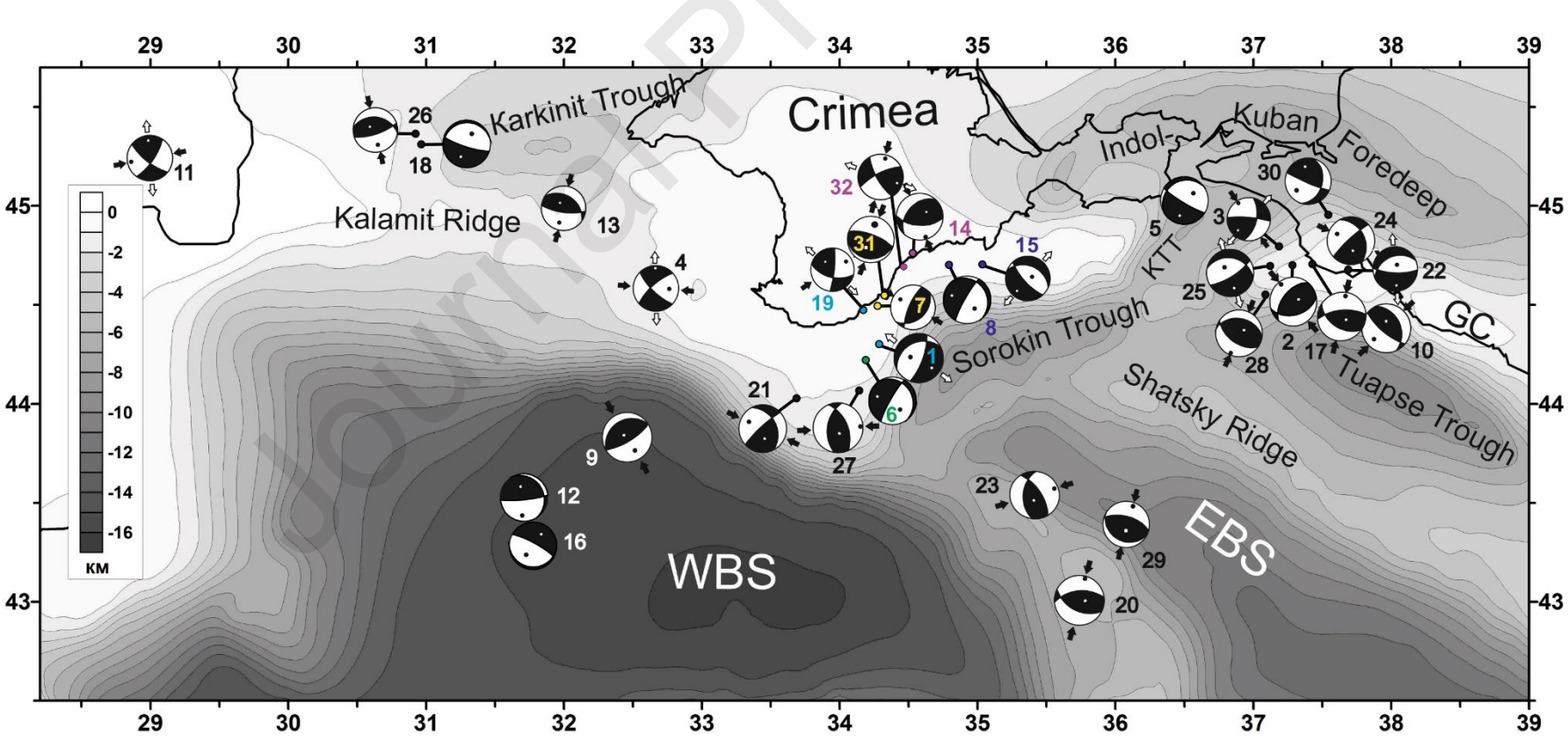
