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Paratethys response to the Messinian salinity crisis

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Abstract:

The Black Sea and Caspian Sea are the present-day remnants of a much larger epicontinental sea on the Eurasian continental interior, the Paratethys. During the late Miocene Messinian Salinity Crisis (MSC), a unique oceanographic event where 10% of the salt in the world's ocean got deposited in the deep desiccated basins of the Mediterranean, the Paratethys Sea was connected to the Mediterranean Sea. Unlike the Mediterranean, no salt is known to have been deposited in the Paratethys region at this time, yet a similar mechanism of deep desiccation (with a water level drop of up to 2km occurring at 5.6 Ma) has been proposed in the past to explain the late Miocene and Pliocene Paratethys basin evolution.

Here, we review the basin evolution, stratigraphy and subsurface data of the four main Paratethyan sub-basins to investigate the response to the Mediterranean Messinian event. We show that hypotheses of a Paratethys-wide desiccation synchronous to the Messinian Salinity Crisis climax at 5.6 Ma do not hold. Determinations of the magnitude of the sea level drop appear to have regularly been overestimated by speculative basin-to-margin interpretations, and often been disproven by increased age model resolution.

In the Euxinian (Black Sea) Basin, the most recent estimates for the magnitude of sea level drop vary between 50 and 500 m, yet the timing is debated. Marginal outcrops in the Dacian Basin highlight multiple switches from shallow basinal to littoral and fluvial environments during the MSC interval, but no major water level drop coincides with the 5.6 Ma event. The Paleo-Danube deposits filling in the Pannonian Basin do not indicate any influence by the MSC and show prograding patterns into the deepwater lake Pannon. The dramatic expansion of the Paleo-Volga delta in the Caspian Sea is shown to be younger than the MSC, while estimates of the amount of water level drop vary widely due to the poorly understood contribution of tectonic processes.

These changing perspectives and decreasing estimates of water level lowering are not surprising given the vast northern drainage of the Paratethys region. Precipitation and runoff from the Eurasian continent ensures a much more positive hydrological budget under isolated conditions than the vast negative hydrological budget of the Mediterranean Sea which requires constant compensation by inflowing oceanic waters.

Keywords (max. 6)

Paratethys; Messinian Salinity Crisis; Pontian; Hydrological budget; Semi-isolated seas; Late Miocene

1. Introduction

The first hypothesis that the Mediterranean Sea desiccated during the Messinian Salinity Crisis (MSC) came from cores of the Deep Sea Drilling Project (DSDP) Leg 13 (Fig. 1). Messinian aged evaporites were discovered in the deep Mediterranean Basin and these were claimed to have been deposited in shallow waters or even subaerially (Hsü et al., 1973; Ryan et al., 1973). These observations were considered highly controversial and resulted in much discussion and the proposal of different hypotheses, ranging from "shallow basin-shallow water" (Nesteroff, 1978) and "deep basin-deep water" (Debenedetti, 1976) models. Geophysical evidence combined with paleontological studies later concluded that the Mediterranean was already a very deep basin before the MSC (Ryan, 1976; Hsü et al., 1977). Detailed investigations combining sedimentology, integrated stratigraphy, geochemistry, geophysics and geochronology, have since led to a general consensus on the paleoenvironmental changes that took place before, during, and after the MSC in the Mediterranean (CIESM, 2008; Roveri et al., 2014a).

The first suggestion that the Black Sea desiccated in concert with the Mediterranean MSC came from sediment cores retrieved during DSDP Leg 42B (Hsü, 1978; Hsü and Giovanoli, 1979). Drilling in the Black Sea Basin (Fig. 1) revealed a peculiar sedimentary layer (Unit IVd of DSDP Hole 380A) somewhere in the Upper Miocene, pointing to shallow, supratidal and intertidal sediments in the otherwise deep water sequence (Ross et al., 1978). This layer was interpreted as a sudden drastic (~1600 m) lowering of the Black Sea water level and was proposed to be linked to the Messinian event, mainly because "the MSC was the only unusual event that might be responsible for those unusual sediments" (Hsü and Giovanoli, 1979). The hypothesis of a Messinian desiccation event in the Black Sea was not generally accepted (Ross, 1978a; Stoffers et al., 1978), and especially the timing was seriously questioned (Kojumdgieva, 1979, 1983). Seismic profiles of the Black Sea Basin show evidence of several major erosional surfaces, which were generally correlated to the MSC event, despite poor chronostratigraphic control (Dinu et al., 2005; Gillet et al., 2007; Munteanu et al., 2012).

A major Messinian base level drop has furthermore been proposed for the Dacian Basin (Clauzon et al., 2005), the Pannonian Basin (Csató, 1993) and the Caspian Sea (Jones and Simmons, 1996; Reynolds et al., 1998), suggesting that the entire Paratethys domain was seriously affected by the Mediterranean MSC event. However, the timing, magnitude and, in some cases, even the existence of these water level changes are seriously questioned and subject to ongoing controversy (e.g. Popov et al., 2006; Magyar and Sztanó, 2008; Krijgsman et al., 2010; Stoica et al., 2013). The alternative scenario is that the Paratethys only experienced a minor sea level drop during the MSC and did not experience any widespread desiccation event. In particular, the common presence of brackish water fauna and phytoplankton of Paratethyan affinity (ostracods, molluscs and dinoflagellates) in the post-evaporitic 'Lago Mare' sediments of the Mediterranean suggests that Paratethys almost continuously provided brackish water to the Mediterranean during its so-called Lago Mare phase (Cita et al., 1978; Orszag-Sperber, 2006; Marzocchi et al., 2016; Stoica et al., 2016).

In this paper, we discuss the most significant data and arguments that deal with the question of how the Paratethys basins were influenced by the Mediterranean MSC event. We do this to find both time-tested and innovative approaches used to understand comparable events in a wide array of scientific disciplines by the otherwise largely separated scientific communities of the different basins. The response of the Paratethys domain to Mediterranean desiccation is still one of the main outstanding controversies regarding the MSC (CIESM, 2008; Roveri et al., 2014a; Flecker et al., 2015) and is of major scientific and economic interest. For instance, the timing and nature of water exchange between the Paratethys and Mediterranean are crucial components in quantitative studies on Messinian paleocirculation and precipitation processes of MSC evaporites (Meijer and Krijgsman, 2005; Flecker and Ellam, 2006; Marzocchi et al., 2016; Simon et al., 2017). Also, a significant Messinian sea level fall would have caused shelf exposure, slope incision, and large amounts of sediment being transported into the deep basin. In the Caspian Basin, the huge "deltaic" systems of the Productive Series that the Pliocene Volga River formed (during a period of sharp sea level lowering) are today the main hydrocarbon reservoir units for the South Caspian Basin oil province.

A full understanding of the MSC events in the Paratethys region, is however seriously hampered by confusing stratigraphic correlations and nomenclature in the Paratethys. A large part of the nomenclature was originally defined as regional biostratigraphic zonations based on (mostly) mollusc fauna endemic to the Paratethyan (sub) basins. The use of this terminology of typically marginal biozonations in a lithostratigraphic and seismic sense has contributed to the confusion. In addition, limited communication between researchers from different countries has in some cases led to different (boundary) definitions of the same stratigraphic units. Therefore we will provide an overview of the stratigraphic framework for both the Mediterranean and Paratethys regions during the late Miocene and Pliocene, and the progress that has been made since the first hypotheses on MSC desiccation were introduced.

2. Stratigraphic framework

2.1. The late Miocene – Pliocene in the Mediterranean

A reliable chronologic framework is a 'conditio sine qua non' to correlate the sedimentary sequences of the Mediterranean and Paratethys in detail and to unravel and quantify their water exchange. The Geological Time Scale (GTS) for the upper Miocene-lower Pliocene has experienced revolutionary changes since the DSDP drilling in the 1970s. In the widely used GTS of Berggren et al. (1985), the Messinian Stage was dated between 6.5 and 4.8 Ma, mainly based on integrated biostratigraphic correlations. The MSC was thought to be a single evaporitic event, roughly subdivided into "Lower Evaporites" and "Upper Evaporites" according to the stratigraphic subdivision made on Sicily (Decima and Wezel, 1973). The duration of the MSC was highly uncertain and estimated to last for several 100 kyrs.

Chronostratigraphic control for the fresh to brackish water deposits of the Paratethys domain was even poorer and largely based on biostratigraphic data from endemic molluscs and ostracods. The complex paleogeographic configuration of the region resulted in the development of different biozonations and different chronostratigraphic stages for individual Paratethyan basins. Correlations of the Messinian Stage to the biochronologies of the Paratethys were highly ambiguous, but it was generally assumed that the MSC event corresponded to the uppermost part of the Pontian regional stage of the Euxinian Basin (Hsü and Giovanoli, 1979; Stevanovic et al., 1989).

In the 1990's a spectacular breakthrough in dating sedimentary successions was made by the tuning of sedimentary cycles to astronomical (Milankovitch) curves of solar insolation and/or precession (Shackleton et al., 1990; Hilgen, 1991a, 1991b). Research focused on the Mediterranean domain where the sedimentary record is particularly sensitive to astronomically induced changes in past climate (Hilgen et al., 1995; Lourens et al., 1996; Krijgsman et al., 1999). The cyclicity of the Mediterranean successions underlies the construction of the astronomically calibrated time scale, which has now become the global standard for the Neogene. Recently, this astrochronology was also used to synchronize radiometric dating techniques (Kuiper et al., 2008). In the most recent GTS (ATNTS 12) the Messinian is now astronomically dated and constrained to between 7.25 and 5.33 Ma (Fig 2; Hilgen et al., 2012).

Astronomical tuning of the sedimentary cyclicity in Messinian deposits allowed a direct comparison of MSC sequences all over the Mediterranean (Hilgen et al., 1995; Krijgsman et al., 1999) and of the Mediterranean successions to the Atlantic record (Hodell et al., 2001; Krijgsman et al., 2004). Detailed cyclostratigraphic, biostratigraphic and magnetostratigraphic investigations of the pre-evaporite sequences showed that the onset of the MSC occurred at 5.96 ± 0.02 Ma, synchronously between the western and eastern Mediterranean (Krijgsman et al., 1999, 2002). Recently, this was refined to 5.97 Ma (Manzi et al., 2013). Opening of the Strait of Gibraltar and the subsequent re-flooding of the Mediterranean terminated the MSC at the beginning of the Pliocene at an astronomical age of 5.33 Ma (Van Couvering et al., 2000).

Detailed sedimentological and stratigraphic studies on the Messinian evaporites and post-evaporites led to the reconstruction of a robust high-resolution stratigraphic framework for the intra-MSC successions (Fig. 3a; Roveri et al., 2008; Manzi et al., 2009; Lugli et al., 2010). This has clearly highlighted regional differences throughout the Mediterranean basin. The present chronostratigraphic framework for the MSC generalizes the various lithological sequences into three main units (Roveri et al., 2008, 2014a, Manzi et al., 2009, 2011): 1) PLG (Primary Lower Gypsum) unit (5.97-5.60 Ma) consisting of shallow-water evaporites that accumulated only in semi-isolated marginal basins under full Mediterranean-Atlantic connectivity, 2) Acme interval of the MSC (5.60-5.55 Ma) with deposition of Halite and RLG (Resedimented Lower Gypsum), which is probably linked to glacial peaks TG10 and TG12 that potentially caused disconnection from the Atlantic Ocean, desiccation and erosion (Fig. 3a), 3) Post-evaporitic ("Upper Evaporites") unit (5.55-5.33 Ma) characterized by highly stratified brackish and fresh water environments hosting hyposaline Paratethyan faunal assemblages.

Numerous seismic profiles have become available that allow a detailed registration of the MSC event in the deep offshore domain of the Mediterranean (Lofi et al., 2005; Bertoni and Cartwright, 2007; Urgeles et al., 2011). These profiles confirm the basinward extension of erosion and enable the mapping of distinctive seismic markers. The Messinian Erosional Surface (MES) is generally interpreted as a subaerial feature that correlates to the partial desiccation phase of the MSC. Seismic profiles from the Gulf of Lion furthermore show that submarine gravity flows occurred prior to the halite deposition, which may partly account for the parallel reflectors of the so-called Lower Unit (Fig. 3b; Lofi et al., 2005).

In recent years, alternative interpretations have been proposed, arguing for a Mediterranean that remained full of water during the MSC. According to numerical modeling, the volume of evaporites in the Mediterranean are best explained by a basin that remained connected, albeit restricted, to the open ocean (Krijgsman and Meijer, 2008; Simon and Meijer, 2015). Observed simultaneous basin-wide paleoenvironmental changes during both the first stage (PLG: Lugli et al., 2010) and the last stage (Lago Mare: Stoica et al., 2016) indicate that both the onset and end of the salinity crisis did not necessarily correspond to major variations in sea level. Furthermore, a mechanism linking dense shelf water cascading into the deep basin has been proposed to explain the erosional surfaces related to the deep basin salt deposits, further questioning a major sea level drop (Roveri et al., 2014b). The important missing link at the moment is new scientific drilling of the MSC deposits in the deep basin, to allow for stratigraphic confirmation of proposed hypotheses.

2.2. The late Miocene-Pliocene in the Paratethys

In the last decade, numerous integrated bio-magnetostratigraphic studies have been performed on the Mio-Pliocene sedimentary successions of the Eastern Paratethys, which resulted in revised chronological frameworks for the Dacian and Euxinian (Black Sea) basins (Vasiliev et al., 2004, 2011, Stoica et al., 2007, 2013; Krijgsman et al., 2010; Van Baak et al., 2015b, 2016b). This allows high-resolution stratigraphic correlations between the individual Paratethys sub-basins and to the MSC of the Mediterranean (Fig. 2). However, there are still points open to discussion. The main uncertainty in the high-resolution Paratethys stratigraphic correlations is the lack of independent absolute age constraints throughout the MSC interval. Paleomagnetically, the entire MSC (5.97-5.33 Ma) falls within the reverse chron C3r (6.033-5.235 Ma). As a result, magnetostratigraphic studies rely on assuming constant sedimentation rates across the entire MSC equivalent interval, which, with changing paleoenvironmental conditions, leads to additional uncertainties. Identification of depositional hiatuses and estimating their duration therefore heavily relies on additional biostratigraphic arguments. Volcanic ash layers which could provide independent age constraints are only found in the South Caspian Basin (Van Baak et al., 2016b). Finally, the vast majority of the bio-magnetostratigraphic studies were performed in surface outcrops, and the correlation of the age data with the subsurface (well and seismic) database is not straightforward.

The base of the Pontian regional stage is dated throughout the Eastern Paratethys at 6.1 Ma and is thus slightly older than the onset of the MSC (Radionova and Golovina, 2011a; Chang et al., 2014; Van Baak et al., 2016b). A basin-wide transgression took place during the earliest Pontian, probably due to the establishment of a connection to the Mediterranean through the Aegean Basin (Stevanovic et al., 1989; Popov and Nevesskaya, 2000; Krijgsman et al., 2010; Radionova and Golovina, 2011a). This is marked by marine planktonic and benthonic foraminifera and calcareous nannofossil assemblages that are correlative to Subzone NN11b. Diatoms assemblages are correlative to Zone *Thalassiosira convexa*, subzone A (of tropical oceanic zonation, according to Burckle, 1972). Both groups confirm the ingression of marine water (Marunteanu and Papaianopol, 1995; Radionova and Golovina, 2011a; Stoica et al., 2013). This transgression is followed by the migration of faunal elements from the Pannonian Basin and the Aegean region into the Black Sea domain (Stevanovic et al., 1989; Popov et al., 2006; Grothe et al., 2017).

The Lower Pontian (Novorossian) substage largely correlates to the PLG unit of the MSC and relates to a general high-stand in the Paratethys with interbasinal and Mediterranean connections (Stoica et al., 2013). Locating the time equivalent of the acme interval of the MSC is a matter of much debate. To fully appreciate this, we will discuss this separately per basin in the respective sections. The Mio-Pliocene boundary is not evidently recognized throughout the Paratethys. A reestablishment of normal marine conditions as is the case in the Mediterranean Sea is not known anywhere in the Paratethys. Instead, the general trend across the MSC in many of the Paratethys basins is one of decreasing salinity (Van Baak et al., 2015b; Krezsek et al., 2016). This fits with the Paratethyan basins becoming disconnected from the Mediterranean Sea under (generally) positive hydrological budget conditions.

The late Miocene time scale of the Central Paratethys (Pannonian Basin, Vienna Basin and Transylvanian Basin) is still equivocal, lacking the integration of chronostratigraphic units defined outside its geodynamic domain. The commonly used “Pontian Stage” (sensu Stevanovic et al., 1989) is now omitted from the regional chronostratigraphic framework, because it is clearly demonstrated that the Pontian of the Central Paratethys did not correspond in time to the Pontian of the Eastern Paratethys (Magyar et al., 1999; Piller et al., 2007; Mandic et al., 2015). The proposed Transdanubian Stage (Sacchi and Horváth, 2002) was not accepted either, resulting in an extended Pannonian Stage that now covers the entire late Miocene in the most recent stratigraphic frame (Hilgen et al., 2012). A large portion of the Central Paratethys, including the Vienna and Transylvanian basins, had been filled up with sediment to base level (and even above) by the Messinian (Magyar et al., 1999). For the Pannonian Basin proper, an extensive seismic correlation system was developed, calibrated with biostratigraphic, magnetostratigraphic and geochronological data from wells (e.g. Teleki et al., 1994; Vakarcs et al., 1994). For the Tortonian interval a good system of biozones (Magyar and Geary, 2012) and authigenic $^{10}Be/^{9}Be$ data (Šujan et al., 2016) are available, but the biostratigraphic resolution of the Messinian is insufficient due to a limited amount of available data and radioisotopic ages are also missing from this interval. Correlations to the MSC events thus remain problematic or tentative at best.

3. Euxinian (Black Sea) response to the Messinian Salinity Crisis

As the nearest Paratethyan basin, the Euxinian or Black Sea Basin has since the first discovery of salt in the Mediterranean been the logical first stop to study MSC related events. As early as 1975 a DSDP drilling expedition was organized to (among others) retrieve Mediterranean aged sediments to answer a crucial question: What was the Paratethys response to the Messinian Salinity Crisis? However, all data for over forty years have highlighted another, more intriguing question: Why are there no Messinian evaporites in the Black Sea?

3.1. Basin evolution and depositional environments

The Euxinian (Black Sea) basin represents a back arc basin that opened during the Early Cretaceous to early Paleogene related to the northward subduction of the Neo-Tethys below the Balcanides-Pontides volcanic arc (Letouzey et al., 1978; Zonenshain and Le Pichon, 1986; Nikishin et al., 2003, 2015a, 2015b). Post-rift Paleogene and Neogene deposits cover almost the entire Black Sea shelf (Popov et al., 2004; Dinu et al., 2005). A thick unit of mudstones with organic-rich shale intervals was deposited during the Oligocene-Early Miocene, indicating the presence of euxinic environments (Popov and Stolyarov, 1996; Georgescu, 2003; Nevesskaya et al., 2003; Sachsenhofer and Schulz, 2006; Vincent et al., 2016). These organic-rich shales, grouped together under the Maykop Series, are the main source rock for hydrocarbons in the regions in and surrounding the Black Sea and South Caspian Basin (Sachsenhofer et al., 2015, 2017). The Middle Miocene mainly comprises shales, marls, limestones and fine-grained sandstones (e.g. Nevesskaya et al., 2003). The Volhynian (Lower

Sarmatian) is marked by an increased influx of fresh water in the Eastern Paratethys domain, causing a westward directed overspill and accentuated salinity fluctuations all over the Paratethys (Kojumdgieva, 1983; Piller and Harzhauser, 2005; Popov et al., 2006; Palcu et al., 2015). This resulted in deposition on the margins of white and white-yellow limestones in the circum-Black Sea area (Bulgaria, Romania, Moldova). These limestones are highly fossiliferous bearing numerous large mollusc shells endemic to the Paratethys region (e.g. *Mactra* limestones) and yielding rich foraminiferal assemblages dominated by long-ranging cosmopolitan species (Georgescu, 2003).

At the end of the Khersonian (Late Sarmatian sensu Barbot-de Marny, 1869), a major drop in sea level, estimated to be in the order of 300 m, has been reported in several places in the Euxinian (Black Sea) domain (Tugolesov et al., 1985; Robinson et al., 1996; Popov et al., 2010). Unconformities between the brackish (or hypersaline) Khersonian and semi-marine Maeotian deposits are very common, suggesting drying up of large areas (Popov et al., 2006, 2010). This short-term drop in sea level was followed by a transgression during the Early Maeotian that significantly enlarged the Eastern Paratethys domain (Stevanovic and Ilyina, 1982; Nevesskaya et al., 1986; Popov et al., 2004). The Early Maeotian (Tortonian, late Miocene) was inhabited mainly by endemic species and subspecies of euryhaline Mediterranean genera. Ephemeral marine ingressions from the Mediterranean Sea took place as evidenced by the presence of nannoplankton (Lyul'eva in Semenenko, 1987) and diatom associations. During the late Maeotian, semi-marine flora and fauna were again replaced by brackish mollusc associations and brackish and fresh water diatoms and dinocysts. This freshening trend was suggested to be related to the gradual closure of Paratethys-Mediterranean connections (Popov et al., 2006).

The beginning of the Pontian stage is marked by another reconnection to the open ocean. The Zheleznyi Rog section on the Taman Peninsula has proved to be the key section for detailed study of the Late Miocene and Pliocene environmental change in the Black Sea (Popov et al., 2016). In the lowermost part of the Pontian interval, anoxic conditions are encountered with lithologies alternating between bedded bituminous clays, non-laminated clays with benthic fauna and laminated diatomites. Diatom assemblages and dinocysts in this interval and within the individual diatomite beds indicate a trend of decreasing salinity (Radionova and Golovina, 2011a; Popov et al., 2016). Molluscs found in the lowermost part of the Pontian interval indicate a connection with the Aegean Basin (the Eupatoria beds) (Stevanovic et al., 1989; Popov and Nevesskaya, 2000). Upwards, the bedded bituminous clays disappear, to be replaced by marly clays rich in typical Pontian molluscs (e.g. *Paradacna abichi, Valenciennius* sp.) that originated in the Pannonian and Aegean basins. After the last diatomite occurrence the lithology passes into alternations of lighter and darker marly clays (Fig. 4), indicating regular changes between higher and lower carbonate contents. The clays are rich in Pontian molluscs and ostracods, indicative of deposition in the photic zone.

A pronounced forced regression took place at the end of the Pontian interval, which led to the emergence of the northern outer shelf (Popov et al., 2006). In the Zheleznyi Rog section, a conspicuous change from Pontian marls to Kimmerian coastal sediments with reddish sands/oolites/pisoliths and minerals indicative for evaporitic conditions suggest a minimum sea level drop in the order of 50-100 m (Krijgsman et al., 2010).

For the Pontian interval of the Taman Peninsula, two contrasting age models exist (Fig 4). Paleomagnetically these two options cannot be distinguished, given both options are possible within the duration of the C3r reversed chron (Fig 4). Both models agree on the age of the base of the Pontian. Magnetostratigraphic dating, biostratigraphic studies on nannofossils and diatoms, and cyclostratigraphy indicate that the Novorossian transgression took place at 6.1 Ma, just before the onset of the MSC in the Mediterranean (Krijgsman et al., 2010; Radionova and Golovina, 2011a; Chang et al., 2014; Popov et al., 2016). Both options recognize the need for one or more hiatuses at some point in the Pontian-Kimmerian succession, yet they differ on the position, timing and duration in the section. This has implications for the interpretation of the record and especially for the age of the condensed coastal sediments in the section.