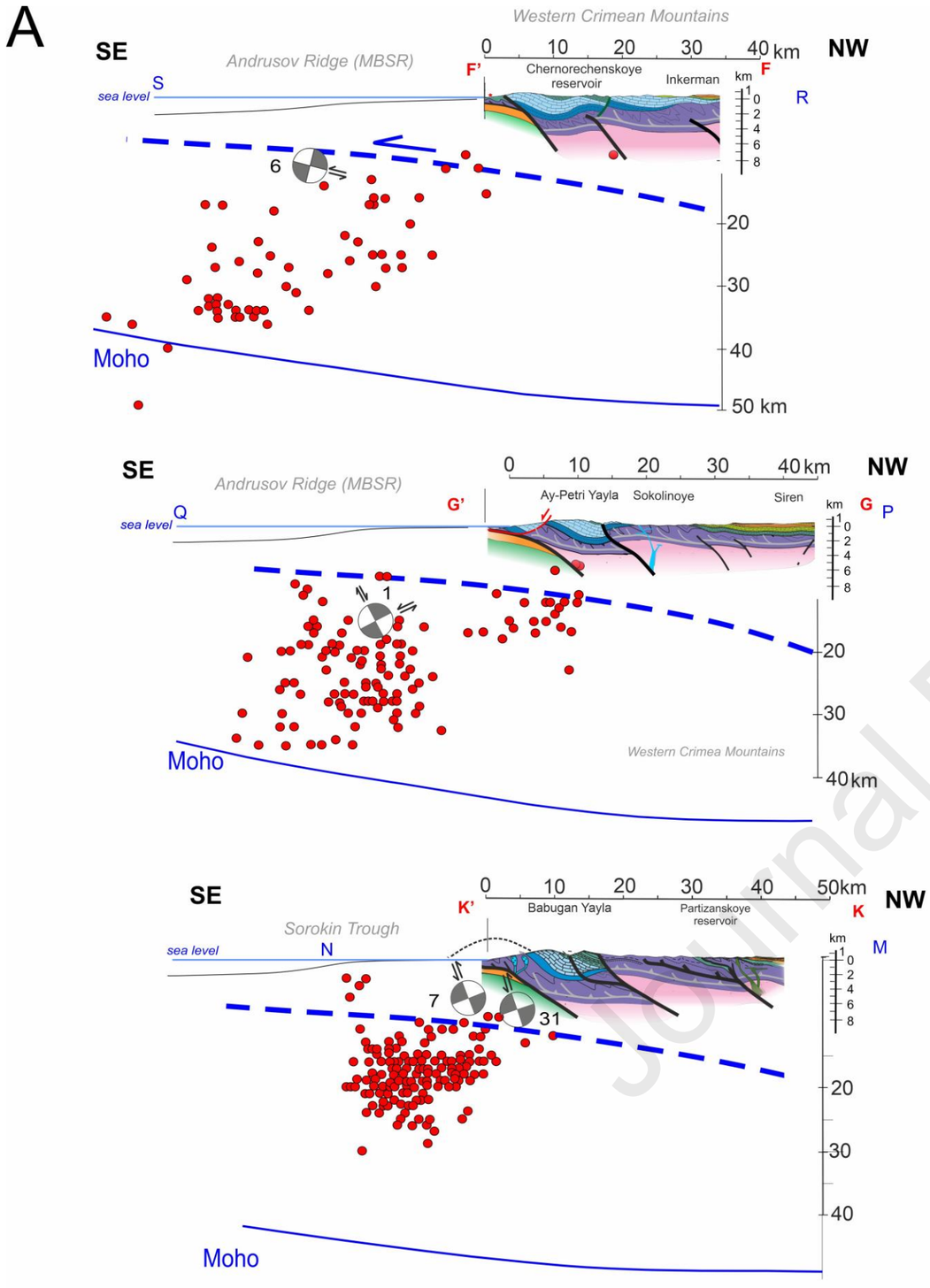
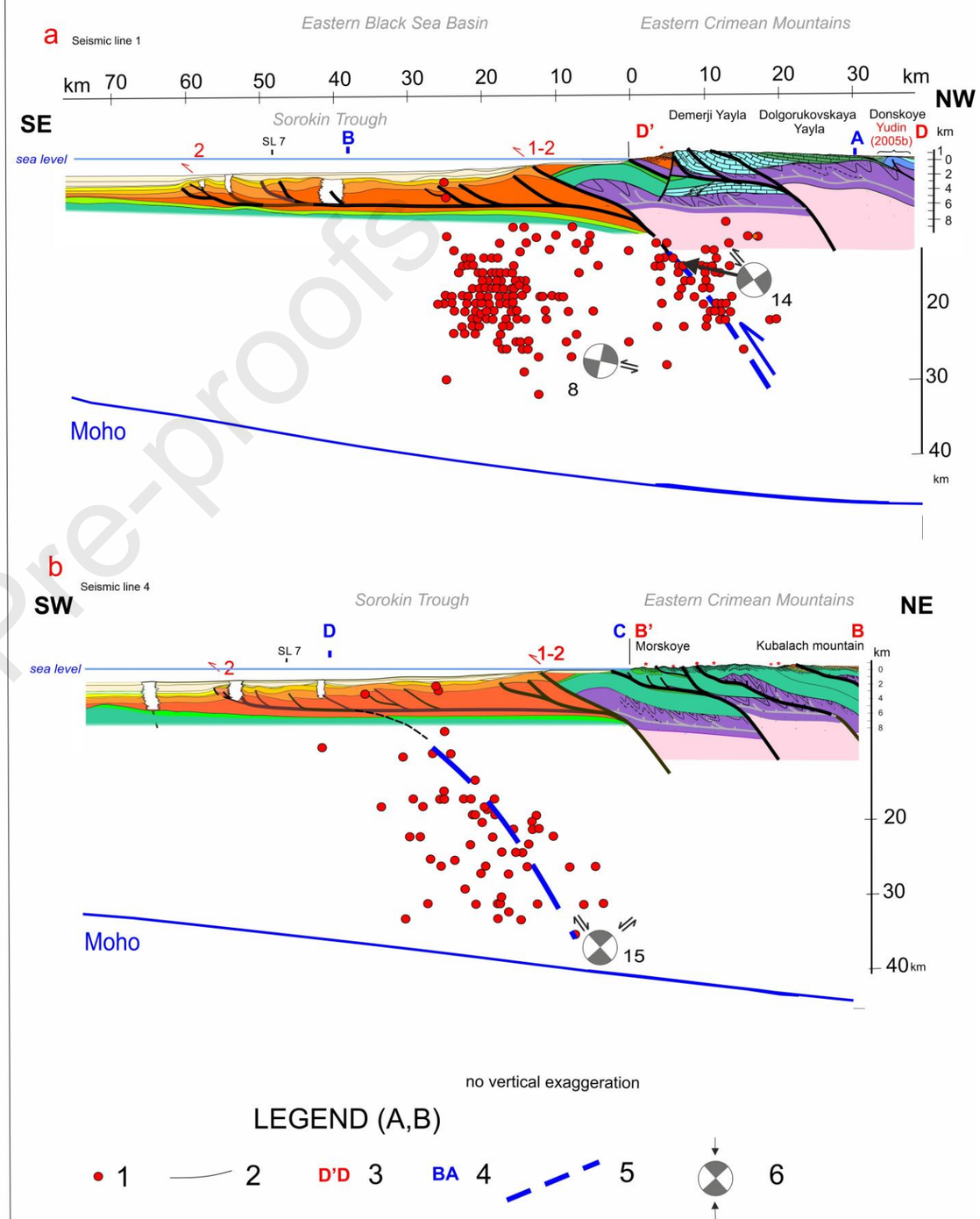
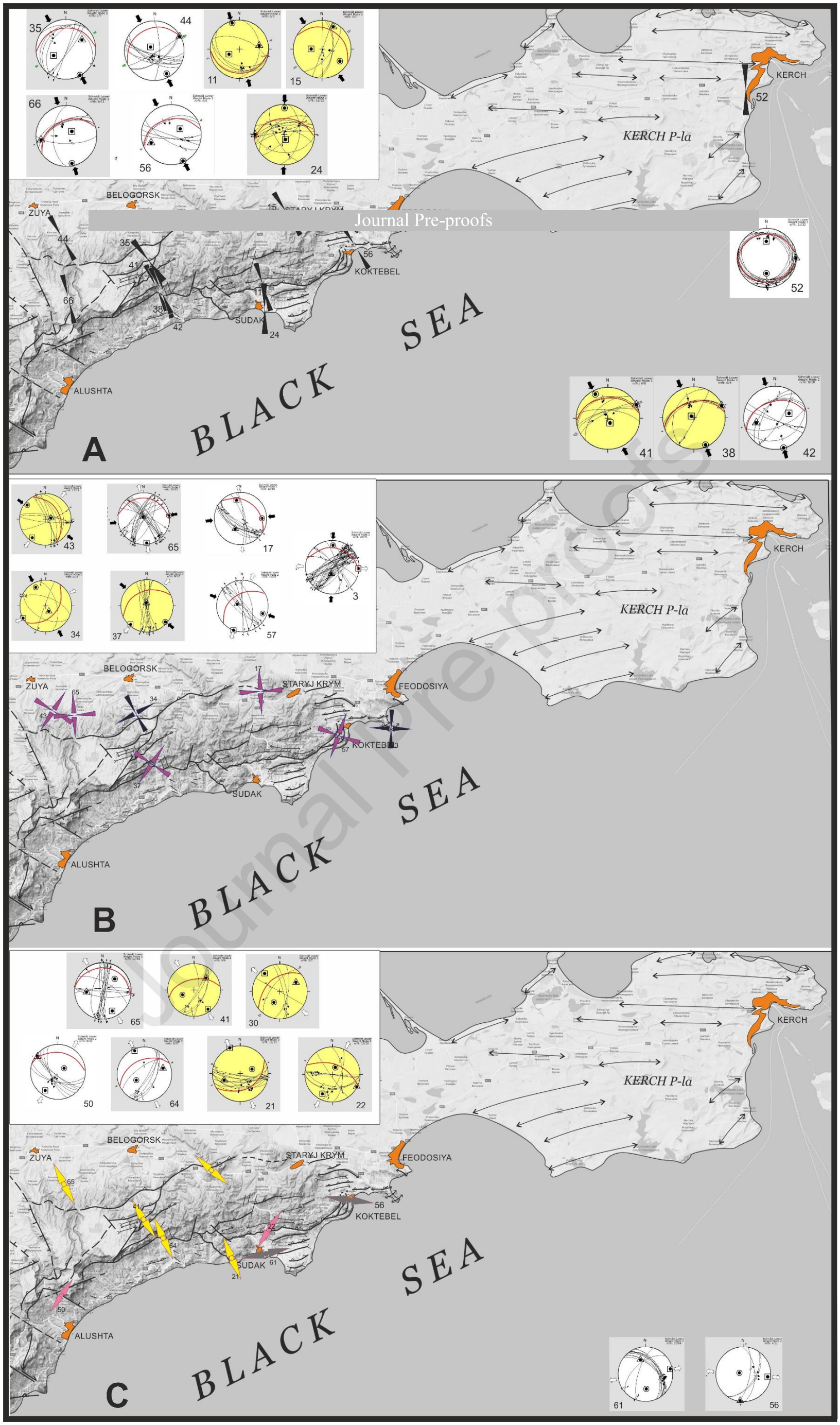