The first option interprets the Pontian marly clay succession to be deposited without a major hiatus, and places a condensed section above the marls at the change to the red coastal sands (Krijgsman et al., 2010; Vasiliev et al., 2011; Chang et al., 2014). The age of the sea level fall responsible for this change should in this case be ~5.6 Ma, and therefore approximately corresponding in time with the Mediterranean sea level drop of the MSC (stage 2).

The second age model proposes an additional hiatus in the Pontian marly clays at a lumachelle level (Rostovtseva and Rybkina, 2014, 2017; Popov et al., 2016). At this level Andrusov (1917) defined the biostratigraphic boundary between Lower and Upper Pontian in the Russian part of the Euxinian Basin. Overlying the unconformity, the higher part of the Pontian section is in this case the equivalent of MSC stage 3.2 (the Lago Mare) and the lowermost Pliocene (Rostovtseva and Rybkina, 2014, 2017). The second hiatus (overlying the Pontian) then correlates to the lower Pliocene. This approach proposed that the climax of the MSC corresponds to the hiatus at the lumachelle level.

3.2. Changing perspectives of DSDP results suggesting desiccation

To get a full appreciation of the MSC in the Black Sea, it is important to investigate the deep basinal records. In 1975, three sites (DSDP Leg 42, Sites 379, 380/380A and 381) were drilled in the Black Sea (offshore the Bosporus - Fig. 1) and more than 2000 m of the Neogene sedimentary rocks were continuously cored (Ross et al., 1978). These DSDP drillings recovered Neogene sediments that are mainly deep-water hemipelagic, lacustrine, or brackish marine muds and marls. The sediment cores from Hole 380A, however, also revealed a peculiar interval at 864-884 m that became known as the

Pebbly Breccia unit. This unit contains stromatolitic dolomites associated to oolitic sands, conglomerates and pebbly breccias (Shipboard Scientific Staff, 1978). The horizontally laminated stromatolitic dolomite consists of pellets, ooids, with partially filled vugs, and looked similar to dolomite crusts formed by diagenesis in supratidal environments (Stoffers and Müller, 1978). Thin sections of the dolomites revealed a great variety of features (e.g. intraclasts, algal mats, crusts, pellets, oolites) indicating a shallow depositional environment with occasional subaerial exposure and evaporitic and meteoric diagenetic alterations (Stoffers and Müller, 1978). The stromatolitic structure was interpreted to be formed by trapping of carbonate muds by algal mats in intertidal environments. The well-sorted oolitic calcarenites are considered typical of deposition in a shallow subtidal environment. All observations suggested that the rocks in this Pebbly Breccia (IVd) unit were deposited in shallow-water and/or subaerial environments. This led to the hypothesis that the Black Sea had experienced a ~1600 m lowering of water level, under the assumption that the stromatolitic dolomite was not transported downslope but had formed autochthonously on the Black Sea floor during a significant sea level drop (Hsü and Giovanoli, 1979).

Paleontological and paleomagnetic evidence in the 1970s, combined with estimates of constant sedimentation rate, suggested that the unusual sediments of the Pebbly Breccia unit are late Miocene to Pliocene in age (Ross, 1978a). However, diagnostic fossils were practically absent and diagenetic formation of magnetic greigite (Fe_3S_4) minerals complicated reliable magnetostratigraphic dating (Hsü and Giovanoli, 1979). The tentative dating of the more than 1000 m of Black sea sequence was thus considered a working hypothesis at best (Ross, 1978b). Nevertheless, the Pebbly Breccia unit was correlated to the Messinian Salinity Crisis event of the Mediterranean (Fig. 5; Hsü, 1978; Hsü and Giovanoli, 1979), an opinion that was not shared by all shipboard scientists of Leg 42B (Ross, 1978a; Stoffers et al., 1978).

An alternative correlation was presented by Kojumdgieva (1979, 1983), who agreed on the shallow, subaerial, depositional conditions of the stromatolitic dolomites, but claimed that these deposits were of latest Khersonian (Late Sarmatian) age (Fig. 5). The specific composition of the evaporites (dolomites and other carbonates instead of sulfides and chlorides) were considered a consequence of the specific alkaline composition of Khersonian waters in the Black Sea basin (Kojumdgieva, 1983).

Based on diatom studies, a refinement of the age of the sediments overlying the Pebbly Breccia unit was proposed by Jousé et al. (1980) and later by Khursevich and Mukhina (1995). They showed multiple Pliocene zones in what was described by Kojumdgieva (1979, 1983) as Late Miocene. Further refinement based on diatoms and nannoplankton was made by Radionova and Golovina (2010), who indicated that unit IVe, below the Pebbly Breccia unit, belongs to the Maeotian Stage, instead of the Sarmatian Stage.

DSDP Hole 380A has been re-investigated for palynological studies (Popescu, 2006; Popescu et al., 2010; Grothe et al., 2014). The pollen data reveal coastal vegetation in association with a deltaic environment in the oldest sediments recovered, and large amplitude climatic variations with both humid thermophilous forests and dry steppes indicators directly underlying the Pebbly Breccia unit and in the aragonitic mud (Unit IVc) overlying the Pebbly Breccia unit (Popescu, 2006). An early Zanclean age was inferred for the sediments overlying the Pebbly Breccia unit because the observed climatic differences were in the same range as those between the late Miocene and early Pliocene of Sicily. Additionally, the early Zanclean nannofossil marker species *Triquetrorhabdulus rugosus* and *Ceratolithus acutus* were reported from this interval (Popescu et al., 2010). However, a later restudy of the same interval could not confirm the presence of these species (Van Baak et al., 2015b, 2016a; Popescu et al., 2016).

Dinocyst assemblages revealed two important events above the Pebbly Breccia unit: 1) the first occurrence (FO) of *Galeacysta etrusca* and the FO of *Caspidinium rugosum* (Grothe et al., 2014). These two events are also observed in the magnetostratigraphically dated Zheleznyi Rog section (Taman Peninsula, Russia) where they correspond to the Pontian flooding interval at 6.1 Ma. Furthermore, the presence of oceanic diatom species *Thalassiosira convexa* and *Thalassiosira praeconvexa* in subunit IVC (850-864m) permits dating this interval as 6.3-6.1 Ma (Radionova and Golovina, 2011b). The pebbly mudstone / breccia unit is consequently older than 6.3 Ma and it is therefore not possible to link the hypothesis of Black Sea desiccation to the MSC event (Grothe et al., 2014).

The continuity of the overlying Unit IV in DSDP Hole 380A is questioned by the interpretation of seismic data of Tari et al (2015), who referred to this interval as part of a mass transport complex due to gravitational collapse higher on the margin. In their interpretation, DSDP Hole 380A does not represent a continuous stratigraphy, but rather multiple shorter intervals between imbricated thrusts. As a result, they conclude that the DSDP Hole 380A record cannot be used to properly assess the impact of the MSC in the Black Sea.

Multiplication of section is however not immediately clear from the high-resolution (bio) stratigraphic records available for the core (e.g. Fig. 6). The available biostratigraphic and paleoenvironmental records of the core do not indicate obvious repeated shifts. This is especially clear in the most complete biostratigraphic record to date, the diatom record of Schrader (1978). Renewed attempts at dating the cored material also argues for the continuous nature of the core. A volcanic ash layer at 706.8 mbsf was dated at 4.36 ± 0.19 Ma using $^{40}Ar/^{39}Ar$ geochronology (Van Baak et al., 2015b). The magnetostratigraphic polarity pattern from here down to the 6.1 Ma biostratigraphic tie-point at 850.3 mbsf of Grothe et al. (2014) suggests a constant sedimentation rate of ~8 cm/kyr in this interval (Van Baak et al., 2015b, 2016c). Using this assumption of constant sedimentation rate, previously published proxy records for this core were then plotted in the time domain, allowing discussion of the paleoenvironmental evolution of the Black Sea throughout the MSC. Due to the present uncertainty on the continuity of the record, here we will show and discuss these records primarily in depth rather than in time (Fig. 6).

Across the supposed MSC interval, the diatom-based sea-surface salinity record, published by Schrader (1978), indicates a general trend of decreasing salinity, with exclusively fresh water conditions from 795 mbsf (5.6 Ma) upwards. Low abundances of halophyte pollen, indicating increased distance to the shoreline (Popescu, 2006), is interpreted to represent the Pontian marine flooding at 6.1 Ma (850 mbsf) (Fig. 6). Furthermore, this coincides with marine, Black Sea-like hydrogen isotopic (δD) values recorded by alkenones (biomarker for haptophyte algae) (Vasiliev et al., 2015).

A first change occurs at ~835 mbsf, but detailed study is complicated by a large core gap. Across this interval salinity and water level show a clear decrease, continental temperatures go down while sea surface temperatures go up. Most striking is that the coldest continental temperatures coincide with a second peak in hydrogen isotopic (δD) values recorded by alkenones (at 827 mbsf). This is interpreted to represent dry/evaporative conditions that could correlate to the glacial peaks of TG22 and TG20 at 5.8 Ma (Vasiliev et al., 2015).

A major change in this record occurs around 800 mbsf. In time, assuming constant sedimentation rates throughout this entire interval, this would approximate the 5.6 Ma water level drop in the Mediterranean Sea. A significant increase in halophyte pollen indicates a more proximal position of the shoreline and afterwards surface water salinities go to fresh. An absolute amount of sea-level drop can however not be determined based on this record. A potential mechanism to explain (near) simultaneous lowering of salinity and lowering of water level would be that at this time the connection to the Mediterranean was lost due to Mediterranean water level lowering. This would, even under a positive water budget in the Black Sea, lower the Black Sea water level to the height of the sill with the Mediterranean. In this case it seems important to note that the lowering of salinity appears to slightly postdate the lowering of the water level. Removing the influx of saline Mediterranean waters into the Black Sea would cause the gradual freshening of the Black Sea.

The DSDP records drilled in the 1970's have significantly progressed the understanding of the Black Sea evolution. However, the limitations of these cores in terms of recovery, storage and sampled locations (see also the next paragraph on seismic interpretations) hamper a full understanding of the deep basin response of the Black Sea to the MSC. This may only become possible through a renewed effort to obtain new cored material from the basinal Black Sea that is available for academic purposes.

3.3. Seismic interpretations

Comparing interpretations of seismic studies in the Black Sea are complicated by the use of different nomenclature between different sectors. In the Romanian offshore, biostratigraphically defined regional stages are preferred. For the MSC interval, the most relevant of these are the Maeotian, Pontian and Dacian. These are however not the same as used on the Russian margin, where instead of the Dacian, the Kimmerian regional stage is used. Boundary definitions differ between the two concepts. The Pontian-Kimmerian and Pontian-Dacian boundaries do not have the same age (e.g. Krijgsman et al., 2010), with the result that what on the Romanian margin is called Upper Pontian is the equivalent of the Kimmerian on the Russian margin. In the Turkish sector of the Black Sea, chronostratigraphic epochs are preferred. This divides the relevant time interval into Miocene, Pliocene and Quaternary. Due to the limited availability of published biostratigraphic data from the wells in which these epochs are determined, it is difficult to assess the correlation between sectors in detail.

Initially, geophysical surveys of the Black Sea Basin revealed the presence of several subbottom reflectors, but none of them seemed to be correlative with the Messinian M-refector (Letouzey et al., 1978). The seismic profiles furthermore revealed that, in contrast to the Mediterranean, no salt units were detected in the Miocene of the Black Sea Basin. This led to the conclusion that brackish sedimentation continued during the Late Miocene and Pliocene and that the Pebbly Breccia unit was probably related to prograding deltaic deposits (Letouzey et al., 1978).

In past decades, the understanding of the Euxinian (Black Sea) basin evolution has significantly improved by a large number of offshore seismic lines with increasingly high resolution. The seismic profiles of the Blason 1 & 2 surveys of the French Research Institute IFREMER displayed several major unconformities in the Neogene successions of the Black Sea (Gillet et al., 2003). Seismic lines offshore the Bosporus link the DSDP holes 380 and 381. They show that the top of the Pebbly Breccia unit at Hole 381 correlates with a clear erosional surface that corresponds to a strong amplitude reflector (M reflector) underlined by several erosional truncations. At Site 380, the top of the Pebbly Breccia unit does not correlate with any meaningful reflector. It was concluded that erosion did not reach the location of Hole 380, or that it was too weak to give a strong signal on the seismic data (Gillet et al., 2007). However, in their final interpretation, given the presence of the Pebbly Breccia unit in the DSDP wells, they assumed a dramatic lowering of the water level.

A Black Sea-wide 2D seismic dataset with a total length of almost 8900 km was acquired in 2011 as part of the project *Geology without Limits* (Graham et al., 2013; Nikishin et al., 2015c, 2015b). Especially the southwestern part of the Black Sea shows a complex pattern of deposition and erosion. One seismic line was specifically shot across the location where DSDP Sites 380 and 381 were drilled. This allowed assessment of the correlation between the two sites with significantly improved resolution (Tari et al., 2015). It was shown that the presence of gravity driven depositional processes has resulted in major packages of instantaneously deposited sediment. These can obtain thicknesses in the order of 100's meters. Both DSDP wells have penetrated such deposits, indicating that some of the sediment disturbance originally described as drilling-induced might be related to these large-scale depositional events. More importantly, this higher resolution data indicates that the magnitude of the MSC sea level drop is likely to be significantly less than 1600 m (Tari et al., 2015).

The Romanian shelf has been subject to multiple seismic studies in the last decades, based on both academic and industry datasets. Many of these studies favored an interpretation based on the hypothesis that the Black Sea experienced a major sea level drop of up to 1700 m during the MSC (Gillet et al., 2003, 2007; Dinu et al., 2005; Munteanu et al., 2012; Matenco et al., 2016). Recently this has become increasingly questioned, particularly with the acquisition of 3D seismic datasets (Tari et al., 2015, 2016; Krezsek et al., 2016; Schleder et al., 2016). These recent papers indicate that the initial interpretation of a subaerial erosional surface occurring in the deep basin was likely incorrect. Importantly, around the Black Sea no evidence is found for major river incisions during the MSC (Tari et al., 2016). Instead, the vast majority of the major erosional features present on seismic data are better interpreted as submarine canyon systems (Tari et al., 2016). As a result, basin-scale interpretations following the early claims for deep desiccation of the Black Sea during the MSC (Hsü and Giovanoli, 1979), need to be revised. The interpretation of Suc et al (2015) of fluvial processes as the cause for major Messinian erosion in the SW Black Sea should also be reconsidered (Tari et al., 2016).

What was previously interpreted as subaerial erosion on the Romanian margin, is now interpreted as the rugose top of a submarine mass transport complex similar to the ones observed near the DSDP wells (Fig. 7; Tari et al., 2015; Krezsek et al., 2016). This is part of a larger gravitational-collapse system detached on top of Maikopian clays (Schleder et al., 2016). The gravitational failure is interpreted to be triggered by a large sea-level drop. After the gravitational failure, thick sand-prone units, without age-diagnostic fauna but younger than Lower Pontian and older than Upper Pontian, are found as the infill of mini-basins created in the resultant topography (Krezsek et al., 2016).

Two important erosional surfaces are present in this sequence. First, a major unconformity cuts into the preceding shelf with a total relief of ~500 m (BES in Fig. 7). Second, erosion occurred across the infill of the mini-basins (MES in Fig. 7), expressed as a planar surface near the base of the BES incision on the slope. These erosional surfaces are very similar to the MSC erosional surfaces known from the Gulf of Lions in the Mediterranean (Krezsek et al., 2016). Crucially, two widely different interpretations have been given for this sequence. Bache et al. (2009) interpret this as a wave ravinement surface (peneplanation), formed after a major sea level drop. Alternatively, Roveri et al. (2014b) favor an interpretation in which dense shelf waters cascading along the slope into the deep basin cause significant erosion. The implication of this interpretation would be that no desiccation needed to have occurred and only a minor sea level drop would be required to explain the seismic observations. Either way, early claims of a drop of 1700 m are largely disproven and current thinking would involve a drop of 500-600 m at most. However, how a 500-600 m base level drop would not have caused dramatic river incisions similar to the Mediterranean Sea, is a point to be studied in future.

For a detailed chronological correlation of the above seismic sequences to the MSC events in the Mediterranean it becomes extremely important which definition of Upper Pontian is used. The use of the Pontian-Kimmerian stages as prescribed from the onshore sections on the Taman Peninsula is uncommon on the Romanian margin. Instead, Krezsek et al. (2016) adopt a Lower Pontian and Upper Pontian Pliocene. It is unclear how their boundaries correlate in detail to the stratotypical sections exposed on the Russian margin. Only a strict correlation using the Russian terminology would mean the events are all during the MSC, yet this would not give certainty at a cyclostratigraphic resolution for the age of the base level drop. Further complications arise when the Dacian Basin stratigraphic record is used for correlation as there is good evidence for multiple lowstands in this basin. This will be discussed in the next section. Correlation of a base level drop in the Black Sea to the climax of the Mediterranean MSC is therefore far from proven.

4. Dacian Basin response to the Messinian Salinity Crisis

The Dacian Basin is a satellite basin of the Euxinian (Black Sea) Basin. In the Late Miocene these two basins were separated by a shallow sill north of the Dobrodgea region (Fig. 8a). As a silled basin, the Dacian Basin is likely to record smaller amplitude base level variations than the Euxinian Basin. Combined with the active foreland basin setting, it provides an important record of events during the MSC. The large number of excellent kilometer-scale outcrops have inspired many generations to study Late Miocene events in increasingly higher levels of detail.

4.1 Basin evolution and depositional environments

The Dacian Basin is located in between the Southern Carpathians, Pre-Balkan area and the Dobrogean High (Hamor and Halmai, 1988; Matenco et al., 1997, 2003). It existed as a transitional basin between the Pannonian Basin and the Euxinic (Black Sea) basin. The Dacian Basin evolved as a satelite basin of the Eastern Paratethys since the Middle Bessarabian (Middle Sarmatian (s.l.); ~11.6-11.3 Ma), when the Carpathian orogen was tectonically uplifted to become a barrier separating the Pannonian and Transylvanian basins of the Central Paratethys from the Dacian Basin and Black Sea Basin of the Eastern Paratethys (Vasiliev et al., 2010; ter Borgh et al., 2013).

The distribution of Neogene sediments in the Dacian Basin is well documented by lithofacial maps, integrating the available well data with information from outcrop studies (Saulea et al., 1969). The isopach lines traced on these maps show that two separate areas of sediment distribution existed from the Sarmatian (s.l.) to the Pontian and that the thickest sediments are located near the Carpathians (Fig. 8b). The detrital influx came from two separate source areas: for the western Dacian Basin the main source was located in the surrounding South Carpathians (Leever et al., 2010, 2011; Fongngern et al., 2015), for the East Carpathian foreland (e.g. Focsani Depression), the main sediment was sourced from the north (Jipa and Olariu, 2009).

The sediment transport in the Dacian Basin evolved in three stages (Jipa and Olariu, 2009): 1) during Sarmatian (s.l.) and Maeotian times, sediments accumulated within two separated depocenters located near the two source-areas (Fig. 8b); 2) during the Pontian a large-scale westward progradation process occurred with clastics derived from the Carpathian source area. The Pontian progradation was first revealed through isopach line analysis (Jipa, 1997), but is also clearly shown by seismic profiles with inclined bedding (Tărăpoancă et al., 2003); 3) southward advancement of the depocenter took place during Dacian-Romanian times (Fig. 8b). In the sediment filling process, southern source-areas and the Dobrodgea region to the east provided additional detrital material into the Dacian Basin (Jipa and Olariu, 2009).

The thick and continuous Mio-Pliocene sedimentary successions of the Focşani Depression are excellently suited to revealing the palaeoecological changes in the Dacian Basin during MSC times (Stoica et al., 2013). These successions form the basis of high-resolution magneto-biostratigraphic studies that allow a detailed correlation to the standard Geological Time Scale (Vasiliev et al., 2004; Krijgsman et al., 2010; Van Baak et al., 2015a). However, due to frequent revisions of stratigraphic terminology and concepts, the Pontian stage as used in the Dacian Basin is different from the Black Sea.

The late Maeotian depositional environment in outcrops is characterized by shallow waters and littoral to fluvio-deltaic sediments (Fig. 9). A significant transgression took place at the Maeotian-Pontian boundary (6.1 Ma), marked by marine foraminifera and nannofossils (Stoica et al., 2013). This Odessian transgression is the equivalent event that has been reported from the Black Sea Basin and narrowly pre-dates the MSC onset. The Lower Pontian continues with fine pelitic sediments, deposited in deeper water basinal environments. In the Lower Pontian of the Dacian Basin the introduction of Pannonian Basin fauna into the Eastern Paratethys is evident (Stevanovic et al., 1989; Stoica et al., 2013; Grothe et al., 2017). Ostracod assemblages indicate this occurred within the photic zone. The ostracod assemblage also indicates salinity decreased to ~8‰ probably because of a positive water balance determined by the increased influx of continental waters.

The Middle Pontian (Portaferrian; 5.8-5.5 Ma) is marked by a widespread relative sea level lowering resulting in dominant fluvio-deltaic depositional conditions. These outcrop studies on the Pontian Stage of the Dacian Basin demonstrate that the main MSC sea level drawdown in the Mediterranean Sea (at 5.6-5.5 Ma) occurred in Portaferrian times. Palaeoenvironmental indicators show that the water level in the Foçsani Depression dropped less than 100 m during the MSC event (Stoica et al., 2013). During the MSC, the Dacian Basin remained filled with water, suggesting a positive hydrological balance for the region. This is compatible with the presence of a shallow barrier at Dobrogea (the Galati passage), separating the Dacian Basin from the Black Sea Basin during the late Miocene.

An alternative interpretation of the sedimentary response as normal regression is less satisfactory. In outcrop, mud-prone prodelta deposits are immediately overlain by delta plain and fluvial sediments, lacking any distinct intermediate features which are to be expected for a slower regressive transition. Higher in the sequence, the Dacian regional stage does show hundreds of meters of such intermediate environments being preserved during the final regressive sequence of the Dacian Basin (Van Baak et al., 2015a).

The Portaferrian-Bosphorian boundary (~5.5 Ma) is marked by a second transgressive event, with the overlying Bosphorian deposits accumulating in a more distal setting with finer pelitic sediments (Fig. 9). These shallow basinal conditions are in turn replaced by a second period of littoral and fluvial conditions, suggesting a second lowstand. This occurs near the top of chron C3r (prior to 5.2 Ma). Whether this transition still occurred during the MSC (i.e. prior to 5.33 Ma) or in the lowermost Pliocene is unclear. For correlation to the seismic records offshore Romania, this may however be very important. Both lowstands could be the equivalent of the sand-prone unit observed on seismics.

A third transgressive event is recorded during the Thvera subchron at around 5.1 Ma (Fig. 9). This is the final of multiple switches from littoral and fluvial environments towards shallow basinal conditions during the 6-5 Ma time interval. The overlying Pliocene and Pleistocene part of the basin infill sees the progressive filling of the basin (Van Baak et al., 2015a). All these base level fluctuations remained in the order of 100 m. It is therefore remarkable that in the western Dacian Basin a widely different series of events got proposed.