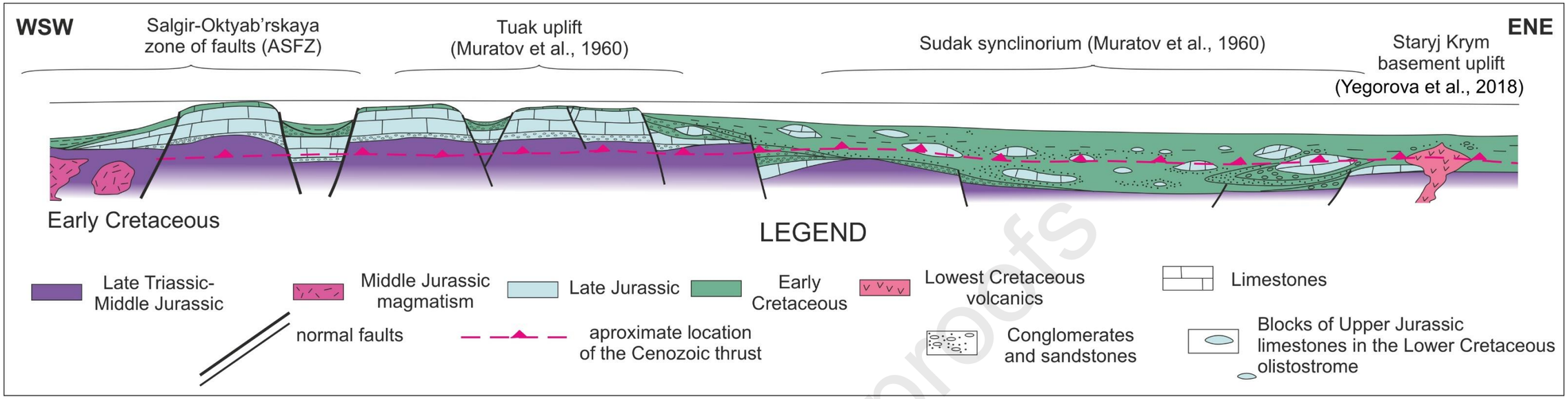