4.2 Sedimentation patterns suggesting desiccation

In the extensive Romanian literature and the detailed geological maps of the Carpathian foredeep, most areas in the Dacian Basin experienced continuous sedimentation during the latest Messinian time interval without any evidence of desiccation (Saulea et al., 1969; Jipa, 1997; Jipa and Olariu, 2009). Magneto-biostratigraphic dating of the sedimentary sequences of the eastern Dacian Basin

(Focsani Depression) showed that the entire MSC interval (lower-middle part of C3r) corresponds to a ~700m thick succession of fossiliferous fluvio-deltaic deposits (Vasiliev et al., 2004; Panaiotu et al., 2007; Stoica et al., 2013). Also in the southern (Prahova Valley) and western (Turnu Severin regon) parts of the Dacian Basin the Pontian sedimentary sequences are continuous and show no evidence for major hiatuses. One exception is the Arges-Topolog region (central part of southern Carpathians) where the Upper Pontian (Bosphorian) is found trangressively upon the Maeotian, indicating the presence of a major unconformity (Stoica et al., 2007). This hiatus is considered to be caused by a combination of erosional processes during the Middle Pontian sea level drop and local tectonic processes (Floroiu et al., 2011).

A complete desiccation of the Dacian Basin during the MSC was proposed by Clauzon et al., (2005), based on the sedimentary successions at the outlet of the Iron Gates (Fig. 8c). This is where the present-day Danube River cuts through the Southern Carpathian mountains and enters the Dacian Basin. Nearby, the Danube River cuts through a thick series of conglomerates that show an average dip 20 degrees eastward. These conglomerates had earlier been interpreted, based on fossil intercalations, as tectonically tilted strata of Middle Miocene age (Marinescu, 1978). However, Clauzon et al., (2005) interpreted the same conglomerates as the foreset beds of a very large Gilbert-type delta, suggesting an intense sea level fall and the desiccation of the Dacian Basin. It was concluded that these conglomerates must be of Zanclean age because the MSC was the only relatively brief event of such sea level change. Following up on this, a second Gilbert-type delta in the Turnu Severin area has been proposed in Suc et al. (2011), following similar argumentation as in Clauzon et al. (2005).

One of the characteristic features of a Gilbert-type delta is a tripartite structure, consisting of bottom sets, unconformably overlain by foreset beds and finally topset beds (Gilbert, 1890). In the graphical image presented in Clauzon et al., (2005), the threefold structure of the Gilbert-type delta at the Iron Gates is clearly present (Fig 10). However, the stratigraphic relationships between the three units is nowhere visible in the field. The localities that comprise these three individual units are all about 20-30 km apart (Jipa et al., 2011). Recent micropaleontological reinvestigations of the Iron Gate region confirmed again that the tilted conglomerates contain foraminiferal assemblages of Middle Miocene age, and that they are overlain by sandy deposits of Maeotian age (Jipa et al., 2011; ter Borgh et al., 2014). This makes these sediments up to 8 million years older than the Messinian Salinity Crisis. In addition, resampling of the assumed bottomset unit showed that this corresponds to the Lower Pontian (6.0-5.8 Ma) and that this is thus also younger than the, hypothesized overlying, conglomerate unit (Stoica et al., 2013).

The hypothesis of the Dacian Basin experiencing a major desiccation and reflooding event linked to the Mediterranean MSC thus remains highly speculative and is not supported by robust stratigraphical and sedimentological data.

4.3 Seismic interpretations

Sequence stratigraphic interpretations of seismic studies in the Dacian Basin suffer from poorly documented biostratigraphic control, partially because of proprietary reasons. As a result, correlations between subsurface and outcrop are ambiguous. In part lithostratigraphic arguments have been used to correlate the regional stage nomenclature into the deeper basin. The diachronous nature of some of the studied lithological transitions (e.g. deltaic progradation in the proximal areas), adds to this uncertainty.

A seismic transect across the western part of the Dacian Basin revealed a significant base level drop during the Middle Pontian, roughly coeval with the MSC (Fig. 8d; Leever et al., 2010). In the shallow part of the basin, two seismic sequences, separated by an unconformity (SB2), were distinguished: a Maeotian-Lower Pontian (SSQ1) and a Middle Pontian-Dacian (SSQ 2). Both sequences are characterized by well-developed clinoforms, arranged into prograding bodies progressively filling the basin (Fongngern et al., 2015). In the eastern part of the record, the early Pontian transgression is clearly evident where the prograding upper Maeotian unit is followed by a partly aggrading lower Pontian, indicating the flooding of the shelf. The sequence boundary SB2, associated in time with the MSC, is an erosional surface during the deposition of the forced regressive prism and a sedimentary bypass surface (Leever et al., 2010). The SB2 unconformity was traced to the Arges-Topolog region of the central Carpathians where it corresponds to a major Maeotian-Upper Pontian hiatus in the stratigraphic succession (Stoica et al., 2007; Floroiu et al., 2011). The base level fall associated with SB2 is roughly estimated at 100m, based on the elevation difference between the inferred shelf margins above and below the sequence boundary, although the seismic transect had to compensate for (unknown) tectonic activity along the Getic Fault (Leever et al., 2010). An overlooked seismic contribution to vertical movements may therefore significantly alter the interpretation of a 100 m base level fall.

In the eastern part of the Dacian Basin, a large database consisting of several seismic lines and numerous wells shows in detail the architecture of the Miocene sedimentary sequences of the Focsani Depression (Tărăpoancă et al., 2003). The Maeotian deposits are widespread over the region and their thickness reaches 1.5-1.6 km. Some Maeotian deposits were subsequently removed by erosion, as observed from an intra-Maeotian unconformity in the seismic profiles (Tărăpoancă et al., 2003). During the Pontian, the sediment depocenter shifts to the central part of the Focsani Depression, where thicknesses of 1.5 to 1.6 km are attained. The Pontian does not contain any remarkable seismic structures and mainly shows continuous sedimentation in a tectonically quiet environment, without any sign of a Messinian unconformity (Tărăpoancă et al., 2003).

5. Central Paratethys response to the Messinian Salinity Crisis

As a large endorheic lake, the Late Miocene Lake Pannon was similar to the present-day Caspian Sea. Contrary to the Dacian Basin, the Central Paratethys region does not have a spectacular MSC-related outcrop record. However, this is compensated by a very complete understanding of the subsurface through numerous studies which integrate seismic and borehole data for typical prograding Late Miocene deposits. In this isolated environment an endemic fauna developed, one which found its way into the Mediterranean during the final stage of the MSC, the Lago Mare. As such, the importance of the Central Paratethys appears to be in its connectivity history. This makes it even more remarkable that also here a major Messinian lowstand has been proposed.

5.1 Basin evolution and depositional environments

In Messinian time the once extensive Central Paratethys was restricted to the southern part of the Pannonian Basin (Fig. 11; Popov et al., 2006). This large intramontane basin opened as a back-arc basin during the Early to Middle Miocene (Horváth and Royden, 1981; Horváth, 1995). Extension was manifested through a set of low angle normal and strike-slip faults, resulting in the formation of a highly complex basement topography with several depressions and basement highs (Horváth and Tari, 1999; Krézsek and Filipescu, 2005). Due to the uplift of the Alpine-Carpathian orogen, the intra-Carpathian area became separated from the Eastern Paratethys by the beginning of the Late Miocene (ter Borgh et al., 2013). A large brackish water lake developed, known as Lake Pannon (Fig. 11; Magyar et al., 1999).

Post-rift thermal subsidence started in the Late Miocene (Horváth and Tari, 1999) and led to the formation and maintenance of deep-water sub-basins. These basins host several thousand meter thick sediments of Lake Pannon, deposited as deep water marls, turbidites, slope shales and deltaic sandstones to fluvial feeder systems (Bérczi and Phillips, 1985; Juhász, 1991).

Two distinct phases of partial basin inversion were postulated to have taken place during the post-rift evolution of the basin (Horváth, 1995). The first occurred in the middle/late Miocene (=Sarmatian/Pannonian) boundary, and resulted in a widespread unconformity at the base of the lacustrine sediments. The second was thought to be effective during the latest Pliocene and Quaternary, causing both uplift and accelerated subsidence (Horváth and Cloetingh, 1996). Opinions on the beginning of the latter inversion vary in more recent literature: Sacchi et al. (1999) put it to the Miocene/Pliocene boundary, whereas Bada et al. (2007) and Uhrin et al. (2009) suggest approximately 7 Ma as starting point of inversion, at least in the western part of the basin. During the

Quaternary, differential subsidence and large-scale lithospheric folding continued and the present-day topographic relief of the intra-Carpathian area evolved (Horváth and Cloetingh, 1996; Horváth et al., 2006).

The endorheic Lake Pannon (as a remnant of the Central Paratethys) persisted between 11.6 and 4 Ma (in the Late Miocene and early Pliocene) and hosted a variety of depositional environments, from shallow to several hundred meters deep water. Initially, Lake Pannon was only a mosaic of interconnected shallow lakes with patches of dry land in between. Due to the combination of post-rift subsidence and a period of increased precipitation (cf. Böhme et al., 2011), the volume of the lake basin as well as the volume of lake water increased, and most of the islands became inundated (Magyar et al., 1999). Deep underfilled troughs and shallow sublacustrine highs with condensed sedimentation were formed, thus calcareous marls with relatively high organic content accumulated over the lake floor.

As a consequence of the uplift of the Alps and Carpathians, enormous amounts of sediment were shed into Lake Pannon through large rivers from 10 Ma onwards (Magyar et al., 2013). Despite high (albeit decreasing) subsidence rates, sedimentation ultimately outstripped accommodation, leading to regression of Lake Pannon (Fig. 11). Major rivers entering the lake via deltaic lobes (mostly from NW and NE) built an extensive shallow water shelf bordered by a slope, transmitting sand for the turbidite systems in the depressions. This shelf-margin slope, having an average height of 400-600 m, and appearing as clinoforms on seismic profiles, advanced continuously from N to S from 10 to 4 Ma ago (Magyar et al., 2013). Data from well logs, cores and outcrops indicate that water depth on the shelf rarely exceeded 50 m, so these deposits appear as the upper horizontal portion ("topset") of the clinoforms. The overall shelf-edge geometry is expected to faithfully mirror any major (i.e. larger than 30-50 m in amplitude) lake-level falls (e.g. Helland-Hansen and Hampson, 2009). A water level drop over 100 m amplitude, for instance, should obviously result in descending shelf-edge trajectory and formation of a set of offlaping or small height clinoforms at different locations of the same age.

5.2 Sedimentation and seismic patterns suggesting dry climate and base-level drop in the late Messinian

In the non-marine sedimentary successions of the Pannonian Basin, the Miocene-Pliocene boundary is generally difficult if not impossible to detect. This problem was avoided by the introduction of the term "Pannonian Stage" (Roth, 1879). The Pannonian Stage includes (and thus unites) the Upper Miocene and the Pliocene. This approach was maintained by the Hungarian Stratigraphic Committee even after the Pontian, Dacian, and Romanian regional stages were officially introduced in other countries bordering the Pannonian Basin.

The idea that the Mediterranean Salinity Crisis left its fingerprint on the sedimentary record of the Pannonian Basin, however, was articulated from the 1990s in various fields of research. One line of arguments came from paleoclimatological studies. The presence of fossils of “extreme steppe elements such as gazelles and *Epimeriones* (desert mouse)” in the mammal stage “Bérbaltavárium” (Kretzoi, 1987) was considered by some authors to indicate semi-desert or almost desert climate. The occurrence of red and reddish brown varnish-coated desert crusts and dreikanters (ventifacts), i.e. three-edged rock faces formed by blowing sand, was also considered as additional evidence of an arid environment at the end of the Miocene and beginning of the Pliocene (Schweitzer, 1997, 2001). Whether the alleged desert conditions extended southward beyond the Pannonian Basin, and whether they were the cause or the consequence of the MSC, was never discussed. In addition, both the interpretation and the dating of some of the above features were questioned by other researchers (e.g. Jámbor, 1992, 2002; Csillag et al., 2010).

Another body of evidence for the effect of the MSC was expected to come from seismic sequence stratigraphic studies. Seismic studies in the Pannonian Basin identified unconformities that were interpreted as sequence boundaries (dated by correlation with several magnetostratigraphic borehole profiles). The interpretation of the seismic patterns, however, has been equivocal. Some studies claimed that the hiatuses represented by the sequence boundaries are local and are related to continuous delta lobe switching, thus representing 4th order stratigraphic cycles (Mattick et al., 1994). Others argued that among the many unconformities there are a few regional, third-order sequence boundaries, and that they correspond in time to the global late Neogene sea level lowstands of Haq et al. (1987) (Pogácsás et al., 1988, 1994). A third opinion was expressed by Csató (1993) who stated that he had found only one major unconformity in eastern Hungary, and he dated it “base Messinian” (Fig. 12). Vakarcs and Várnai (1991), Ujszászi and Vakarcs (1993), and Vakarcs et al. (1994) combined the two latter views by claiming that in the Upper Miocene – Pliocene basin fill, there are several basin-wide third-order sequence boundaries which mostly correlate in time with the global sea level drops, and that the most dramatic water level fall (ca. 200 m), the same as in Csató (1993), occurred at 6.3 Ma and was probably connected with the MSC. This age was determined using the global polarity time scale of Berggren et al. (1985), but as pointed out by Vakarcs (1997), if correlated with the improved GPTS of Berggren et al. (1995) this equates to 6.85 Ma. Recalculating this same level with the present-day astronomically tuned GPTS (Hilgen et al., 2012), would result in an age of 7.05 Ma.

Finally, Sacchi et al. (1999) conducted seismic sequence stratigraphic analysis in the southwestern Hungarian part of the Pannonian Basin, and came to the conclusion that there are several third-order cycles in the late Neogene basin fill, but these were dated quite differently from those of Vakarcs et al. (1994). Sacchi et al. (1999) observed a major angular unconformity across the

Miocene-Pliocene boundary, and attributed it to the tectonic inversion of the Pannonian Basin rather than the environmental impact of the MSC.

The systematic sedimentological analysis of continuously cored boreholes from the Neogene basins of Hungary has also provided essential contribution to the Messinian debate (Juhász et al., 1996, 1997, 1999). These studies found that the Late Neogene sediments of the Pannonian Basin are divided into two third-order sequences: an Upper Miocene and a Pliocene one. The two sequences are separated by a significant regional unconformity. At the top of the Miocene, flood plain fine sand, silt and clay were deposited and palaeosols were formed. The subaerial exposure was indicated by yellow and white mottles, calcareous nodules, and abundant root casts. The thickness of the altered zone below the unconformity ranged from 1 to 10 m. The Pliocene sequence, on the other hand, started with coarse channel sands and flood plain marls. Magnetostratigraphic data indicated a 1.5-2.0 Ma hiatus between the two sequences. This result was fully consistent with the suggestion of Kretzoi and Krolopp (1972), who found that there was a regional unconformity representing a 2-million-year hiatus between the Miocene and Pliocene in many boreholes in eastern Hungary based on mammal and mollusc biostratigraphy. Juhász et al. (1996, 1999) suggested that the unconformity at the top of the Late Miocene sequence reflected a major global or Mediterranean event, and thought that it might have been correlated in time, and probably casually, to the MSC and the Lago Mare event.

More recently Csató et al. (2007) re-addressed the “Messinian problem”. They changed the seismic interpretations of Csató (1993), without any justification or explanation, so that their major unconformity correlated with the end of the Messinian instead of the “base Messinian” (Fig. 12), and claimed that the presence of the dinoflagellate alga *Galeacysta etrusca* near the intra-Messinian unconformity confirms its late Messinian age. They also demonstrated by building stratigraphic models that the Messinian unconformity in the Pannonian Basin was a combined effect of tectonic inversion and an absolute lake-level drop, and concluded that it was “likely linked to the desiccation of the Mediterranean” (Csató et al., 2007, 2013; Suc et al., 2011). In a most recent paper, however, Csató et al. (2015) formulate a more permissive conclusion. They claim that the major unconformity in Eastern Hungary was “most likely formed by a contribution of base-level fall and tectonic motions”, and that “the exact age of the Messinian unconfomity is not known in the Pannonian Basin”, therefore “it is not possible to tell whether it was coeval or not with the salinity crisis in the Mediteranean”.

5.3 Seismic and stratigraphic evidence (and/or interpretations) questioning a significant late Messinian base-level drop

The latest Miocene through earliest Pliocene is represented by a thick sediment package in eastern Hungary, but its chronostratigraphic control is rather poor. In particular, the 4.6 to 6.8 Ma interval is not represented in any of the interpreted magnetostratigraphic profiles (Csató et al., 2007; Magyar and Sztanó, 2008). Consequently, any magnetostratigraphy-based correlation with the events of the Mediterranean MSC must be considered an estimation at best.

The only alleged biostratigraphic evidence for such a correlation is the presence of *Galeacysta etrusca*, a Mediterranean Lago Mare dinocyst (Popescu et al., 2009) near to the "intra-Messinian unconformity" of the Pannonian Basin (Csató et al., 2007). This form, however, like many other Lago Mare organisms, has a much longer stratigraphic record in the Pannonian Basin, spanning the entire Messinian and part of the Tortonian (Süto-Szentai, 1994; Grothe et al., 2017).

The seismic evidence for a large end-Messinian lake-level fall remains equivocal. Although Csató et al. (2007, 2013, 2015) established that the stratigraphic pattern in eastern Hungary could be simulated only if a significant lake-level drop was postulated in addition to the tectonic inversion, the pattern that they reconstructed was their own interpretation of the seismic profiles. That the seismic data can be interpreted in various ways, however, was demonstrated when they changed the seismic interpretation of Csató (1993) so that the intra-Messinian unconformity now correlates with the top of the Messinian and not with the base of the Messinian as in the original paper (Fig. 12). Additional alternatives for the interpretation of the same seismic pattern, offered by Sztanó et al. (2007) and Magyar and Sztanó (2008), do not imply a base-level drop below the shelf-edge. Instead, they suggest that this unconformity is a local phenomenon, caused by onlaps on the shelfmargin clinoforms by turbidites that arrived from another source (Fig. 12).

There is no doubt, however, that the unconformity reported from borehole sequences by Juhász et al. (1996, 1997, 1999) is significant on a regional scale, and it can be observed in seismic profiles as well (Fig. 13). There are a lot of scattered biostratigraphic (molluscs, mammals) and magnetostratigraphic data indicating that the sequence below the unconformity is Miocene, whereas the sequence above the unconformity is Pliocene in age (Magyar and Sztanó, 2008). This unconformity is associated with the second late Neogene inversion phase of the Pannonian Basin, and formed as differential subsidence in some depressions and/or uplift of the neighboring highs took place. Sediment accumulation in extended alluvial plains, ponds and marshes continued in the depressions, thus forming a gradual onlap on the topographically higher parts of the unconformity.

There is no agreement, however, on the basinward correlation and tracing of this unconformity (Fig. 12). If we accept the present-day geochronological framework of the Pannonian Basin, the latest Messinian shelf margin is located in southeastern Hungary and northern Serbia, where undisturbed progradation-aggradation cycles developed during the Messinian through Pliocene; there is no sign of any detectable lake-level drop (descending shelf-margin trajectory or pronounced truncation surface) in the shelf-margin clinoform architecture (Fig. 14; Pigott and Radivojevic, 2010; Sztanó et al., 2013; ter Borgh et al., 2015). This pattern confirms that the (approximately) Miocene-Pliocene unconformity developed only in the wide hinge zones of the subsiding basins, and that it does not imply any direct hydrological relationship with the Messinian drawdown of the Mediterranean Basin (Magyar and Sztanó, 2008).

The largest and most spectacular late Miocene erosional surface known to date in the Pannonian Basin is associated with a buried canyon system in central Hungary (Fig. 14; Juhász et al., 2013). This perplexing feature roughly and partly correlates in time with the major sequence boundary ("Pa-4") of Vakarcs et al. (1994), which is the same as the "intra-Messinian unconformity event" of Csató et al. (2007, 2013). Both phenomena are early Messinian in age (ca. 6.8 Ma). They could only be correlated with the MSC if we suppose that the geochronological framework of the Pannonian Basin is highly unreliable, and that the latest Messinian shelf-margin should be looked for in a much more northerly position than it is thought today (Csató et al., 2015) (Fig. 14). Such a deep revision of Pannonian Basin chronostratigraphy is not inconceivable, but only if supported by independent data; an *a priori* correlation between the MSC and any erosional feature in the Pannonian Basin, especially if contradicting the present-day chronostratigraphic knowledge, can be highly misleading.

6. Caspian Basin response to the Messinian Salinity Crisis

Of all the Paratethyan basins, the Late Miocene Caspian Sea events are probably the most spectacular. Highlights include an 800 km basinward shift of the Paleo-Volga deltaic system and the complete isolation of the Caspian Sea for the majority of the Pliocene. Naturally, such a dramatic event has been correlated to the Messinian Salinity Crisis in the Mediterranean Sea and clear similarities exist between the two events. There are however also crucial differences and uncertainties. Here we will discuss both similarities and differences to better differentiate the internal and external drivers of the Late Miocene Caspian basin evolution.

6.1 South Caspian Basin evolution and depositional environments

The Caspian Sea is divided into three sub-basins. The North Caspian Basin contains the oldest deposits (predominantly Upper Paleozoic to Permian) and forms a stable platform with low relief throughout the Late Cenozoic (Kroonenberg et al., 2005). Deposition continued, with regular interruptions during erosive periods, into the Quaternary (Popov et al., 2010; Yanina, 2014). The Middle (or Central) Caspian Basin has the thinnest sedimentary cover and is most likely underlain by continental crust (Green et al., 2009). The Apsheron Peninsula and submarine Apsheron Ridge (the eastward continuation of the Greater Caucasus range) form the divide between the Middle and South Caspian Basins (Fig. 15). The South Caspian Basin is the main depocenter of the Caspian Sea during the Late Miocene and Pliocene.

Sediments in the South Caspian Basin have a maximum total thickness of 20 km, mostly represented by Cenozoic deposits (Brunet et al., 2003). The basement is formed by either thinned continental crust or oceanic crust (Mangino and Priestley, 1998; Abdullayev et al., 2015). Thinning of the crust probably occurred in a back-arc setting (Allen et al., 2002). The age is poorly constrained, but is most likely Jurassic (Zonenshain and Le Pichon, 1986; Brunet et al., 2003; Green et al., 2009). Half of the total sedimentary thickness of the South Caspian Basin is deposited in the Late Miocene and Pliocene.

Similar to the Black Sea region, deposits of the Oligocene-Lower Miocene Maikop series represent the initial stages of the Eastern Paratethys due to the (partial) isolation from the global ocean. Onshore records in the Gobustan region of Azerbaijan show deep-water facies (Popov et al., 2008). The highly bituminous deposits form the main source-rock for the South Caspian oil-province (Hudson et al., 2008, 2016; Johnson et al., 2009; Bechtel et al., 2013). In a tectonic sense the Maikop Series act as a major ductile detachment level for Pliocene structures (Allen et al., 2003). Middle and Late Miocene deposits in the South Caspian region are generally referred to as the Diatom Suite. This stage encompasses the Karaganian, Konkian, Volhynian, Bessarabian, Khersonian and Maeotian regional stages as known from the Black Sea (Jones and Simmons, 1996). Brackish-marine, anoxic/disoxic conditions prevailed during most of the Diatom Suite and the associated mudstones have source-rock potential. Erosive valleys are found at this time on the northern margin of the Caspian Sea, interpreted to represent short-lived sea-level falls in which rivers cut valleys down to a depth of several hundred meters (Popov et al., 2010).

The Pontian regional stage of the Caspian Sea forms the final highstand moment before the northern margin of the Caspian Sea moved 800 km southwards from a position north of the Central Caspian to the South Caspian (Fig. 15). A second depocenter probably remained in the North Caspian Basin (Sidnev, 1985). In the center of the South Caspian Basin up to 5000-7000 m of non-marine clastic sediments were deposited in fluvial and deltaic environments during the Pliocene (Reynolds et al., 1998; Hinds et al., 2004; Abdullayev et al., 2012). Three major rivers (Volga, Kura and Amu Darya) sourced these large amounts of clastic non-marine sediments. These sediments are referred to as the Productive Series in Azerbaijan and the Red Colour Series in Turkmenistan (Abreu and Nummedal, 2007; Torres, 2007). Together they form the main reservoir levels for the South Caspian oil province (Devlin et al., 1999).

The Late Miocene base-level drop in the Caspian region has been ‘tentatively correlated’ in time to the MSC events based on the Pontian deposits underlying it (Jones and Simmons, 1996). As a result the connection to the Black Sea was lost and the Caspian Sea became isolated. Estimates of the total fall in sea-level vary between 50 m and 2000 m (Jones and Simmons, 1996; Reynolds et al., 1998; Buryakovsky et al., 2001; Allen et al., 2002; Green et al., 2009; Popov et al., 2010; Abdullayev et al., 2012) (Fig. 16a). The widely varying estimates for the water level drop associated to this event have led to significantly different hypotheses for Late Miocene and Pliocene basin evolution. Most notably, the enormous amounts of Pliocene and Pleistocene siliciclastic deposits in the South Caspian Basin have both been interpreted as the cause and the effect of the onset of high Pliocene subsidence rates in the South Caspian Basin (Allen et al., 2002; Brunet et al., 2003; Green et al., 2009).

Brunet et al. (2003) assume a 2.5 km water depth in the South Caspian Basin for the Oligocene in their subsidence model. This is then gradually reduced to the present-day maximum depth of ~1000 m due to sediments filling the depression. A 1000 m decrease in the water depth is placed at the onset of the Productive Series (Fig. 16b). They however note the large uncertainty on depositional water depths which is their largest error in the calculations, and emphasize this should be refined in future. In this scenario, a fall of the sea-level is caused by an externally forced MSC event resulting in the desiccation of the basin and the supply of Productive Series sediments is the trigger for high subsidence rates. Due to the base level fall, the basin profile gets steeper, allowing for larger amounts of erosion and a higher sediment-flux into the basin.

In contrast, Allen et al. (2002) infer a water depth of only 50 m during the Pontian for a backstripping exercise. A MSC drawdown is not invoked. Instead, they argue for a large increase in tectonic subsidence in the South Caspian Basin to explain the sudden onset of high accumulation rates. They interpret the subsidence to be the effect of a 5 ± 2 Ma onset of subduction of the South Caspian Basin to the north underneath the Middle Caspian region (Allen et al., 2002; Jackson et al., 2002; Knapp et al., 2004). The accommodation space created by tectonic subsidence was then rapidly filled with Pliocene coarse clastic sediments.

More recent work applies both concepts to explain the crustal structure and subsidence history. Compression related tectonic subsidence appears to have important regional effects, but the impact differs significantly throughout the South Caspian Basin (Abdullayev et al., 2015). However, no consensus has been reached whether the trigger for high sedimentation rates in the Pliocene relates to enhanced tectonic subsidence due to compression or subduction, or to an externally forced Messinian drawdown (Egan et al., 2009; Green et al., 2009; Abdullayev et al., 2012). The difference in bathymetry between the two endmember models is substantial, and advances in field-based studies have provided important constrains.

6.2 Sedimentation patterns suggesting a minor sea-level drop during the MSC

The main feature in Late Miocene-Pliocene basin evolution of the Caspian Sea is the switch from Late Miocene highstand conditions (up to and including the Pontian regional stage) to the lowstand of the Pliocene Productive Series (Fig. 17). Sedimentological studies have mostly focused on Productive Series outcrops on and close to the Apsheron Peninsula in Azerbaijan, where over 100 years of hydrocarbon exploration has resulted in a very complete description of the local (sub-) surface (Baganz et al., 2012). In addition to field observations, well log data are used in two comprehensive studies describing the sedimentology of the Productive Series (Reynolds et al., 1998; Hinds et al., 2004). The total thickness of the Productive Series on the Apsheron Peninsula is around 1500 m, considerably less than in the basin center where it may become up to 5 km thick (Hinds et al., 2004). The Productive Series is divided into a lower, middle and upper subdivision, with most of the oil-reservoir sands concentrated in the middle part. The many smaller subdivisions in this scheme are interpreted to be divided by regionally extensive sequence boundaries and flooding surfaces (Reynolds et al., 1998).

Due to the lack of (macro-) fauna in the Productive Series, biostratigraphic correlations cannot be made to the other margins in the South Caspian Basin nor with other basins (Jones and Simmons, 1996). Instead, stratigraphic subdivisions are made on a lithostratigraphic basis. While this allows for basin-scale correlation (Abdullayev et al., 2012), it does not allow for direct absolute age determinations and correlation to events outside the basin.

Sediments of both pre- and post-sea-level drop are found in the middle of the Gobustan region, referred to as the Adzhiveli section (Trubikhin, 1989). Here, Pontian sediments are continuously exposed in a series of outcrops (Van Baak et al., 2016b). A depositional water depth of 50 to 200 m is determined based on ostracod and mollusc assemblages in the section. A cyclostratigraphic age model at this locality allows detailed (resolution <100 kyr) correlation to MSC events in the Mediterranean Sea and shows that deposition occurred continuously between 6.2 Ma and ~5.4 Ma. Water depth estimates based on ostracod assemblages show that a sea-level drop in the order of 100 m occurred at 5.6 Ma (Van Baak et al., 2016b). The Pontian deposits in outcrop are overlain by a hiatus of unknown duration possibly related to a second water level drop.

The Gobustan record shows an intra-Pontian water level drop at ~5.6 Ma in the order of 100 m (Van Baak et al., 2016b), which correlates in time to the sea-level drop in the Black Sea (Krijgsman et al., 2010; Van Baak et al., 2015a) and the Mediterranean Sea during the acme of the MSC (Roveri et al., 2014a). The magnitude of this water level drop is in agreement with the postulated drop of ~50 m by Allen et al. (2002), but far less than the commonly cited 1500-2000 m sea-level drop in the Caspian Sea that has previously been proposed in relationship with the complete isolation of the basin (Jones and Simmons, 1996; Brunet et al., 2003; Abdullayev et al., 2012). Most importantly, there is no field evidence for a desiccation of the Caspian Sea during the acme of the MSC at 5.6 Ma.

This is not to say that a large base level drop never occurred. It is possible that the hiatus on top of the Pontian deposits is the expression of a much bigger water level drop, but in that case it is younger than the MSC event in the Mediterranean Sea.

6.3 Seismic studies of the Mio-Pliocene interval

Seismic studies are used to further constrain the magnitude of a late Miocene / Pliocene sea-level drop in the Caspian Sea. The evidence presented in literature in support of a 1500-2000 m water level drop, and how it relates to the MSC is also disputed for the South Caspian Basin. Pre-base level drop (late Miocene) deposits are evident on the northern margin of the middle Caspian basin where they form a highstand clastic shelf margin (Reynolds et al., 1998; Abdullayev et al., 2012). This is overlain by a basin-wide series of unconformities which indicate the 800 km shift of the coastline to the south (Fig 15). This exposed much of the Middle Caspian and caused deposition to concentrate in the South Caspian Basin. Stratigraphically this represents the change from Pontian to Productive Series (Reynolds et al., 1998; Abdullayev et al., 2012).

The interpretation of the complete subaerial origin of the basal erosional surface is the main argument for a complete desiccation of the Caspian Sea at the onset of the Productive Series. A commonly used argument to highlight the extent of the base level fall is the horizontal distance between the position of the Late Miocene and Pliocene margins. The first onlapping sediments of the overlying Pliocene Productive Series can be found 700-800 km south of the preceding margin. While this horizontal difference is impressive, it does not constrain the vertical component of the associated sea-level drop. For instance, the Late Pleistocene Caspian Sea repeatedly experienced shoreline shifts of similar horizontal magnitude, yet base level changes were in the order of 150-200 m (Fig. 18; Yanina, 2014). The depositional gradient on which the Productive Series sediments were deposited was extremely gentle throughout the Pliocene, with no evidence for shelf-break processes occurring near the Apsheron Peninsula (Baganz et al., 2012). The large horizontal movements of the Late Miocene and Pliocene delta systems are therefore not necessarily associated with large base level variations.

The height of clinoforms of the late Miocene clastic shelf margin should be able to approximate the water depth in which it formed, as used to study paleo-water depths in the Central Paratethys (e.g. Magyar et al., 2013) and Dacian Basin (Fongngern et al., 2015). However, no evidence has been given that a single clinoform of 1500-2000 m height is present in the Caspian Basin prior to the base level fall. Instead, in the Middle Caspian, late Miocene clinoforms are typically 150-200 m high (Kholodov et al., 1992). These are overlain by the regional erosional surface (surface A in Fig. 19) separating Miocene from Pliocene deposits. This basal erosional surface represents a significant duration during the Pliocene, a much longer time period than just the MSC climax at 5.6 Ma. The maximum incision of this surface is about 500-600 m (e.g. Abdullayev et al., 2012). Correlation to the South Caspian outcrop records would suggest a correlation to the post 5.4 Ma water level drop. The basal infill is overlain by Plio-Pleistocene Akchagyl and younger sediments which record multiple moments of downcutting and infill.

The Middle Caspian clinoforms show a minimal water depth of 150-200 m. Adding to this the 500-600 m of maximum incision in the Middle Caspian region (e.g. Abdullayev et al., 2012) would at maximum get to 600-800 m of base level fall. Correctly extrapolating the amplitude of the base level fall from the Middle Caspian towards the south into the South Caspian Basin depends on the relative impact of water level drop and tectonic subsidence. As shown previously, there is no consensus on this topic, with the result that calculating the relative contributions is highly impacted by over- or underestimations of the amount of tectonic subsidence.

A model based on a scenario of only water level drop, thereby largely excluding tectonic subsidence, has been proposed by Green et al. (2009) and Abdullayev et al. (2012). Based on backstripping the sedimentary section, water depths of up to 2000 m were calculated for the Pontian (Abdullayev et al., 2012). In this reconstructed basin profile, the first onlapping sediments may have been 1500 meters lower than the previous water level (Green et al., 2009). This would indicate that the gentle ramp fluvio-deltaic sediments of the Productive Series observed at and around the Apsheron Peninsula were deposited in a land-locked basin 1500 m below contemporary global sea-level (Green et al., 2009). However, this is an order of magnitude deeper than the outcrop data indicating depositional water depths prior to base level fall of a maximum of 200 m (Van Baak et al., 2016b). Clearly this is not a satisfying solution and needs to be better constrained in future work.

6.4 The Paleo-Volga canyon

One of the key arguments in this discussion is the northward extension of the canyon of the Paleo-Volga river. This canyon at the scale of the Grand Canyon formed in the Late Miocene / Pliocene and has since been completely filled with sediment to now form a buried feature in the Russian subsurface (Sidnev, 1985; Matoshko et al., 2004; Kroonenberg et al., 2005). Erosional features can be traced northward from the Middle Caspian Basin to north of the latitude of present-day Moscow (over 2000 km). The study of this major erosive feature and its infill has, in part, been possible due to a large number of wells drilled for the construction of hydroelectric dams during the 1940's-1980's. Depths of almost 600 m were obtained while drilling before reaching the underlying basement (Sidnev, 1985) (Fig. 18). In the Middle Caspian Basin, this conspicuous seismic feature cuts 600 m deep into Paleogene and Mesozoic bedrock (Abdullayev et al., 2012).

In the context of this review, we had a closer look at the work of Sidnev (1985) who provides (in Russian) a comprehensive overview of the state of knowledge in the 1980's of the Paleo-Volga canyon. We georeferenced the maps and digitized the paleo-relief contours to form a digital 3D model of the erosive relief of the Paleo-Volga region (Fig. 18a). For the canyon in the Middle Caspian basin, the depth map of Svitoch & Badyukova (2004) was used, as presented by Kroonenberg (2013). From the model we extracted both the Mio-Pliocene depth, and present-day elevation profiles along the trace of the entire Paleo-Volga canyon (Fig. 18b). This approach leads to some interesting insights of the Paleo-Volga canyon.

The Paleo-Volga canyon is not one continuous 2000 km long erosive feature, but it should be subdivided into at least two separate erosive valleys: 1) the Middle Volga region (and areas further upstream) with a maximum depth of incision of ~200 m, and 2) the Middle Caspian Basin erosive feature. In between these two erosive features a basin was present, the Northern Pre-Caspian Basin. The combination of rapid deepening and multiple parallel erosive valleys on the northern flank of this Northern Pre-Caspian Basin suggest that it was actively being eroded on the margins. Sediments overlying the erosive contact indicate a lake basin remained in the Northern Pre-Caspian Basin in the Kimmerian (Sidnev, 1985). These sediments provide valuable information on the nature of (part of) the infill of the Paleo-Volga canyon and support the interpretation of the Northern Pre-Caspian Basin as a lake basin during (part of) the Kimmerian (Sidnev, 1985).

The oldest sediments found in the North Caspian Pliocene are typically referred to as the Kinel Fm (Danukalova, 2012). Kinel Fm deposits are interpreted to represent a range of fresh-water environments, deposited after the maximum erosion of the erosive valleys (Sidnev, 1985). Uncertainty exists about the basal age of this formation and resulting from that, the age of maximum incision of the Paleo-Volga erosive features (Sidnev, 1985). The first option is Late Pontian (?) – Early Kimmerian, which would make it roughly time equivalent to the peak of the MSC. Evidence for this is however rather poor and seems to be mostly based on correlation to Late Pontian regressive features elsewhere. Based on paleomagnetic studies of the infill of the canyon in the Kama / Belaya region, another option is given for the age of the oldest deposits (Sidnev, 1985 and references therein). Normal polarities in the lowermost deposits were correlated to the fifth paleomagnetic epoch, representing chrons C3An.1n and/or C3An.2n. If this correlation is correct, the age of the oldest sediments above the Paleo-Volga erosion is between 6.0 and 6.7 Ma, which corresponds to the Maeotian, and predate the MSC events. This agrees with the idea that erosion started during the Late Sarmatian (Sidnev, 1985), and makes at least parts of the Paleo-Volga canyon a pre-existing feature during the MSC.

Kroonenberg et al. (2005) point out that a large base level drop may not be necessary to explain the Paleo-Volga Canyon. If the basin during the Pontian was shallow, only a limited amount of water would have been present in the Caspian Sea. The onset of rapid subsidence at 5±2 Ma may therefore have triggered sea-level fall and erosion of a canyon by the Paleo-Volga river. Subsequent sea-level falls may therefore have been of similar depth, but since it happened in a steeper basin profile, they did not necessarily reach the same southern extent. The northward expansion during the Productive Series is also easily explained by a constant sea-level and a steepening basin gradient over time. This hypothesis is however the opposite of the basin profile reconstruction recently published by Abdullayev et al. (2012) who assumes the basin gradient to become lower during deposition of the Productive Series.

In the context of both shallow and deep South Caspian Basin hypotheses, the Volga canyon is supposed to have formed completely subaerially. However, the interpretation of any erosional feature as having a subaerial origin for the entire formation history is an oversimplification of actual geological processes. For the Volga Canyon, no reference is made to possible submarine erosion, even though this is a common feature throughout the world's oceans (Jobe et al., 2011; Harris et al., 2014), on the margins of the present-day and Late Miocene Black Sea (Tari et al., 2015, 2016; Krezsek et al., 2016), Mediterranean Sea (Canals et al., 2006), and on the seafloor of the present-day Caspian Sea (Richardson et al., 2011). For the Mediterranean MSC, canyons of the rivers Nile (Chumakov, 1973), As Sahabi in Libya (Nicolai, 2008; Bowman, 2012), Ebro (Urgeles et al., 2011), and Rhone (Clauzon, 1982) have all been proposed as important arguments for subaerial erosion processes and hence deep desiccation. Here an alternative has also been proposed, explaining these major canyons as submarine erosion as the result of dense shelf water cascading into the deep basin (Roveri et al., 2014b, 2016). As mentioned before in the Pannonian Basin discussion, the Late Miocene Alpar canyon in Hungary has a similar depth to the Volga canyon of up to 600-700 m, yet is interpreted to be largely formed due to submarine erosion (Juhász et al., 2013).

Large parts of the northern Caspian region submerged during the Pontian highstand only to emerge during the Kimmerian and the deposition of the Productive Series. This means that in a large part of the Paleo-Volga canyon system both submarine and subaerial erosion processes should have been active in the Late Miocene and Pliocene. As a result, the maximum incision of the canyon (prior to the deposition of the infilling Kinel Fm) cannot simply be related to subaerial erosion and the lowering of the base level. Only in the northern part of the canyon (north of Saratov and therefore north of the area flooded at the base of the Pontian) a case could be made that subaerial erosion was dominant. In that case, the maximum depth of erosion is roughly 200 m, which would constrain the base level fall for the Northern Pre-Caspian Basin.

7. Hydrological budget and Paratethys response to the Messinian Salinity Crisis

When connected, base level and environments in the Paratethys are directly linked to the evolution of the Mediterranean Sea. Changes in basin connectivity can lead to different mechanisms of lowering base level in the Paratethys. As long as the sea-level in the Mediterranean is above the height of the sill, Mediterranean and Paratethys will be connected and the Paratethys will have an inflow of both marine waters (from the Mediterranean) and freshwater (from the continent). Salinity will be between freshwater and marine depending on the relative contribution of the two components. Paratethys water level will be at the height of the Mediterranean. When however Mediterranean sea-level drops below the sill, due to uplift of the sill or a drop of Mediterranean sea-level, the Paratethys water level will depend on the regional hydrological budget. In case of a positive local budget (river input + precipitation > evaporation) base level will be at, or slightly above the height of the sill. The salinity of the Paratethys basins will go down and become essentially freshwater. If the local hydrological budget in the Paratethys is negative, the water level will drop below the sill and stabilize at a level where the evaporation is equal to the total inflow. This will convert the Paratethys to an endorheic lake, with local base level highly dependent on climatic changes in the drainage basin, similar to the present-day Caspian Sea. These mechanisms work both on the Paratethys as a whole, and between the individual Paratethys sub-basins.

For example, isolating the Mediterranean Sea from the Atlantic Ocean today would create a major negative hydrological budget, resulting in a base level drop of 2.5 kilometer within 6 kyrs (Meijer and Krijgsman, 2005). It is possible that this happened to the Mediterranean in the Late Miocene. In the current chronostratigraphic framework this would be limited to a 50 kyr interval between 5.6 and 5.55 Ma around glacial peaks TG12 and TG14 (Roveri et al., 2014a). Given the highly negative hydrological budget of the Mediterranean, even this short period would be enough for a serious base level drop.

The hypothesized time equivalent base-level fall in the Paratethys is thought to be caused by the resulting isolation of the Paratethys from the world ocean (Jones and Simmons, 1996). However, this seems contradictory when using the present-day Caspian Sea as an analogue. The Caspian Sea has no connection to the global ocean and is the world's largest lake, with river water as the only input into the basin (Dumont, 1998). The inflow of river water is sufficient to stabilize the Caspian Sea at a level of -26 meters. The Volga river delivers ~80% of the influx of fresh water with the catchment of the Volga extending into the humid mid-latitudes. Here moisture is supplied by westerly winds and thus the main moisture source should be looked for in the North Atlantic and not in the Mediterranean Sea (Arpe and Leroy, 2007; Renssen et al., 2007). Similarly the Black Sea receives water from this northern area through rivers like the Don, Dniepr and Dniestr.

If the present-day Volga river is capable of preventing the Caspian Sea from total desiccation, why would the Pliocene Paleo-Volga River, which had to be at least equal in volume to be able to deposit the Productive Series sediments, not be capable of keeping the Caspian Sea at a relatively stable high water level? A similar argument has previously been voiced by Kroonenberg et al. (2005), who argued that the total volume of sediment of the Productive Series can only originate from a humid drainage area. Similarly in the northwestern Black Sea, a major northern sourced delta system was present, depositing the Balta Fm (Matoshko et al., 2016).

Regionally higher humidity for the latest Miocene compared to the present-day has both been argued for in climate models (Gladstone et al., 2007; Marzocchi et al., 2016), and from paleo-precipitation proxies (Böhme et al., 2010, 2011). Having regionally higher humidity during the Late Miocene makes desiccation more difficult. It therefore seems counterintuitive to imagine that a dessication of the Paratethyan basins could have occurred simply as the result of isolation from the global ocean.

The link between the hydrological budget and resulting sea-level drop in the Late Miocene Paratethyan basins was recently explored in a modeling study (De La Vara et al., 2016). The model setup allowed to explore at which level the base level would stabilize for a full range of different hydrological budgets. For this, both the river-runoff and precipitation-evaporation (P-E) could be varied individually. In the model, the Pontian paleogeographical reconstruction of Popov et al. (2006) was converted to bathymetry and sea-surface area profiles. Using the Late Miocene climate model results from Marzocchi et al. (2016) realistic numbers could be attributed to the different components of the hydrological balance. The model showed that under isolated conditions, the Late Miocene Paratethys would have a positive hydrological budget, i.e. more water coming in than going out through evaporation over the sea-surface area. Isolating the late Miocene Paratethys from the world ocean is therefore unlikely to have caused desiccation.

According to these calculations, the most realistic way of significantly lowering the water level in the Caspian Sea would be to not have the drainage of the Volga contribute towards the hydrological budget. This would require the paleo-Volga River to be captured at an earlier stage, for instance in the North Caspian basin, or drain towards the Black Sea instead. Only in this case a water level drop of over 200 m was calculated for the Caspian Sea. This however leads to a problem when trying to explain the paleo-Volga derived sediments of the Productive Series at the bottom of a desiccated Caspian Sea. For the deep desiccation to have happened, no Volga water was allowed to enter the basin while for the sediment, the water into the basin was a necessity. Clearly, this shows that the three pillars of the Late Miocene Caspian Sea basin evolution, 1) a major base level fall, 2) large amounts of sediments delivered to the South Caspian Basin and 3) a humid northern catchment with big rivers, do not appear to have been able to act simultaneously.

Very similar processes acted in the Pannonian Basin during the MSC. Here, another major river (the Danube) supplied large amounts of freshwater to Lake Pannon as evidenced by basin-wide prograding sediment systems. The infill of this basin was a continuous process in the Late Miocene and Pliocene (Magyar et al., 2013), with no indication of a major desiccation related to the MSC.

8. Conclusion and outlook

Since the first discovery of salt deposits in the deep basins of the Mediterranean Sea, the understanding of the Messinian Salinity Crisis has significantly increased. In many cases this was a stepwise process, where new data conflicted with previously established models (Rouchy and Caruso, 2006; Ryan, 2009; Roveri et al., 2014a). This urged the need to rethink previously established hypotheses, which inevitably led to multiple steps forward in the total understanding. An astronomically tuned chronology has proven to be the vital basis for the present understanding of this unique oceanographic event. Many questions however remain and it has become increasingly clear that a complete understanding of the Mediterranean Sea is impossible without a similarly detailed knowledge of the Paratethys region to the north.