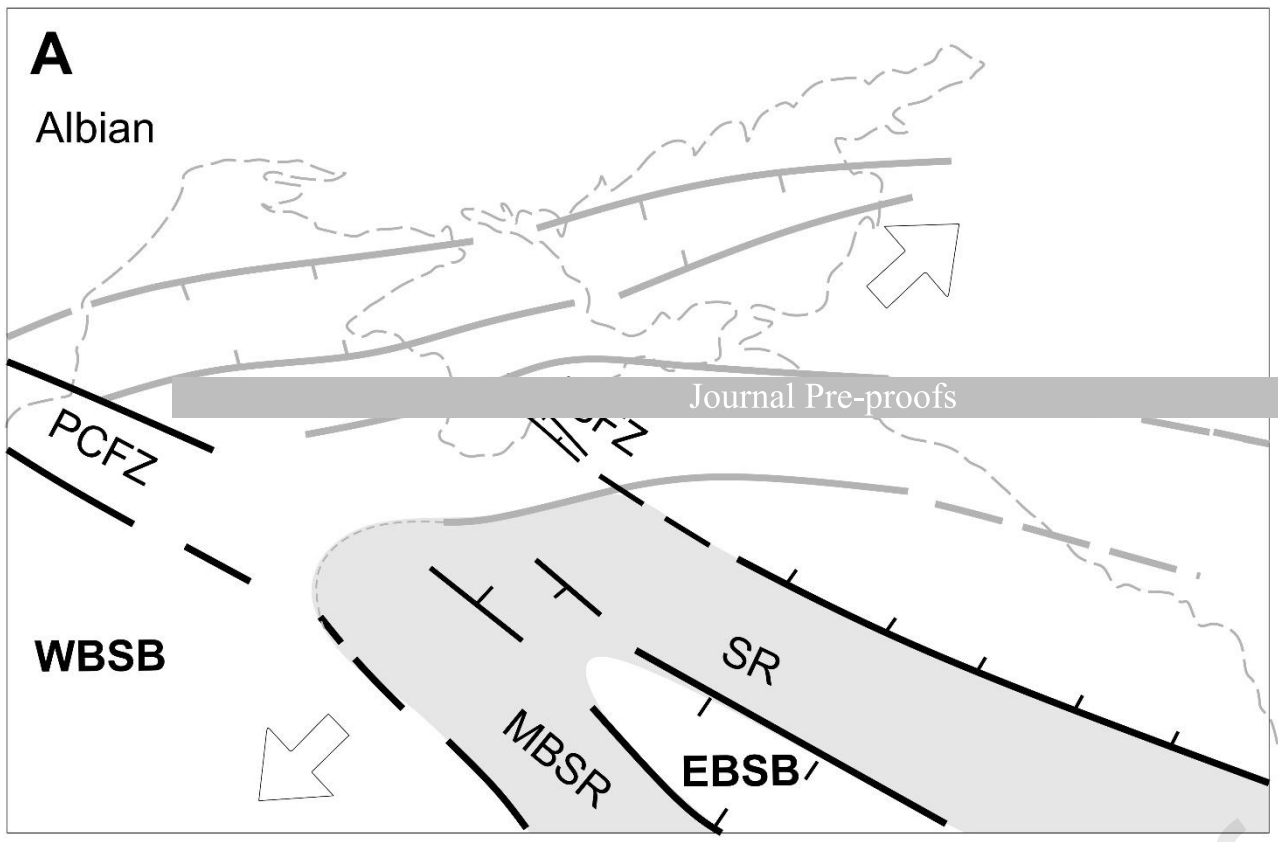


Table 1. Parameters of focal mechanisms of strong earthquakes in the Black Sea's northern coastal area (selected and supplemented from Gobarenko et al. 2016, Pustovotenko and Eredzhepov, 2017). Serial number corresponds to number of earthquake mechanism in Fig. 7 and

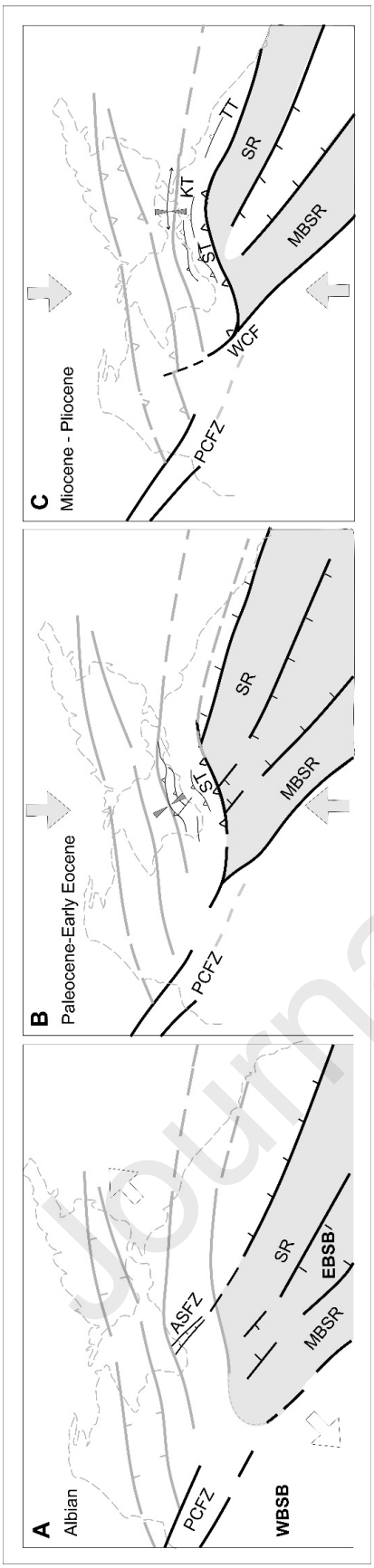
No	Date	Latitude, N	Longitude, E	Depth, km	Azimuth and dip of contraction axis, P	Azimuth and dip of intermediate axis, N	Azimuth and dip of extension axis T	Magnitude, Mw	Deformation mechanism	Structural position
1	11.09.1927	44.3	34.3	15	263/62	28/17	125/22	6.9	Normal fault	Continental slope, south Crimea
6	1955-1975	44.2	34.2	15	134/54	29/11	292/34	2.5	Thrust	Continental slope, south Crimea
7	1955-1975	44.5	34.3	5	301/29	204/13	92/58	2.5	Reverse	Continental slope, Southern Crimea
8	1955-1975	44.7	34.8	20	134/54	29/11	292/34	2.5	Thrust	Continental slope, South-Eastern Crimea
14	02.07 1990	44,78	3453	14	160/8	254/28	55/61	4.7	Reverse	Mountain Crimea
15	16.08.1990	44.7	35.06	28	81/67	317/14	223/19	4.7	Normal	Continental slope, South-Eastern Crimea
19	22.11.1996	44.51	34.16	10	61/16	173/52	320/34	3.2	Strike-slip	Mountain Crimea
31	15.10.2013	44,55	34,35	7	16/27	108/6	209/62	3.4	Reverse	Mountain Crimea
32	13.05.2016	44,63	34,41	13	10/200	67/315	20/106	3.5	Strike-slip	Continental slope, south Crimea

Fig. 8.

Table 2. Reverse, strike-slip, and extensional regimes obtained for the ECM, illustrated in Fig. 4. Nd: Number of fault data, R: Stress ratio (σ2 - σ 3)/(σ 1 - σ 3), CD: Counting Deviation in %, (Delvaux and Spencer, 2003).

Compressional regime with NW-SE to N-S σ_1	Ref. No	Place	Latitude/ Longitude In degrees	Stratigraphic age	Lithology	Nd	R	σ_1		σ_2		σ3		CD %
								Dir.	Plung.	Dir.	Plung.	Dir.	Plung.	
	35	Krasnoselovka	34,635167 44,918167	LOWER CRETACEOUS	Flysch	7	0.44	148	8	59	37	243	53	24.6
	41	Belogorsk-Privetnoe	34,6345 44,9011	LOWER CRETACEOUS	Flysch	8	0.75	334	13	64	0	155	77	23.8
	42	Belogorsk-Privetnoe	34,6351 44,901	LOWER CRETACEOUS	Flysch	13	0.33	164	2	255	38	72	52	23.7
	56	Koktebel	35,278133 44,969050	LOWER CRETACEOUS	Flysch	6	0.25	156	4	248	29	58	61	19.7
	11	Sudak-Dachnoye	35,000633 44,886433	LOWER CRETACEOUS	Flysch, conglomerates	6	0.17	345	7	79	31	244	58	23.3
	15	Strayj Krym	35,0905 45,047017	LOWER CRETACEOUS-UPPER CRETACEOUS	Conglomerates, limestones	5	0.7	123	27	26	14	272	59	18.8
	24	Kapsel-Sudak	34,997 44,8373	LOWER CRETACEOUS	Flysch	12	0.6	1	13	270	4	164	77	24.5
	38	Belogorsk-Privetnoe	34,634 44,911667	LOWER CRETACEOUS	Flysch	6	0.58	156	7	66	4	310	82	15.9
	66	Karaby	34,486667 44,966667	UPPER JURASSIC	Limestone	6	0.58	168	9	261	17	50	71	20.5
	44	Mezhgorye	34,415 44,991	LOWER CRETACEOUS	Limestone	12	0.71	152	14	61	4	316	75	21
	52	Kerch, Arshyncevo	36.47 45.30	MIOCENE	Limestone	21	0.52	181	47	90	1	359	43	24
Strike-slip regime (SSR) with E-W σ_1	17	Grushevka	35,005383 45,0162	UPPER JURASSIC	Limestone	12	0,29	114	28	207	54	348	30	22.2
SSR with NW-SE σ_1	37	Belogorsk-Privetnoye road	34.635 44.91	LOWER CRETACEOUS	Flysch, limestones	9	0.5	131	8	350	80	222	6	14.1
SSR with NW-SE σ_1	43	Mezhgorye	34.415 44.991	LOWER CRETACEOUS	Limestone, flysch	11	0.55	308	12	76	71	215	14	19.7
SSR with	57	Karadag	35.2151 44.9546	LOWER CRETACE	Volcanics,	13	0.5	110	17	292	73	200	01	20