In the Mediterranean Sea, high-resolution, biostratigraphically controlled, absolute age determinations are key to critically assess hypotheses on basin evolution. Originally, theories were put forward suggesting the coeval desiccation of the individual Paratethyan basins. At the same time however, the introduction in the Mediterranean of Paratethyan fauna in the final Lago Mare stage of the MSC suggested Paratethys had a positive hydrological budget. Improvements in chronology have shown that hypothesized MSC events in the Paratethys are not necessarily coeval with the MSC events in the Mediterranean Sea. By increasing the age model resolution for records throughout the Paratethys, the classical model of a desiccating Paratethys during the MSC has been seriously challenged in the last decade. As a general rule, the a priori correlation between regionally significant erosional surfaces and the MSC can be highly misleading. Determinations of late Miocene erosional surfaces as caused by the MSC have in many cases been disproven by increased age model resolution.

The long history of both academic and industry focus has led to a very well established scheme for inter- and intra-basinal correlations. However, as a result there is a significant risk of pigeonholing new observations into previously established hypotheses. In many cases previous hypotheses may be underlain by data obtained with less accurate methods than presently available and should therefore not necessarily be taken as fact. Quality control and critical utilization of data in studies with an integrated approach is of vital importance to avoid favoring chance encounters over solid data. Reproducible data should always be the starting point for hypotheses.

Complications have arisen in cases where subsurface chronostratigraphy is based on lithostratigraphic arguments. This has resulted in studies using the chronostratigraphic terminology established in outcrops for unrelated subsurface sequences. To avoid the potential confusion, the ongoing aim should be a better integration of biostratigraphic and seismic data. The Pannonian Basin is a good example how this close integration acted as a game changer for understanding the regional subsurface.

Correlation of any erosional feature to subaerial erosion can be highly misleading given that similar processes occur in submarine settings. A good example is the recent re-interpretation of erosional features in the Black Sea, indicating no evidence is found for major river incisions during the MSC (Tari et al., 2016). Instead, many of the major erosional features present on seismic data are better interpreted as submarine canyon systems (Tari et al., 2016). As a result, basin-scale interpretations following the early claims for deep desiccation of the Black Sea during the MSC (Hsü and Giovanoli, 1979), will need to be revised.

More generally, the observation of a regressive system does not necessarily need to be interpreted as a forced regression (i.e. a base level drop). The Paratethys system is a tectonically active system of semi-isolated basins in which major rivers supply large amounts of sediment. Both can cause regression of the depositional system under stable base level conditions. Basin-to-margin correlations are vital to establish the evolution of the system and with high enough temporal resolution should allow the distinction between these options.

The gradual separation of individual Paratethyan sub-basins throughout the Neogene has led to a system which is extremely susceptible to climatic changes in the hydrological budget and to tectonics in the shallow gateway regions. Distinguishing between these internal (geodynamics, tectonic uplift or subsidence, basin infill) and external (climate, glacio-eustatic sea-level change) forcing mechanisms acting on sea-level and basin connectivity of the Paratethyan semi-isolated seas is therefore an ongoing challenge requiring integration of many sub-disciplines of geoscience. All of this shows the need to keep increasing age model resolution to better understand basin evolution.

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Figure Captions

Introduction

1. Location map of the present-day Mediterranean Sea with (in blue) the Paratethys Pontian (late Miocene) paleogeography of Popov et al. (2004). In yellow is the present-day drainage of the Black Sea and in grey the Caspian Sea and Aral Sea drainage. Major rivers and mountain ranges are indicated. White dots indicate locations of cored material during DSDP Leg 13 (Sites 124 and 134) and DSDP Leg 42B (Sites 380A), yellow dots outcrop localities described later in the text. Abbreviations: PB: Pannonian Basin, DB: Dacian Basin, FD: Focsani Depression, ZR: Zheleznyi Rog section (Taman Peninsula, Russia), AZ: Azerbaijan sections.

   (b) Photographs and original caption of the core material used as early evidence for the Messinian dessication theory of the Mediterranean Sea (after Hsü et al., 1973). The correct orientation of this core has been debated (Lugli et al., 2015).

   (c) Photograph and original caption of the core material used as early evidence for the Messinian dessication theory of the Black Sea (after Hsü and Giovanoli, 1979).

2. Time scales for the Mediterranean and Paratethys basins for the late Miocene and early Pliocene and their relationship with the MSC stages (after Krijgsman et al., 1999). The Paratethyan basin time scales are biostratigraphically defined regional stages and substages. The exception to this are the Pliocene Productive Series of the Caspian Sea, due to a lack of fauna, this is purely defined on a lithostratigraphic basis. Dashed lines indicate uncertainties of the regional stage boundaries or boundaries under discussion. These will be discussed in the dedicated sections of this paper. Abbreviations used are PB: Pannonian Basin, DB: Dacian Basin, BS: Black Sea (Euxinian Basin), CS: Caspian Sea, Od: Odessian, Pf: Portaferrian.

3. (a) The CIESM - Commission Internationale pour l'Exploration Scientifique de la mer Méditerranée (2008) Messinian salinity crisis stratigraphic framework with the 6 key-surfaces used for the definition of the MSC stages (after Van Baak et al., 2016b; adapted from Manzi et al., 2013) allowing for the correlation between marginal and deep basins.

   (b) Deep basin seismic expression of the MSC in the western (Gulf of Lions) and eastern (Levant) basins of the Mediterranean Sea (Lofi et al., 2005; Bertoni and Cartwright, 2007; Roveri et al., 2014a). Basin to margin correlations are complicated and rely on tracing polygenic erosional surfaces. These examples from the deep basin show the two typical expressions of the MSC and their conformable surfaces. In the western basin a clear distinction can be made between three units, the Lower Unit (equivalent to RLG and/or partially the PLG); the Mobile Unit, corresponding to the Messinian salt; and the Upper Unit, the top of which has been drilled for scientific purposes and contains Paratethyan ostracod fauna. In the eastern basin only the Mobile Unit is found.

Euxinian (Black Sea) Basin

4. The Pontian outcrop at Zheleznyi Rog, Taman Peninsula, Russia and its age model. The field photograph (N45.114867°, E 36.749472°, looking southwest) shows the dominant regular alternations of darker and lighter marls typical for the Pontian. Two key surfaces are indicated, the lumachelle level at the biostratigraphic boundary between Lower and Upper Pontian, and the coastal sands overlying the Pontian.

   Two age models have been proposed for the Zheleznyi Rog section, here shown next to each other using the log of Chang et al. (2014), and correlated to the MSC stages and astronomical reference curves (Laskar et al., 2004). Maker beds are numbered according to the publications of Chang et al. (2014) and Popov et al. (2016) and logs have been stretched to best line up with geological time. The main difference between the two age models is the additional 200 kyr hiatus at the lumachelle level as proposed by Rostovtseva and Rybkina (2017). As a result, the age of the Upper Pontian is significantly different between the two models.
5. The evolution of the age model of DSDP 380/380A in the southwestern Black Sea. The original hypothesis of Hsü and Giovanoli (1979) explained the coarse deposits in Unit IVd (the Pebbly Breccia unit; 865-885 mbsf) as the result of a MSC desiccation event (in red). This was disputed by subsequent analysis, and this specific level is likely older than the MSC.
6. Previously published proxy records from DSDP 380/380A documenting the continental and marine paleoenvironmental changes associated with the MSC and the transition from marine to fresh water environments. The horizontal colored bars indicate the position of the MSC stages when assuming a constant sedimentation rate throughout chron C3r. In orange the position of the main intervals of highly enriched deuterium isotopes of Vasiliev et al. (2015). The sea surface salinity tolerance of the diatom record of Schrader (1978) gives in blue the calculated highest and lowest known salinity tolerances of the species assemblage. In red an additional upper salinity estimate based on recalculating salinity with a more marine estimate of 17 ppm for *Coscinodiscus* (?) *stokesianus*, as proposed by Schrader (1978).

   Continental climate temperatures show a first peak at ~840 mbsf, afterwards stabilizing at colder conditions. Notably, an inverse trend is observed in the sea surface temperatures (SST), showing a significant increase and stabilizing afterwards. This increase in SST appears to coincide with a notable decrease in the distance to the coast (i.e. a water level drop) and a decrease in sea surface salinity (SSS) from ‘marine’ to ‘brackish’. Although not reflected in the SST, a further freshening (this time from 'brackish' to 'fresh') during a water level drop occurs at 805 mbsf. From roughly this point onwards, the trophic level becomes increasingly more eutrophic. A return of the position of the coastline to that of the 840-805 mbsf interval is observed at 790 mbsf, yet this does not appear to be related to an increase in SST or SSS.

7. Seismic architecture of the northwest Black Sea shelf (Romania) showing key Pontian events (adapted from Krezsek et al., 2016). (a) Well log in a similar marginal basinal position as the Zheleznyi Rog section on the Taman Peninsula, Russia. On the left the stratigraphic units as originally proposed, and on the right 'translated' to the terminology as used on the Russian Black Sea margin. (b) Seismic section showing that highstand Lower Pontian deposits have been eroded down to a level 500-600 m (~600 ms) below its top shelfal position. To the east-southeast this erosional feature continues as a horizontal surface covering a gravitational collapse system. In this sequence the down going surface (BES) is interpreted to represent the onset of a lowstand, and the horizontal (MES) to mark the beginning of transgression (Krezsek et al., 2016). (c) Location map with the approximate location of the seismic line (NWBS), the star indicates the location of the Focsani Depression (FD) in the Dacian Basin.

Dacian Basin

8. Maps of Dacian Basin. (a) Spatial variations in uplift and erosion along the Romanian Carpathians inferred from geothermochronology studies and thickness of foredeep sediments in the Romanian Carpathians foreland (adapted by Cloetingh et al., 2004; partly after Sanders et al., 1999). Time of uplift (elliptical boxes) and main moment of subsidence (square boxes) are indicated. (b) Isopach maps of the Dacian Basin at different stages from late Miocene to Pliocene (adapted from Saulea et al., 1969). (c) Dacian Basin paleogeographic maps for the Pontian (adapted from Marinescu and Papaianopol, 1989). Note that these maps show lithofacies only. (d) Paleogene to recent seismic sequences in the western Dacian Basin Getic Depression (after Leever et al., 2010). For location of the line see Fig. 8a. Two seismic sequences, characterized by well-developed clinoforms were distinguished in the post-orogenic sediments of the Getic Depression. They are of Maeotian - Lower Pontian (SSQ1) and Middle Pontian - Dacian (SSQ2) age (Leever et al., 2010). SSQ1 is bounded at the top by an erosional unconformity (SB2) **(**150-195 km). In the northern part of the section, transtensional and subsequent compressional deformation of the Uppermost Cretaceous – Middle Miocene sequence of the Getic Depression is observed (Leever et al., 2010) with activity along the Getic Fault likely continuing into the post-orogenic phase. Here, the Miocene-Quaternary sediments are gradually tilted up to 20 degrees due to late stage uplift and fault reactivation (Răbăgia et al., 2011). This could indicate that additional erosion

due to this tectonic activity affected the proximal parts of SB2, which was otherwise largely a sediment bypass surface during deposition of LST2.

9. Ostracod range chart of the Rimnicu Sarat section in the Focsani Depression of the Dacian Basin, Romania, and the inferred paleoenvironmental evolution across the MSC (modified from Stoica et al., 2013). For this publication, ostracods have been reorganized according to their first occurrence. This clearly indicates three main intervals of shallow basinal depositional conditions, the lower two occur during paleomagnetic chron C3r, a third starts during the overlying Thvera subchron. These shallow basinal conditions are separated in the section by intervals with littoral, fluvial and lacustrine conditions, as reflected in the poor ostracod assemblages and from sedimentological observations in the section (Stoica et al., 2013).
10. (a) The Iron Gates "Gilbert-Type delta" model of Clauzon et al. (2005). Indicated are the outcrops with their stratigraphic contacts on which the model is based. (b) Geological map of the study area, with the locations of the outcrops used in (a) to propose the Gilbert-Type delta model (Jipa et al., 2011). bn: Badenian; sm: "Sarmatian"; me: "Maeotian"; pt: "Pontian"; dc: Dacian; rm: Romanian. (c) Reinterpretation of the depositional model of same outcrops as proposed by Jipa et al. (2011) and Ter Borgh et al. (2014). Instead of one connected system representing a Gilbert-Type delta formed during desiccation and reflooding of the Dacian Basin caused by the MSC event, these outcrops represent unrelated parts of the Middle Miocene to Pliocene basin evolution (modified from Jipa et al., 2011).

Central Paratethys

11. The Pannonian Basin with the known maximum extent of Lake Pannon in the mid-Tortonian, and the reconstructed profundal region of the lake in the early Messinian and early Pliocene. Dotted lines indicate the southward shift of the shelf break due to normal regression. A, B, and C indicate positions of seismic profiles in the following figures.
12. The seismic architecture in eastern Hungary which is most often referred to in MSC debates; profile (a) and horizon slice (c) from a 3D cube and various stratigraphic interpretations of the phenomena (b). (a) The red horizon represents an originally almost horizontal paleosurface in the shelf just below the Miocene/Pliocene unconformity; it was extended into a 3D horizon, and the horizon was shifted 150 ms downwards so that it cuts clinoforms in the entire 3D area (see the yellow horizon). (b) Interpretations of the seismic profile in "a". *: unconformity, coinciding with the only falling shelf-edge trajectory on the section; much older than the MSC event. **: IMU of Csató et al. (2007, 2013), indicated here because of the

"onlapping LST wedge", although the corresponding shelf-edge trajectory is flat to rising with a set of tectonically tilted parallel paleo-horizontals (shelf reflections). The shelf edge belonging to IMU is also older than the MSC. Age data is based on Magyar et al, (2013), fs: flooding surface, mfs: maximum flooding surface. (c) A map view of the yellow horizon in "a", indicating the strike of clinoforms. Location of profile in Fig. 11a.

13. Seismic profile focused on the shallow subsurface in eastern Hungary. The unconformity indicated by yellow arrows runs about 280-300 m below the surface and separates the Upper Miocene from the Plio-Pleistocene. According to seismic correlations, it corresponds to the updip continuation of the red horizon in Fig. 12a and the Miocene/Pliocene unconformity in Fig. 12b. Location of profile in Fig. 11b.
14. A depositional dip directed composite seismic line and its interpretation from SE Hungary. (a) Interpretation of the original seismic profile. (b) The original seismic profile, flattened to a delta plain surface below the M/P unconformity which before deformation might have been approximately "horizontal". (c) Interpretation of the flattened profile. Age data refer to the shelf break. The Late Messinian is characterized by prograding and aggrading clinoform sets. No descending shelf-margin trajectory is observed to indicate major lake-level fall. Location of profile in Fig. 11c.

Caspian Sea

15. Middle and South Caspian Basin paleogeographical reconstructions prior to, and after the expansion of the Paleo-Volga delta into the south basin with respect to the present-day Caspian Sea and major rivers. The Pontian map and the active tectonic thrusts are based on Reynolds et al. (1998), the Productive Series (PK) map and the outline of the Paleo-Volga Canyon is based on Abdullayev et al. (2012). What becomes clear is that the outline of the erosional feature is largely confined to the area that was submerged during the Pontian, following the deep basinal facies towards the south. On the other hand, the deposition of the Productive Series sediments is confined to the South Caspian Basin, limited to the north by active thrust tectonics at the eastward prolongation of the Greater Caucasus mountains. Depositional facies have been as best as possible converted from the original publications to show the same in both panels. The star indicates the location of the outcrops near Baku, Azerbaijan on which most of the fieldwork studies are based.
16. (a) Estimates from different papers of the amount of water level drop in the Caspian Sea region related to the MSC. (b) Contrasting bathymetry records proposed for the Apsheron Peninsula region back in time to the Eocene (0-50 Ma) on which basin evolution models are

based. The Brunet et al. (2003) curve invokes a large 1200 m water level drop during the MSC to explain the switch to Productive Series coarse clastics. Alternatively, the Allen et al. (2002) interpretation is that the basin was shallow throughout this time period, having near-completely filled up with sediment prior to the MSC. Renewed tectonic subsidence at ~5 Ma is in this model proposed as main driver in the formation of the Productive Series. The Buryakovsky et al. (2001) model is more in line with a minor water level drop, although it indicates a minimal drop of ~200 m.

17. Simplified stratigraphy of the Late Miocene-Pliocene Pontian-Productive Series time period in the Apsheron Peninsula region. The age model is not to scale and based on Van Baak et al. (2016). According to this age model, the climax of the MSC in the Mediterranean region does not coincide with a major water level drop. The main change and expansion of the Paleo-Volga delta postdates the Mediterranean water level drop, although the exact timing is unknown.
18. (a) Volga Canyon depth map based on the erosional contour maps of Sidnev (1985) for the Upper Volga to Lower Volga region and Svitoch and Badyukova (2004). The zero depth contour is indicated in red, with increasingly darker blue colors indicating greater depth to the base of Pliocene erosion. Well depths as originally reported by Sidnev underlie this model, but are not shown in this figure to reduce clutter. On top of the erosional contour map are the present-day water bodies and rivers, and the reconstruction of the late Miocene Pontian highstand (dark dotted after Popov et al., 2004). The Northern Pre-Caspian Basin clearly shows up, bounded by shallower maximum erosion levels to the north and south. It reaches maximum depths exceeding 700 m and a lake remained here after the water level drop (Sidnev, 1985). We therefore tentatively add a second Late Miocene coastline (dotted in light grey) around the north of the basin, to indicate this as an alternative position of the late Miocene shoreline.

    (b) Canyon depth profile along the line A-A’ for both the present-day elevation/bathymetry and the Mio-Pliocene erosion surface as extracted from the map. For the present-day profile, the bandwidth of Late Pleistocene Caspian Sea level variability is plotted. This 200 m of vertical variation corresponds to a nearly 1000 km horizontal change. The maximum depth of incision in the northernmost 1000 km is roughly limited to -200 m, before rapidly deepening going into the NPCB to a maximum depth of -600 m. The shallowing at around 1500 km into the canyon corresponds to a tectonically active zone in the Lower Volga region around Astrakhan. The second rapid deepening starting at ~2000 km coincides with the northernmost position of the Paleo-Volga canyon of Fig. 15. This is in roughly the same position as the present-day shelf margin, suggesting an important tectonic control on the position of this margin.

19. Seismic section oriented NE-SW across the Middle Caspian showing erosional structures and infill. The Pliocene regional seismostratigraphic horizon A incises the underlying bedrock, in this location down to the Oligo-Miocene Maykop. Further south, the main erosional valley cuts down into the Cretaceous (Abdullayev et al., 2012). The initial infill of canyon in the Middle Caspian remains largely undated. Drilling of the basal infill of the canyon suggests sandy lithologies, riverine conditions without datable features. Given the regional context, this likely represents the top of the retrogradational unit of the Paleo-Volga delta system. This backstepping of the Paleo-Volga delta to the north in the Late Pliocene flooded the Middle Caspian Paleo-Volga canyon (Abdullayev et al., 2012). Also note that in the higher sequences (e.g. Turkanian stage) multiple other erosional features indicate the recurrent erosion in this part of the basin.

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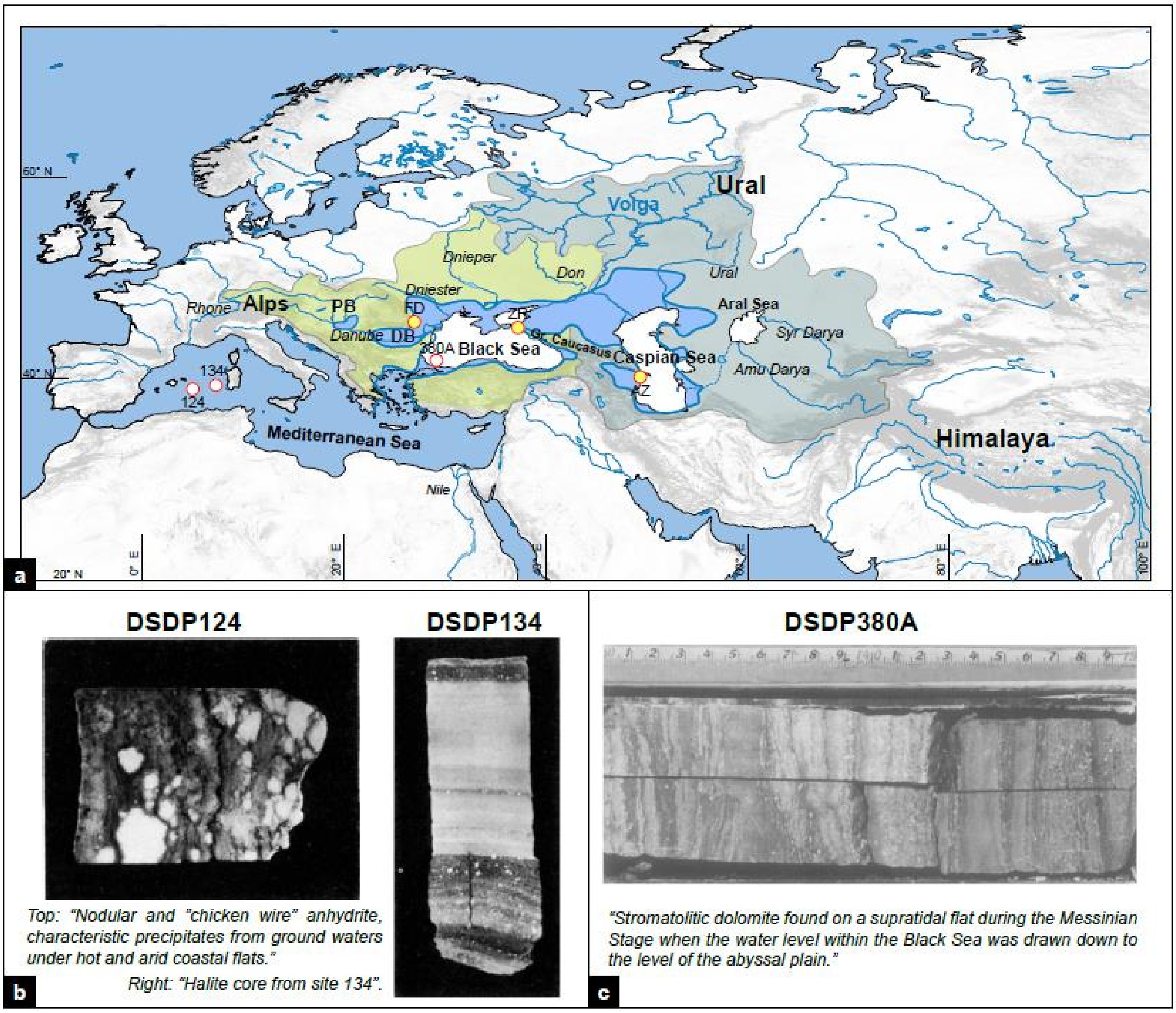


Ural
Volga
Dnieper
Don
Ural
Dniester
Rhone
Alps
PB
FD
ZR
Aral Sea
Syr Darya
Danube
DB
380A
Black Sea
Gr. Caucasus
Caspian Sea
Amu Darya
134
124
Z
Mediterranean Sea
Himalaya
Nile
60° N
40° N
20° N
0° E
20° E
80° E
100° E
a
DSDP124
DSDP134
DSDP380A
Top: "Nodular and "chicken wire" anhydrite, characteristic precipitates from ground waters under hot and arid coastal flats."
Right: "Halite core from site 134".
b
"Stromatolitic dolomite found on a supratidal flat during the Messinian Stage when the water level within the Black Sea was drawn down to the level of the abyssal plain."
c

Figure 1

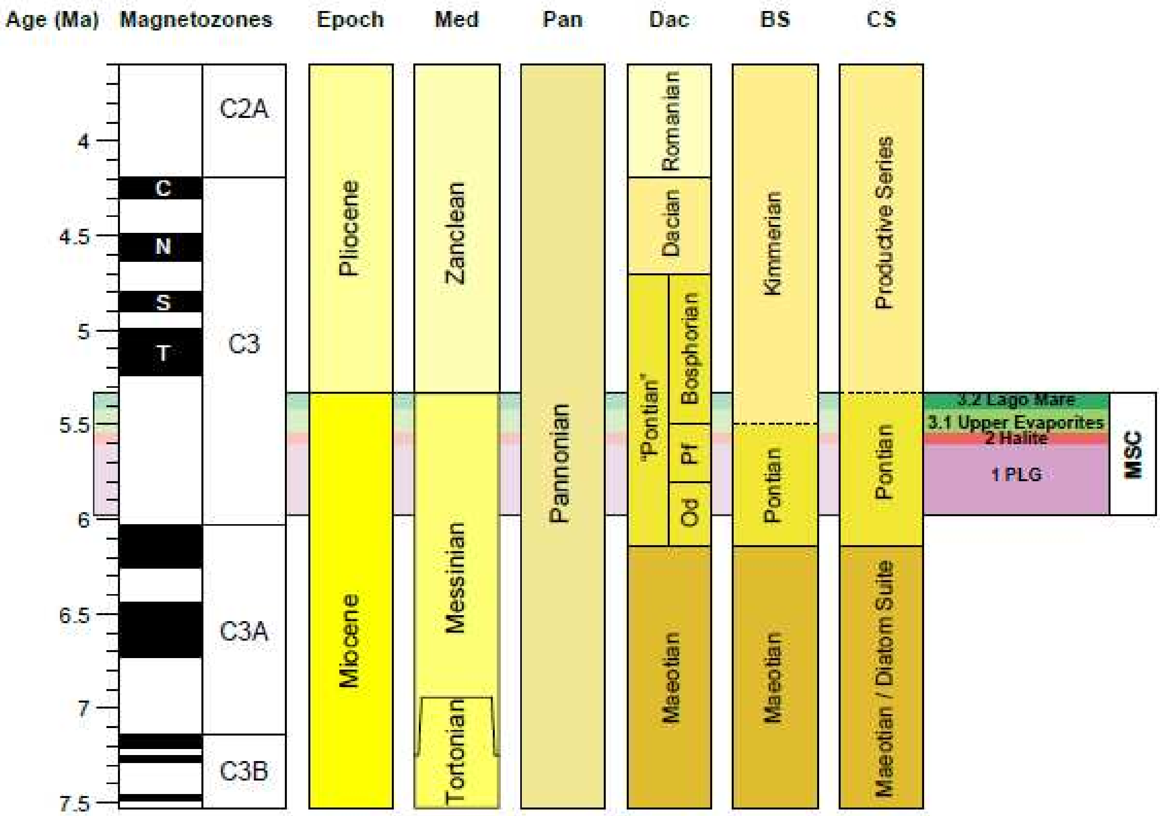
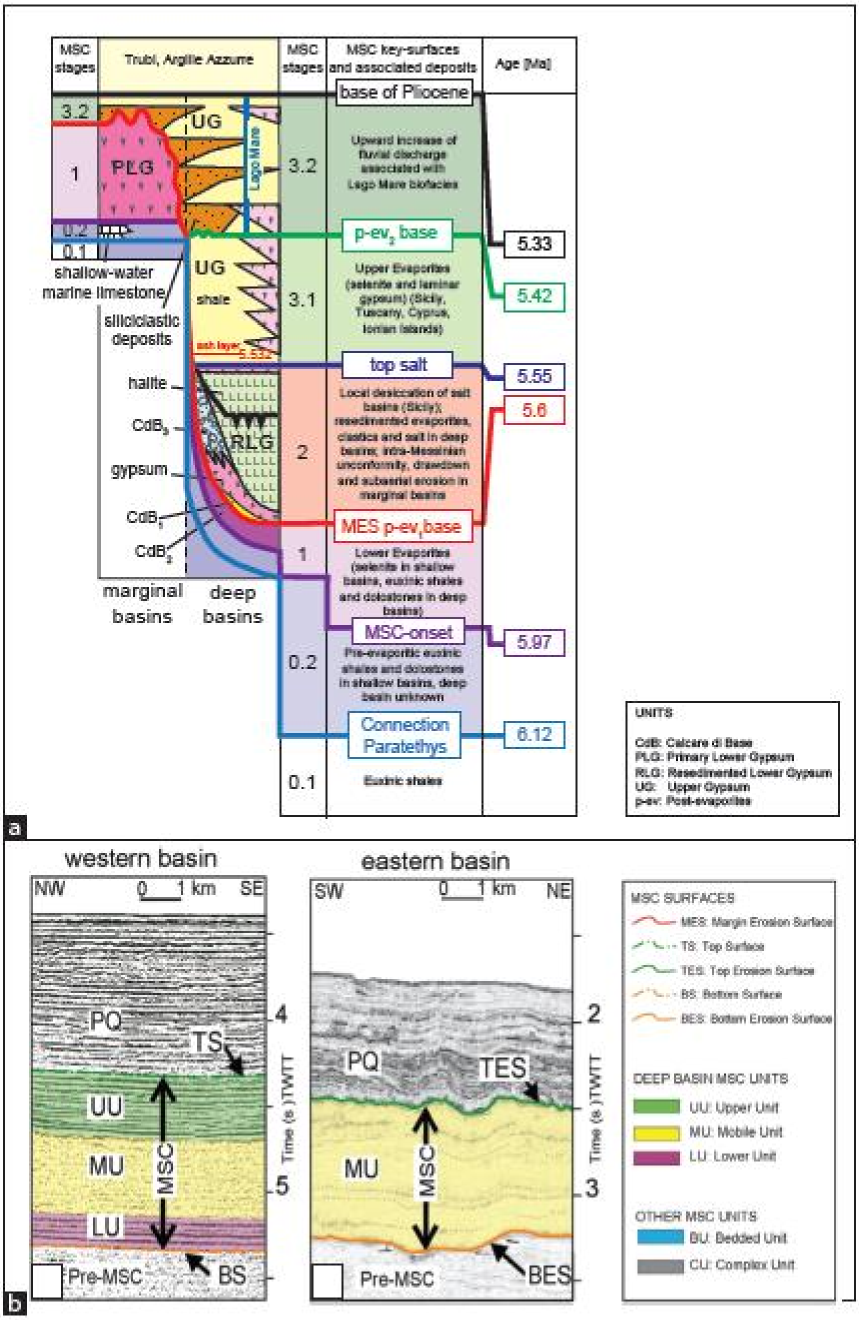

Age (Ma)
Magnetozones
Epoch
Med
Pan
Dac
BS
CS
4
4.5
5
5.5
6
6.5
7
7.5
C
N
S
T
C2A
C3
C3A
C3B
Pliocene
Miocene
Zanclean
Messinian
Tortonian
Pannonian
Romanian
Dacian
"Pontian"
Bosphorian
Pf
Od
Maeotian
Kimmerian
Pontian
Maeotian
Productive Series
Pontian
Maeotian / Diatom Suite
3.2 Lago Mare
3.1 Upper Evaporites
2 Halite
1 PLG
MSC

Figure 2



MSC stages
Trubi, Argille Azzurre
MSC key-surfaces and associated deposits
Age [Ma]
base of Pliocene
Upward increase of fluvial discharge associated with Lago Mare biofacies
p-ev2 base
Upper Evaporites (selenite and laminar gypsum) (Sicily, Tuscany, Cyprus, Ionian Islands)
top salt
Local desiccation of salt basins (Sicily); resedimented evaporites, clastics and salt in deep basins; intra-Messinian unconformity, drawdown and subaerial erosion in marginal basins
MES p-ev1 base
Lower Evaporites (selenite in shallow basins, euxinic shales and dolostones in deep basins)
MSC-onset
Pre-evaporitic euxinic shales and dolostones in shallow basins, deep basin unknown
Connection Paratethys
Euxinic shales
5.33
5.42
5.55
5.6
5.97
6.12
3.2
3.1
2
1
0.2
0.1
PLG
UG
RLG
Lago Mare
shale
ash layer 5.532
shallow-water marine limestone
siliciclastic deposits
halite
CdB3
gypsum
CdB1
CdB2
marginal basins
deep basins
UNITS
CdB: Calcare di Base
PLG: Primary Lower Gypsum
RLG: Resedimented Lower Gypsum
UG: Upper Gypsum
p-ev: Post-evaporites
a
western basin
NW
0 1 km
SE
PQ
TS
UU
MU
LU
MSC
Pre-MSC
BS
Time (s) TWT
4
5
eastern basin
SW
NE
TES
BES
2
3
MSC SURFACES
MES: Margin Erosion Surface
TS: Top Surface
TES: Top Erosion Surface
BS: Bottom Surface
BES: Bottom Erosion Surface
DEEP BASIN MSC UNITS
UU: Upper Unit
MU: Mobile Unit
LU: Lower Unit
OTHER MSC UNITS
BU: Bedded Unit
CU: Complex Unit
b

Figure 3

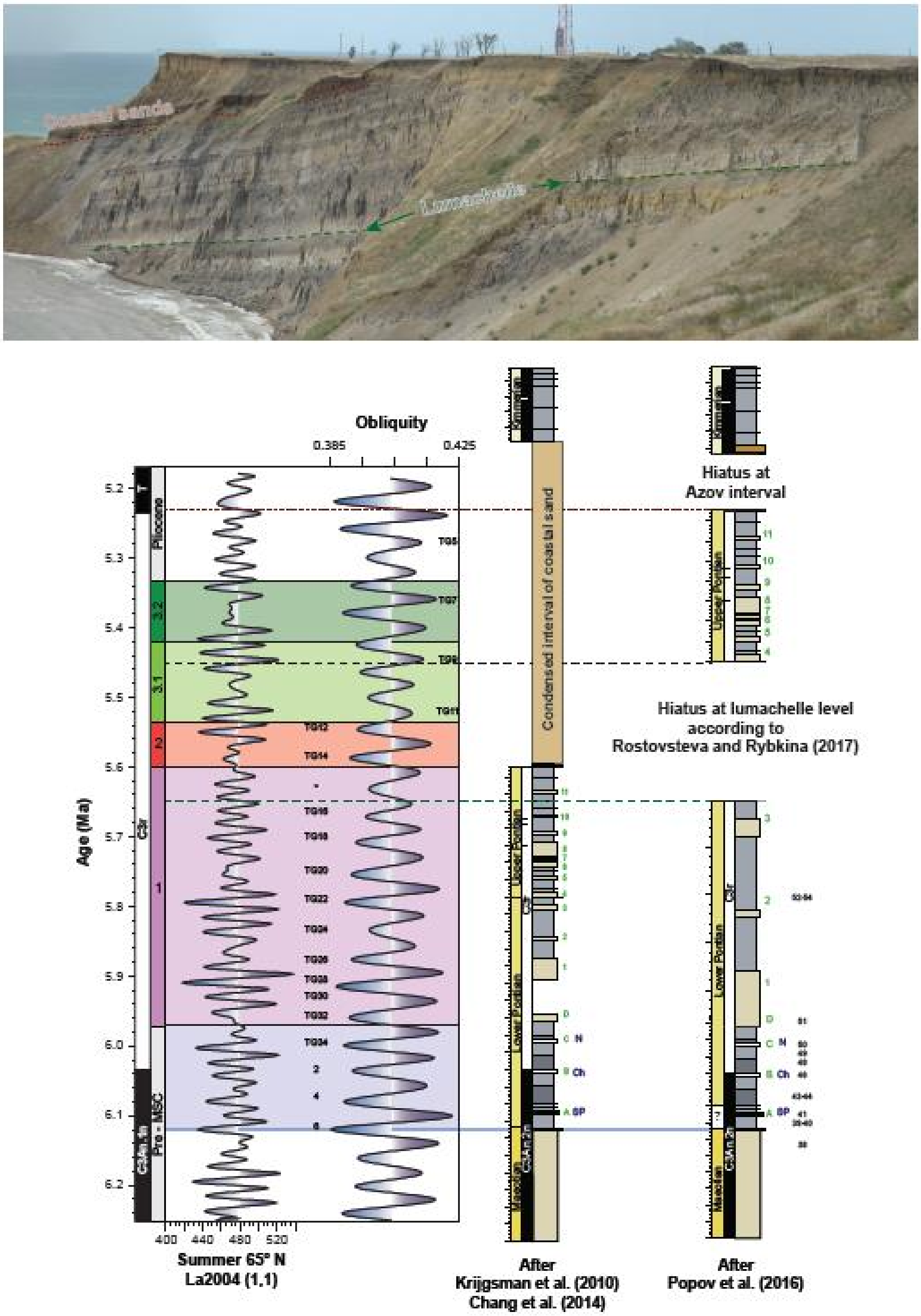
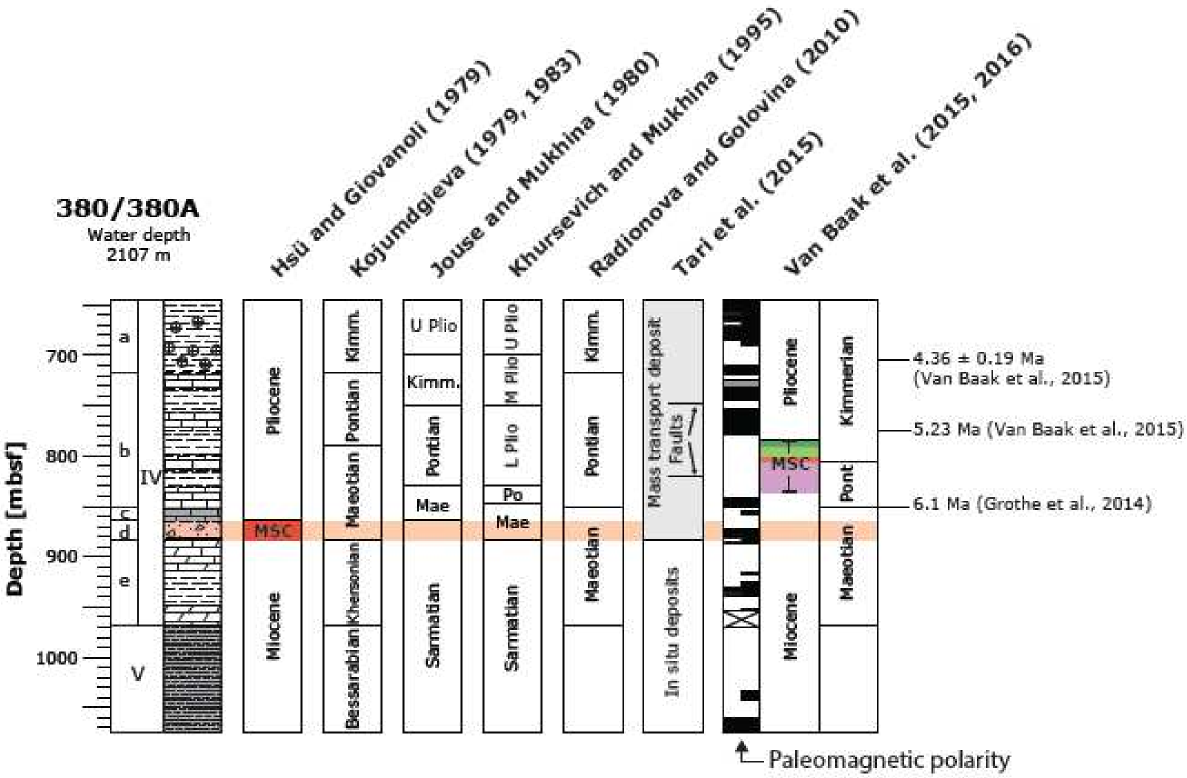
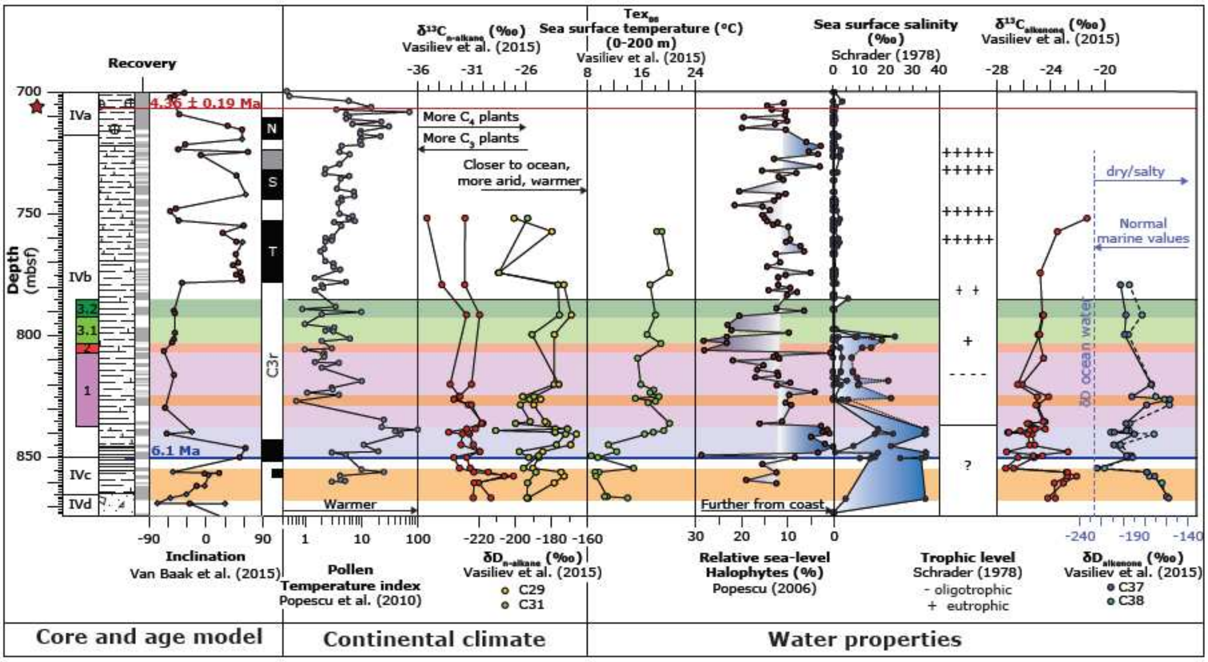

Lumachelle
Obliquity
0.385
0.425
Age (Ma)
Pliocene
Pre - MSC
Summer 65° N
La2004 (1,1)
Condensed interval of coastal sand
Hiatus at
Azov interval
Hiatus at lumachelle level
according to
Rostovsteva and Rybkina (2017)
Upper Pontian
Lower Pontian
Maeotian
Kimmerian
After
Krijgsman et al. (2010)
Chang et al. (2014)
After
Popov et al. (2016)

Figure 4



380/380A
Water depth
2107 m
Hsü and Giovanoli (1979)
Kojumdgieva (1979, 1983)
Jouse and Mukhina (1980)
Khursevich and Mukhina (1995)
Radionova and Golovina (2010)
Tari et al. (2015)
Van Baak et al. (2015, 2016)
Depth [mbsf]
700
800
900
1000
a
b
c
d
e
IV
V
Pliocene
MSC
Miocene
Kimm.
Pontian
Maeotian
Khersonian
Bessarabian
U Plio
Kimm.
Pontian
Mae
Sarmatian
U Plio
M Plio
L Plio
Po
Mae
Sarmatian
Kimm.
Pontian
Maeotian
Mass transport deposit
Faults
In situ deposits
Pliocene
MSC
Miocene
Kimmerian
Pont
Maeotian
4.36 ± 0.19 Ma
(Van Baak et al., 2015)
5.23 Ma (Van Baak et al., 2015)
6.1 Ma (Grothe et al., 2014)
Paleomagnetic polarity

Figure 5



Recovery
Depth (mbsf)
4.36 ± 0.19 Ma
6.1 Ma
IVa
IVb
IVc
IVd
3.2
3.1
2
1
N
S
T
C3r
Inclination
Van Baak et al. (2015)
Pollen
Temperature index
Popescu et al. (2010)
Warmer
δ13C n-alkane (‰)
Vasiliev et al. (2015)
More C4 plants
More C3 plants
Closer to ocean, more arid, warmer
δD n-alkane (‰)
Vasiliev et al. (2015)
C29
C31
Tex86
Sea surface temperature (°C)
(0-200 m)
Vasiliev et al. (2015)
Relative sea-level
Halophytes (%)
Popescu (2006)
Further from coast
Sea surface salinity (‰)
Schrader (1978)
Trophic level
Schrader (1978)
- oligotrophic
+ eutrophic
δ13C alkenone (‰)
Vasiliev et al. (2015)
dry/salty
Normal marine values
δD ocean water
δD alkenone (‰)
Vasiliev et al. (2015)
C37
C38
Core and age model
Continental climate
Water properties

Figure 6

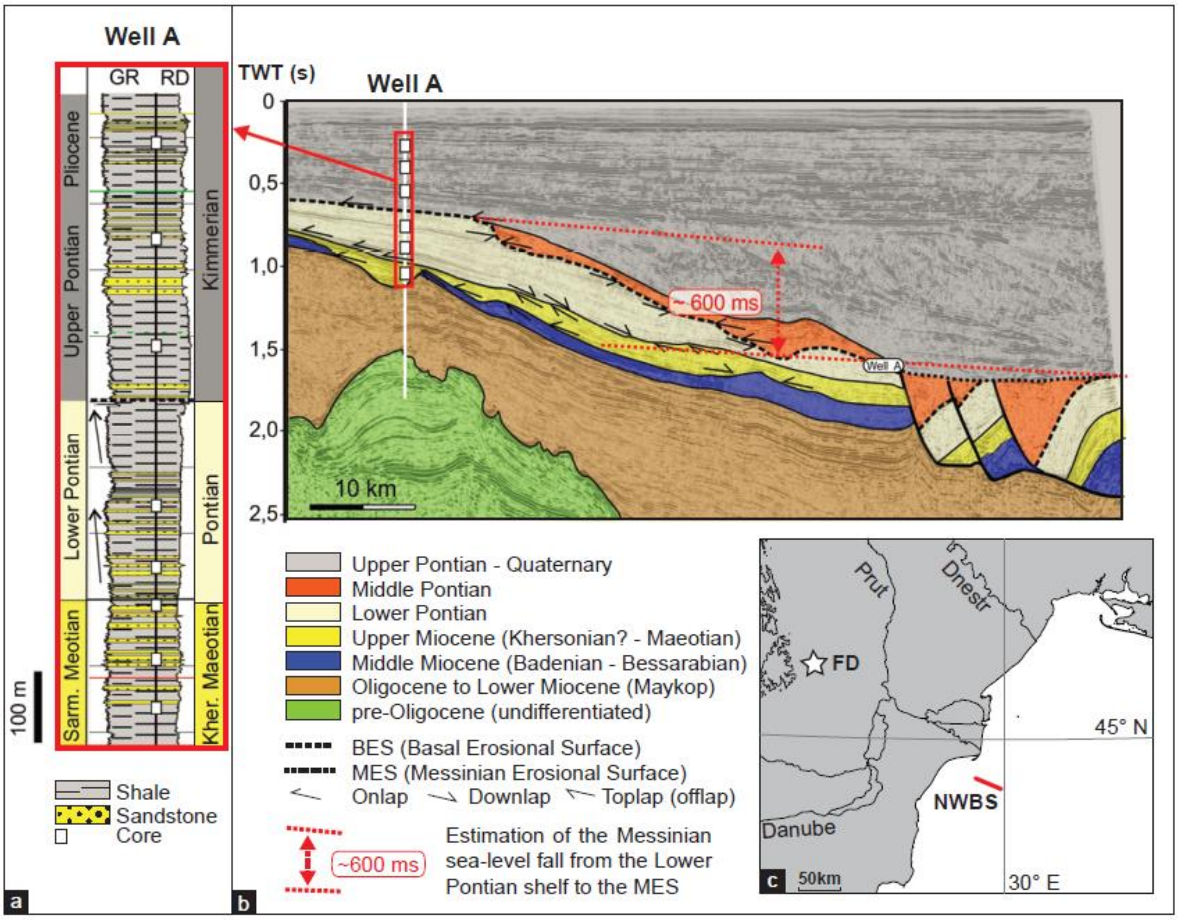

Well A
GR
RD
Pliocene
Upper Pontian
Lower Pontian
Sarm. Meotian
Kimmerian
Pontian
Kher. Maeotian
100 m
Shale
Sandstone
Core
a
TWT (s)
Well A
0
0,5
1,0
1,5
2,0
2,5
~600 ms
10 km
b
Upper Pontian - Quaternary
Middle Pontian
Lower Pontian
Upper Miocene (Khersonian? - Maeotian)
Middle Miocene (Badenian - Bessarabian)
Oligocene to Lower Miocene (Maykop)
pre-Oligocene (undifferentiated)
BES (Basal Erosional Surface)
MES (Messinian Erosional Surface)
Onlap
Downlap
Toplap (offlap)
~600 ms
Estimation of the Messinian sea-level fall from the Lower Pontian shelf to the MES
Prut
Dnestr
FD
45° N
NWBS
Danube
50km
30° E
c

Figure 7

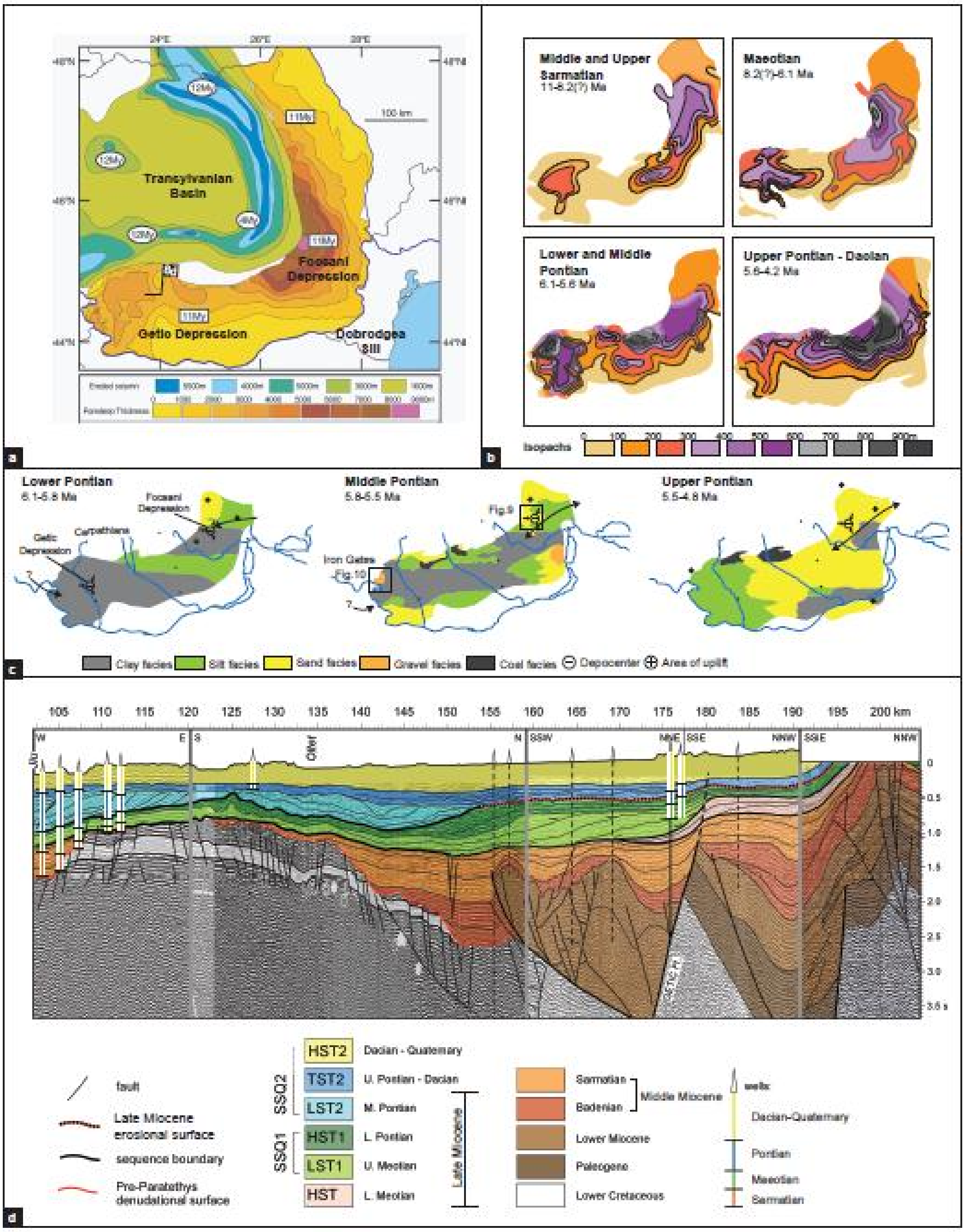


FIgure 8

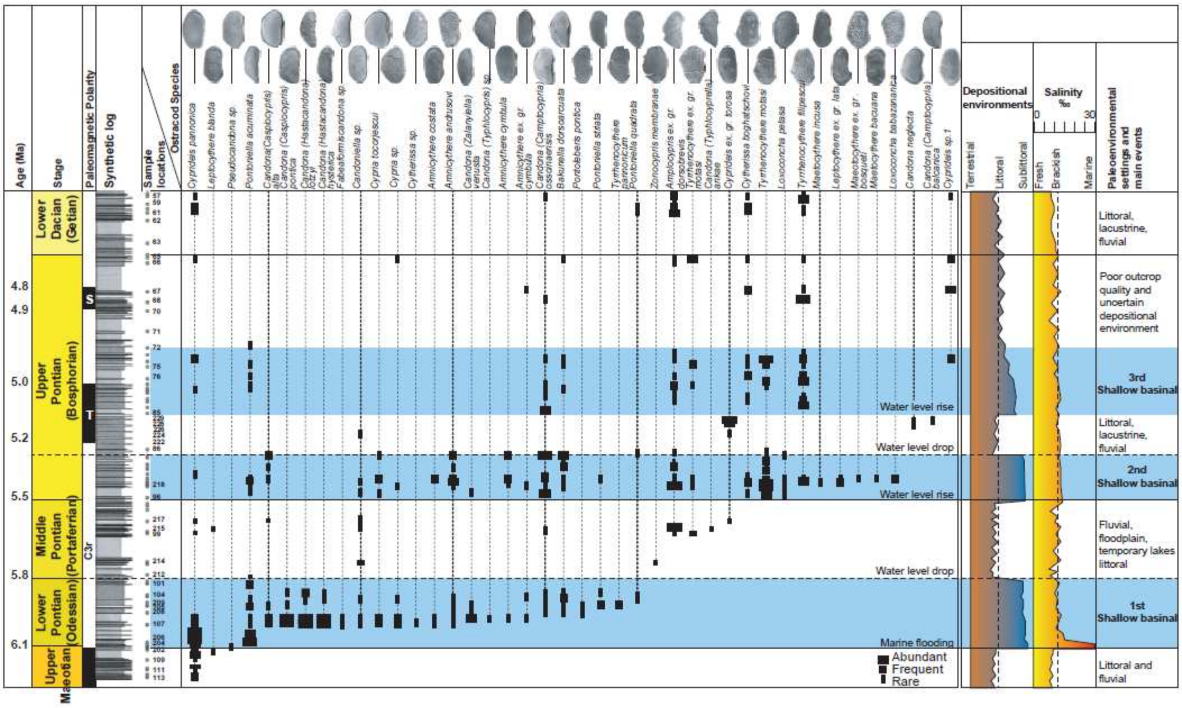
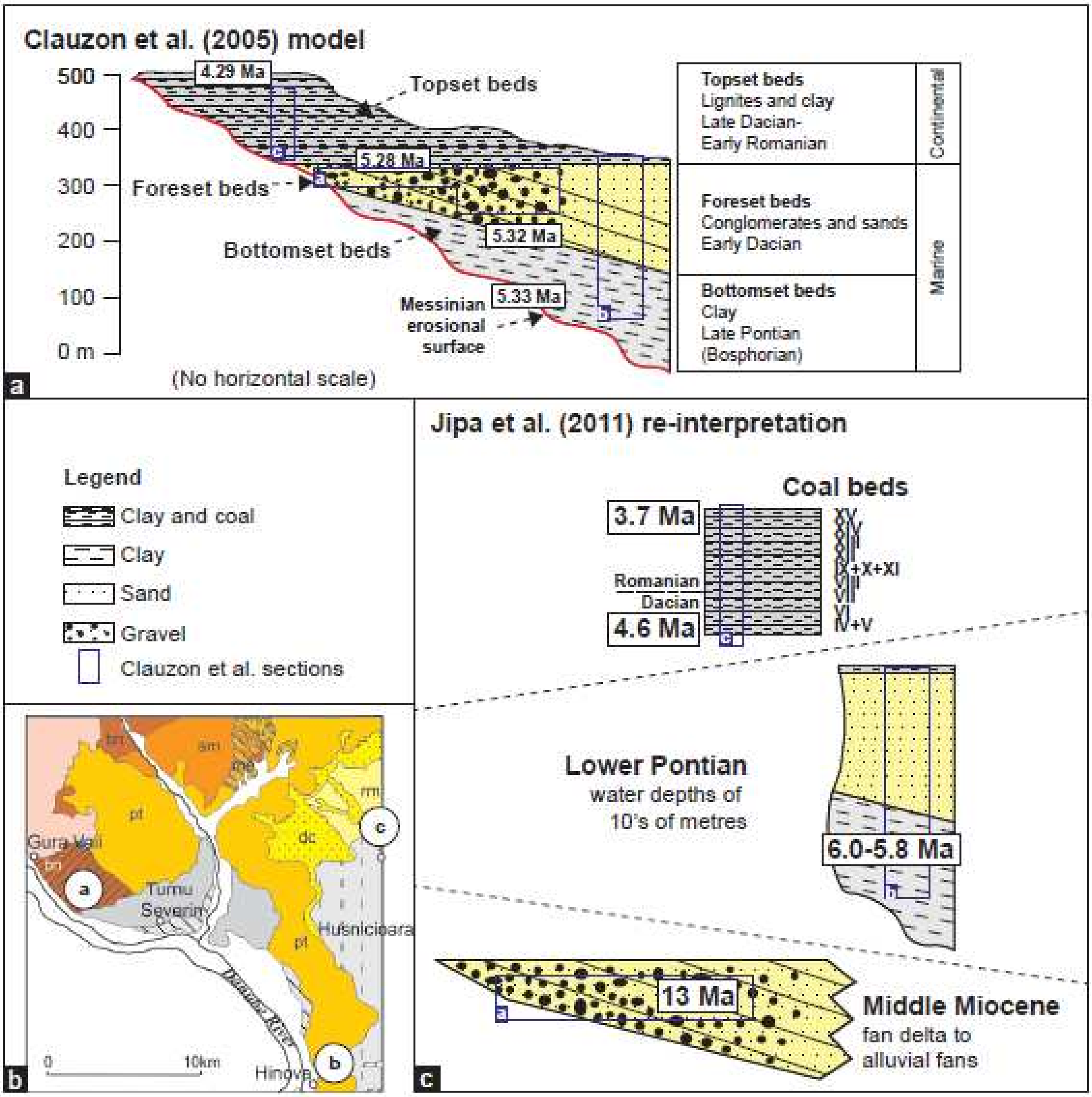
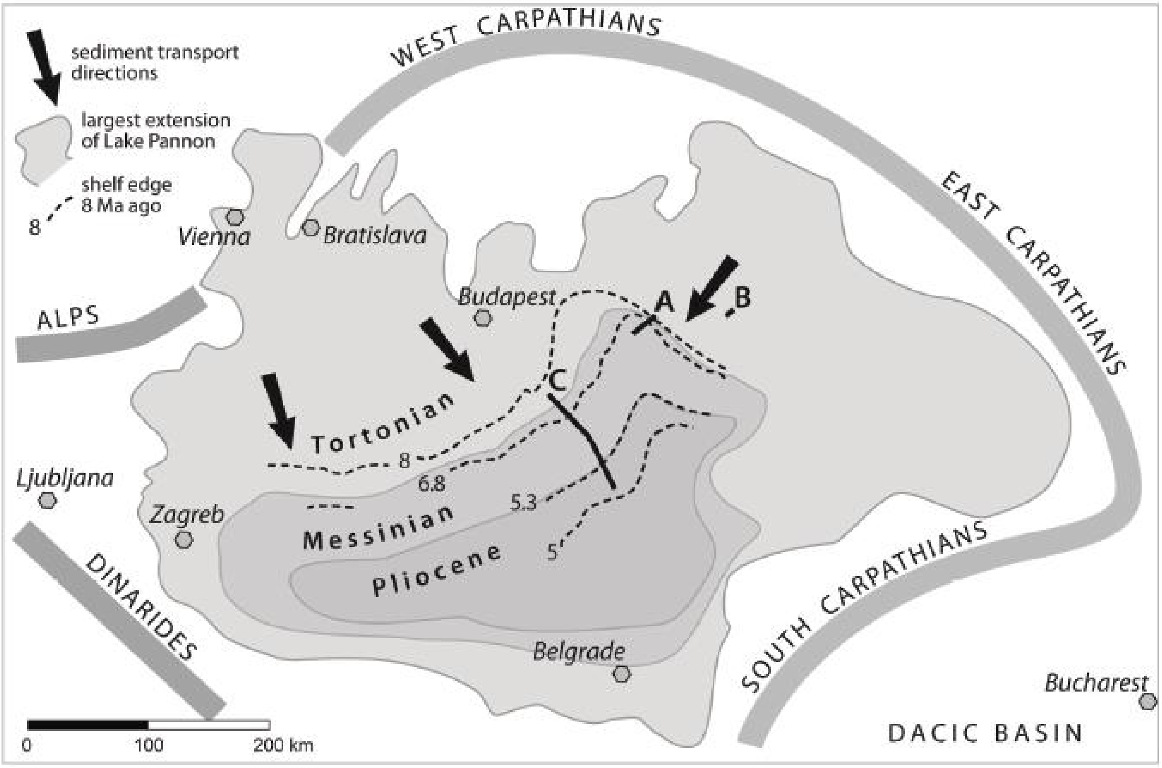

Age (Ma)
Stage
Paleomagnetic Polarity
Synthetic log
Sample locations
Ostracod Species
Depositional environments
Terrestrial
Littoral
Sublittoral
Salinity ‰
Fresh
Brackish
Marine
Paleoenvironmental settings and main events
Lower Dacian (Getian)
Upper Pontian (Bosphorian)
Middle Pontian (Portaferrian)
Lower Pontian (Odessian)
Upper Maeotian
4.8
4.9
5.0
5.2
5.5
5.8
6.1
C3r
Littoral, lacustrine, fluvial
Poor outcrop quality and uncertain depositional environment
3rd Shallow basinal
Littoral, lacustrine, fluvial
2nd Shallow basinal
Fluvial, floodplain, temporary lakes littoral
1st Shallow basinal
Littoral and fluvial
Water level rise
Water level drop
Water level rise
Water level drop
Marine flooding
Abundant
Frequent
Rare

Figure 9



Clauzon et al. (2005) model
500
400
300
200
100
0 m
4.29 Ma
Topset beds
5.28 Ma
Foreset beds
5.32 Ma
Bottomset beds
Messinian erosional surface
5.33 Ma
(No horizontal scale)
Topset beds
Lignites and clay
Late Dacian-
Early Romanian
Continental
Foreset beds
Conglomerates and sands
Early Dacian
Bottomset beds
Clay
Late Pontian
(Bosphorian)
Marine
a
Legend
Clay and coal
Clay
Sand
Gravel
Clauzon et al. sections
Gura Văii
Turnu Severin
Husnicioara
Danube River
0
10km
Hinova
b
Jipa et al. (2011) re-interpretation
Coal beds
3.7 Ma
Romanian
Dacian
4.6 Ma
IX+X+XI
IV+V
Lower Pontian
water depths of
10's of metres
6.0-5.8 Ma
13 Ma
Middle Miocene
fan delta to
alluvial fans
c

Figure 10



sediment transport directions
largest extension of Lake Pannon
shelf edge 8 Ma ago
8
WEST CARPATHIANS
EAST CARPATHIANS
SOUTH CARPATHIANS
ALPS
DINARIDES
DACIC BASIN
Vienna
Bratislava
Budapest
Ljubljana
Zagreb
Belgrade
Bucharest
A
B
C
Tortonian
Messinian
Pliocene
8
6.8
5.3
5
0
100
200 km

Figure 11

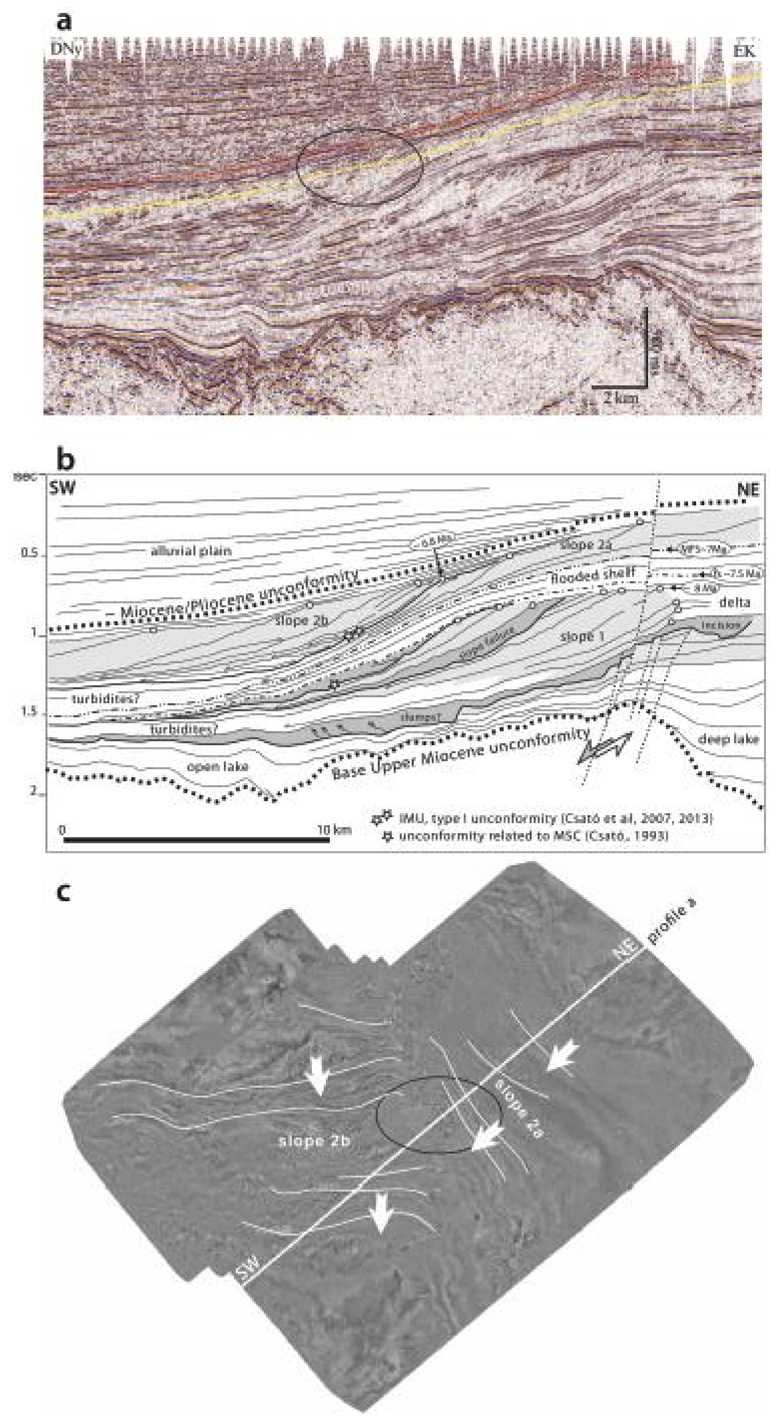

a
DNy
EK
2 km
b
SW
NE
0.5
1
1.5
2
alluvial plain
Miocene/Pliocene unconformity
slope 2a
slope 2b
flooded shelf
delta
incision
slope 1
turbidites?
turbidites?
slumps?
deep lake
open lake
Base Upper Miocene unconformity
IMU, type I unconformity (Csató et al, 2007, 2013)
unconformity related to MSC (Csató, 1993)
0
10 km
c
NE
profile a
slope 2a
slope 2b
SW

Figure 12

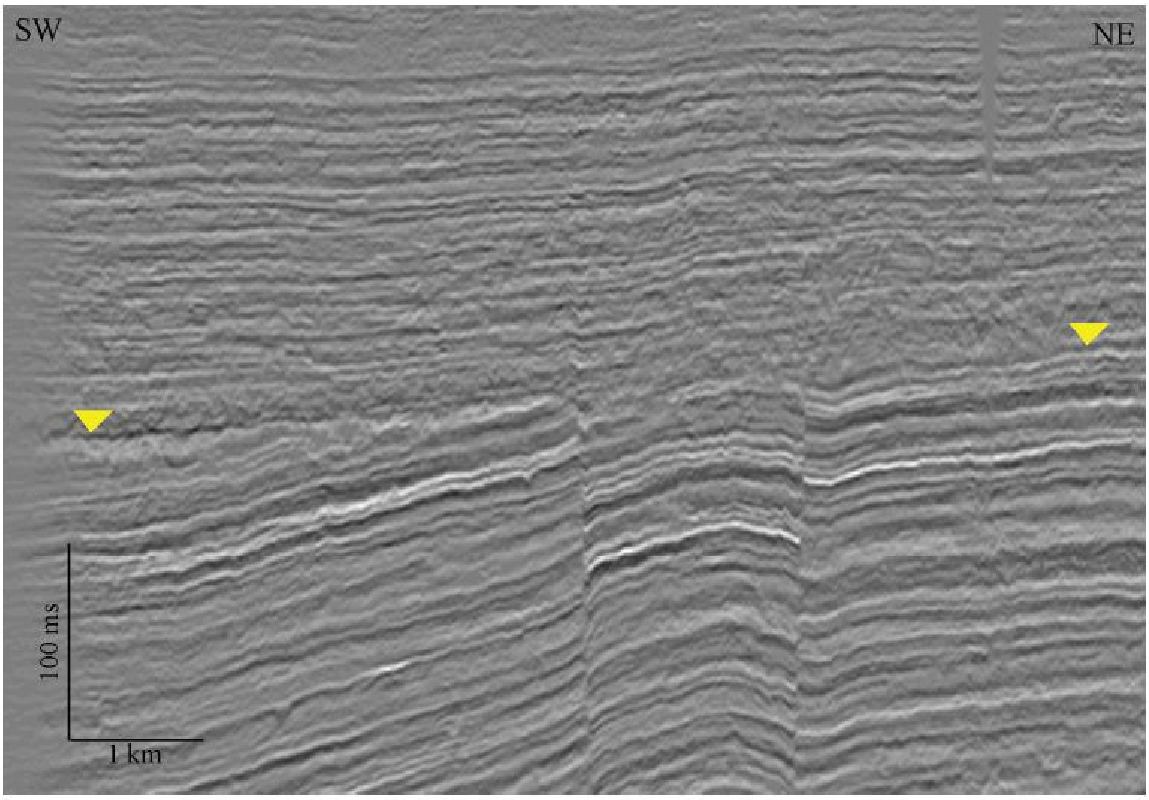
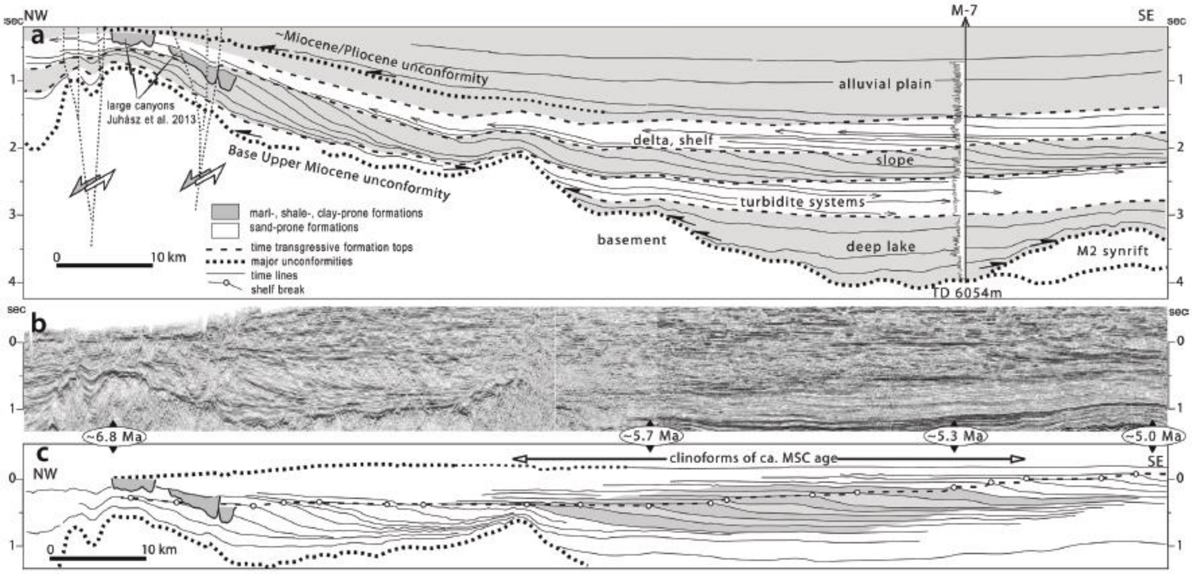


FIgure 13



a
b
c
NW
SE
sec
M-7
~Miocene/Pliocene unconformity
alluvial plain
large canyons
Juhász et al. 2013
delta, shelf
slope
Base Upper Miocene unconformity
turbidite systems
marl-, shale-, clay-prone formations
sand-prone formations
time transgressive formation tops
major unconformities
time lines
shelf break
basement
deep lake
M2 synrift
0
10 km
TD 6054m
~6.8 Ma
~5.7 Ma
~5.3 Ma
~5.0 Ma
clinoforms of ca. MSC age

Figure 14

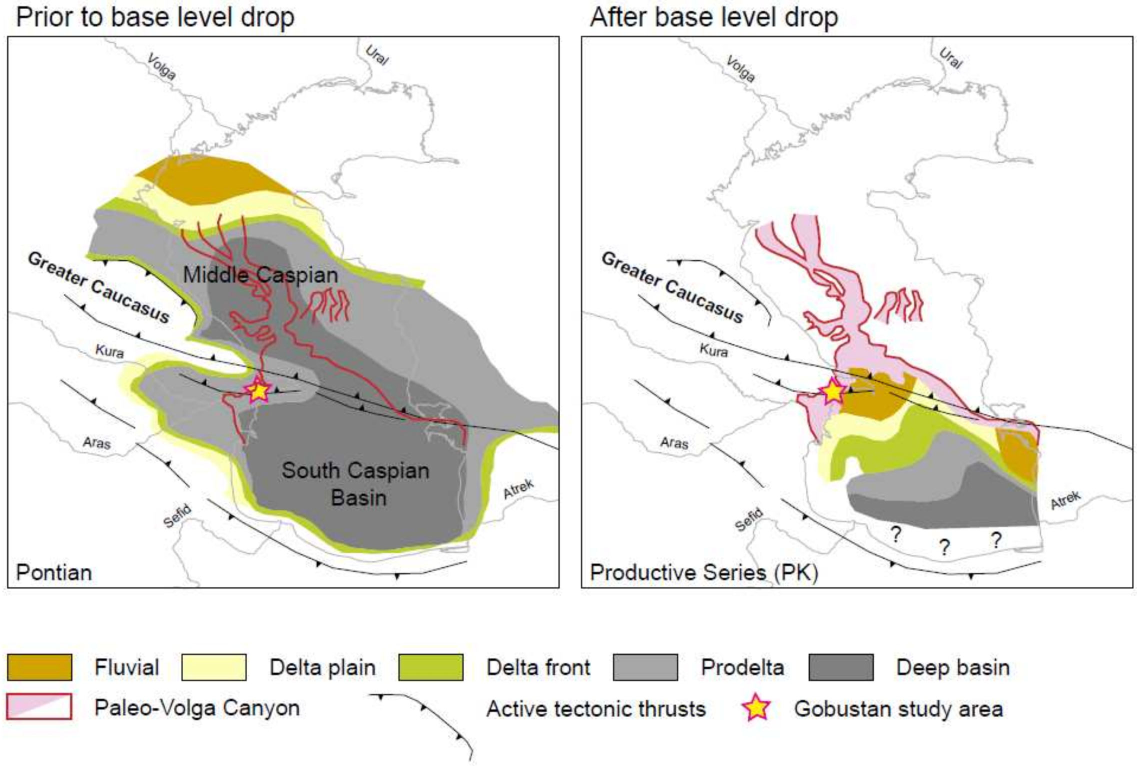
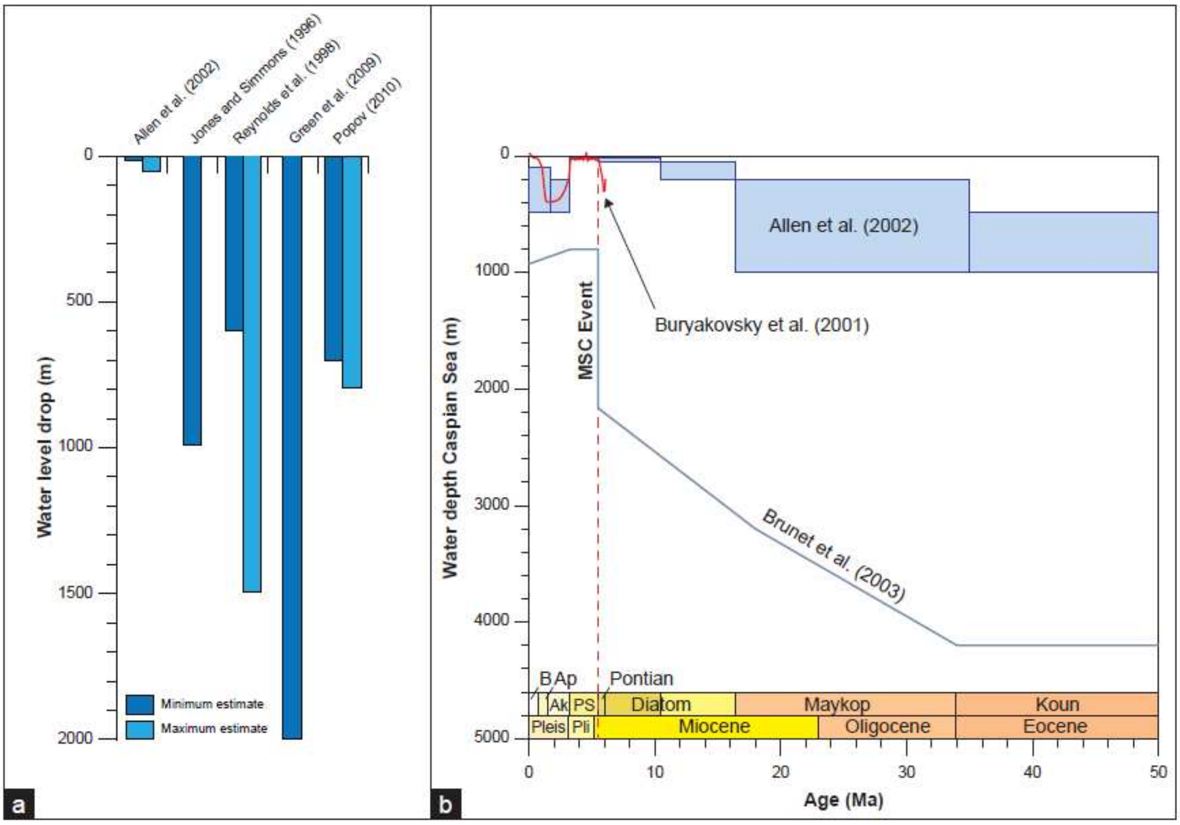

Prior to base level drop
After base level drop
Volga
Ural
Greater Caucasus
Middle Caspian
Kura
Aras
South Caspian
Basin
Sefid
Atrek
Pontian
Productive Series (PK)
? ? ?
Fluvial
Delta plain
Delta front
Prodelta
Deep basin
Paleo-Volga Canyon
Active tectonic thrusts
Gobustan study area

Figure 15



Allen et al. (2002)
Jones and Simmons (1996)
Reynolds et al. (1998)
Green et al. (2009)
Popov (2010)
Water level drop (m)
0
500
1000
1500
2000
Minimum estimate
Maximum estimate
a
Water depth Caspian Sea (m)
0
1000
2000
3000
4000
5000
Allen et al. (2002)
Buryakovsky et al. (2001)
MSC Event
Brunet et al. (2003)
B
Ap
Pontian
Ak
PS
Diatom
Maykop
Koun
Pleis
Pli
Miocene
Oligocene
Eocene
0
10
20
30
40
50
Age (Ma)
b

Figure 16

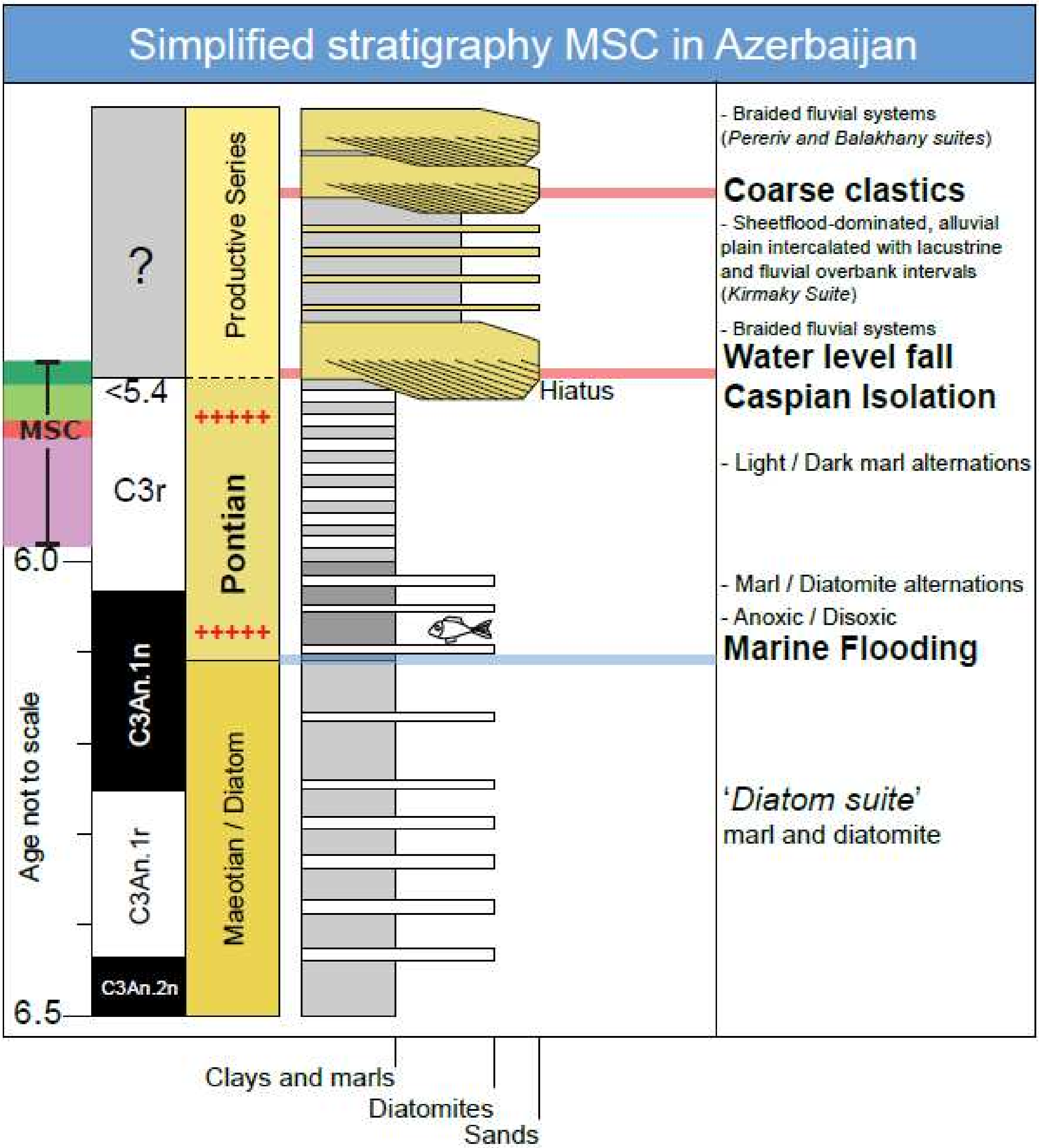
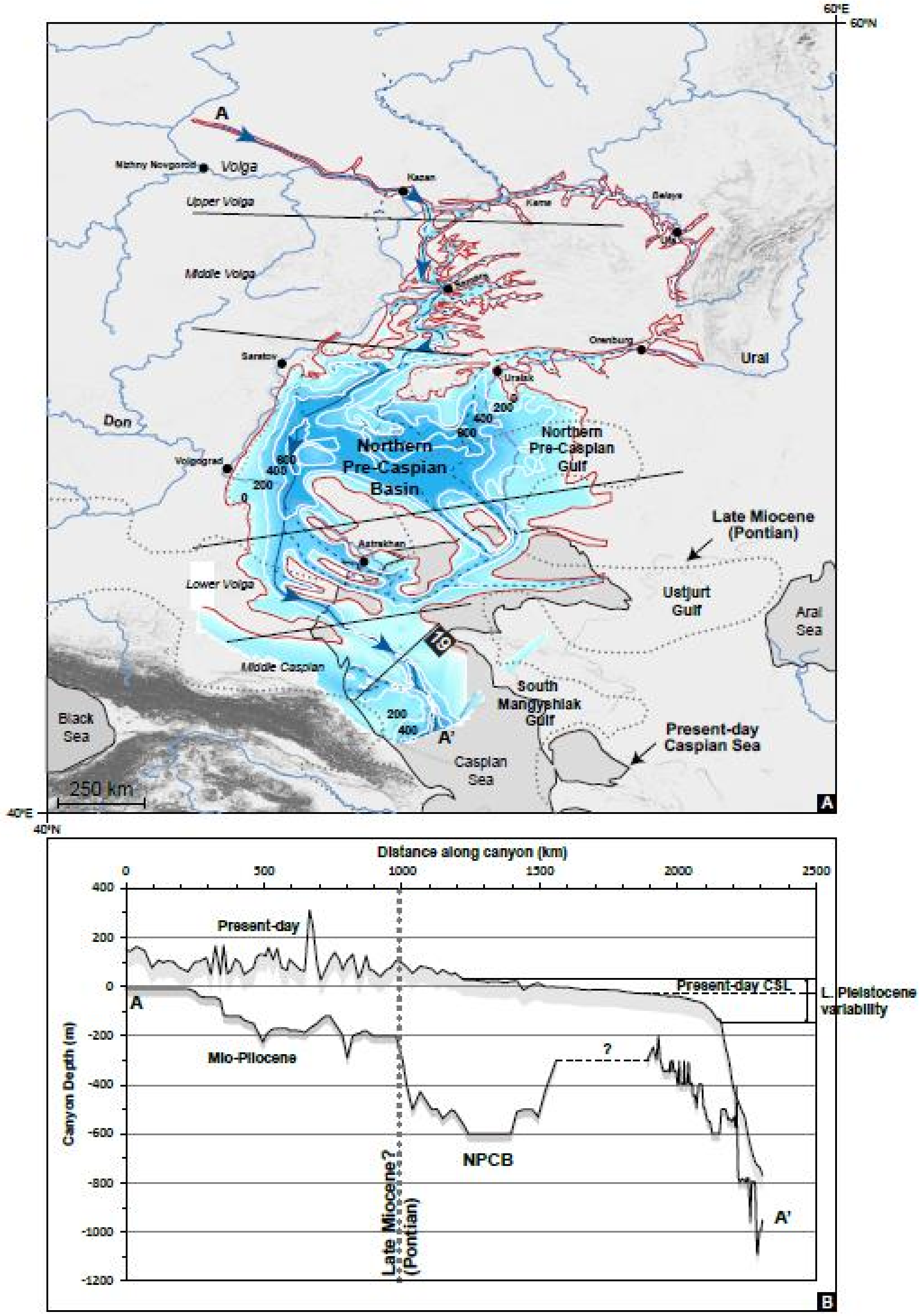

Simplified stratigraphy MSC in Azerbaijan
?
Productive Series
<5.4
MSC
C3r
Pontian
6.0
C3An.1n
C3An.1r
C3An.2n
Maeotian / Diatom
6.5
Age not to scale
Hiatus
- Braided fluvial systems
(Pereriv and Balakhany suites)
Coarse clastics
- Sheetflood-dominated, alluvial plain intercalated with lacustrine and fluvial overbank intervals
(Kirmaky Suite)
- Braided fluvial systems
Water level fall
Caspian Isolation
- Light / Dark marl alternations
- Marl / Diatomite alternations
- Anoxic / Disoxic
Marine Flooding
'Diatom suite'
marl and diatomite
Clays and marls
Diatomites
Sands

Figure 17



60°E
60°N
A
Nizhny Novgorod
Volga
Kazan
Upper Volga
Kama
Belaya
Middle Volga
Orenburg
Ural
Saratov
Uralsk
Don
200
400
600
Northern Pre-Caspian Basin
Northern Pre-Caspian Gulf
Volgograd
0
Late Miocene (Pontian)
Astrakhan
Lower Volga
Ustjurt Gulf
Aral Sea
19
Middle Caspian
South Mangyshlak Gulf
Black Sea
A'
Present-day Caspian Sea
Caspian Sea
250 km
40°E
40°N
A
Distance along canyon (km)
0
500
1000
1500
2000
2500
400
200
0
-200
-400
-600
-800
-1000
-1200
Canyon Depth (m)
Present-day
Present-day CSL
L. Pleistocene variability
Mio-Pliocene
?
NPCB
Late Miocene? (Pontian)
A'
B

Figure 18

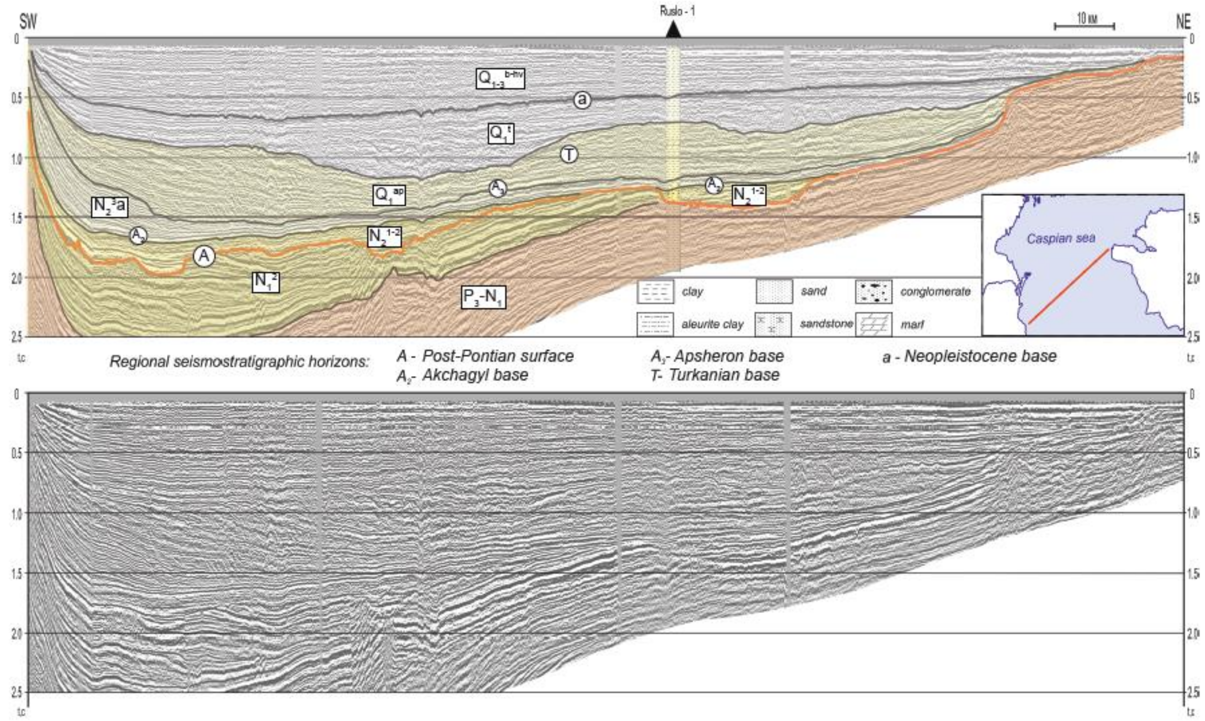

SW
NE
Ruslo - 1
10 км
Q_{1-3}^{b-hv}
a
Q_1^{t}
T
Q_1^{ap}
A_3
A_2
N_2^{1-2}
$N_2^{3}a$
A
N_1^{2}
P_3-N_1
clay
sand
conglomerate
aleurite clay
sandstone
marl
Caspian sea
Regional seismostratigraphic horizons:
A - Post-Pontian surface
A_2- Akchagyl base
A_3- Apsheron base
T- Turkanian base
a - Neopleistocene base

Figure 19