NW-SE σ_1				OUS	limestone									
SSR with E-W σ_1	65	Mezhgorye	34.415 44.991	UPPER JURASSIC-LOWER CRETACEOUS	Limestone, flysch	26	0.56	83	1	339	84	173	6	17.7
SSR with N-S σ_1	3	Ordjonikidze and Feodosiya	35.34911 44.96933	LOWER CRETACEOUS	Flysch	40	0.57	2	19	189	71	93	2	21.2
SSR with NW-SE σ_1	34	Ulyanovka, open pit	34,634 45,0101	UPPER JURASSIC-LOWER CRETACEOUS	Limestone	5	0.4	332	19	150	71	242	1	23.3
Extensional regime (ER) ER with NW –SE σ3	21	Novyj Svet	34,90055 44,82065	LOWER CRETACEOUS-UPPER JURASSIC	Limestone, flysch	12	0,67	69	52	248	38	338	0	21.4
	30	Bogatoye-Bor Kaya	34,8529 45,0127	PALEOCENE, UPPER CRETACEOUS	Limestone, marls, sandy limestones	5	0.3	208	53	61	32	320	16	19.6
	41	Belogorsk-Privetnoye	34,6345 44,9011	LOWER CRETACEOUS	Flysch	5	0.5	247	51	45	37	330	29	18.7
	64	Belogorsk-Privetnoye	34,639 44,909	LOWER CRETACEOUS	Flysch	6	0.75	235	60	72	29	338	7	20.2
	65	Mezhgorye	34.415 44.991	LOWER CRETACEOUS-UPPER JURASSIC	Limestone, Flysch	17	0.68	44	40	266	42	155	22	23
ER with NE – SW σ3	22	Dachnoye	35.0006 44.88643	LOWER CRETACEOUS	Conglomerates	10	0.72	22	67	117	2	207	23	23.6
	50	Demerji	34,3958 44,7519	PALEOCENE (?)	Conglomerates	8	0.56	58	69	307	8	42	15	14.6
ER with E-W σ3	56	Koktebel	35,278133 44,969050	LOWER CRETACEOUS	Flysch	4	0.38	292	78	186	3	96	11	13.4
	61	Alchak-Kapsel	34,992050 44,836683	UPPER JURASSIC	Limestone	12	0.5	176	45	341	44	78	7	13.3



Highlights

1) The Crimean Mtns (CM) are divided into two parts by a NW-SE oriented main fault zone
2) Two systems of E-W and of NW-SE oriented faults were reactivated since the Cretaceous.
3) The reactivated structures control the Cenozoic inversion of CM
4) The model shows the major role of indenter during the Cenozoic inversion

CRediT author statement

YEVGENIYA SHEREMET: Conceptualization, Methodology, Investigation, Writing - Original Draft

MARC SOSSON: Supervision, Funding acquisition, Project administration

ANNA MUROVSKAYA: Investigation, writing precisions

VALENTINA GOBARENKO: Methodology, Investigation

TAMARA YEGOROVA: Project administration, writing precisions

Declaration of interests

☒ 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.

☐The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: